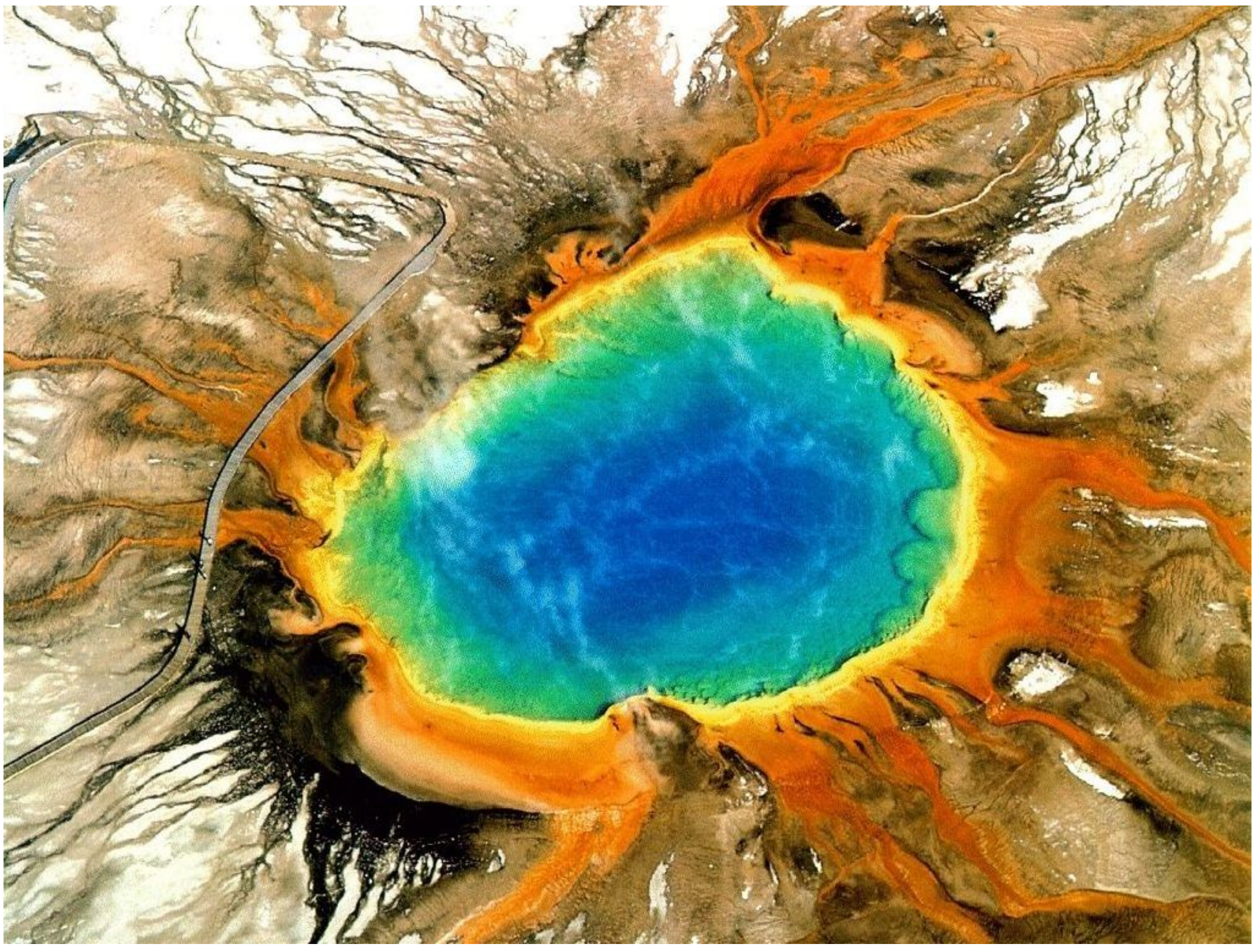
HOFSTRA UNIVERSITY
FIELD TRIP GUIDEBOOK

GEOLOGY 280C

Grand Teton and Yellowstone

Summer Session Two – 22-29 July 2006



Aerial view of the Grand Prismatic Spring, Yellowstone National Park. The intense colors are naturally produced by varieties of thermophillic cyanobacteria. (NPS Image.)

Field Trip Notes by Charles Merguerian

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Hofstra University Geology 280C
22-29 July 2006

Field Trip Guide by:
Dr. Charles Merguerian

Introduction

A land of ***"Fire and Ice"***, at least for the past few million years! Yet the geological history of NW Wyoming stretches back to the Archean Eon, roughly 2.7 billion years ago. This year's field trip will visit the most thermally active area on the Earth's surface – above the Yellowstone hot spot. Here the heat of a buried hot mantle plume and associated cooling pluton at a mile deep have created a huge domal uplift scattered with over thousands of active hot springs and other thermal features (see cover). Our trip will visit Yellowstone National Park to see all of the wonders of an active caldera complex and the surface expression of buried heat. Just to the south, near Jackson Hole, the Grand Teton range was launched from the Earth's interior by ~9 Ma (million years) of extensional faulting and seismogenic uplift activity along the range front Teton fault zone. Here, earthquakes are quite common – a testimonial to active crustal fracturing and neotectonic stress. Evidence for recent glaciation and recent volcanic activity truly creates a land of "Fire and Ice", for those with eyes tuned to see.

The guidebook you hold is simply an attempt to acquaint you with the region and to provide a ready source of information as we work in the field. The guide will also help in completion of your projects for graduate credit. The Geology 280C field guide is arranged by providing an itinerary, a general section on the geology of NW Wyoming, then two larger sections on the specifics of the geology of Grand Teton National Park and Yellowstone National Park. The guide ends with an appendix - a primer on geological structure. Naturally, many commercial guides are available for these regions and the Selected Reference list at the back of the field guide will help out in that area as will computer searches on key words.

So, sit back in your airline seat and prepare to read about the great wonders you are about to witness and experience in the week ahead. By comparison to our past field trips to examine a geological transect of central California and the great variety of geological features across northern Arizona, this year's field trip to NW Wyoming will focus on a much smaller but nonetheless interesting region. Prepare to land in the Jackson Hole Airport, find and board the happy van, place a few bets, hear the first of a long string of bad jokes, and start looking at rocks. Get 'yer hiking boots ready.

Purpose and Goals

Developed for teachers and travelers, Geology 280C is a graduate level introduction to geologic field interpretation of igneous, sedimentary, metamorphic, glacial, seismogenic, and thermal features of the Middle Rocky Mountain Physiographic Province in the vicinity of

Jackson Hole, WY. Emphasis will be placed on the field relationships of the various rock units found in the area and what conclusions about geological processes can be inferred from the field data gathered. The primary goal of this field trip is to provide hands-on instruction of geologic field observation and to explore the methodology used to develop conclusions based integrated academic learning and observed data.

Acknowledgements

No endeavor of this scope is possible without the help of many individuals and agencies. CM would like to thank **Dr. Janice Koch** and **Eloise Gmur** of the IDEAS Institute and **Dean Steven Costenoble** of Hofstra University for their direct help in the logistical planning for air and ground travel and for our overnight accommodations and for their support of this field program. Help from people on the ground in WY is always an important part of any off-campus experience. In this connection, my colleague **Starr Lanphere** has provided input into the localities and resources I have included in this field trip guidebook and he will hopefully join us during parts of our trip. None of these trips would have been possible without the assistance and fortitude of **John Gibbons** of Hofstra University who first implored me to organize and conduct such undertakings. The assistance of the **National Park Service** in granting admittance into Grand Teton National Park and Yellowstone National Park is gratefully acknowledged. In addition to these individuals and agencies, the staff at Duke Geological Laboratories and especially **H. Manne Vb** have been instrumental in completing the guidebook in a timely manner.

Since confession is good for the soul, web resources were used extensively in the preparation for this guide and some areas of text were cut and pasted in true plagiaristic style. In this connection, special thanks are due Hofstra Geology major, **Hallie Thaler** for her timely help in downloading and organizing web material and to **J. Mickey Merguerian** for file transcription. The following web resources were used in preparation and production of this guide:

http://en.wikipedia.org
http://rmg.geoscienceworld.org
http://visibleearth.nasa.gov
http://www.americanparknetwork.com
http://www.nps.gov
http://www.yellowstone.net
http://www.yellowstonenationalpark.com
http://www2.nature.nps.gov

Field Trip Itinerary

17 July (Mon.) – Meet at Gittleson Hall (162) for pre-trip information (10:00 AM).

22 July (Sat.) – Arrive at LaGuardia airport at 4:30 AM for 6:00 AM United Airlines flight #115 to Denver. Arrive Denver 8:05 AM. Switch to United Airlines Flight 1037 (9:05 AM) from Denver and arrive Jackson Hole WY at 10:29 AM. Secure van and provisions. Overview of southern Grand Teton region, Wind River Range and the Gros Ventre landslide area near Kelly, WY. Stay in Jackson Lake Lodge for the first of three evenings. Moose watch begins before sundown. Evening lecture programs available for those who can stay awake.

Jackson Lake Lodge: http://www.gtlc.com/lodgeJac.aspx; Phone: 307-543-1900

View of Grand Tetons and Snake River Plain. Image by C. Merguerian (1981).

23 July (Sun.) – Hike south from String Lake to Jenny Lake to trail head at boat dock for hike up Cascade Canyon to examine geological features. Discuss Precambrian geology of hike up to Inspiration Point. Continue small group projects and explore central portion of park including Signal Mountain Road drive. Stay in Jackson Lake Lodge for second. Moose calls and evening lecture programs. Pack for early AM departure.

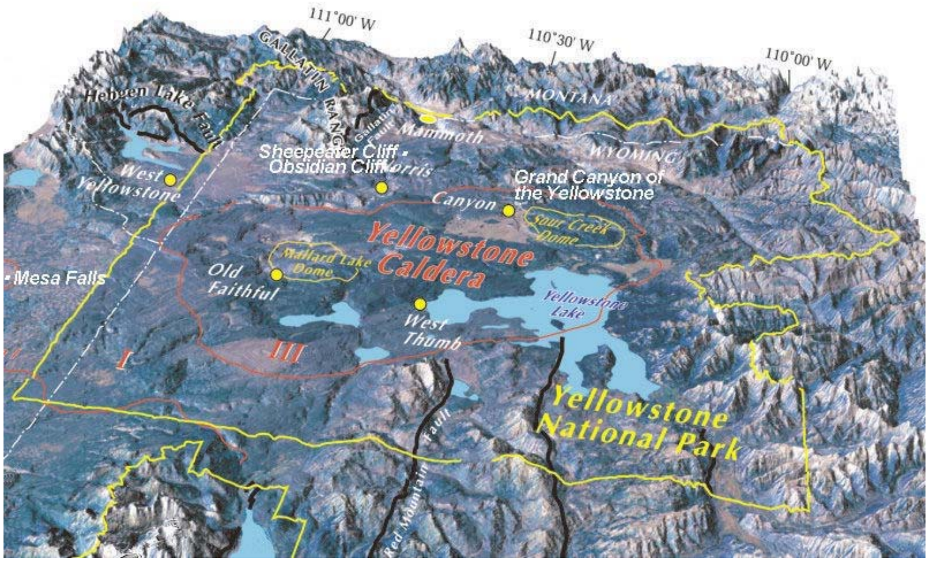
24 July (Mon.) – Discuss Jackson Buttes as horsts, then take aerial tram from Teton Village up to alpine zone of Grand Tetons for geological overview and short alpine hike. Discuss structural geology of Tetons in context of regional tectonics and suggest more group projects. In afternoon, descend to valley by tram and hike Valley Trail across the Tetons fault. Take Jenny Lake Scenic Drive and stay at Jackson Lake Lodge for third night. More Moose!

Jenny Lake area of Grand Tetons, NPS.

View of Cascade Canyon of the Grande Tetons, NPS.

25 July (Tues.) – In morning drive north on Routes 89, 191, and 287 across the Continental Divides toward the West Thumb entrance for Yellowstone National Park (~75 mi.). We will spend the next three days examining geological features in and overall clockwise fashion through the park. Today we concentrate on Cenozoic volcanic stratigraphy, caldera formation, and thermal features below West Yellowstone entrance area, including Old Faithful, Upper Geyser and Middle Geyser basins. Stay in West Yellowstone, MT for evening. Evening program (7:30 pm) available at Grizzly and Wolf Discovery Center, near hotel.

Stay at Best Western Cross-Winds Motor Inn. *Note: there is a pool and hot tub.*
Web - http://book.bestwestern.com/bestwestern/selectHotel.do; Phone: 406-646-9557.



Physiographic view of Yellowstone National Park, NPS.

26 July (Wed.) – Backtrack south along Grand Loop Road to Grand Prismatic Spring for early morning oblique lighting and from there continue clockwise northward as we plan to marvel at the depositional forms of silicic geyser activity (geyserite). Stops at Monument Geyser Basin, Beryl Spring, and the Norris Geyser Basin are planned. Students continue to record data and add to their field notes. In the late afternoon we head northward along Grand Loop Road to see the Obsidian Cliffs. Stay in Gardiner, MT (just outside of north park entrance) for evening. Evening program (9:30 pm) at Mammoth Campground Amphitheater.

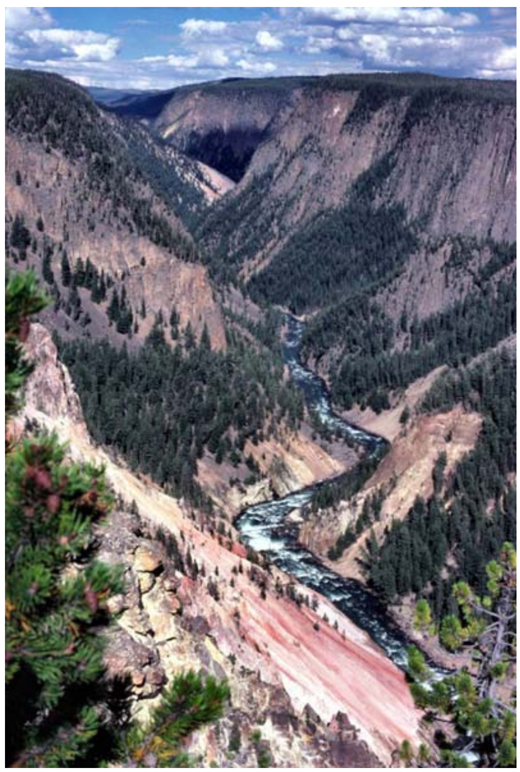
Stay at Yellowstone Village Inn: Note there is an indoor pool and sauna.
Web - http://www.yellowstonevinn.com; Phone: 800-228-8158.

27 July (Thurs.) – Start at Mammoth Hot Springs for calcareous geyser forms and head eastward along Grand Loop Road toward Tower-Roosevelt and include a visit to the famous Petrified Tree. Discuss valley formation, age relationships of lava flows and contrast Yellowstone and Grand Teton geomorphology. Stay at Yellowstone Village Inn again by backtracking. Evening program (different topic) available at same place as last night.

28 July (Fri.) – Complete loop southward past then eventually southward toward the Grand Canyon of the Yellowstone. Short hikes to see Yellowstone Canyon from different vantage points. Examine lower falls of the Yellowstone and hike along the mud volcano loop. Time permitting we may detour eastward at Lake Junction eastward to Lake Butte on into the Absaroka Range. Regroup afterward for the drive down to Jackson Hole for a final visit to Grand Teton National Park. Stay in Jackson Hole for evening.

Stay at Ranch Inn, the closest hotel property to the "famous" Jackson Town Square.
Web - http://www.ranchinn.com; Phone: 800-348-5599.

Mammoth Hot Springs area of Yellowstone National Park. Image taken by author in 1981.



Grand Canyon of the Yellowstone River. Image taken by author in 1981.

29 July (Sat.) – Morning wrap up of Grand Teton area. After breakfast, drop off van and arrive at Jackson Hole airport for flights back to New York. United Airlines flight # 1040 leaves Jackson airport at 3:21 PM and arrives Denver at 4:47 PM. From Denver, at 6:01 PM we'll take United Airlines flight # 574 and arrive at LaGuardia at 11:34 PM.

15 Aug (Tues.) – Meet in Gittleson 162 for class Powerpoint presentations and pot-luck dinner (6:30 PM).

The Geology of Wyoming

Location and Physiography

As planned, our geological field course will be held in NW Wyoming, a part of the Middle Rocky Mountain Division of the United States (Figure 1). The landscape of NW Wyoming forms a huge bulge or dome, so much so that major rivers and their tributaries drain away from the region in a radial fashion. Clockwise fashion from the northwest corner of the state these drainages include the Yellowstone, Bighorn, Platte, Green, Snake, and Missouri Rivers. Wyoming varies greatly in climate and relief from low and desert-like in the northeast (Great Plains) to mountainous, extending from the northwest corner into the interior. The mountains give way to the Wyoming-Bighorn Basin to the south and east (Figure 2).

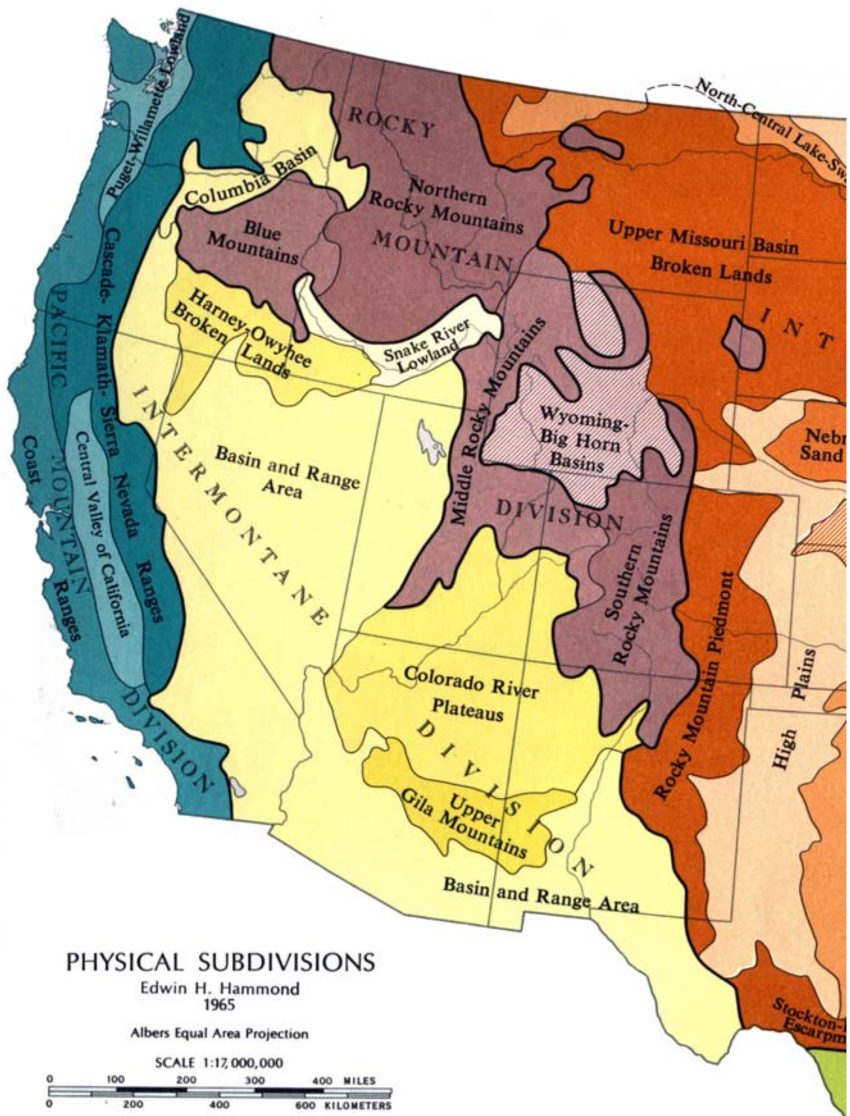


Figure 1 – Physical subdivisions of the Cordilleran belt of western North America (Hammond 1965).

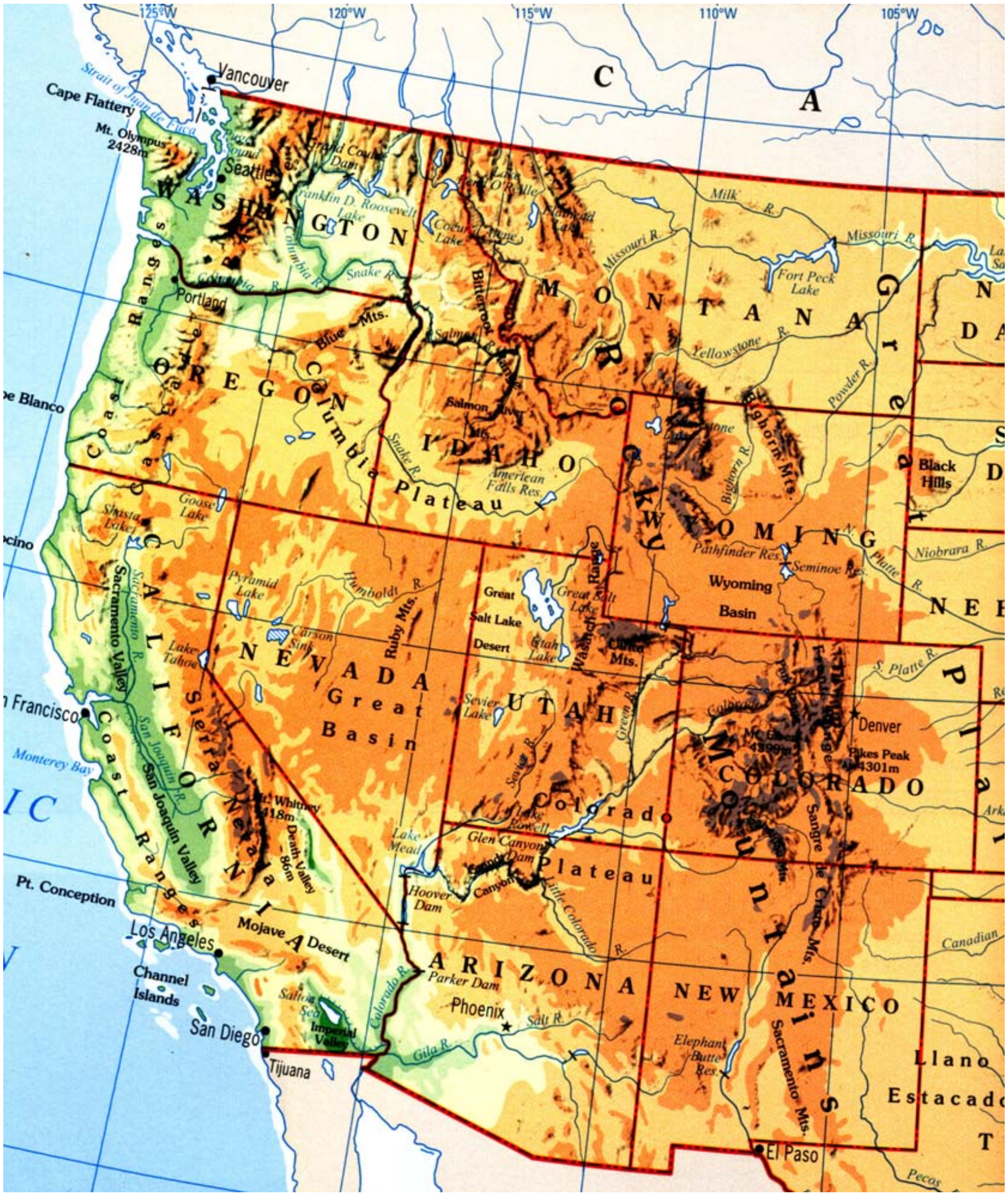


Figure 2 – Physiographic map of the western United States.

The most imposing Wyoming mountain ranges include the Beartooth, Absaroka, Wind River, and Bighorn with peaks reaching above 12,000' and relief of 5,000'-7,000' above adjacent basin floors. Direct connections with the Rocky Mountains of southern Montana, the Black Hills, the Colorado Rockies, and the Uinta Range surround the state. With the exception of the Eocene Absaroka "volcanic" range, the Wyoming Rockies consist of folded and faulted Mesozoic and Paleozoic sedimentary strata cored by Proterozoic and Archean metamorphic and igneous rocks. These rocks are exposed in broad folds (typically thrust-bound antiforms) or on tilted fault blocks. (Compare Figures 3 and 4.)

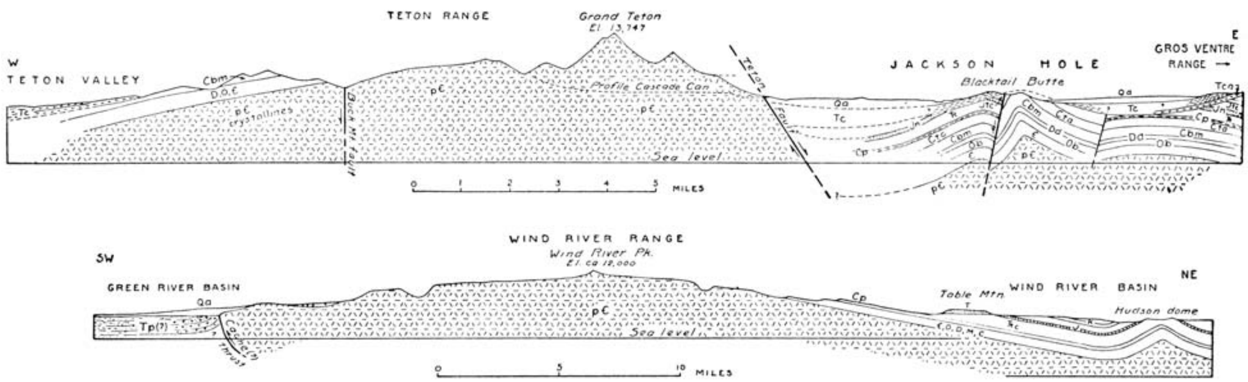


Figure 3 – Cross sectional views of the contrast between fault block mountains (Teton Range) vs. Laramide fold and thrust mountains (Wind River Range). (Adapted from Eardley, 1951, fig. 203, p. 348.)

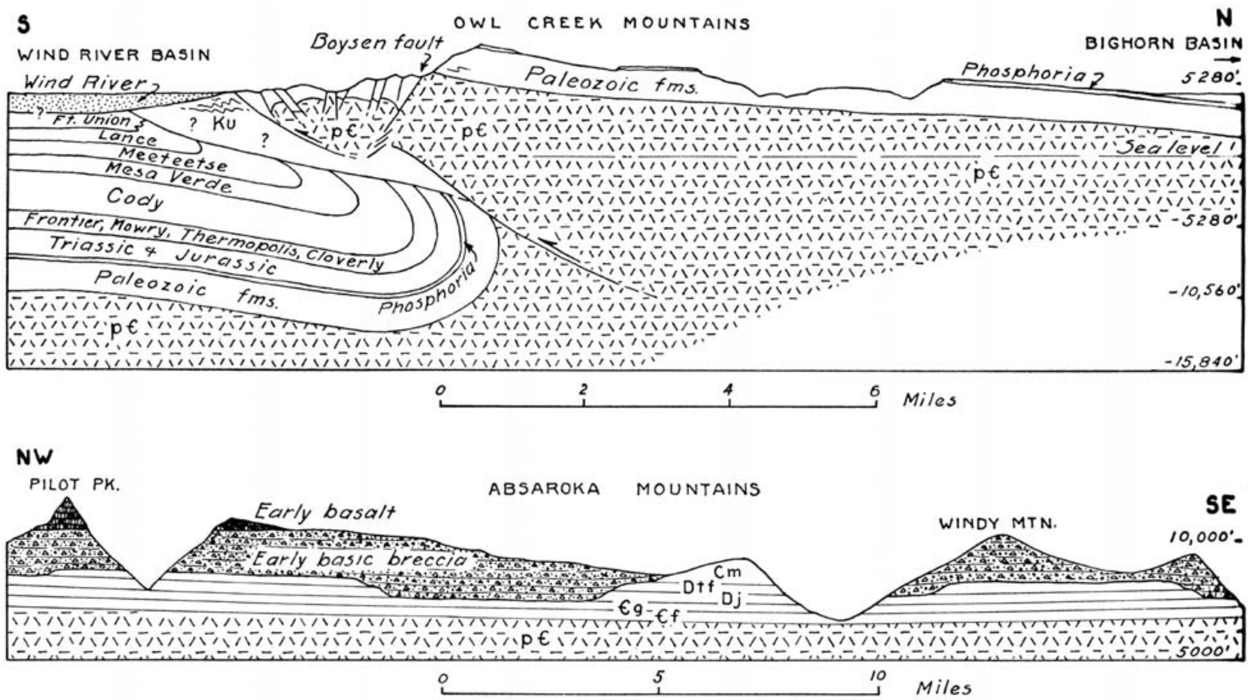


Figure 4 – Cross sectional views of the Owl Creek Mountains near the Wind River Canyon, consisting of backthrusted (Laramide) Proterozoic and older crystalline cored overthrusts above tightly folded Paleozoic and Mesozoic strata (Ku = Upper Cretaceous). By contrast, the lower section exposes the structure of the Eocene Absaroka Range on the SW side of Clark Fork Valley. pꞒ = Proterozoic and older basement rocks, Ꞓf = Flathead Quartzite, Ꞓg = Gallatin Shale, Dj = Jefferson and Bighorn Formations, Dtt = Three Forks Formation, Cm = Madison Limestone. Eocene basalt breccia and basalt flows cap the Paleozoic strata with disconformity. (Adapted from Eardley, 1951, fig. 205, p. 350.)

Snake River-Teton-Gros Ventre-Wind River Element

Trending southeastward from Idaho, the Snake River, Teton, Gros Ventre, and Wind River ranges are in general alignment for over 240 km (150 mi). Supporting a number of isolated modern glaciers, the ranges are of great relief and exquisite beauty. The Grand Teton is 13,747' in elevation, second only to Gannett Peak in the Wind River Range at 13,785' in elevation. Three of the four ranges have simple northeastern flanks consisting of dipping strata consistent with anticlinal arches such as found in the Big Horn and Black Hill ranges. (See Figures 3, 4.) By contrast, their southwestern flanks are characterized by steep upturning, attenuated strata and overthrusts, all produced during the Laramide Orogeny. In the case of the Teton range, strata on the western slopes dip west and the raw eastern escarpment has been created by ~9 Ma of extensional seismogenic fault activity along the Teton range front fault system. These majestic ranges are superimposed on ancestral roots of the earlier Teton-Gros Ventre, Washakie, and Owl Creek Uplifts, all produced during the Sevier Orogeny. These building blocks of the Rockies are exposed along the Targhee Uplift (Figure 5), a broader element of the Laramide orogenic event.

