GEOLOGIC RESOURCES INVENTORY

of

GLACIER NATIONAL PARK

by

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Executive Summary

Glacier National Park encompasses about 1545 square miles (4100 square kilometers) of rugged, glaciated, mountainous terrain in northwestern Montana. The Park sits adjacent to Canada, straddling the continental divide over its entire length. The Park is dominated by two roughly parallel mountain ranges trending northwest-southeast, the Livingstone Range to the west and the Lewis Range to the east. With the courageous leadership of Montana Congressman Charles Pray, and the insistence of the early environmentalist George Bird Grinnell, the legislation designating the park was signed into law on May 11, 1910. When Glacier National Park was created, all the fuss about would-be mineral wealth had subsided in the face of hard economic reality. All that was left was what had always been present, a marvelous rugged mountain wilderness with breathtaking scenery. The National Park Service (NPS) Geologic Resource Inventory (GRI) for Glacier National Park will develop a digitized geologic map, a geologic resources bibliography (NRBIB), and this report to assist Park staff with understanding and managing Glacier National Park's geologic resources.

Glacier National Park is located within the northeastern section of Belt terrane, and contains incredible exposures of Precambrian age Belt Supergroup sedimentary rocks, which in Tertiary time were displaced eastward onto Cretaceous rocks by the Lewis thrust fault. These remarkable ancient rocks are relatively undeformed and unchanged by the forces of pressure and heat and remain among the finest examples of Precambrian sedimentary rocks in the world. They record an ancient Belt basin (sea) environment, opening and closing intermittently over many millions of years. Throughout this time period, water level fluctuations in the basin caused the shoreline and the accompanying deposition environments to shift position and change. The rocks record a rather quiescent northwestern margin to the North American craton.

The rocks, which were undoubtedly deposited atop the Belt rocks, of Paleozoic, Mesozoic, and Cenozoic age are missing at Glacier National Park; long ago eroded from the jagged peaks now present. This leaves a rather unique, if enigmatic geological setting. The Precambrian Era is notoriously difficult to decipher, much of the rock record necessary for tectonic correlation is missing or buried beneath younger rocks. For this reason study of the Belt Supergroup rocks at Glacier National Park is crucial for understanding at least one setting from the Precambrian, but the rock record gap leaves a heavy reliance on the study of adjacent areas to determine the tectonic setting between the Precambrian and the Tertiary.

The Belt Supergroup strata range in age from about 1,325 million years ago (m.y.) to about 900 m.y. They contain the Purcell Lava, which is a basalt flow (1,075 m.y.), as well as gabbroic sills and some dikes of similar age all of which provide isotopic dating opportunities. The Precambrian strata are mostly reddish-brown and greenish-gray argillite and siltite with some quartzite and carbonate (Prichard, Grinnell and Snowslip Formations). They were deposited in near and far shore environments. The dark, fine-grained layers of the Prichard Formation record a relatively deep-water depositional environment. The layered sandstones and mudstones of the Grinnell and Snowslip Formations record a shoreline environment with intermittent exposure and desiccation combined with submersion and occasional storm layers across the Belt basin. The Empire Formation is recognized in the Park, and it is a transitional unit as much as 1,170 ft (375 m) thick between the under lying Grinnell Formation and overlying Helena Formation. It records a transgressional environment in which the water of the basin deepened and the shoreline approached the craton to the southeast. The Altyn, Helena, and Shepard Formations compose the Belt carbonate units. They depict deeper water depositional settings with a carbonate saturated water chemistry. These layers are usually massive and the Helena Formation is of prevalent exposure along Going-to-the-Sun road. The stromatolites in these carbonate units record the only traces of Proterozoic life in the Belt basin. These unique features are formed by blue-green algae, primarily responsible for the oxygen rich atmosphere, which we breathe today.

The Cretaceous age rocks, present in limited exposure beneath the Lewis thrust sheet and east of the fault in Glacier record a myriad of marine environments present during the intermittent progradations and regressions of the Cretaceous Interior Seaway. These units include the Kootenai, Blackleaf, Telegraph Creek, Two Medicine, St. Mary, and Willow Creek Formations, the Marias River and Bearpaw Shales, and the Horsethief and Virgelle Sandstones. The fossils contained in these rocks indicate incredible biodiversity during this time. The Tertiary age Kishenehn Formation forms the thick valley fill units in the Park. These valleys formed as a response to tectonic extension in the region, where large blocks of rock bounded by normal faults, downdropped relative to the surrounding block. These dramatic features contain fill sediment several thousand meters thick. Quaternary age rocks in Glacier are represented by the glacial deposits from the

Pleistocene and Holocene and recent alluvial gravel deposits, present along Glacier's myriad of streams and rivers. Landslide and slope deposits are also prevalent in recent sediments due to the spectacular relief of the Park.

As might be expected by its name, glaciers played a dominant role in shaping Glacier National Park. Nearly all of the physical features in the Park have been chiseled, gouged, engraved, or trimmed by glacial ice. Glaciers flowed ever so slowly down drainages that now contain narrow lakes. The numerous mountain valleys that wind their way between the high peaks are U-shaped, the typical product of glacial sculpting. The Park has been the setting for at least ten periods of widespread glaciation, the last major advance occurring during the Pinedale Glaciation. During the mid-19th century, the more than 150 glaciers in Glacier National Park advanced to their furthest extent since the Pinedale. Since this event, now called the Little Ice Age, the remaining 50-60 glaciers are greatly reduced in size, some to the point of becoming snowfields, no longer capable of downslope movement.

Knowledge of the geologic resources should directly impact resource management decisions regarding the economic resources associated with Glacier National Park, potential geological issues pertaining to the Park, future scientific research projects, and interpretive needs. Economic resources are present at Glacier National Park in the form of copper ore and natural gas.

The unique geology of Glacier National Park lends itself to many possible scientific research projects. Some of these include:

- The Belt Supergroup records rare Precambrian age environments, yet a detailed correlation of the rocks present in the Park and those found to the south, north and west has not been completed.
- Ongoing discussion of life during the Precambrian age demands more detailed, microscopic study of the preserved sedimentary Belt rocks of Glacier.
- The erodability difference between Conophyton and Baicalia stromatolites
- Research of how observed glacier changes might affect streams and surface characteristics across a mountain landscape is of interest to ecosystem modeling and climate change research. Further work in the Glacier National Park area is needed to complete regional assessment of glacial recession, and address climatological and ecological implications.
- Continued study of why the northern Rocky Mountains exist so far inland from a continental margin. How compressional stresses were translated so far inland remains a geologic enigma.

Because of the nature of the landscape, several potential geological issues need to be considered with regard to land-use planning and visitor use in the Park. Along with a detailed geologic map and road log, a guidebook that would tie Glacier National Park to the other Parks in the Pacific Northwest area could enhance a visitor's appreciation of the geologic history and dynamic processes that not only created Glacier but also created the spectacular landscape of the entire region. Strategically placed wayside exhibits can help explain the geology to the visitor.

Geologic processes initiate complex responses that give rise to rock formations, surface and subsurface fluid movement, soil, and alcove formation. These processes develop a landscape that welcomes or discourages our use. The geology attracted ancient native peoples to the Glacier National Park area for hunting and ceremonial reasons. The geology inspires wonder in visitors to Glacier National Park, and emphasis of geologic resources should be encouraged so as to enhance the visitor's experience.

Key Words: glaciation, Belt Supergroup, northwest Montana geology, resource management

Introduction: The NPS Geologic Resources Inventory

The Natural Resource Inventory and Monitoring (I&M) Program is charged with helping to prevent the loss or impairment of significant natural resources in 273 of the approximately 377 units of the National Park System. Air and water pollution, urban encroachment, and excessive visitation are only some of the unfavorable influences that may adversely impact many of the natural resources in the system. Gathering information about the resources and the development of techniques for monitoring the ecological communities in the National Park System (NPS) are the principal functions of the I&M Program. Ultimately, the inventory and monitoring of natural resources will become integral parts of park planning, operation and maintenance, visitor protection, and interpretation. An inventory of natural resources and subsequent monitoring of these resources will enhance the preservation and protection of natural resources and improve the stewardship of natural resources in the NPS. An inventory of natural resources is imperative in order to identify the links between changes in resource conditions and the causes of any negative changes. Without identifying these links, the elimination or mitigation of such causes will be neither systematic nor definitive.

