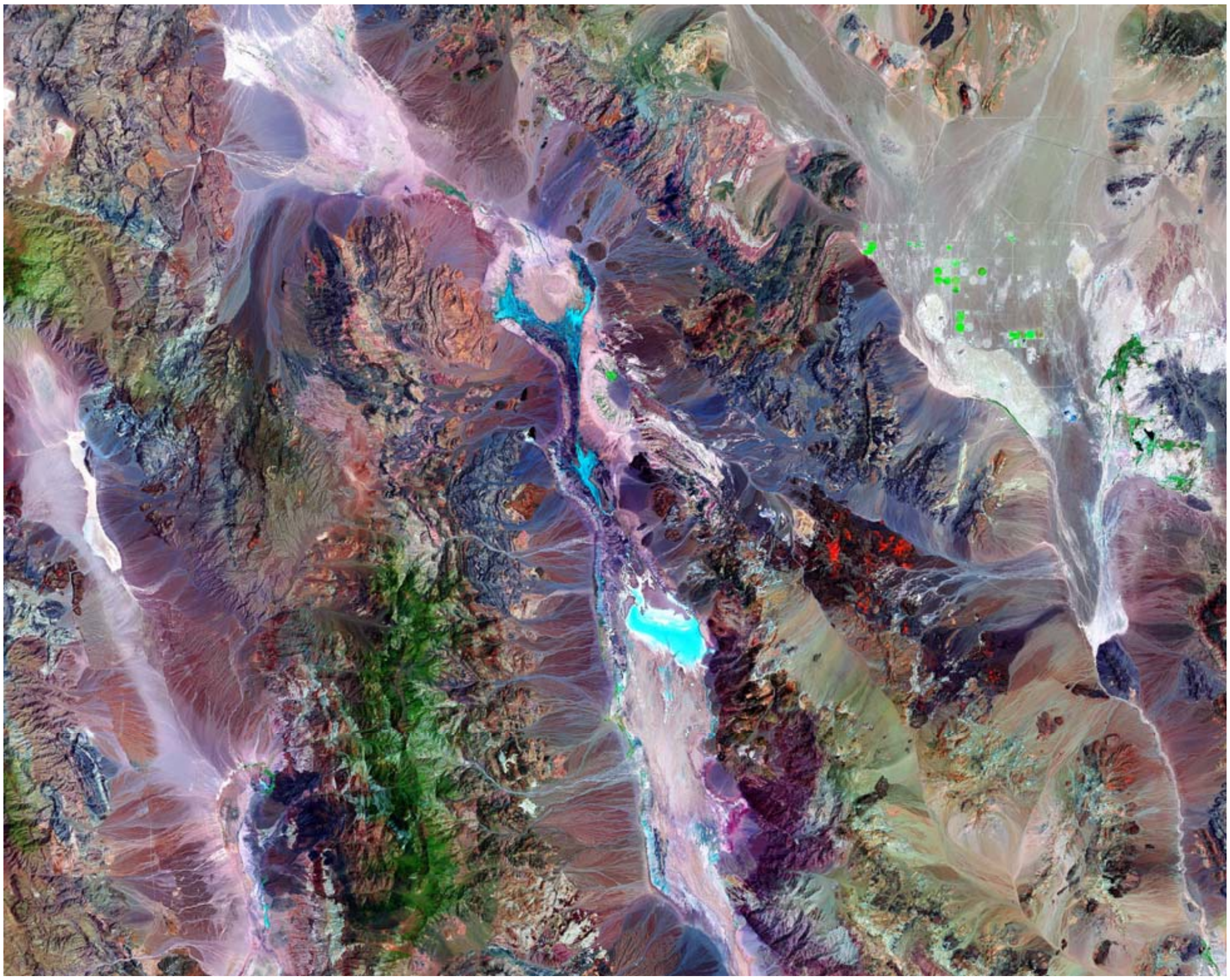
HOFSTRA UNIVERSITY
FIELD TRIP GUIDEBOOK

GEOLOGY 143D – Field Geology of California/Nevada

Spring Semester – 04-11 April 2009



Death Valley region of California and adjacent Nevada, the focus of our trip. Pleistocene glacial melt created a 600' deep freshwater lake in Death Valley that persisted into the Holocene at a depth of 30'. Today, saltpans (three to five feet thick) mark evaporitic remnant pools of the lake. *(Landsat 7 image: http://landsat.gsfc.nasa.gov/images/archive/f0007.html)*

Field Trip Led by Dr. Charles Merguerian

Hofstra University Geology 143D
Field Geology of California/Nevada, 04-11 April 2009

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Charles Merguerian

Introduction

This guidebook has been provided to discuss some of the geological background needed to enhance your experience on the Geology 143D field trip to Nevada and California including the Sierra Nevada (Snowy Range). Our itinerary will be found after some general discussion of the history, physiography, general geology and tectonics of the state of California. As I envision it, our field trip is divided into four major geological segments that will be described in depth in a later section. These regions include the **Basin and Range** and **Great Basin** of Nevada and California (Days 1, 2), **Death Valley**, California (Days 3 and 4), **Mono Craters and the Inyo "Face" Mountains** (Day 5), the **Sierra Nevada range** and **Yosemite National Park** (Days 6 and 7).

The choice of these four regions was made to best exemplify the geology of southeastern California and adjacent Nevada and to develop a transect view of the geology for students to better understand the plate tectonic assembly of the various terranes that, over geologic time, have created the region. This field course dovetails nicely with Geology 143B (Field Geology of Central California) and students interested in the terrane concept should consider taking both. This guidebook provides the essential background information you will need to fully understand the rock exposures that we will visit during our trip. Each day we will examine natural rock exposures in the field, employing and learning the methods and observations of field geologic science. Use of the guidebook in the field is essential as discussions will be based upon the maps and diagrams that have been reproduced for you, under the threat of copyright arrest and protracted intellectual torture (drive-by insults).

California harbors a rich though recent history by comparison to areas of the east coast of the United States. According to Chamberlain (1977), the conquests of Commodore Robert A. Stockton and Colonel John C. Frémont were paramount in signing over the land of California to the American Republic on 02 February 1848, thus ending Mexican rule. The Spanish and Mexican domination that lasted centuries had come to a halt that day in Hidalgo, Mexico with the slash of a pen. Many Americans feared that the land would be uninhabited. The gold rush changed all that and by 1850 nearly 100,000 people had arrived there. Physical barriers to the conquest of the gold seekers (Argonauts) came in the form of snow-covered high Sierran passes for access, steep canyons that to be crossed and controlled, and lack of food and supplies.

Stretching 800 miles in long dimension and up to 375 miles wide, California covers an area of 158,693 mi^2 and is the second largest state (Texas, of course being #1). Elevations in the Sierra Nevada range peak at 14,495' (Mt. Whitney) yet those in the Death Valley are the lowest in North America at -281.9' (below sea level, in fact, way below sea level). Now you can see why it's a **great relief** for me to be back in California! Let's learn something about the geology.

Geology of California

The Cordillera of the western United States is a 1,000-mile wide zone of mountain building that records the results of over 350 Ma of continual plate tectonic activity along the embryonic western margin of North America. Older geologic activity is not recorded in the area of our trip. Yet, Proterozoic and older rocks are exposed in southern California and in the southern Coast Range where they have been splintered and transported NW along the San Andreas fault system and form a component of the Salinian Block to the south of Monterey. These older rocks, such as those found to occupy the inner gorge of the Grand Canyon, are associated with NE-trending Proterozoic and older Archean rocks exposed in Arizona and adjacent areas (Figure 1). The Cordillera (Figure 2) consists of a vast zone of imbricated and in some cases far-traveled terranes, accrued from continual Paleozoic, Mesozoic, and Cenozoic marginal plate activity. It would appear that there was always some form of active subduction zone, volcanic arc(s), shear zone, or brittle faulting taking place across California throughout that interval. The active mobile belts of the Indonesian region come to mind when attempting to derive an analog to the paleogeography of the active California margin (Hamilton 1978).

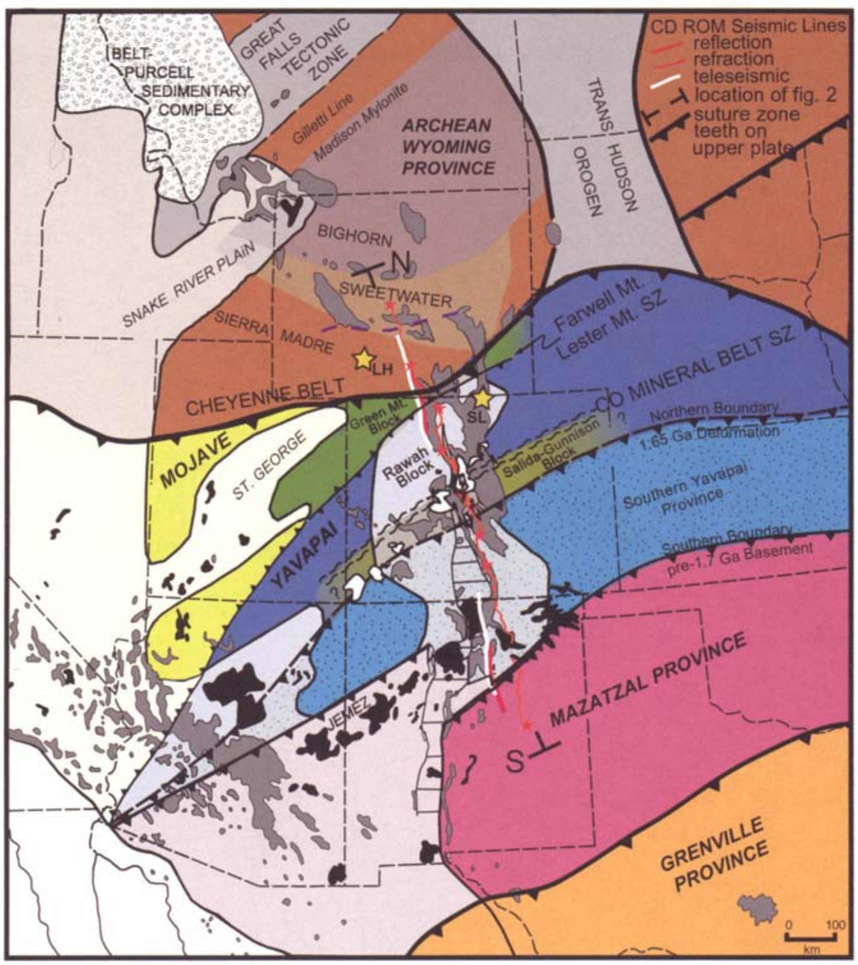


Figure 1 – Terrane map of the SW Cordillera showing the convergence of the Mohave, Yavapai, and Mazatzal provinces and their extension into southern California. (From CD-ROM Working Group, 2002.)

Figure 2 – Index map of the western United States showing the broad area underlain by the Cordilleran mountain range. Elements of this 1,000-mile wide mountainous region extend from California eastward to New Mexico, Colorado, Wyoming, and Idaho.

Figure 3 – Shaded relief map of the same area as Figure 1 showing the topography and major drainage systems. Note that California consists of two NNW-trending mountainous regions, the Coast Range on the west and Sierra Nevada Range on the east. The ranges, key targets of our Geology 143B trip, are geologically different and separated by a broad sediment-filled flat valley known as the Great Valley.

A shaded relief map of the Cordillera (Figure 3) shows a rugged juvenile terrain dissected by major drainages that flow predominately toward the Pacific Ocean. Indeed, the Continental Divide is located near the eastern edge of the Cordillera, extending through western Wyoming near Yellowstone Park southward through Colorado, and New Mexico. We have visited Yellowstone National Park (home of Yogi Bear), and the Grand Tetons (Geology 143C). Water falling to the east of the Continental Divide flows ultimately into the Atlantic Ocean. Note the broad sediment fans (Figure 3) that cover the boundary between the Rocky Mountain Division and the Piedmont and High Plains of the central United States. Physiographic subdivisions of the Cordillera are shown in Figure 4. Our Geology 143A field trip to the Intermontane Division of the Cordillera through northern Arizona where we studied the Grand Canyon of the Colorado River Plateau. This year's trip to Nevada and California will concentrate on the Basin and Range province westward from Las Vegas, Nevada into California, the influence of the San Andreas fault system, and the Sierra Nevada Range of California. Time permitting we will take a peek at some of the Paleozoic rocks of the Sierran foothills, west of Yosemite.

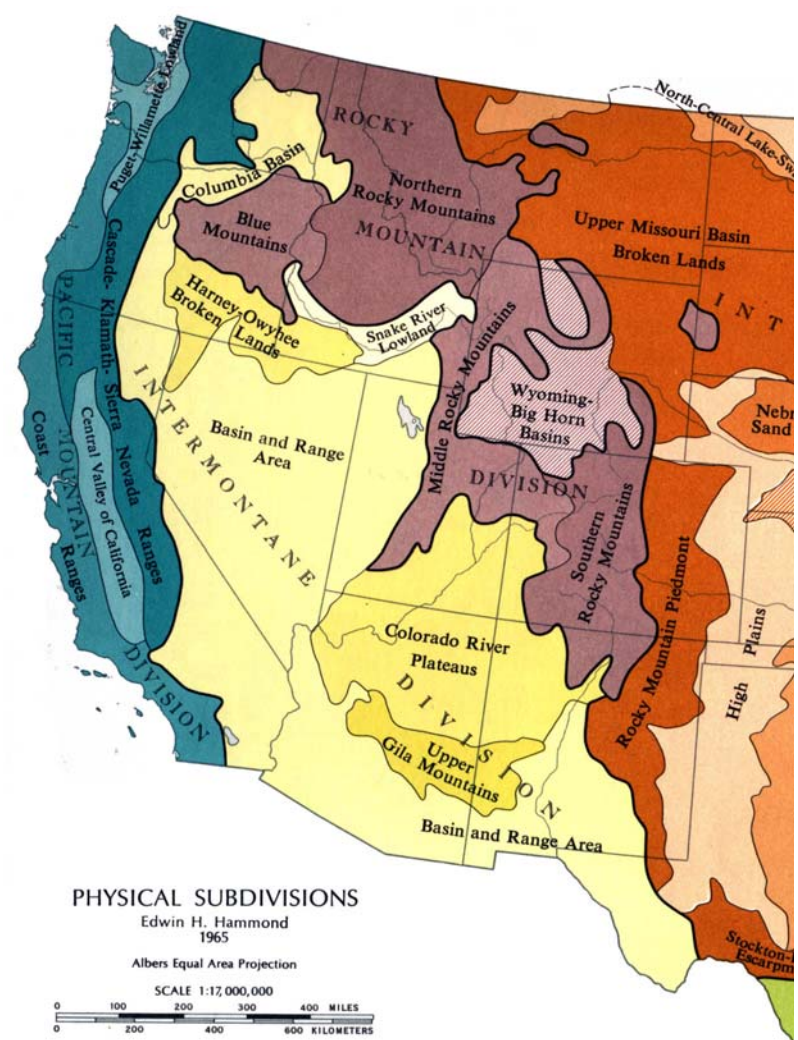


Figure 4 – Physical subdivisions of the Cordilleran belt of western North America (Hammond, 1965).

Figure 5 shows the three major NNW-trending physiographic elements of California. A physiographic diagram of California for help with place names by famed artist Erwin Raisz is reproduced as Figure 6.

The Coast Ranges

The Coast Ranges constitute a series of subdued, elongate hills consisting of highly sheared, faulted, and disarticulated Late Mesozoic to Early Tertiary oceanic rocks (mostly slate) of the Franciscan Complex. The Franciscan includes exotic blocks of lithic sandstone, chert, and eclogite (recycled mantle rocks) and show evidence for blueschist metamorphism. To the east, the Coast Ranges are in thrust fault contact with dominantly flat-lying marine and intercalated volcaniclastic strata of the Great Valley. The Great Valley of California, land of fruits and nuts, was buried by sediment during westward tilt of the Sierra Nevada block that began in earnest in Pliocene time (~2 Ma). Rapid downcutting of westward-flowing mountain streams produced steep v-shaped valleys that severely dissected the Sierran and the Foothills metamorphic terranes (Figure 7). Abundant sediment, some of it auriferous, was deposited above older volcaniclastic strata of the Great Valley, a former fore-arc basin. The Sacramento and San Joachin rivers flow from the tilted Sierra block westward across the Great Valley where they empty into San Francisco Bay, a drowned composite river mouth. A simplified geological section of the Coast Ranges is shown as Figure 8.

Figure 5 – Physiographic relief map of California showing, from west to east, the Coast Ranges, the Great Valley and the Sierra Nevada Range.

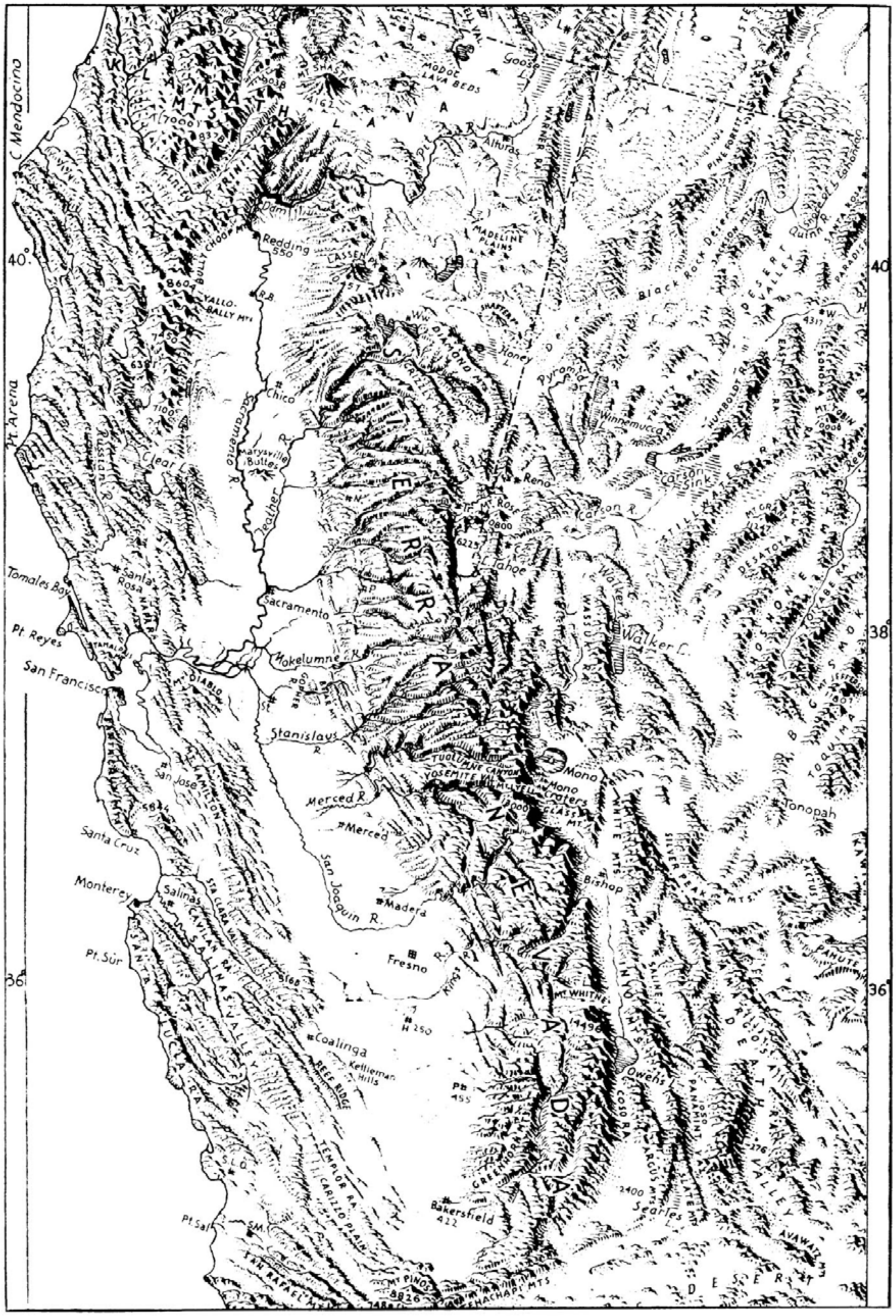


Figure 6 – Physiographic map of California by Erwin Raisz.

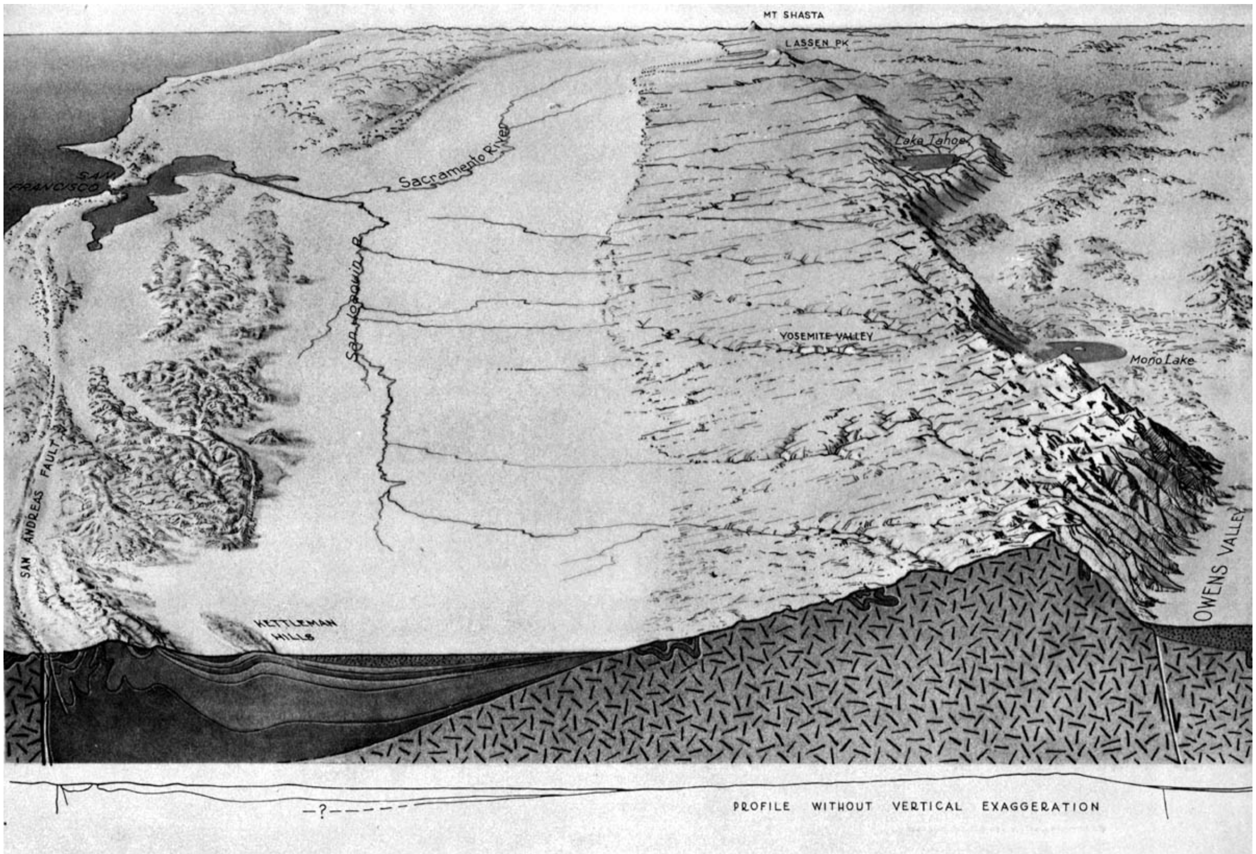


Figure 7 – Drawing from Shelton (1966) showing the westward tilt of the Sierra Nevada block that began in earnest (no, not Ernest P. Worrell) in Pliocene time (~2 Ma). Rapid downcutting of westward-flowing streams produced steep v-shaped valleys that severely dissected the Sierra and Foothills metamorphic belts.

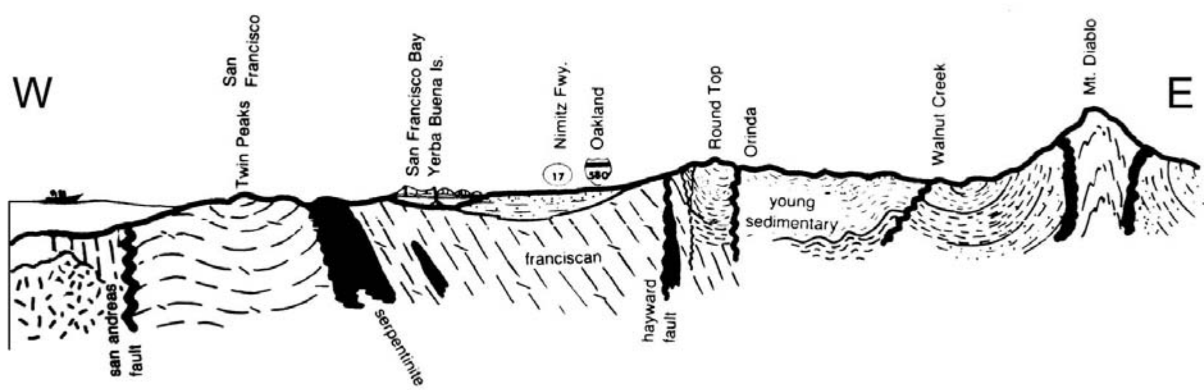


Figure 8 – Geologic section across the eastern part of the Coast Range at the latitude of our trip routes showing the effects of the San Andreas fault zone and internal structure of the Franciscan Complex. (From Alt and Hyndman, 1975, p. 27.)

Great Valley Sequence

Along its western boundary, the Great Valley Sequence is highly deformed and overthrust above the Franciscan of the Coast Range along a zone of broken serpentinite that delineates the Coast Range thrust (Figure 9). To the east, the Great Valley strata unconformably overlie the Sierra Nevada Range – a complex metamorphic and igneous complex of variable age, lithology, and former tectonic setting. The former tectonic setting of the Great Valley was that of a forearc basin (marked v in Figure 10), situated between the trench complex on the west (Franciscan Complex) and the volcanic arc (now eroded away from the uplifted Sierran block except for a few remnant stringers of lava and deposits of lahar and ash that survived the erosion and weathering dilemma). As discussed earlier, the younger strata of the Great Valley are derived from weathering and erosion of the arc complex and its apron that were formerly situated above the present Sierra Nevada, the eroded plutonic roots of that arc.

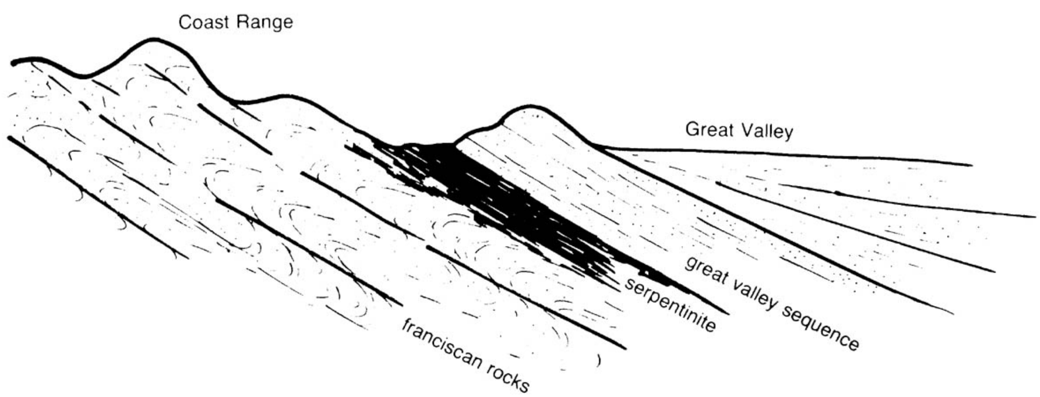


Figure 9 – Cross section showing the Coast Range thrust fault, a major fault that separates the Coast Ranges on the west from the Great Valley Sequence on the east.

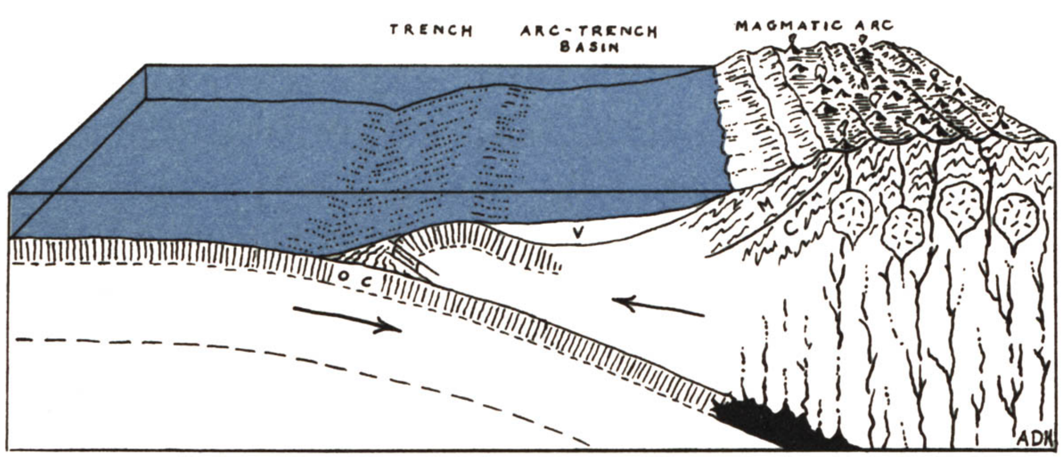


Figure 10 – Block diagram showing a typical Andean arc-trench basin, an analog for the Coast Ranges, Great Valley, and Sierra Nevada of California. Landward of the trench a broad zone of highly sheared oceanic strata or

outer non-volcanic high gives way to predominately volcaniclastic strata of the Great Valley. Note that the forearc basin strata overlap the arc basement to the right of the diagram. (From Howard, 1979, Fig. 7, p. 29.)