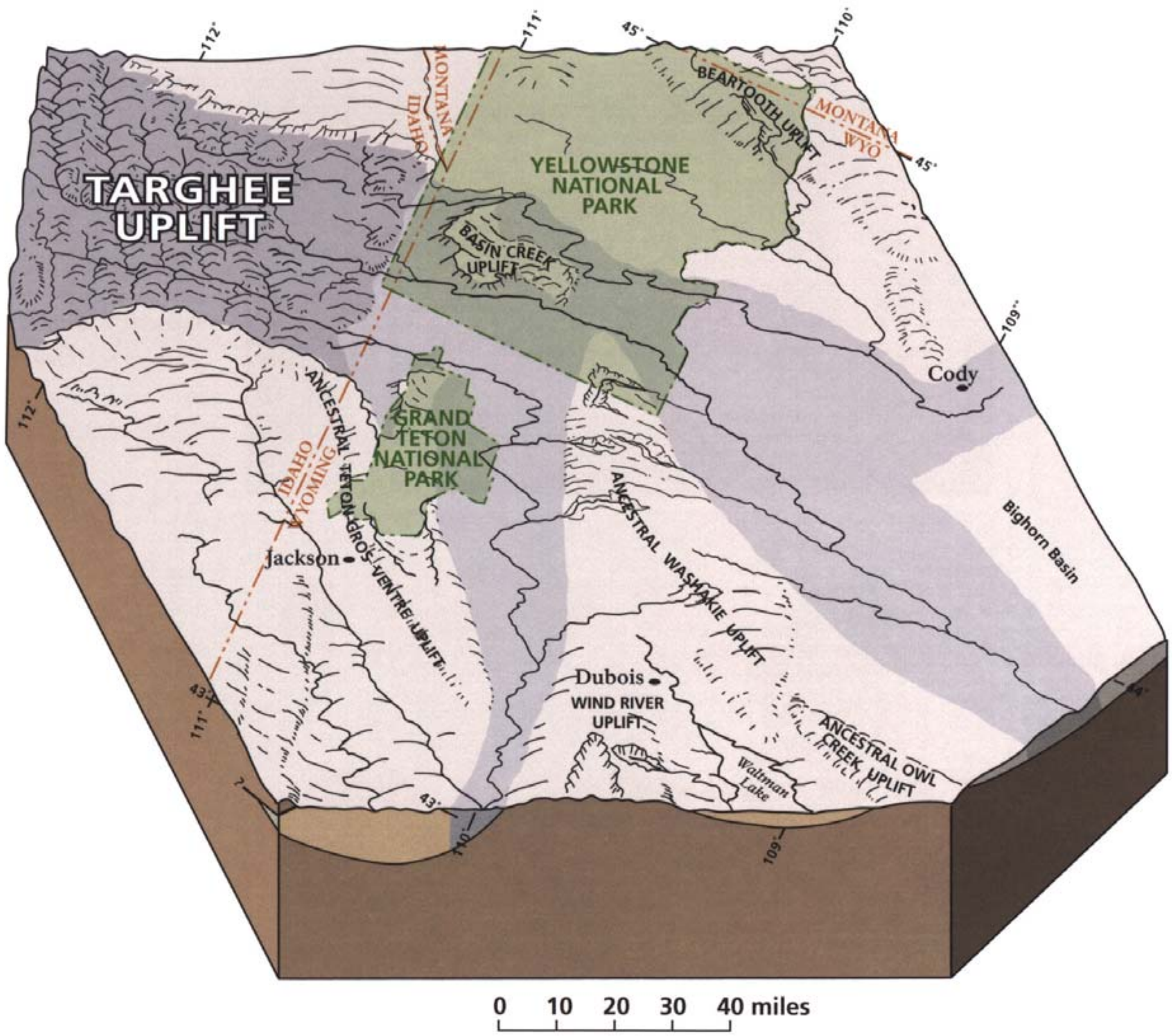
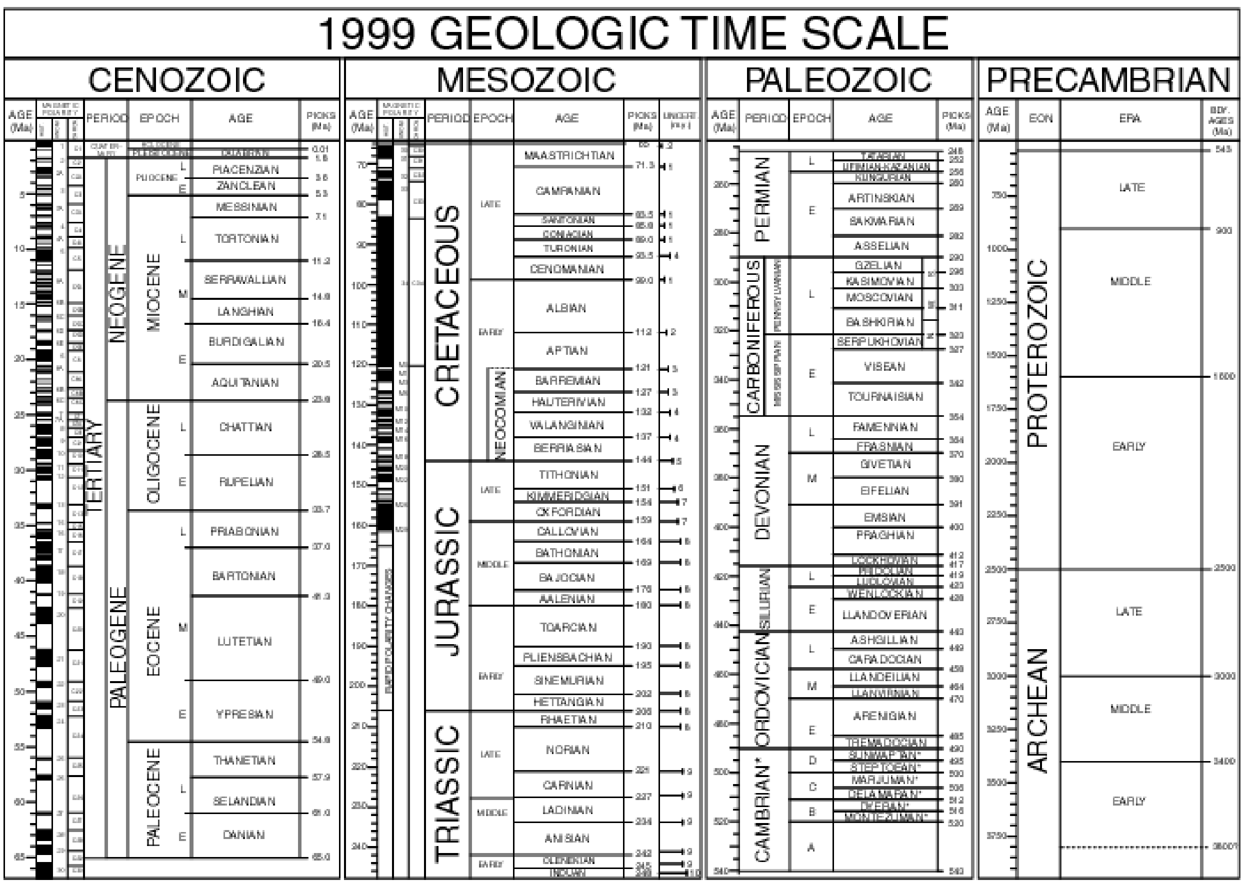


Figure 5 – Block diagram showing pre-Paleozoic crystalline rocks exposed in the Targhee Uplift (darker gray) and depositional belt of the 67-55 Ma Pinyon Canyon Conglomerate (lighter gray). (Adopted from Love et al., 2003, fig. 54, p. 60.)

The mountainous ranges of the Rockies are the direct result of the combined effects of Sevier (145-73 Ma) and Laramide (70-50 Ma) orogenies, both respectable compressive mountain building events. Mountain building produced a protracted series of fold and thrust belts across the Cordillera in response to mid-Mesozoic to lower Cenozoic Andean-type subduction along the active margin (Table 1). A general discussion of Cordilleran Tectonics is timely since all events are recorded in the study area, either as active tectonic elements or as periods of depositional change. Timely also, is the introduction of the geological timescale, a document central to all of our rocky thinking (Figure 6).

Table 1 is a geological timescale and chronology of major tectonic and glacial events for the Grand Teton and Yellowstone National Parks Region, compiled from many sources. It will be a basis for discussion in the field and is included in this section. The time scale indicates the era, period, and epoch subdivisions. The geologic development of NW Wyoming spans more than two and a half a billion years (~2.7 Ga). As a result, you will consult this table often.



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*International ages have not been established. These are regional (Laurentian) only.

Sources for nomenclature and ages: Primarily from Gradstein, F., and Ogg, J., 1996, *Episodes*, v. 19, nos. 1 & 2; Gradstein, F., et al., 1995, SEPM Special Pub. 54, p. 95–128; Berggren, W. A., et al., 1995, SEPM Special Pub. 54, p. 129–212; Cambrian and basal Ordovician ages adapted from Landing, E., 1998, *Canadian Journal of Earth Sciences*, v. 35, p. 329–338; and Davidek, K., et al., 1998, *Geological Magazine*, v. 135, p. 305–309. Cambrian age names from Palmer, A. R., 1998, *Canadian Journal of Earth Sciences*, v. 35, p. 323–328.

Figure 6 – Geological timescale. (From Geological Society of America, 1999.)

Table 1 - Geological Timescale and Chronology of Major Tectonic and Glacial Events for the Grand Teton and Yellowstone National Parks Region

2006 – Geology 280C field class invades NW Wyoming, 22-29 July.

1959 – **Madison River Canyon "Hebgen Lake" earthquake** occurred on 17 August at 11:37 PM. The Richter magnitude 7.5 temblor shook for 8 seconds and unleashed an eighty-million-ton wall of debris that slid down canyon to create a natural dam ~1 mile long and 0.75 mile wide.

1949 – ***Holocene/Obscene boundary.***

1927 – Stressed by snow melt, the water-saturated and undercut Gros Ventre debris dam broke loose resulting in draining of an 8 km long 60 m lake which flooded the town of Kelly, WY and resulted in eight deaths (six humans and two jackalopes).

1925 – **Gros Ventre landslide** (possibly related to a small earthquake) resulted in river damming with 10 km^3 of rock displaced from Sheep Mountain into the Gros Ventre River valley.

1,250 - 1,850 A.D. – Little Ice Age heralded by 0.25°-1.0° decrease in global temperatures. Perhaps related to the **Maunder Minimum** – a dramatic decrease in sunspot activity between 1645 and 1715. Rivers froze in Europe and drought patterns form in North America.

10,000 – 6,000 yBP (years before present) – Maximum summer temperatures experienced in NW Wyoming the result of altithermal effect.

11,200 yBP – ***Pleistocene/Holocene boundary***. (=10,000 radiocarbon years BP.)

13,200 yBP – Trees invade Yellowstone and Clovis people arrive in Wyoming with Tupperware and Bic lighters.

70,000 - 15,000 yBP – Pinedale glaciation; maxima at 25,000 yBP.

125,000 yBP – Worldwide warm interglacial climate.

150,000 yBP – Bull Lake glacier extends to west of West Yellowstone, MT.

500,000 - 70,000 yBP – Lavas flow in and around Yellowstone Caldera.

0.64 Ma (million years ago) – Extensive Yellowstone volcanism and formation of Lava Creek Caldera.

1.3 Ma – Oldest dated glaciation in Yellowstone and Grand Teton region. Extensive Yellowstone volcanism and formation of Island Park Caldera.

1.6 Ma – ***Pliocene/Pleistocene boundary***. Starts interval of "Fire and Ice".

2.1 Ma – Extensive Yellowstone volcanism and formation of Huckleberry Ridge Caldera.

5 Ma – ***Miocene/Pliocene Boundary***.

9 - 6 Ma to Present – Extensional seismogenic uplift and westward tilting of Teton Range along Teton fault zone results in roughly 30,000' of composite offset. Thus, the offset has been ~1' per 300 yrs or 1.0 cm per year for you nit pickers out there.

15 Ma to Present – **Basin and Range Extension.**

16 Ma – Yellowstone hotspot initiates near the Nevada-Oregon border. Interval of "Fire" begins.

24 Ma – ***Oligocene/Miocene Boundary***.

24 -2.05 Ma – Interval of Tertiary volcanic and non-volcanic deposition forming an aggregate thickness of 18,850'. From the base upwards these include the Miocene Colter, Teewinot, and Camp Davis formations covered by 6 Ma volcanic rocks and the 5.5 Ma Conant Creek Tuff. These are overlain by Pliocene volcanic and lacustrine deposits known as the Kilgore Tuff (4.45 Ma), the Shooting Iron Formation, and the Huckleberry Ridge Tuff (2.05 Ma).

37 Ma – ***Eocene-Oligocene Boundary***.

54 - 42 Ma – **Absaroka Volcanism** covers NW Wyoming during Eocene, absarookly everywhere.

58 Ma – ***Paleocene/Eocene Boundary***.

70 - 50 Ma – **Laramide Orogeny** – Late Cretaceous to early Eocene period of continued but lower angle Andean subduction and compression that results in back-arc, eastward directed thrusts involving thermally weakened basement. Creates thrust bounded broad uplifts such as the Teton-Gros Ventre, Washakie, Beartooth, and Targhee Uplifts.

66 Ma – ***Cenozoic/Mesozoic boundary***.

112 - 99 Ma – **Great Cretaceous Seaway** – Deposition of vast deposits (4,675' thick) of shale, sandstone, and subordinate bentonite, coal, and conglomerate in non-marine deep water to coastal conditions. Formations in Teton region, from oldest to youngest include Thermopolis Shale, Mowry Shale, Frontier Formation (Sandstone, Shale, Bentonite), Cody Shale, and Bacon Ridge Sandstone.

144 Ma – ***Jurassic/Cretaceous boundary.***

145 - 73 Ma – **Sevier Orogeny –** Period of Cretaceous Andean subduction that results in continentward displacement of vast sheets of overthrust strata (including Cretaceous Seaway) to form the Rocky Mountain Fold and Thrust belt. Produced extensive highlands (Caribou, Salt River, Wyoming, and Snake River Ranges) that shed sediment eastward into Teton region.

205 Ma – ***Triassic/Jurassic boundary.***

270 - 160 Ma **– Sonoman Orogeny** and **Nevadan Orogeny** create unrest along active margin to the west of the study area. During active margin phase, change from passive margin sedimentation to deposition of sandstone, siltstone, shale, phosphate, and gypsum in a marginal seaway fluctuated with coastal continental deposition. Formations from this period in the Teton region exceed 4,275’ in aggregate thickness. From oldest to youngest the strata include the Amsden (Miss.), Tensleep, Phosphoria, Dinwoody, Chugwater, Nugget, Gypsum Spring, Sundance, Morrison, and Cloverly (Cret.) formations.

245 Ma – ***Mesozoic/Paleozoic boundary***.

495 - 354 **– Carbonate Platform Stage.** Deposition of shallow water marine strata above the uplifted, eroded, and submerged passive margin starting with Cambrian Flathead Sandstone, Gros Ventre Formation(Shale and Limestone), and Gallatin Limestone, followed by Ordovician Bighorn Dolostone, Devonian Darby Formation (Dolostone and Shale), and Mississippian Madison Limestone. In the vicinity of NW Wyoming the Paleozoic strata exceed 2,860’ in thickness.

545 Ma – ***Paleozoic/Proterozoic boundary***.

1,400 - 765 Ma – Intrusion of Mid- to Late Proterozoic mafic dikes such as found at the east face of Mount Moran and the Middle Teton and elsewhere cutting Archean rocks of the Tetons. The larger dikes are clearly visible from the Jenny Lake and String Lake areas. The mafic injections were followed by an extended period of uplift and erosion.

2,500 Ma – ***Proterozoic/Archean boundary***.

2,680 - 2,545 Ma – Age of metamorphic rocks found in the crystalline core of the Teton Range. They are polydeformed, of sedimentary and igneous parentage, and consist of layered gneiss, metagabbro, granite, pegmatite, and lesser amphibolite, schist, and serpentinite. Granitic intrusives at around 2500 Ma marks a period of global granitization of continents known as the Algoman Intrusive Episode. Studies indicate deformation of core rocks of the Tetons range at former depths of 10-15 km and 575°C. As the base is not exposed, the thickness of the crystalline basement is moot.

Cordilleran Tectonics

The modern plate tectonic setting of the western United States is depicted in Figure 7 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 lithosphere along the California margin has created it all. Today, with the development of a triple junction in the Baja California area, subduction has been converted to transcurrent (strike-slip) motion along most of the western margin of North America and certainly along the south half of California. Basin and Range extension is concentrated in areas of eastern California, Arizona, and Nevada and extends all the way to the Grand Tetons and west Texas (the Rio Grande Rift). Such profound extension in the last 15 Ma mark the lithospheric response to ancient subduction of the buried but still spreading East Pacific Rise. To the north of the Mendicino fracture zone, modern Andean subduction is responsible for the Cascade volcanic district which includes Mt. Shasta and Lassen in California.

Figure 7 – 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 Andreas Fault. (From Tharp, 1969.)

The Cordillera of the western United States forms a 1,600 km wide zone of mountain building that records the results of over 350 Ma of continual, predominately compressive plate tectonic activity along the embryonic western margin of North America. The Cordillera 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 the Cordillera throughout that entire interval. The active mobile belts of the Indonesian region come to mind when attempting to derive a modern paleogeographic analog to active Cordilleran margin.

Proterozoic and Archean rocks are largely buried but underlie the Cordillera in NE-trending fault-bounded belts (Figure 8). We have seen these older rocks on our previous trips – they occupy the inner gorge of the Grand Canyon and occur in southern California and in the southern Coast Range where they have been splintered and transported NW along the San Andreas fault system. They are plainly visible in the internal zones of the Rocky Mountain fold and thrust belt, and along the elevated sides of fault-block mountains, such as the Grand Tetons.

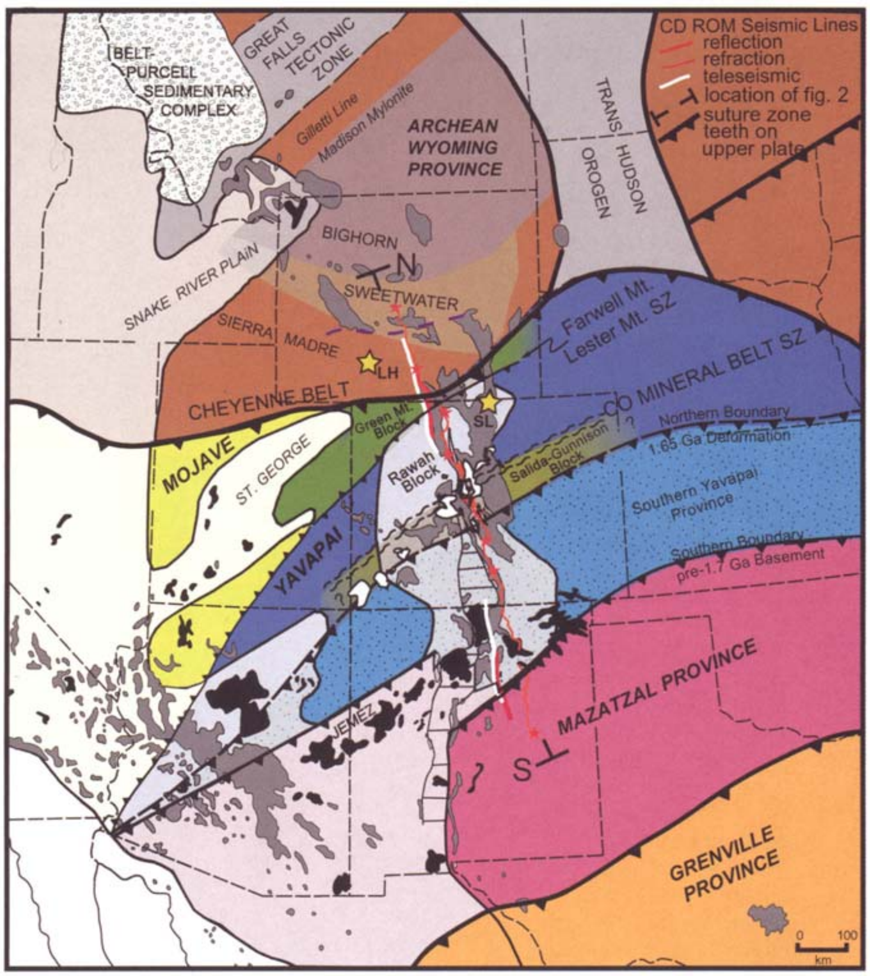


Figure 8 – Terrain map of the SW Cordillera showing the convergence of the Mojave, Yavapai, and Mazatzal provinces and their extension across Wyoming into southern California. (From CD-ROM Working Group, 2002.)

Pre-Antler Passive Margin

To best explain the ins and outs of our modern ideas on the tectonics of the Cordillera, I will explain things geologically – that is, from the bottom up. So, bottoms up! Here we go! In the aftermath of nearly two billion years of lithospheric upheaval and development, continental assembly was interrupted by a period of rifting and creation of a trailing edge, passive continental margin by early Paleozoic time. Indeed, throughout the early Paleozoic Era, development of an open-ocean passive margin formed an extensive open-ocean miogeosynclinal and offshore deep-water eugeosynclinal couple (Figure 9). Although facing in the opposite direction, the plate tectonic setting of the Cordillera was a mirror image to the developing Appalachian passive margin throughout most of the early Paleozoic.

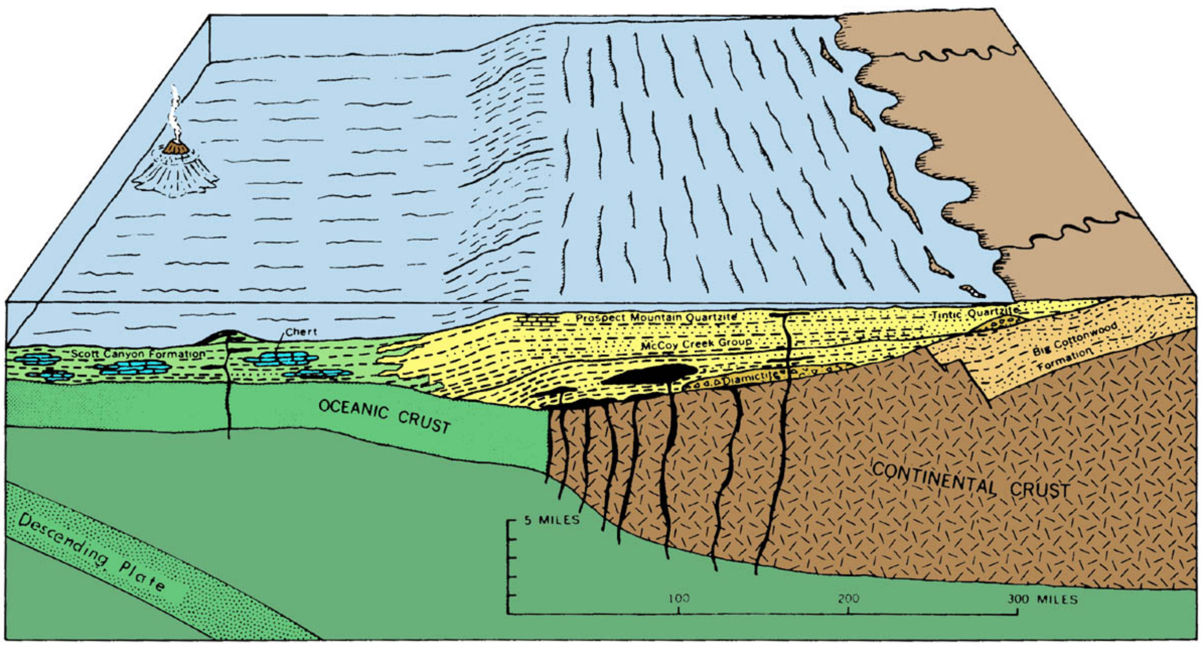


Figure 9 – Pre-Antler open ocean view of the passive Cordilleran margin during early Paleozoic time. The eventual position of Wyoming is somewhere near the top right edge of this diagram.

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 continentward to sweep across areas as far as Arizona and western Wyoming. As time passed, the offshore depositional environments shifted continentward above shallow water strata with trangression of the Cambrian sea. This resulted in a fining upward sequence of strata where Cambrian sands were overlain by shale and then by carbonate, a sequence reminiscent of the Tonto Group in Arizona (Tapeats-Bright Angel-Mauv Limestone). Carbonate deposition in the Cordillera continued into the late Ordovician. A regionally impressive Cambrian to early Ordovician transgression is also recorded in the Appalachian belt, by the way.

Antler Orogeny

Starting with the **Antler Orogeny** in late Devonian – early Mississipian time, a series of convergent margin events began to affect the western Cordillera. An arc-continent collision was responsible for the Antler Orogeny (Figure 10). 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 1) and/or a cold one from the ice cooler. In the area of our trip, the Antler is expressed by a period of non-deposition followed by deposition of the Mississipian Madison Limestone. Up to 4,000' of Paleozoic rocks are found to drape the crystalline basement.

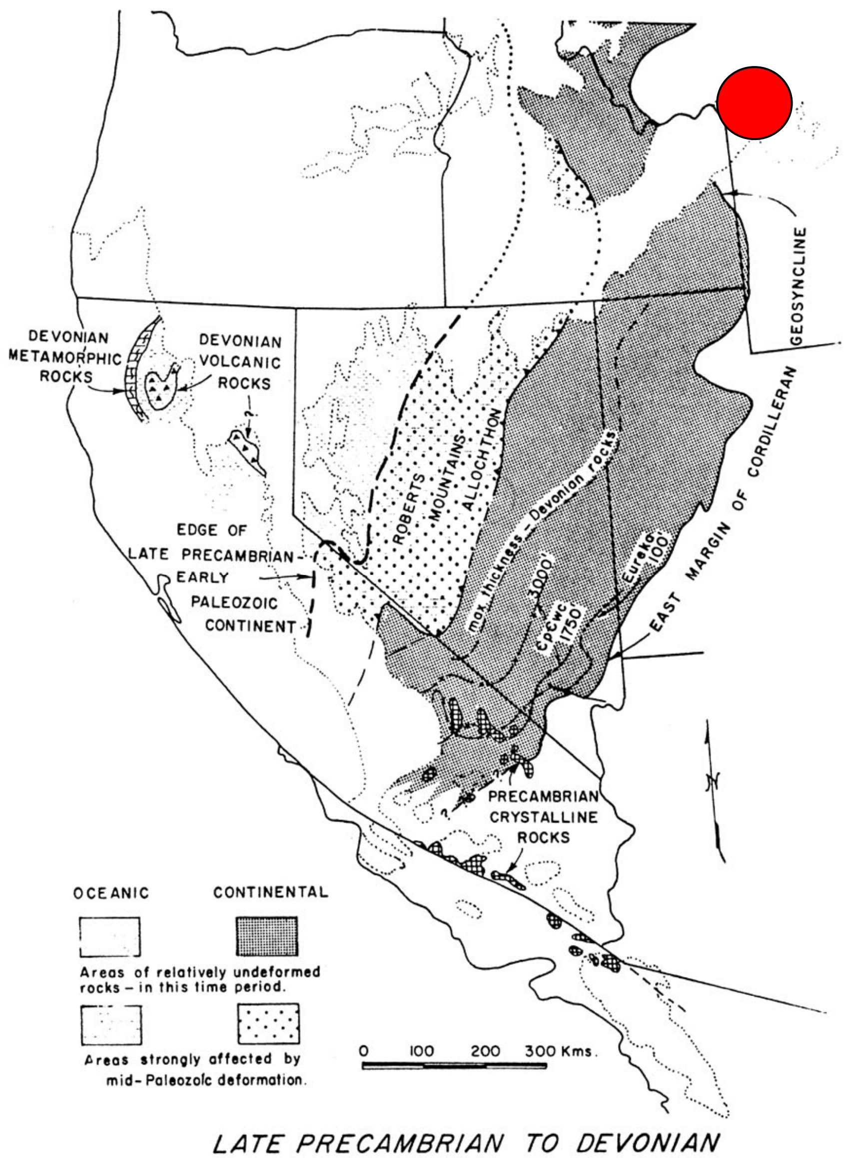


Figure 10 – Map showing Late Proterozoic to Devonian isopachs and position of Roberts Mountain allochthon. Area of field trip circled in red.

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 11 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. The area of our field trip is outside the view of Figure 11, located diagonally away from the NE corner of the sketchmap.

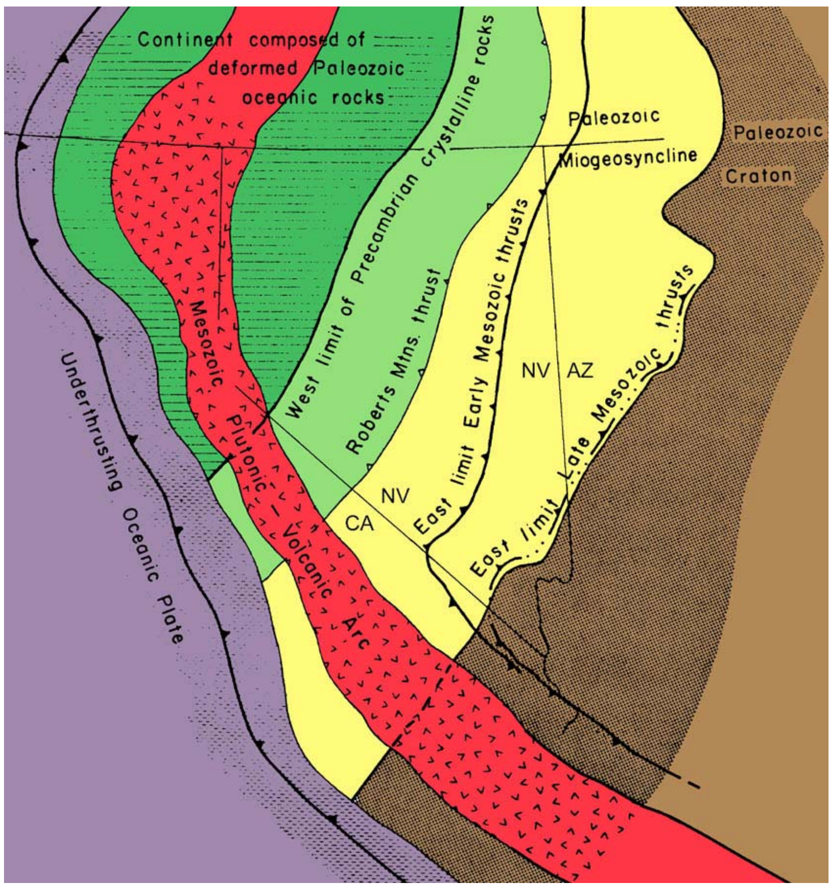


Figure 11 – Geologic sketchmap showing the truncation of Antler, Sonoman, and Sevier thrusts. (Burchfiel and Davis, 1972, Fig. 7.) Area of field trip just off the upper right margin of diagram.

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 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 12). 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 13).

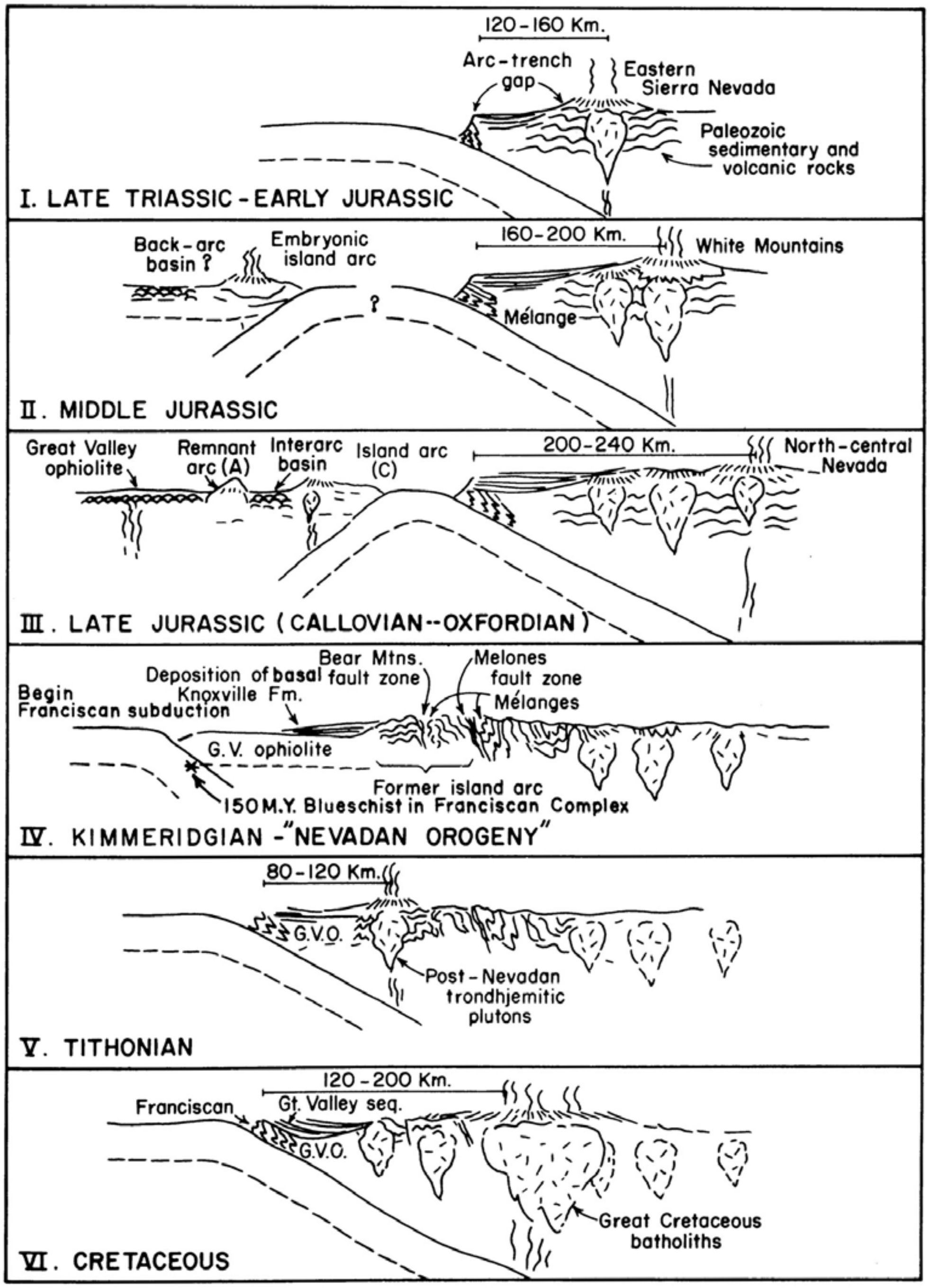


Figure 12 – Plate models to explain Mesozoic tectonics of the SW Cordillera. (From Schweickert and Cowan, 1975.)

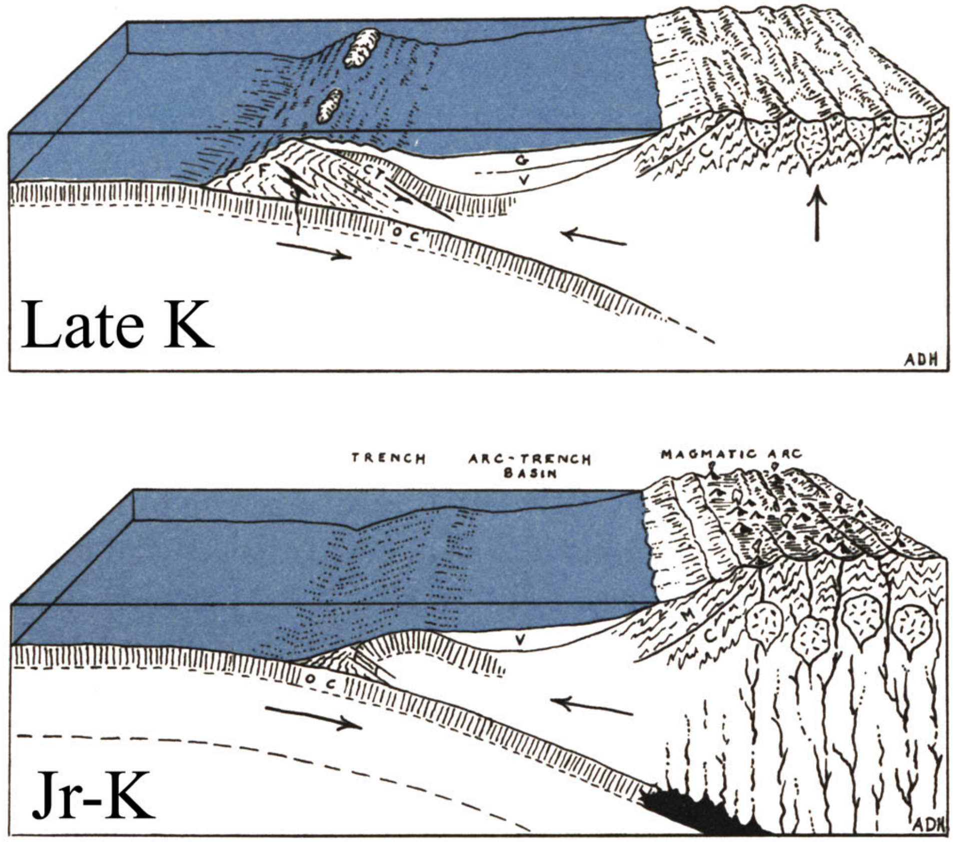


Figure 13 – Block diagrams depicting the growth and evolution of an Andean arc-trench system during Mesozoic time. (Composite diagram adapted from Howard, 1979.)