Geologic resources constitute the foundation of all ecosystems within the NPS units. Bedrock and surficial geology is basic to the evolution of soils, vegetation, water quality, surface and groundwater distribution, and wildlife. Knowledge of the geologic structures and stratigraphy is critical to park management with regards to construction and maintenance of roads, buildings, water wells, and trail systems. Documentation of the geologic resources requires a comprehensive geologic map.

Part of the NPS Geologic Resources Inventory (GRI) includes bedrock and surficial geologic maps. Geologic maps provide the foundation for studies of hydrology, geomorphology, soils, prospective paleontology and archaeology sites, and environmental hazards. Examples of how better understanding of geology has aided management include the use of geologic data to construct fire histories and to identify habitat for rare and endangered plant species, areas with cultural and possibly paleontological resources, and the location of potential hazards for park roads, facilities, and visitors. Geologic maps describe the underlying physical habitat of many natural systems and are an integral component of the physical inventories stipulated by the NPS in its Natural Resources Inventory and Monitoring Guideline (NPS-75) and the 1997 NPS Strategic Plan. The NPS Geologic Resources Inventory is a systematic, comprehensive inventory of the geologic resources in National Park System units and is being implemented through the cooperation of the NPS Geologic Resources Division, the Inventory and Monitoring Program of the Natural Resource Information Division, the U.S. Geological Survey, and state geological surveys. For each of the 273 designated NPS units, the GRI will produce the following products:

- A bibliography of geologic literature and maps called GRBib
- An evaluation of available geologic maps, resources, and issues
- Digital map products and accompanying information
- A report documenting basic geologic information, hazards and issues, and existing data and studies.

Prior to a GRI report, a workshop is conducted at the individual NPS unit wherein specific geologic issues and maps are discussed with regards to management needs and potential research projects. The results from these workshops along with basic geologic data are included in the GRI report for that NPS unit.

Geologic Resources of Glacier National Park

The Crown of the Continent.

- George Bird Grinnell, instrumental in the formation of Glacier National Park

A Sense of Place: The Relevance of Geology to the Park

The Glacier National Park area includes about 1545 square miles (4000 km^2) of rugged, deeply glaciated mountainous terrain of northwestern Montana. The Park is defined by sweeping U-shaped, glaciated valleys bounded by dramatic, shear rock faces topped by pinnacles of layered rock (figure 1). The many streams and lakes, following the paths of previous glaciers descend outward from the continental divide, which runs up the middle of the Park. Glacier National Park (GLAC) was created in 1910 to preserve and protect the astounding natural beauty and myriad of ecological systems found there (figure 2). Ancient Native Americans considered the area to be sacred and used it for hunting range. Though scarcely preserved, ancient campsites and other archaeological finds are present to denote this use. Later fur trappers and prospectors took advantage of glaciers wildlife abundance and potential economic resource wealth. It is because of the spectacular geology that people of every time period have found this land to be special.

Figure 1: Layered Belt Supergroup Rocks comprise Clements Mountain, near the Logan Pass Visitor Center. NPS Glacier National Park photo.

Though used intermittently by Native American tribes as hunting range for thousands of years, the first Europeans to explore the area were probably French fur traders. The expedition by Lewis and Clark in the early 1800's skirted the area to the south in favor of an easier mountain crossing. After the Civil War, the region, like much of the American West, embodied the ideals of untamed capitalism, and was seen as a sea of endless possibilities for settlement, mining, and later, fossil fuel extraction. The first big gold strike in the region occurred in 1862, in southwest Montana. Two years later there were enough people in the region to achieve territorial status.   Between 1865 – 1869, the gold strikes in nearby British Columbia and Saskatchewan caused more attention in the area by prospecting parties. These hopefuls faced a substantial deterrent to exploration, the Blackfeet Nation.

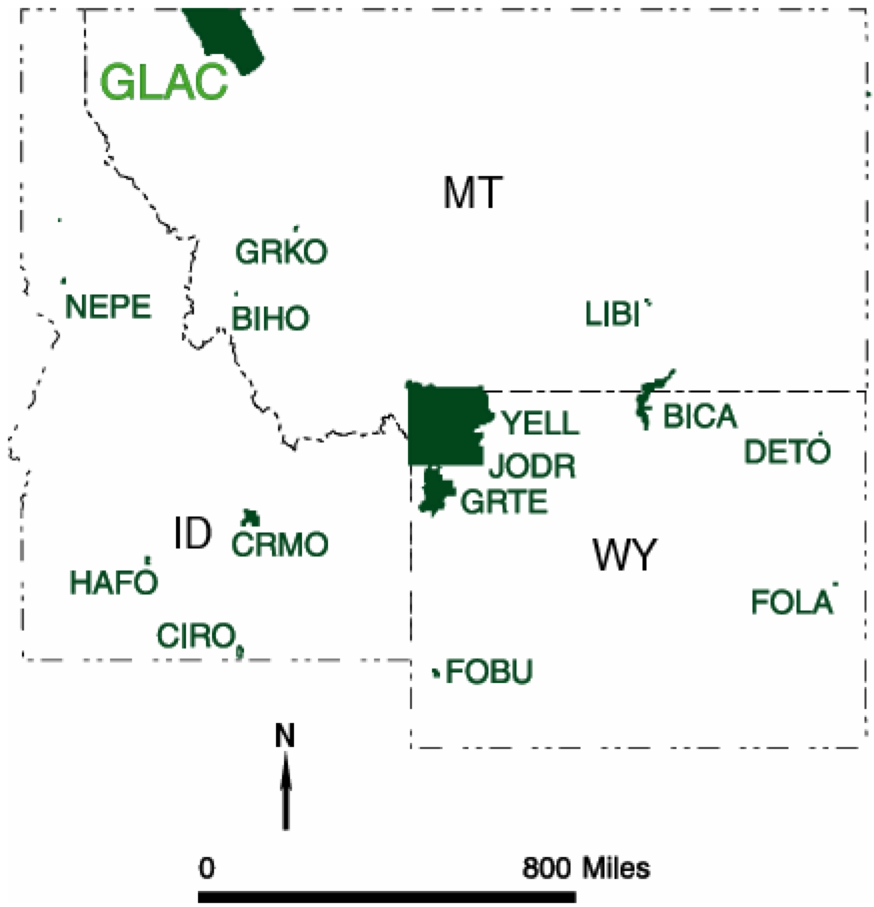


Figure 2: National Park units located in Montana, Idaho, and Wyoming. Glacier National Park is located in northwest Montana. See Appendix B for a list of Park unit abbreviations.

The Blackfeet Indian reservation was established in 1855 and included the area east of the continental divide of present-day Glacier National Park. The constant probing of prospectors and explorers caused serious tension that escalated to the controversial "Baker Massacre" or the retribution against the Blackfeet for the killing of an intruding settler resulting in the slaughter of nearly 200 native individuals. This event and the devastating smallpox epidemic ended Blackfeet resistance along the Montana Front Ranges. When Dutch Lui's illegal prospecting efforts bore fruit (copper ore) east of the continental divide, word spread and soon other squatters were disregarding the reservation laws. The incoming Great Northern Railroad increased interest and accessibility of the region. The promise of instant wealth from mining led Montana residents to put pressure on Congress to open up the land for legitimate staking of mining claims. In 1895, George Bird Grinnell, William C. Pollack, and Walter M. Clements were appointed to negotiate with the Blackfeet Nation over the sale of the mountainous land east of the continental divide of their reservation (figure 3). The Blackfeet, desperate and pressured, sold the "ceded strip" for $1.5 million in 1896; an Act of Congress officially changed the land on April 15, 1898. The area of the ceded strip was nearly half the area of the present day Park.