Foothills Metamorphic Belt

The western metamorphic belt of the Sierra Nevada range (Clark, 1964, 1976) is roughly 350 km long, 60 km wide and extends from lat. 40°15'N south-southeasterly to lat. 37°N. It is overlain by Cretaceous to Cenozoic rocks of the Great Valley sequence to the west and is intruded by the Sierra Nevada batholith to the east (Figure 11). South of lat. 39°N it is composed of three ductile-fault bounded tectonostratigraphic units known as the Jurassic belt, the Calaveras Complex, and the Shoo Fly Complex. These contiguous belts show an eastward increase in age, metamorphic grade, and structural complexity with abrupt lithologic, structural, and metamorphic truncations occurring at the Melones, Sonora, and Calaveras-Shoo Fly faults (Merguerian 1985a,b; Schweickert, Merguerian and Bogen, 1988).

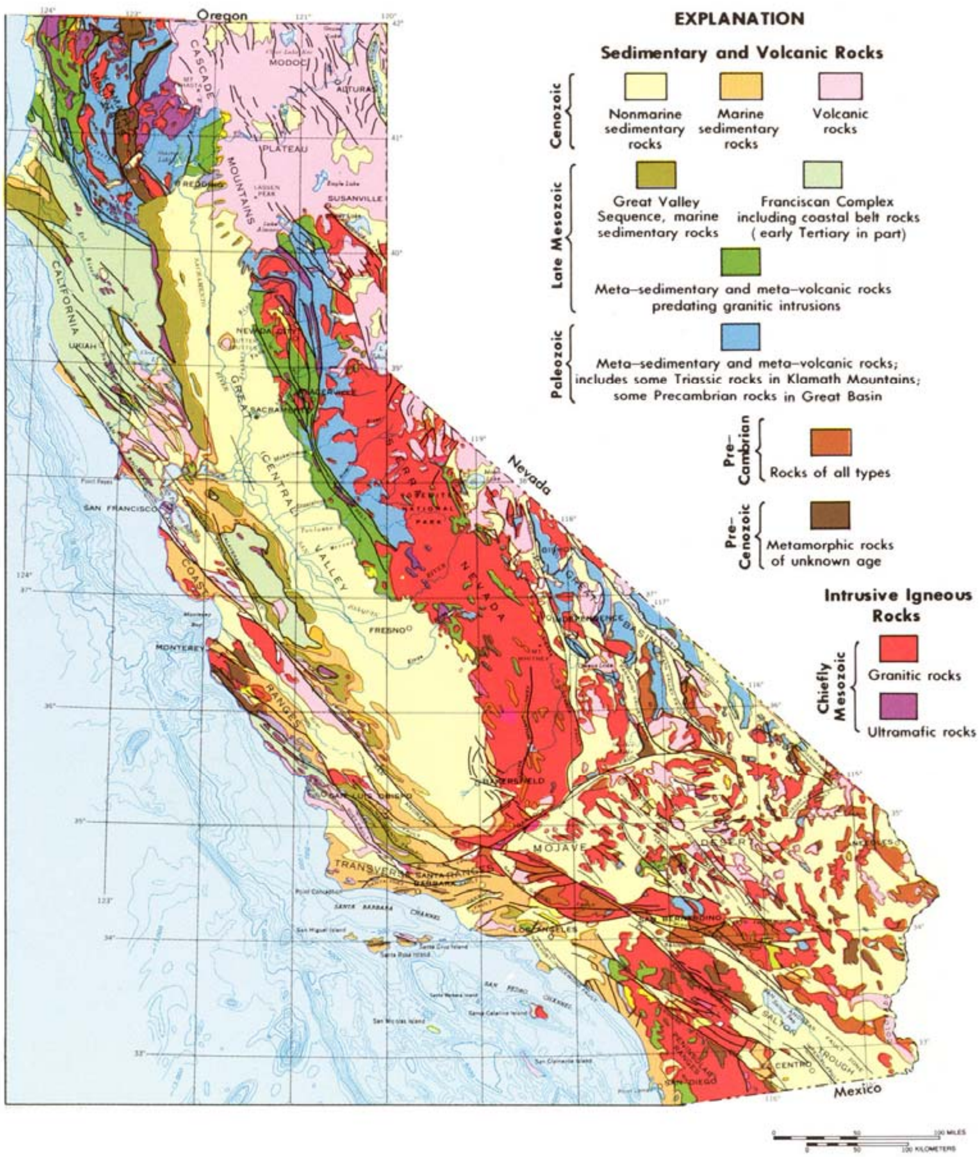


Figure 11 – Simplified geological map of California. (California Division of Mines and Geology, Centennial Map, 1980.)

The lower Paleozoic **Shoo Fly Complex** of Unit 1 underlies a terrane 330 km in length and 6-20 km wide which terminates southward near lat. 37°30'N in Mariposa County (Merguerian, 1981) where batholithic rocks cut across trend (Figure 12). Along its strike the Shoo Fly Complex shows a marked southward increase in structural complexity and metamorphic grade. North of lat. 39°N it consists of weakly metamorphosed quartzose sandstone, graywacke, slate, chert and limestone (Clark, 1976). South of lat. 38°05'N (Figure 13) it is a multiply deformed (seven superposed phases of deformation) and sheared assemblage of quartzite, quartzofeldspathic gneiss, granite-, syenite-,and gabbroic augen gneiss, garnet schist, calc-silicate rock, marble, and rare amphibolite (Merguerian, 1985a, b).

The Shoo Fly is in ductile fault contact with the Permo-Triassic **Calaveras Complex** (Unit 2) along a 1-2 km wide zone of mylonite, intense flattening, imbricated rock units and transposition with overprinting metamorphism of older fabric elements (Merguerian, 1983). The Calaveras, which forms the lower plate of' the east-dipping thrust, is a chaotic assemblage of massive argillite and siltstone, marble, massive and rhythmicallybedded-chert, talc-schist, basalt, and rare sandstone layers. The age of the Calaveras remains uncertain, but Permo-Carboniferous fossils from limestone olistoliths (Turner, 1894; Schweickert and others, 1977) indicate a minimum late Paleozoic age for the sequence. Descriptions of the Calaveras in this region can be found in Schweickert and Wright (1975), and Wright and Schweickert (1977), and Schweickert and others (1977).

Dominated by Jurassic island arc volcanic and volcaniclastic strata and deep-water flysch-type sediment, the **Jurassic Belt** (Unit 3) has been well studied. Cropping out east of the Great Valley where flat-lying strata predominate, tilted and cleaved metasedimentary and metavolcanic rocks of Jurassic age form an impressive former volcanic arc terrane now exposed in dissected low rolling hills that reflect the trend of the lithologic units and bounding faults (Figure 12). These rocks were isoclinally folded and cleaved under chlorite to greenschist grade metamorphic conditions in contrast to the blueschist metamorphism found in the Coast Range (Figure 13).

Cut by major faults in the vicinity of Jamestown, where slivers of serpentinite exhibit shearing and multiple offsets, the Jurassic Belt is in fault contact with deformed Permo-Triassic marble and chaotic argillite of the Calaveras Formation along the Melones or Sonora faults (Figure 12). The Calaveras, in turn, is in ductile fault contact with the lower Paleozoic Shoo Fly Complex along the Calaveras-Shoo Fly thrust (Figure 13). Thus, in cross sectional view the foothills metamorphic belt consists of a sequence of fault steeply dipping fault bounded terranes of increasing age, metamorphic grade, and structural complexity from west to east (Figure 14).

In the period 1978-1980, CM was a part of a huge NSF-funded research effort to better understand the intricacies and structural sequence across the Foothills Metamorphic belt in the vicinity of Sonora, California. In the summer of 1981, CM mapped at 1:24,000 scale, ten 7.5-minute quadrangles south of the Sonora-Twain Harte area for the California Division of Mines and Geology (Merguerian, 1981). Along with graduate advisor Dr. Richard A. Schweickert and fellow Columbia University graduate student Nicholas Bogen, focused study and integration of results from the Shoo Fly, Calaveras, and the Jurassic belt along the same general latitude resulted in new geological data and interpretation. The combined results of those studies can be found in papers by Bogen (1983, 1984a, b), Merguerian (1981, 1985a, b), Schweickert and others (1984), Merguerian and Schweickert (1987), Schweickert and Bogen (1983), and Schweickert, Merguerian, and Bogen (1988) and in new maps of the region published by the California Division of Mines and Geology (Wagner and others, 1981, 1990). During our field trip we will study the various components of the Foothills metamorphic belt and examine the terrane bounding fault zones and try to find some mistakes that CM made the first time.

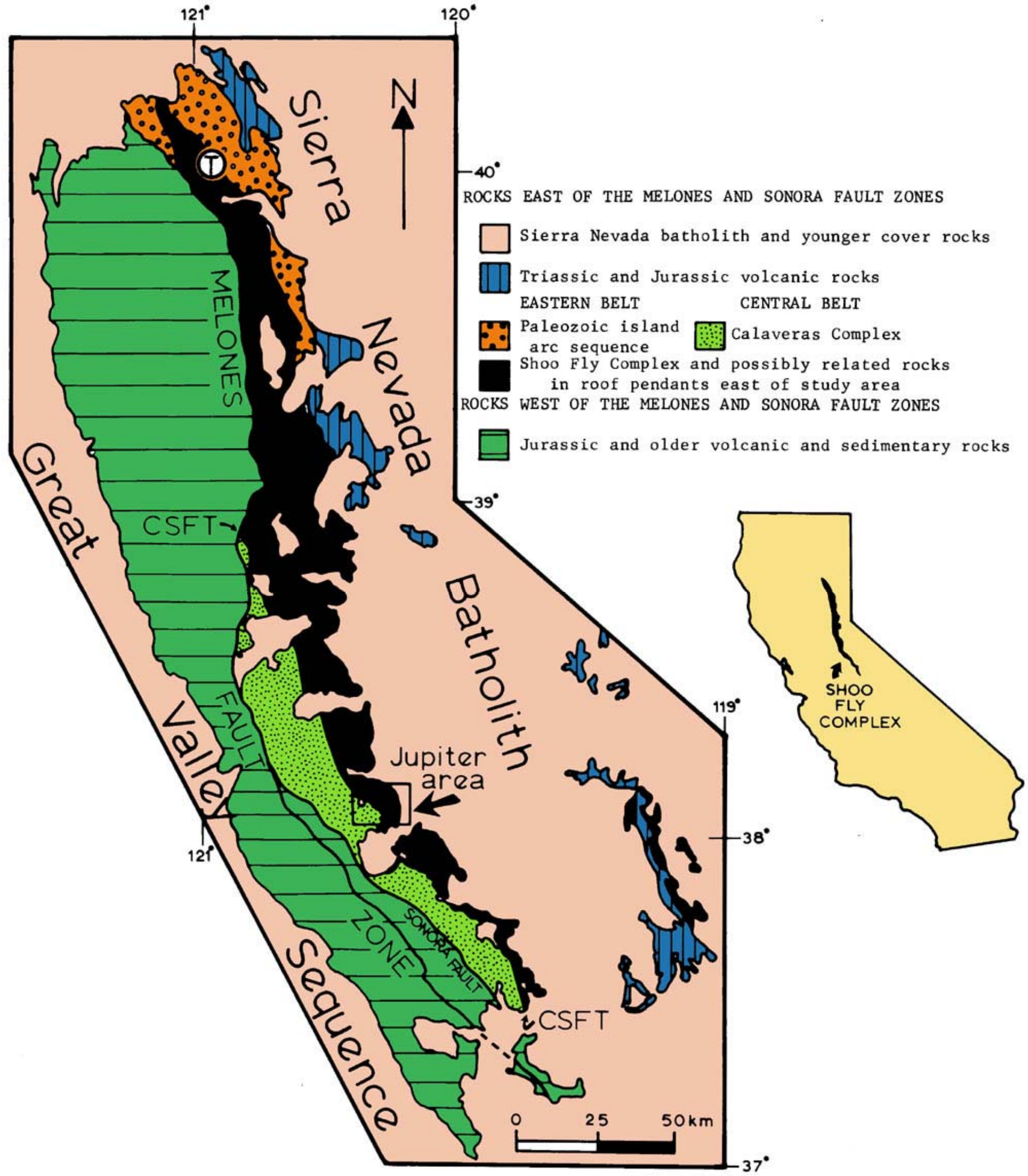


Figure 12 – Geologic map showing the components of the Foothills Metamorphic belt of the Sierra Nevada. The area underlain by the Lower Paleozoic Shoo Fly Complex is shown in black on map and inset. (From Merguerian, 1985.)

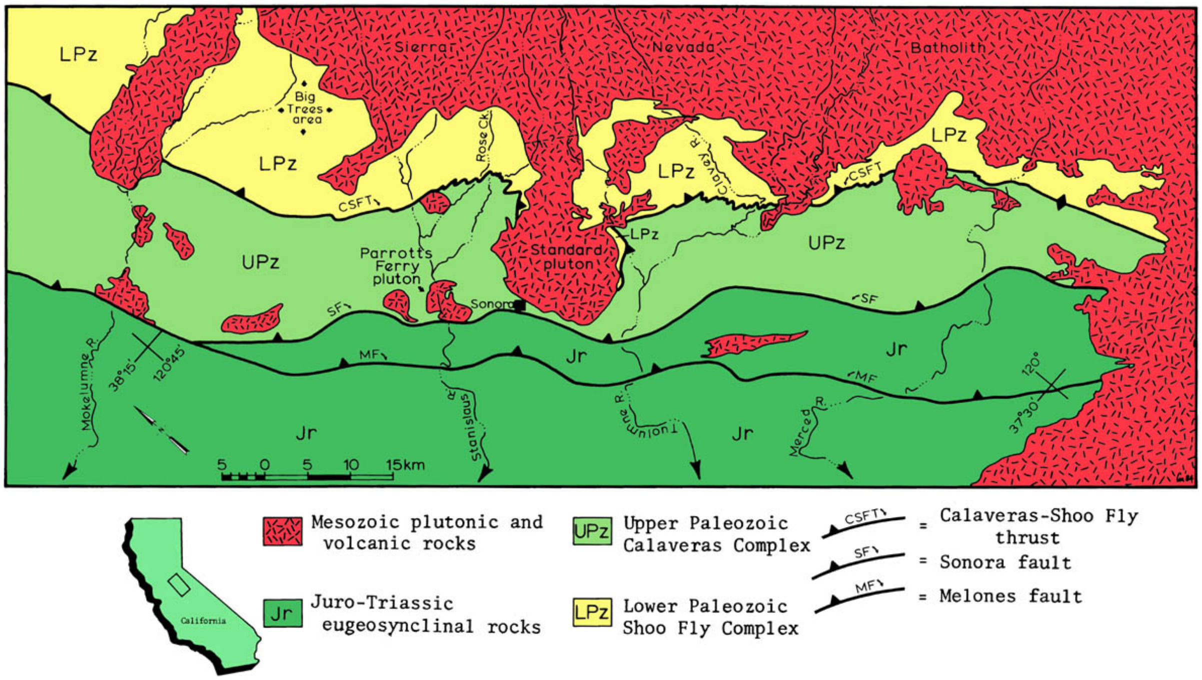


Figure 13 – Geologic map of the southern part of the Foothills Metamorphic belt of the Sieerra Nevada showing the Melones and Sonora faults and the Calaveras-Shoo Fly thrust. These major tectonic boundaries separate Juro-Triassic arc strata (Unit 3), from the Calaveras Complex (Unit 2), from the older Shoo Fly Complex (Unit 1). (From Merguerian, 1985.)

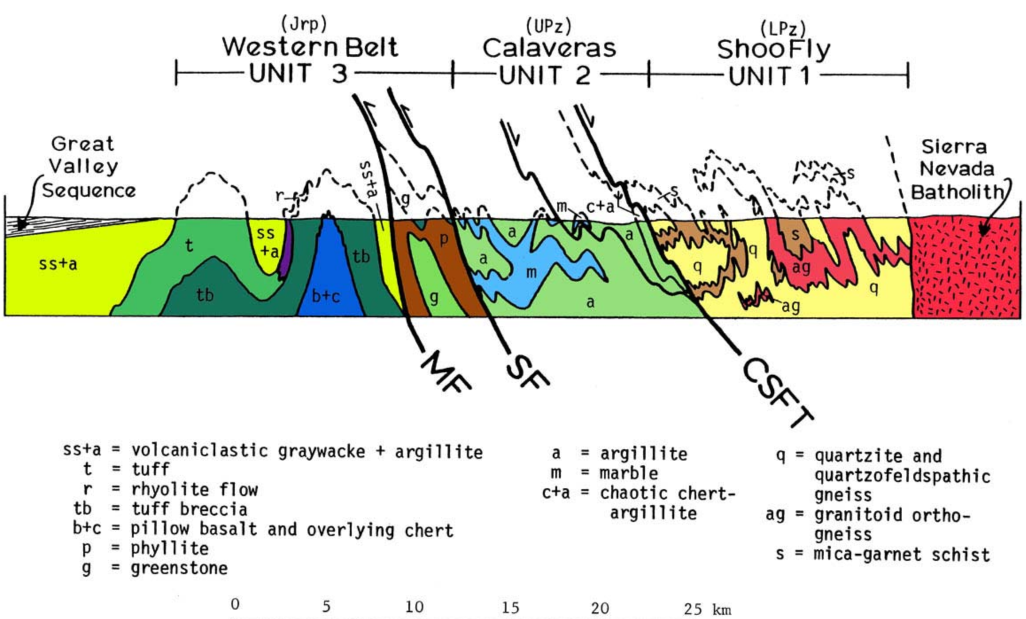


Figure 14 – Geologic section of the foothills metamorphic belt showing the across strike internal structure and stratigraphy. (From Merguerian, 1985.)

Phanerozoic Tectonic History of California

The modern plate tectonic setting of California is depicted in Figure 15 in the glorious physiographic map produced by Tharp (1969). In this view the cause for the broad zone of Cordilleran deformation is clearly evident. Subduction of most of the east half of the Pacific Ocean plate along the California margin has created it all. Today, with the development of a triple junction in Baja California, the subduction has converted to strike-slip motion along most of the western margin of North America and certainly along the south half of California. Basin and Range extension concentrated in areas of eastern California, Arizona, and Nevada and extending all the way to west Texas (the Rio Grande Rift) are the extensional lithospheric response to subduction of the spreading East Pacific Rise. To the north of the Mendicino fracture zone, Andean subduction is responsible for the Cascade volcanic district which includes Mt. Shasta and Lassen in California.

Figure 15 – Physiographic relief map of the Pacific ocean showing the modern plate tectonic configuration of North America. Note the obvious subduction of the East Pacific Rise beneath the Rocky Mountain range of the western Cordillera. In addition note that the rise separates Baja California from the mainland – a place where the east Pacific

Rise changes from an extensional boundary to a transform fault known as the san Andsreas Fault. (From Tharp, 1969.)

All of the elements of the modern tectonic setting of California have occurred in the past. Indeed the assembly of California requires a rich history of continuous active margin plate tectonic activity since the early Paleozoic Era. To best explain the ins and outs of our modern ideas on the tectonics of California, I will explain things geologically – that is, from the bottom up. So, bottoms up! Here we go.

In the blink of a geological eye we draw your attention to the time scale below (Figure 16) that is specific to California. The time scale indicates the era, period, and epoch subdivisions. The geologic development of California spans at least a half a billion years (0.5 Ga) and elements of great age are caught up in the highly faulted terranes of southern California where they have been slivered off and transported northward. (See Figure 11.) To simplify a complex story, I have taken the liberty of subdividing the geologic history of California into major Phanerozoic orogenies. In much the same way that a layer cake is made, we start with the older events and move upwards in time and stratigraphy to the present. The details of the ingredients of each layer are described individually below where many stratigraphic charts and tables are provided for use in the field.

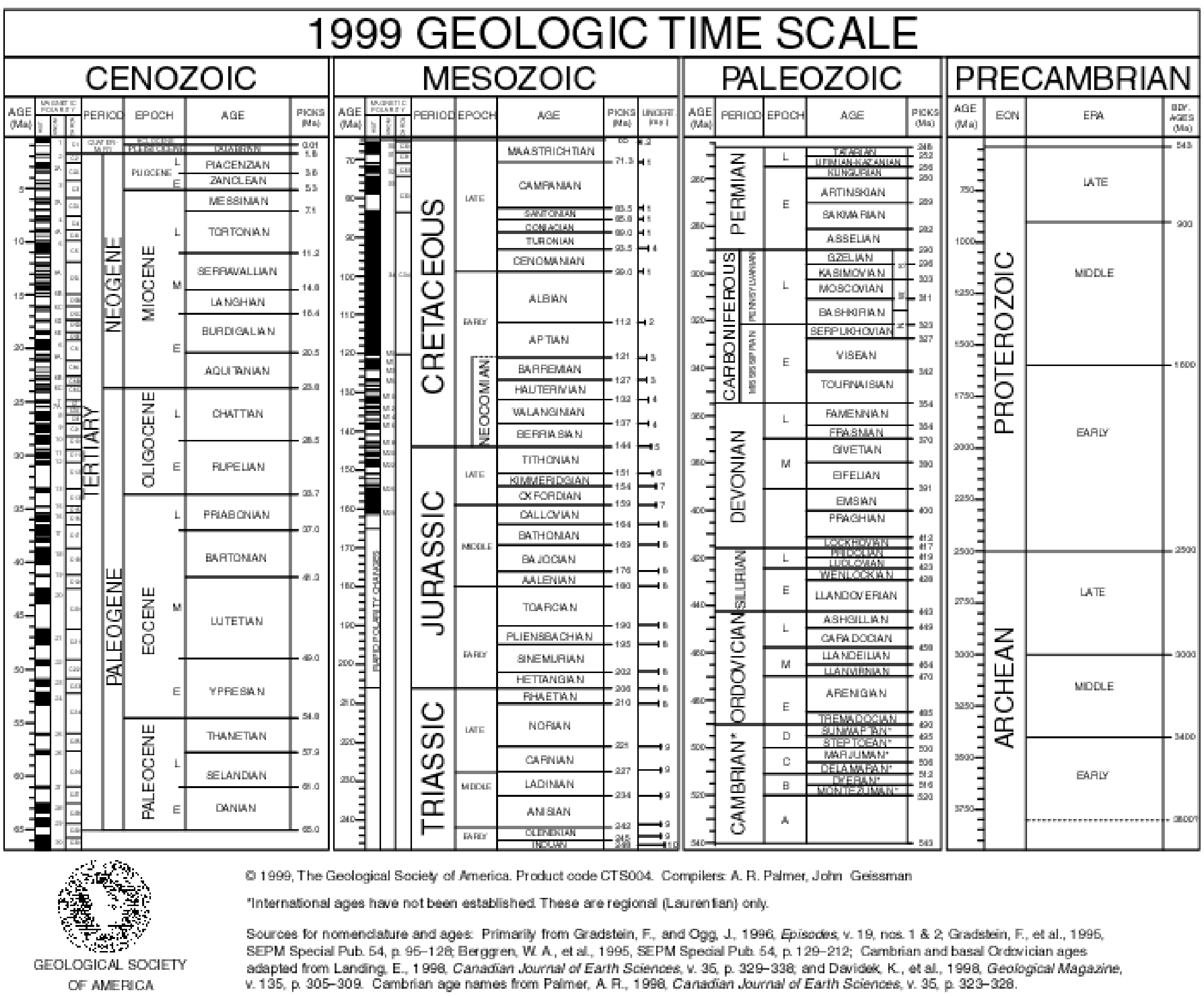


Figure 16 – Geological timescale. Geological Society of America, 1999.

You've heard of Tora! Tora! Tora! Well, Subduction!, Subduction!, Subduction! That would be the easiest way to describe the Phanerozoic tectonics of the western Cordillera. Throughout the early Paleozoic, development of an open-ocean passive margin formed an extensive open ocean miogeosynclinal and deep-water eugeosynclinal couple formed adjacent to embryonic western North America (Figure 17). Although facing in the opposite direction, the plate tectonic setting of the Cordillera was a mirror image of the developing Appalachian passive margin throughout the early Paleozoic.

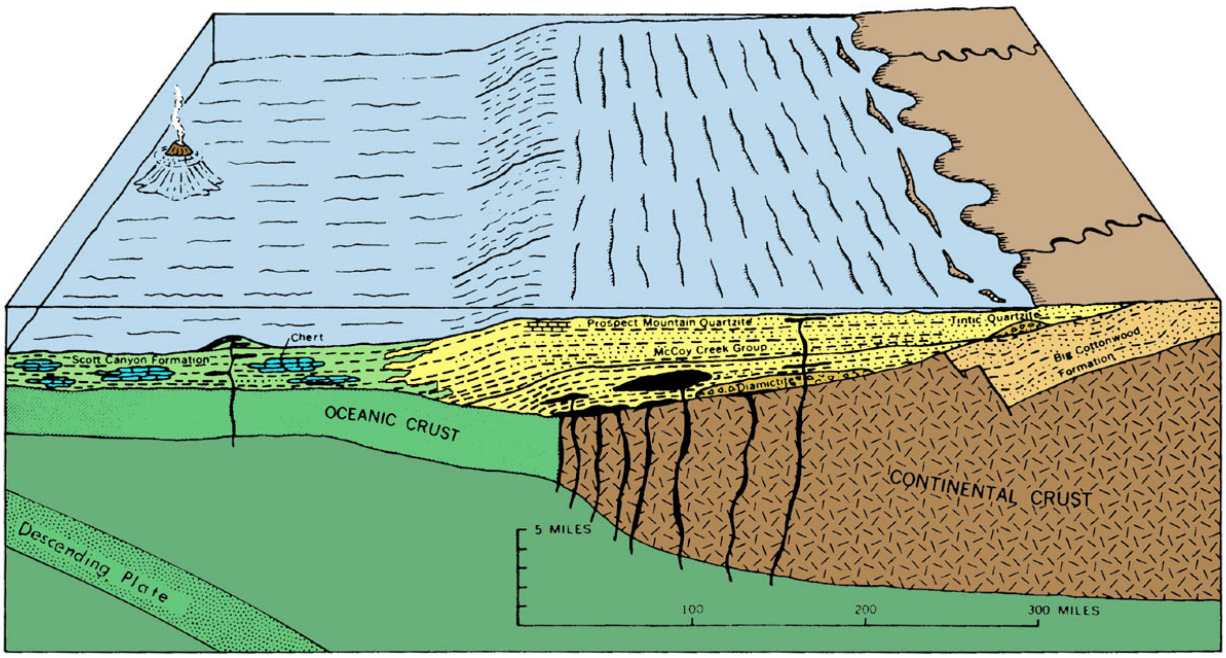


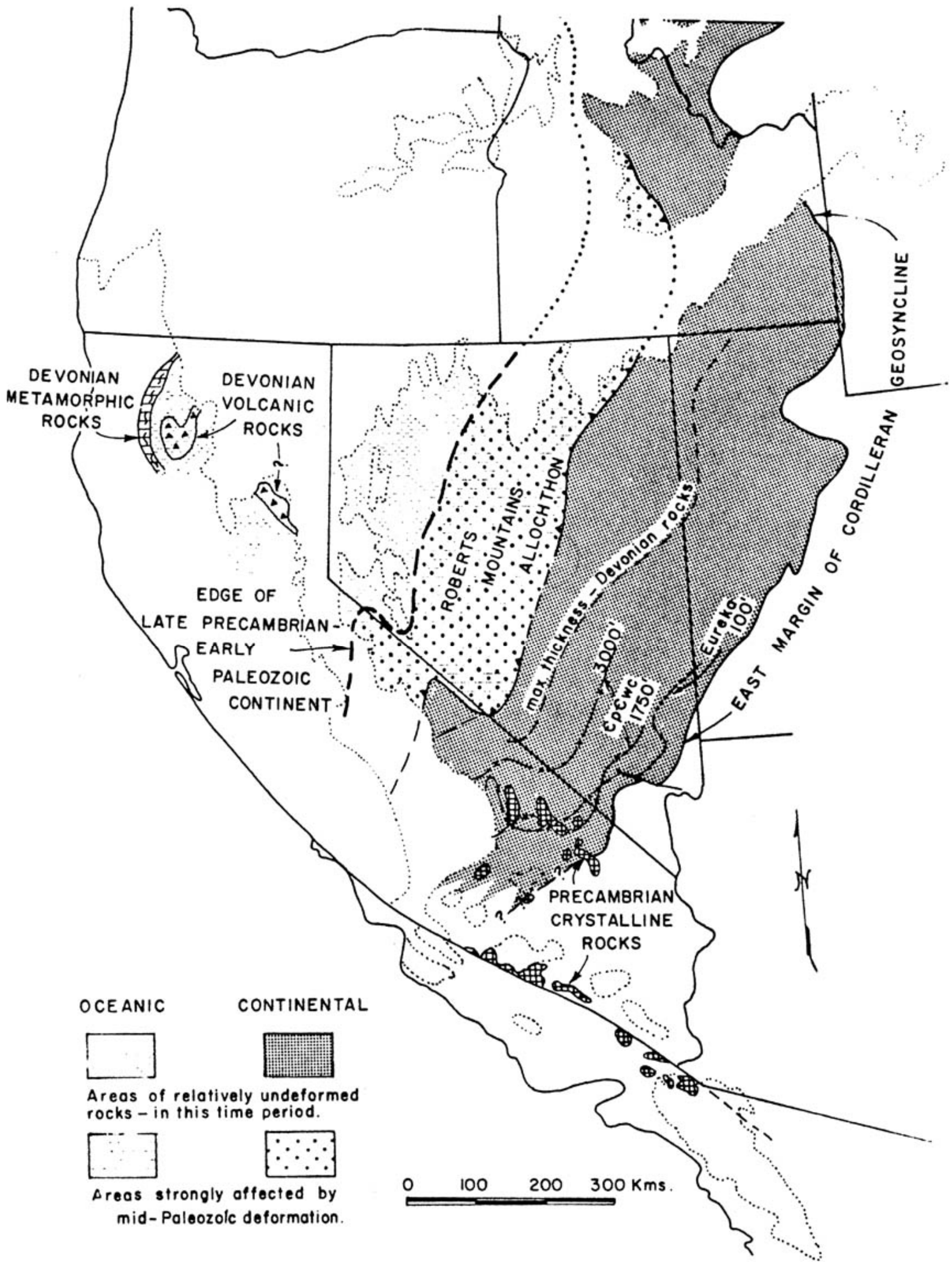
Figure 17 – Pre-Antler open ocean view of the passive Cordilleran margin during early Paleozoic time. The eventual position of California is somewhere near the middle of this diagram.