As 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 11.) The Cenozoic geology of the Cordillera is the result of continued subduction and eventual changeover to a San Andreas type margin, the product of interaction with the subducted spreading East Pacific Ridge (Figure 14).

Sevier and Laramide Orogenies

The **Sevier and Laramide Orogenies** affected the SW Cordillera between earliest Cretaceous and early Eocene time. Wyoming 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 to granulite facies gneisses in massive overthrust sheets. Younger Basin and Range faulting has obscured many of the geologic relationships.

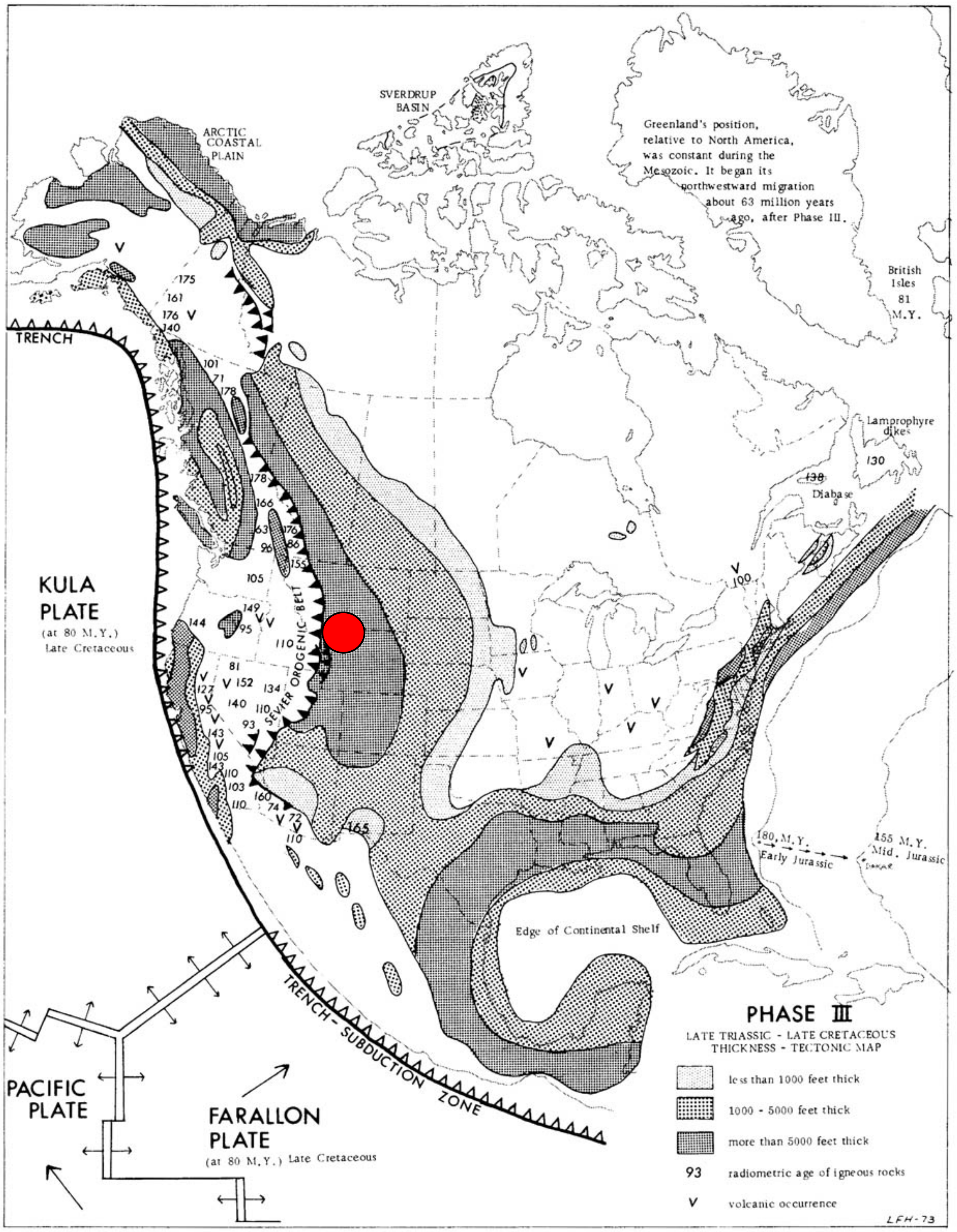


Figure 14 – Paleotectonic map of the Cordillera from Late Triassic to Early Cretaceous time. Area of trip circled.

As far as Wyoming goes, by early Cenozoic time subduction was in a distant Andean setting with development of an elongate volcano-plutonic complex parallel to the present coastline. With flattening of subduction angles during the Cenozoic the volcanism broadened to include the Great Basin and parts of the study area. Here, volcanism absarookly includes the Eocene Absaroka range (54-42 Ma) followed by an interval of periodic volcanity to the present day. 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 and heralded a changeover to extensional tectonics.

By 15 Ma, 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 extensional seismicity. Indeed, the result of subduction of the Pacific Ridge had far-reaching geological consequences, including perhaps development of the Yellowstone hot spot. 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, some with low-angle ductile normal faulting, were formed by thermal weakening of the back arc region.

Basin and Range extensional faulting and the creation of the San Andreas transcurrent fault 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 in the Cordillera. In the Teton-Yellowstone area, the combined influence of active earthquake faulting and an active supervolcano might make one feel as though they are walking around the brim of an upturned, unstable mantle cannon.

Effects of Cordilleran Tectonic Events in Wyoming

In the region of our field trip, Rocky Mountain phase structural elements are shown in Figure 15 in yellow (Sevier) and orange (Laramide). The main difference between Sevier and Laramide structures is their opposition in dip and the fact that Laramide structures are commonly basement cored backthrusts, reflecting an overall weakening of subcrustal strength resulting from prolonged, low-angle Andean subduction along the Cordilleran margin.

During latest Cretaceous and earliest Tertiary times, broad uplifts traversed the region along the same general trend and in the same area as future Laramide structures. One of these, the Teton-Gros Ventre Uplift, is outlined in purple in Figure 15. Broken by post-Laramide faults (red line in Figure 15), the former extent of the Teton-Gros Ventre Uplift is unknown but most workers extend the uplift in toward the Snake River Plain of Idaho. The Cache Creek thrust (Laramide structure) marks the southwestern margin of the Teton-Gros Ventre Uplift and places the terrain above shingled Sevier thrusts of the Snake River Range as shown in the section of Figure 16.

The changeover from compressional tectonics to Basin and Range extension in roughly 9 Ma is responsible for tilting of the Teton block and inducing over 30,000' of composite offset.

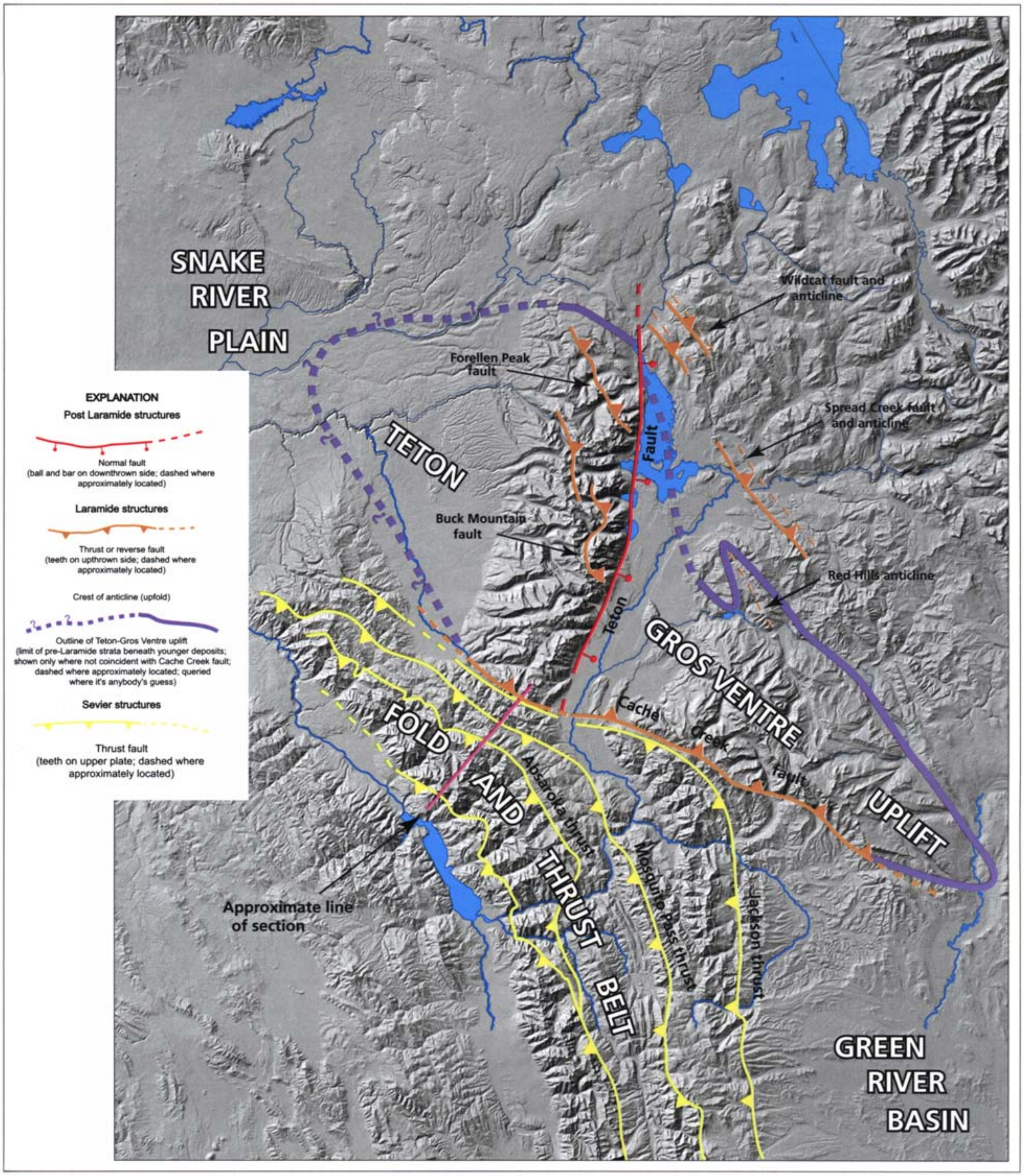


Figure 15 – Shaded relief map showing the extent of the Teton-Gros Ventre uplift (purple line). Lying at the juncture of Sevier (yellow) and Laramide (orange) thrust sheets, the Tetons-Gros Ventre Uplift is bounded on the southwest by the Cache Creek fault. The Jackson thrust delimits the northeastward most extent of Sevier thrusts. Younger extensional faulting along the Teton fault system (red line) cuts the Teton-Gros Ventre Uplift and has resulted in uplift of the Teton block (Adopted from Love et al., 2003, Fig 49A, p. 55.)

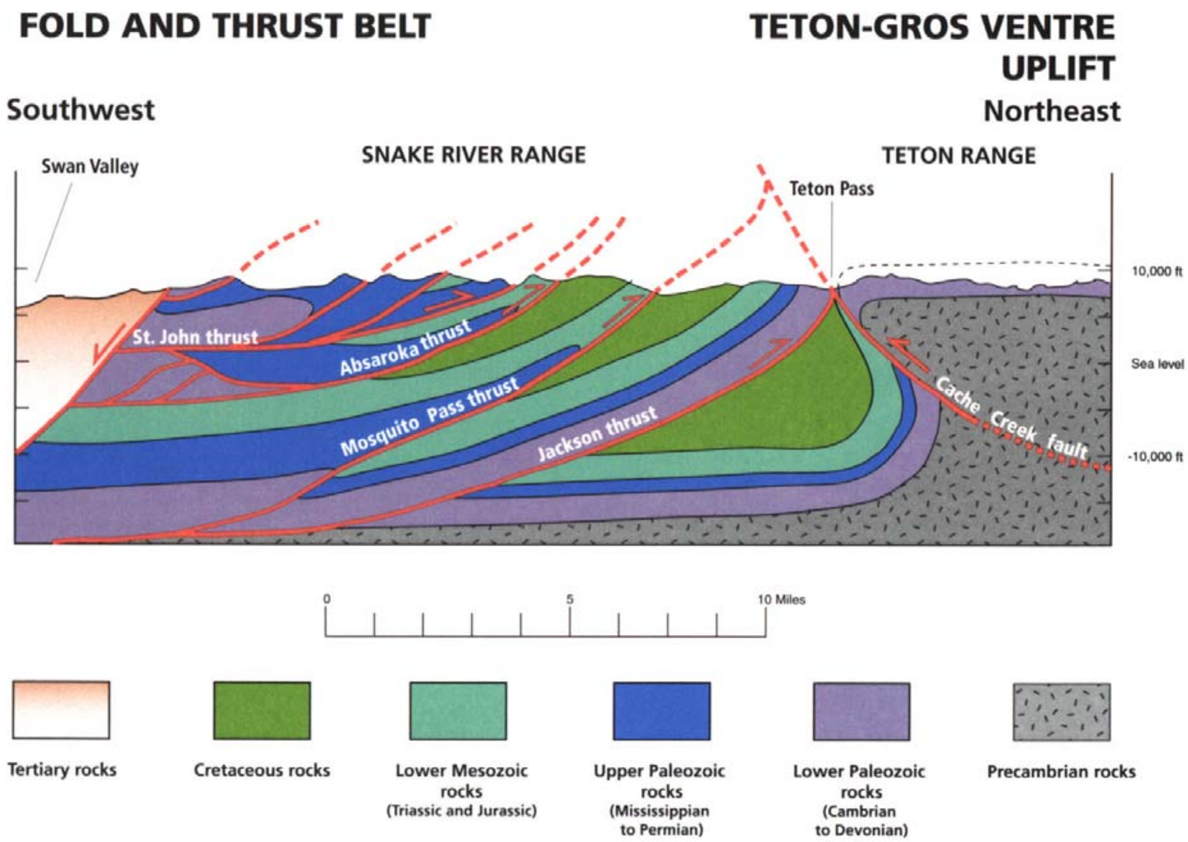


Figure 16 – Geologic section across the Snake River Range from Swan Valley across to the Teton Range showing the staked sequence of thrust sheets, typical of those produced during the 145-73 Ma Sevier Orogeny. (Adapted from Love et al., 2003, fig. 49B, p. 56.)

The Grand Tetons

Located within the Middle Rocky Mountain physiographic province, the Grand Tetons are the youngest and perhaps most awesome range in the Rockies (Figure 17). They form the youngest section of the Rocky Mountains, yet conversely their uplift exposes some of the oldest rock formations in North America. Reminiscent of the Himalayas in this view and with a history of over 2.7 billion years of activity, the area is a delight for visitors and a religious experience for devotees of the rock hammer, boots, and handlens.

Grand Teton rises to 13,770 feet (about 4,200 meters), and twelve of the Teton peaks in the range rise above 12,000 feet (3,660 m), more than a mile above the valley floor below (known as Jackson Hole). The gradual slope on the western side of the peaks reflects the uplifting and tilting of the section of the North American crust shown here. The Snake River originates here, draining into and out of Jackson Lake. Tributaries on the western side of the range also drain into the Snake River where Henry's Fork merges with the main river south and west of the park near Idaho Falls. The Snake River forms the headwaters for the Columbia River, the largest river in the Pacific Northwest.

The Teton Range was set aside as a National Park in 1929 after decades of contention with local ranchers and other land interests, and at that time only included the Teton Range and glacial lakes at their base. The area around Jackson Lake, including national forest and other federal lands, were declared as a national monument in 1943. The two were merged, and a 35,000-acre section added by donation by John D. Rockefeller, Jr., to form the current Grand Teton National Park in 1950.

Figure 17 – Landsat view of Grand Teton National Park and its immediate surroundings from an elevated camera looking at the park from the northeast. Visualization combines simulated natural color imagery from the Landsat 7 Enhanced Thematic Mapper Plus (ETM+) instrument with elevation data derived from the Shuttle Radar Topography Mission (SRTM) and the United States Geological Survey's Digital Terrain Elevation Data (DTED). The elevation extrusion shown here is to proper scale with no vertical exaggeration. These ETM+ data were acquired on September 23, 2002.
(Source: http://earthobservatory.nasa.gov/Newsroom/NewImages/Images/landsat_grand_teton_lrg.jpg.)

Geology of Grand Teton National Park

Geologic structures in Grand Teton National Park are located within the Middle Rocky Mountain physiographic province and result from tectonic activities associated with the Laramide orogeny and continuing through recent times. The Washakie and Absaroka ranges are located northeast of the park and consist of thrust faulted, asymmetric anticlines and Eocene andesitic volcaniclastic rocks, respectively. The Pinyon Peak and Mount Leidy Highlands, east of the park, include conglomerates of late Cretaceous and early Tertiary age. The uplands east of Jackson Hole are underlain by Precambrian through Cenozoic rock strata.

The Gros Ventre Range, located southeast of the park, is composed of thrust faulted sedimentary rocks of Mesozoic and Paleozoic age, including Carboniferous units which form local karst. By contrast, the Teton Range is an upthrown, tilted fault-block that contains more than 5,000 feet of Paleozoic sedimentary strata including significant deposits of karst forming limestones and dolostone. (See Figures 3 and 4.)

Jackson Hole is a structural basin as much as 5 km deep formed by a tilted, downthrown normal fault block hinged to the east (Figure 18).

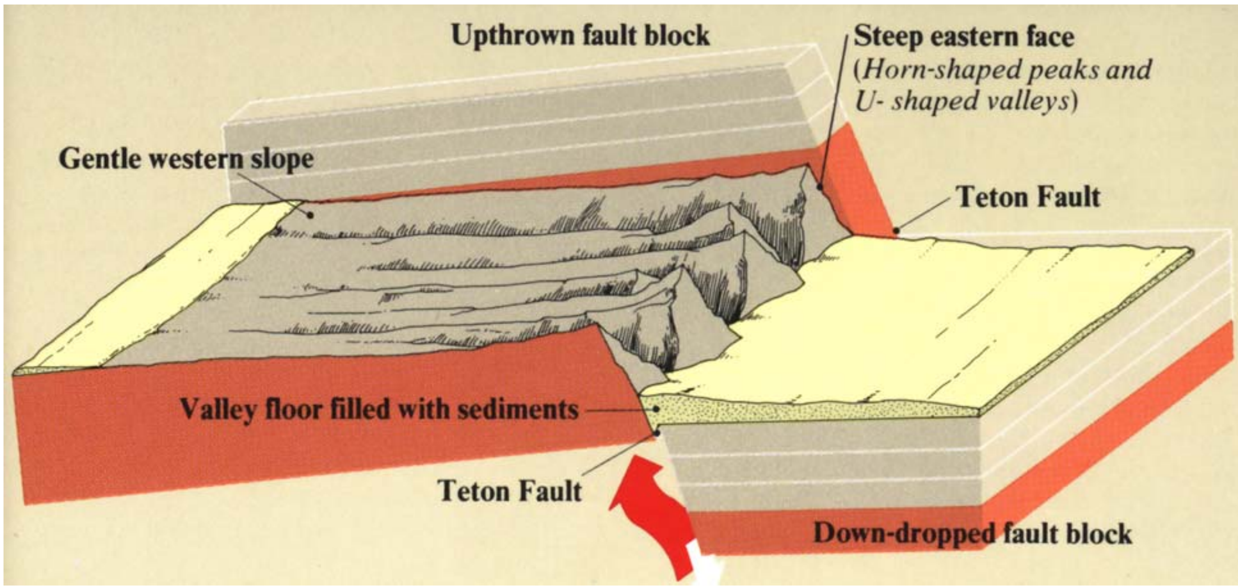


Figure 18 - Block diagram illustrating the generalized structure of the Teton range, a fault block mountain. (Adapted from Handbook 122, p. 31, Official National Park Handbook, Grand Teton National Park, 2002, 96 p.)

Pre-Paleozoic History

The geologic story of the Teton Range starts with the formation of rocks that make up the core of the mountains, rocks far older than the mountains themselves known as the Wyoming Province. (See Figure 1.) Perhaps 3000 million years ago in Archean time, sand, limey ooze, silt and clay were deposited in a marine trough (accurate dating is not possible, because of subsequent recrystallization). Interbedded volcanic deposits indicate the presence of a nearby island arc. These strata were later lithified into sandstones, limestones, and various shales. Studies indicate deformation of core rocks of the Tetons range took place 2800 to 2700 Ma at former depths of 10-15 km and temperatures of about 575°C. Orogenic folding and metamorphism, creating alternating light and dark banded gneiss, amphibolite, schist, and minor serpentinite. Today, alternating light and dark layers identify banded gneiss, readily seen in Death Canyon and other canyons in the Teton Range. The green to black serpentine created was used by Native Americans to make bowls and stream-polished serpentine pebbles are locally called “Teton jade”.

Next, at around 2500 Ma, granitic magmas forced its way up through cracks and zones of weakness in the gneiss. This igneous rock intruded as light-colored dikes and larger plutons inches to hundreds of feet thick. Collectively, the granitic intrusives at 2500 Ma mark a period of global granitization of continents known as the Algoman Intrusive Episode. Extensive exposures of this rock are found in the central part of the range. Look for larger dikes as you view the mountains from the Jenny Lake and String Lake areas. Uplift and erosion have exposed the granites that now form the central peaks of the range.

About 1400 to 785 Ma, 5' to 200' thick (1.5 to 60 m) thick black diabase dikes intruded the Algoman granites, forming the prominent vertical dikes seen today on the faces of Mount Moran and Middle Teton (the dike on Mount Moran is 150' [46 m] wide). The mafic dike on Mount Moran exhibits a diabasic texture and protrudes from the face because the gneiss surrounding it eroded faster than the diabase. A diabase dike on the Middle Teton is recessed because the granite of the central peaks erodes more slowly than the diabase. Some of the large dikes can be seen from the Jenny Lake and String Lake areas. A long period of erosion then began, creating a major gap in the geologic record called an unconformity. The landscape was eventually reduced to a flat plain or expanse of rolling hills.

Paleozoic History

Deposition resumed in the Cambrian period and continued throughout the Paleozoic era, creating nine major formations which together are over 4000' (1200 m) thick. Indeed, the only geologic period in the Paleozoic not represented is the Silurian. Local pre-Pennsylvanian formations are listed in Table 2. The Paleozoic strata were deposited in a shallow sea and later became a discontinuous mix of dolomite, limestone, sandstones, and shale (Figure 19). The strata are relatively non deformed for their age even though periodic upwarp exposed them to erosion, creating local unconformities. Fossilized brachiopods, bryozoans, corals, and trilobites are found in the carbonate rock layers with the best examples found outside the park in the Alaska Basin. The most complete examples of this unit are found to the west, north, and south of park borders. Geological maps and sections for the park are in Figures 20, 21, and 22.

The shallow seas that covered the Teton region 600 million to 65 million years ago have left sedimentary formations, still visible at the north and south ends of the Teton Range and also on the west slope of the mountains. The seas repeatedly advanced and retreated. Marine life, especially tiny trilobites, corals and brachiopods, flourished in the shallow seas that covered this area. During Mesozoic retreat of the Paleozoic seas, the area became a low-lying coastal plain frequented by dinosaurs. Fossilized bones of a horned dinosaur, the Triceratops, have been found east of the Park near Togwotee Pass.

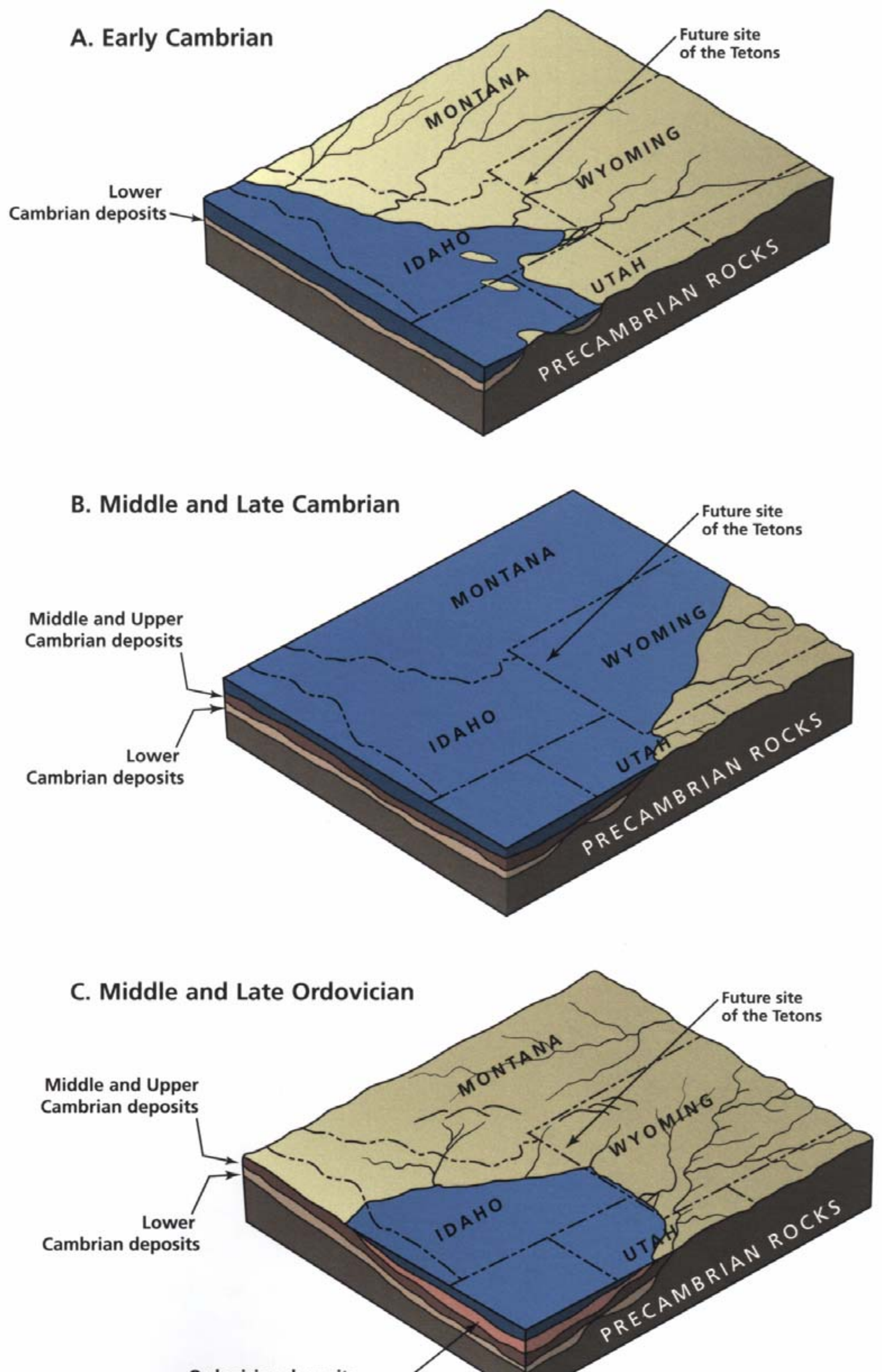


Figure 19 – Cambrian to Ordovician paleogeography of the study area. (Adapted from Love et al., 2003, fig. 34, p. 43.)

Early Paleozoic strata of the Teton region

Column labeled "Map symbol" refers to the symbol used for the unit on the Geologic map of Grand Teton National Park (Love, Reed, and Christiansen, 1992)

AGE (Numbers show age in millions of years)	FORMATION (Thickness in feet)	ROCKS AND FOSSILS	WHERE FOUND	MAP SYMBOL
MISSISSIPPIAN (354-366)	MADISON LIME-STONE (about 1100)	Uniform thin beds of blue-gray limestone and sparse very thin layers of shale. Brachiopods, corals, and other fossils abundant	Northwestern, western, and southern flanks of the Teton Range. Also on Blacktail Butte, the Gros Ventre Buttes, and in the Gros Ventre Range	Mm
LATE AND MIDDLE DEVONIAN (368-366)	DARBY FORMATION (about 350)	Thin beds of gray and buff dolomite interbedded with layers of gray, yellow, and black shale. A few bra-chiopods, corals, and bryozoans	Northwestern, western, and southern flanks of the Teton Range. Also on Blacktail Butte, the Gros Ventre Buttes, and in the Gros Ventre Range	Dd
LATE AND MIDDLE ORDOVICIAN (454-445)	BIGHORN DOLOMITE (about 450; Leigh Dolomite Member about 40 feet thick at top)	Thick to very thin beds of blue-gray or brown dolomite, white on weathered surfaces. A few broken brachiopods, bryozoans, and horn corals. Thin beds of white fine-grained limestone at top are Leigh Member	Northwestern, western, and southern flanks of the Teton Range. Also on Blacktail Butte, the Gros Ventre Buttes, and in the Gros Ventre Range	Ob
LATE CAMBRIAN (506-495)	GALLATIN LIMESTONE (180)	Blue-gray limestone mottled with irregular rusty or yel-low patches	Northwestern, western, and southern flanks of the Teton Range. Also on Blacktail Butte and in the Gros Ventre Range	€g, €gg[2]
MIDDLE CAMBRIAN (510-506)	GROS VENTRE FORMATION (605)	PARK SHALE MEMBER AT TOP (220 FT): Gray-green shale containing beds of platy limestone conglomerate. Trilobites, brachiopods, and algal heads DEATH CANYON LIMESTONE MEMBER (285 FT) IN MIDDLE: Two thick beds of dark-blue-gray limestone separated by 15-20 ft. of shale that locally contains abundant bra-chiopods and trilobites WOLSEY SHALE MEMBER (100 FT) AT BASE: Soft greenish gray shale containing beds of purple and green sandstone near base. A few brachiopods	Northwestern, western, and southern flanks of the Teton Range and in the Gros Ventre Range	€gg[2] €gd €gf
	FLATHEAD SANDSTONE (175)	Brown, maroon, and white sandstone, locally containing many pebbles of quartz and feldspar. Some beds of green shale at top	Northwestern, western, and southern flanks of the Teton Range and in the Gros Ventre Range	€f
PRECAMBRIAN (OLDER THAN 540)		Granite, gneiss, and pegmatite		

Table 2 – Early Paleozoic strata of the Teton region. (From Good et al., 2003, Table 3, p. 41.)

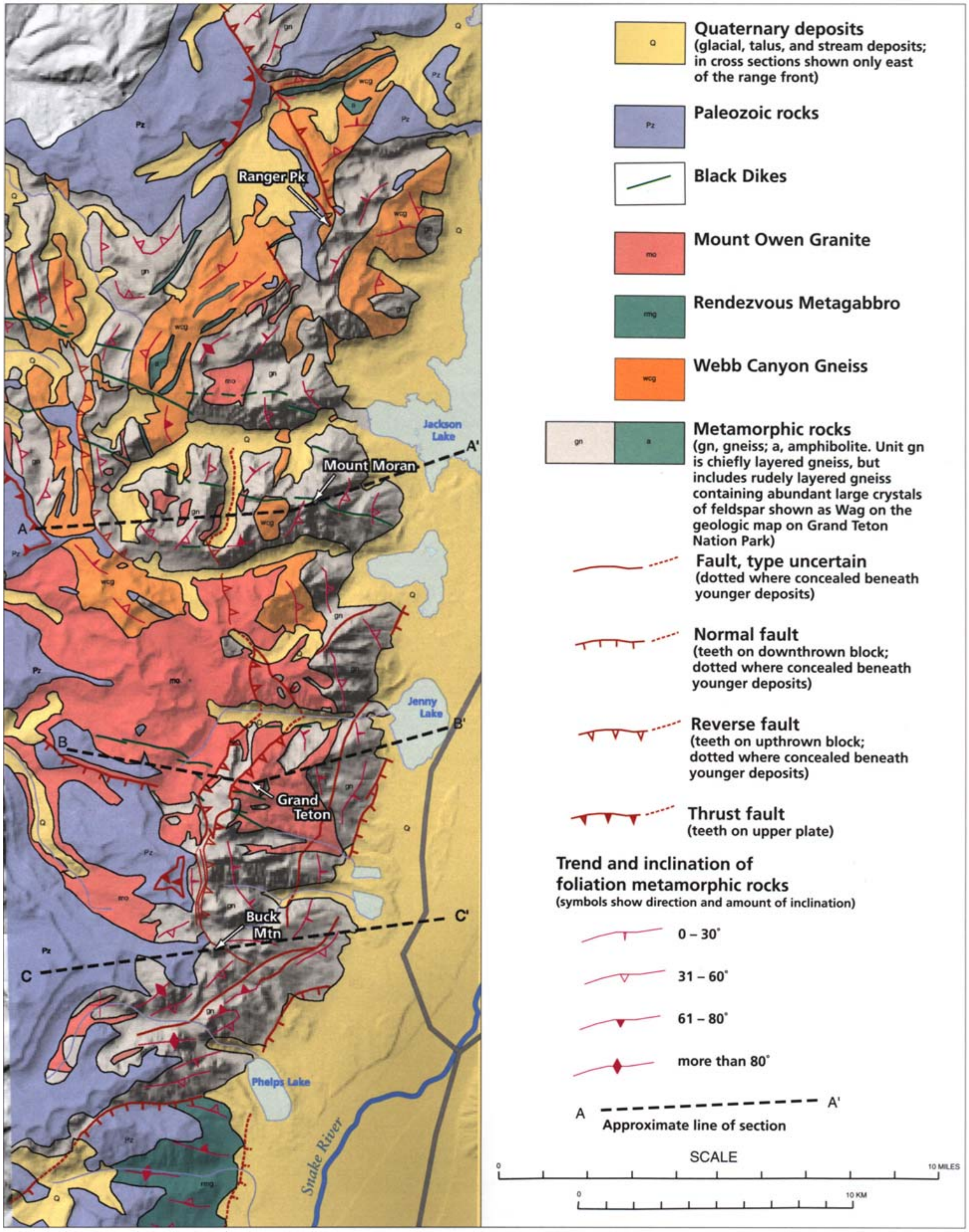


Figure 20 – Generalized geologic map of the Grand Teton showing the distribution of Archean and younger units of the range. Lines of section are keyed to Figure 21. (Adapted from Love et al., 2003, fig. 17, p. 31.)