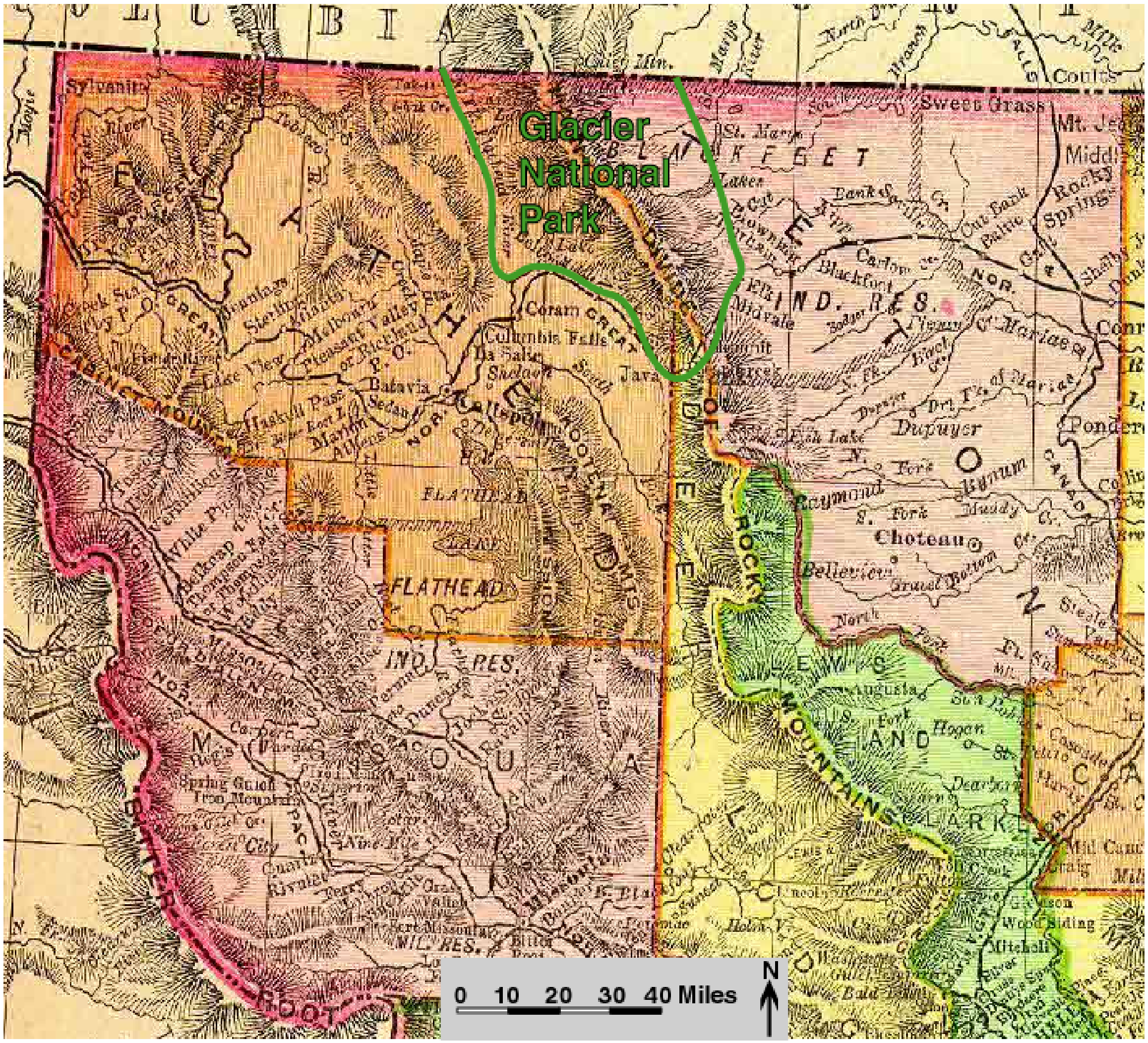


Figure 3: Excerpt from Rand, McNally & Co.'s 1895 atlas of Montana, showing the Glacier National Park area prior to the purchase of the "ceded strip" and ultimately, the creation of the Park. Blackfeet Reservation is outlined in Teton County (pink area). Park outline and scale added for reference.

On April 15, 1898, the rush officially started. Prospectors began pouring into the area. The most lucrative mining claims were clustered at Rose Creek, Boulder Creek, Cracker Lake in the Swift Current Valley, and above Slide Lake. Some reports estimate over 2,000 mining claims were staked in what is now Glacier National Park by 300 or so individuals. There was some talk of a railroad spur to Swiftcurrent Valley, but that never saw the light of day. By 1902, the ceded strip, so alluring to impatient prospectors, was largely abandoned because no economically viable deposits of ore were uncovered. The prospectors all moved to the next "hill" in search of instant riches and the ceded strip was left to sit.

In December 1907, Montana Senator Thomas Carter first introduced legislation to authorize the land as a national park to an outcry of objection from many Montanans, most notably from those living closest to the proposed park. Local newspapers, the *Inter Lake*, the *Kalispell Journal* and the *Kalispell Bee* editorialized against the legislation mostly on economic grounds. They saw the creation of a national park as a loss of harvestable timber, development opportunities, and mining jobs. The *Inter Lake* editorialized its concerns about "throngs of wandering tourists." Fortunately for future generations, Montana Congressman Charles Pray, noted environmentalist George Bird Grinnell and James J. Hill of the Great Northern Railroad, fought the uphill battle to pass the legislation designating the park signed on May 11, 1910. By 1910, the fuss about would-be mineral wealth had petered in the faced of hard economic reality and the rugged mountain wilderness with the breathtaking scenery, rich in geologic wonders, was claimed by the federal government as a protected national park.

The name *Glacier* National Park is somewhat of a misnomer. Many visitors come to the park expecting to see enormous glaciers akin to those in Alaska. While those undoubtedly existed at one time in Glacier as evidenced by the dramatic U-shaped valleys, hanging valleys, cirques and other glacial-specific features, today the warmer, drier climate could not support such ice masses. There are glaciers in the park, most have retreated back to their source cirques, but many are still viewable by the public. What does remain of the past ice ages, in spectacular fashion, are the sculpted rocks and remnants left behind when these massive glaciers melted some 10,000 years ago.

The story of this corner of America known as Glacier National Park is the story of geology and ice: the natural ice of snow, storms and glaciers and the rocks that form the basis for the entire ecosystem. Glacier National Park lies in the leeward rainshadow of the Continental Divide. The distribution of alpine tundra and upper treeline environments in Glacier National Park is a complex pattern resulting form interactions of wind-redistributed snow, soil moisture conditions, slope steepness, aspect and exposure, and catastrophic slope processes. The elevation of alpine tundra is therefore variable, but is typically encountered above ca. 2,000 m (6562 ft). Far below in the sweeping U-shaped valleys, a myriad of ecosystems exists from wetlands, old growth cedar-hemlock forests, and drier lodgepole pine forests in the west to windswept, stunted-growth slopes in the east, just to name a few. The complex interactions of climate and geology create the necessary conditions for this variety to exist.

The dramatic vistas, abundance of plant and animal life, and spectacular ice sculpted rocks record dynamic ecological relationships. To fully appreciate the natural splendor of Glacier National Park, however, one needs to understand the significance of the rocks and why they are unique to Glacier both in composition and form. The experience of

Glacier National Park begins with the geology, with the processes that established the groundwork from which the present-day environments and scenery arise.

For more on the history of Glacier National Park, visit the GLAC website: http://www.nps.gov/glac/index.htm

Regional Geologic Setting

Located in the northwestern corner of Montana, Glacier National Park is part of a geological feature called the Belt Terrane (figure 4). Covering parts of Montana, Idaho, Washington, and southern Canada, the Belt Terrane contains some of the oldest preserved sedimentary rocks in North America. Even in its structurally foreshortened, compressed state, the Belt Supergroup covers are large part of the U.S. Rocky Mountains. The region is almost entirely mountainous with the Lewis thrust fault front forming the abrupt eastern edge. The original western and southwestern limits of Belt deposition are, however, problematic. To the southwest in central Idaho lie the Middle Proterozoic Yellowjacket Formation and Lemhi Group, which contain rock types similar to those of the Belt, and may in fact, represent the southwest extent of the Belt basin. The western lip of the modern day Belt terrane is marked by combinations of: 1) Late Proterozoic Windermere Supergroup unconformably overlying the Belt, 2) Mesozoic accreted terranes, 3) Mesozoic granitic complexes, 4) Eocene extensional core complexes containing Early Proterozoic metamorphic rocks cut by Phanerozoic granite, and 5) overlapping Miocene Columbia Plateau flow basalt. Thus the Proterozoic western margin of the Belt basin cannot be positively identified in the rock record and was probably removed by latest Proterozoic crustal rifting (Winston 1989).