Pre-Antler

Roughly 550 Ma, with formation of an open ocean to the west, Cambrian clastics spread out over the edge of North America under shallow water conditions. Directly analogous to the transgressive lower Paleozoic facies of the Appalachian belt, with rising sea level, subsiding continental edge, or both, the Cambrian shoreline migrated eastward to sweep across areas as far as Arizona. As time passed, the offshore depositional environments shifted continentward along with the Cambrian trangression of the seas. This resulted in a fining upward sequence of strata where Cambrian sands (Tapeats Sandstone) were overlain by shale (Bright Angel Shale), then by carbonate (Muav Limestone) of the Tonto Group. A regionally impressive Cambrian to early Ordovician transgression is also recorded in the Appalachian belt.

Antler Orogeny

Starting with the **Antler orogeny** in late Devonian – early Mississipian time, a series of convergent margin events affected the western Cordillera. An arc-continent collision was responsible for the Antler orogeny (Figure 18). The Roberts Mountain Allochthon places deep-water facies atop coeval miogeoclinal rocks through central Nevada along east-directed overthrusts. Such overthrusts were produced within the walls of an accretionary prism that collided with the passive margin of North America. Seek geologic timescale (Table 2). In the area of our trip, the Shoo Fly Complex records the effects of Antler orogenesis superimposed on the lower Paleozoic passive margin sequence.



LATE PRECAMBRIAN TO DEVONIAN

Figure 18 – Map showing Late Proterozoic to Devonian isopachs and position of Roberts Mountain allochthon.

Sonoman Orogeny

After a period of uplift, erosion and extension, a marginal basin formed that filled with Pennsylvanian and Permian sediment. A piece of the late Devonian Antler volcanic arc that had already collided rifted away from the suture zone, leaving a small oceanic “marginal” basin behind. Closure of that marginal basin along a west-dipping subduction zone resulted in a collision with the old Antler arc that had rifted away earlier. The **Sonoman orogeny** (late Permian - early Triassic) produced a collision that resulted from the composite arc that formed in the upper plate of the new subduction zone. Figure 19 shows, in green, the trend and position of both the Antler (Roberts Mountain) and Sonoman (Golconda allochthon) belts in Nevada and their presumed extension into California. Note the high-angle truncation of the older Antler and Sonoman tectonic trend with the late Triassic and younger NW trend.

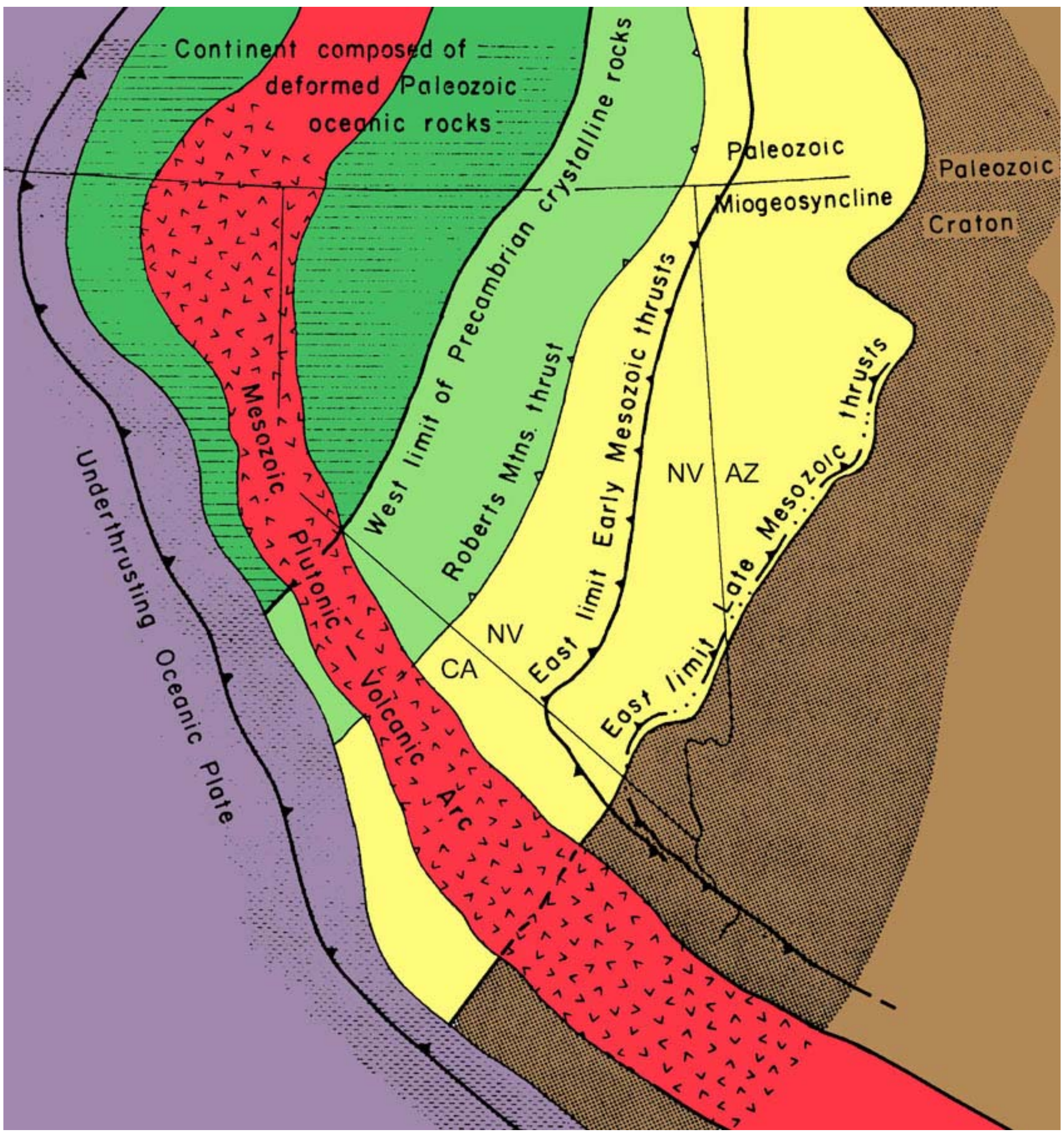


Figure 26 – Geologic sketchmap showing the truncation of Antler, Sonoman, and Sevier thrusts. (Burchfiel and Davis, 1972, Fig. 7.)

Nevadan Orogeny

By the late Triassic, after some significant shifts in polar wander paths (indicating rapid plate reorganizations), the SW Cordillera looked quite different. Development of a NW-trending megashear zone cut across the Antler and Sonoma trends and prepared the newly arranged margin for an unprecedented epoch (Jurassic to present) of continuous continentward subduction. During the middle Jurassic **Nevadan orogeny**, island arcs were swept into the Cordilleran margin and subduction flips were common (Figure 20). The scattered volcanic island arcs of the SW Pacific may offer a modern analog to the conditions that must have prevailed along the active edge of the western Cordillera throughout the Mesozoic (Figure 21).

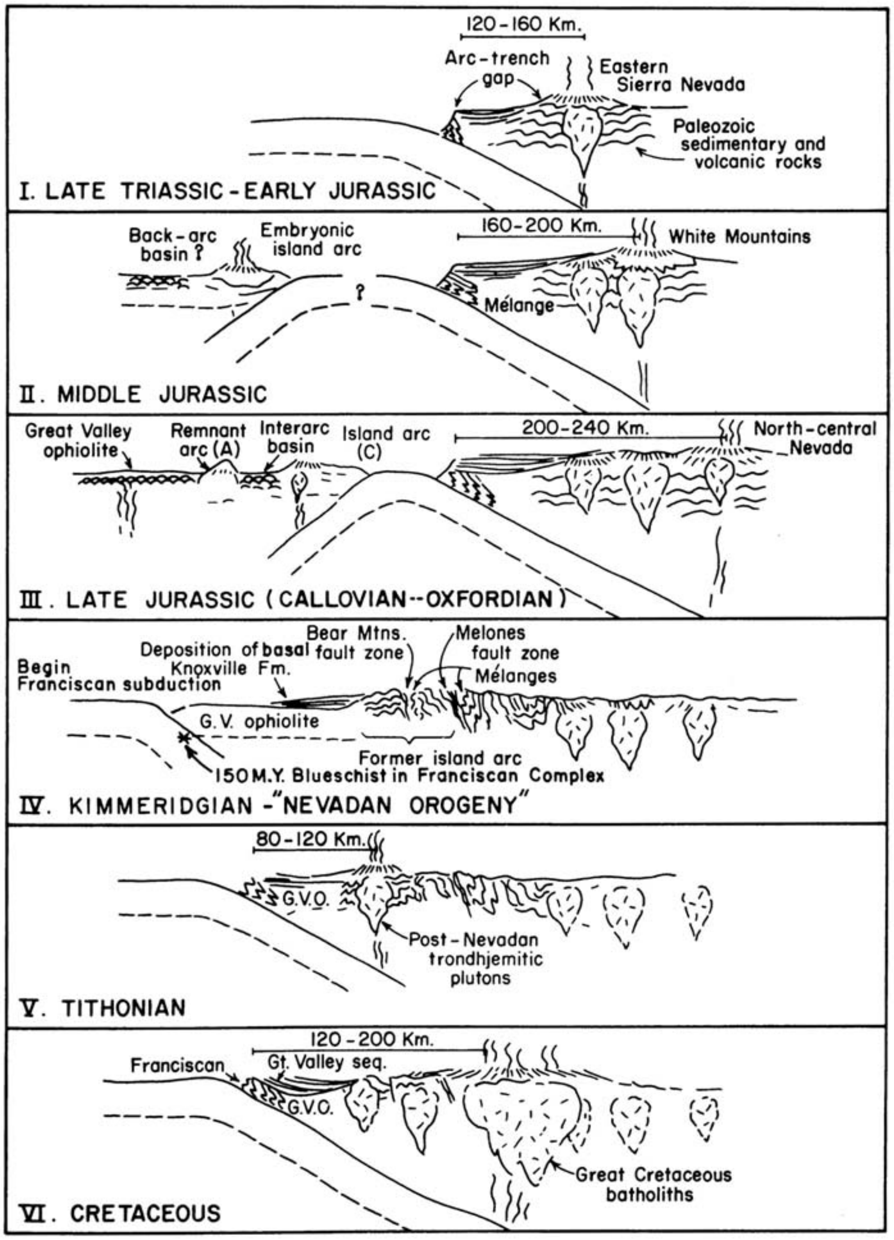


Figure 20 – Plate models to explain Mesozoic tectonics of the SW Cordillera. (Schweickert and Cowan, 1975.)

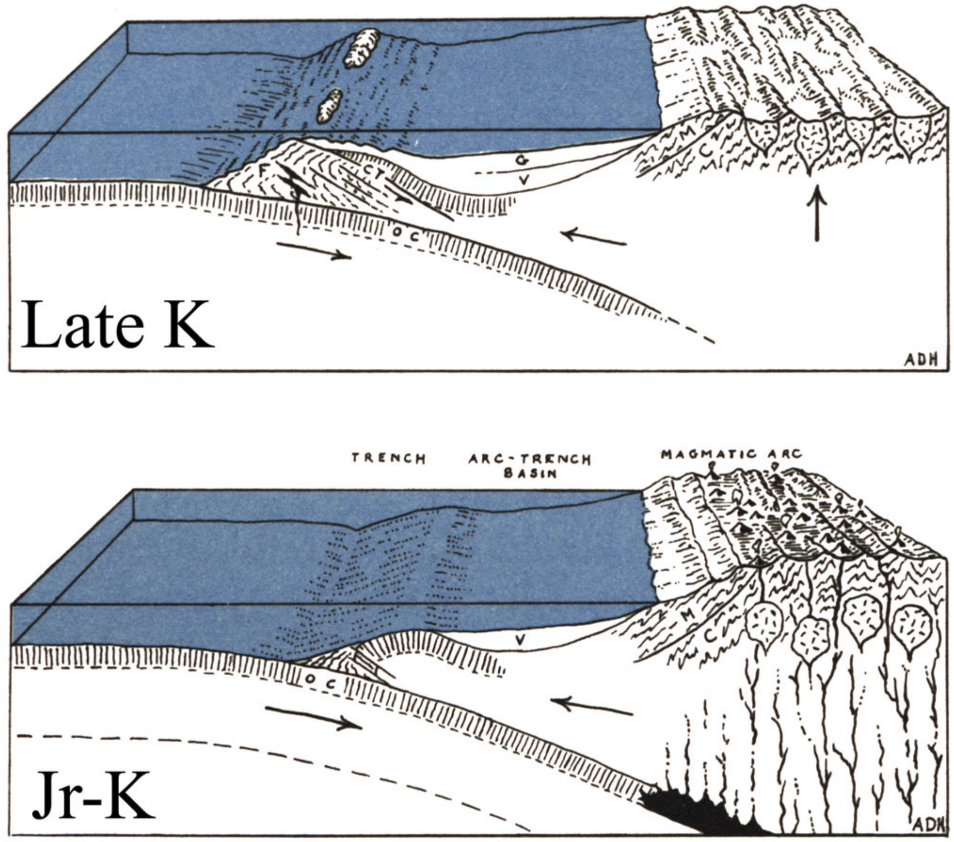


Figure 21 – Block diagrams depicting the growth and evolution of an Andean arc-trench system during Mesozoic time. (Composite diagram adapted from Howard, 1979.)

As we had mentioned earlier, the Mesozoic was a period of continuous subduction but many changes in tectonic plate subduction. For example, between the late Triassic to Cretaceous rapid tectonic suturing of exotic terranes and development of a broad tectonic welt occurred in the SW Cordillera. Thus, starting in Kimmeridgian (middle Jurassic) time and extending through the Cretaceous to present, continentward subduction has resulted in the truncation of all older geologic trends including Antler (Roberts Mountain) and Sonoman overthrusts. (See Figure 20.) The Cenozoic geology California is the product of continuous subduction against a complexly deformed Phanerozoic mobile belt and eventual changeover to a San Andreas type margin, the product of interaction with a subducted spreading ridge (Figure 22).

Laramide Orogeny and Beyond

The Laramide orogeny affected the SW Cordillera between late Cretaceous and Eocene time. Arizona was tectonically active with upwarps and downwarps of the region controlling sediment patterns. That instability was followed by intermediate to silicic eruptive volcanism in response to continuous subduction along the active Andean margin to the west. Lithospheric softening and NE-directed compression resulted in thermal collapse of internal core zones of the

Laramide overthrusts where ductile thrusts show maximum offset of 100 km. Known as metamorphic core complexes to some, these areas exhibit mobilized amphibolite facies gneiss in massive overthrust sheets. In addition, large Cenozoic plutons invaded portions of southern Arizona but Basin and Range faulting has obscured many of the geologic relationships.

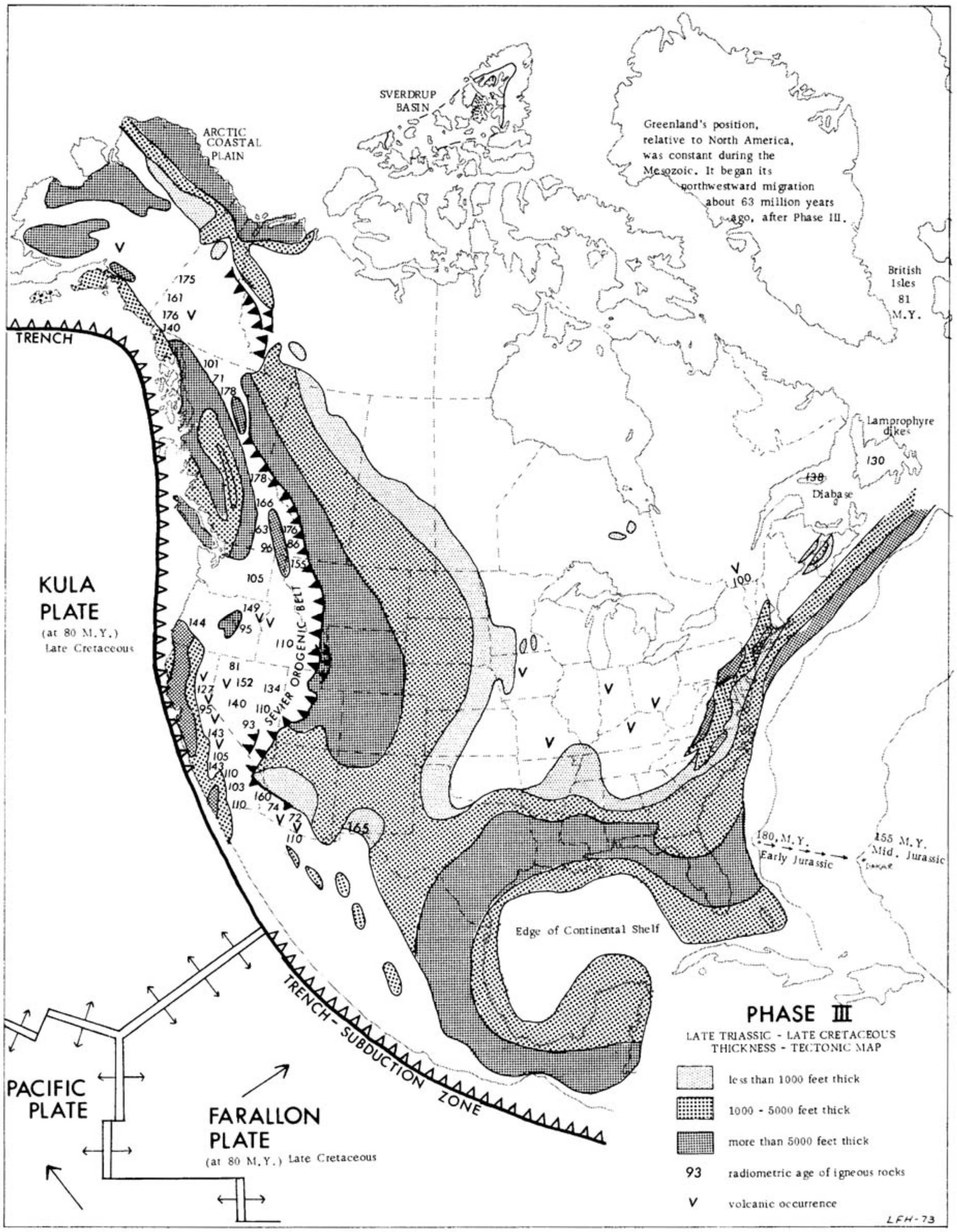


Figure 22 – Paleotectonic map of the Cordillera from Late Triassic to Early Cretaceous time.

Far as California goes, by early Cenozoic time subduction was in an Andean setting with development of an elongate volcano-plutonic complex parallel to the present coastline. With flattening of subduction angles in the Cenozoic, volcanogenesis broadened to include the Great Basin. Starting at about 30 Ma, destruction of the Farallon plate and development of the San Andreas plate boundary has dominated the tectonics of the region (Figure 23).

Figure 23 – Aerial view of the San Andreas fault. (From Shelton, 1966.)

A view of the Cenozoic tectonics of the region helps to understand the products of Cenozoic geologic development of California. Subduction of the east half of the Pacific Ocean plate during the Mesozoic and Cenozoic resulted in the formation of the western Cordillera of the United States. Subduction of the Farallon plate and parts of the East Pacific Rise began during the later part of the Cenozoic Era. By the Eocene (roughly 50 Ma) the plate configuration involved subduction of an offset ridge crest that ultimately resulted in fragmentation of North American crust and development of the right-lateral San Andreas fault system (Figure 24). By 30 Ma, the NE-corner of the transform-ridge Farallon segment had collided with northern Mexico and from that noble area subduction of western half of the Pacific Ocean plate officially began. Parades, celebrations, and parties were held throughout the western Cordillera.

By 15 Ma the partying had faded as the consequence of the subduction of an active ridge crest had become obvious – decreased volcanicity at first then continuous volumes of lava and explosive volcaniclastic debris all in the midst of active seismicity. Indeed, the result of subduction of the Pacific Ridge had far-reaching geological consequences. High heat flow and resulting volcanicity, extensional faulting, uplift, and seismicity can all be attributed to the consequences of subduction and active ridge crest. The development of Cenozoic metamorphic core complexes, with low-angle ductile normal faulting, were formed by thermal weakening produced by the same mechanism.

Starting about 5 Ma in Baja California our present plate configuration has produced a sliver of North America (the Salinian block) that moves northwestward along the Cordilleran margin. As demonstrated by studies of the western Canadian Rockies, this scenario has played out many times in the past. Much of the western Canadian Rockies form a collage of accreted terranes that are linked to the SW Cordillera of the United States and Mexico. The Cenozoic dislocation and drift of the Salinian block provides a candidate for the next accreted terrane when the future tectonics change to convergence. A summary diagram is provided as Figure 25.

The San Andreas system continues to evolve along the SW Cordillera. Because of the relative motion of the Pacific Ocean plate, the extension produced by the spreading of the subducted Pacific Ridge has resulted in an overall pattern of seismicity (Figure 26).

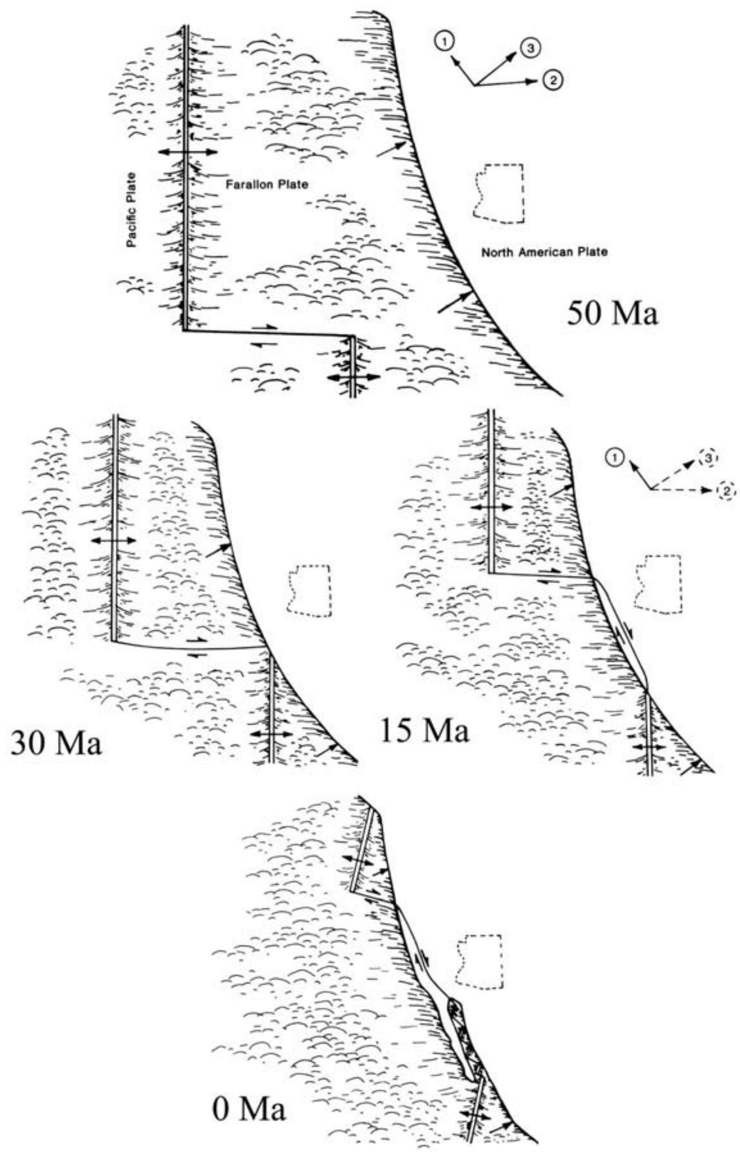


Figure 24 – Four views of the post-Eocene plate tectonic evolution of the SW Cordillera. The position of Arizona is shown as a dotted outline. (Adapted from Nations and Stump, 1981, Figure 7-6, p. 75-77.)

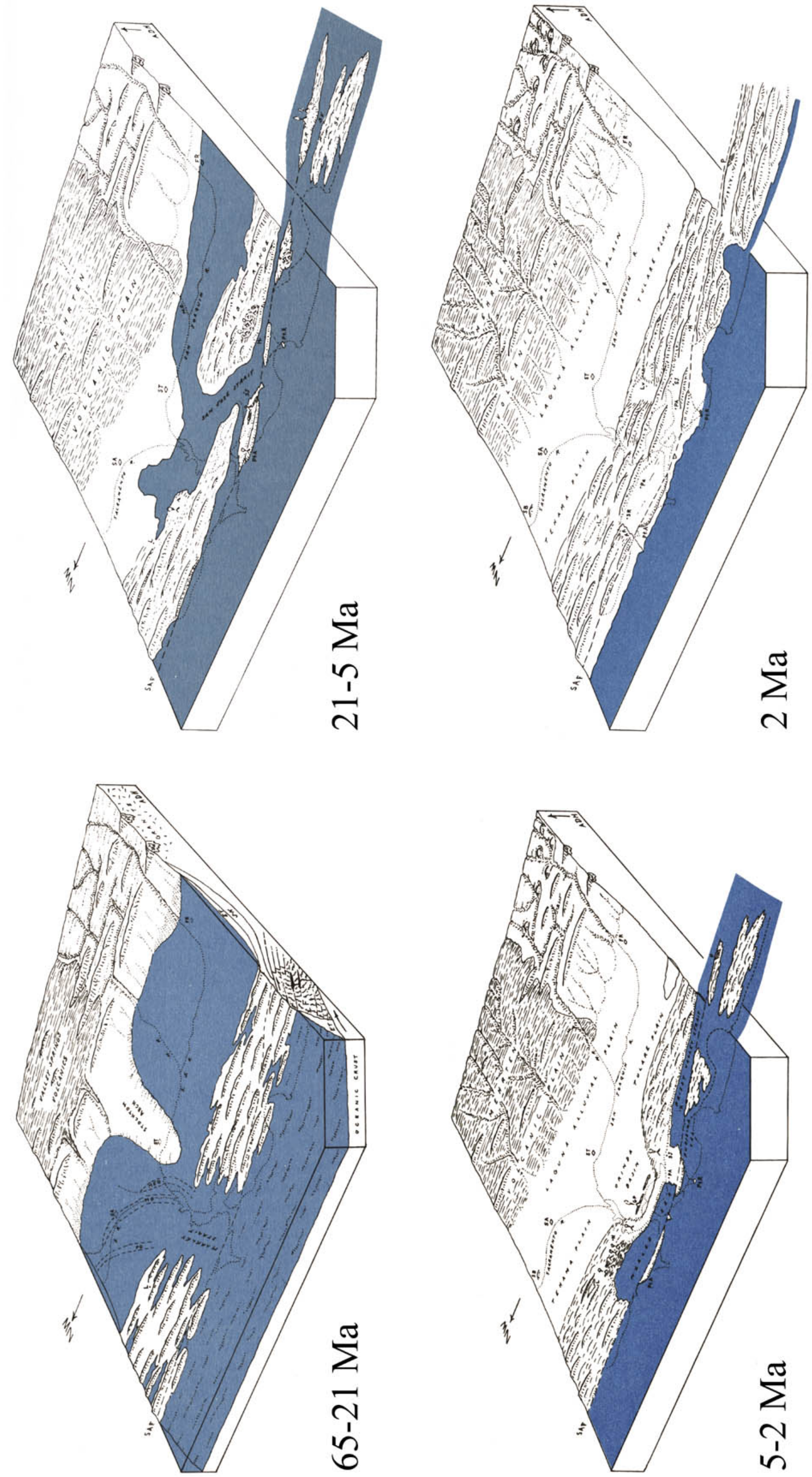


Figure 25 – Block diagrams showing the NW migration of the Salinian block and physiographic changes taking place during the last 65 Ma. (Composite diagram adapted from Howard, 1979.)