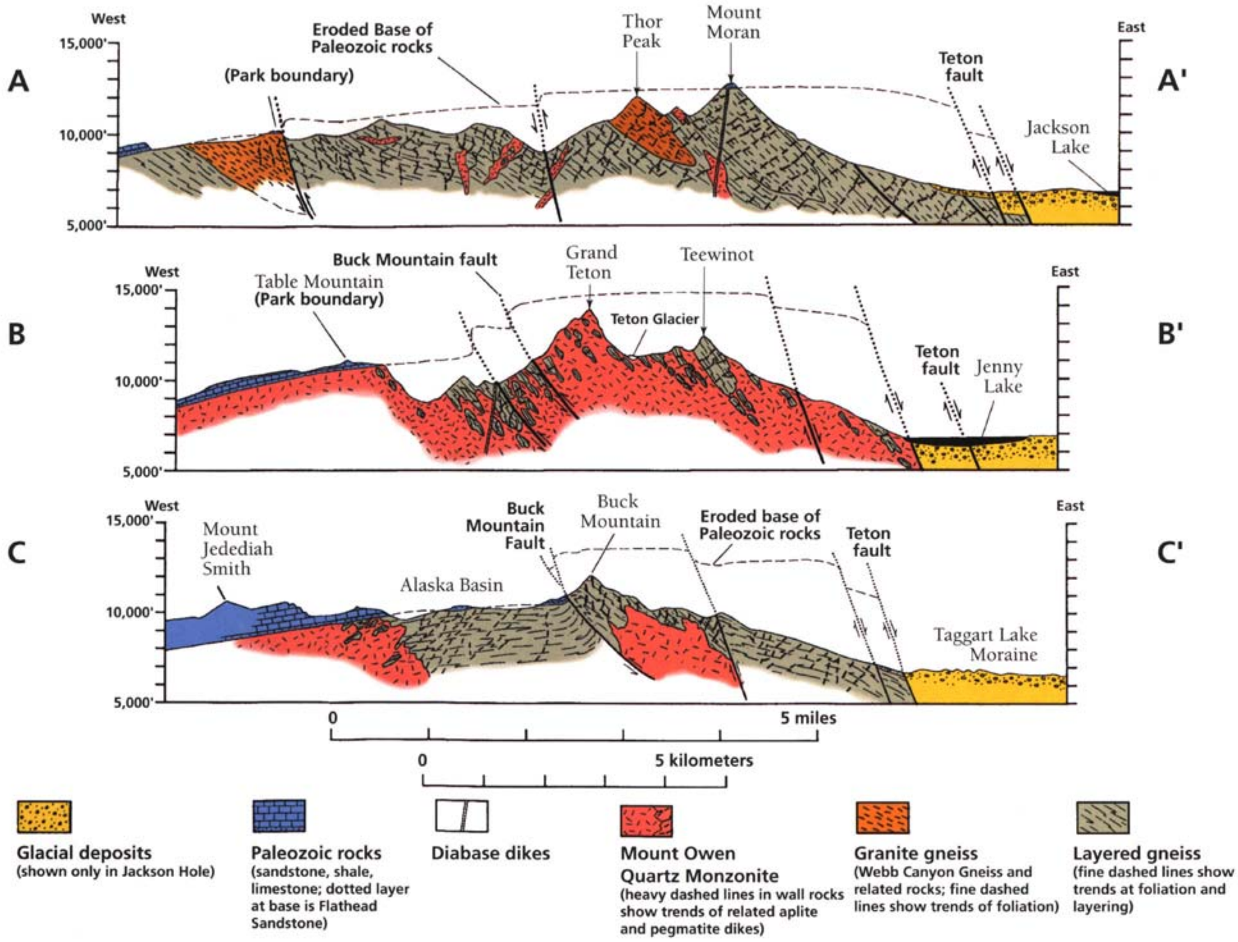


Figure 21 – Geological section keyed to map of Figure 20 showing the extent and structure of Archean and Proterozoic rocks of the Grand Teton Range. (Adapted from Love et al., 2003, fig. 17, p. 30.)

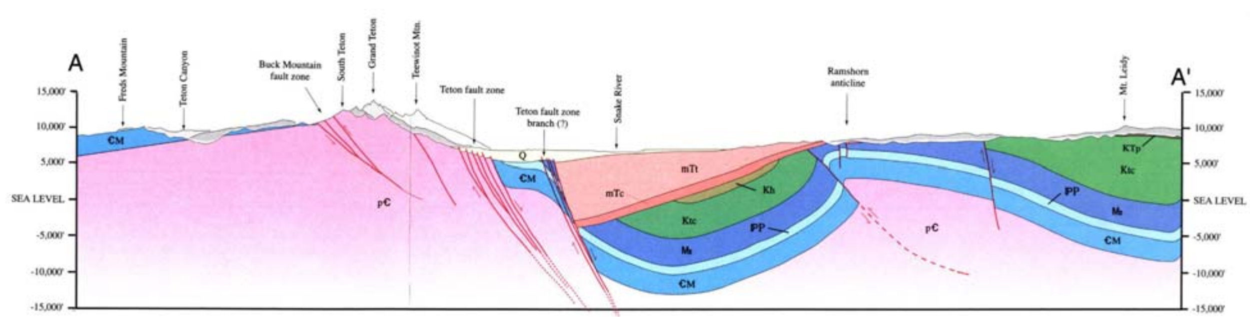
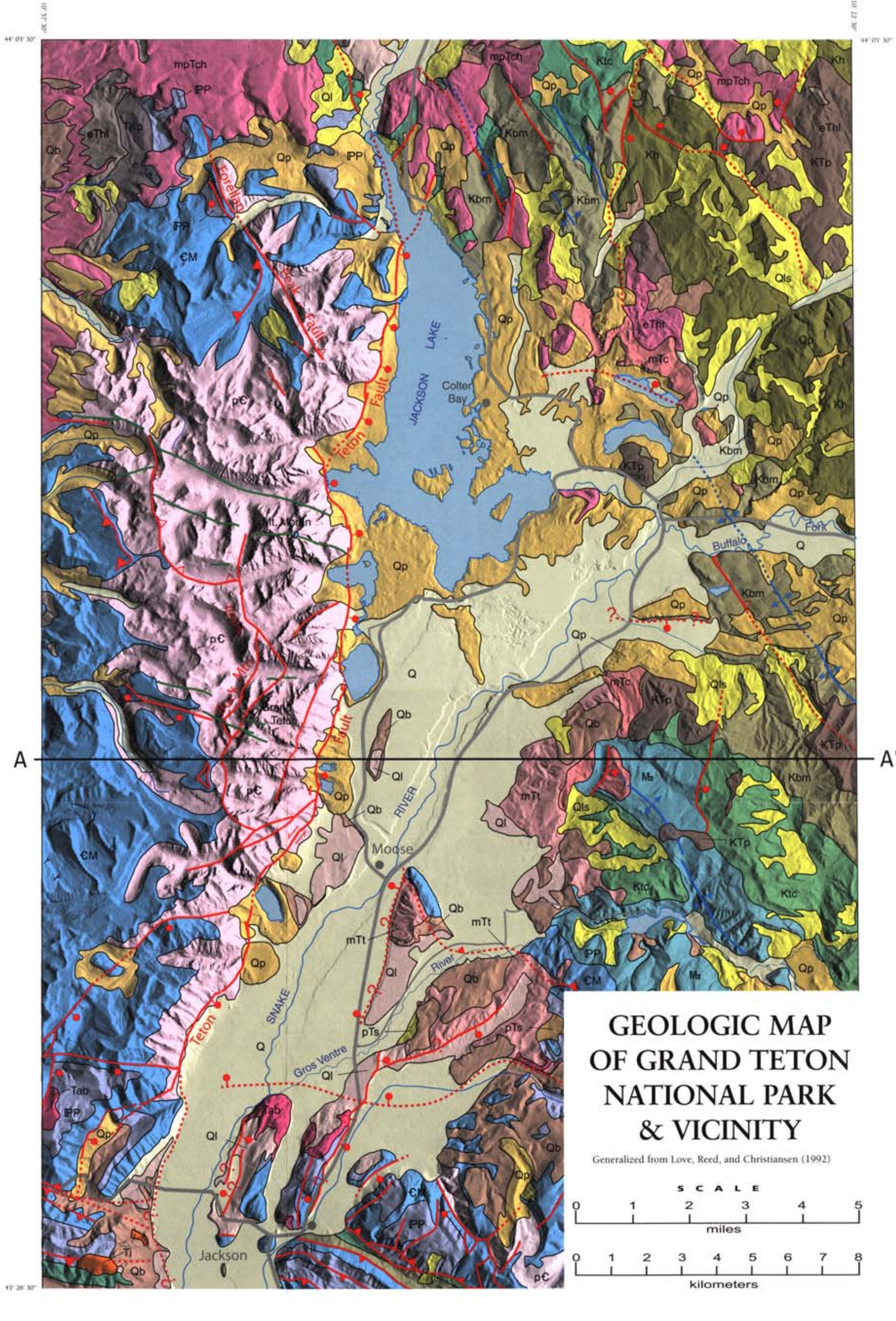


Figure 22 – Generalized geologic section and map (continued on next page) of the Grand Teton Range. (From Love et al., 2003, Plate 1.)


GEOLOGIC MAP
OF GRAND TETON
NATIONAL PARK
& VICINITY
Generalized from Love, Reed, and Christiansen (1992)
SCALE
0 1 2 3 4 5
miles
0 1 2 3 4 5 6 7 8
kilometers
A
A'
JACKSON LAKE
Colter Bay
Teton Fault
Moose
Jackson
SNAKE RIVER
Gros Ventre River
Buffalo Fork
Mt. Moran
Grand Teton

Mesozoic History

Mesozoic deposition changed from primarily marine to a mix of marine, transitional, and continental that varied over time as crustal conditions altered the region. By the close of this era, 10,000’ to 15,000’ (3000 to 4500 m) of sediment accumulated in 15 recognized formations. The most extensive non-marine formations were deposited in the Cretaceous period when the eastern part of the Cretaceous Seaway (a warm shallow sea that periodically divided North America in that period) covered the region. Their composite blanket of sediment came from rock eroded from mountains east of the seaway with ash from volcanos west of the seaway in the Sierran Arc (a long volcanic island chain like the modern Andes Mountains but in island form). This interbedded ash eventually became bentonite, a clay which expands in water and thus causes landslides in the park.

Regional uplift in latest Cretaceous time caused the seaway to retreat and transformed the Grand Teton area into a low-lying coastal plain that was frequented by dinosaurs. Coalbeds were eventually created from the swamps and bogs left behind after the last stand of the seaway retreated. Coal outcrops can be found near abandoned mines in and outside of the eastern margin of the park. Outcrops of older Mesozoic-aged formations can be found north, east, and south of the park.

Tertiary History

The uplift responsible for erasing the Cretaceous Seaway and fusing the Sierran arc to the rest of North America was caused by the Laramide Orogeny (Figure 23). Starting 70 million years ago and lasting well into the first half of the Cenozoic era, the Laramide was the main mountain-building episode creating the Rocky Mountains. Compressive forces from this orogeny created north-south-trending thrust faults along with general regional uplift. Erosion of the Targhee uplift north of park borders was driven by steepened stream gradients. Gravel, quartzite cobbles, and sand from this erosion eventually became the 5000’ (1500 m) thick Harebell Formation seen today as various conglomerates and sandstones in the northern and northeastern parts of the park.

In the Paleocene epoch large amounts of clastic sediment derived from uplifted areas covered the Harebell Formation to become the Pinyon Conglomerate. The lower members of this formation consist of coal beds and claystone with conglomerate made of quartzite from the Targhee uplift above.

Uplift intensified and climaxed early in the Eocene epoch when large thrust and reverse faults created small mountain ranges separated by subsiding basins. One of the reverse faults, the north-south trending 10 mile (16 km) long Buck Mountain Fault, elevated what is today the central part of the Teton Range and one of the basins created was Jackson Hole. Subsidence of Jackson Hole and other basins in the area provided a resting place for more and more sediment (subsidence kept up with deposition at Jackson Hole). Erosion of the up-faulted ranges provided much of that sediment.

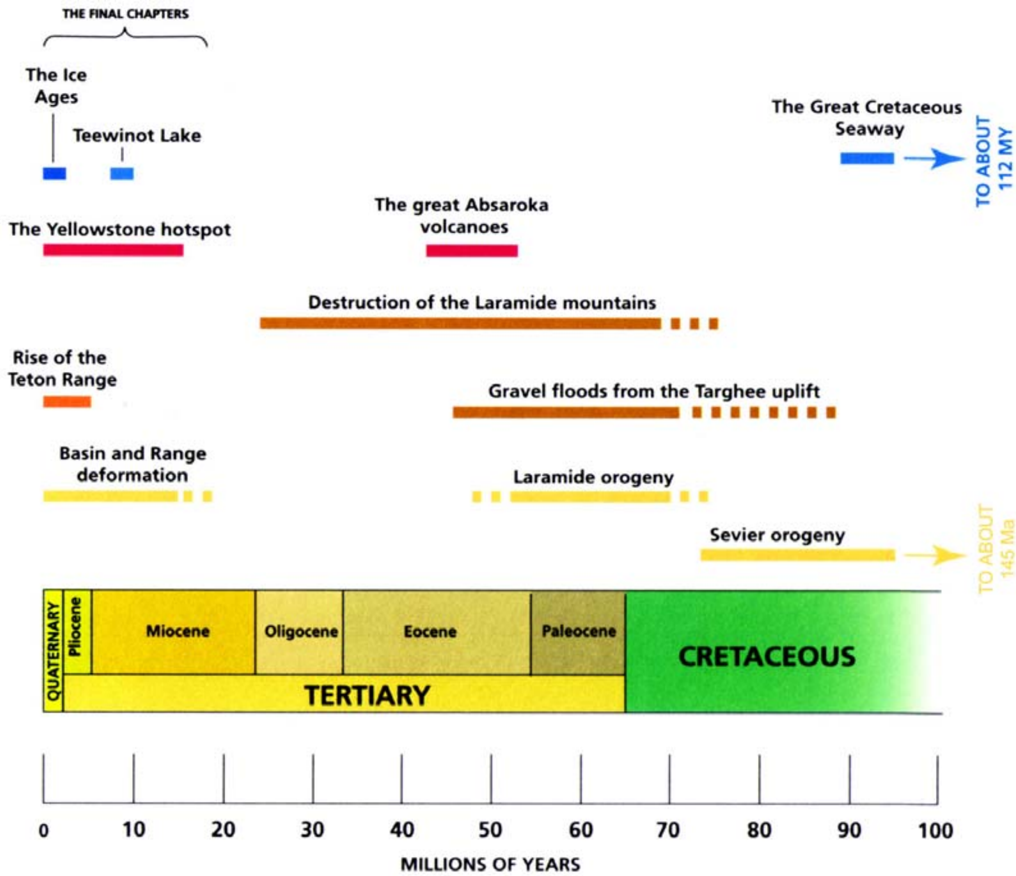


Figure 23 – Geologic events during Cretaceous and younger time in the study area. (Adapted from Love et al., 2003, fig 55, p. 62.)

Waning of the Laramide orogeny in the Eocene coincided with volcanic eruptions from two parallel volcanic chains separated by a long valley in the Yellowstone-Absaroka area to the north. Huge volumes of volcanic material such as tuff and ash accumulated to great depth in the Grand Teton area, forming the **Absaroka Volcanic Supergroup,** absarookly. Additional eruptions east of Jackson Hole deposited their own debris in the Oligocene and Miocene epochs.

During the Miocene, roughly starting 15 Ma, development of the Basin and Range province began in earnest with a broad zone of extensional tectonics largely west of the Tetons (Figure 24). Also during Miocene time, about 9 million years ago, a 40 mile (64 km) long steeply east dipping normal fault system deepened the Jackson Hole basin and started to uplift the westward-tilting eastern part of the Teton Range (the youngest mountain range in the Rocky Mountains). Eventually all the Mesozoic rock from the Teton Range was stripped away and the same formations in Jackson Hole were deeply buried. A prominent outcrop of the pink-colored Flathead Sandstone exits 6,000 feet (1830 m) above the valley floor on the summit of Mount Moran. Drilling in Jackson Hole found the same formation 24,000' (7300 m) below the valley's surface, indicating that the two blocks have been displaced 30,000' (9100 m) from each other. Thus, an average of one foot of movement occurred every 300 years (1 cm per year on average).

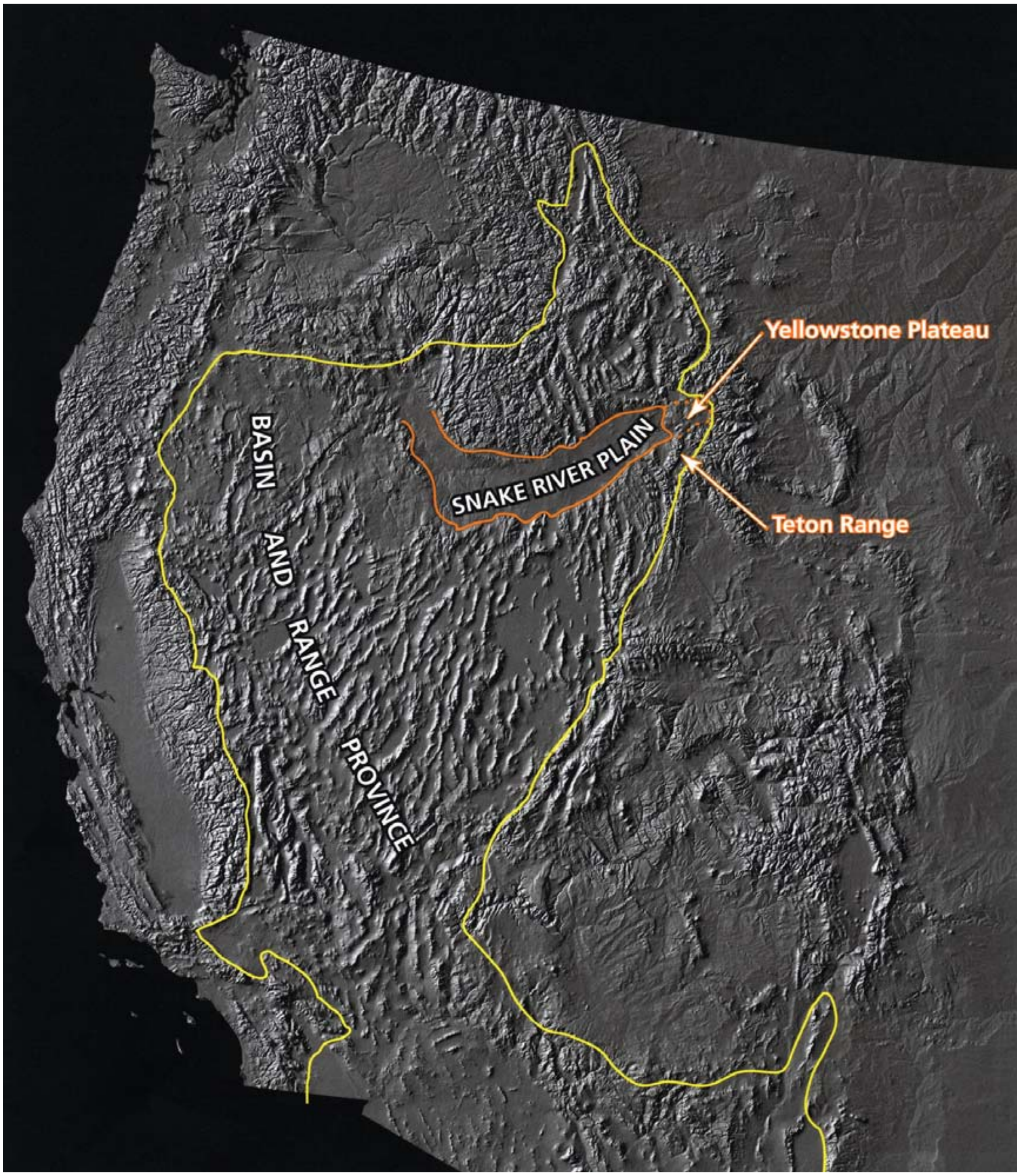


Figure 24 – Shaded relief map of the western United States showing the extent of the Miocene to Present Basin and Range Province. (Adapted from Love et al., 2003, fig. 63, p. 73.)

During Pliocene time, Jackson Hole's first large freshwater lake was impounded by east-west fault movement in what is today the southern end of the park. Geologists call this fault-scarp dammed body of shallow water **Lake Teewinot** and it persisted for around 5 million years. Fossilized clams and snails are found in the lakebed sediments. All told, sediments in the Tertiary period attained an aggregate thickness of around 6 miles (10 km), forming the most complete non-marine Tertiary geologic column in the United States. Most of these units within the park are, however, buried under younger deposits.

Stretching and thinning of the earth's crust caused movement along the Teton fault to begin about 9 Ma. Every few thousand years, when the elasticity of the crust stretches to its limit, a fault (or break) of about 10’ (3 m) occurs, relieving stress in the earth's crust. The blocks on either side of the fault move with the west block swinging skyward to form the Teton Range, the youngest and most spectacular range in the Rocky Mountain chain. The eastern block dropped downward, forming the valley called Jackson Hole. The valley block under your feet has actually dropped down four times more than the mountain block has uplifted.

Early nineteenth century fur trappers referred to high mountain valleys as "holes". When they named this valley Jackson Hole, they were geologically correct! Today the sheer east face of the Teton Range, rising abruptly more than a mile above the valley, captures our attention more than the valley does. Rocks and soil, thousands of feet thick, transported into the valley over the past several million years, mask the subsidence of the valley.

Some of the deposits filling Jackson Hole contain innumerable rounded rocks varying in color from white to pink and purple. These quartzite rocks were eroded from an ancestral mountain range probably located 20 to 70 miles northwest of the Teton Range (the Targhee Uplift). Rivers rounded the quartzite into cobblestones as they carried the rocks into this area.

Quaternary History – Period of “Fire and Ice”

Massive volcanic eruptions from the Yellowstone Volcano northwest of the area occurred 2.2 million, 1.3 million, and ~ 630,000 years ago. Each catastrophic caldera-forming eruption was preceded by a long period of more conventional but less voluminous eruptions. One such event sent large amounts of rhyolitic lava into the northern extent of Teewinot Lake. The resulting obsidian (volcanic glass) has been potassium-argon dated to 9 Ma and was used by native Americans thousands of years ago to make arrowheads, knifes, and spear points. The lake was dry by the time a series of enormous pyroclastic flows from the Yellowstone area buried Jackson Hole under welded tuff. Older exposures of this tuff are exposed in the Bivouac Formation at Signal Mountain and Pleistocene-aged tuffs are found capping East and West Gros Venture Buttes (both the mountain and buttes are small fault blocks).

Climatic conditions in the area gradually changed through the Cenozoic as continental drift moved North America northwest from a sub-tropical to a temperate zone by the Pliocene epoch. The onset of a series of glaciations in the Pleistocene epoch saw the introduction of large glaciers in the Teton and surrounding ranges, which flowed all the way to Jackson Hole during at least three ice ages.

The first and most severe of the known glacial advances in the area was caused by the **Buffalo glaciation**. In that event the individual alpine (mountain valley) glaciers from the east side of the Tetons coalesced to form a 2000’ (610 m) thick apron of ice that overrode and abraded Signal Mountain and the other three buttes at the south end of Jackson Hole (Figure 25). Similar dramas were repeated on other ranges in the region, eventually forming part of the Canadian Ice Sheet, which at its maximum, extended into eastern Idaho. This continental-sized glacial system stripped all the soil and vegetation from countless valleys and many basins, leaving them a wasteland of bedrock strewn with boulders after the glaciers finally retreated. Parts of Jackson that were not touched by the following milder glaciations still cannot support anything but the hardiest plants (smaller glaciers deposit glacial till and small rocks relatively near their source, while continental glaciers transport all but the largest fragments far away).

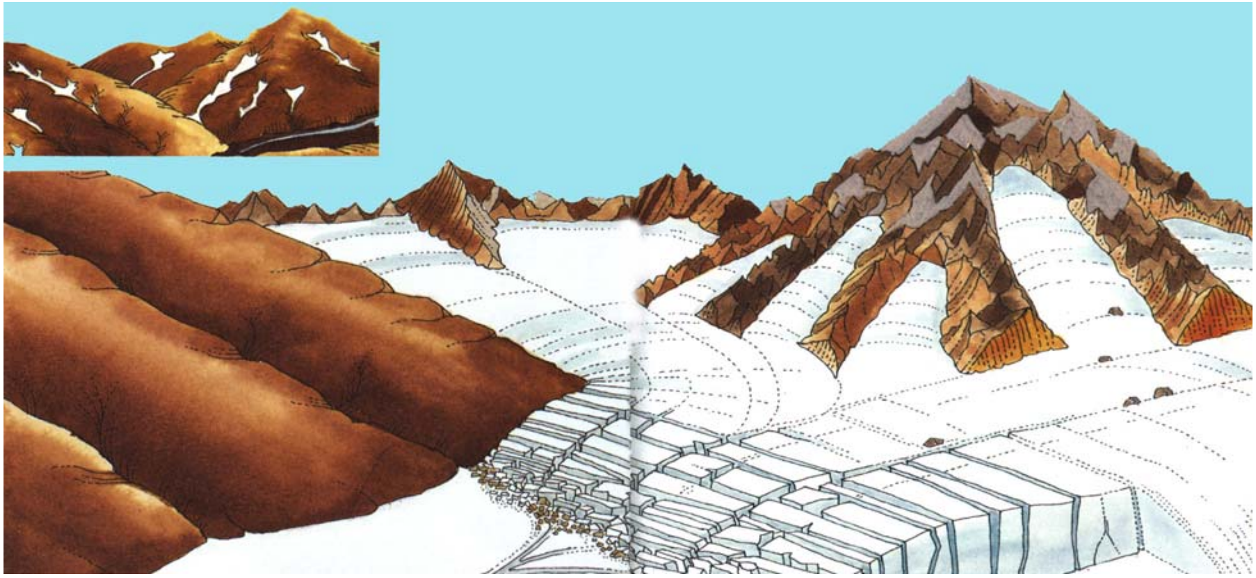


Figure 25 – As massive Ice Age glaciers sculpted the landscape and flowed through the Teton Range, the ice changed steep v-shaped water-cut valleys (inset) into the distinctive U-shape seen in the Teton canyons today. (Adapted from Handbook 122, p. 30-31, Official National Park Handbook, Grand Teton National Park, 2002, 96 p.)

A less severe glaciation, known as **Bull Lake Glaciation**, started sometime between 160 to 130 thousand years. Bull Lake helped repair some of the damage of the Buffalo event by forming smaller glaciers which deposited loose material over the bedrock. In that event, the large glacier which ran down Jackson Hole only extended just south of where Jackson, Wyoming now sits and melted about 100,000 years ago.

Then from 25,000 to 10,000 years ago the lower volume **Pinedale Glaciation** carved many of the glacial features seen today (Figures 26, 27, 28). Burned Ridge is made of the terminal moraine (rubble dump) of the largest of these glaciers to affect the area. Today this hummocky feature is covered with trees and other vegetation. Smaller moraines from a less severe part of the Pinedale were formed just below the base of each large valley in the Teton Range by alpine glaciers.

Many of these piles of glacial rubble created depressions that in modern times are filled with small but deep lakes (Leigh, String, Jenny, Bradley, Taggart, and Phelps). Jackson Lake is the largest of these and was impounded by a recessional moraine left by the last major glacier in Jackson Hole. A collection of kettles (depressions left by melted stagnant ice blocks from a retreating glacier) south of the lake is called the Potholes. The basins that hold Two-Ocean Lake and Emma Matilda Lake were created during the Bull Lake glaciation. Since then humans have built a dam over Jackson Lake's outlet to increase its size for recreational purposes.

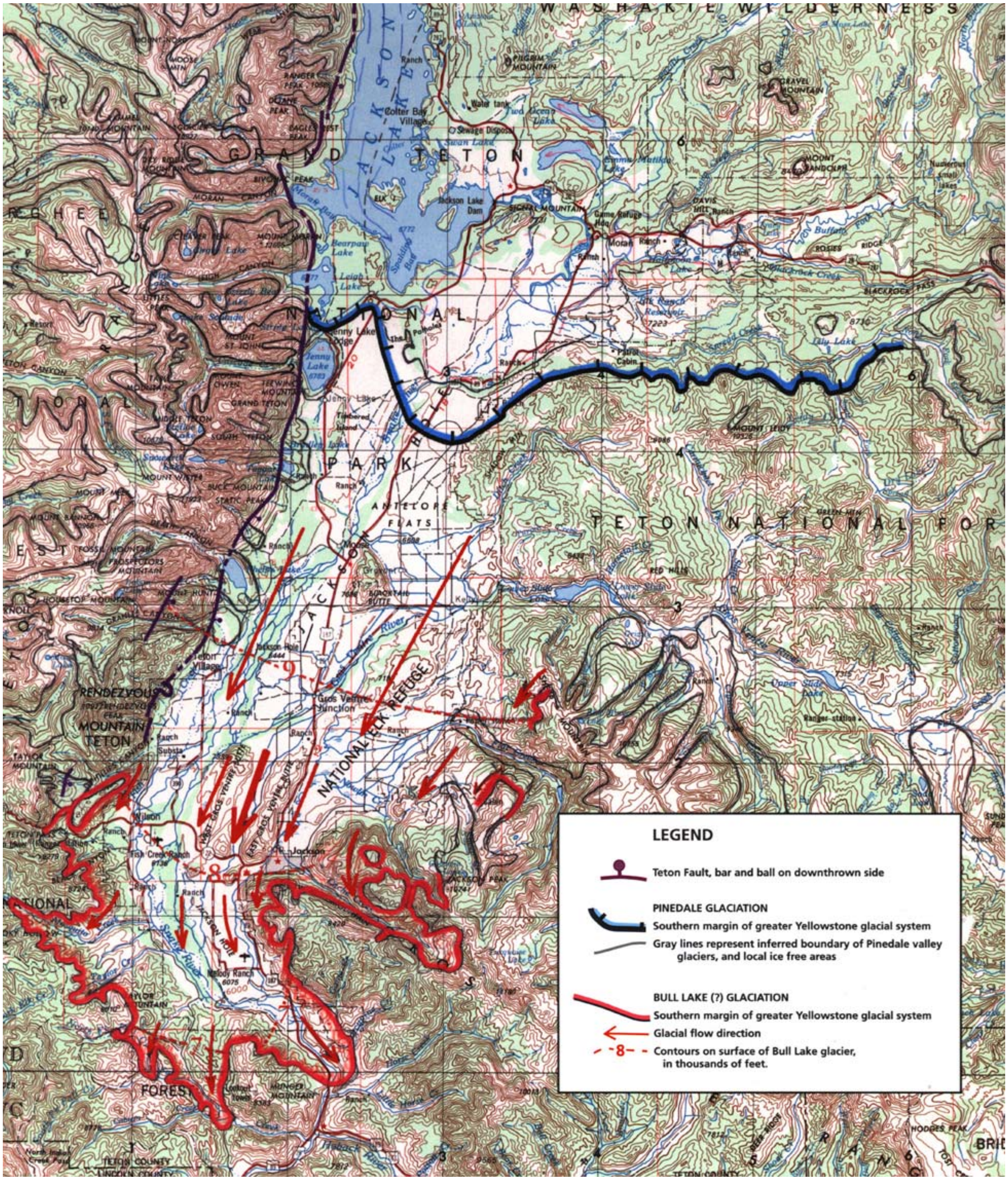


Figure 26 – Map showing the extent of the Bull Lake Glaciation (red) that filled Jackson Hole. Contours show ice thickness in thousands of feet. The blue line shows the southern extent of the greater Yellowstone (Pinedale) glaciation and indicates that it stopped 30 miles short of its Bull Lake extent. (Adapted from Good et al., 2003, fig. 82, p. 94.)