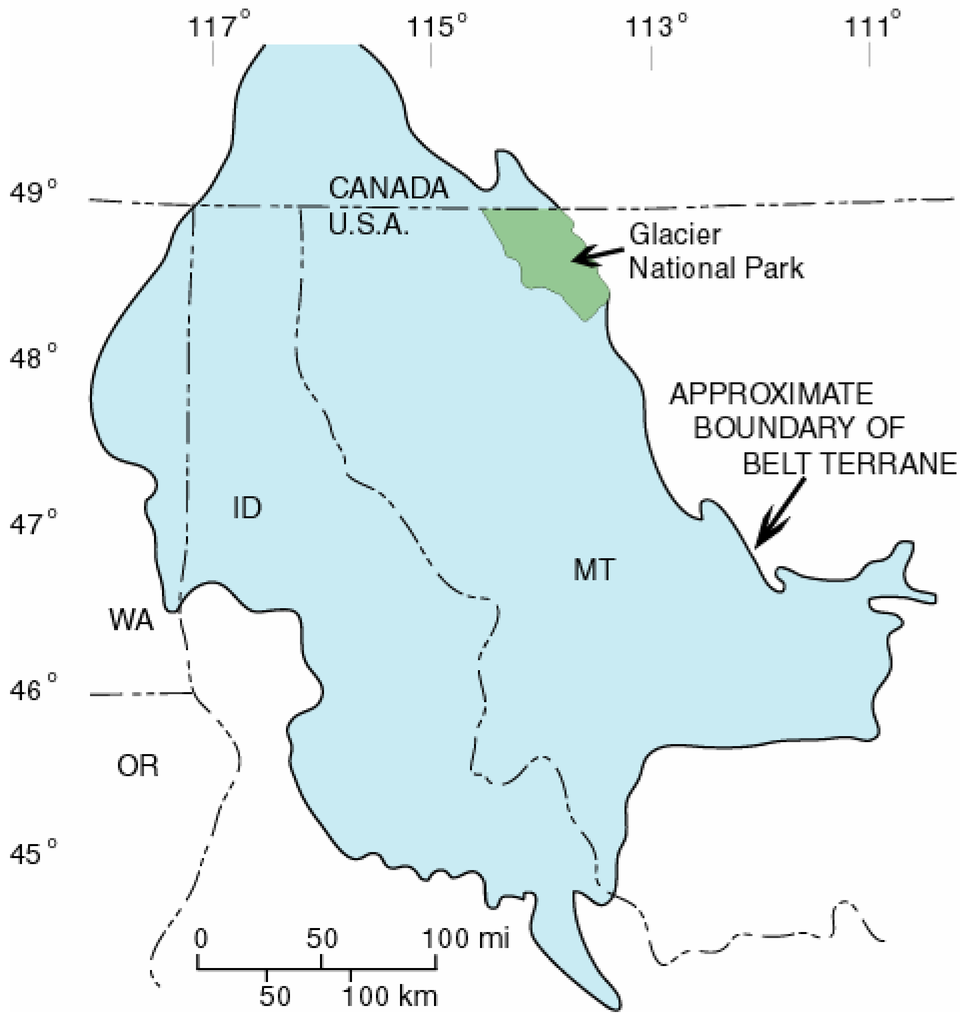


Figure 4: Location of Glacier National Park relative to the entire Belt Terrane.

Southeast of the former Belt basin lies crystalline basement of the Archean age Wyoming province of the North American craton onto which Belt sediments lapped and over which Belt rocks have been thrust. Farther south, Belt rocks lie in an autochthonous (formed in the place where now found) east-trending graben, called the Helena embayment, which may extend in the subsurface as far east as the Williston Basin near the Montana-Dakotas boundary. During the Middle Proterozoic, the Belt basin was bound on the southeast by the uplifted Dillon block of the Archean Wyoming province, along the east-trending Perry fault line (figure 5) (Winston 1989).

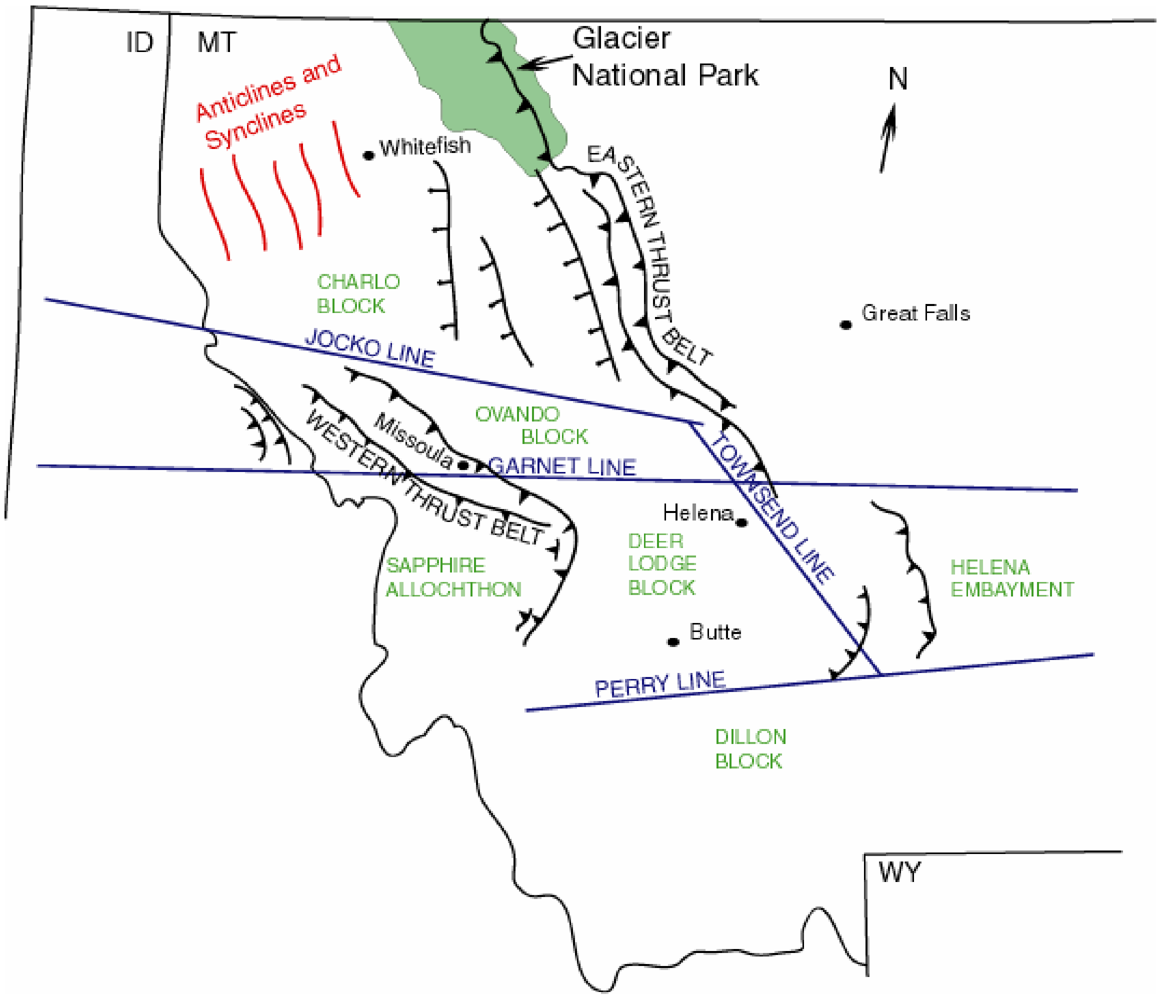


Figure 5: Map of present day structural features in the Glacier National Park area superimposed on Mid-Proterozoic fault lines (blue) with cratonic blocks labeled in green. Red lines indicate fold axes, black lines with teeth indicate thrust faults with teeth on overriding plate, black lines with bar and ball show normal faults with the symbol on the downthrown block. Adapted from Winston 1989.

Parts of Glacier National Park and the adjacent areas are in what is called the northern disturbed belt of Montana. This is a narrow north-south trending strip of heavily fractured and deformed rock. The area east of the mountains contains thrust-faulted and folded Upper Cretaceous strata; it is equivalent to the Foothills structural province in southern Alberta. The area southeast of the Park contains thrust-faulted and folded Jurassic and Cretaceous rocks, which locally are cut or transected by northeasterly trending normal faults. These strata plunge northwest beneath the Lewis thrust plate and are not clearly exposed in Glacier National Park or southern Alberta and British Columbia. The Lewis thrust plate, or mass of material above the fault face, itself is deformed by numerous folds and small normal and thrust faults. The major structure in the plate is a northwesterly trending, doubly plunging syncline, called the Akamina syncline. The largest normal fault in Glacier Park is the Blacktail fault, which extends northwestward into British Columbia as the Flathead fault. West of it are other northwesterly trending normal faults bounding ranges such as the Swan Range, the Whitefish Range and the Mission Range. The measured minimum easterly translation of the Lewis is 15 mi (24 km), but it may have moved at least 40 mi (64.4 km) (Mudge 1977; Winston 1989; Elias 1996).