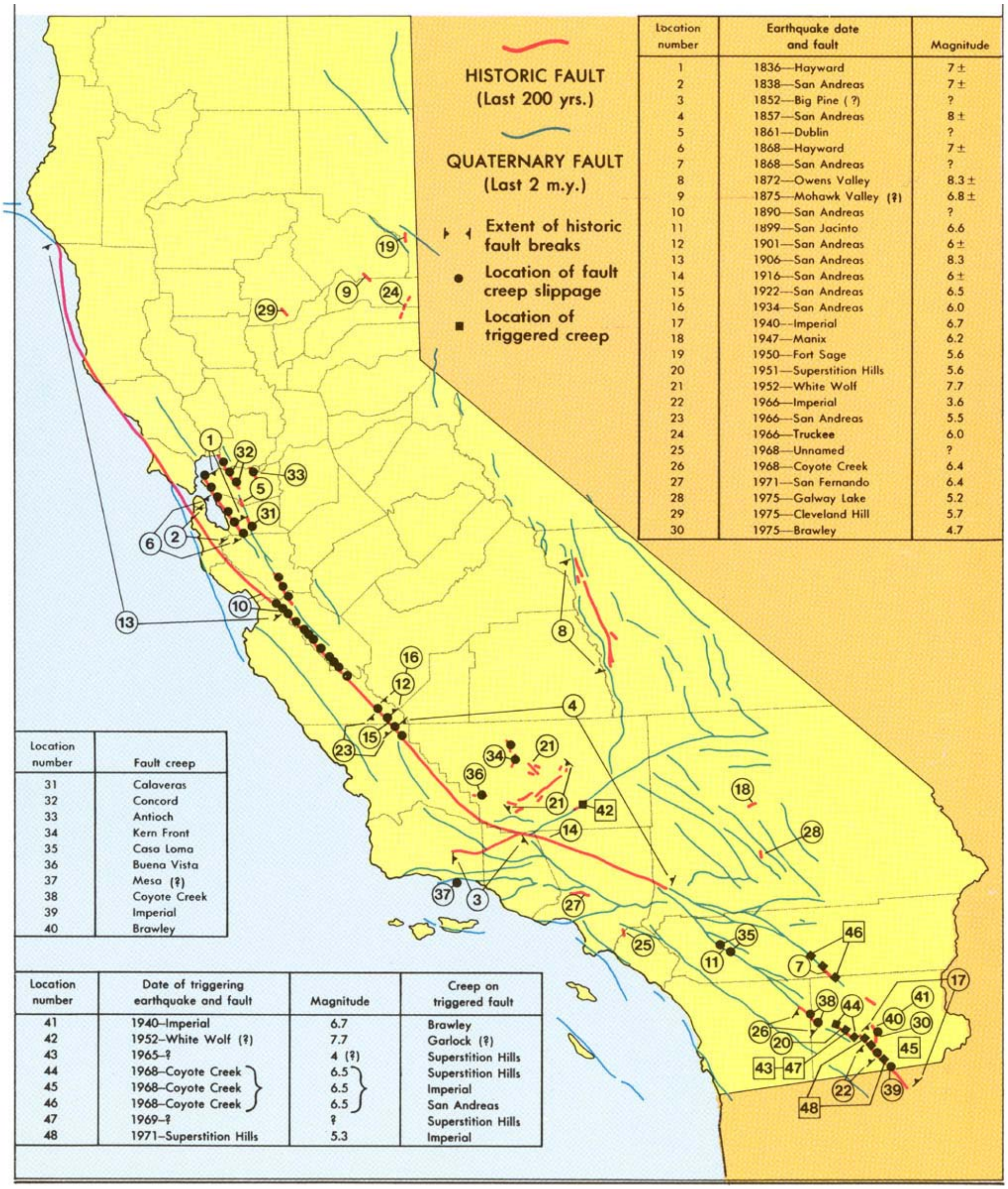


Figure 26 – Fault map of California showing the locations of major faults and a listing of major events. (From California Division of Mines and Geology, 1980.)

Thus, we end the description of the geological background and tectonics of California and we can look over the itinerary (below), then move on to the day by day details of the geology and history of the four segments of our trip across California. Described separately below in the appropriate section, the Coast Range and Great Valley, Foothills Metamorphic Belt, Sierra Nevada, and Mono-Inyo Basin form the four geological segments of our 10-day excursion. Appendix A includes a Geological Primer. Reading the primer will help you brush up on the terminology and major concepts of geology that we need to apply to our outcrop scale study of this impressive terrane.

Itinerary for Geology 143D – Field Course in Nevada/California

We will make every attempt to go to as many places listed in the itinerary as possible and will certainly stop at other places along the way if deemed necessary to develop the field trip message. Detailed geologic descriptions of the four trip segments follow. The following pages will be referred to often in the field during our trip. Our entire trip route will take us through the Basin and Range, Death Valley, Mono Craters, and the Sierra Nevada mountains in Nevada and California, a part of the Intermontaine Division of the SW Cordillera of the United States. (See Figure 4.) The landscape varies greatly in relief and climate from desert in the south and central valley to gentle coastal and rugged interior mountains where elevations exceed 16,000'. (See Figures 5, 6).

Day 1 (Saturday, 04 April 2009) – Red Rock Canyon State Park

Jet Blue Airlines Flight 187 from JFK to Las Vegas
Depart: 7:55 AM JFK; Arrive 10:42 AM (Las Vegas) [~5.75 hour flight].

Meet Starr Lanphere; Secure Van; Secure food for camp and get lunched.
Out of the airport we'll find Hwy 159 to Hwy 160 West to Red Rock Canyon State Park where we will set up camp for two nights and begin the scenic drive with a geology hike planned for the afternoon to see the keystone thrust in afternoon (time permitting).

Stay overnight (actually two nights) in Red Rock Canyon State Park.

Day 2 (Sunday, 05 April) – Red Rock Canyon State Park

Continue with the scenic drive stopping for a geology hike in AM.
Lunch and continue with the scenic drive stopping for a geology hike in PM.

Same camp as last night.

Day 3 (Monday, 06 April) – Death Valley

Break Camp early AM and drive west via Hwy 160 past Mountain Springs about 15 miles. Turn L onto Old Spanish Trail Road and head into Tecopa and set up camp (2 Days).
After we get settled plan to do geology tours and hikes in the Death Valley area by taking Hwy 127 north through Shoshone into Death Valley Junction. From there take Hwy 190 West to Death Valley National Park entrance. Follow Guidebook.
Sunset Hike.

Camp at **Ranch House Inn and Hostel** (760 852-4580; Coleen) in Tecopa, CA for two nights.

Day 4 (Tuesday, 07 April) – Death Valley

All Day Geology Tour and Hikes (Pack Food).
Follow Guidebook. Stay again at **Ranch House Inn and Hostel**

Day 5 (Wednesday, 08 April) – Mono Craters – Inyo "Face" Mountains

Break Camp and Drive north on 127 then west on Hwy 190 through Panamint Springs and the Inyo range onto Hwy 136 through Keeler and join Rte 395 north to Bishop. Continue on Rte 395 North to the Devil's Postpile and for hikes and geologic tours in the Mono Craters area.

The higher parts of Mono craters are difficult to reach and we will rely on a guide by Tierney, (1995). Spend the day here. On the South shore is Panum crater a 600-700 year old obsidian dome and tuff ring. Continue south out of the Mono Lake area to Mammoth to see a 30-mile to long caldera valley, roughly 780,000 years old. The Bishop Tuff is exposed there.

Camp overnight in Bishop area, perhaps at the Grandview campground in the White mountains.

Day 6 (Thursday, 09 April) - Yosemite

Break Camp early.
Take Rte 120 West over through Tioga Pass and climb Sierran crest with some field stops along the way. Eventually, get to Yosemite for tour and hikes and we will stay there for the evening.

Camp in Yosemite Valley at Curry Village in tent cabins for the night (it will be cold!)
Reservation confirmation #s (76M9PK and 76M9PJ) and phone (209 372-8502).

Day 7 (Friday, 10 April) – Yosemite and Tioga Pass

After Yosemite backtrack eastward out over the High Sierra. Cut up Canyon Road and take Big Oak Flat Road to Crane Flat then E to Tioga Pass Road. It's about 50 miles from Yosemite to Tioga Pass. Stop at Yosemite Creek – High Country scenery and scenery. Drive to Olmstead Point for view back at Yosemite Half Dome granodiorite and view the effects of glaciers.

About 1-2 miles east of Olmstead Point, Tenaya Lake (a Sierran mountain lake) can be found where the contact is exposed between the Half Dome Granodiorite and the Cathedral Peak granite (w/ large K-spar). At Tioga Pass we are in the Saddlebag Lake roof pendant. Option: At Tioga Pass take road N from highway to Saddle Bag Lake. Glacial features are visible to the west as well as an angular unconformity and Triassic Caldera (Schweickert and Lahren, 1999). The caldera is exposed along the highway near Tioga Lake. Time permitting, plan to do some hiking in areas to the east side of Tioga Pass. About 9 miles E of Tioga Pass we get to Lee Vining Canyon (before the town of Lee Vining).

Start drive back to Las Vegas in early afternoon to ensure time for dropping off van and for making travel connections.

Jet Blue Flight 196; Vegas to JFK
Depart 9:50 PM Las Vegas Airport; Arrive JFK at 5:41 AM.

Day 7 (Saturday, 11 April) – Back to New York

Field Trip Descriptions

Segment 1

Red Rock Canyon State Park, Nevada

Red Rock Canyon National Conservation Area in Nevada is a 195,819 acre (801 km²) area managed by the Bureau of Land Management as part of its National Landscape Conservation System, and protected as a National Conservation Area. The conservation area showcases a set of large red sandstone peaks and walls called the Keystone Thrust. The walls rise up to 3,000 ft (1,000 m) high, making them a popular hiking and rock climbing destination. The highest point is La Madre Mountain, at 8,154 ft (2,485 m).

Red Rock Canyon is a relatively small and narrow valley that is situated around 15 miles west of downtown Las Vegas. Red Rock Canyon NCA (National Conservation Area) is one of the most scenic spots to visit in Southern Nevada. 'Red Rock' as it is known to the locals, gets its name from the red colored sandstone formation embedded in the mountains that form the western and northern margins of this small valley. The distinct red colored layers of sandstone streak horizontally across mountains known as the Wilson Cliffs and can be seen from almost anywhere in and around the Las Vegas desert.

While most of the Wilson Cliffs are visible from the Las Vegas Valley, much of Red Rock Canyon is nestled below them and out of sight from the city of Las Vegas. The red sandstone that makes Red Rock so unique, is part of the same geologic formation, the Navaho sandstone Formation, that is found in the Valley of Fire, Zion National Park and throughout many parts southern Nevada and the southwestern United States including the Grand Canyon.

The underlying structure of Red Rock valley is typical of the north-south trend of the ranges and valleys of the Great Basin found throughout Nevada and many parts of the American southwest. The spectacular Wilson Cliffs consist of the Bridge Point, Bridge, Rainbow, Mount Wilson, Indecision Peak and Sandstone mountains. The eastern boundary of the valley consists of a series of low hills collectively known as Blue Diamond Hills. Between these mountains and hills is the valley that State Highway 159 runs through. The southern part of the valley is home to the Blue Diamond mining community.

The easiest way to see Red Rock is to just drive through the place. A one-way loop road, 13 miles (21 km) long, provides vehicle access to many of the features in the area. Several side roads and parking areas allow access to many of the trails located in the area. When stormy, the visual effect of the clouds against the mountains and cliffs is dramatic. When the weather is good, hiking and picnicking are recommended because you will have a lot to choose from. There are picnic areas and hiking trails in Calico Basin, the Scenic Drive, Oak Creek and the Spring Mountain Ranch. The best time for photography is very early in the morning. Exactly at and just after sunrise, the light is perfect to shoot pictures of the Wilson Cliffs. The sunlight, at that angle, highlights the red colors in the cliffs because they face east and reveals the subtle colors in the sandstone best.

Geological History

The Red Rock Area has a complex geological history, which over millions of years, helped to create the dramatic landscape that characterizes the region. For much of the past 600 million years, the land that is now Red Rock Canyon NCA was the bottom of a deep ocean basin and the western coast of North America was in present day western Utah. A rich variety of marine life flourished in those waters and left behind deposits of shells and skeletons more that 9,000 feet thick which were eventually compressed into limestone and similar carbonate rocks.

Beginning approximately 225 million years ago crustal movements caused the sea bed to slowly rise. Streams entering the shallower waters deposited mud and sand which later consolidated into shale and marine sandstones. Changing land and sea levels trapped large bodies of water which later evaporated leaving behind layers of salt and gypsum in some areas. Exposure of the sediments to the atmosphere allowed some of the minerals to oxidize, resulting in red and orange colored rocks. Streams meandering across the broad plain deposited sand, mud, gravel and amassed other debris such as tree limbs and logs. In some cases, minerals in the groundwater replaced the organic materials in the buried logs forming petrified wood. Petrified wood is one of the few fossil remains found in the rocks at the foot of the Wilson Cliffs.

About 180 million years ago the area was completely arid, much as the Sahara Desert is today. A giant dune field stretched from this area eastward into Colorado, and windblown sand piled more than half-a-mile deep in some spots. As the wind shifted the sands back and forth, old dunes were leveled and new ones built up leaving a record of curving, angled lines in the sand known as "crossbeds". These shifting sands were buried by other sediments, and eventually cemented into sandstone by iron oxide with some calcium carbonate. This formation, known locally as the Aztec Sandstone, is quite hard and forms the prominent cliffs of the Red Rock escarpment. In some areas the iron minerals in the rocks have been altered and concentrated giving the rock its red color.

The most significant geologic feature of Red Rock Canyon NCA is the Keystone Thrust Fault. The Keystone Thrust is part of a large system of thrust faults that extends north, into Canada and began to develop approximately 65 million years ago. A thrust fault is a fracture in the earth's crust that is the result of compressional forces that drive one crustal plate over the top of another. This results in the oldest rocks on the bottom of the upper plate resting directly above the youngest rocks of the lower plate. At Red Rock Canyon NCA, the gray carbonate rocks of the ancient ocean have been thrust over the tan and red sandstone in one of the most dramatic and easily identified thrust faults to be found. The Keystone Thrust Fault extends from the Cottonwood Fault along State Route 160 north for 13 miles along the crest of the Red Rock escarpment. It then curves east along the base of La Madre Mountain before it is obscured by very complex faulting north of the Calico Hills.

Native Americans

The first humans were attracted to the Red Rock area due to its resources of water, plant and animal life that could not be easily found in the surrounding desert. This made Red Rock Canyon NCA very attractive to hunters and gatherers such as the historical Southern Paiute and the much older Archaic, or Desert Culture Native Americans.

As many as six different Native American cultures may have been present at Red Rock over the millennia. The following chronology is an approximation:

* Southern Paiute 900 A.D. to modern times
* Patayan Culture 900 A.D. to early historic times in the 1800s
* Anasazi 1 A.D. to 1150 A.D.
* Pinto/Gypsum (Archaic) 3,500 B.C. to 1 A.D.
* San Dieguito 7,000 to 5,500 B.C.
* Paleo-Indians (Tule Springs) 11,000 to 8,000 B.C.

Petroglyphs. Numerous petroglyphs as well as pottery fragments remain today throughout the area. In addition, several roasting pits used by the early Native Americans provide further evidence of human activity in the past at Red Rock.

Links:

http://www.blm.gov/nv/st/en/fo/lvfo/blm_programs/blm_special_areas/red_rock_nca.html
http://www.blm.gov/nv/st/en/fo/lvfo/blm_programs/blm_special_areas/red_rock_nca/red_rock_s_unique/red_rock_geology.html
http://www.localhikes.com/Hikes/KeystoneThrust_4120.asp
http://www.friendsofredrockcanyon.org/keystone.php
http://www.birdandhike.com/Hike/Red_Rocks/Keystone/_Keystone.htm

Keystone Thrust Hike

Trailhead: 18 miles W of Las Vegas Blvd. on Charleston Ave. (SR 159) to Red Rock entry (fee) station. Take 13 mile scenic loop past Sandstone Quarry and past the Highest Point Overlook to the signed turn for White Rock Hills parking area. If you are in a passenger vehicle you might want to park at the lower parking area (wide dirt area on L side of road as soon as you turn off paved scenic loop), and begin your hike here. It's a short but rocky drive up the dirt road to the upper parking area, where a vault toilet and trail signage is located. There is very limited parking at upper White Rock Hills and rock climbers use it as well as hikers so be considerate and park efficiently. This parking lot is too small and the road is not suitable for trailers or long wheel base vehicles, but if you realize this too late, the road ends in a loop so you can drive back down to the lower parking area and walk up to the trailhead. There is signed agave roasting pit just above the upper parking lot.

During the winter months when the days are short, it's nice to have a few small hikes available that can be enjoyed in just a couple of hours. One of these is the Keystone Thrust Trail. It's easily accessible, packed with interesting history and a pretty good workout, too! To find this trail, drive about six miles around the scenic loop until you see a sign announcing White Rock. From here, a rocky dirt road leads north another mile to a fenced parking area that serves the White Rock and Keystone Thrust trails. This will be at least a three-mile hike, depending on the return route that you decide to take, so don't forget to pack a bottle of water for each person. Even in these cooler winter months it's important to stay hydrated while you're in the desert.

At the north end of the parking lot you'll find an interpretive sign briefly explaining the Keystone Thrust trail. As it turns out, this is a unique and important area for geologists, and they come from all over the world to study this thrust fault. You'll discover why as we get a little farther up the trail. Follow the rock-lined path north across the wash until you come to a sign directing you up a picturesque little hill with railroad-tie steps. At the top of the hill the trail intersects an old road. Turn left and follow the road around Hogback Ridge and toward the La Madre Mountains (Figure 27). Legend has it; in the early days of Las Vegas this remote desert road was a popular place for stolen cars to be stripped. Those old vehicles were removed long ago, but if you look carefully you may be able to spot some small metal parts slowly rusting away among the rocks and shrubs.

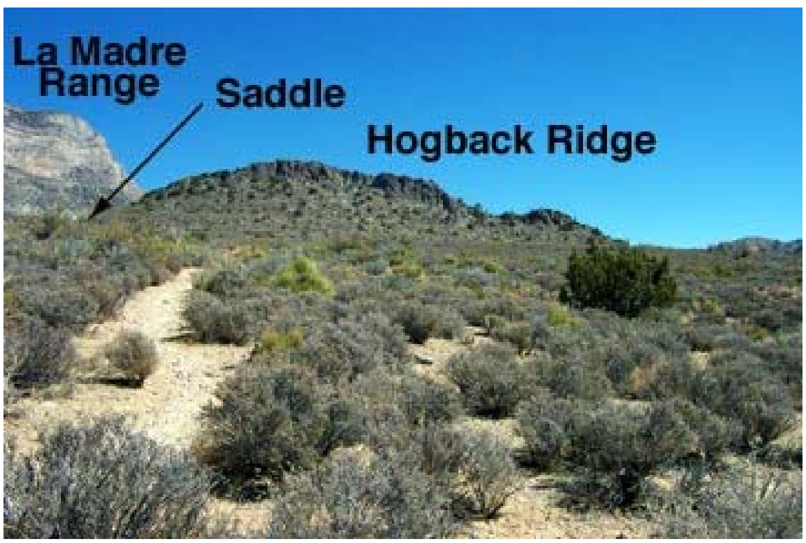


Figure 27 - Keystone Thrust (view east from the saddle on the northern edge of Hogback Ridge).

Once you've passed Hogback Ridge you'll come to another sign directing you toward the Keystone Thrust. This is the highest point of the trail and the views are really nice with Turtlehead Peak and the Calicos to the east, the La Madre Mountains to the north, and White Rock Mountain to the west. The trail now leaves the old road, turns sharply east and descends into a wide red and gray canyon. This is what the geologists go bonkers over. Two exposed tectonic plates! These are gigantic slabs of the earth's crust that move around to create earthquakes, mountains and continents. It's actually possible to stand with a foot on each plate. Keystone Thrust Fault is one of the only places on earth where this can be done! Normally, this kind of exciting rock action is taking place miles underground. Even for the non-geologist, it's interesting to see the red sandstone on the west side of the canyon smashing into the grey limestone on the east side forming new mountains before our very eyes. Try to think in rock-years to get the big picture.

For most people, this is the grand finale. Time to turn around, hike back the way you came in and call it a day. But as a Boot Tracks reader, you'll want to know that there are some hidden treasures to be found just a little farther south and deeper into the canyon. The rocks become very dramatic and in the winter you'll find pools of water reflecting the sky and trees. The jagged east side of the Hogback looms darkly overhead. Pick your way around a high waterfall area and carefully follow the canyon through dense shrubs and colorful boulders until the canyon slowly gives way to the open desert. The wash eventually leads back to the scenic loop drive, depositing you about a mile east of the White Rock turnoff. If you decide to take this route, please stay in the wash and avoid shortcutting across the desert to your car. This is sensitive ground with areas of active cryptobiotic (living) soils. Use your tread lightly skills and be sure to leave no trace!

The three-mile round trip to the Keystone Thrust is a fun outing and a good workout with a major educational bonus at the end. It deserves six boots all on its own, but if you have the extra time and inclination, exploring the canyon a little deeper "thrusts" the overall rating to seven boots!

Segment 3

Death Valley Area

Death Valley (Figures 28, 29) lies between the Panamint and Amargosa mountain ranges in eastern California. The valley is the lowest, driest and hottest place in the United States with an immense range in elevation. A portion of the Badwater saltpan is 282 feet below sea level while nearby mountain peaks reach over 11,000 feet. The annual precipitation in Death Valley is a mere 1.9 inches and temperatures typically soar to 130° F on summer days and can fall below freezing on winter nights. The harsh environmental extremes of Death Valley make human existence there difficult. The valley was named by a small group of 49ers that survived crossing the valley en route to the gold mines of California in 1849.

The landscape is so varied and extreme, one cannot help to wonder what events have taken place to create such a harsh and foreboding environment. From Badwater, the lowest point in the western hemisphere at 282 feet below sea level, to the 11,049 foot high Telescope Peak, this rugged topography, as well as sand dunes, volcanic craters and flood-carved canyons indicate a lengthy and complex geologic history. This history begins in Proterozoic time.

Figure 30 is a geological map of the Death Valley region and Figure 31 is a stratigraphic time chart, both useful in comprehending the following descriptions. In addition, I provide a time chart that lists the formative geological events of the region as Figure 32.

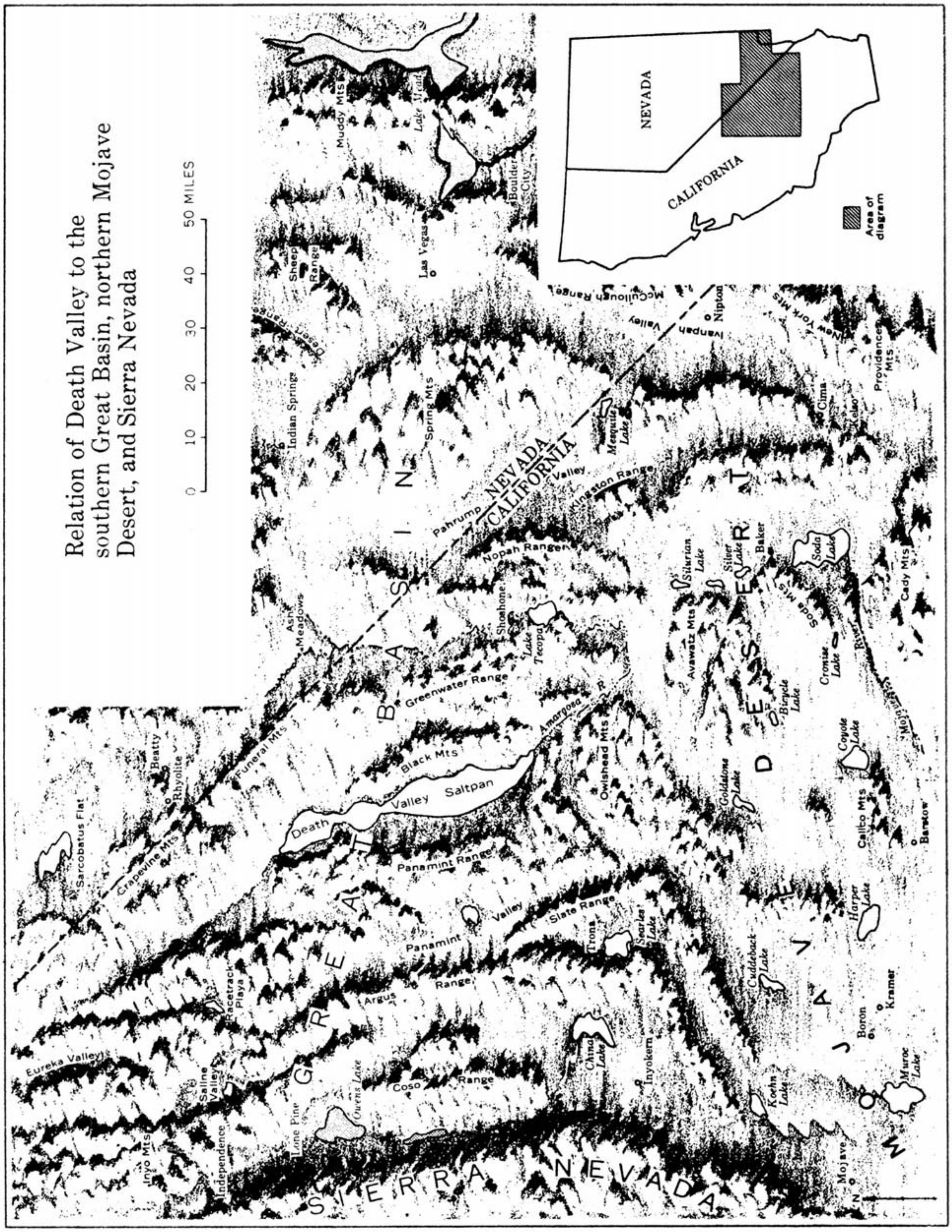


Figure 28 – Physiographic diagram of the Great Basin area of Nevada and California. (Adapted from Webb, 1969.)

Figure 29 – Index map of place names, roads (red), and interesting sites (yellow squares) in the area of Death Valley National Park (outlined in purple).

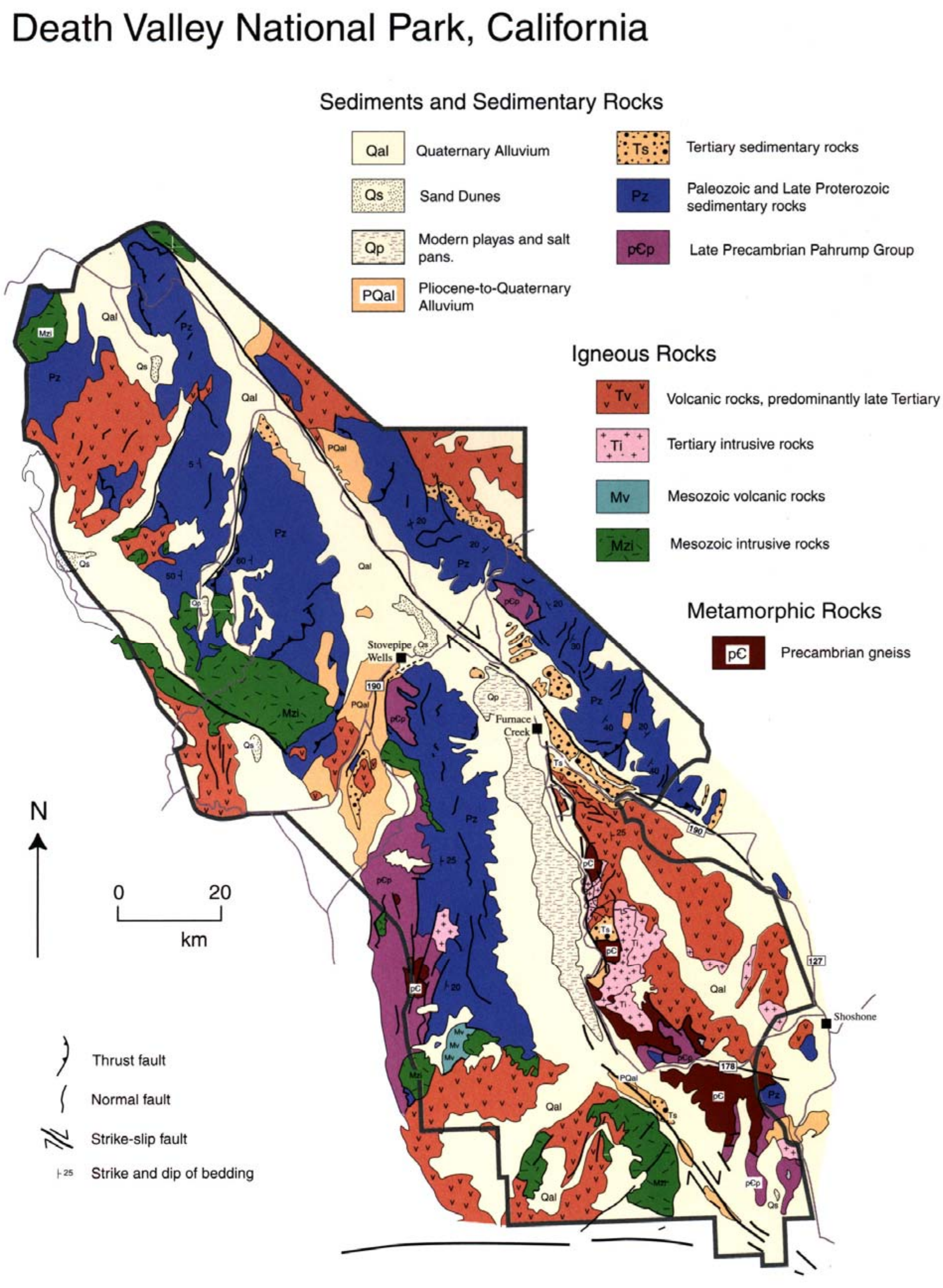


Figure 30 - Geological Map of the Death Valley region. (From Miller and Wright, 2007, Fig3A, p. xi.)