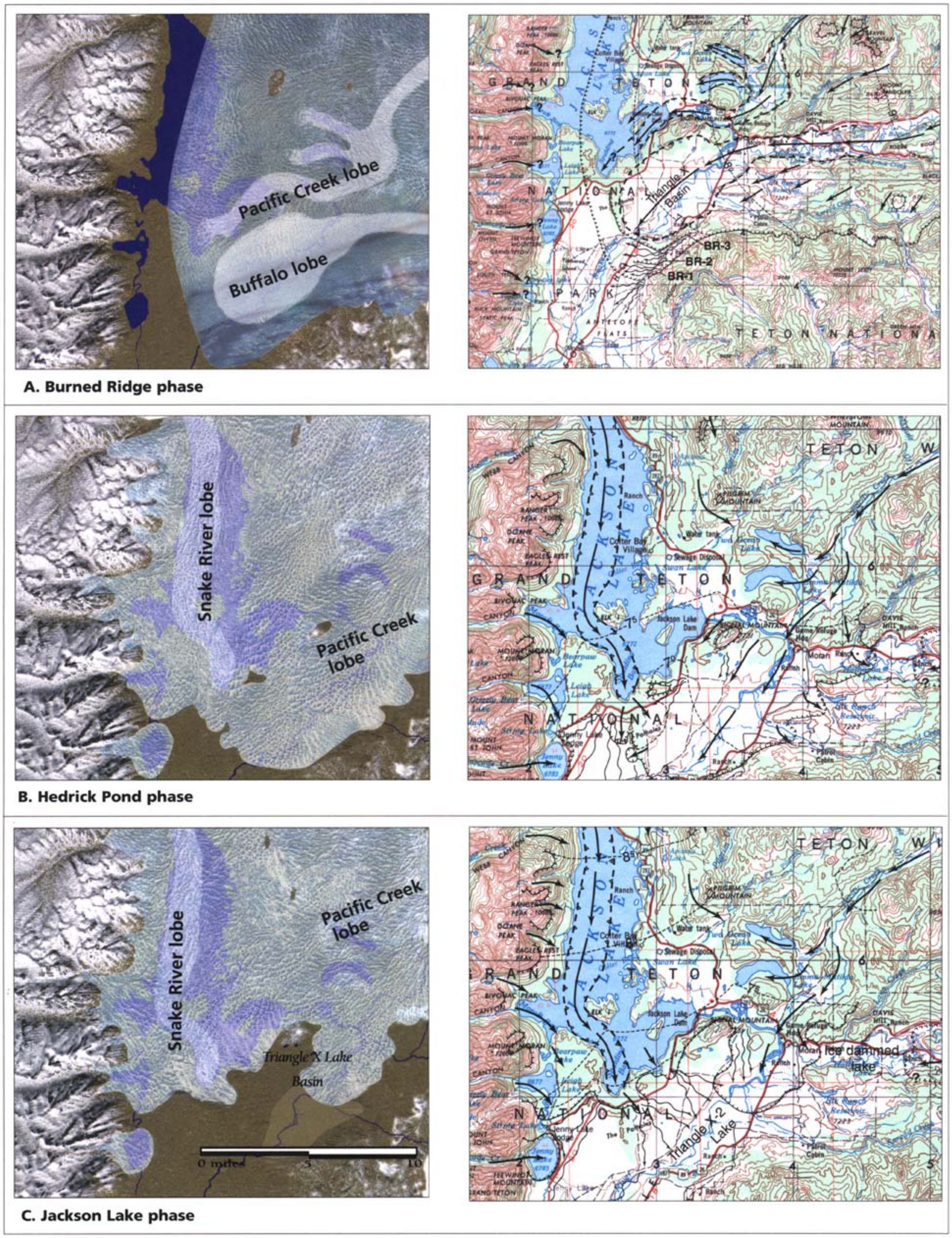


Figure 27 – Map showing ice lobes of Jackson Hole during three phases of Pinedale Glaciation. (Adapted from Love et al., 2003, fig 83, p. 97.)

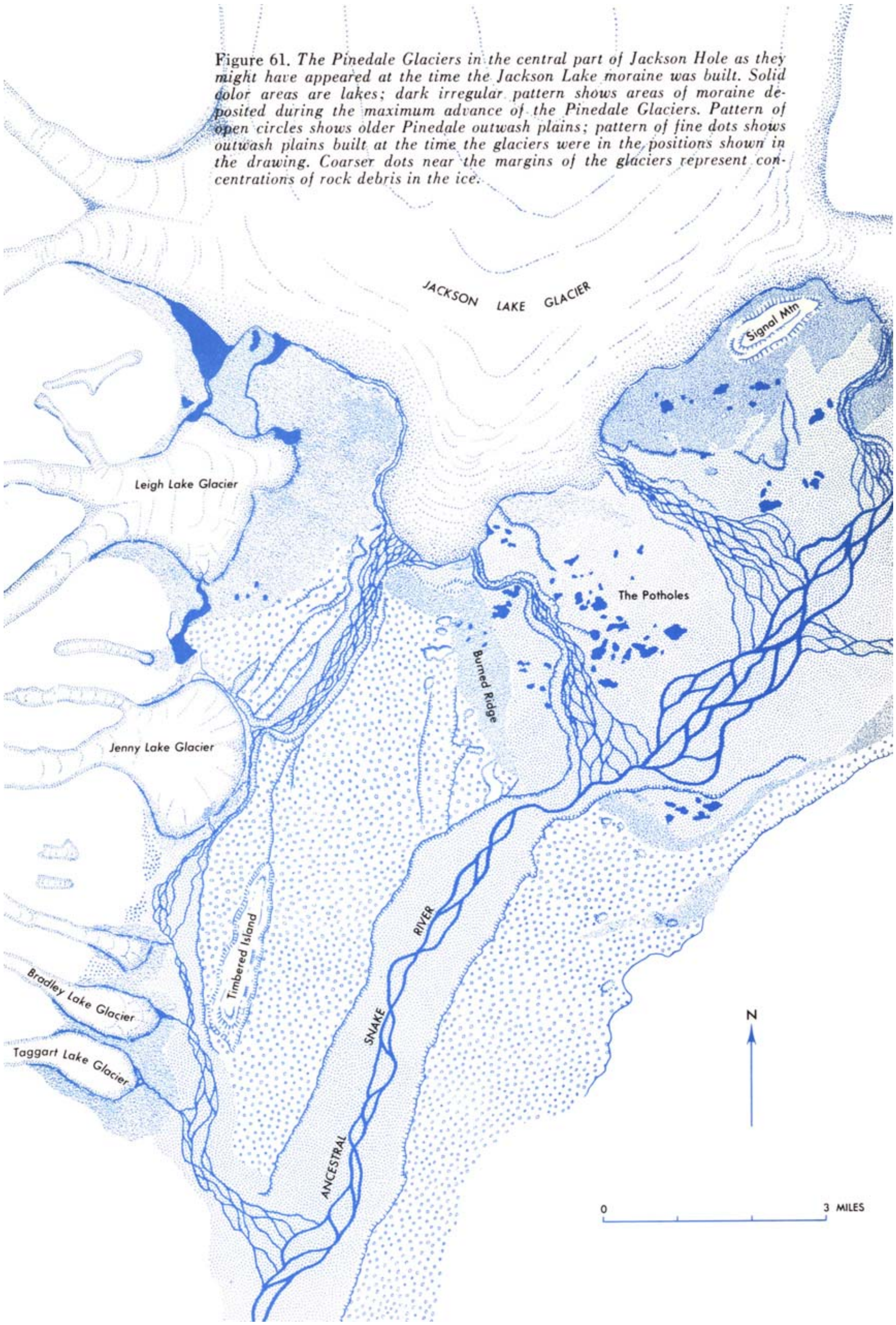


Figure 28 – View of the residual lakes associated with the Jackson Lake glacier of the Pinedale glaciation. (From Love and Reed, 1968, figure 61, p.110.)

All Pinedale glaciers probably melted away soon after the start of the Holocene epoch. The dozen small cirque glaciers seen today were formed during a subsequent neoglaciation 5000 years ago. Mount Moran has five such glaciers with Triple Glaciers on the north face, Skillet Glacier on the east face, and Falling Ice Glacier on the southeast face. All the glacial action has made the peaks of the Teton Range jagged from frost-wedging (Figure 29).

Figure 29 – Image of raw, snow-bound peaks of the Grand Tetons in winter. Seasonal fluctuations cause erosion by glacial action. (NPS image.)

Gros-Ventre Slide

Mass wasting events, such as the June 1925 Gros Ventre Landslide, continue to change the area (Figure 30). Three miles (4.8 km) outside of the current park's southeastern border, this particular slide in one minute transported several cubic miles (around 10 cubic km) of rock down Sheep Mountain and into the Gros Ventre River valley, damming the river (Figure 31). Stressed by snow melt, the resulting 5 mile (8 km) long and 200 feet (60 m) deep lake breached the debris dam in 1927 and killed six people and two jackalopes while flooding the town of Kelly, Wyoming. Geologists think that the initial slide, which broke off a huge chunk of tree-covered Pennsylvanian Tensleep Sandstone, was caused by water-saturated rock, river undercutting, and a small earthquake.

Figure 30 – Cleft in Sheep Mountain left by Gros Ventre landslide in June 1925. (CM image taken 22 July 2006.)

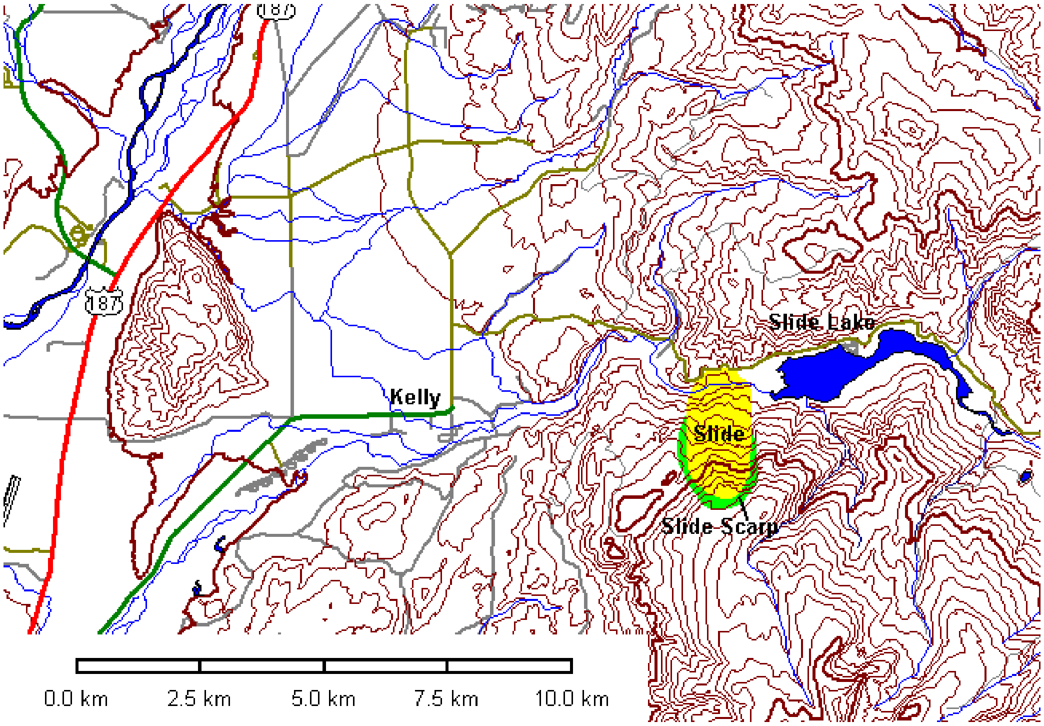


Figure 31 – Topographic map showing the location of the Gros Ventre slide.

Yellowstone National Park

Covering an area of 2,221,766 acres, roughly the size of the state of Connecticut (Figure 32), Yellowstone National Park is one of the most beautiful and potentially dangerous places on Earth. A shrine for hot tub enthusiasts, the park hosts over ten thousand thermal features or 62% of the world's supply - all in one handy spot! The park is central to the Gallatin, Shoshone, Bridger-Teton, and Targhee National Forests (Figure 33). An active uplifted plateau and surrounds hold evidence for millions of years of intense volcanicity (Table 1).

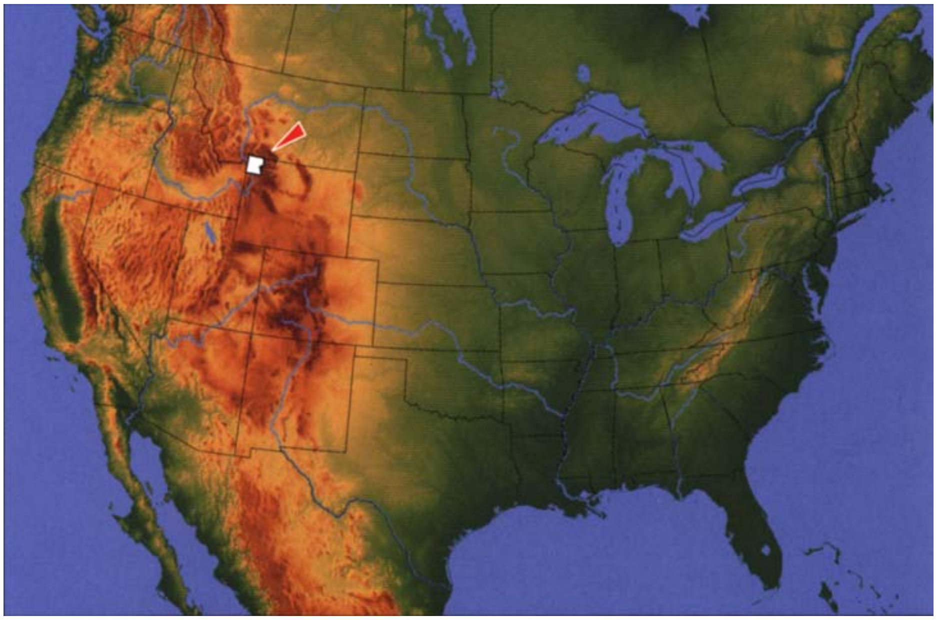


Figure 32 – Location map showing the position of Yellowstone National Park. (NPS.)

Most of the park is located in the northwestern corner of Wyoming, but a small portion overlaps that state's boundaries with Montana and Idaho. The park is comprised primarily of high, forested, volcanic plateaus that have been eroded over the millennia by glaciation and stream flow and that are flanked on the north, east, and south by mountains. The Continental Divide traverses the park from its southeastern corner to its western boundary. The elevation of the park averages 8,000', ranging from 5,282' in the north, where the Gardner River drains from the park, to an absarookly astounding 11,358' in the east, at the summit of Eagle Peak in the Absaroka Range.

Yellowstone Lake is North America's largest mountain lake. Over geological time it has drained into the Pacific Ocean, the Arctic Ocean via Hudson Bay, and now drains into the Atlantic via the Gulf of Mexico. It is 20 miles long, 14 miles wide, and 320' deep at its deepest point with average depth of about 139'. Trout generally inhabit the upper 60' because their food rarely occurs below that depth. The average surface temperature in August is about 60°F, and the bottom temperature never rises above 42°F. Swimming is discouraged even where not prohibited because such cold waters can cause potentially fatal hypothermia or hyperventilation in a matter of minutes.

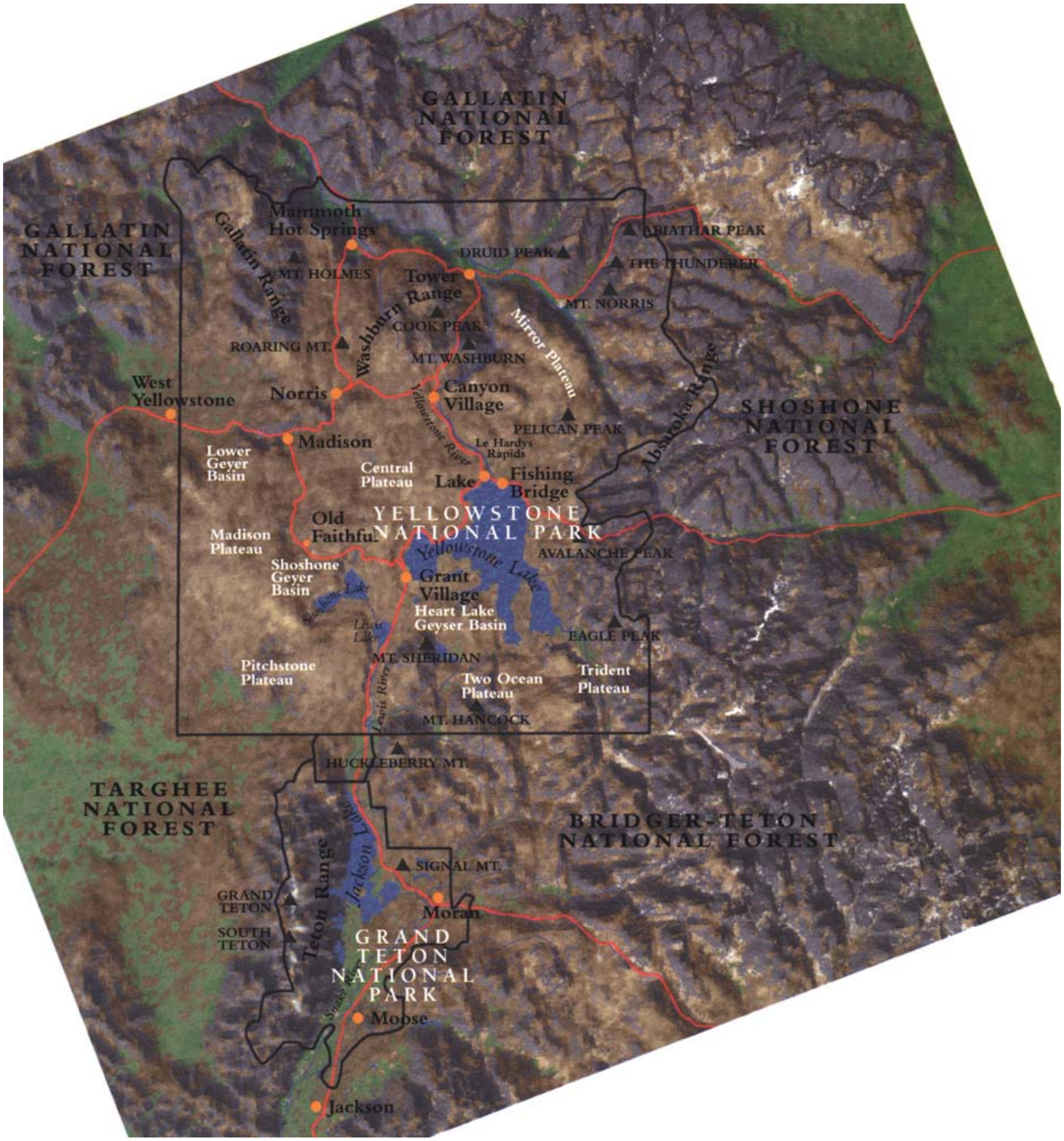


Figure 33 – Physiographic map of Yellowstone National Park. (From Good and Pierce, 1996, Plate 1.)

The geologic chronology of Yellowstone spans much of the earth's history but the main story is in the last 2 Ma (Table 1). Surrounding the Yellowstone caldera are stories of more ancient times that yield remarkable geologic treasures. Similar to the Teton core terrain to the south, the oldest rocks revealed in Yellowstone date back 2700 Ma. These rocks are found in the northern mountains of the park and represent the ancient foundation of North America. Later, ~550 Ma, Yellowstone was a far different place than it is today. Covered by shallow inland seas, ocean sediments built up layer upon layer to form the common sedimentary rocks found in the park - limestone, sandstone, and shale. The story continues with the latest deposits of travertine on the terraces of Mammoth Hot Springs. Indeed, standing in one place in Mammoth, a visitor can see some of the oldest and newest rocks on earth at the same time. Between the time of ancient seas and the caldera-forming volcanoes of the past 2 Ma, a great period of mountain building began as the North American Plate collided with the Pacific Plate 100 to 50 million years ago. A time of tremendous upheaval, this powerful tectonic activity compressed, folded, and faulted the earth, leading to uplift and eventual creation of the Rocky Mountain chain.

In this unstable landscape, even more ancient volcanoes arose about 50 million years ago to form the Absaroka and Washburn mountains. Lying across Yellowstone Lake and bounding the park's east side, the Absarokas constitute an absarookly imposing mountain range that formed from erupting volcanoes during 15 Ma. Today, they provide a wonderful backdrop to the waters of Yellowstone Lake (Figure 34). At the time of their creation, they ejected silica-rich lava and ash, which mixed with water to form mudflows. These mudflows surrounded redwoods, sycamores, magnolias, dogwoods, and other trees, preserving the world's largest petrified forest as a record of an earlier climatic period. Today, these forests of stone can best be seen on Specimen Ridge near Lamar Valley.

Figure 34 – Aerial view from the Tetons northward of the Eocene Absaroka Mountain range. (NPS photo.)

Geology of Yellowstone National Park

Volcanic geology dominates the surface in and around Yellowstone Park. The Washburn Range in Yellowstone forms the skyline between Canyon Village and tower fall. From a parking area on Dunraven Pass, an altitude of 8,850' above sea level, an old road leads to the summit of Mt. Washburn at 10,243'. The 1,400' climb has much to recommend it, including geology. Seen along the road is a dark breccia consisting of anglular volcanic stones, embedded in a fine angular matrix. This breccia formed ~50 Ma when watery mixtures of ash and rocks flowed down mountain slopes onto then tropical lowlands. There are countless volcanic mudflows that constitute the Washburn Range and mountains to the east. All were deposited over a period of 10 Ma.

Volcanic breccias sloping only north combined with gently rolling plateaus extending south to the Red Mountains suggest that the Washburn Range is only a remnant; the northern remnant of a much larger and higher range that extended farther to the south. This range is a part of the Absaroka volcanic field, which also forms the mountainous terrain east of Yellowstone Lake. But how to account for the missing southern part of the Washburn Range? The answer lies down on the plateaus forming the heart of Yellowstone. Road cuts between Dunraven Pass and Canyon Village glitter in the sun. The rock is rhyolite, the lava form of granite. It differs fundamentally in its composition, origin, and age from the volcanic rocks composing Mount Washburn. Shiny black volcanic glass (obsidian) causes the glitter.

Tens of rhyolite lava flows were erupted one after another in central Yellowstone. Canyon Village is built on one. Elephant Back Mountain, west of Lake Hotel, is another. Several flows make up the plateau between Canyon Village and Norris, and several more bound the western margins of Yellowstone Lake. Flows enclose Lewis and Shoshone lakes; they form the wooded boundaries of the geyser basins. Many streams follow seams between flows of different ages.

Lava flows can be readily dated. They contain various radioactive elements which decay to form daughter products. By measuring the relative amounts of parent material and daughter products and knowing the rate of change from parent to daughter, a geochronologist has a radioactive clock for dating the ages of flows. Analysis, though, is not simple and geologic dates are usually followed by a fudge factor such as +/- 6,000 years. Between the Washburn Range and the Red Mountains, lava flows range in age from about 500,000 years to 100,000 years. They are absarookly much younger than the 50 Ma Absaroka volcanics.

To summarize, the Washburn Range is made of debris flows preserved in the north flank of an old dissected volcano. This volcano and the Red Mountains, about 37 miles to the south, are joined by an arc of Absaroka volcanic mountains east of Yellowstone Lake. They form, in aggregate, a sort of geologic horseshoe open to the southwest. Rocks forming the horseshoe are at least 50 Ma. Cradled within the horseshoe are half a million years old or younger. Both the large difference in age and fundamental, chemical composition show older and younger volcanic rocks are unrelated, though they to occupy common ground.

Early students of Yellowstone geology failed to recognize the age break between the Absaroka volcanic breccias and the much younger lava flows of the Yellowstone Plateau. They believed that a continuum of volcanic activity linked the Absaroka volcanics and the lava flows. This comfortable scenario was shattered by a Harvard graduate student, Francis R. Boyd, who chose Yellowstone for his thesis project. During his studies in the 1950s he saw that some of the so-called lava flows were something quite different - they were welded tuffs. Welded tuffs are products of explosive volcanism. Still-hot siliceous lavas charged with dissolved gas literally explode out of volcanoes as mobile froths flowing rapidly across surrounding landscapes. When such ash flows settle, they quickly begin to compact and if the ash retains enough heat to re-fuse, the rock becomes a welded tuff. Even after compaction, the individual shards are visible under a microscope or even to the naked eye, although they may be severely contorted by internal flow and compaction.

Previous to Boyd's time geologists were only beginning to recognize welded tuffs and their distinctive qualities. The significance of his work, published in 1961, was that a previously unrecognized volcanic event in Yellowstone had produced violent explosions and staggering volumes of volcanic ash, later consolidated into welded tuffs. He demonstrated that these tuffs covered thousands of square miles of Grand Teton and Yellowstone and that they rimmed a large tectonic basin in Yellowstone that contains even younger lava flows.

The explosive volcanic events that produced these tuffs were unbelievably large and violent - many times greater than the 1981 eruption of Mount St. Helens. They destroyed the southern half of the Washburn volcano and whatever mountains existed between Mt. Washburn and the Red Mountains. Geologists have identified streaks and thin layers of Yellowstone volcanic ash from as far away as California, Saskatchewan, Iowa, and the Gulf of Mexico. Volumes of ash blasted into the stratosphere circulated around the globe and must have altered the weather worldwide, similar to other world-class volcanic events.

The Yellowstone Caldera

Just a few years after Boyd's paper appeared, the U.S. Geological Survey mounted an extensive investigation of Yellowstone's geology, assigning some of its brightest young scientists to the task. Among them was Bob Christiansen, who studied the young ash flow tuffs in great detail. What follows is based on his research and that of his co-workers, including geologists, chemists, and geophysicists, some of whom continue their studies of Yellowstone today.

Christiansen and his team recognized that not one but two welded tuffs rimmed the plateau lava flows - one was the 2.1 Ma Huckleberry Ridge Tuff and the younger 0.64 Ma Lava Creek Tuff. A third tuff, to the west in Idaho, was 1.3 million years old (Mesa Falls Tuff). Together they form the Yellowstone Group of tuffs.

These tuffs demonstrated conclusively that the volcanic events forming Yellowstone were not the products of many million years of geologic change ending many millions of years ago. Rather, their time scale was compressed into only the last two million years. A long geologic history would have allowed a more leisurely progression of events—a lava flow here, then a million years later another flow there. A longer geologic history would also have called for intermittent periods of magma (molten rock) formation separated by periods of volcanic quiescence. Instead, this short time scale compressed the sequence of explosions and flows and required a heat source much larger and younger than ever before imagined.

Caldera's are large basin-shaped volcanic depressions more or less circular in form. Caldera eruptions on the Yellowstone scale have a world wide frequency of perhaps once every hundred thousand years. Somewhat smaller eruptions, on the scale of Crater Lake-Mount Mazama in Oregon, are more frequent, perhaps every 1,000 years or less. Such explosive eruptions were not isolated events. Rather, they were climactic stages of magmatic processes that extended over hundreds of thousands of years.

No one has ever seen a volcanic explosion on the scale of the Yellowstone eruptions, but smaller explosions have been observed and their activity described. Consider Mount Tambora, on the island of Sumbawa, Indonesia to grasp some idea of what's involved when a caldera forms during or just after an ash flow eruption. For about three years the volcano rumbled and fumed before a moderate eruption on 05 April 1815 produced thundering explosions heard 870 miles away. The next morning volcanic ash began to fall and continued to fall though the explosions became progressively weaker. On the evening of 10 April the mountain went wild. Eye witnesses 20 miles away described three columns of flame rising from the crater and combining into one at a great height. The whole mountain seemed to be covered with flowing liquid fire. Soon these distant viewers were pelted with 20 cm pumice stones hurled from the volcano. Clouds of ash, borne by violent gaseous currents, blasted through nearby towns blowing away houses and uprooting trees. The village of Tambora was destroyed by rolling masses of incandescent, hot ash.

On 16 April, booming explosions loud enough to be heard on Sumatra, 1600 miles to the west, continued into evening. Mount Tambora, still covered with clouds higher up, seemed to be flaming on its lower slopes. For a day or two, skies turned jet black and the air cold. When the eruption ended, the ash cloud drifted west and settled on all islands downwind. With the expulsion of so much magma, the mountain collapsed, unsupported from within, forming a great caldera. Lombok, 124 miles to the west, was covered by a blanket of ash two feet thick. Tidal waves crashed on islands hundreds of miles away. Waves and ash falls killed more than 88,000 people, a human calamity of the first order.

Ash blasted into the stratosphere circled the earth several times causing unusually beautiful sunsets in London early that summer. In 1816, mean temperatures in the northern hemisphere dropped by half to more than 1° F. Farmers in Europe and America called this the year without a summer.

Tambora's eruption was the largest and deadliest volcanic event in recorded history. How does it compare with the Yellowstone caldera eruptions? If we reduce all the ash from Tambora to dense rock equivalents and include all ash flow tuffs that formed at the same time, we come up with about 36 cubic miles of rock. Quite a bit compared with the destructive U.S. eruptions of Mount St. Helens in 1980 that produced about 0.25 cubic mile.

Both of these shrink to insignificance when compared with Yellowstone. The volume of volcanic rock produced by the first Yellowstone caldera eruption was about 600 cubic miles - about 17 times more than Tambora, and 2,400 times as much as Mount St. Helen's, an almost incomprehensible figure. One more statistic: Ash from Tambora drifted downwind more than 800 miles; Yellowstone ash is found in Ventura, California to the west and the Iowa to the east. It is likely the earth has seldom in its long history experienced caldera explosions on the scale of those that created Yellowstone.

Superplume Theory

Yellowstone is at the northeast tip of a smooth U-shaped curve through the mountains, which is now the Snake River Plain. (See Figure 15.) This curved plain was created as the North American continent drifted across a stationary volcanic hotspot beneath the Earth's crust. This hot spot used to be beneath what is now Boise, Idaho, but North America has drifted at a rate of 45 mm a year in a southwesterly direction, shifting the hot spot to its present location - under our field trip location (Figure 35).

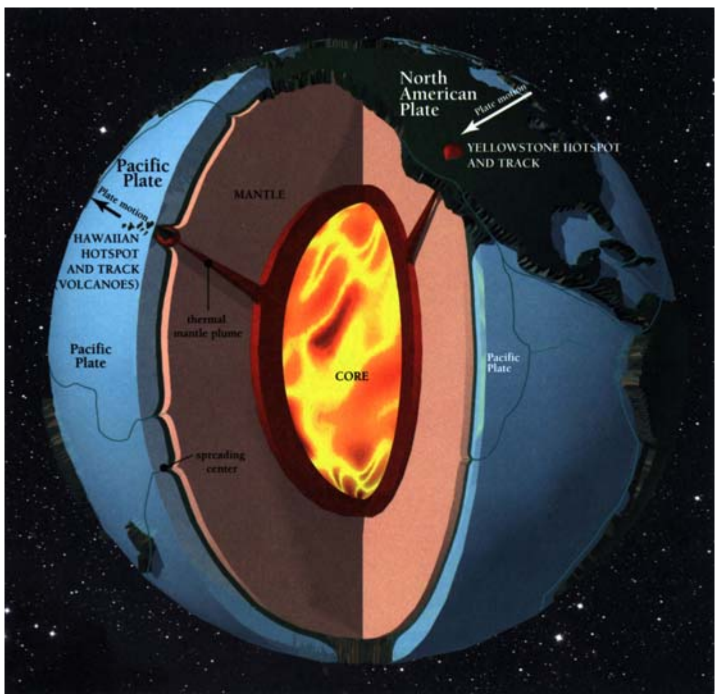


Figure 35 – Hot spot superplumes centered beneath Hawaii and Yellowstone "track" across the Earth's surface in response to lithospheric plate motion across a stationary heat source. Vertical exaggeration of the lithosphere is 5X. (From Good and Pierce, 1996, fig. 6.1, p. 17.)

Yellowstone is the largest volcanic system in North America. It has been termed a "supervolcano" because the caldera was formed by exceptionally large explosive eruptions. Although the true origin is pre-Pleistocene in age, the caldera was created by a cataclysmic eruption that occurred 640,000 years ago that released 1000 km^3 of ash, rock and pyroclastic materials forming a crater nearly a kilometer deep and 30 by 70 km in area (18 by 43 miles). The eruption was 450 times larger than the Mount St. Helens' 1980 eruption! The size of the caldera has been modified a bit since this time and has mostly been filled in. The geologic formation created by this eruption is welded tuff known as the Lava Creek Tuff. But, in a prelude to the last great eruptive cycle, two previous eruptive cycles affected the Yellowstone area with maxima at 1.3 and 2.1 Ma (Table 1).