The Park is in a southwesterly trending, structurally low area that is bounded on the north and south by southwest-trending structures or arches. The low (bowl-shaped depression) structural configuration of the basement rocks that influenced the present setting of the Glacier Park area was very likely established by the end of Cretaceous time, and it probably controlled, at least in part, the present northwest trending structural pattern in the area, a pattern resulting from the Laramide (early Tertiary) orogeny (Mudge 1977). The Park is dominated by two linear mountain ranges trending northwest-southeast. On the west side of the Park, the Livingstone Range runs about 35 km (22 miles) from the Park boundary to Lake McDonald. To the east, the Lewis Range extends the full length of the Park, a distance of about 100 km (61 miles) from the International Park boundary on the north to Marias Pass on the south. The Continental Divide runs along the crest of the Lewis Range from Marias Pass to about 16 km (10 miles) south of the U.S. – Canada border; then it swings west to the crest of the Livingstone Range into Canada (Elias 1985).

Glacier National Park contains the eastern front of the Rocky Mountains of the United States (figure 6). It is also part of a series of northwest trending mountain ranges in Western Montana including the Whitefish Range (immediately west), the Flathead, Swan and Mission Ranges (to the southwest). The Sweetgrass Hills (part of the Sweetgrass Arch) bound the Park on the east. The Purcell Anticlorium, to the west of the Whitefish Range, contains the Purcell Mountains and the Salish Mountains in a region of broad folds that forms a large, elongate dome structure in far western Montana and eastern Idaho. The Bitterroot Mountains, which form the border between Montana and Idaho were mistaken by the Lewis and Clark survey team in 1805 as the continental divide. They followed the Bitterroots instead of the Anaconda Range north into Montana when surveying the boundary giving Idaho its "panhandle" and Montana some spectacular scenery.

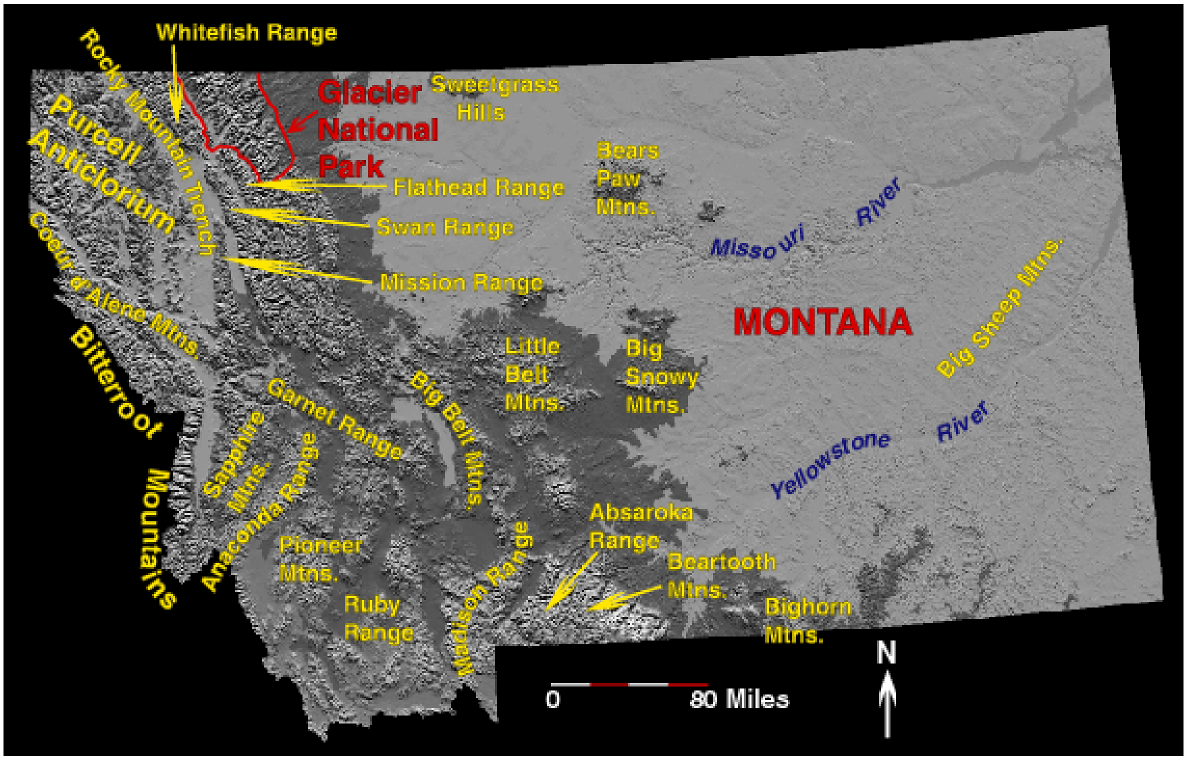


Figure 6: Map of Montana with physiographic features surrounding Glacier National Park. Modified from map from Perry-Castañeda Library Map Collection.

Glacier National Park contains the best exposed and most complete sections of the Belt Supergroup rocks. A partial section of the Belt Supergroup is about 2900 m (9514 ft) thick in the central and northeastern margin of the Park, with the base being in fault contact with underlying Cretaceous strata and the top having been removed by erosion. Neither the base nor the top of the Belt Supergroup is present anywhere in Glacier National Park. Although these strata were subjected to lowermost greenschist-facies metamorphism (~300ºC and 2 Kbar), details of sedimentary structures and fine sedimentary laminae are extremely well preserved, and for the purposes of sedimentary study are considered to be essentially unmetamorphosed (Horodyski 1983). Because they were deposited before the evolution of complex plants and animals, the Belt Supergroup provides a unique context to study ancient sediments undisturbed by the bioturbation ubiquitous with younger rocks. The only organisms present to leave their mark on Belt rocks are stunningly variable stromatolites or algal mats present in many rock formations of Glacier National Park (figures 7 and 8).