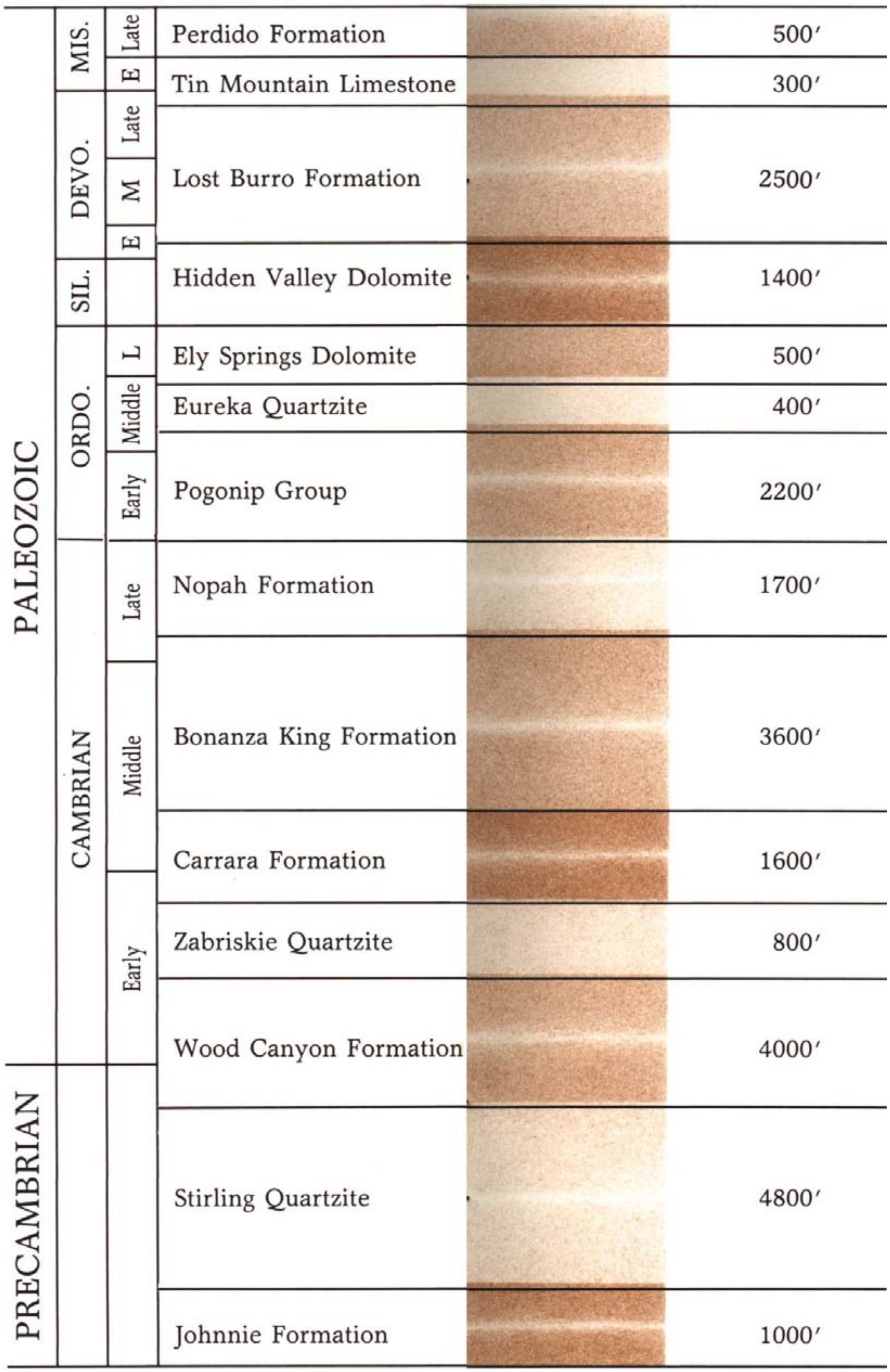


Figure 31 – Stratigraphic tiem chart of the Death Valley region. (From Collier, 1990, p. 44.)

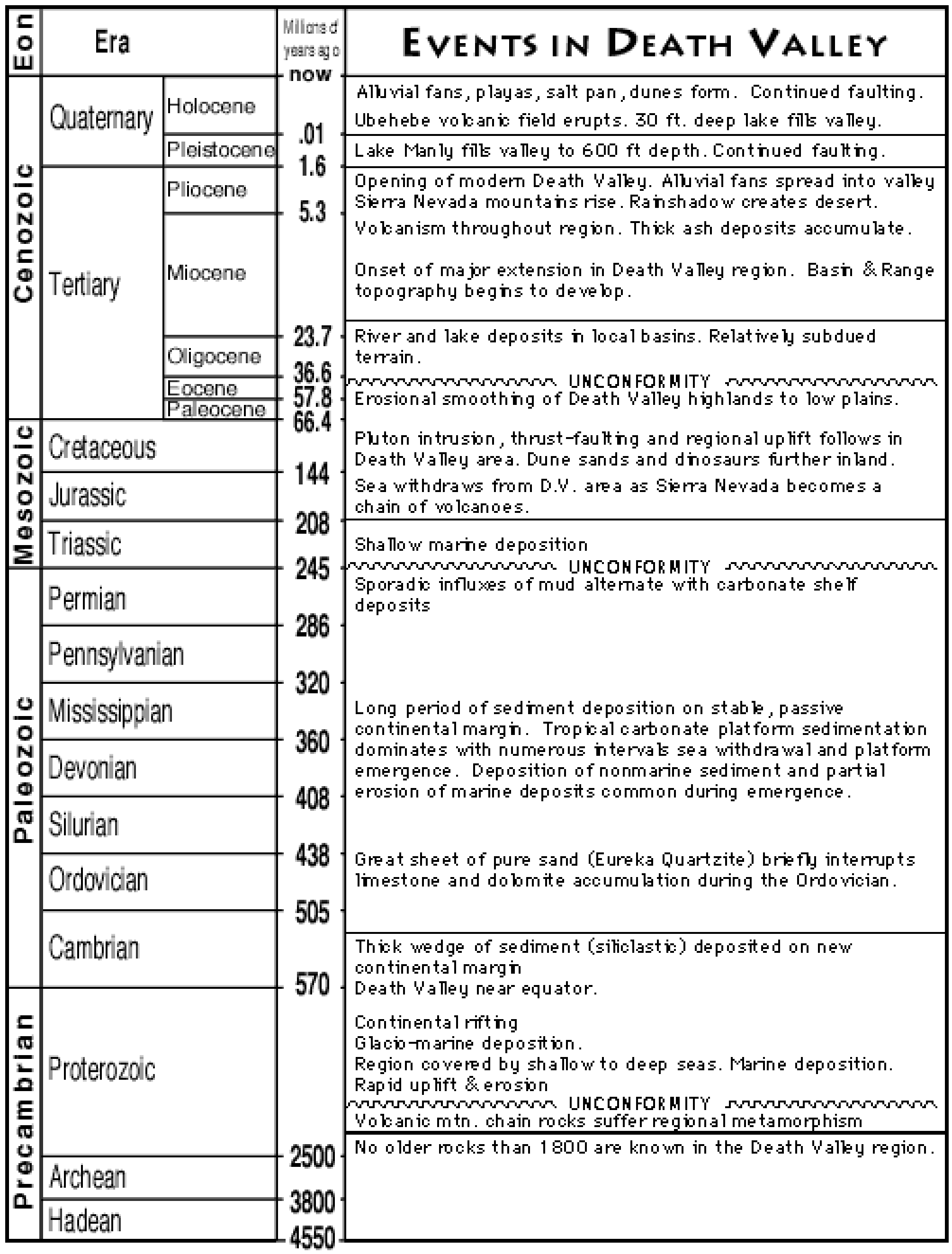


Figure 32 – Time chart and list of geological events that resulted in Death Valley. (U.S. Geological Survey.)

Proterozoic Geology

Death Valley's rocks, structure and landforms offer a wealth of information about what the area may have looked like in the past. It is apparent that there has not always been a valley here. Death Valley's oldest rocks, formed about 1.8 billion years ago have been so severely altered and suffered so many changes the their history is nearly unreadable. The oldest rocks are relics of the Proterozoic, the remnants of an ancient volcanic mountain belt with flanking deposits of mud and sand. About 1.8-1.7 billion years ago, these volcanic and sedimentary rocks were severely metamorphosed—altered, recrystallized, and partially remelted by the Earth's internal heat and by the load of overlying younger rocks. The original rocks were transformed to contorted gneiss, making their original parentage almost unrecognizable. Metamorphic schist and gneiss, and intrusive granitic bodies are exposed in the Black Mountains at Dante's View and Badwater, and in the Panamint Mountains north of Hanaupah Canyon.

Rocks of the crystalline basement are metasedimentary with granitic intrusions and include schists and gneisses. One coarse facies of gneiss contains large feldspar crystals known as augen (Hunt, 1975, p. 96). These rocks are mapped as epC on the Geologic Atlas of California (1958) and listed as Panamint metamorphic complex and World Beater porphyry. These rocks may be seen in the steep front of the Black Mountains, close to the highway near Badwater and farther south at the head of Galena Canyon to the east front of Panamint Range north of Hanaupah Canyon (Hunt, 1975, p. 94).

Neoproterozoic

The Neoproterozoic is thought to span the time period from 900 to 543 Ma. The initial stage in the evolution of the California continental margin was the deposition of the miogeoclinal sediment wedge along a passive continental margin of Atlantic type (Stewart, 1972) but the rifting that initiated deposition was accompanied or preceded by basaltic volcanism at about 850 Ma (Dickinson, 1977) though continental separation was probably not complete until about 650 Ma according to Stewart and Suczek (1977). The succession of Precambrian-Cambrian rock units from the southeast near Las Vegas to the northwest in the Inyo-White mountain ranges show facies changes from continental deposits through shallow water or continental margin deposits in the death Valley area to deep water deposits in the Inyo-White Mountains.

Pahrump Series

The Pahrump series contains slightly metamorphosed sedimentary rocks of the Crystal Spring, Beck Spring and Kingston Peak formations. They were deposited in the Amargosa aulacogen bordered by the Nopah upland on the north and the Mojave upland to the south. In general, Pahrump time seems to correspond with "Belt" time of deposition in the Precambrian Belt Basin of western Montana and adjacent Canada. Details follow:

Crystal Spring Formation

The Crystal Spring formation, about 2,000 feet thick, "...consists of a basal conglomerate overlain by quartzite that grades upward into purple shale and thinly bedded dolomite; upper part, thick bedded dolomite, diabase, and chert .." (Hunt, p. 93) "... rests with profound unconformity on older Precambrian metamorphic and igneous rocks" (Roberts, 1976, p. 35). The formation contains a number of individual members.

• Upper member • Chert member • Algal Member • Dolomite member • Mudstone member • Feldspathic sandstone member • Arkose member

Roberts (1976) analyzed the middle and lower members and concluded that it was deposited during the initiation of a late Precambrian intracratonal trough. "... contains Baicalia stromatolites ... suggests a middle Riphean age (1,350 to 950 m.y.) for the algal member ... also contains Conophyton stromatolites ... the Baicalia-Conophyton association is typically middle Riphean ... age may be 1,350 to 1,200 m.y. and is intruded by sheets of diabase ... the carbonates are altered to talc ... extensively mined" (Hunt, 1975, p. 96).

Beck Spring Dolomite

The Beck Spring dolomite is the middle formation of the Pahrump series. Hunt (1975, p. 93) notes that it "... outcrops to the west. Blue-gray cherty dolomite; thickness estimated at 500 feet; identification uncertain". However, Labotka and Albee (1977) report that the "... Beck Spring dolomite lies conformably on the Crystal Spring throughout most of the Telescope Peak Area ... consists of 200 to 300 meters of massive bedded dolomite and sandy dolomite ... upper part ... on Sentinal Peak is predominately clastic and interfingers with quartz-rich arenite".

Kingston Peak Formation

The Kingston Peak Formation is "... mostly conglomerate, quartzite, and shale; some limestone and dolomite near middle. At least 3,000 feet thick" (Hunt, 1975, p. 93). It is "...northward thinning ... well displayed in exposures in the Black Mountains north of the highway between Jubilee Pass and the Ashford Mill site" (Wright, Troxel, Williams, Roberts and Diehl, 1976, p. 9). A slight angular unconformity above the Pahrump group is indicated by the truncation of the uppermost units in the Kingston Peak Formation by the Noonday Dolomite" (Labotka and Albee, 1977).

Noonday Dolomite

The Noonday is exposed in the southern Panamint range. Williams, Wright and Troxel (1976) discuss it in detail. Consists of "... dolomite in indistinct beds; lower part cream colored, upper part gray. Thickness, 800 feet" (Hunt, 1975, p. 92). Farther north, where mapped as Noonday (?) dolomite, it contains much limestone which is colored tan and white, and some limestone conglomerate. Its thickness there is about 1,000 feet (Hunt, 1975, p. 92).

Many more Neoproterozoic units are known from this area but beyond the scope of our investigation. Suffice to say that throughout the end of Proterozoic time, the Cordilleran margin became a trailing edge continental margin, basically mirroring the Appalachian geosyncline on the other side of the continent.

Paleozoic Miogeosyncline

The importance of the trailing edge Cortdilleran geosyncline has been discussed earlier in a regional context. (See Figure 23.) A clearer picture of the Paleozoic is presented by rocks dating about 500 million years ago. These Paleozoic rocks, found in the Panamint and Funeral Mountains are made of sandstone and limestone indicating it was the site of a warm, shallow sea throughout most of the Paleozoic Era (570-250 Ma).

Time passed and the sea began to slowly recede to the west as land was pushed up. This uplift was due to the movement occurring far beneath the Earth's surface. Scientists have discovered that the Earth's crust is composed of a series of interconnected sections, or plates. The site of Death Valley lies adjacent to the boundary separating two of these plates. As the plates slowly moved in relation to each other, compressional forces gradually folded, warped and fractured the brittle crust. This widespread rock deformation and faulting occurred throughout most of the Mesozoic Era (250-70 million years ago). Active mountain building periods alternated with quiet times when forces of erosion worked to break down the mountains that had formed.

Most of the exposed Paleozoic rocks in the region are Cambrian in age. They include marine deposits about 8,000 feet thick. Fossils found include trilobites. Exposed in the Panamint Mountains in Trail Canyon and on Tucki Mountain. Also found in the Funeral Mountains.

Wood Canyon Formation

Diehl (1976) provides a good description of the Wood Canyon formation. Nolan (1924) named the Wood Canyon Formation and described it from exposures in the northwest part of the Spring Mountains, Nevada. Hazzard (1937) extended the use of the name to the southern Nopah Range ... where he measured a section of Precambrian to Cambrian strata including those of the Wood Canyon Formation. Later, Stewart (1966 ...) divided the formation into lower, middle, and upper informal members. Still more recently, Stewart (1970), in a systematic regional stratigraphic study of the Wood Canyon and associated formations of the southern Great Basin, California and Nevada.

The Wood Canyon Formation is of particular interest because it has diverse lithologies, occupies a stratigraphic position athwart or near the Proterozoic-Paleozoic boundary, and its areal distribution is well documented. The section that contains the Wood Canyon and the underlying Stirling Quartzite, Johnnie Formation, and Noonday Dolomite in the southern part of the Nopah Range has been suspected as a possible stratotype for the Proterozoic-Phanerozoic boundary (Cloud, 1973).

The Wood Canyon Formation has been divided previously into three informal members that are recognizable over a large area of eastern California and western Nevada.

...The lower member is composed of interbedded siltstone (36 percent), thinly laminated to platy bedded, fine- to medium-grained feldspathic and micaceous sandstone (50 percent), and laminated, siliceous dolomite (14 percent). The latter occurs primarily as three subunits in the member... Mudcracks, trace fossils, loading features, scoured surfaces, and discontinuous, convex sandstone bodies that are cross-laminated suggest shallow water deposition…

...The middle member contains a lower sub-unit composed of arkosic conglomerate and grades up-section into cyclically bedded subarkosic sandstone and maroon siltstone.

The interbedded light olive-gray to tan siltstone (23 percent) and dark brown to pale red weathering, fine- to very fine-grained quartzitic sandstone (49 percent) of the upper member are similar to those in the lower member with respect to distribution and primary sedimentary features. Sandstone beds are 1 to 5 feet thick, well laminated to massive with occasional cross-laminations. The dolomite (12 percent) in the upper member is contained mostly in one sub-unit, which is characteristically cyclically bedded. Distinctive millimeter-sized platelets of echinodermal debris are found in the dolomite beds.

Zabriskie Quartzite

This unit probably correlates with the Prospect Mountain Quartzite in the Craton facies.

Carrara Formation

The primary reference for the Carrara formation seems to be: Cornwall, H. R. and F. J. Kleinhampl, 1961, Geology of the Bare Mountain quadrangle, Nevada, U. S. Geological Survey Map GQ?157, scale 1:62,500. The unit is an alternation of shaly and silty members with limestone members; transitional between underlying clastic formations and overlying carbonate ones. Thickness about 1000 feet but variable because of shearing" (Hunt, 1975, p. 92).

McAllister (1970) provides this summary of the Carrara formation: The Carrara formation, of early and middle Cambrian age, displays the transition from siliceous clastic to carbonate rocks, apparent even from a distance, by a diversity of colors and varied relief of different layers. In the lowest part of the Carrara, light colored quartzite and dark shale resemble rocks in the underlying Zabriskie and Wood Canyon formations, whereas silty limestone in the highest part resembles limestone in overlying formations. Three prominent sets of limestone beds, separated by predominantly shaly and silty beds, stand out as ridges or as ribs on steep slopes. The two lower sets are conspicuously gray, but the highest is partly subdued by brown weathering clastic material. Much of the limestone, either silty or purer, contains concentrically structured ovoids as much as two inches long, generally attributed to the alga Girvanella.

Diagnostic trilobites in particular beds well spaced through the Carrara make it one of the most significantly fossiliferous formations in the section. They demonstrate that the lower part is early Cambrian and the upper part is middle Cambrian.

In the Dry Mountain quadrangle, the Carrara formation "... crops out only in the low hills 4 1/2 miles east of Dry Mountain ...", along the west side of Racetrack Road, and "... along the western boundary of the quadrangle in the Saline Range" (Burchfiel, 1969, p. 2).

Burchfiel (1969, p. 2) discusses the Carrara formation as it crops out near Racetrack Road as follows: East of Dry Mountain, the lower 600 feet of the Carrara Formation is composed of interbedded green and brown shale, orange-weathering silty limestone, gray to dark gray coarse-grained limestone, and brown calcareous sandstone and quartzite. The limestone beds range in thickness from six inches to two feet and may be grouped into ledge-forming limestone units up to 40 feet thick. Algal (?) pisolites up to two inches in diameter are common in the limestones. Overlying the basal shales are 900 feet of Carrara rocks containing two distinct limestone units. The lower limestone unit is about 250 feet thick and consists of dark gray thick-bedded coarse- to medium-grained limestone containing pisolites. Orange-weathering silty limestone is present at the top and bottom of this unit and grades into the green shale below and red, brown and green shaly siltstone above. The upper limestone unit is about 200 feet thick, consisting of gray and light gray limestone and dolomite. A prominent cream-colored band of dolomite occurs near the top.

Ordovician

Thickness 2,500 feet. Exposed in the Panamint Mountains in Trail Canyon and on Tucki Mountain. A coiled mollusc fossil was found in Ordovician(?) carbonate rock near the mouth of Trail Canyon. A quartzite of unknown age was found in the mouth of Trail Canyon.

Silurian

Silurian rocks contain fossil corals, crinoids and brachiopods. Present day corals require a water temperature of 68 F, a water depth of less than 150 feet and clear water. Exposed on Tucki Mountain in the Panamint Mountains.

Devonian and younger

All sedimentary rock, mostly limestone. Fossils include corals, crinoids and brachiopods. Exposed in the Panamint Mountains on Tucki Mountain. Post-Cambrian Paleozoic rocks also occur in the southern part of the Funeral Mountains.

Mesozoic

At the beginning of the Mesozoic, the continents of the Earth formed a supercontinent called Pangea. The western or northwestern coast of the supercontinent was the Panthalassa Ocean. During the Triassic fossiliferous marine sediments were formed. Some volcanic rocks may be exposed in the southern Panamints. Triassic age Hunter Mountain quartz monzonite is also known from the region. By the Jurassic mostly granitic intrusive rocks exposed in the Black Mountains and the southern Panamint Range.

Cenozoic

The next phase in Death Valley's development was primarily influenced by volcanic activity that spanned much of the Tertiary Period (70-3 million years ago). As mountain building stretched the land surface, the crust was weakened. Hot, molten material beneath the surface welled up and erupted at these weak points. The seething volcanos first appeared to the northeast, in Nevada, and blanketed the Death Valley region with numerous layers of ash and cinders. The topography at that time consisted of gently rolling hills. Over time, the center of volcanic activity moved progressively westward, finally producing a chain of volcanos running from Furnace Creek to Shoshone. Secondary results of the ash and cinder eruptions include the vivid colors of the Artist's Palette and Death Valley's infamous borate mineral deposits.

The oldest Tertiary deposits are exposed in the Titus Canyon area in the Grapevine Mountains and in the Funeral Mountains. Here the Oligocene to Miocene deposits are represented by the Artist's Drive formation. Composed of volcanic rocks, sedimentary rocks and playa deposits, the composite thickness is about 13,000 feet. Fossils include fresh water diatoms. Faulted and folded against the Precambrian core of the Black Mountains. In the northern Black Mountains, the Artist's Drive formation is overlain by the Furnace Creek formation of Pliocene age.

Approximately three million years ago, the dynamics of crustal movement changed, and Death Valley proper began to form. At this time, compressional forces were replaced by extensional forces. This "pulling apart" of the Earth's crust allowed large blocks of land to slowly slide past one another along faults, forming alternating valleys and mountain ranges. Badwater Basin, the Death Valley salt pan and the Panamint mountain range comprise one block that is rotating eastward as a structural unit. The valley floor has been steadily slipping downward, subsiding along the fault that lies at the base of the Black Mountains. Down-dropping continues today. Evidence of this can be seen in the seismicity and the fresh scarps exposed near Badwater.

Concurrent with the down-dropping has been slow, but continuous erosion. Water carries rocks, sand and gravel down from surrounding hills and deposits them on the valley floor. Beneath Badwater lies more than 9,000 feet of accumulated sediments and salts.

In addition to structural changes, Death Valley has been subjected to major climatic changes throughout the last three million years. During North America's last major Ice Age, the valley was part of a system of large lakes. The lakes disappeared approximately 10,000 years ago when the climate warmed. A shorter Ice Age, about 2,000 years ago, resulted in a smaller lake system. When this water evaporated, vast fields of salt deposits were left behind.

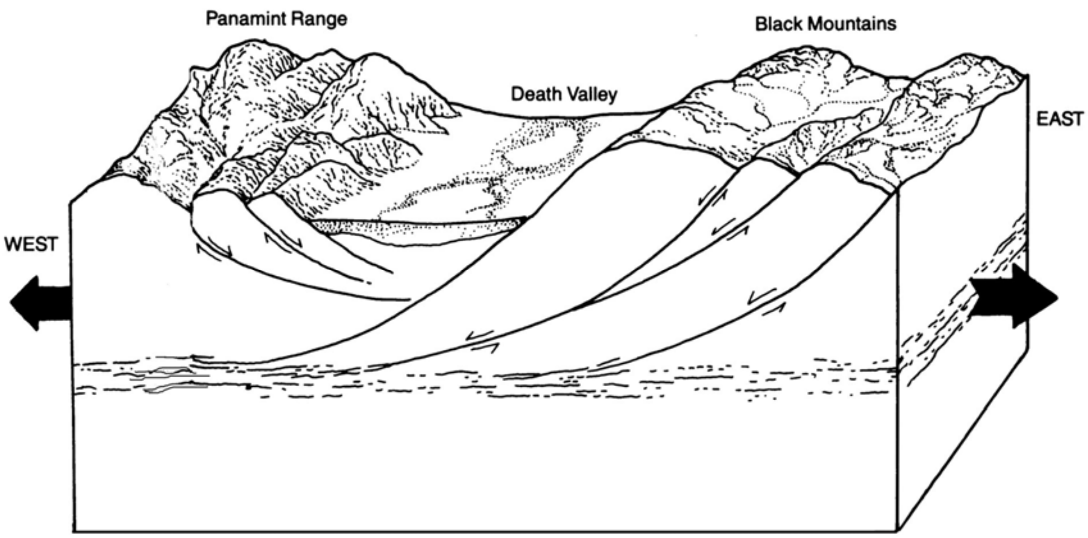


Figure 33 – Diagrammatic sketch showing extension in the subsurface of Death Valley and the resulting tilting of large blocks of the Earth's crust along lystric normal faults.

Recent signs of volcanic activity exist in northern Death Valley at Ubehebe Crater. Caused by violent steam explosions, the craters formed only about 2,000 years ago when hot, molten material came in contact with groundwater. These large depressions indicate that not all is quiet beneath the surface of the Earth.

Links:

http://digital-desert.com/death-valley-geology/

http://digital-desert.com/mojave-preserve/geology/08.html

http://geomaps.wr.usgs.gov/parks/deva/devatime.html

http://3dparks.wr.usgs.gov/deathvalley/index.html

http://www.schweich.com/dvfA.html

Segment 3

Mono Basin and Inyo Craters and the Long Valley Caldera

The third segment of our trip is to visit the Mono Basin volcanic district on the east side of the Sierra block. Recent seismicity along the Sierra Nevada – Great Basin Boundary Zone indicate the youthful tectonics of this region where significant right-lateral offset has been demonstrated. Clearly, the effects of right lateral shear of the San Andreas system are not restricted to the area immediately surrounding the fault. Locked transform-type tectonic boundaries, such as the San Andreas, form deep-seated lithospheric boundaries that distribute strain over a broad area. As such, right-lateral and resultant extensional forces are at play here, as is typical of the Basin and Range province of Nevada and southern Arizona. The heat derived from the subducted East Pacific Rise and lithospheric fracturing and extension write a sweet recipe for active volcanism. The area of Mono and Inyo Craters is the place to see it (Figure 34).

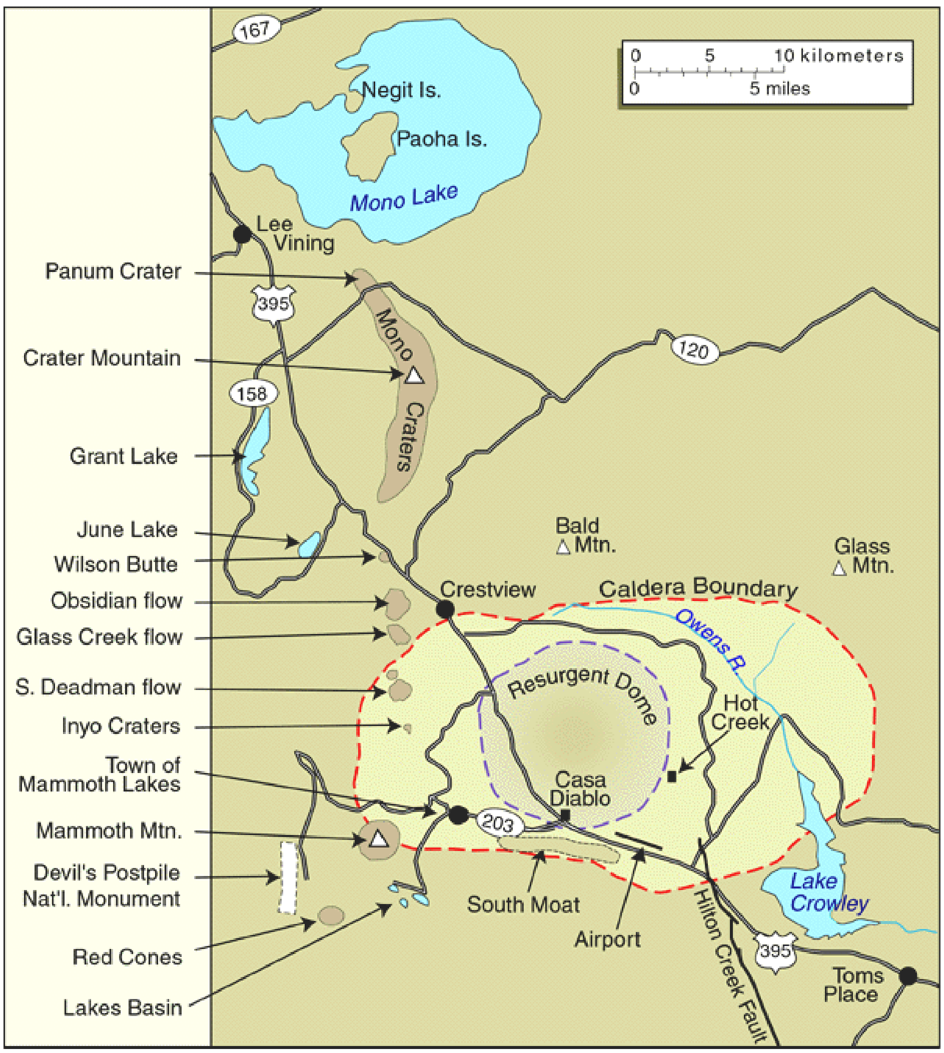


Figure 34 – Location map of the Mono-Inyo Craters area. From: *quake.wr.usgs.gov/VOLCANOES/LongValley/.*

Over the past 5,000 years, volcanic eruptions of small to moderate size have been reported from the area of Mammoth Mountain and the Mono Basin (Figure 35). Volcanic activity repeats every 250-700 years in that interval from sites along the Mono-Inyo craters volcanic chain. Steam blasts are responsible for large pits or craters and explosive eruptions typically spread hot ash as pyroclastic flows. These events are usually followed by lava doming and at other times lava flows. Of all of the areas we shall visit on our trip route, (with a runners up award given to San Francisco), this is the most geologically active.