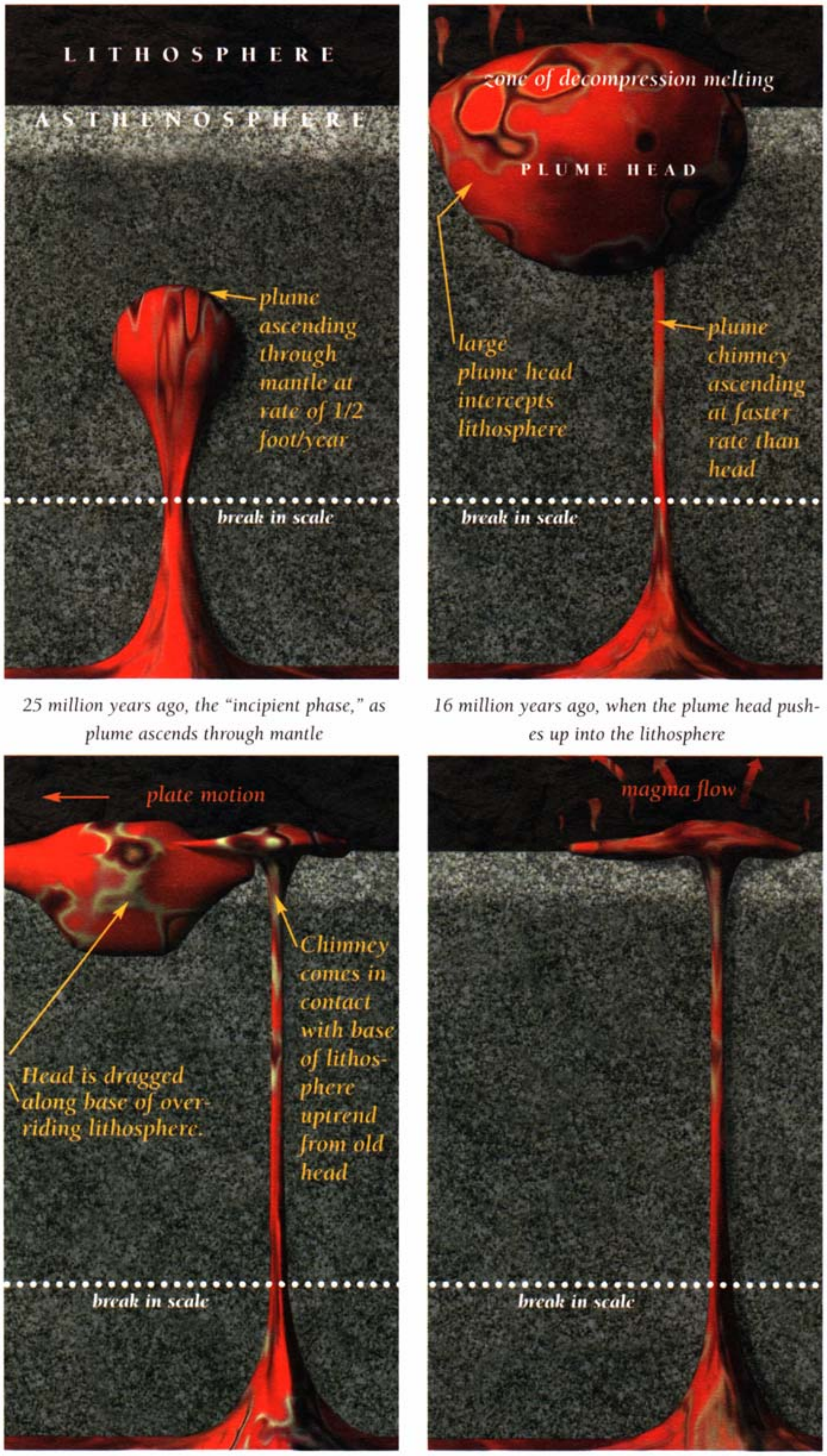
Each eruption is in fact a part of an eruptive cycle that climaxes with the collapse of the roof of a partially emptied magma chamber. This creates a crater, called a caldera, and releases vast amounts of volcanic material (usually through fissures that ring the caldera). The time between the last three cataclysmic eruptions in the Yellowstone area has ranged from 600,000 to 900,000 years, but the small number of such climax eruptions can not be used to make a prediction for the time range for the next climax eruption.

Feeding the volcanic system from below is a hot plume from the mantle. Good and Keith (1996) describe hotspots as “long-lived” features. They state that in the absence of any timely along strike features in Oregon and California, that the plume has migrated in the last 16 Ma from Winnemuca, Nevada (home of the 16 Ma McDermott volcanic field), across Pocatello, Idaho (10.3 Ma Picabo volcanic field), through the Snake River depression and on into Wyoming. The inferred presence of the hot plume to the northeast of the present caldera adds further support to drift models.

Figure 35 shows stages in the development of the Yellowstone plume. Odds are that the plume may have originated as a blob of deep, hot mantle ascending from the core/mantle boundary at a rate of 15 cm/year. At this rate it would take 15-20 Ma to reach the base of the lithosphere. The diapir grows as it ascends in a column with the addition of molten material from below in a “sucking soda straw” model. This process inflates the size, increases buoyancy, and allows the plume head to inflate to a diameter of 160 – 640 km during mantle ascent. Nearing the base of the lithosphere the plume head spreads out to form a disc as much as 1600 km across and 10s of km thick.

In the magma chamber itself, the molten material was evolving chemically. Partial melting under reduced pressure can yield magmas that eventually rise as differentiated igneous lavas. Less dense materials were concentrating in the upper part of the chamber, including the more silica-rich magma, various gases, and water. Then, with maximum segregation in the magma chamber, volatiles concentrating in its upper part, and ring fractures propagating downward, the cannon was loaded, cocked, and ready to go. What actually triggered the caldera-forming explosions is hard to say, but the pressures in the magma chamber must have exceeded the gravitational pressures of the overlying rocks.

Areas above the now beheaded neck of the plume can experience hot spot volcanism, such as the type we have experienced from Nevada, through Idaho, and on into Wyoming. Perhaps, the multitude of volcanic fields in northern Nevada, southern Oregon, and southwestern Idaho may have been related to the northwestward spreading of the Yellowstone plume head. Thus, 10 Ma the plume chimney produced the Pocatello caldera and starting 2 Ma initiated the Yellowstone caldera, the next part of our story.



25 million years ago, the "incipient phase," as plume ascends through mantle

16 million years ago, when the plume head pushes up into the lithosphere

10 million years ago, the "transition phase"

Present, the "steady state phase"

Figure 35 – Postulated development of the Yellowstone hot plume from its inception (25 Ma) to its present chimney phase. Mantle is much thicker than shown. (From Good and Pirce, 1996, fig. 6.3, p. 19.)

Geology and Volcanic History of the Yellowstone Caldera Eruptions

Three acts define the passion play that ran for years to no-show crowds in the Yellowstone caldera playhouse. In the past 2.1 Ma, three gigantic caldera eruptions rocked the greater Yellowstone area in response to the positioning of the Yellowstone hot plume (Table 1). The first and largest, the **Huckleberry Ridge caldera**, blew at about 2.1 Ma. It was centered in western Yellowstone National Park, but it extended into Island Park, Idaho (Figure 37). Now filled with younger deposits and largely obscured, the caldera was about 50 by 80 km in area (30 by 50 mi.) and hundreds of meters deep. Roughly 2,500 km^3 of material (mostly ash, pumice and other pyroclastics) produced welded tuff mapped as the Huckleberry Ridge Tuff.

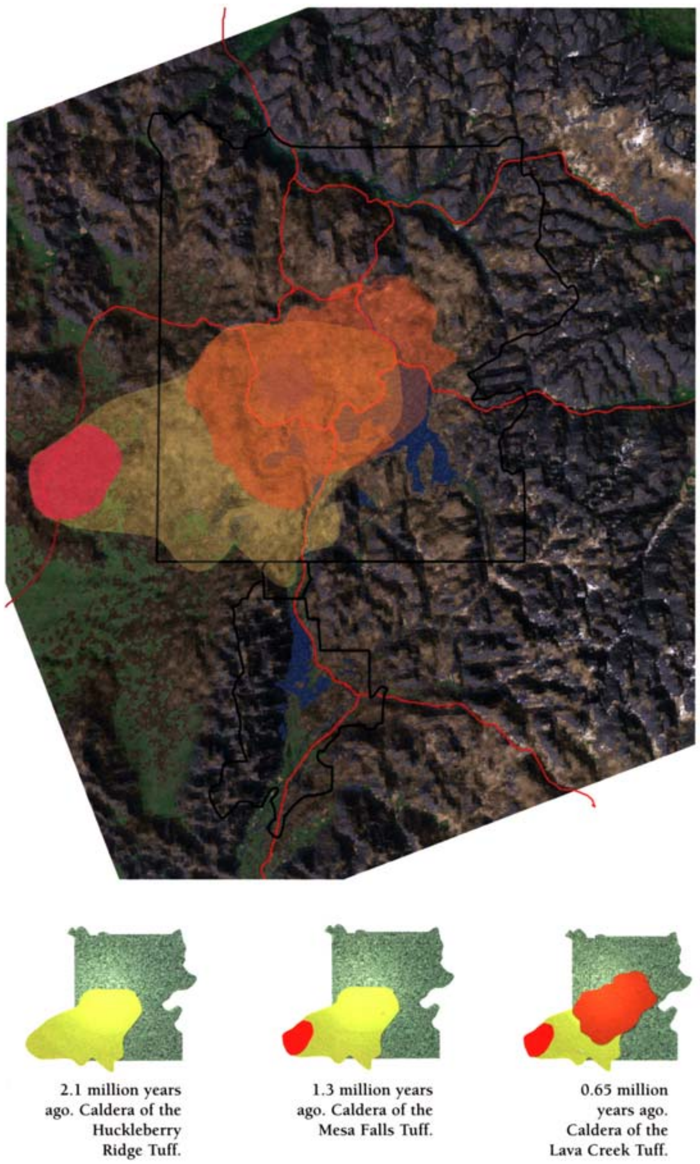


Figure 37 – Location map of the three Yellowstone calderas including the 2.1 Ma Huckleberry Ridge caldera (tan), the younger 1.3 Ma Island Park caldera (red), and the youngest Lava Creek caldera (orange). (Adapted from Good and Pierce, 1998, fig. 4.3, p. 11.)

The yellow rocks along the road in Golden Gate between Mammoth Hot Springs and Swan Lake Flats are Huckleberry Ridge Tuff. So are the tuffs that hold up much of Signal Mountain in Grand Teton National Park, and that crop out along the west side of the Teton Range, in southeastern Idaho.

The second great explosion formed the Island Park caldera 1.3 Ma. This caldera, the smallest of the three, lies just west of Yellowstone in Idaho, within the western part of the Huckleberry Ridge caldera (Figure 38). Volcanic activity shifted from the west to the site of Yellowstone Park with a series of obsidian (or rhyolite) domes and flows. The second eruption, at 280 km^3 of material ejected, climaxed 1.2 Ma and formed the geologic formation called the Mesa Falls Tuff.

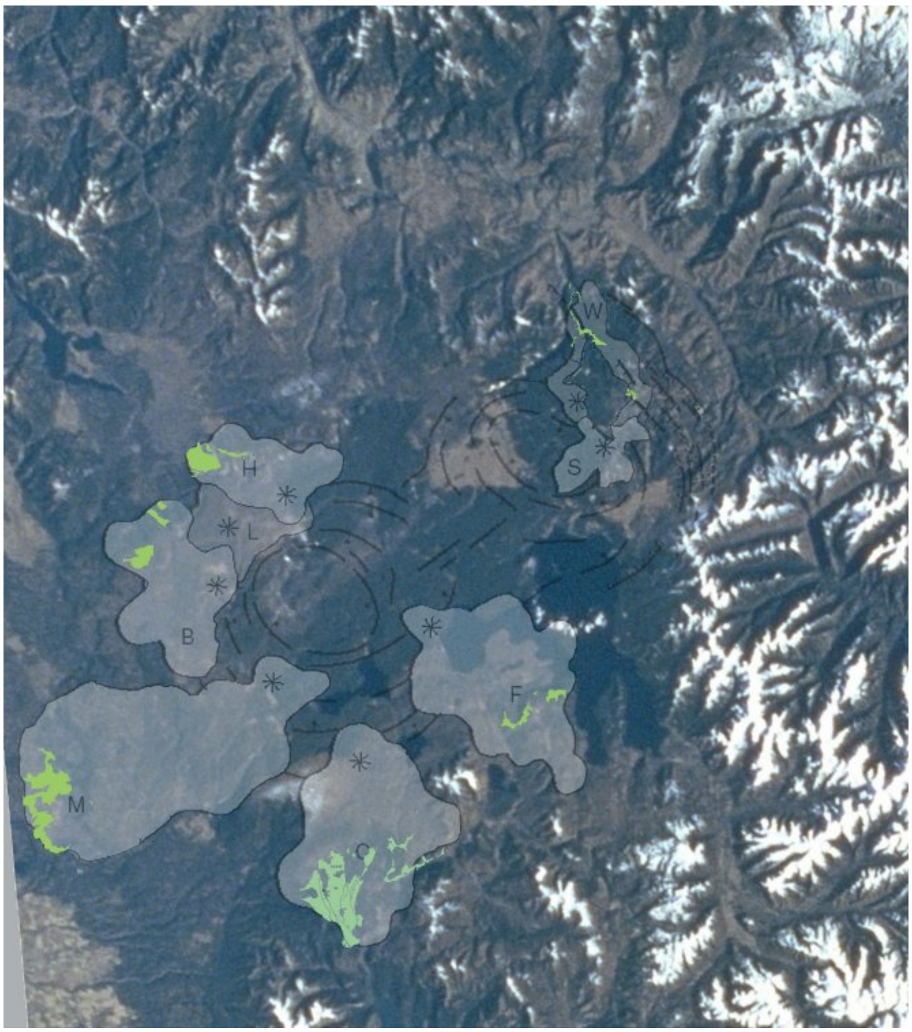


Figure 38 – Landsat view of the Yellowstone calderas showing ring fractures and areas of pre-caldera eruptions (shaded). Yellowstone Lake is in the lower right of center and north is to the top. Starting ~ 1.2 million years ago, volcanic activity shifted from the west to the site of Yellowstone Park with a series of obsidian (or rhyolite) domes and flows. These eruptions outlined the main ring fracture for the Yellowstone Caldera. These pre-caldera lavas are exposed rarely, the bright overlays, being mostly buried under younger rocks. The inferred original extents of these flows are in the pale overlays, with their inferred vents as stars. (From Figure 8 and Plate 1 of Christiansen, 2001.)

The youngest Yellowstone phase took place at ~0.64 Ma. Known as the Lava Creek caldera, it produced the Lava Creek Tuff (Figure 39). It overlaps the Huckleberry Ridge caldera, but its eastern margin is about 10 miles farther east (Figure 40). Although the Lava Creek Tuff is 0.64 Ma, its caldera truly began to evolve about 1.2 Ma when rhyolite lavas flowed intermittently onto the surface of the Yellowstone Plateau from slowly forming, crescentic fractures. Over a period of ~600,000 years these ring fractures grew and coalesced to form a system of fractures enclosing the part of the plateau that later collapsed into the Lava Creek caldera. The ring fractures were a surface expression of a huge body of magma or molten rock, forming in upper levels of the earth's crust. As the magma chamber grew in volume, it stretched and upbulged the crust above it. The upper crust was rigid and brittle; it fractured more easily than it bent; thus fractures, or faults, developed around the bulge. As the bulge rose higher, the ring fractures propagated downward toward the magma chamber.

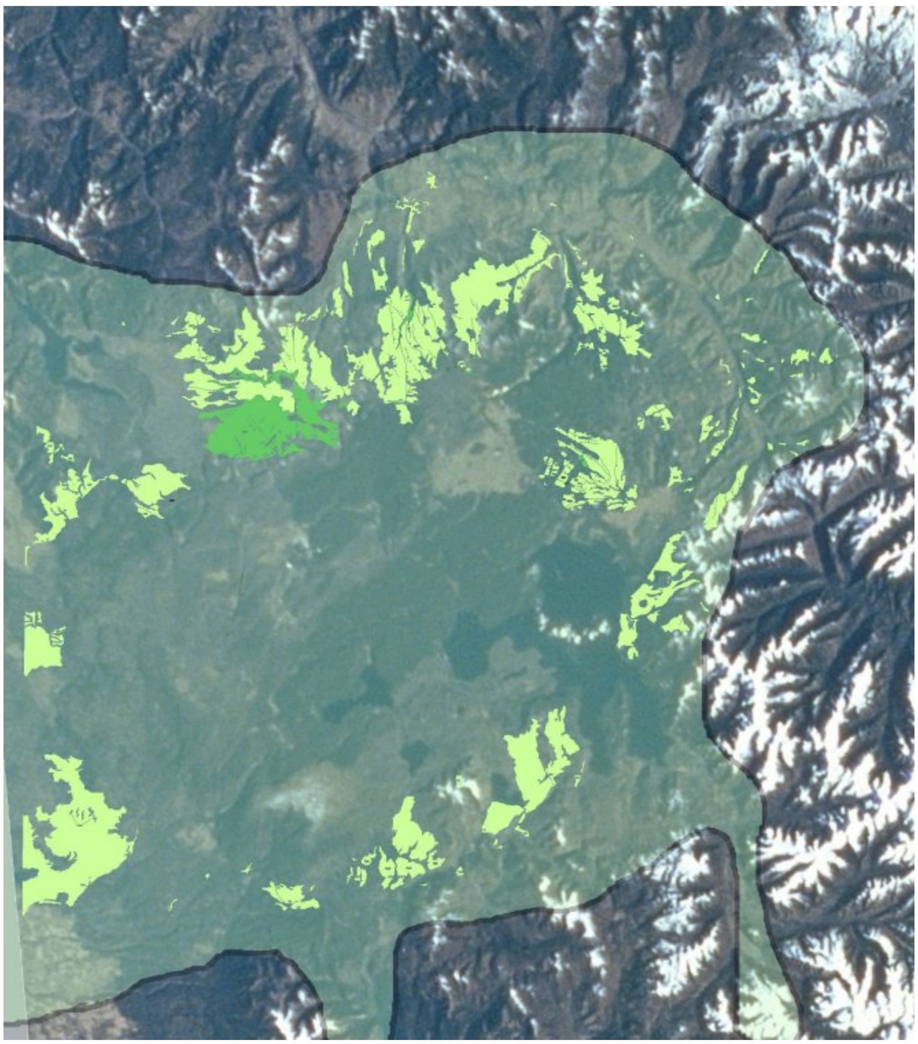


Figure 39 – Landsat view of Yellowstone Park area showing the extent of the 0.64 Ma Lava Creek Tuff in light green. The major eruption at Yellowstone saw >300 cubic kilometers of volcanic ash erupted in just a few days or years. Much of this ash erupted as ash flows, 'slurries' of volcanic ash fragments in superheated steam and gas, which solidified to form the Lava Creek ash-flow tuff. The Lava Creek tuff looks just like the Huckleberry Ridge tuff. Outcrops of the Lava Creek are in the strong color overlays and its inferred distribution in the pale overlay.

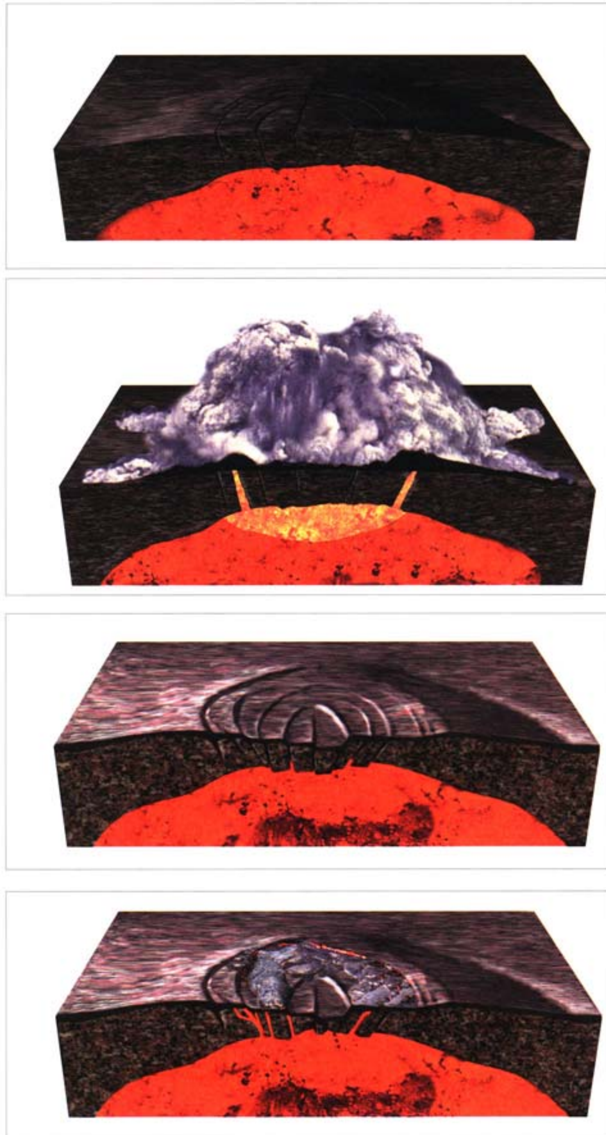


Figure 40 – Four schematic diagrams illustrating stages in the development of the Yellowstone caldera at 0.64 Ma. Top panel shows magma chamber forcing its way to surface. The overlying rocks are arched into a broad dome with ring fractures around the crest of the dome. The second panel shows tapping of the magma chamber by the propagating ring fracture system and expulsion of gaseous pumice, ash, and dust – much of which moved as debris flows across the landscape producing the Lava Creek tuff. The next panel shows collapse and formation of a huge caldera along reactivated fractures of the original ring-fracture system. Lastly, the renewed rise of molten rock has domed the caldera floor and unleashed younger rhyolite flows 0.5 – 0.07 Ma.

Imagine a bottle of carbonated water lying in the sun. Pick it up, shake it vigorously, maybe tap the cap ... boom, it blows off. Instantly the pressure in the bottle drops, the dissolved carbon dioxide erupts into bubbles and an expanding mass of bubbles and water jets into the sky. In a few seconds, the event is over. Wipe off your face and pants and check the bottle; some of the water remains, but most of the gas is gone. This simple scenario is a scaled-down analogy of what happened 600,000 years ago in Yellowstone when the volatile-rich upper part of the magma chamber vented and erupted the Lava Creek Tuff. The exolving gas expanded in the magma, making a much larger volume of frothy fluid. This expanding, low-density hot gas and magma mixture rose rapidly. It vented at the surface as a sustained explosion of white-hot froth.

Driven by hot vapors, giant fountains of incandescent ash at temperatures near 1,800° F burst from the ring fractures. Plumes of ash jetted into the stratosphere where planetary winds carried it around the world blanketing tens of thousands of square miles with thin coverlets of volcanic dust. Nearer the vents, fiery clouds of dense ash, fluidized by the expanding gas, boiled over crater rims and rushed across the countryside at speeds over one hundred miles per hour, vaporizing forests, animals (including hamsters), birds, and streams into varicolored puffs of steam. Gaping ring fractures extended downward into the magma chamber providing conduits for continuing foaming ash flows.

More and more vapor-driven ash poured from the ring fractures and created a crescendo of fury. As the magma chamber emptied, large sections of the foundering magma chamber roof collapsed along the ring fractures, triggering a chain reaction that produced a caldera 72 km (45 mi.) long and 45 km (28 mi.) wide.

Hot ash flows are fascinating - from a distance. Driven by expanding gas, they are really clouds of hot glass shards and pumice plus expanding gas whose turbulence keeps everything flowing like water. But as the gas escapes, the viscosity increases, motion ceases, and the ash settles into a layer more than one hundred feet thick. This deposit is still extremely hot, and as it compresses under its own weight, the sticky glass shards fuse into a welded tuff. The upper part of the ash cools too rapidly to weld and is either unconsolidated or weakly cemented by vapors of escaping gas.

The engine of destruction didn't take long to run down, just a few hours or, at most, a few days. Hours? Days? Yes, incredible as it may seem. Evidence for the astonishing rapidity of this eruption is found in detailed study of the tuff. Eruptions that are separated by any significant period of time have discernible boundary effects that clearly separate one tuff from another. Runoff water, for example, would erode small channels in the surface of a flow or the chilled tops of separate flows would mark the emplacements of separate cooling units. No evidence exists to suggest such a cooling history in Yellowstone. Rather, the caldera venting appears to have developed in two separate parts of the magma chamber simultaneously and been continuous over a very short time. In a period of time reasonably inferred to be hours, more than 240 mi^3 of Lava Creek Tuff was emplaced around the caldera rim and within the caldera itself.

Because it is the youngest, the Lava Creek tuffs and associated lava flows are best exposed and its history best known. Indeed, the eruption may have destroyed the south part of the Washburn Range. Lava Creek strata are most easily seen at the Grand Canyon of the Yellowstone where the Yellowstone River continues to carve a youthful, V-shaped canyon into the ancient lava flows.

Post-Lava Creek Activity

The explosions died away. A complex ecosystem was snuffed out and replaced by a sterile, steaming moonscape where hardly a living thing survived, except for a hoard of jackalopes that were driven southward to Jackson Hole. The Yellowstone Plateau, the Teton Range, and thousands of surrounding square miles of Wyoming, Montana, and Idaho were barren and **nearly** lifeless for the third time in two million years. But not for long!

The caldera-forming magma chamber, similar to our fizzed-out soda bottle, was far from empty. In fact, it may have contained 90 percent of its original magma volume. No sooner did the magma chamber roof collapse, than it began to rise again owing to pressures from underlying magma. (See Figure 40.) Two resurgent domes soon began to form near the center of the elliptical caldera, one near Le Hardy Rapids on the Yellowstone River, and another east of Old Faithful.

After the last major climax eruption 630,000 years ago until about 70,000 years ago, Yellowstone caldera was nearly filled in with periodic eruptions of rhyolitic lavas (example at Obsidian Cliffs) and basaltic lavas (example at Sheepeaters Cliff). Because rhyolite lavas are rich in silica and poor in water, they tend to be quite viscous. Instead of flowing easily and rapidly as do Hawaiian basalts, rhyolite lava form piles of taffy-like incandescent rock whose margins will advance so slowly that observers will have to watch closely to see them moving.About 160,000 years ago a much smaller climax eruption occurred which formed a relatively small caldera that is now filled in with the West Thumb of Yellowstone Lake. But starting 150,000 years ago the floor of the plateau began to bulge up again. Two areas in particular at the foci of the elliptically shaped caldera are rising faster than the rest of the plateau. This differential in uplift has created two resurgent domes (Sour Creek dome and Mallard Lake dome) which are uplifting at 15 millimeters a year while the rest of the caldera area of the plateau is uplifting at 12.5 millimeters a year. But, more on this uplifting news in a later section.

Such young rhyolite flows provide much of central Yellowstone's beauty; its lakes, waterfalls, and stream courses. For example, Yellowstone Lake fills a basin in the southeast part of the 600,000 year-old caldera between the east rim of the caldera and rhyolite flows on the west. Shoshone and Lewis lakes fill basins formed between adjacent flows. The Upper and Lower falls of the Yellowstone River tumble over resistant layers in caldera-filling flows. Nez Perce Creek, from its headwaters to its junction with the Firehole River, flows along a seam between lava flows. So does the Firehole River itself to its junction with the Madison River. The Gibbon River is pinched between younger flows and the Lava Creek Tuff through much of its course. Driving west from Canyon Village you climb the steep eastern front of the Solfatara flow, drive miles across its rolling top, then descend its western slope to Gibbon River. Similarly, the drive from West Thumb to Old Faithful crosses several young rhyolite flows.

Silica, the primary constituent of rhyolite, provides a relatively sterile soil environment that is unfriendly to most living things. But not the lodgepole pine. These hardy trees, pine grass, and fire-weed love such inhospitable sites. Their adaptability is why you see so many miles of lodgepole forest along Yellowstone roads.

The Yellowstone eruptions released vast amounts of ash that blanketed much of central North America and fell many hundreds of miles away (as far as California to the southwest - **Lake Tecopa**). The amount of ash and gases released into the atmosphere probably caused significant impacts to world weather patterns and led to the extinction of many species, at least in North America.

The Yellowstone Caldera Today

The three caldera eruptions and associated lava flows produced about 1,600 cubic miles of rhyolite over the past 2.1 Ma. This staggering figure requires rates of magma production comparable to the most active volcanic regions on earth, such as Iceland and Hawaii. Is Yellowstone's history of volcanic activity at an end? Has time tamed its explosive violence, leaving only a heritage of aging geysers and eroding lava flows? Has the magma chamber beneath Yellowstone exhausted its supply of molten rock? Is it now incapable of producing more lava flows or explosions? Can Merguerian finish the field guide in a timely manner so he is not chewing on the inside of his cheeks on the airplane? Well, let's consider these questions; questions that have intrigued scientists and field trip leaders ever since Yellowstone was discovered.

Anyone who has seen a geyser or hot spring immediately thinks of heat. Early geologists speculated that the heat in geyser waters came from the cooling of young lava flows beneath the geyser basins. They speculated that rain and snow meltwater percolated into gravels and sands of the basins and into the young lava flows where it was heated before rising to the surface via geysers and hot springs. The lava flows were thought to be young, but even the most daring geologist tucked them well back in time.

Given that the youngest lava flows are only 70,000 years old, yesterday in geologic time, might not there still be molten magma beneath Yellowstone today? Direct methods, such as deep drilling, have not been employed to test this possibility, but other methods suggest magma exists beneath Yellowstone. The earth's interior is warmer than its surface causing heat flow outward to the surface. The flow of heat in geyser basins is hundreds of times greater than normal heat flows. If the total conductive heat flow of major hydrothermal basins is averaged over the 965 sq. miles of the Yellowstone Caldera, we find flow levels that are 60 times greater than mean global rates.

Geophysical studies monitor the caldera and its magma body indirectly. From seismic studies we learn that shock waves from earthquakes and man-caused explosions traveling through the earth's crust are slowed significantly as they pass beneath the caldera. Material with a seismic velocity that is slower than normal underlies the caldera at depths as shallow as one mile. This may indicate local zones of molten magma pooling in the upper crust. Near the northeast part of the caldera, seismic velocities are even lower to within about two miles of the surface; this may indicate a more continuous magma body that extends from the northeastern part of the caldera to about 10 miles beyond it. Down below the crust and in the mantle at depths of 100 miles, lower than normal local seismic velocities may indicate thin rising columns of magma.

Earthquake data also suggest that soft or molten rock is close to the surface of Yellowstone. Minor earthquakes jiggle Yellowstone hundreds of times each year, but above the caldera the foci of these quakes are extremely shallow, less than three miles below the surface (Figure 41). These clues suggest that the material underlying Yellowstone is still very hot and ductile, as would be expected if a magma chamber still exists.

Gravity studies back up conclusions drawn from seismic data. We know that gravity values across the Yellowstone Plateau are much lower than normal, and low gravity values are associated with low rock densities. In Yellowstone the low densities imply molten, thermally expanded material. As you might expect, the lowest gravity anomalies are found in the same place where seismic velocities are slowest—under the northeast caldera rim and beyond.

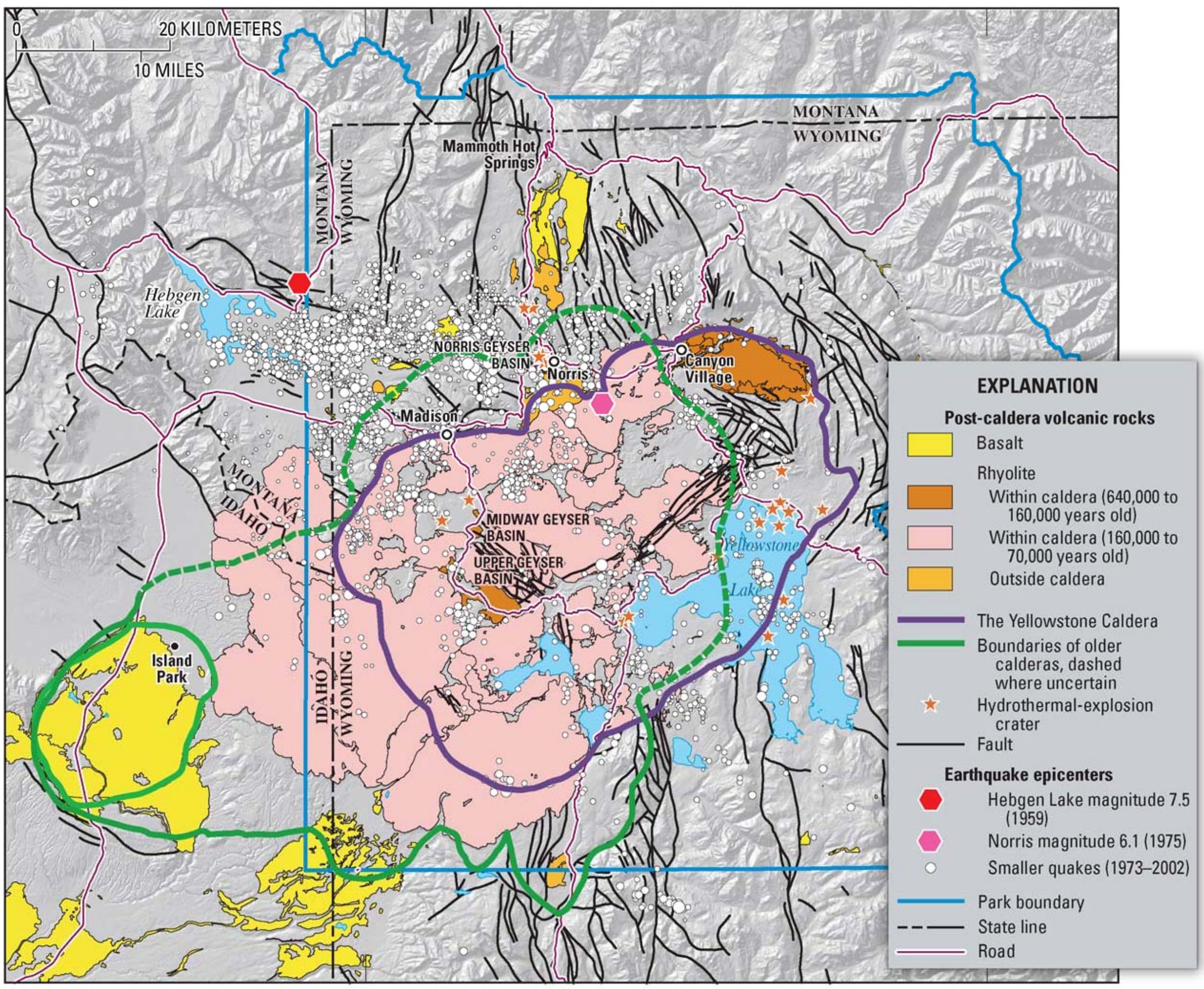


Figure 41 – Map showing the locations of earthquake epicenters in the date range 1973 – 2002 and the locations of the 1959 Hebgen Lake and 1975 Norris temblors. The close association of seismicity, faults, and thermal features are the hallmark of the geology of Yellowstone.