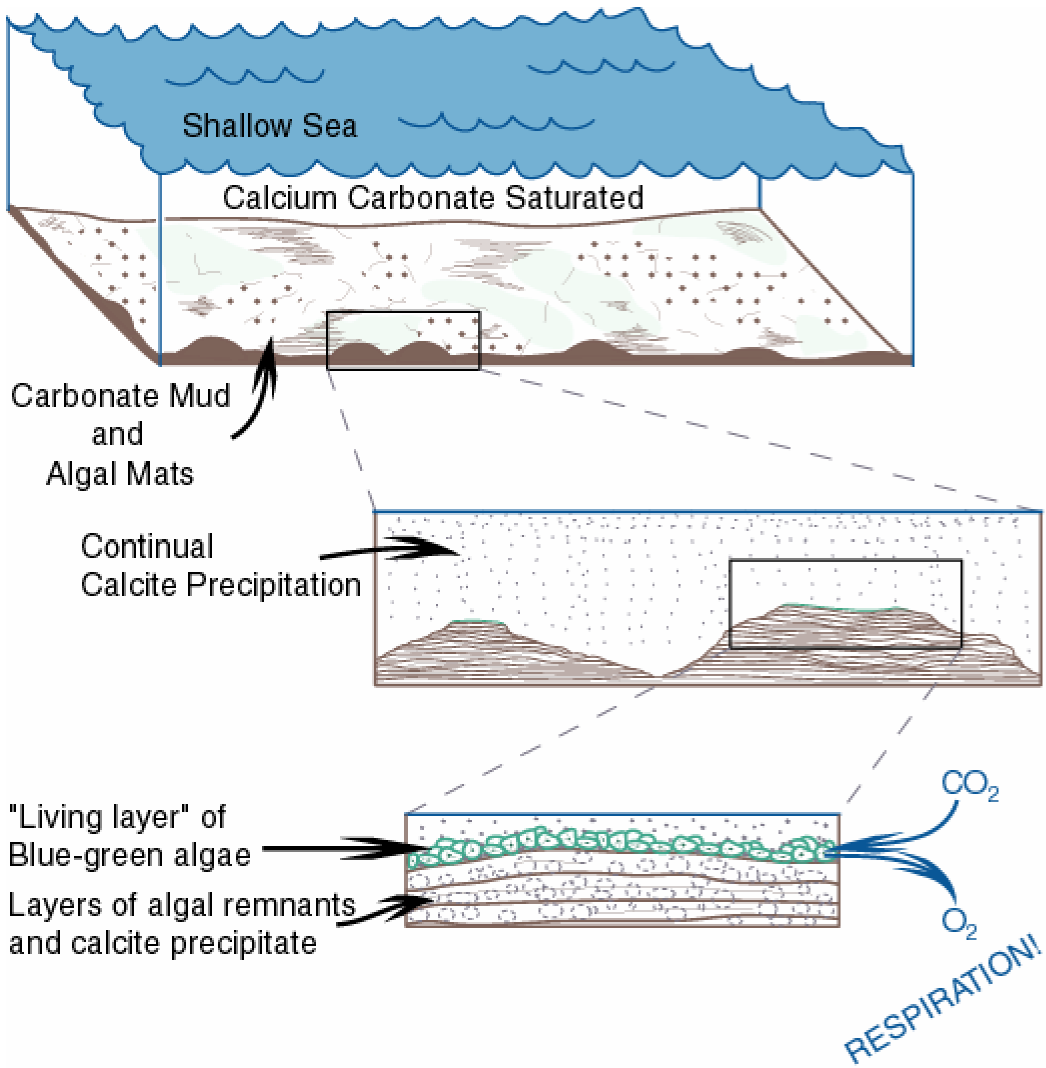


Figure 7: Formation of stromatolites from a combination of precipitation of calcium carbonate from seawater and layers of blue green algae.

FORMATION NAME	DESCRIPTION OF ROCK FORMATION	DEPOSITIONAL ENVIRONMENT (interpreted)	STROMATOLITES PRESENT
SHEPARD FORMATION	Green and greenish gray argillite, dolomitic argillite, muddy sandstone, and muddy dolostone. Purcell Lava present locally near the base of the formation.	Shallow subtidal to intertidal	Stromatolites are relatively uncommon. Dolomitic, mound-shaped structures 5-40 cm high.
SNOWSLIP FORMATION	Green and red argillite, sandy argillite, and muddy sandstone, commonly somewhat dolomitic. Subordinate muddy dolostone.	Largely intertidal. Parts shallow subtidal	Stromatolites are locally common. Calcitic, mound-shaped structures 5-50 cm high. Formed by in situ carbonate precipitation. Some may be largely nonbiogenic.
HELENA FORMATION	Muddy and sandy dolostone, muddy and sandy dolomitic limestone, muddy and sandy limestone, dolomitic and calcitic argillite, and dolomitic and calcitic sandstone. Rare mudstone.	Subtidal to intertidal. Occasional elevated salinity indicated by halite casts 250-370m above the base of the formation.	Stromatolites are common throughout the formation. Partially silicified dolomitic stromatolites occur 110-180m above the base of the formation. Calcitic and calcitic-dolomitic mound-shaped stromatolites 5-200 cm high are common in the middle and upper part of the formation. Prominent, 24-32 m thick unit composed of Baicalia and Conophyton occurs 200m below the top of the formation.
GRINNELL AND EMPIRE FORMATIONS	Red argillite, sandy argillite, siltite, and muddy sandstone. White quartzose sandstone present and locally abundant. Subordinate green argillite and sandy argillite.	Largely alluvial plain with deposition by sheetflooding or flooding of extremely shallow channels. Portions at margin of lacustrine or marine shoreline.	Two discontinuous stromatlite horizons, 67 and 84m above the base of the formation. Dolomitic, mound-shaped stromatolites 5-20 cm high.
APPEKUNNY FORMATION	Green argillite, sandy argillite, siltite, and muddy fine-grained sandstone. Quartzose sandstone locally forms prominent units. Subordinate red-colored strata and rare black pyritic mudstone.	Largely offshore and below wave base. Portions shallow subtidal to intertidal. Possibly offshore of and marginal to an alluvial plain.	No stromatolites are known from this unit.
ALTYN FORMATION	Sandy dolarenite and silty and clayey dololutite.	Shallow subtidal to intertidal.	Stromatolites are locally common. Dolomitic, Baicalia-like columnar stromatolites and highly elongate, inclined, unbranched columnar stromatolites.

Figure 8: Brief summary of some of the formations and stromatolites in the Belt Supergroup of Glacier National Park. Adapted from Horodyski 1985.

The middle Proterozoic Belt Supergroup of Glacier National Park is a sequence of red and green argillites and siltites, quartzite, and thick carbonate rocks that were thrust over

Cretaceous age shales in the Glacier Park area, but elsewhere unconformably overlie the Archean Wyoming province The Supergroup thickens southwestward to a maximum of 20 km (12.4 mi) in western Montana. It was deposited in either (1) a large intracratonic rift basin, like the East African Rift Zone, or (2) a passive margin, such as that along the present day eastern United States (McGimsey 1985; Moe et al. 1996). The Proterozoic units are named, from oldest to youngest, the Prichard, Altyn, Appekunny, Grinnell, Empire, Helena, Snowslip, Shepard, and Mt. Shields Formations, the Bonner Quartzite, and the McNamara Formation. In either hypothesized depositional environment, the Belt Supergroup was deposited in water, with a variety of near and far shore environments recorded in Glacier's rock formations. The carbonate units record deeper water environments as do fine-grained clay rich rocks. Sandstones and desiccated mudstones record beach and other near shore subaerially exposed deposition sites. The Belt basin was a dynamic system intimately tied to the surrounding tectonic setting.

These Proterozoic rocks are now exposed at the surface because of the deformation associated with the Sevier-Laramide orogeny. The Rocky Mountains of the western United States, including Glacier National Park, were formed during these Cretaceous-Eocene orogenic mountain building events (Ehrlich 1999). The deformation was manifested as a series of roughly north-south trending anticlinal arches and adjacent basins in northwest Montana. In Glacier, the Lewis thrust fault not only underlies the entire park, but is responsible in part for the rugged alpine scenery. The fault initiated intermittent deformation some 200 million years ago and ceased some 15 million years ago. It is responsible for the juxtaposition of older rock over younger rock. Since the end of thrust faulting, several extensional normal faults have offset the Lewis thrust sheet along lines such as the Flathead-Blacktail fault and the Roosevelt fault. These structural features are responsible for deforming the characteristic column of rock at Glacier. Structural features alone however, do not create the fantastical rugged mountainscape of Glacier National Park; glaciers also played a vital role.

During Pleistocene glaciation (1.6 Ma to 10,000 years ago), when large ice sheets covered much of North America, in Glacier Park large alpine glaciers scoured the mountains into peaks, knife-edge ridges, and shallow mountain lakes. Glaciers form when the average snowfall exceeds the rate of average melting. When this happens, snow begins to accumulate from year to year in what is called an *accumulation zone*. The snow turns to ice, and as the snow pack thickens, the ice undergoes plastic deformation (a permanent change in shape of a solid that does not involve failure by rupture) and begins to creep downslope under the force of gravity. No other erosional process on earth is as effective as glacial erosion. As glaciers move downslope, they pluck underlying rocks from the hillside and carry them along in the glacial ice. The boulders act as abrasives to gouge and scour and polish the floor and walls of the valley down which the glacier flows.

Glacier National Park is the only park in the conterminous U.S. Rocky Mountain region with a climate suitable for maintaining substantial glaciers since the end of the Pinedale glacial maximum some 10,000 years ago. This climate exists there for two reasons. One is that the Park is situated far enough north and its mountains are high enough to keep relatively cool in the short summers. The other reason is that the mountains in the Park are high enough to capture significant precipitation from moist Pacific air moving inland. The impressive glaciers that existed in northwestern Montana at the turn of the twentieth century, along with the incredible mountain scenery and wildlife, led to the creation of the Park. When the Park was founded in 1910, the glaciers there were in the process of thinning and retreating or melting back from their inter-glacial maximum extent in about 1850, when they were bigger than they had been in the last 10,000 years (Elias 1996).

At the height of the Pinedale Glaciation, about 20,000 years ago, only the highest ridges and peaks in the Park were free of ice. These ice-free regions are called nunataks; today these appear as horns and arêtes. Mountain glaciers that formed on the western slope of the Continental Divide in the Livingstone Range flowed down the valley of the North Fork and the Flathead River. These glaciers joined with ice coming from the eastern slope of the Whitefish Range to the west, and the combined glaciers flowed south, overriding the lower Apgar Mountains, near West Glacier, and contributing ice to the Flathead Lobe of the much larger Cordilleran Ice Sheet. The Flathead Lobe gouged out a deep depression, and the terminal moraine left by this glacial lobe dammed the Flathead River about 20,000 years before present (B.P.), creating present day Flathead Lake, the largest naturally occurring freshwater lake west of the Mississippi River in the conterminous U.S. (Elias 1996; Karlstrom 2000).