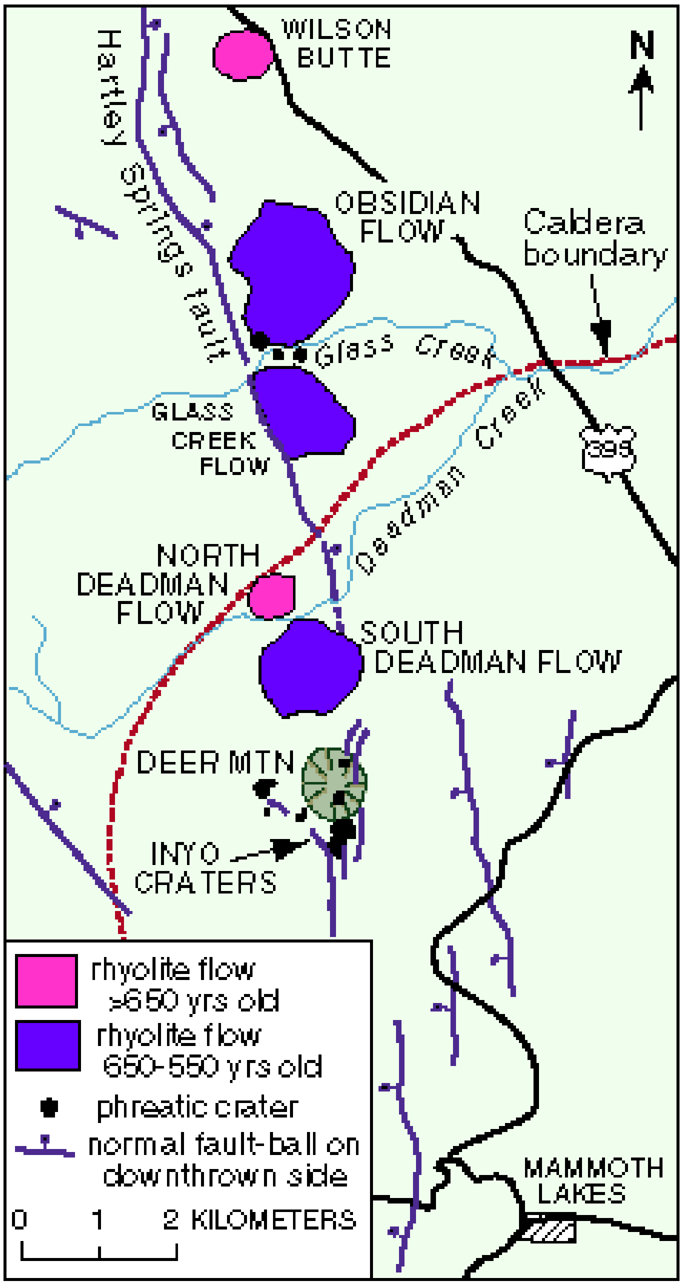


Figure 35 – Geological map of the Inyo Craters area showing the extent and ages of volcanism and fault zones. From: *quake.wr.usgs.gov/VOLCANOES/LongValley/.*

The **Mono Craters** are a 17-km-long chain of rhyolite domes and flows that were erupted from 35,000 to about 600 years ago (Figure 36). All but four of the 24 exposed domes and flows of the Mono Craters are less than 10,000 years old (Figure 37). The most recent eruptive episode occurred between A.D. 1325 and 1365, during which time there were several explosive eruptions and five separate lava flows that oozed onto the surface, including Panum Dome and North Coulee flow (Figures 38, 39). Unless indicated otherwise, the bulk of the material in this segment is derived from ***quake.wr.usgs.gov/VOLCANOES/LongValley/,*** including images and figure captions. See this website for individual photographer credit.

Figure 36 – View to the southwest of Mono Lake showing the Panum Crater (R) and other features. (From Shelton, 1966.)

Extending from eastern California and Nevada, the basin and range province extends into Arizona and bordering Mexico. In a nutshell, the province is dominated by normal fault blocks of upper crust produced by oblique slip and tensional fragmentation (Hamilton and Myers, 1966). The crust has been thinned by tension but thickened by Cenozoic surface volcanism and by intrusion at depth. Renowned for the stark character of the resultant landscape, Basin and Range topography is rooted in Cenozoic crustal extension related to the development of the San Andreas fault system, inland development of horst and graben structure, and coseismic erosion of the ranges and infilling of the basins. In southern California and Arizona the trend of the ranges are NW-SE, essentially parallel to the San Andreas trend.

The plan here is to follow the field trip route suggested by Tierney (1995) and for CM to wing it as he has never set foot in the region before. We will use index (See Figure 34) and geological maps (See Figure 35).

Figure 37 - Aerial view of a lava dome nestled within Panum Crater, which was built by the explosive ejection of tephra. The explosive eruptions also generated three pyroclastic flows and a surge, and built the ring of tephra that surrounds the Crater. The dome was the last lava flow erupted during Mono eruptive episode between A.D. 1325-1365. From http:/quake.wr.usgs.gov/VOLCANOES/LongValley.

Figure 38 – Aerial view towards the north from above the South Coulee (dark, irregular surface at bottom of photo). The light, smooth dome in the center of the photo is mantled with thick layers of tephra. Between the South Coulee and the smooth dome are 4 small explosion craters. Mono Lake is at the top of photo. The hills beyond Mono Lake are the Adobe Hills, a tertiary volcanic complex. From http:/quake.wr.usgs.gov/VOLCANOES/LongValley.

Figure 39 - View from the crest of the Mono Craters is looking toward the north. Mono Lake is visible on the right side of the photograph; the Sierra Nevada are the snow-covered peaks on the left. The original blocky surface of the red-colored lava dome sticks up through tephra erupted from one of the North Coulee vents; a remnant of a crater rim cuts through the photograph. The blocky lava outcrops are also active fumaroles releasing water vapor (steam) and carbon dioxide gas. From http:/quake.wr.usgs.gov/VOLCANOES/LongValley.

Long Valley Caldera (From *quake.wr.usgs.gov/VOLCANOES/LongValley/)*

The Caldera. Long Valley Caldera is a 15- by 30-km oval-shaped depression located 20 km south of Mono Lake along the east side of the Sierra Nevada in east-central California. (See Figures 19 and 20.) This area of eastern California has produced numerous volcanic eruptions over the past 3 million years, including the massive caldera-forming eruption 760,000 years ago. The most recent eruption occurred just 250 years ago in Mono Lake at the north end of Mono-Inyo Craters volcanic chain.

Volcanic Unrest. In May of 1980, a strong earthquake swarm that included four magnitude 6 earthquakes struck the southern margin of Long Valley Caldera associated with a 25-cm, dome-shaped uplift of the caldera floor. These events marked the onset of the latest period of caldera unrest that continues to this day. This ongoing unrest includes recurring earthquake swarms and continued dome-shaped uplift of the central section of the caldera (the resurgent dome) accompanied by changes in thermal springs and gas emissions.

USGS Monitoring. In 1982, the U.S. Geological Survey under the Volcano Hazards Program began an intensive effort to monitor and study geologic unrest in Long Valley caldera. The goal of this effort is to provide residents and civil authorities in the area reliable information on the nature of the potential hazards posed by this unrest and timely warning of an impending volcanic eruption, should it develop. Most, perhaps all, volcanic eruptions are preceded and accompanied by geophysical and geochemical changes in the volcanic system. Common precursory indicators of volcanic activity include increased seismicity, ground deformation, and variations in the nature and rate of gas emissions. Maybe we'll get lucky and experience a temblor.

Segment 4
Yosemite and the Sierra Nevada Range

Segment 4 involves driving up the east slope of the Sierra Nevada range from a few hundred feet in Mono Craters area to elevations in excess of 10,000’ in the vicinity of Tioga Pass (Figure 40), so hold onto your hats. The “discovery” of Yosemite Valley in 1864 created quite a stir and brought attention to the preservation of natural areas for the public. Indian legend had long told of a magic valley nestled high up in the mountains. It took the adventurous spirit of James Mason Hutchings (Figure 41) to discover and popularize the region in his Hutchings Magazine.

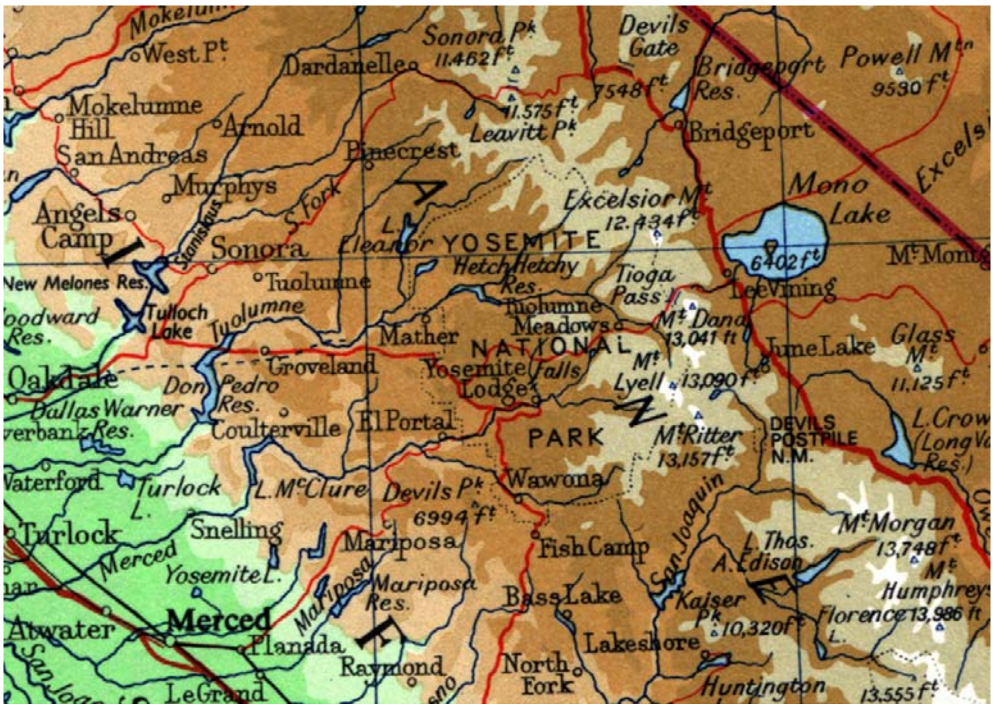


Figure 40 – Index map of Segments 3 and 4.

Discovery of Yosemite Valley

Yosemite Valley, a former mysterious Indian stronghold, was popularized in the October 1859 edition of Hutchins’ California Magazine. After bloody Indian uprisings in 1851 and 1852, word spread through Mariposa or fabulous waterfalls and immense, steep cliffs along long valleys. James M. Hutchings, an Englishman, became intrigued with the stories. Enough so that he engaged an artist, Thomas A. Ayers to assist in preparing illustrations for his new magazine about California. With Indian guides, in June 1855 Hutchings, Ayers, and two companions set foot on the trail to find and document Yosemite Valley. They spent five days in the valley, with Ayers recording the beauty with hand drawings (Figures 42, 43). Three other parties were organized to the same end and by the end of the year (1855) forty-two visitors had seen Yosemite. Fifty years later 10,000 visitors came and after a century the number of visitors had topped a million.

Figure 41 – James Mason Hutchings, miner and publisher of “The California Magazine” circa 1900.

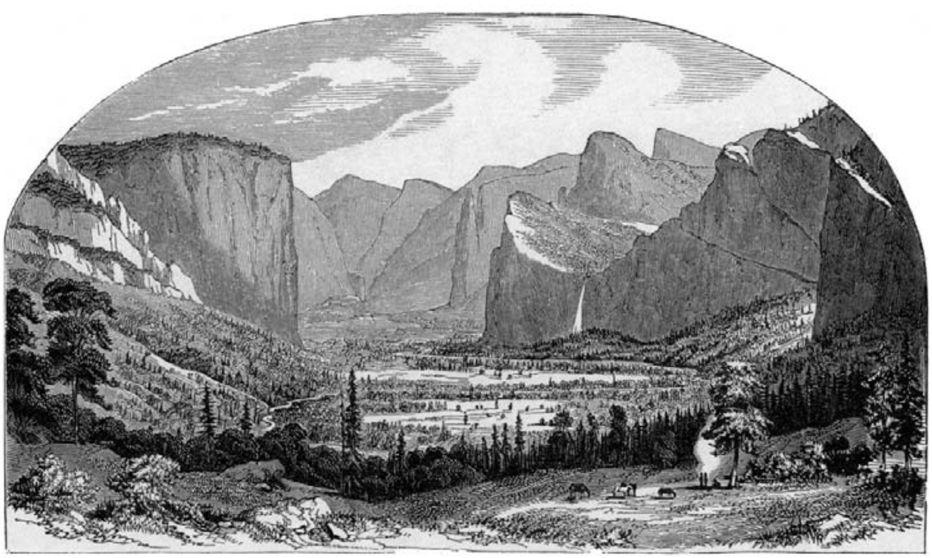


Figure 42 – View of the Yo-Semite Valley from Open-eta-noo-ah (Inspiration Point) on the Mariposa trail. Hutchings California Magazine, October 1859.

Figure 43 – View of the Yo-Semite Falls. Hutchings California Magazine, October 1859.

Geology of the Yosemite Valley

Details on the igneous geology of the Sierra Nevada have been extensively discussed above. (See Figures 17, 18; Table 1.) The geomorphic development of the landforms are covered in this section. A relief map of the region shows as a long gash in a dissected mountainous region with a flat floor (Figure 44).

Figure 44 – Shaded relief map of the Yosemite Valley.

The valley was formed by the combined effects of rapid downcutting of the Merced river and modification of that channel by a valley glacier. Figure 45, which I have digitally rearranged and enhanced, shows six stages in the development of the valley from Matthes (1950). The **Broad Valley Stage** (A) ended the Miocene the result of uplift to the east and stream erosion. The ancient Merced river occupies the base of a valley less than 1,000' deep. The **Mountain-valley Stage** (B) ended the Pliocene and witnessed rapid uplift and subsequent downcutting of the Merced to produce a steep walled valley 1,800' deep that featured hanging valleys.

The **Canyon Stage** (C) was in the early Pleistocene when a third and largest ratchet upward of the Sierran block took place along the eastern border faults. The Merced cut a 3,000' canyon in a raging torrent of Sierran meltwater and many hanging valleys were formed. The Ice age was starting and cirque glaciers may have formed. The **El Portal Glacial Stage** (D) saw a thick valley glacier flow through Yosemite and create a broad U-shaped valley (Figure 46). All of the northern uplands with the exception of El Capitan and Eagle Peak were ice covered. Half Dome extended 700' above the ice flow (Figure 47).

During the **Wisconsinan Glacial Stage** (E) valley glaciers returned to Yosemite but were not as thick as the previous stage and did not overtake the valley. The glacier descended to about the position of Brideveil Falls, based on the position of the moraine. This event modified the valley to its present dimension. The Modern Stage (F) witnessed the formation of ancient lake Yosemite, fed by glacial meltwater. Over time, coarse sediment from the Merced river built out a delta and eventually filled the lake to produce its present flat valley floor.

An index map to the features of Yosemite Valley (Figure 48) will be useful as we examine the rock structures, sentinels of time and geological change.

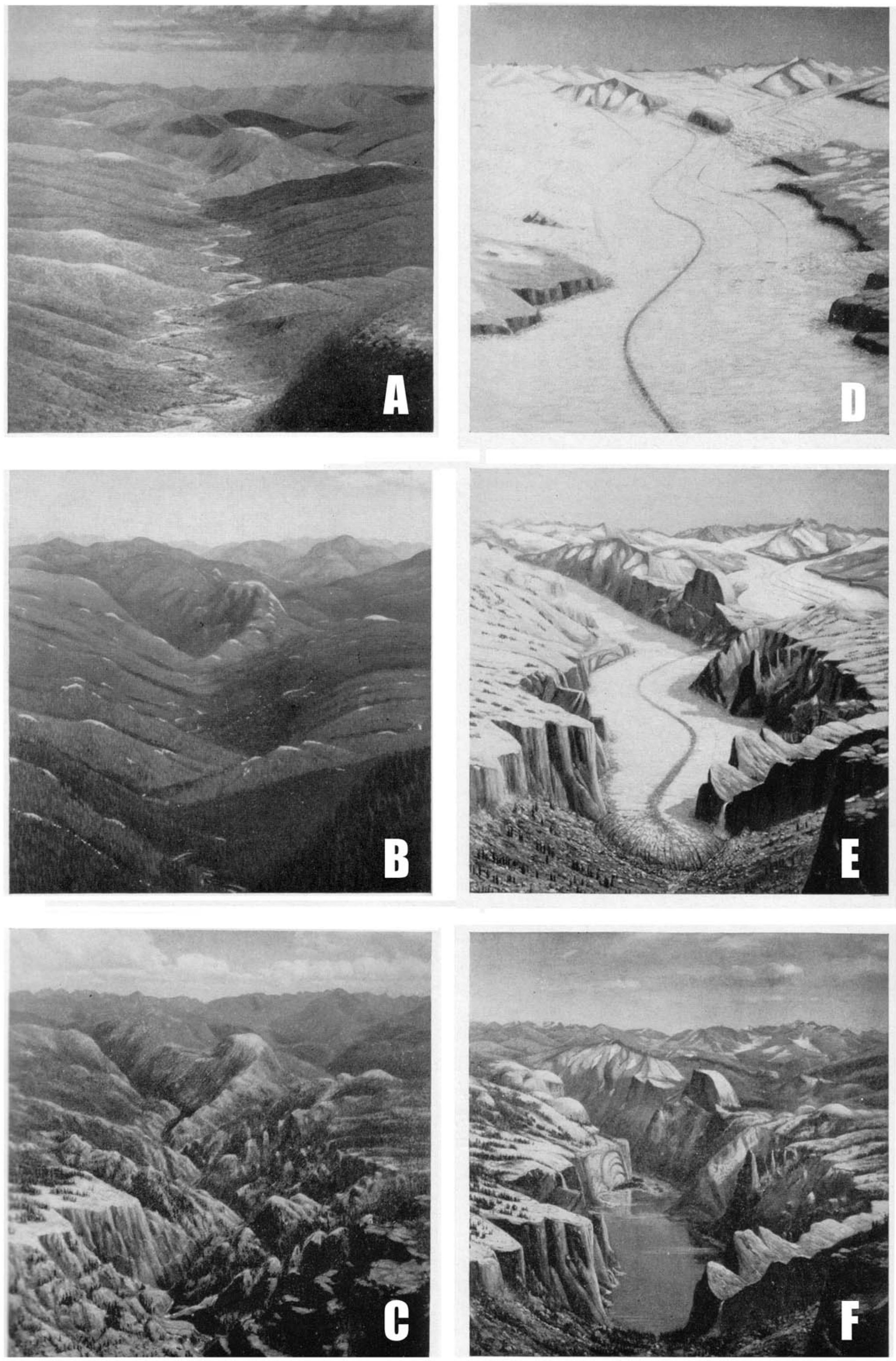


Figure 45 – Composite diagram showing the six stages in the formation of Yosemite Valley. (From Matthes, 1950, Plates 2-7.)

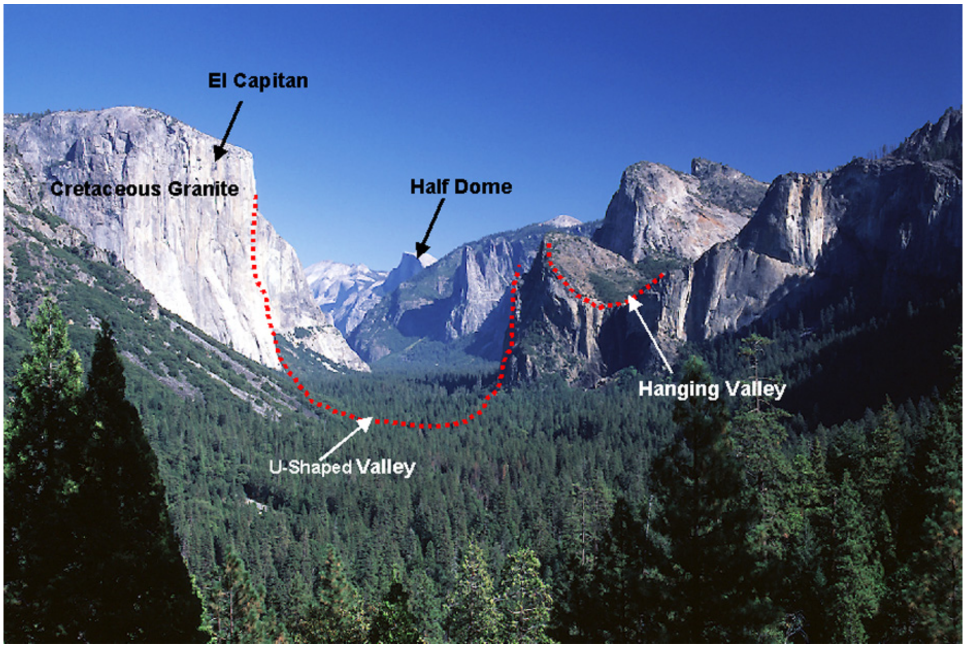
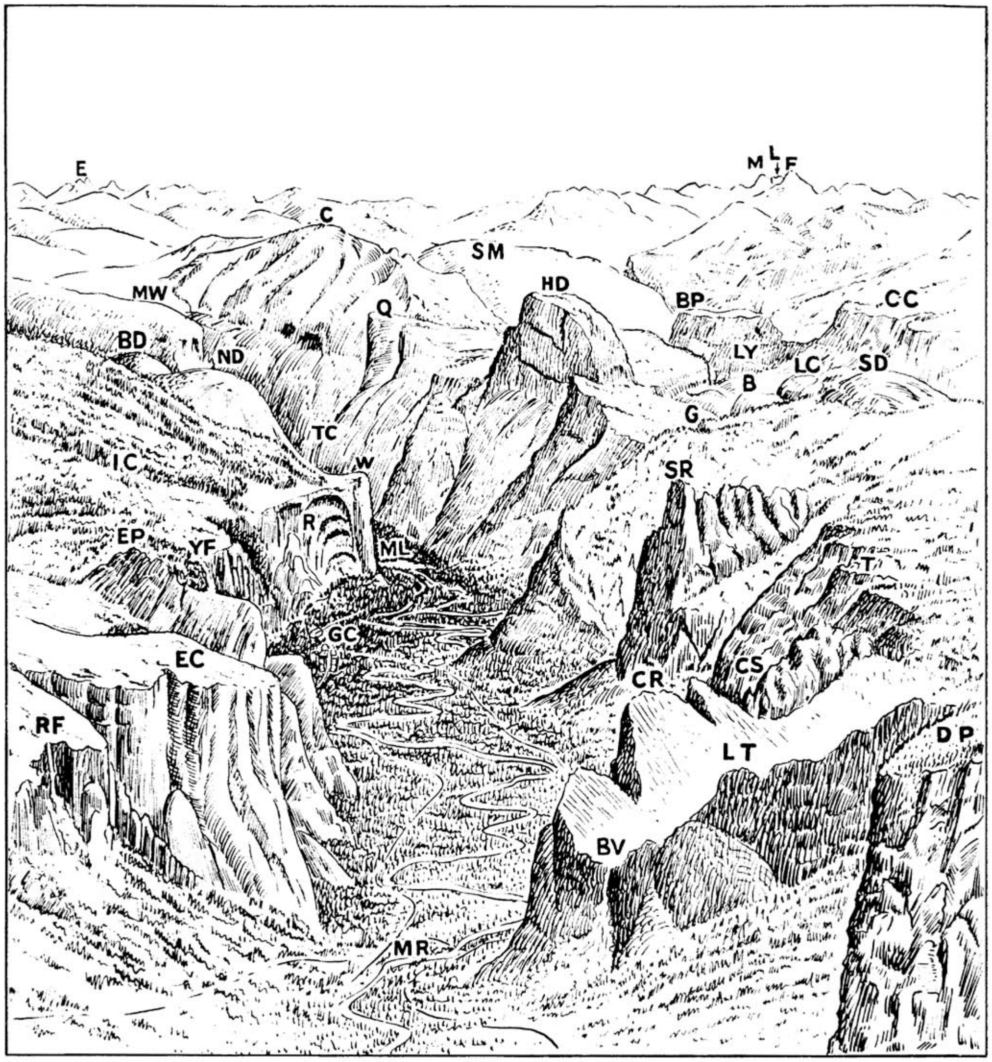


Figure 46 – View of Yosemite Valley showing the broad U-Shaped valley form created by glacial ice. Note the hanging valley to the right, the result of truncation of a former drainage by the ice. (From Harris and Tuttle, 2002.)

Figure 47 – Aerial view toward the WNW of Half Dome and the surrounding peaks of the Sierra with the Tuolumne River drainage visible in the distance. Half Dome extended 700' above the top of the El Portal glacier. (From Harris and Tuttle, 2002.)



RF	Ribbon Fall	MW	Mount Watkins	LC	Liberty Cap
EC	El Capitan	E	Echo Peaks	B	Mount Broderick
EP	Eagle Peak	C	Clouds Rest	SD	Sentinel Dome
YF	Top of Yosemite Falls	SM	Sunrise Mountain	G	Glacier Point
GC	Government Center	Q	Quarter Domes	SR	Sentinel Rock
IC	Indian Creek	HD	Half Dome	T	Taft Point
R	Royal Arches	M	Mount Maclure	CS	Cathedral Spires
W	Washington Column	L	Mount Lyell	CR	Cathedral Rocks
TC	Tenaya Canyon	F	Mount Florence	BV	Bridalveil Fall
ML	Mirror Lake	BP	Bunnell Point	LT	Leaning Tower
ND	North Dome	CC	Cascade Cliffs	DP	Dewey Point
BD	Basket Dome	LY	Little Yosemite	MR	Merced River

Figure 48 – Index map of the various geological and other features of Yosemite Valley. View to the west. (From Matthes, 1930, 1950.)

The Sierra Nevada Batholith

The Foothills Metamorphic belt is truncated by plutonic rocks of the Sierra Nevada batholith. (See Figure 13.) The foothills belt was intruded by a number of isolated middle Jurassic plutons that were early harbingers of the massive intrusive series that was to follow in late Jurassic through Cretaceous time. The Sierran granites dominate the landscape of California forming imposing mountains of eroded granitoid rock sculpted by rapid uplift and subsequent glaciation. During this interval the Tuolumne Intrusive series and related rocks were emplaced as a series of coalescent plutons the invaded an area over 40 miles in width (Figure 49). The coarse-grained igneous rocks include granite and granodiorite mostly but also include intermediate, mafic, and ultramafic phases (listed in order of decreasing abundance). Uplift and erosion of the arc sequence of the Sierra Nevada, a former Andean or marginal volcanic arc has exposed the plutonic roots of that arc (Figure 50). The details of the geology of this sequence will be developed below in the actual trip description section.