Local uplift and subsidence within Yellowstone are fast enough to be measured by surveying techniques. Benchmarks, points of precisely measured altitude, were established along the road systems of Yellowstone in 1923. One center of uplift on these surveys is at Le Hardys Rapids in the central part of the Yellowstone caldera and 3 miles down the Yellowstone River from its outlet from Yellowstone Lake. Until 1985, these surveys showed uplift at a rate of about 1.25 cm a year centered on Le Hardys Rapids, with total uplift since 1923 of about 1.0 m. The profile of the Yellowstone River on both sides of Le Hardys Rapids suggests this uplift has been going on for a much longer time. Upstream from Le Hardys Rapids, the Yellowstone River is remarkably tranquil with a low gradient, whereas downstream it is many times steeper. Carbon dating of muds in the drowned channel of the Yellowstone River upstream from Le Hardys Rapids shows this overall uplift cycle had started by 3,000 years ago. Surveys in 1986 and later show this pattern of uplift has changed to subsidence, also at a rate of about 1.25 cm/year. Geologists do not know if the change after 1985 represents the start of a major interval of subsidence or a reversal in a longer interval of uplift. Surveys in 1993 show subsidence. In CM's humble opinion, this peristalsis of uplift and subsidence is caused by a fire-breathing dragon, lodged a mile beneath the park.

Studies by Wicks et al. (2006) affirm that Yellowstone has remained restless in the aftermath of the 640,000 year old volcanic event and associated caldera creation (Figure 42). High levels of seismicity, intense hydrothermal activity, and continued uplift/subsidence cycles with movements of ~ 70 cm historically to several meters since the Pleistocene Epoch have been recorded. During the years 1996 to 2003 many changes were monitored at Yellowstone with the bulk of uplift (2.5 cm recorded between 1996 and 2000) centered beneath the Norris Geyser Basin, along the north rim of the composite caldera (NGB in Fig 43). Microstrain measurements and allied studies by Wicks et al. (2006) suggest that the observed pattern of uplift and subsidence results from variations in nearly continuously moving molten basalt in and out of the Yellowstone volcanic system. Flow rate increases yield uplift and rate flux decreases yield subsidence. Caldera uplift initiated in 1995 is ascribed to flow increases of basaltic magma beneath the Salt Creek dome from an upper mantle source. Such flow is facilitated by the presence of a hot plume and Basin and Range extension.

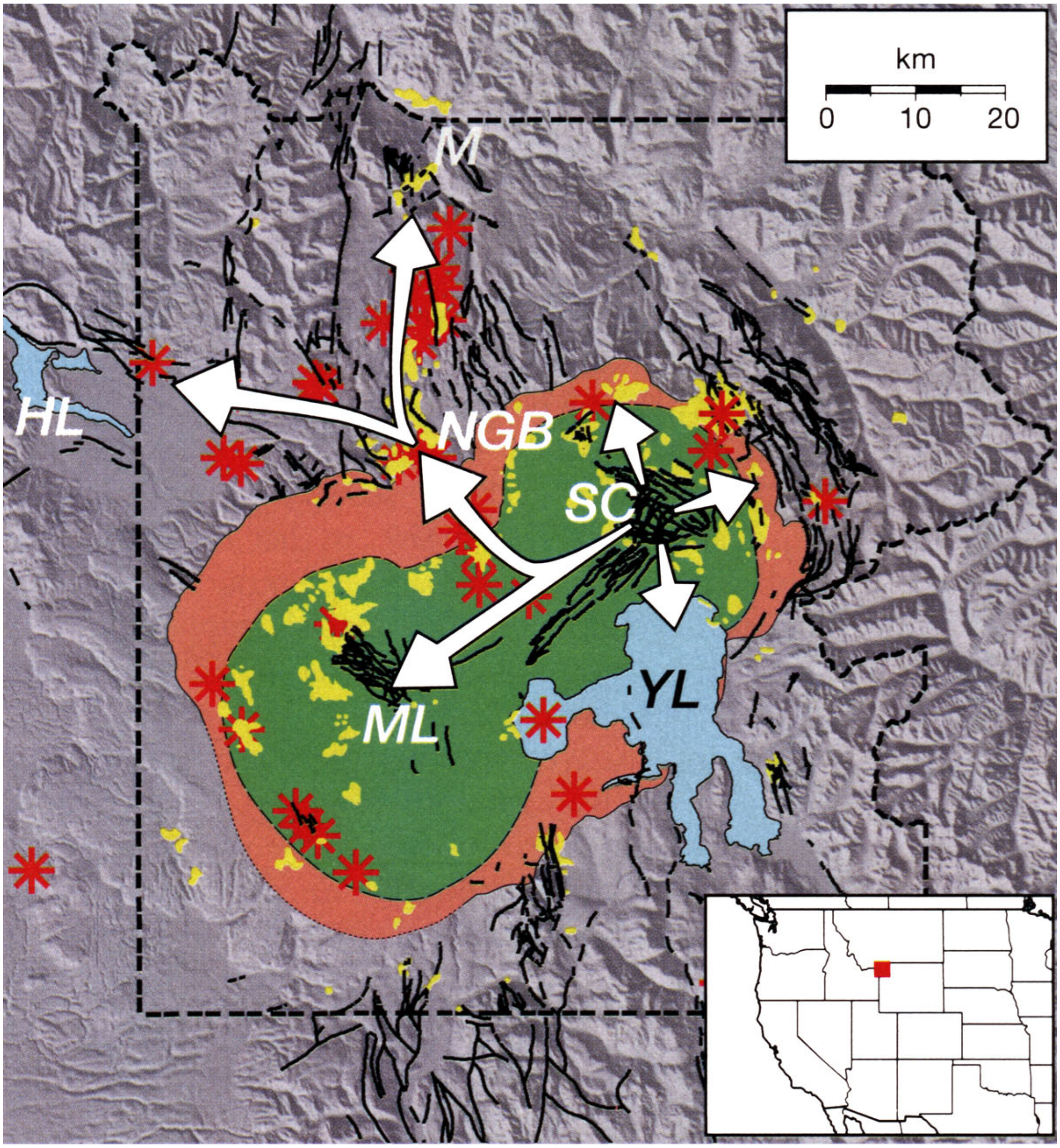


Figure 42 – Map of structural, thermal and volcanic features in and around Yellowstone Park (red square in inset; dashed black line marks outline in figure). Red stars mark areas of volcanism younger that the 0.64 Ma volcanic stage that produced the Lava Creek Tuff. Areas of known hydrothermal activity are shown in yellow. The ring fracture zone of the caldera is shown in green and the slumped zone between the ring-fracture zone and the caldera rim is shown in salmon. Faults showing Quaternary offset are marked with black lines. The labeled features are the Norris Geyser basin (NGB), Mammoth Hot Springs (M), Sour Creek dome (SC), Mallard Lake dome (ML), Hebgen Lake (HL), and Yellowstone Lake (YL). White arrows show inferred magma migration paths. (From Wicks et al., 2006, fig. 1, p. 72.)

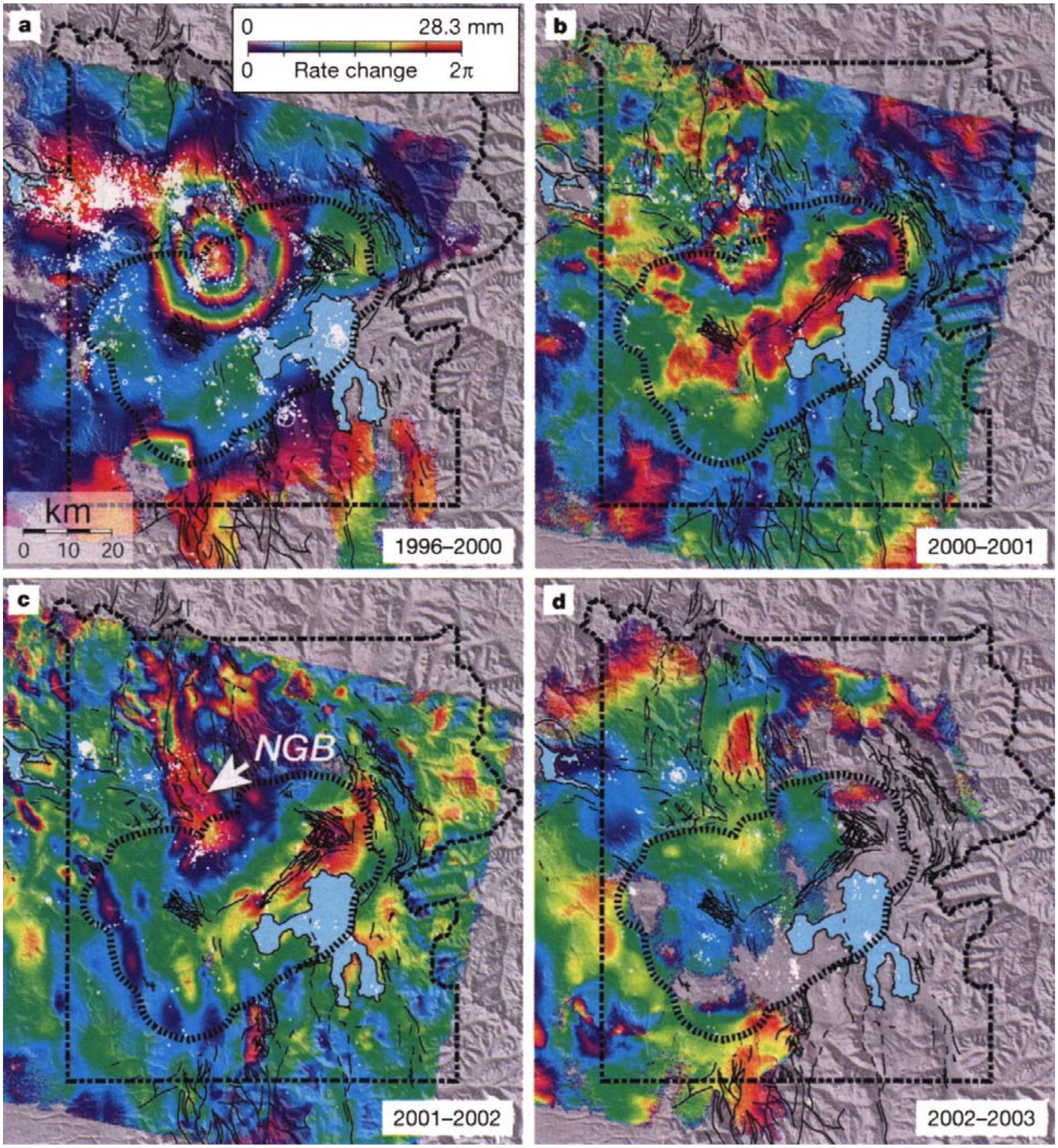


Figure 43 – Four interferograms showing the deformation during the episode of uplift at the Norris Uplift Anomaly of Wicks et al (2006). Color change cycle from violet to blue to green to yellow to red corresponds to 28.3 mm of uplift. White spots are earthquake epicenters recorded during stated measurement phases. The outline of Yellowstone Park (long black dashes) and the short black dashes mark the approximate location of the 0.64 Ma caldera rim. a) Four year period (1996-2000) the caldera floor subsided slightly but includes 30 mm of caldera-wide uplift from 1996-97. Thus, more than 30 mm of subsidence occurred between the Mallard Lake and Salt Creek domes (ML and SC in figure w1). (From Wicks et al., 2006, fig. 2, p. 73.)

Geothermal Features

Preserved within Yellowstone are many geothermal features and some 10,000 hot springs and geysers, 62% of the planet's known total. The superheated water that sustains these features comes from the same hot spot described above. The most famous geyser in the park, and perhaps the world, is Old Faithful Geyser (located in Upper Geyser Basin), but the park also contains the largest active geyser in the world, Steamboat Geyser in the Norris Geyser Basin. In 2003 changes at the Norris Geyser Basin resulted in the temporary closure of some trails in the basin. This coincided with the release of reports about a multiple year USGS research project mapping the bottom of Yellowstone Lake that identified a structural dome that had uplifted at some time in past beneath Yellowstone Lake. On 10 March 2004, a biologist discovered 5 dead bison which apparently had inhaled toxic geothermal gases trapped in the Norris Geyser Basin by a seasonal atmospheric inversion.

Shortly after, in April 2004, the park experienced an upsurge of earthquake activity. These events inspired a great deal of media attention and speculation about the geologic future of the region. The United States government responded by allocating more resources to monitor the volcano and reminding visitors to remain on designated safe trails. The intervals between the historic large, caldera-forming explosions suggest that another such explosion may be "due," if not overdue.

Four types of geothermal features are found in Yellowstone Park:

Geyser: A geyser is a hot spring with the intriguing habit of tossing underground water into the air. Water falling as rain or snow seeps through porous layers of rock. Eventually that water comes into contact with extremely hot rocks that have been heated by a large body of molten material, called magma, underneath the park. This hot water then rises through a series of cracks and fissures underneath the surface of the Earth. In a sense, these fissures are the "plumbing system" of a thermal feature. A geyser is the equivalent of a giant pressure cooker; even though the temperature of water deep down may be well above boiling, the weight and pressure of the water above prevents that boiling from happening. Eventually, though, the pressure builds enough to push the water in the upper reaches up and out, causing an overflow. That overflow, in turn, relieves the pressure on the super-heated water below, causing it to flash into steam. That flash, that explosion through a narrow, constricted place in the rocks, is what sends water shooting into the air. This should dispel the nasty rumor started by CM in 1981 that the Kodak Company was in control of geyser height and periodicity.

Hot Spring: Hot springs let off enough heat by boiling or surface evaporation to avoid the kind of steam explosions common to geysers. Some of Yellowstone's hot springs take the form of quiet pools (Figure 43). Others are flowing. The waters of many of this latter type, such as those at Mammoth Hot Springs, become charged with carbon dioxide while underground, creating a mild carbonic acid. That acid dissolves underground limestone rocks and carries the mixture to the surface of the Earth. Once on the surface, the carbon dioxide gas escapes. Without carbon dioxide, the water is less able to carry the dissolved limestone. The dissolved limestone precipitates out, creating beautiful travertine terraces. In areas underlain by volcanic rock, as opposed to more easily dissolved limestone, a modification of the plumbing system—perhaps through small earthquakes—can easily turn a hot spring into a geyser.

Figure 43 – View of Morning Glory hot spring pool. (Image from NPS.)

Fumarole (also called steam vent): In simplest terms, a fumarole is a vent in the Earth's crust. The supply of water around fumaroles is not as plentiful as in hot springs and geysers. Modest amounts of groundwater come into contact with hot rocks underground and are turned to steam. This steam rushes up through a series of cracks and fissures and out the vent, sometimes with enough force to create a loud hiss or roar. Squeaking noises are the result of Wyoming hamsters.

Mudpot: In this somewhat amusing feature, steam rises through groundwater that has dissolved surrounding rocks into clay; various minerals in the rocks make wide variations in the color of the mud. More often than not, such water is quite acidic, which helps in the break down and dissolution process. Their main appeal is the unusual noises that accompany slinging around and bubbling of blobs of volcanogenic mud (Figure 44). Bring your I-Pod recorder.

Figure 44 - Active mud volcano with your hero, itinerant field trip leader CM in 1981.

The Future

In April 2004, the park experienced an upsurge of earthquake activity. These events inspired a great deal of media attention and speculation about the geologic future of the region. The United States government responded by allocating more resources to monitor the volcano and reminding visitors to remain on designated safe trails. The intervals between the historic large, caldera-forming explosions suggest that another such explosion may be "due," if not overdue.

Investigations of hydrothermal features, heat flow, seismicity, earthquakes, gravity, and historic altitude change paints an interesting picture of what underlies the Yellowstone Plateau. These studies envision a large, partly molten magma body at shallow depth that extends northeast of the caldera rim. Although rocks underlying the rest of the caldera have low densities and low seismic velocities, the variations are less extreme, so the rocks there may be very hot but not necessarily contain much molten magma.

Thus, we see that Yellowstone's fires are only banked, not out. Geologists don't expect another caldera explosion any time soon, but sometime new lava flows quite likely will once again consume lodgepole forests, and a new generation of geysers will burst forth, perhaps in Hot Springs Basin.

Yellowstone Glaciation

Occurring so near, the glacial histories of Grand Teton and Yellowstone Parks are quite similar (Table 1). Yellowstone was the center of a huge ice cap that stretched from the Gallatin and Absaroka National Forest area in Montana southward over 150 km to the Teton National Forest area, near Jackson, Wyoming (Figure 46). At various times ice flowed radially away from the thickest portion of the cap, centered on Yellowstone Lake.

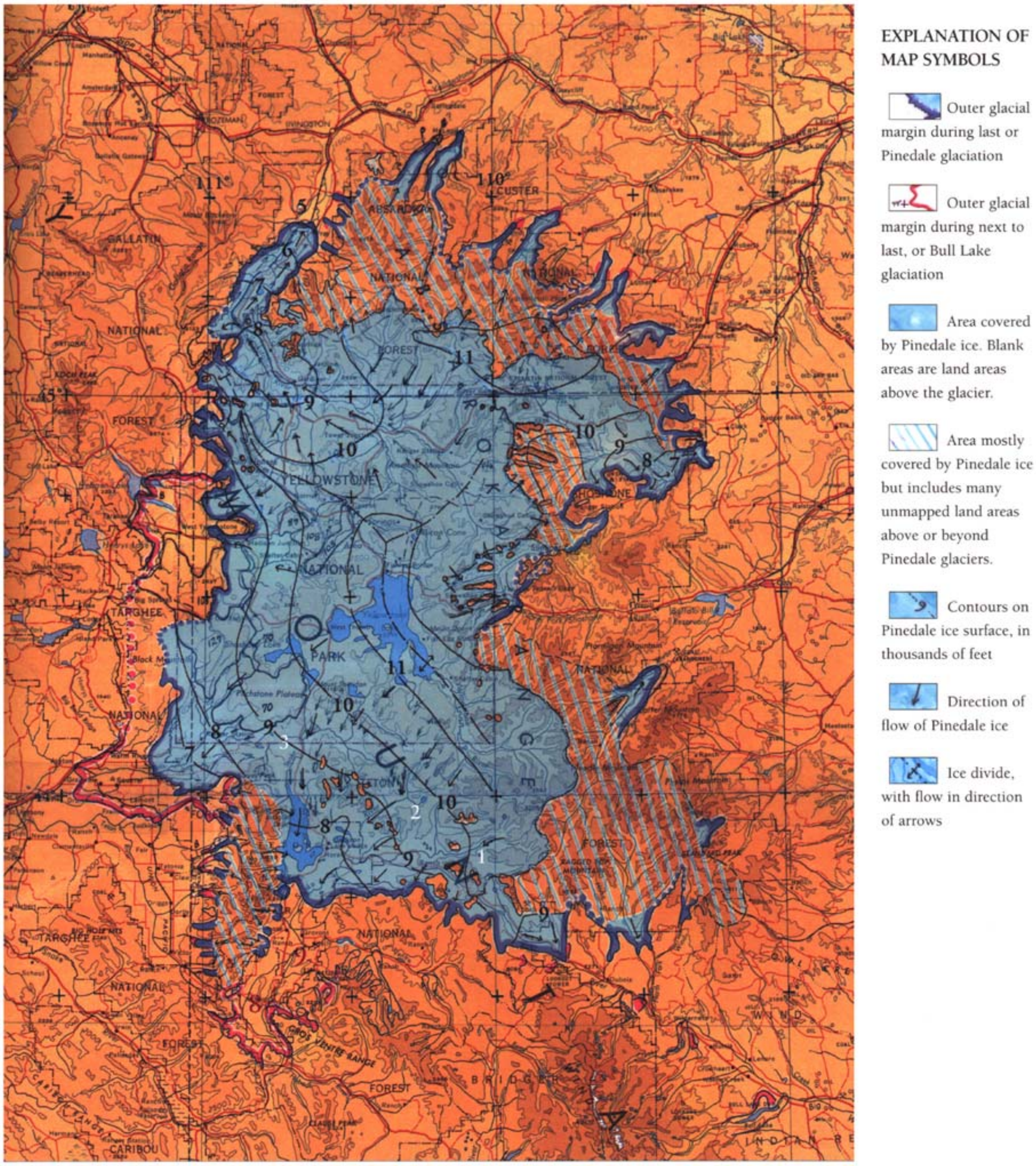


Figure 46 – Yellowstone glacial system showing maximum extent of Pinedale ice sheet. Margins of the older, Bull Lake phase are mapped where known in red. (From Good and Pierce, 1996, fig. 9.4, p. 32.)

The oldest dated glacier in Yellowstone is put at roughly 1.3 Ma. In both Yellowstone and the Grand Teton, the Bull Lake glacier (red outline on figure v) extended farther west than the youngest episode of ice flow, the Pinedale glaciation (Figure 47). The Pinedale accompanied growth of an ice cap exceeding 11,000', much thicker than the height of the Washburn range and many of the divides along the Absaroka ridge crest on the eastern park boundary. As a result, the effects of Pinedale ice can be seen throughout the park unless they are covered by volcanic or thermal features.

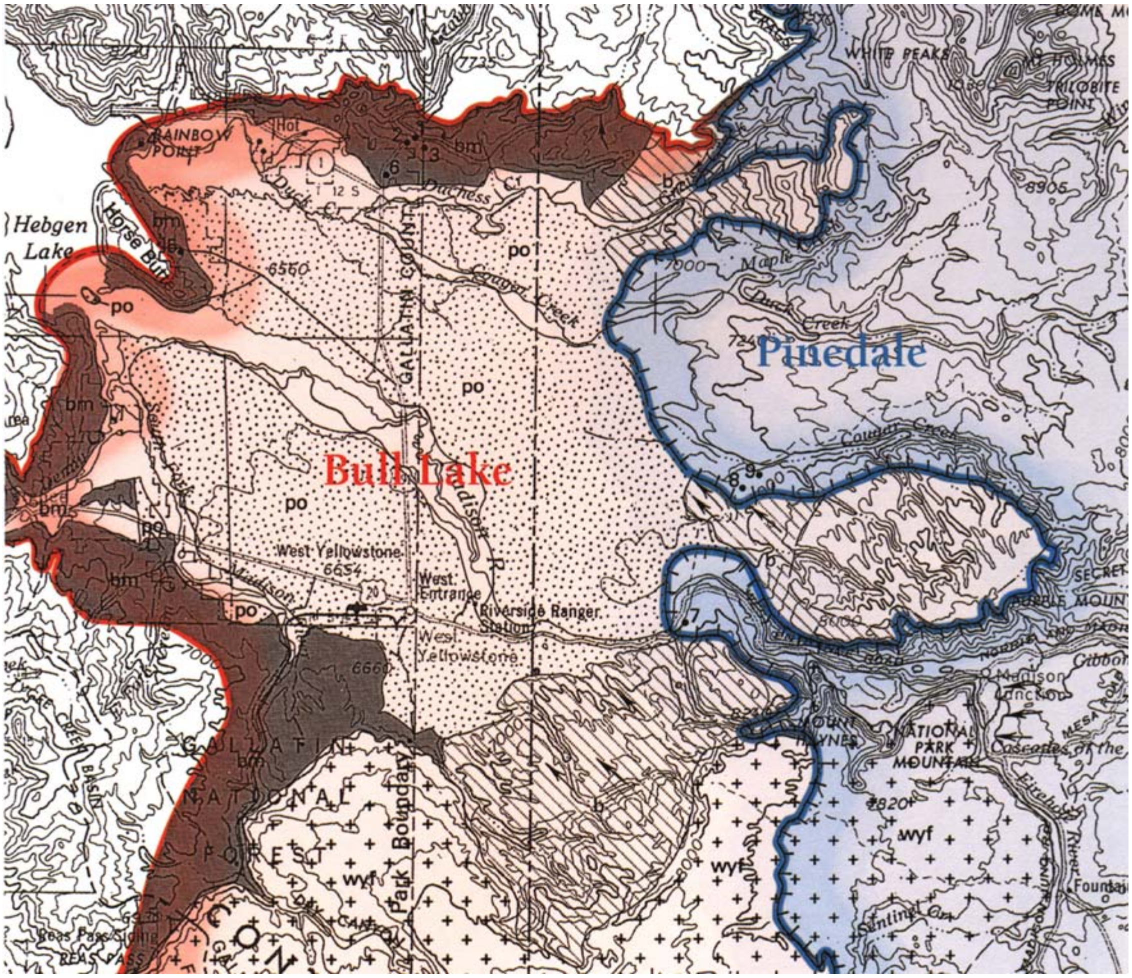


Figure 47 – Bull Lake ice sheet at 150,000 years ago extended about 20 km beyond the position of the 70,000 – 15,000 year old Pinedale ice sheet in the West Yellowstone Basin. By contrast, in the Yellowstone valley, Pinedale ice overrode all earlier deposits. (From Good and Pierce, 1996, fig. 9.8, p. 35.)

According to Ken Pierce, U.S. Geological Survey geologist, at the end of the last glacial period, about 14,000 to 18,000 years ago, ice dams formed at the mouth of Yellowstone Lake. When the ice dams melted, a great volume of water was released downstream causing massive flash floods and immediate and catastrophic erosion of the present-day Yellowstone canyon. These flash floods probably happened more than once. The canyon is a classic V-shaped valley, indicative of river-type erosion rather than glaciation. Today the canyon is still being eroded by the Yellowstone River.

Travelogue – Things to See in Yellowstone

Since so many excellent guides are available, I here provide a brief overview and road map of the outstanding geology features of Yellowstone Park. Consult the Hamilton map (Figure 48) for details of trails and view areas.

Old Faithful to Mammoth Hot Springs

One of the main routes to Old Faithful is from the south via Jackson, Wyoming, and the South Entrance. The park road crosses the Continental Divide three times. Waters flow east of the divide to the Atlantic, or west to the Pacific. This park route passes five geysers West Thumb, Upper (Old Faithful), Midway, Lower, and Norris - on its way to Mammoth Hot Springs. Sampling the world's largest concentration of geysers, you follow the beautiful Firehole and Gibbon rivers. Old Faithful visitor center and museum at Norris tell aspects of the park story.

Old Faithful

Old Faithful Geyser is the world's best-known and most faithful geyser. Eruption intervals have long varied around an average of 65 min., ranging from 30 to 120 min. Recent earthquakes have lengthened the average interval to 78 min. Eruption times are posted.

Old Faithful to Madison

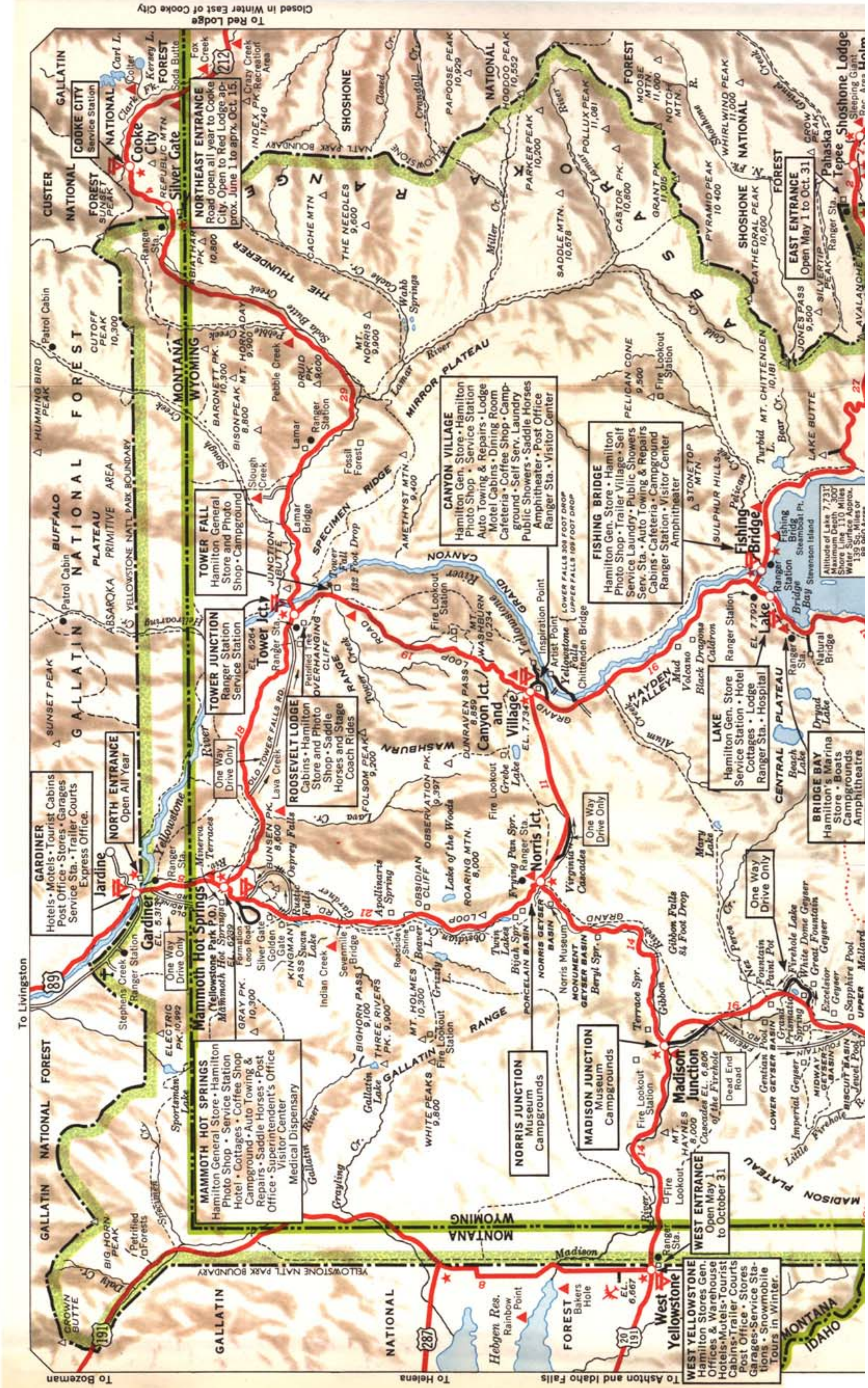
In Black Sand Basin the bright colors of Sunset Lake and Emerald Pool attract photographers. At Biscuit Basin, mineral deposits took on biscuit shapes before a 1959 earthquake triggered changes destroying the biscuits. At Midway Geyser Basin you may walk to Excelsior Geyser Crater and Grand Prismatic Spring. Firehole Lake Drive (one way, northbound) loops off the main road to Great Fountain Geyser, Firehole Lake, a hot pool, and the Three Senses Trail. The Lower Geyser Basin features the Fountain Paint Pots. Fountain Flat Drive exits west and ends at Goose Lake. Firehole Canyon loop drive (one way, southbound), starting south of Madison Junction, passes by Firehole Falls.

The museum at Madison Junction portrays the evolution of the national park idea. Roadside forests are mainly lodgepole pine, some reddened by the feeding of mountain pine beetles. West Yellowstone, Montana is 14 miles west of Madison Junction. From Madison to Norris you drive along the Yellowstone caldera's northwest rim. Gibbon Falls cascade over the caldera wall.

Norris Junction to Mammoth Hot Springs

Norris Geyser Basin's array of thermal features is unparallelled. Steamboat Geyser, the world's largest, erupts at irregular intervals of days to years. Echinus Geyser erupts about once per hour. Porcelain Basin is Yellowstone's hottest exposed area. Exhibits at Norris Museum explain geyser workings. At Norris Junction you can turn east toward the Canyon area. At Canyon you can go north to Tower Junction or south to the Lake Area.

Figure 48 ***(below)*** – Detailed map (2 p.) of Yellowstone National park from Hamilton's guide (1977).



To Livingston
To Bozeman
To Helena
To Ashton and Idaho Falls
To Red Lodge
Closed in Winter East of Cooke City
GARDINER
Hotels • Motels • Tourist Cabins
Post Office • Stores • Garages
Service Sta. • Trailer Courts
Express Office.
NORTH ENTRANCE
Open All Year
MAMMOTH HOT SPRINGS
Hamilton General Store • Hamilton
Photo Shop • Service Station
Hotel • Cottages • Coffee Shop
Campground • Auto Towing &
Repairs • Saddle Horses • Post
Office • Superintendent's Office
Visitor Center
Medical Dispensary
TOWER JUNCTION
Ranger Station
Service Station
TOWER FALL
Hamilton General
Store and Photo
Shop • Campground
ROOSEVELT LODGE
Cabins • Hamilton
Store and Photo
Shop • Saddle
Horses and Stage
Coach Rides
CANYON VILLAGE
FISHING BRIDGE
NORTHEAST ENTRANCE
EAST ENTRANCE
Open May 1 to Oct. 31
WEST ENTRANCE
Open May 1
to October 31
NORRIS JUNCTION
Museum
Campgrounds
MADISON JUNCTION
Museum
Campgrounds
LAKE
BRIDGE BAY
WEST YELLOWSTONE
MONTANA
WYOMING
IDAHO


Hamilton's
Map and guide to facilities in
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SCALE IN MILES
Main Auto Roads
Other Roads
Trails
Campgrounds
Mileages shown between stars
State Highway Numbers
U.S. Highway Numbers
Service Station Locations
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Hamilton Photo Shop • Hotel
Lodge • Cabins • Service Stations • Auto Towing & Repairs
Cafeteria • Post Office • Ranger Station • Visitor Center
WEST THUMB
Hamilton General Store
Hamilton Photo Shop
Service Station
SHOSHONE LAKE AND LEWIS RIVER CHANNEL
Open to hand propelled craft only
SOUTH ENTRANCE
Open May 1 to Oct. 31
COLTER BAY
Motel Cabins • Ranger Sta. • Marina
Campgrounds • Cafeteria • Shops
Post Office • Information Center
Trailer Court • Public Use Center
Grocery Store • Service Station
GRANT VILLAGE
Hamilton's Marina Store
Service Sta. • Auto Towing & Repairs • Campground
Boats • Visitor Center
Ranger Station • Amphitheatre • Self Service
Laundry • Public Showers
BRIDGE BAY
Hamilton's Marina
Store • Boats
Campgrounds
Amphitheatre
LAKE
Hamilton Gen. Store
Service Station • Hotel
Cottages • Lodge
Ranger Sta. • Hospital
FLAT MOUNTAIN, SOUTH AND SOUTH EAST ARMS
Open to hand propelled craft only
EAST ENTRANCE
Open May 1 to Oct. 31
GOFF CREEK LODGE
Service Station
One Way Drive Only
YELLOWSTONE LAKE
Fishing Bridge
Madison Junction
Grant Village
West Thumb
Old Faithful
Lewis Lake
Heart Lake
Jackson Lake
Shoshone Lodge
Pahaska Tepee
Holm Lodge
TARGHEE
NATIONAL
FOREST
SHOSHONE NATIONAL FOREST
TETON NATIONAL FOREST
GRAND TETON NATIONAL PARK
ABSAROKA RANGE
MONTANA
IDAHO
WYOMING

Continuing north of Norris you pass Obsidian Cliff. Obsidian, a volcanic glass excellent for projectile points and cutting tools, was traded across North America by Native Americans. Five miles south of Mammoth Hot Springs, at Swan Lake Flat's north end, a rough one-way dirt road goes around Bunsen Peak. Two miles south of Mammoth Hot Springs the Upper Terrace Loop Drive passes through a fascinating thermal area. Gnarled limber pine trees on some extinct formations are over 500 years old.