Glaciers that flowed from the western side of the Lewis Range south of Lake McDonald merged with ice from the southeastern flank of the Flathead Range to the south. This ice flow, combined with other mountain glaciers southeast of the Park, formed a large body of ice around the southwestern corner of the Park. This body pushed up and over the low divide at Marias Pass and ended up becoming part of the large piedmont glacier called Two Medicine Glacier. This glacier was also fed by ice flowing from the southern end of the east slope of the Lewis Range. Two Medicine Glacier flowed out over the eastern plains for several kilometers, nearly intersecting with the massive continental ice sheet, or Laurentide Glacier, coming from the east. Farther north along the eastern slope of the Lewis range, mountain glaciers flowed out onto the plains to form the Cut Bank and St. Mary's glaciers. Ice from the northwestern flanks of the Lewis Range and the northeastern flank of the Livingstone Range flowed north in to Canada (Elias 1996; Osborn and Gerloff 1997; Karlstrom 2000).

Surface Stratigraphy at Glacier National Park

Rocks surrounding Glacier range in age from Precambrian, the oldest, to Quaternary, the youngest. In Glacier National Park, however, surface exposures consist primarily of Precambrian Belt Supergroup sedimentary rocks (figure 9). In the very eastern edge of the park, scant exposures of the Cenozoic rocks underlying the Lewis thrust sheet exist. Tertiary-age sediments overlie the Belt rocks. Glacial tills and moraines and fluvial gravels have been deposited during the Quaternary. Soils are currently developing on the present Holocene topography.

Precambrian Era

Prichard Formation. The Prichard Formation forms the lowermost sequence of rock units exposed within the Park. It is patchily exposed in the lowest western valleys of the Park. The best exposures are northwest of Lake McDonald on Stanton Mountain, where the aggregate thickness is about 1219 m (4000 ft). The Prichard was previously assigned to the overlying Appekunny Formation (Whipple et al. 1984). In Glacier National Park, the Prichard is subdivided, informally, into lower and upper parts. The lower part is about 1311 m (4300 ft) thick and characterized by thin, even, parallel laminae of rusty-weathering, blackish-gray argillite (rock metamorphosed from mudstone or shale) and light-gray siltite (rock metamorphosed from silt rich sedimentary rock) that contain disseminated pyrite and pyrrhotite (both iron sulfides). Some small-scale cross-lamination is present in siltite laminae. Carbonate (a sediment formed of the carbonates of calcium, magnesium and/or iron) occurs locally near the top of the lower part as pods and nodules of black manganiferous (contains manganese) limestone (Harrison et al. 1997).

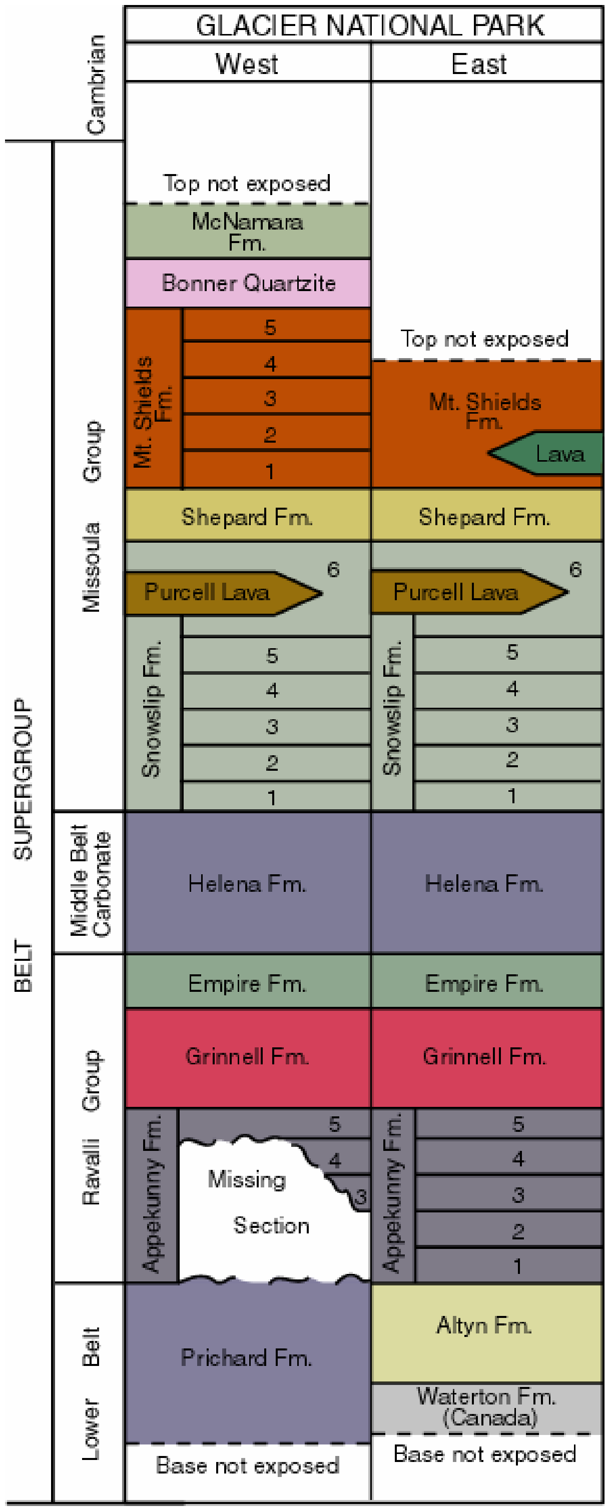


Figure 9: Stratigraphic column of Proterozoic Belt Supergroup rocks for Glacier National Park. See text for explanations of each formation.

The first occurrence upsection of wavy, nonparallel laminae of greenish-gray to medium-gray calcareous siltite marks the upper part of the Prichard Formation. Locally, it contains thin lenticular or lens-like beds of white quartz arenite (rock derived from well-sorted sand-sized fragments with little or no matrix material) and discontinuous beds of fragmental limestone or breccia (rock composed of angular broken rock fragments cemented by a fine-grained matrix) and stromatolitic limestone. This upper part of the Prichard is equivalent to the transition zone of the Prichard as described by Cressman (1989). Its thickness ranges from 244 to 366 m (800 to 1200 ft) (Whipple et al. 1984; Harrison et al. 1997).

Altyn Formation. The Altyn Formation is exposed on the east side of the Park and is the age-equivalent of the Prichard Formation to the west. The formation is completely exposed at Yellow Mountain and northward in the northeast part of the Park. In exposures elsewhere, the base of the Altyn is not exposed and the formation is truncated at the bottom by the Lewis thrust fault. Where the base is present, scant patches of the underlying dolomite-limestone Waterton Formation are exposed in the extreme northeast corner of the Park, nearly in Canada; it is commonly mapped with the Altyn Formation (see Appendix A for Waterton Formation description). In Glacier National Park, the Altyn is informally subdivided into 3 members designated 1 through 3 in ascending order. It consists predominantly of buff-weathering dolomite (calcium-magnesium carbonate), arenaceous dolomite (sandy dolomite), dolomitic limestone, and stromatolitic limestone that commonly form cyclic sedimentation units (Horodyski 1983; Whipple et al. 1984). At Yellow Mountain, the Altyn Formation ranges in thickness from 238 to 256 m (780 to 840 ft) (Whipple et al. 1997). The Lewis thrust cuts upsection in the Altyn Formation to the extent that at Marias Pass, the Altyn is only a few feet thick above the fault (Whipple et al. 1984). The Altyn Formation is not present on the west side of the Park nor in the Whitefish Range. In Canada, the Altyn Formation is underlain by the Waterton Formation and has a maximum thickness of about 1225 ft (Whipple et al. 1984).

Member 1 is mostly thin to thick (1.8 m, 6 ft) beds of yellow- to orange-weathering, dark-gray to black dolomite and thin, lenticular interbeds of fine-grained arenite. Desiccation features, such as cracks, and complete dolomitization (additional magnesium) of carbonate sediments occur in this member (White 1984). Stromatolites are common in the lower part and include subcircular, branched columns and highly elongate columns (Horodyski 1983; Whipple et al. 1997). The member is about 122 m (400 ft) thick (Whipple et al. 1997).