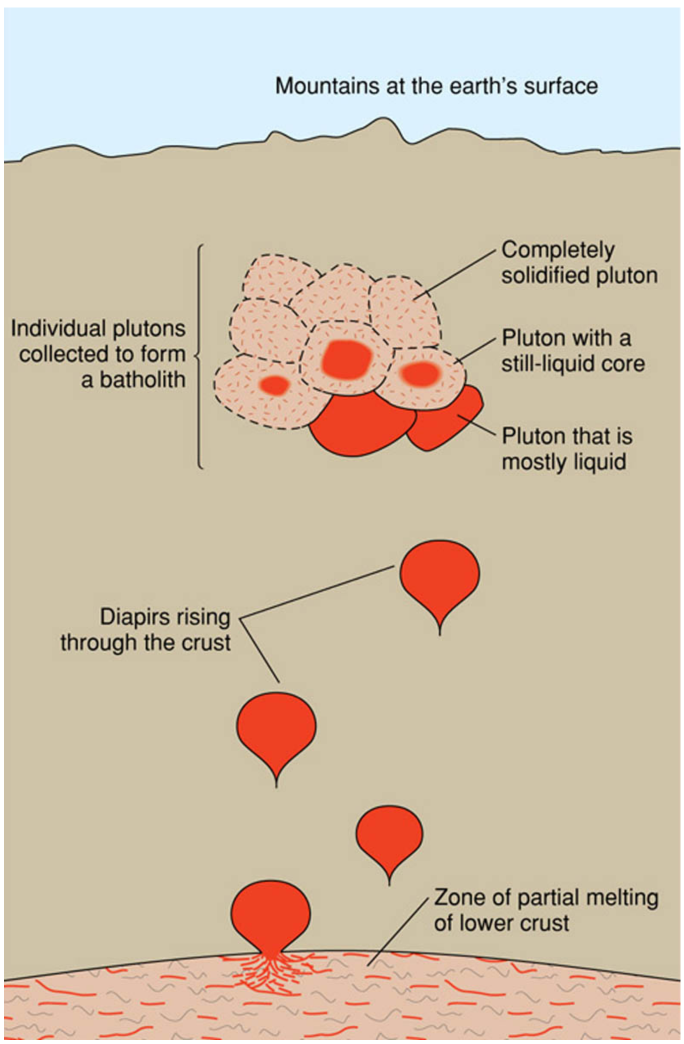


Figure 49 – Development of a batholith in the lithosphere, the result of coalesced diapers of magma that rise from the lower crust and solidify in the upper crust. (From Plummer and McGeary, 2002.)

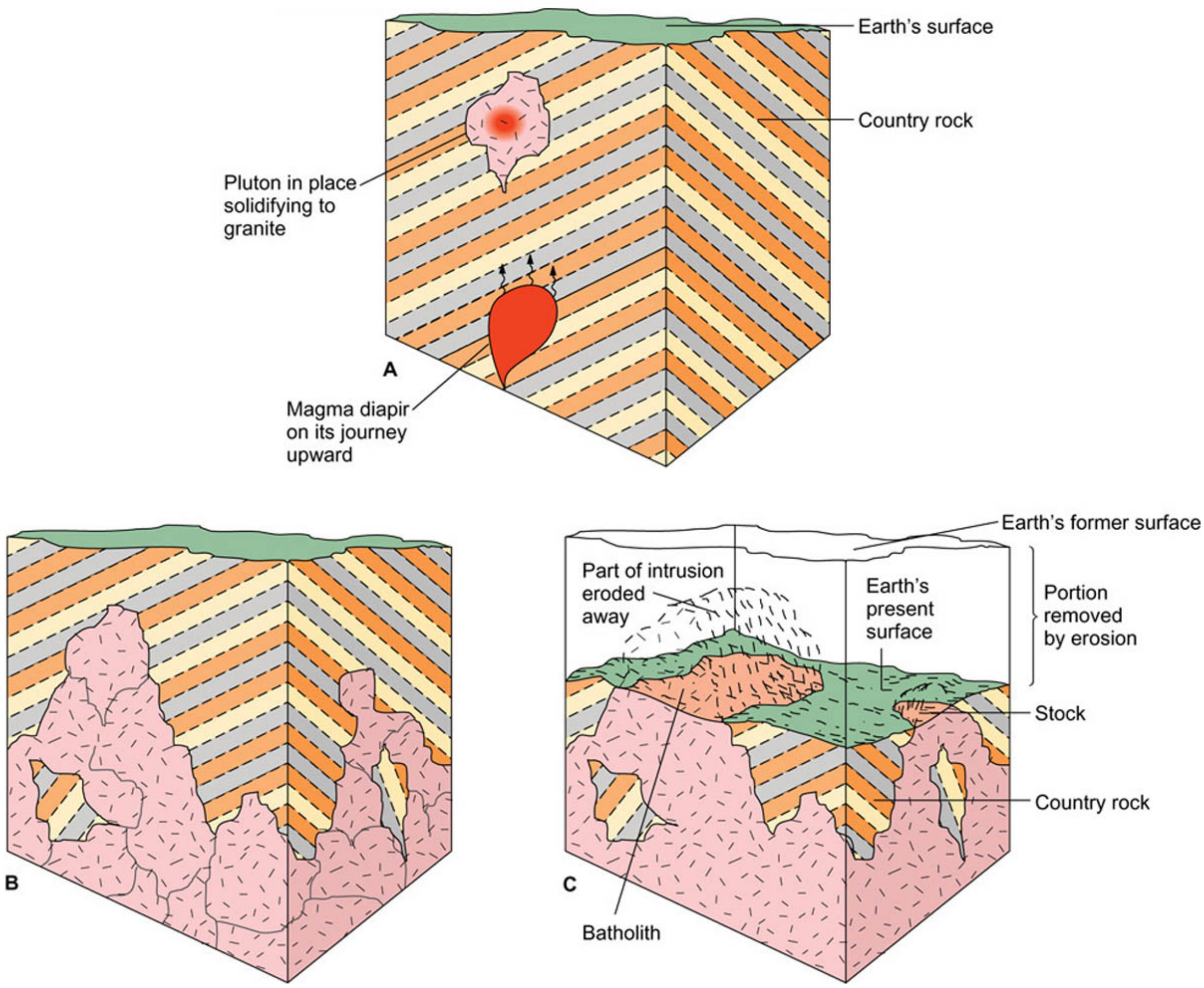


Figure 50 – Composite diagram showing (A) an early stage of intrusion of numerous plutons into the country rock, (B) an intermediate stage where other magma diapers have intruded, coalesced, and crystallized into a solid mass of plutonic rock. After uplift and erosion, the surface expression of batholiths, stocks, roof pendants, screens, and xenoliths results. (From Plummer and McGeary, 2002.)

A geological map of the high Sierra in the vicinity of Yosemite National Park is reproduced as Figure 51. Note the pattern of individual plutons showing cross-cutting relationships and the presence of roof pendants in the central portion of the range. Detailed mapping to establish crosscutting relationships and targeted age dating has resulted in definition of the intrusive pulses of magma that formed the Sierra Nevada batholith in the 32 Ma interval 116 to 84 Ma (Table 1). Middle Jurassic plutonism at roughly 170 Ma (Standard pluton) marks an early stage in the eventual development of the batholithic series to the east.

As we cross over the Sierra at Tioga Pass (Elev. 9,941') we cross over the drainage divide of the Sierra with water falling west of there going to the Pacific. To the east, water enters the Great Basin, the fourth major geological area of our trip route where Cenozoic and Recent tectonics have dominated the geology. Because this creates unusually stark landscapes, we are blessed to have the opportunity to visit this area.

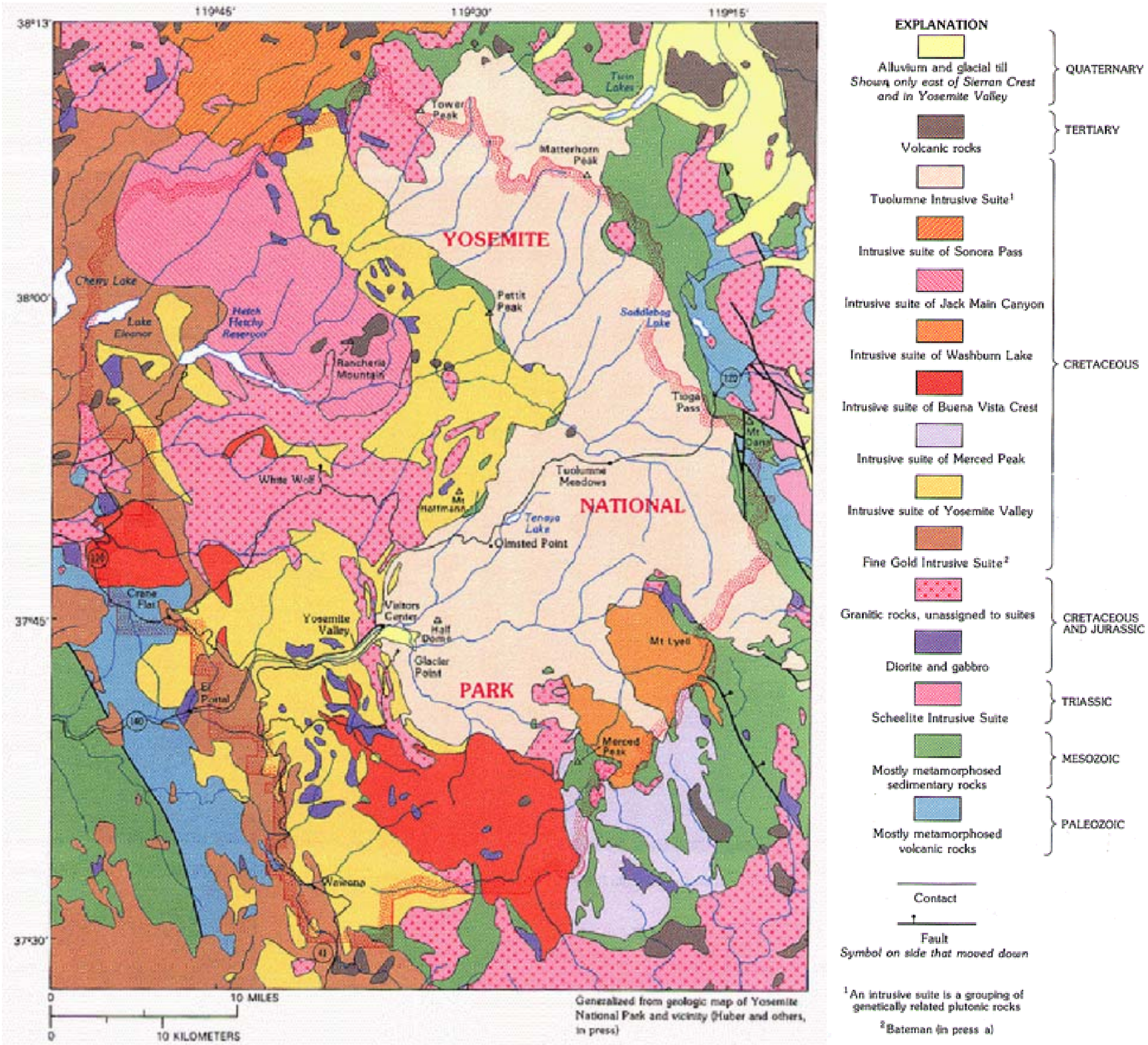


Figure 51 – Geologic map and explanation of Yosemite National Park and vicinity showing the broad zone of intrusive activity and the minor amount of metasediment (green) and volcanic rocks (brown) still preserved as pendants and erosional remnants.

Table 1 - PLUTONIC ROCK UNITS OF YOSEMITE REGION

Tuolumne Intrusive Suite

JOHNSON GRANITE PORPHYRY- 84 Ma? Fine-grained, equigranular granite, locally contains sparse K-feldspar phenocrysts, miarolitic cavities.

CATHEDRAL PEAK GRANODIORITE - 86 - 88 Ma. Medium-grained biotite granodiorite, conspicuous blocky K-feldspar megacrysts - poikilitic.

HALF DOME GRANODIORITE - 88 - 90 Ma. Medium-grained biotite-hornblende granodiorite. Equigranular facies- large euhedral hornblende and biotite, abundant sphene. Porphyritic facies- conspicuous K-feldspar phenocrysts - poikilitic, large euhedral hornblende, seriate texture, abundant sphene.

TONALITE OF GLEN AULIN - 90 - 91 Ma. Dark-gray, variable composition. Fine-grained quartz diorite (west) to medium-grained granodiorite (east).

TONALITE OF GLACIER POINT - 90 - 91 Ma. Medium- to dark-gray, medium-grained, hornblende-biotite tonalite. Conspicuous, medium- to coarse-grained biotite, minor sphene.

GRANODIORITE OF GRAYLING LAKE - 90 - 91 Ma. Fine- to medium-grained, biotite-hornblende granodiorite, abundant sphene, well-foliated.

GRANODIORITE OF KITNA CREST - 90 - 91 Ma. Medium- to dark-gray, medium-grained, hornblende-biotite granodiorite/quartz monzodiorite.

Outboard of Tuolumne Intrusive Suite, but possibly related

SENTINEL GRANODIORITE - 96 Ma. Light- to medium-gray, medium-grained, biotite-hornblende granodiorite, euhedral hornblende, abundant sphene, equigranular to seriate.

GRANODIORTTE OF YOSEMITE CREEK - ? Ma. Dark-gray, medium- to coarse-grained, hornblende-biotite granodiorite to tonalite, locally contains plagioclase phenocrysts.

Intrusive Suite of Buena Vista Crest

BRIDALVEIL GRANODIORITE - ? Ma. Medium-gray, fine-grained, hornblende-biotite granodiorite, granular, `salt and pepper' appearance.

LEANING TOWER GRANITE - 100 Ma. Medium-gray, medium-grained, hornblende-biotite granite, granular, small (<1 cm) clusters of hornblende and biotite -'speckled' appearance.

Intrusive Suite of Yosemite Valley

NORTH AMERICA DIORITE - 103 Ma. Dark-gray, Fine-grained biotite-hornblende quartz diorite. Rich in hornblende.

TAFT GRANITE - 103 Ma. Very light-gray to white, fine- to medium grained, biotite granite to granodiorite. ~quartz – plagioclase - K-feldspar. Biotite scarce. Locally coarse-grained and porphyritic - K-feldspar.

DIORITE OF THE ROCKSLIDES - 104 Ma. Very dark-gray, fine- to medium-grained, biotite-hornblende gabbro to diorite, weakly porphyritic, and coarse-grained, equigranular, hornblende-biotite quartz diorite to tonalite. Conspicuous hornblende.

EL CAPITAN GRANITE - 104 Ma. Light-gray to white, medium- to coarse-grained, equigranular, biotite granite to granodiorite. Locally porphyritic - blocky K-feldspar, conspicuous quartz.

Finegold Intrusive Suite

BASS LAKE TONALITE/ TONALITE OF THE GATEWAY - 114 Ma. Dark-gray, medium-grained, biotite-hornblende tonalite. Locally varies to granodiorite and quartz diorite.

GRANODIORITE OF ARCH ROCK - 116 Ma. Medium- to light gray, medium-grained, biotite granodiorite. Poikilitic K-feldspar.

From Bateman et al. (1983), Bateman (1992), Calkins et al. (1985), Huber et al. (1989), Kistler (1973), Peck (1980), and Ratajeski et al. (2001). Ages are U-Pb ages from zircon and sphere. Ages from Bateman (1992), Coleman and Glazner (1997), Huber et al. (1989), Ratajeski et al. (2001), and Stern et al. (1981).

Appendix 1 - Geologic Structure - a Primer

Geologists use terminology to confuse the layman and to enable them to amass a huge library of terms that are undeniably useless in most social situations. Luckily, our Geology classes and field trips are an exception. We will not try to bury you in a mountain (how about a deeply eroded mountain range?) of terms to help you understand the major types of structures and geologic features that you will read- and hear about today. But, if you are to understand what we are talking about, you need to know some important definitions. In the following section, we describe folds, faults, surfaces of unconformity, sedimentary structures, structures in sedimentary- vs. metamorphic rocks, and tectonostratigraphic units.

We begin with some concepts and definitions based on the engineering discipline known as **strength of materials**. Given today's sophisticated laboratory apparatus, it is possible to subject rocks to temperatures- and pressures comparable to those found deep inside the Earth.

Imagine taking a cylinder of rock out of the Earth and torturing it in a tri-axial compression machine to see what happens. Some geologists get a big charge out of this and tell us (the field geologists) that they really understand how rocks behave under stress. [CM thinks they need to perform these experiments over a longer time frame than a few generations of siblings will allow and thus relies more on field observation and inference than from rock-squeezing data to gain a feel for the complex nature of how rocks are deformed in nature.]

Despite the limitations of the experimental work, measurements in the laboratory on specimens being deformed provide some fundamental definitions. One key definition is the **elastic limit**, which is the point at which a test specimen no longer returns to its initial shape after the load has been released. Below the elastic limit, the change of shape and/or volume (which is known as **strain**) is proportional to the stress inside the specimen. Above the elastic limit, the specimen acquires some permanent strain. In other words, the specimen has "failed" internally. Irrecoverable strain manifests itself in the distortion of crystal lattices, grain-boundary adjustments between minerals composing the rock, and minute motions along cleavage- or twin planes.

When differential force is applied slowly (or, according to CM, over long periods of time), rocks fail by *flowing*. This condition is defined as behaving in a **ductile fashion** (toothpaste being squeezed out of a tube is an example of ductile behavior). Folds are the result of such behavior. If the force is applied under low confining pressure or is applied rapidly (high strain rates), rocks do not flow, but *fracture*. This kind of failure is referred to as rocks behaving in a **brittle fashion** (as in peanut brittle). The result is faults or joints. Once a brittle failure (fracture) has begun, it will propagate and may produce offset thus forming a fault surface. Joint surfaces commonly exhibit distinctive "feathers" which show the direction of joint propagation.

In some cases, during deformation, rocks not only undergo simple strain, but also recrystallize. New metamorphic minerals form and newly formed metamorphic minerals acquire a parallel arrangement. More on metamorphic textures later. From the laboratory studies of rock deformation, a few simple relationships are generally agreed upon regarding brittle- and ductile faulting and these are discussed below.

When subjected to differential forces, under high confining pressures and elevated temperatures, rocks (like humans) begin to behave foolishly, squirming in many directions and upsetting the original orientation of primary- or secondary **planar- and linear features** within them. Geologists try to sort out the effects of deformation by working out the order in which these surfaces or linear features formed using a relative nomenclature based on four letters of the alphabet: D, F, S, and M. Episodes of deformation are abbreviated by (D_n), of folding by (F_n), of the origin of surfaces (such as bedding or foliation) by (S_n), and of metamorphism by (M_n), where n is a whole number starting with 1 (or in some cases, with zero). Bedding is commonly designated as S_0 (or surface number zero) as it is commonly overprinted by S_1 (the first foliation). To use this relative nomenclature to describe the structural history of an area, for example, one might write: "During the second deformation (D_2), F_2 folds formed; under progressive M_1 metamorphic conditions, an axial-planar S_2 foliation developed."

In dealing with the geologic structures in sedimentary rocks, the first surface one tries to identify positively is **bedding** or **stratification**. The boundaries of strata mark original sub-horizontal surfaces imparted to sediments in the earliest stage of the formation of sedimentary rock. Imagine how such strata, buried by the weight of overlying strata and laterally compressed by the advance of lithospheric plates, are subjected to the differential force necessary for folds to form. Contrary to older ideas, we now realize that vertical burial cannot cause regional folds (although small-scale slumping, stratal disharmony, and clastic dikes are possible). Rather, resolved tangential force that creates differential stress must be applied to provide the driving force to bring about folds and faults.

It's now time to turn to some geometric aspects of the features formed as a result of deformation of rocks in the Earth. We start with folds.

Folds

If layers are folded into convex-upward forms we call them **anticlines**. Convex-downward fold forms are called **synclines**. In Figure A1-1, note the geometric relationship of anticlines and synclines. **Axial planes** (or **axial surfaces**) physically divide folds in half. Note that in Figure A1-1, the fold is deformed about a vertical axial surface and is cylindrical about a linear **fold axis** which lies within the axial surface. The locus of points connected through the domain of maximum curvature of the bedding (or any other folded surface of the fold) is known as the **hinge line** (which is parallel to the fold axis). This is geometry folks; we have to keep it simple so geologists can understand it.

In eroded anticlines, strata forming the limbs of the fold *dip away from* the central hinge area or core (axis) of the structure. In synclines, the layers forming the limbs *dip toward the hinge area*. Given these arrangements, we expect that in the arches of eroded anticlines, older stratigraphic layers will peek through whereas in the eroded troughs of synclines, younger strata will be preserved.

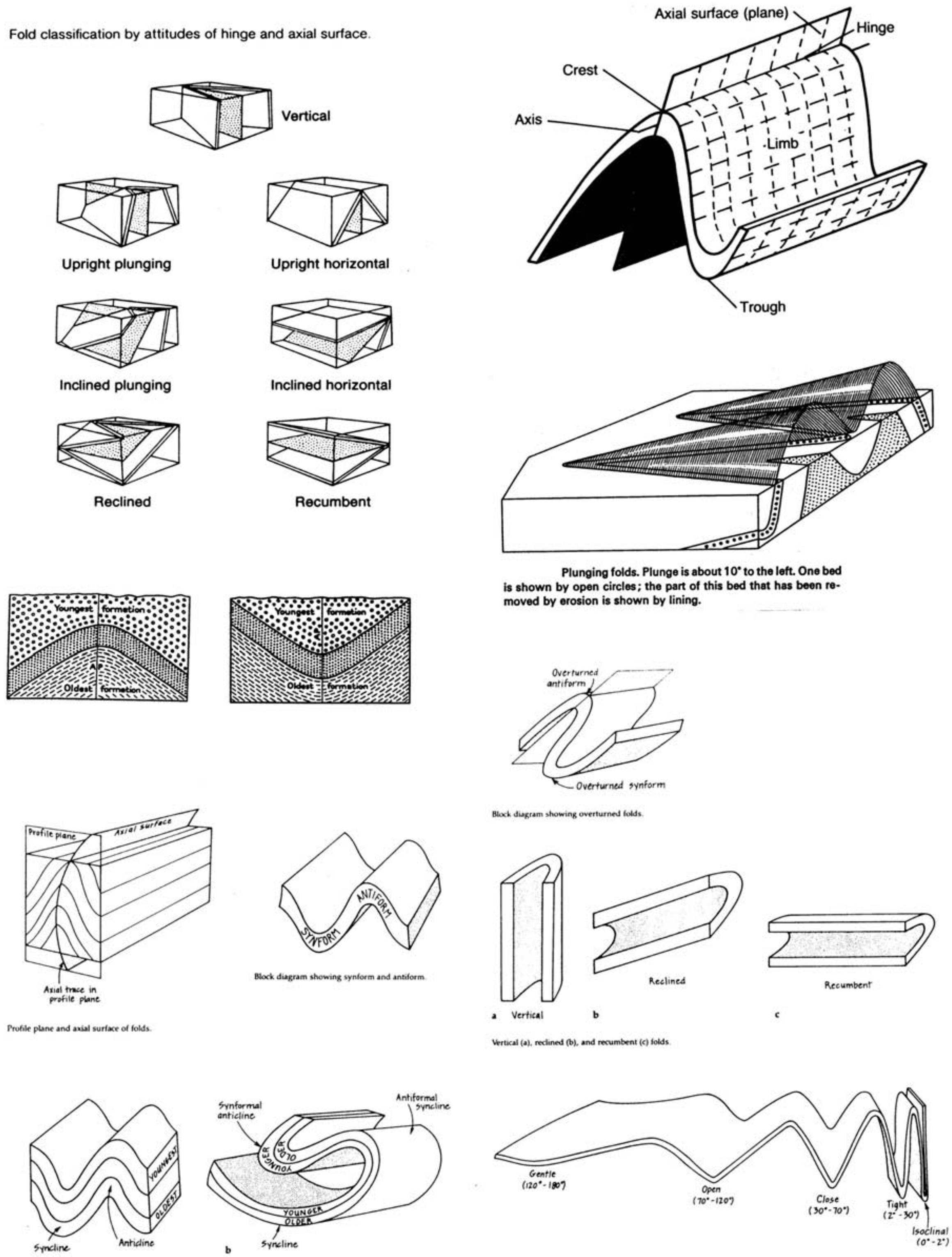


Figure A1-1 - Composite diagram from introductory texts showing various fold styles and nomenclature as discussed in the text.

In metamorphic terranes, field geologists are not always sure of the correct age relationships of the metamorphosed strata. Therefore, it is helpful to make use of the general terms **antiform** and **synform** which describe the folds by whether they are convex upward (antiform) or concave upward (synform) but do not imply anything about the relative ages of the strata within them.

Realize that in the upright folds shown in Figure A1-1, axial surfaces are vertical and fold axes, horizontal. Keep in mind that folding under metamorphic conditions commonly produces a penetrative mineral fabric with neocrystallized minerals (typically micas and amphiboles) aligned parallel to the axial surfaces of folds. Such penetrative metamorphic fabrics are called **foliation**, if primary, and **schistosity**, if secondary. Minerals can also become aligned in a linear fashion producing a **metamorphic lineation**. Such features can be useful in interpreting a unique direction of tectonic transport or flow direction. Because folds in metamorphic rocks are commonly **tight-** to **isoclinal** (high amplitude-to-wavelength aspect ratio) with limbs generally parallel to axial surfaces, a penetrative foliation produced during regional dynamothermal metamorphism will generally be parallel to the re-oriented remnants of stratification (except of course in the hinge areas of folds). Thus, in highly deformed terranes, a composite foliation + remnant compositional layering is commonly observed in the field. Departures from this common norm are important to identify as they tend to mark regional fold-hinge areas.

Folds could care less about the orientation of their axes or axial surfaces and you can certainly imagine that axial surfaces can be tilted, to form inclined or **overturned folds**. Or the axial surfaces may be sub-horizontal, in which case the term **recumbent folds** is used. In both overturned folds and recumbent folds, the fold axes may remain subhorizontal. (See Figure A1-1.) It is also possible for an axial surface to be vertical but for the orientation of the fold axis to range from horizontal to some angle other than 0° (thus to acquire a plunge and to produce a **plunging fold**). Possible configurations include plunging anticlines (or -antiforms) or plunging synclines (or -synforms). **Vertical folds** (plunging 90°) are also known; in them, the terms anticline and syncline are not meaningful. In **reclined folds**, quite common in ductile shear zones, the fold axes plunge directly down the dip of the axial surface.

In complexly deformed mountain ranges, most terranes show the superposed effects of more than one set of folds and faults. As a result of multiple episodes of deformation, the ultimate configuration of folds can be quite complex (i. e., plunging folds with inclined axial surfaces and overturned limbs).

We need to mention one additional point about the alphabet soup of structural geology. Seen in cross section, folds fall into one of three groups, the S's, the M's, and the Z's. Looking down plunge in the hinge area of a northward-plunging anticlinal fold, for example, dextral shearing generates asymmetric Z folds on the western limb and sinistral shearing forms S folds on the eastern limb. Usually only one variety of small, asymmetric folds will be found on a given limb of a larger fold. Therefore, if one notices a change in the pattern from S folds to Z folds (or vice versa), one should be on the lookout for a fold axis. The hinge area is dominated by M folds (no sense of asymmetry).

One final note on folding -- it is generally agreed, in geologically simple areas, that axial surfaces form perpendicular to the last forces that ultimately produced the fold. Therefore, the orientation of the folds give some hint as to the direction of application of the active forces (often a regional indicator of relative plate convergence). In complex regions, the final regional orientation of the structures is a composite result of many protracted pulses of deformation, each with its unique geometric attributes. In these instances, simple analysis is often not possible. Rather, a range of possible explanations for a given structural event is commonly presented.

Faults

A **fault** is defined as *a fracture along which the opposite sides have been displaced*. The surface of displacement is known as the fault plane (or fault surface). The enormous forces released during earthquakes produce elongate gouges within the fault surface (called **slickensides**) that may possess asymmetric linear ridges that enable one to determine the relative motion between the moving sides (Figure A1-2, inset). The block situated below the fault plane is called the **footwall block** and the block situated above the fault plane, the **hanging-wall block**. Extensional force causes the hanging-wall block to slide **down** the fault plane producing a **normal fault**. [See Figure A1-2 (a).] Compressive forces drive the hanging-wall block **up** the fault plane to make a **reverse fault**. A reverse fault with a low angle (<30°) is called a **thrust fault**. [See Figure A1-2 (b).] In all of these cases, the slickensides on the fault will be oriented more or less down the dip of the fault plane and the relationship between the tiny "risers" that are perpendicular to the striae make it possible to determine the relative sense of motion along the fault. Experimental- and field evidence indicate that the asymmetry of slickensides is not always an ironcled indicator of relative fault motion. As such, displaced geological marker beds or veins are necessary to verify relative offset. Fault motion up- or down the dip (as in normal faults, reverse faults, or thrusts faults) is named **dip-slip motion**.

Rather than simply extending or compressing a rock, imagine that the block of rock is sheared along its sides (*i. e*., that is, one attempts to rotate the block about a vertical axis but does not allow the block to rotate). This situation is referred to as a shearing couple and could generate a **strike-slip fault**. [See Figure A1-2 (c).] On a strike-slip-fault plane, slickensides are oriented subhorizontally and again may provide information as to which direction the blocks athwart the fault surface moved.

Two basic kinds of shearing couples and/or strike-slip motion are possible: **left lateral** and **right lateral**. These are defined as follows. Imagine yourself standing on one of the fault blocks and looking across the fault plane to the other block. If the block across the fault from you appears to have moved to the left, the fault is **left lateral** [illustrated in Figure A1-2 (c)]. If the block across the fault appears to have moved to the right, the motion is **right lateral**. Convince yourself that no matter which block you can choose to observe the fault from, you will get the same result! Naturally, complex faults show movements that can show components of dip-slip- and strike-slip motion, rotation about axes perpendicular to the fault plane, or reactivation in a number of contrasting directions or variety. This, however, is no fault of ours.