At Mammoth Hot Springs the terraces are spectacular travertine (calcium carbonate) formations deposited daily. Most new rock from Yellowstone's geysers is called geyserite, a noncrystalline mineral chemically similar to glass. Exhibits at Albright Visitor Center portray the park's early history and wildlife and tell how the U.S. Army protected the park from 1886 to 1916. Park headquarters is in the buildings of Fort Yellowstone, a late 19th-century cavalry post. Gardiner, Montana, lies five miles north. The Yellowstone River flows north, eventually to join the Missouri.

To Tower Junction and Canyon

The road east from Mammoth Hot Springs leads your four miles to Undine Falls, then 0.2 miles to Lava Creek. Three miles further east look for waterfowl and muskrats at Blacktail Ponds. Next, Blacktail Plateau Drive, a one-way dirt road eastbound, leaves the main road to traverse grass- and sagebrush-covered hills and forests of Douglas-fir, Engelmann spruce, and lodgepole pine. Watch for pronghorn antelope, mule deer, and elk. Scattered groves of quaking aspen trees turn gold in autumn. The next side road leads to a petrified redwood tree. Such trees may be found over hundreds of square miles in northern Yellowstone. Some are still in an upright position.

Tower Junction to Northeast Entrance

Lamar Valley, accessible all year, is winter range for elk and bison. You may camp at Slough Creek or Pebble Creek campgrounds en route to the Northeast Entrance, 29 miles from Tower Junction. Beyond lie Silver Gate (one mile) and Cooke City, Montana (four miles), and the Beartooth Highway climbs to 10,940 feet at Beartooth Pass.

Tower Junction to Canyon

Tower Falls, tumbling 132 feet, was named for the adjacent volcanic pinnacles. Tower Creek flows into the Yellowstone River. South from Tower Falls, as you drive up Mount Washington, look east, downslope, into prime grizzly bear country on Antelope Creek. This area is closed to human travel, to offer the bears refuge.

DO NOT ATTEMPT TO FEED OR APPROACH BEARS,
NO MATTER HOW “CUTE” YOU THINK THEY ARE!

The main road next crosses Dunraven Pass at 8,850 feet elevation, amidst broad-topped whitebark pines and spire-shaped subalpine fir. Meadows produce a profusion of wildflowers during the brief summer. From the Washburn Hot Springs Overlook south of the pass, you can see the Yellowstone caldera. Its north boundary is Mount Washburn and its south boundary is the Red Mountains 35 miles away. You can see the Teton Range on clear days, on the right beyond the Red Mountains.

Canyon

Because of bear activity here, camping is restricted to hard-sided units only. Exhibits at the Canyon Visitor Center explain the park's geology. A 2.5-mile loop road (one way) leads first to a spur road out to Inspiration Point. Here the Grand Canyon of the Yellowstone River plunges 1,000 feet below you. The canyon's colors were created by hot water acting on volcanic rock. It was not these colors, but the river's yellow banks at its distant confluence with the Missouri River, that occasioned the Minnetaree Indian name which French trappers translated as roche jaune, yellow stone. The canyon has been rapidly downcut more than once, perhaps by great glacial outburst floods. Little deepening takes place today.

Grandview Point affords a distant view of the 308-foot Lower Falls. Lookout Point affords a vista of Lower Falls, and a steep trail descends to a closer viewpoint. Back on the main road turn left in 0.3 miles to view the brink of the 109-foot Upper Falls. Back on the main road again, go 0.6 miles south to Artist Point Road and cross Chittenden Bridge to Uncle Tom's Parking Area. Trails here offer close views of the Upper and Lower Falls. South Rim Drive leads to Artist Point for another view of the canyon and Lower Falls.

Hayden Valley

The road here follows the Yellowstone River's meanderings across a former lakebed. After the great glaciers retreated, Yellowstone Lake was much larger than it is today, and this area was then flooded. The lake left fine silt and impermeable clay soil that permits little tree growth but allows rich shrubland that provides food for a variety of wild animals. Waterfowl, including white pelicans and trumpeter swans, abound in marshy areas. In this open parkland you may see moose, bison, and occasionally grizzly bears.

VIEW LARGE ANIMALS ONLY AT A DISTANCE, FROM YOUR CAR OR FROM ROAD-SIDES. Do not stop in roadways; use roadside parking areas for your safety.

No fishing is allowed for a 6-mile section in Hayden Valley. This provides quiet for the animals and views of untrammeled wilderness scenery for you. Stop at Mud Volcano and see the varied thermal features there. You might see spawning cutthroat trout jumping at Le Hardy Rapids, 3 miles north of Lake Junction, in June and July.

East Entrance to Fishing Bridge Junction

Cody, Wyoming, lies 50 miles beyond the East Entrance. As you cross 8,530-foot elevation Sylvan Pass, watch for pikes and yellow-bellied marmots in the rocky debris of talus slopes. You descend the west slope of the Absaroka Mountains, an eroded volcanic range named for the Crow Indians. Near Yellowstone Lake a spur road leads to Lake Butte Overlook for a view of this immense body of water. Yellowstone Lake occupies only the southeast quarter of the Yellowstone caldera. At the overlook you are 4 miles outside the caldera's east boundary.

Just north of the lake the Earth's surface is rising about 0.9 inches per year! This suggests future volcanic activity here. As you drive along the lake's edge, you can see Steamboat Springs. This is a hot spring remnant located on a line of faults, or fractures in the Earth, that also pass through Mary Bay and Indian Pond to the northwest. Bay and pond both occupy geologically recent hydrothermal explosion craters. The bottom sediments in Mary Bay are still very warm. Watch for moose browsing in the sedge meadows and marshes along Pelican Creek flats as you approach Fishing Bridge.

Exhibits at Fishing Bridge Museum feature the park's birds. Fishing Bridge itself spans the Yellowstone River, the lake's outlet. The bridge was closed to fishing in 1973. Fishing Bridge now offers one of the best wild trout spawning shows anywhere for most of the summer. White pelicans feed on the native cut-throat trout. Because of a high level of bear activity, only hard-sided units are allowed to camp in the Fishing Bridge area.

Yellowstone Lake is North America's largest mountain lake. Over geological time it has drained into the Pacific Ocean, the Arctic Ocean via Hudson Bay, and now drains into the Atlantic via the Gulf of Mexico. It is 20 miles long, 14 miles wide, and 320 feet deep at its deepest point. The average depth is about 139 feet. Trout generally inhabit the upper 60 feet because their foods rarely occur below that depth. The average surface temperature in August is about 60°F, and the bottom temperature never rises above 42°F. Swimming is discouraged even where not prohibited: Such cold waters can cause potentially fatal hypothermia or hyperventilation in minutes.

Boating is popular on some park lakes. Permits (required for all watercraft) and advice on canoeing and kayaking can be obtained at ranger stations at Lake Village or at Grant Village. A marina is at Bridge Bay and boat ramps are at Grant Village. Traveling toward West Thumb you may take a rough spur road, starting south of Bridge Bay, to see the natural bridge for which the area is named. Gull Point Drive loops off the Grand Loop Road for a closer view of the lake's edge.

West Thumb and Grant Village

Walk the boardwalk through the geyser basin at lake's edge at West Thumb. Intense heat measured in lake sediments below West Thumb indicates a shallow thermal system underlying this more recent caldera within the Yellowstone caldera. Should the lake level fall just a few feet, an immense steam (hydrothermal) explosion could occur here. That is what created the craters now filled by Mary Bay and Indian Pond, described above. Exhibits at Grant Village Visitor Center, two miles south of West Thumb, feature the park's immense wilderness. Fishing, boating, and backcountry use permits are available at the ranger station.

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.

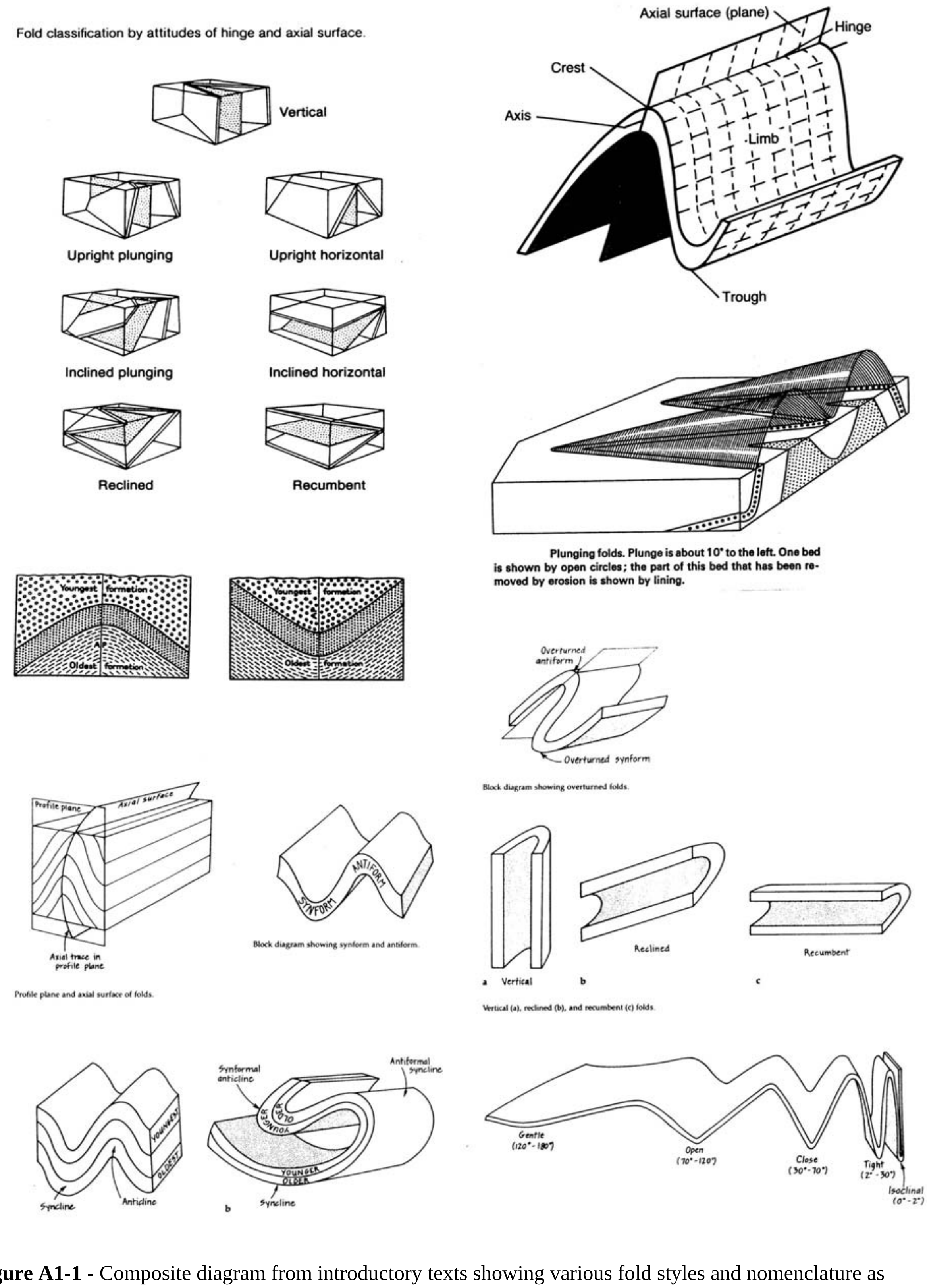

Figure A1-1 - Composite diagram from introductory texts showing various fold styles and nomenclature as discussed in the text.

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.

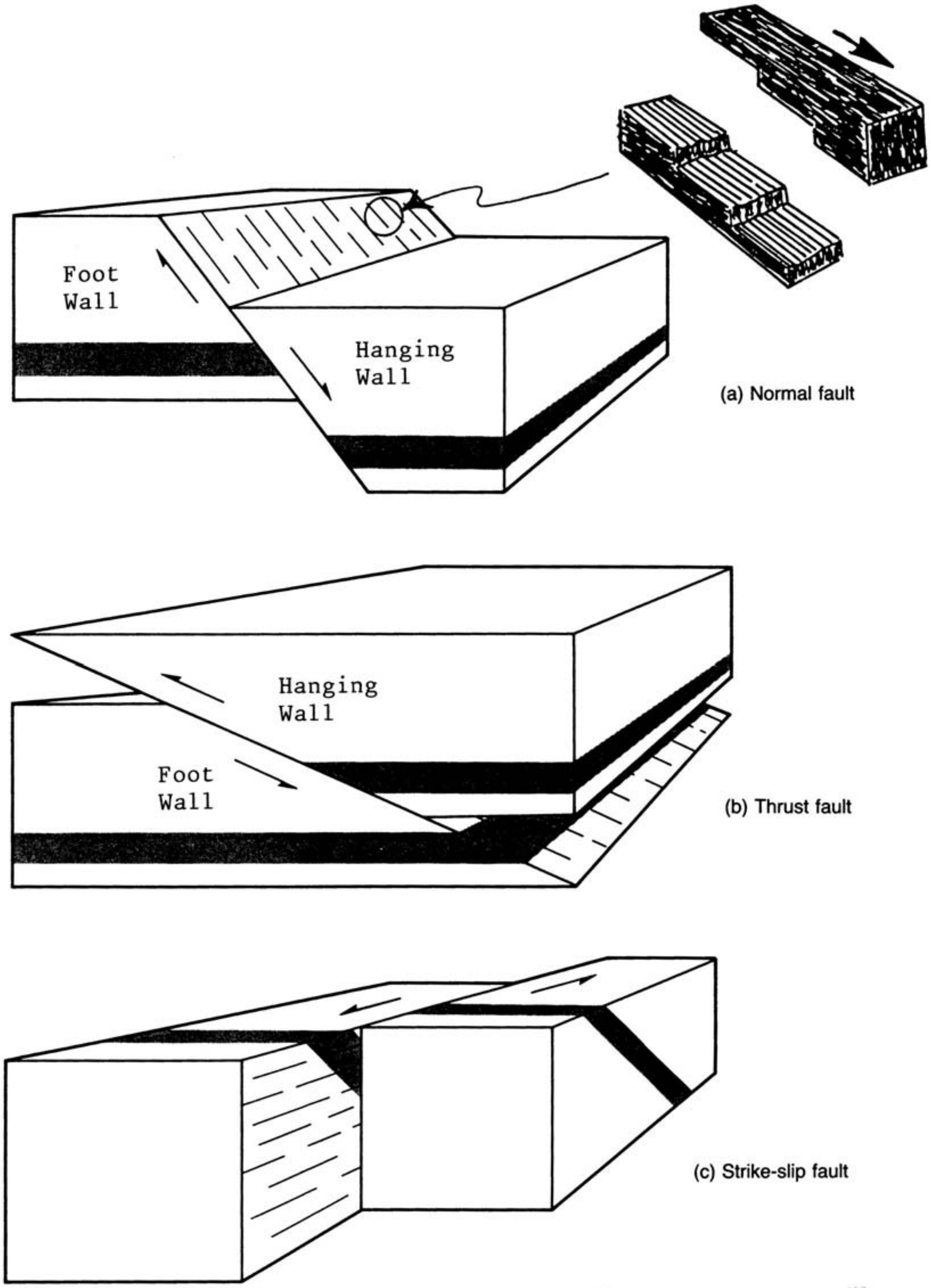

Figure A1-2 - The three main types of faults shown in schematic blocks. Along a normal fault (a) the hanging-wall block has moved relatively downward. On a thrust fault (or reverse fault) (b) the hanging-wall block has moved relatively upward. Along a strike-slip fault (c), the vertical reference layer (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.

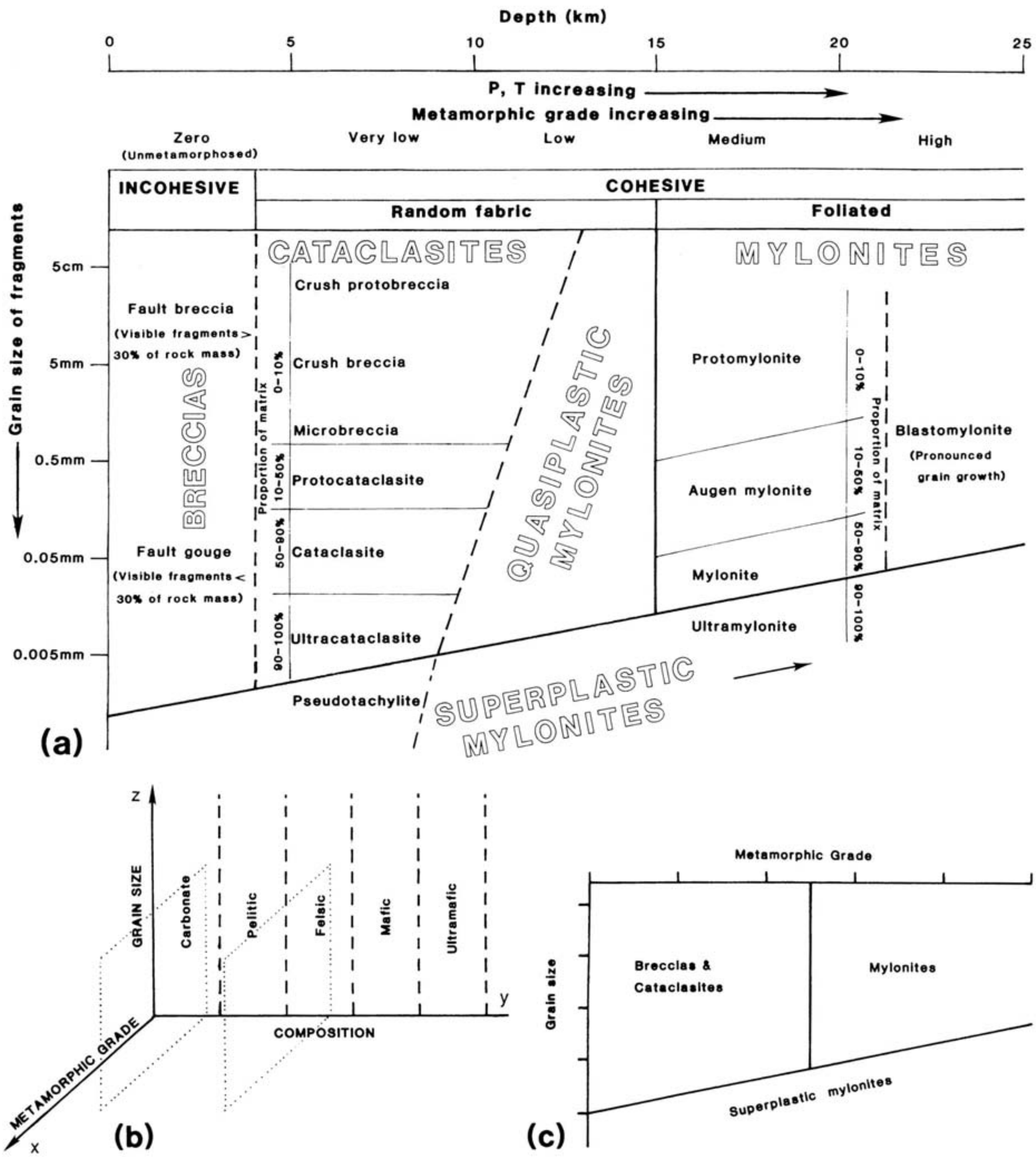

Figure A1-3 - Fault-rock terminology. (a) Classification of fault rocks that have been derived from quartzo-feldspathic lithologies (e. g. granite) (adapted from Sibson, 1977); (b) the grain size - metamorphic grade - lithologic composition grid used for classifying fault rocks (after Hull et al., 1986); (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]).

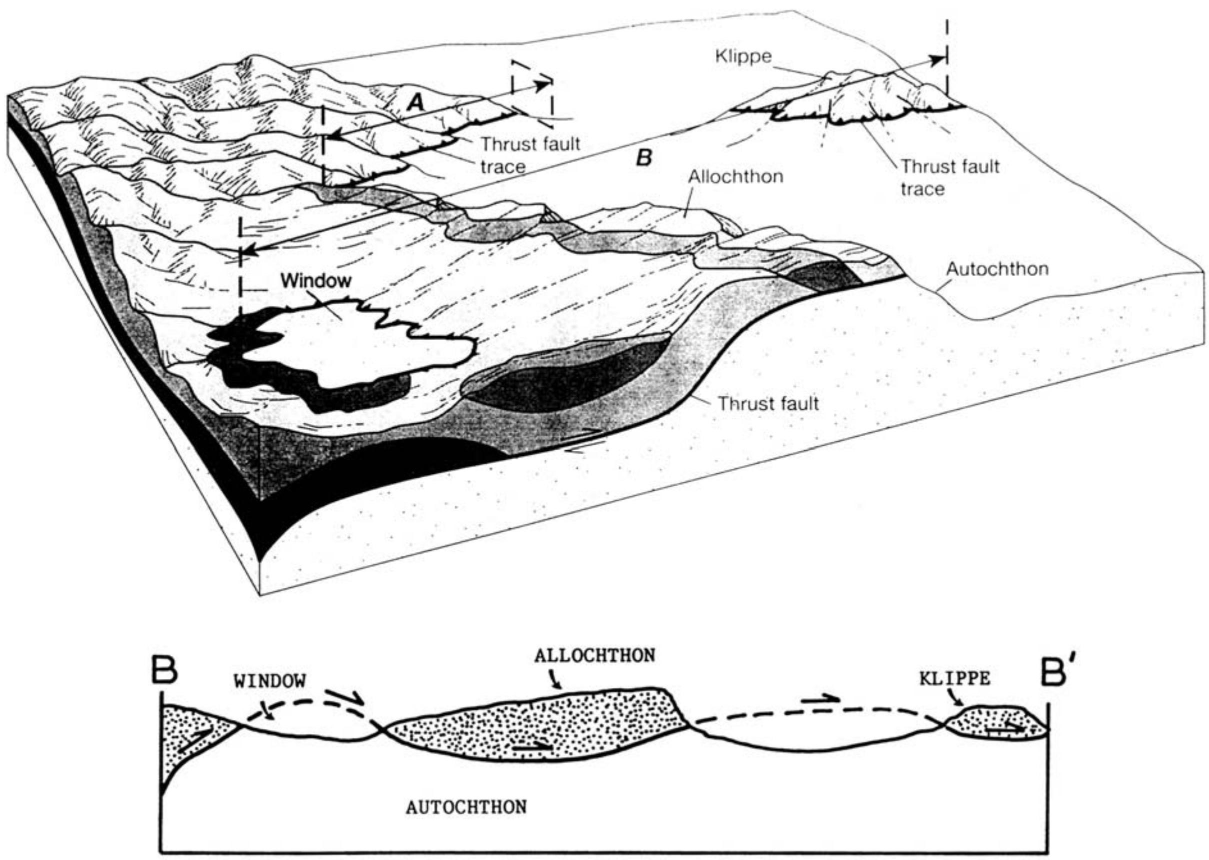

Figure A1-4 - Block diagram illustrating the relationships between allochthons, autochthons, klippen, and windows. (Twiss and Moores, 1992, p. 99) with section B-B' drawn by CM.

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.

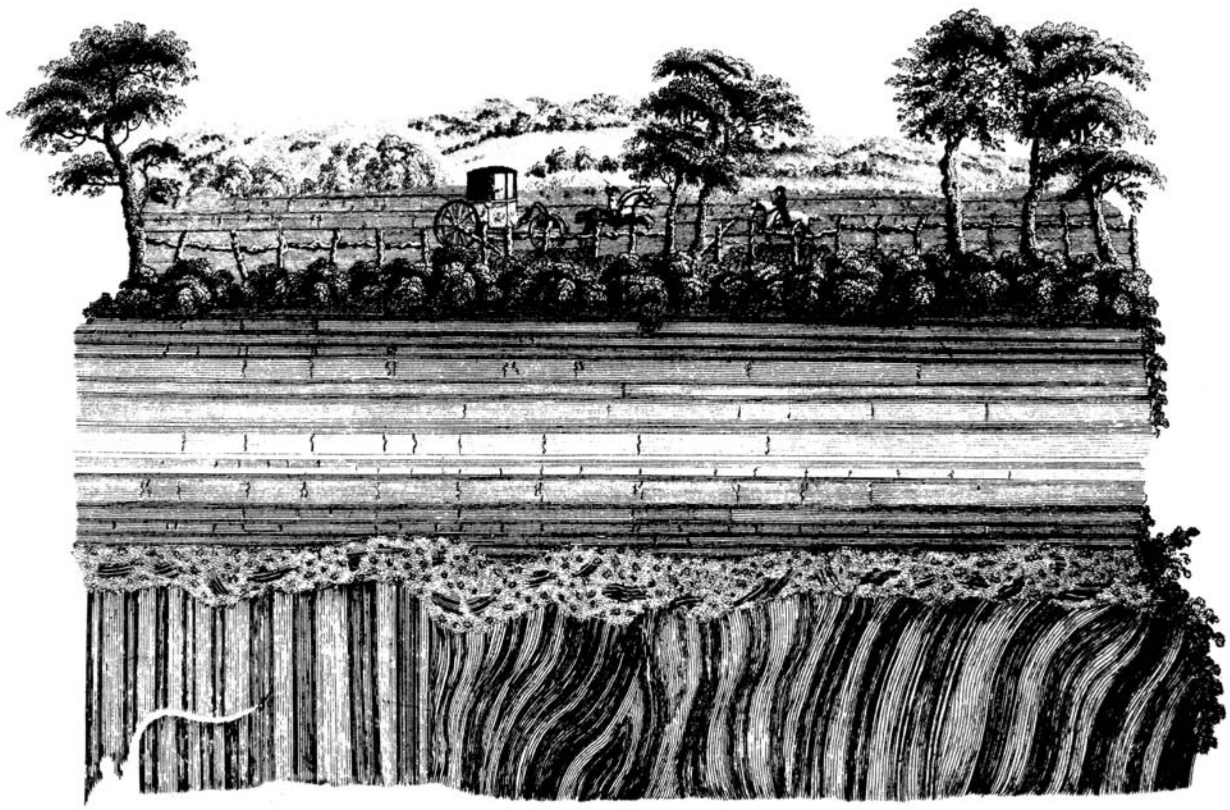

Figure A1-5 - Unconformity with basal conglomerate along the River Jed, south of Edinburgh, Scotland. From James Hutton's "Theory of the Earth", (1795).

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*.

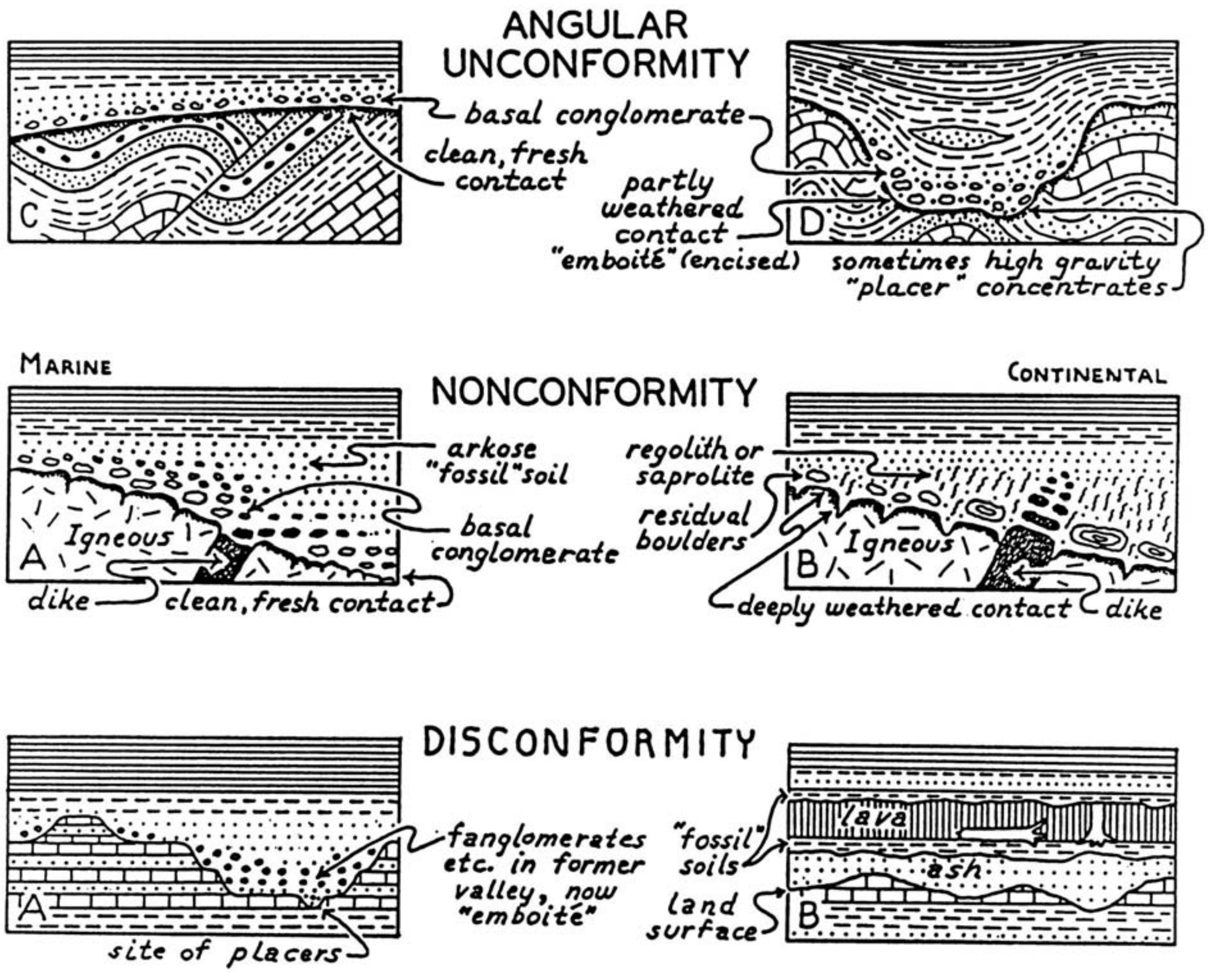

Figure A1-6 - Various types of unconformities, or gaps in the geologic record. Drawings by Rhodes W. Fairbridge.

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.

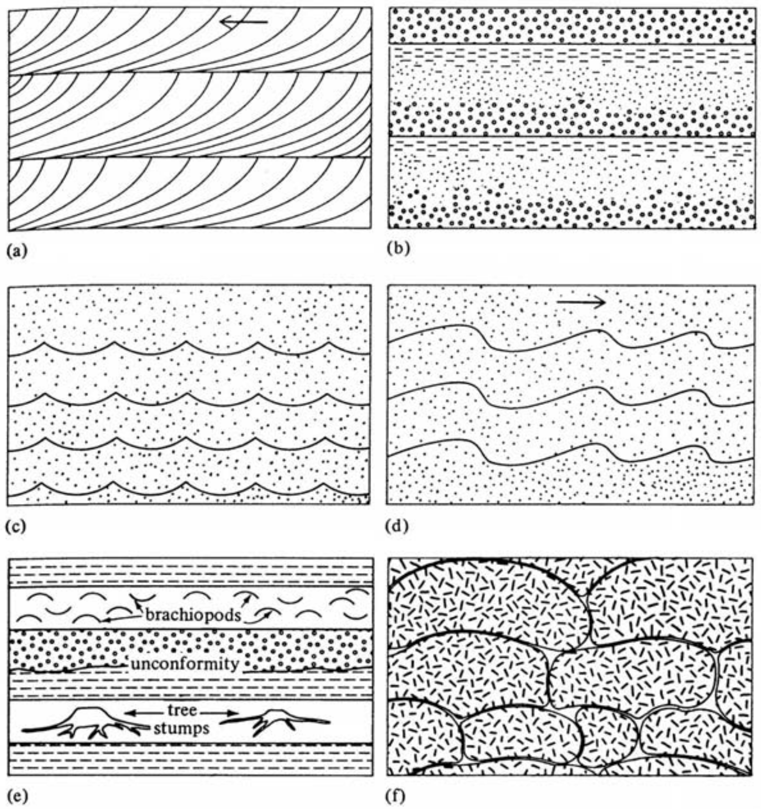

Figure A1-7 - Diagrammatic sketches of primary sedimentary structures (a through e) and cross sections of pillows (f) used in determining topping (younging) directions in rocks.

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