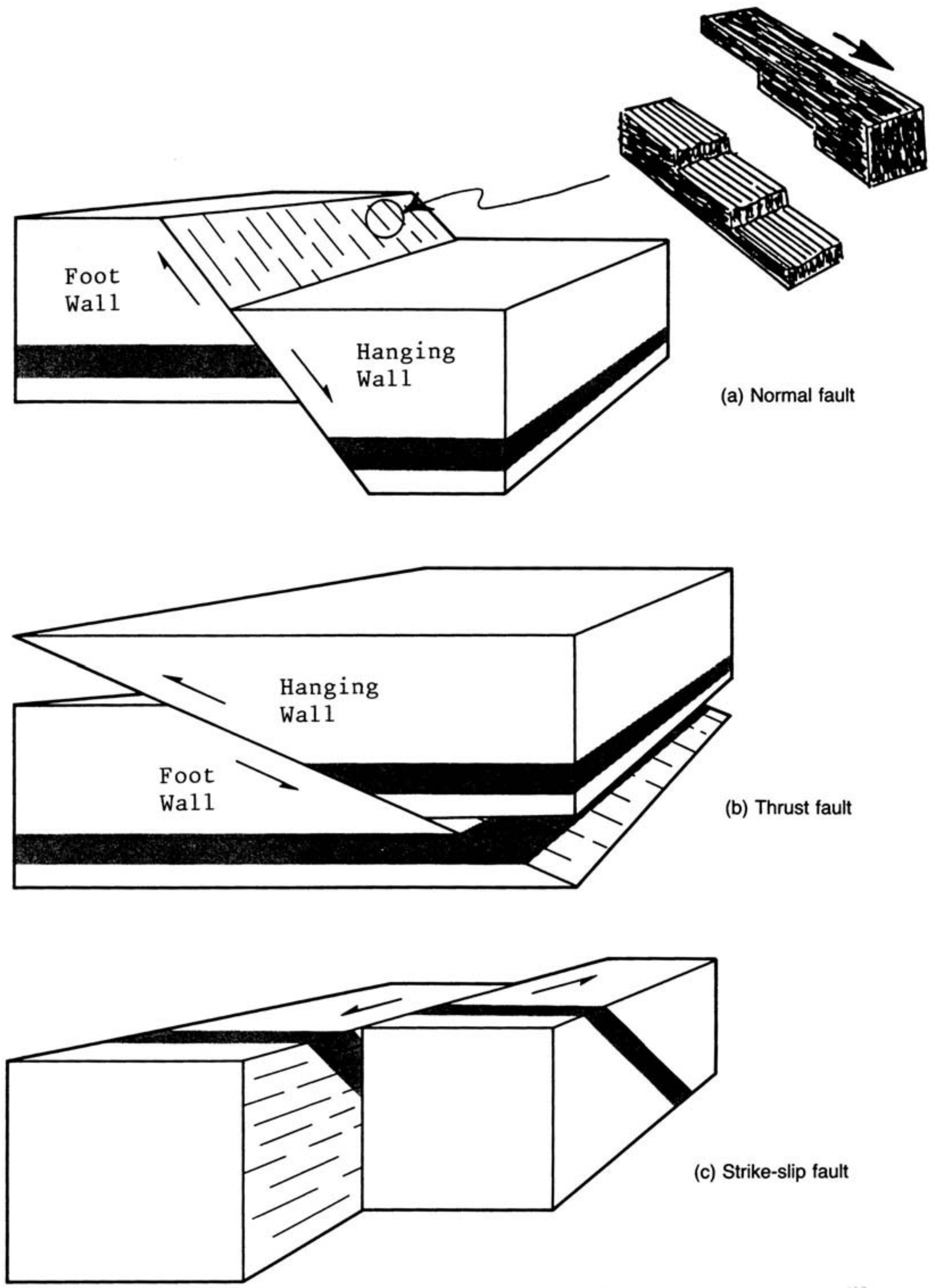


Figure A1-2 - The three main types of faults shown in schematic blocks. Along a normal fault (a) the hanging-wall block has moved relatively downward. On a thrust fault (or reverse fault) (b) the hanging-wall block has moved relatively upward. Along a strike-slip fault (c), the vertical reference layer (black) has been offset by horizontal movement (left-lateral offset shown here). Inset (d) shows segments of two blocks along a slickensided surface show how the jagged "risers" of the stairsteps (formed as pull-apart tension fractures) can be used to infer sense of relative motion. [(a), (b), (c), Composite diagram from introductory texts; (d), J. E. Sanders, 1981, fig. 16.11 (b), p. 397.]

Tensional- or compressional faulting resulting from brittle deformation, at crustal levels above 10 to 15 km, is accompanied by seismicicity and the development of highly crushed and granulated rocks called **fault breccias** and **cataclasites** (including fault gouge, fault breccia, and others). Figure A1-3 lists brittle- and ductile fault terminology as adapted from Sibson (1977) and Hull et al. (1986). Begining at roughly 10 to 15 km and continuing downward, rocks under stress behave aseismically and relieve strain by recrystallizing during flow. These unique metamorphic conditions prompt the development of highly strained (ribboned) quartz, feldspar porphyroclasts (augen), and frayed micas, among other changes, and results in highly laminated rocks called **mylonites** (Figure A1-3).

The identification of such ductile fault rocks in complexly deformed terranes can be accomplished only by detailed mapping of metamorphic lithologies and establishing their geometric relationship to suspected mylonite zones. Unfortunately, continued deformation under load often causes early formed mylonites to recrystallize and thus to produce annealed mylonitic textures (Merguerian, 1988), which can easily be "missed" in the field without careful microscopic analysis. Cameron's Line, a recrystallized ductile shear zone showing post-tectonic brittle reactivation, is an original ductile fault zone (mylonite) having a complex geologic history.

Over the years, field geologists have noted special geologic features associated with thrust faults. Because they propagate at low angles with respect to bedding, thrusts commonly duplicate strata. In addition, thrust faults can displace strata for great distances and wind up transporting rock deposited in one environment above rocks deposited in markedly disparate environments. In such cases, we call the displaced strata of the upper plate above a thrust fault an **allochthon** or describe an entire displaced sequence of strata as an **allochthonous terrane** (see Tectonostratigraphic Units below). In other words, *allochthonous rocks were not originally deposited where they are now found*. By contrast, regions consisting of rock sequences that were originally deposited where they are now found constitute an **autochthon** or **autochthonous terrane**.

Interesting geometric patterns result from the erosion of overthrust sheets of strata that have been folded after they were overthrust. When the upper plate (allochthon) has a "hole" eroded through it, we can peer downward through the allochthon and see the autochthon exposed in a **window**, **inlier**, or **fenster** surrounded by the trace of the thrust fault that was responsible for the dislocation (Figure A1-4). By contrast, if most of the upper plate has been eroded, only a remnant **outlier** or **klippe** may remain. (See Figure A1-4.) Both klippen and windows produce similar map-scale outcrop patterns. The difference is that the thrust surface typically dips *toward* the center of a klippe (a remnant of the allochthon) and *away from* the center of window (which shows a part of the underlying autochthon).

Bedding-plane thrusts are more-localized features but are geometrically the same as thrust faults in that they involve layer-parallel shortening of strata and produce low-angle imbrication of strata. They can easily be "missed" in the field but result in overthickening of strata and can produce anomalous stratigraphic thickness in sedimentary units. The field geologist can identify them by careful bed-by-bed examination of known sequences based on duplication of key- or marker beds and by identification of highly veined dislocation surfaces.

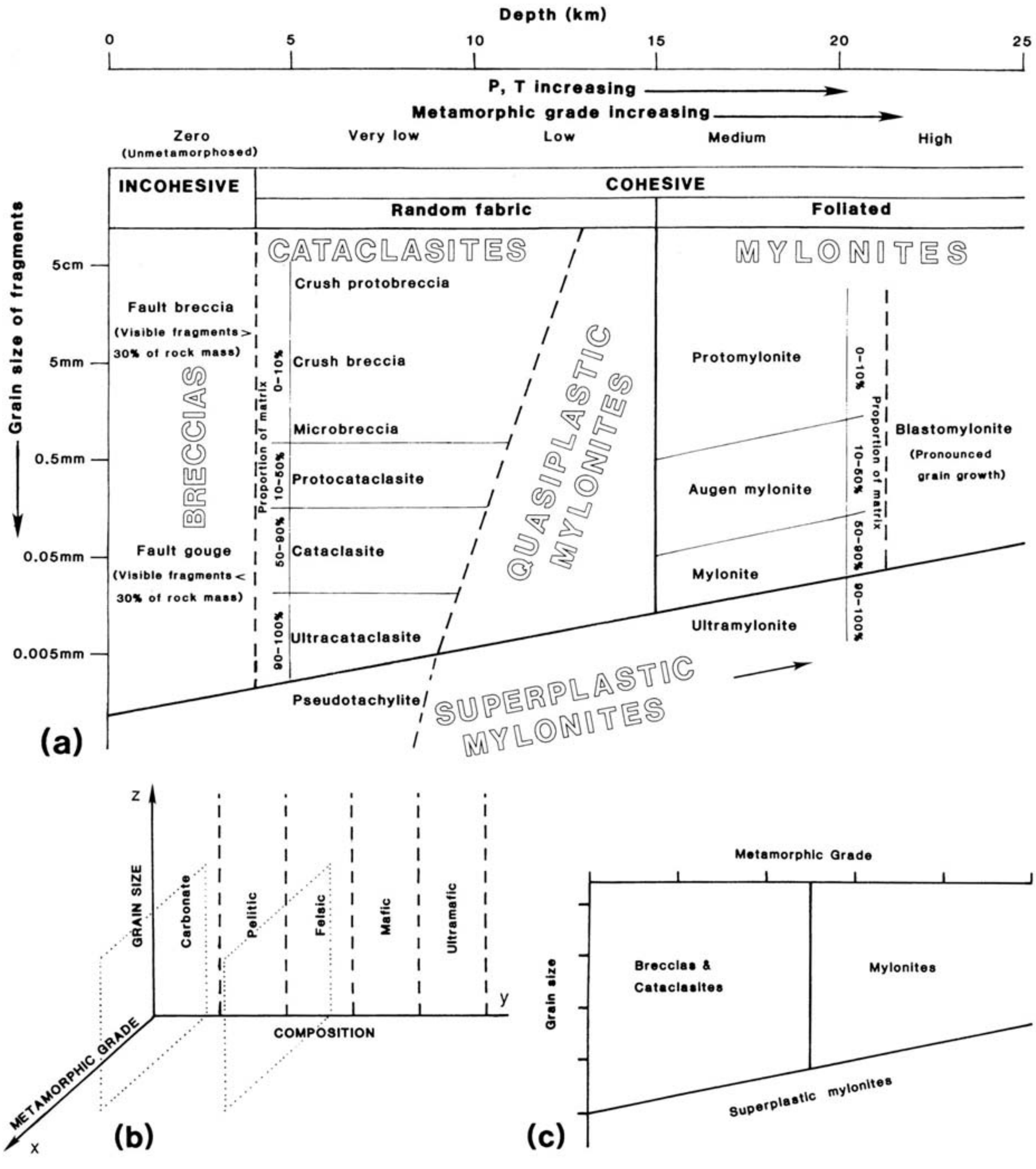


Figure A1-3 - Fault-rock terminology. (a) Classification of fault rocks that have been derived from quartzo-feldspathic lithologies (e. g. granite) (adapted from Sibson, 1977); (b) the grain size - metamorphic grade - lithologic composition grid used for classifying fault rocks (after Hull et al., 1986); (c) fault-rock diagram for marl showing expanded mylonite and superplastic mylonite fields as compared to those shown on the diagram for granite in (a) (from Marshak and Mitra [1988]).

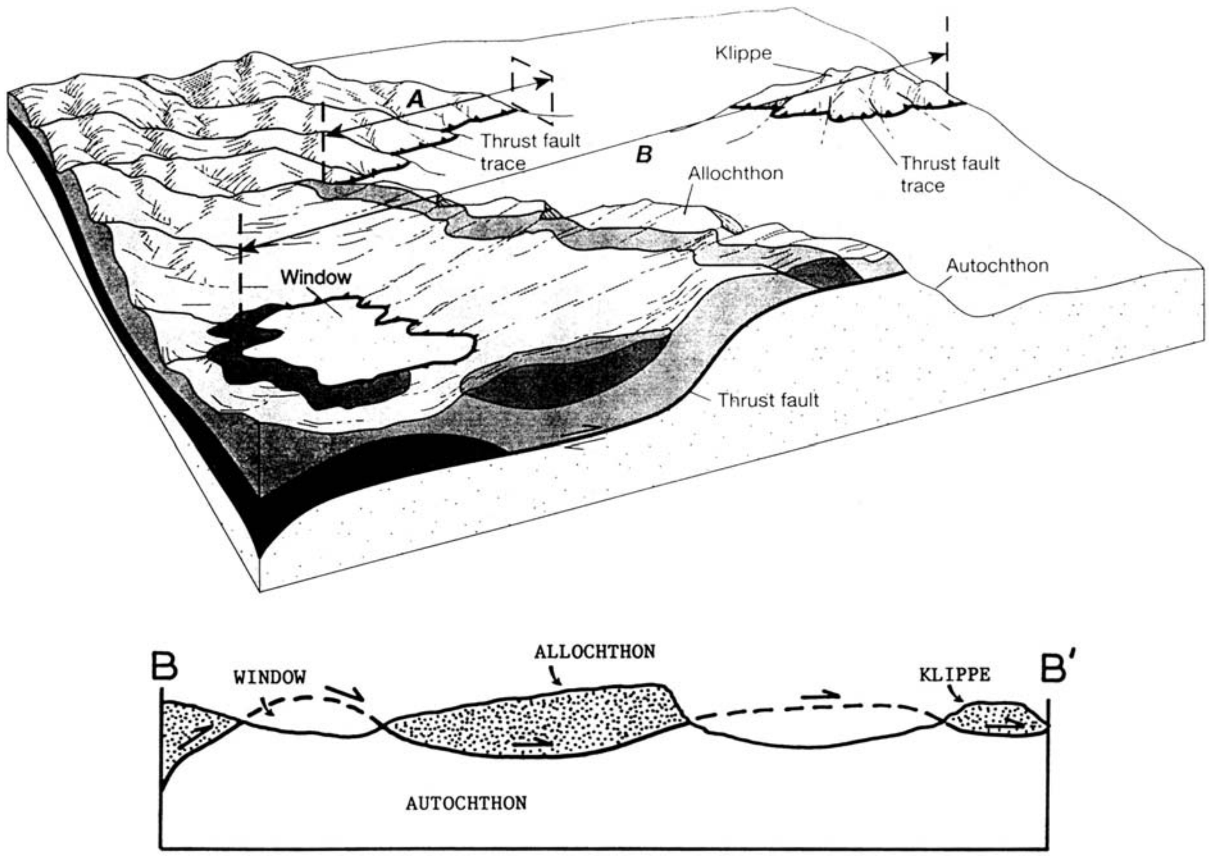


Figure A1-4 - Block diagram illustrating the relationships between allochthons, autochthons, klippen, and windows. (Twiss and Moores, 1992, p. 99) with section B-B' drawn by CM.

During episodes of mountain building associated with continuous subduction and/or collisions near continental margins, thrusting is typically directed from the ocean toward the continent. Accordingly, one of the large-scale effects of such periods of great overthrusting is to impose an anomalous load on the lithosphere that causes it to subside and form a **foreland basin**. These basins receive tremendous quantities of sediment that fill the basin with debris derived from erosion of uplifted areas within the active collision zone. In the late stages of convergence, forces transmitted from the collision zone into the developing foreland basin create a diachronous secondary stage of folding and continent-directed overthrusting of the strata filling the foreland basin. Thus, a thrust may override debris eroded from it.

Surfaces of Unconformity

Surfaces of unconformity mark temporal gaps in the geologic record and commonly result from periods of uplift and erosion. Such uplift and erosion is commonly caused during the terminal phase of regional mountain-building episodes. As correctly interpreted by James Hutton at the now-famous surface of unconformity exposed in the cliff face of the River Jed (Figure A1-5), such surfaces represent mysterious intervals of geologic time where the local evidence contains no clues as to what went on! By looking elsewhere, the effects of a surface of unconformity of regional extent can be recognized and piecemeal explanations of evidence for filling in the missing interval may be found.

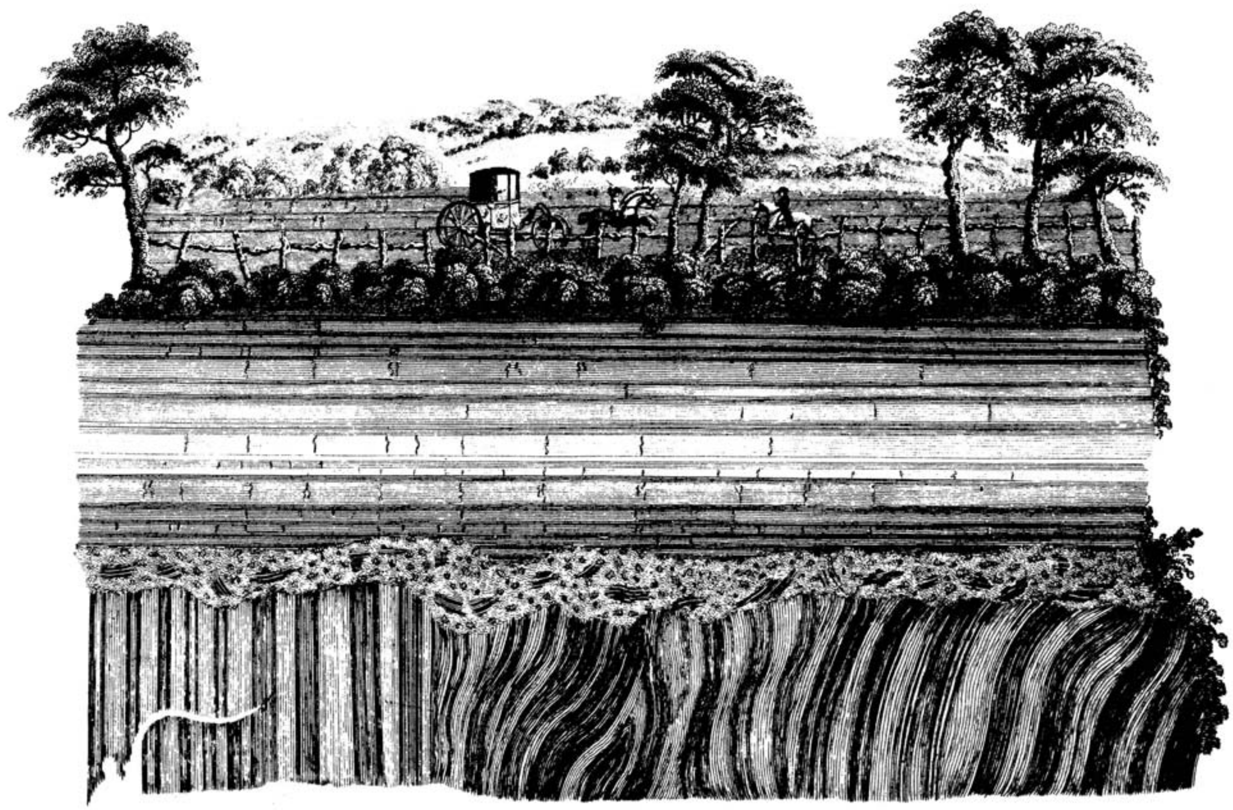


Figure A1-5 - Unconformity with basal conglomerate along the River Jed, south of Edinburgh, Scotland. From James Hutton's "Theory of the Earth", (1795).

Unconformities occur in three basic erosional varieties - angular unconformities, nonconformities, and disconformities (Figure A1-6). Angular unconformities (such as the River Jed) truncate dipping strata below the surface of unconformity and thus exhibit angular discordance at the erosion surface. Nonconformities separate sedimentary strata above the erosion surface from eroded igneous- or metamorphic rocks below. Disconformities are the most-subtle variety, separating subparallel sedimentary strata. They are commonly identified by paleontologic means, by the presence of channels cut into the underlying strata, or by clasts of the underlying strata in their basal part. The strata above a surface of unconformity may or may not include clasts of the underlying strata in the form of a coarse-grained, often bouldery basal facies.

Following the proposal made in 1963 by L. L. Sloss, surfaces of unconformity of regional extent within a craton are used as boundaries to define stratigraphic *Sequences*.

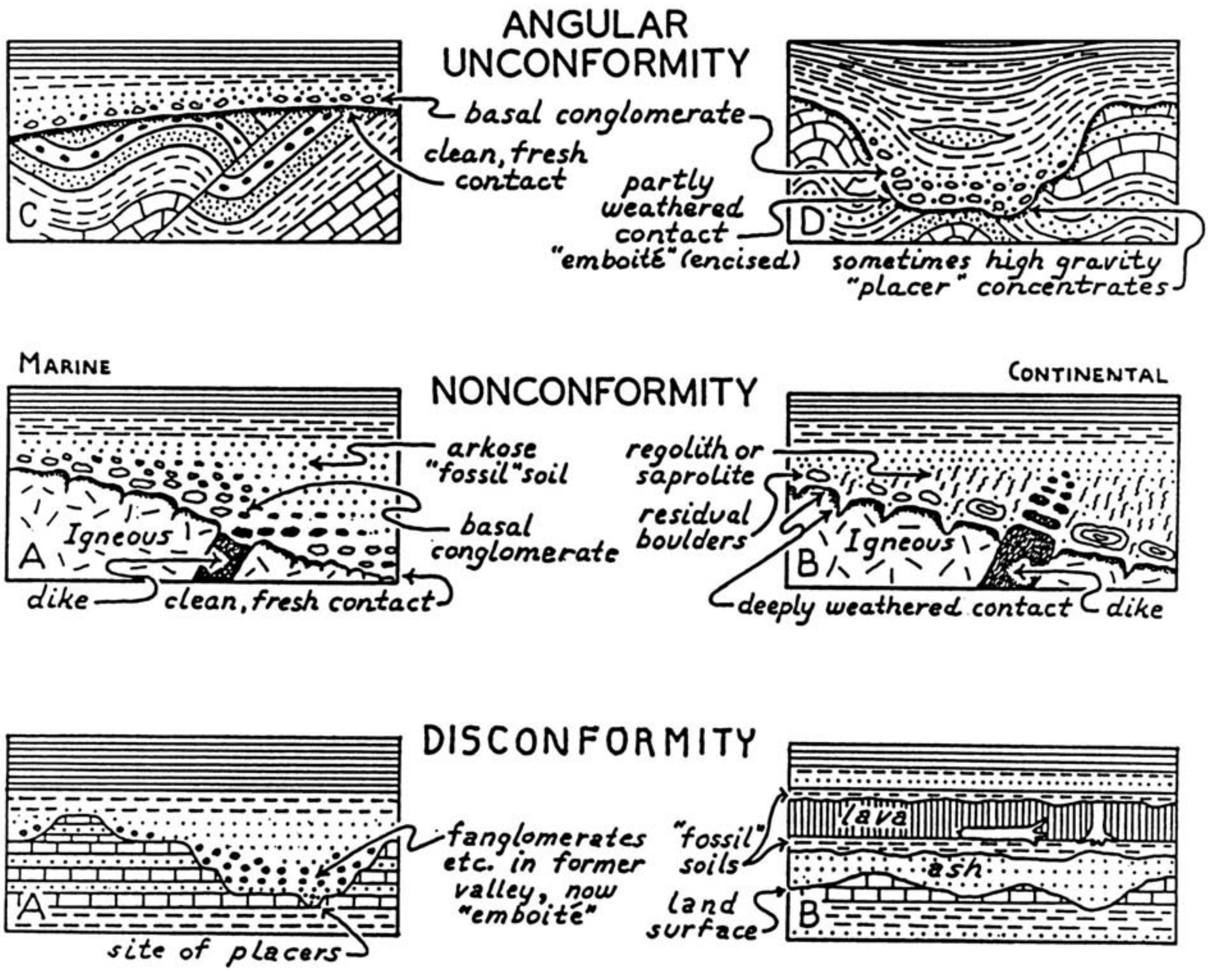


Figure A1-6 - Various types of unconformities, or gaps in the geologic record. Drawings by Rhodes W. Fairbridge.

Sedimentary Structures

During deposition in a variety of environments, primary- and secondary sedimentary structures can develop above-, below-, and within strata. During normal deposition, or settling from a fluid in a rainfall of particles, massive, essentially poorly stratified successions may result. The presence of **strata** implies a change in deposition and as a result most geologists appreciate the significance of layering in sedimentary rocks as marking **CHANGE** in big letters, be it a change in parent area of the sediment, particle size, or style of deposition. Thus, **bedding** can best be viewed as marking the presence of mini-surfaces of unconformity (diastems). During high-energy transport of particles, features such as **cross beds, hummocky strata, asymmetric current ripple marks,** or **graded beds** result. Cross- and hummocky bedding, and asymmetric current ripple marks are deposited by moving currents and help us unravel the paleocurrent directions during their formation. Graded beds result from a kind of a "lump-sum distribution" of a wide range of particles all at once (usually in a gravity-induced turbidity flow).

Thus, graded beds show larger particle sizes at the base of a particular layer "grading" upward into finer particles.

Secondary sedimentary features are developed on already deposited strata and include **mud (or desiccation) cracks, rain- drop impressions, sole marks, load-flow structures, flame structures, and rip-up clasts**. The last three categorize effects produced by a moving body of sediment on strata already in place below. A composite diagram illustrating these common structures is reproduced in Figure A1-7.

Together, these primary- and secondary sedimentary structures help the soft-rock structural geologist unravel the oft-asked field questions - namely.... **Which way is up?** and **Which way to the package store?** The direction of younging of the strata seems obvious in horizontal- or gently tilted strata using Steno's principle of superposition. But steeply tilted-, vertical-, or overturned beds can be confidently unravelled and interpreted structurally only after the true topping (stratigraphic younging) direction has been determined. As we may be able to demonstrate on this field trip, simple observations allow the card-carrying geologist to know "Which way is up" at all times.

Structures in Sedimentary- vs. Metamorphic Rocks

For hard-rock geologists working in metamorphic terranes, simple sedimentary observations will not allow the card-carrying geologist to know "Which way is up" **at all**. Rather, because of intense transposition and flow during ductile deformation, stratification, fossils for age dating, tops and current-direction indicators are largely useless except to identify their hosts as sedimentary protoliths. Thus, according to CM, "*at the outcrop scale, metamorphism can best be viewed as the great homogenizer*." Commonly during metamorphism, the increase in temperature and -pressure and presence of chemically active fluids severely alter the mineral compositions and textures of pre-existing rocks. As a result, in many instances, typical soft-rock stratigraphic- and sedimentologic analysis of metamorphic rocks is not possible.

Tectonostratigraphic Units

In metamorphic terranes, **tectonostratigraphic units** can best be described as large-scale tracts of land underlain by bedrock with similar age range, protolith paleoenvironment, and structure. Such terranes are generally bounded by ductile-fault zones (mylonites), surfaces of unconformity, or brittle faults. Unravelling the collisional plate-tectonic history of mountain belts is greatly facilitated by identifying former cratonic (ancient crustal), continental-margin, continental-slope-, and rise, deep-oceanic, and volcanic-island tectonostratigraphic units. The major distinction in unravelling complexly deformed mountain belts is to identify former shallow-water shelf deposits (originally deposited on continental crust) and to separate them from deep-water oceanic deposits (originally deposited on oceanic crust). The collective adjectives *miogeosynclinal* (for the shallow-water shelf deposits) and *eugeosynclinal* (for the deep-water oceanic deposits) have been applied to the products of these contrasting depositional realms.

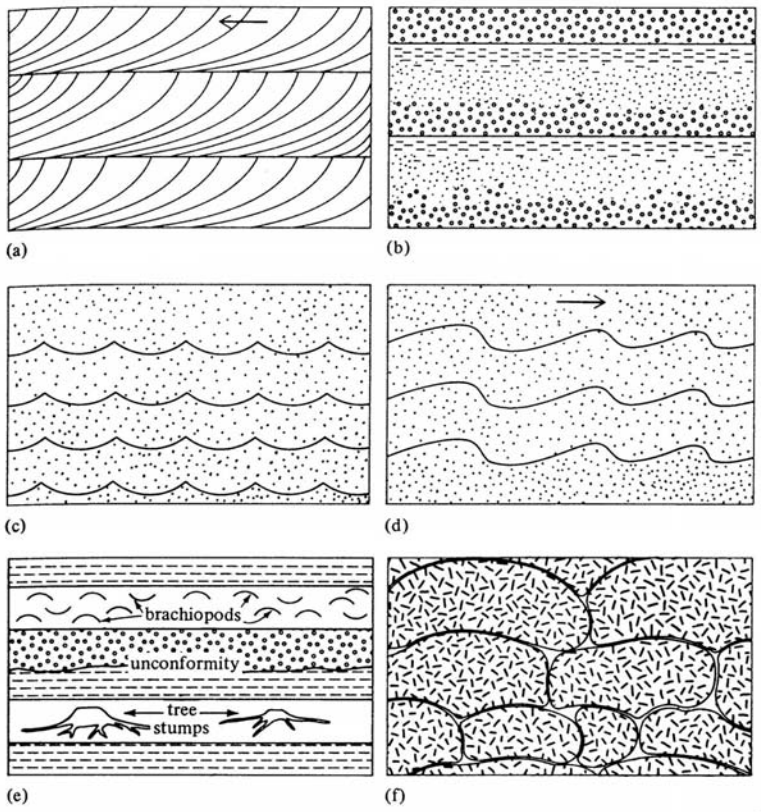


Figure A1-7 - Diagrammatic sketches of primary sedimentary structures (a through e) and cross sections of pillows (f) used in determining topping (younging) directions in rocks.

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