Member 2 is massive, medium- to thick-bedded, white to gray dolomite. Some medium- to coarse-grained, poorly sorted arenite beds are in the upper part. The dolomites and arenites contain quartz and microcline (potassium feldspar mineral) (White 1984). Stromatolites and dark-orange dolomite blebs occur locally in the lower part. The member contains black asphalt-like veinlets near the Lewis thrust fault. The thickness of member 2 ranges form 58 to 69 m (190 to 225 ft) (Whipple et al. 1997).

Member 3 ranges from 58 to 61 m (180 to 200 ft) thick and consists of interbedded and interlaminated light gray to brownish-yellow dolomite, dolarenite (dolomite rich arenite), and arenite. Dolomite beds are 2.5 - 15.2 cm (1-6 in) thick; arenite beds are medium to coarse-grained and commonly cross-bedded with some herringbone lamination (cross-bedding resembles the rib bones of a fish). Stromatolites and stylolites, or the irregular surface formed when rocks are pushed together, are common (Whipple et al. 1997). Member 3 contain mud-cracked shales and terrigeneous (derived from a continent) arenite (White 1984). The Appekunny Formation, and a transition zone of interbedded arenite overlies the Altyn conformably and greenish-gray siltite between the two is placed in the uppermost part of the Altyn Formation (Whipple et al. 1984).

Appekunny Formation. Present stratigraphically all over Glacier National Park, the Appekunny Formation (also known as the Appekunny Argillite) is approximately 671 m (2200 ft) thick, predominantly green-colored, and fine-grained. Stromatolites are not known from the Appekunny (Horodyski 1983; Whipple et al. 1984). In the eastern part of the park, the Appekunny conformably overlies the Altyn Formation. On the west side of the Park, only 91 m (300 ft) of the uppermost Appekunny is present and it unconformably overlies the Prichard Formation. The type locality of the Appekunny Formation is near Many Glacier (Whipple et al. 1984). It is informally subdivided into five members designated 1 through 5 in ascending order (Whipple et al. 1997).

Member 1 contains alternating successions of interlaminated pale-maroon siltite and minor argillite with grayish-green siltite and minor argillite. Bed lamination is even, parallel to nonparallel and curved nonparallel; some beds show broad, low-angle hummocky cross-lamination and small-scale, source-and-fill structures but lack shallow water sedimentary structures. A quartz arenite interval forms a key marker about 55 m (180 ft) above the base of member 1. This interval things gradually northward from about 24 m (80 ft) at Elk Mountain (at south end of Park) to 15 m (50 ft) at Bear Mountain (near U.S.-Canada boundary). Member is about 137 m (440 ft) thick (Whipple et al. 1984; Whipple et al. 1997).

Member 2 closely resembles member 1 except maroon beds are absent and laminae are generally thinner in member 1 than in member 2. Thin arenite beds, 2.5-7.5 cm (0.8-2.3 in) thick, are common in the lower part (Horodyski 1983). The lower contact is placed on top of the uppermost maroon sequence of member 1 and generally coincides with an increase in thickness of siltite laminae in member 2. In areas where maroon beds are absent, the contact between members 1 and 2 may be indistinguishable. Small clots and thin discontinuous laminae of pink ferroan (iron rich) calcite are common in some beds (Whipple et al. 1997). Member 2 is about 168 m (550 ft) thick (Whipple et al. 1984).

Member 3 is about 165 m (540 ft) thick and characterized by interlaminated and interbedded grayish-green siltite, yellowish-brown pyretic (containing iron sulfide) arenite and lesser amounts of grayish-green argillite. Subaqueous shrinkage cracks, load structures (a result of deformation within unsolidified sediments where layers bend and intrude the layers under them), and mud-chip breccia are common and lamination is wavy, nonparallel and arenite beds typically contain pyrite (Whipple et al. 1984). The lower contact is placed at the base of the lowermost bed of pyritic arenite where the pyritic arenite and overlying beds are wavy laminated and contain numerous shallow-water sedimentary structures (Whipple et al. 1997).

Member 4 is poorly exposed because outcrops are mostly cleaved and easily weathered. The member consists of thin to very thin laminae of olive colored siltite and thin lenticular beds of rusty-brown arenite that are commonly stained by iron and manganese oxides. Member 4 is about 137 m (450 ft) thick and is commonly cleaved, folded and sheared by minor thrust faults (Whipple et al. 1984; Whipple et al. 1997).

Member 5, the uppermost member of the Appekunny Formation, contrasts sharply with member 4 and consists of bright-green argillite and lesser amounts of siltite. Lamination is wavy, within nonparallel, fining-upward silt to clay layers. The upper part of these sequences also contain ripple cross-laminated sandy layers intercalated (a juxtaposition of layers with strikingly different types of layers) with silty and pelitic (continental derived mud rich) layers (Horodyski 1983). Mud-chip breccias, fluid-escape structures (resulting structure from water being squeezed from sediment, often resembles spidery lines), and dolomite-filled subaqueous shrinkage cracks are common. Member 5 is about 59 m (195 ft) thick (Whipple et al. 1997).

Grinnell Formation. The Grinnell Formation conformably overlies the Appekunny, and conformably underlies the Empire formation in Glacier National Park. The Appekunny – Grinnell contact was placed where bright green mudcracked argillite and finely coupleted argillite of Appekunny member 5 passes upward to dominantly mudcracked red argillite of the Grinnell. The sharp contact between the Grinnell and Empire formations was recorded as a clean quartz arenite bed, stained bright orange by limonite (brown iron oxide mineral) altering from oolitic pyrite (ooids form from a central nucleus with concentric accretionary layers, resemble fish eggs) comprising the base of the Empire Formation (Kuhn 1987). The Grinnell is 530-790 m (1740-2590 ft) thick in Glacier (Link 1997). Spectacular outcrops of the Grinnell Formation exist in the Two Medicine and Many Glacier areas. In the type area, on the east side of Glacier National Park, the Grinnell is composed of quartzose red-bed (iron rich layered rock) sequences and is about 671 m (2200 ft) thick. The Grinnell on the west side of the Park, north of Lake McDonald, is also about 671 m (2200 ft) thick and contrasts strongly with that in the type area to the east, principally because west-side exposures contain less quartz arenite and have less intense coloration (Whipple et al. 1984; Kuhn 1987). Generally, the Grinnell Formation is a terrigenous unit of argillite, sandy argillite, very fine- to fine-grained muddy sandstone and fine- to very coarse-grained quartzose sandstone and quartzarenite. It is predominantly red-colored, although purple to green-colored strata and white to brown, hematitic (iron oxide rich), cross-bedded beds are present locally (Horodyski 1983; Kuhn 1987).

Willis (1902) first defined the Grinnell Formation in Glacier National Park as follows:

> "A mass of red rocks of predominantly shalely argillaceous character is termed the Grinnell argillite from its characteristic occurrence with a thickness of about 1800 ft in Mt. Grinnell."

It is commonly recognized as continuous red outcrops along the eastern edge of the Lewis overthrust. Detailed stratigraphic study of the Grinnell Formation reveals 5 omnipresent informal units, or 6 informal units, including a transitional unit, present to some extent, with the overlying Empire Formation, from bottom to top: (1) a basal green argillite unit; (2) a lower red and green argillite unit; (3) a middle red argillite unit; (4) a middle green and red argillite unit; (5) an upper sandy unit; and (6) an upper unit transitional with the overlying formation (Horodyski 1983; Link 1997).

A dominant sediment characterizes each unit. Unit 1 (0 – 81 m, 0 – 265 ft) is principally green and minor red argillite consisting of desiccated mud, microlamina, silt to clay beds, and very fine- to fine-grained sand sediments. Scattered beds of medium- to coarse-grained sand sediment occur near the middle, and increase upsection. Most of the sediment in unit 1 is uncracked (Link 1997). Rare stromatolites occur near the top of Unit 1 where the medium- to coarse-grained sand becomes more common and contains tabular, planar, and tangential cross-beds (figure 10). The stromatolites are described as mound shaped dolomitic stromatolites, 5 – 20 cm (1.9 - 7.9 in) high and 10 – 50 cm (3.9 - 19.7 in)wide and are known from a single horizon in the lower Grinnell Argillite along Going-to-the-Sun Road and from two horizons on Mt. Henkel (Horodyski 1983; Kuhn 1987).

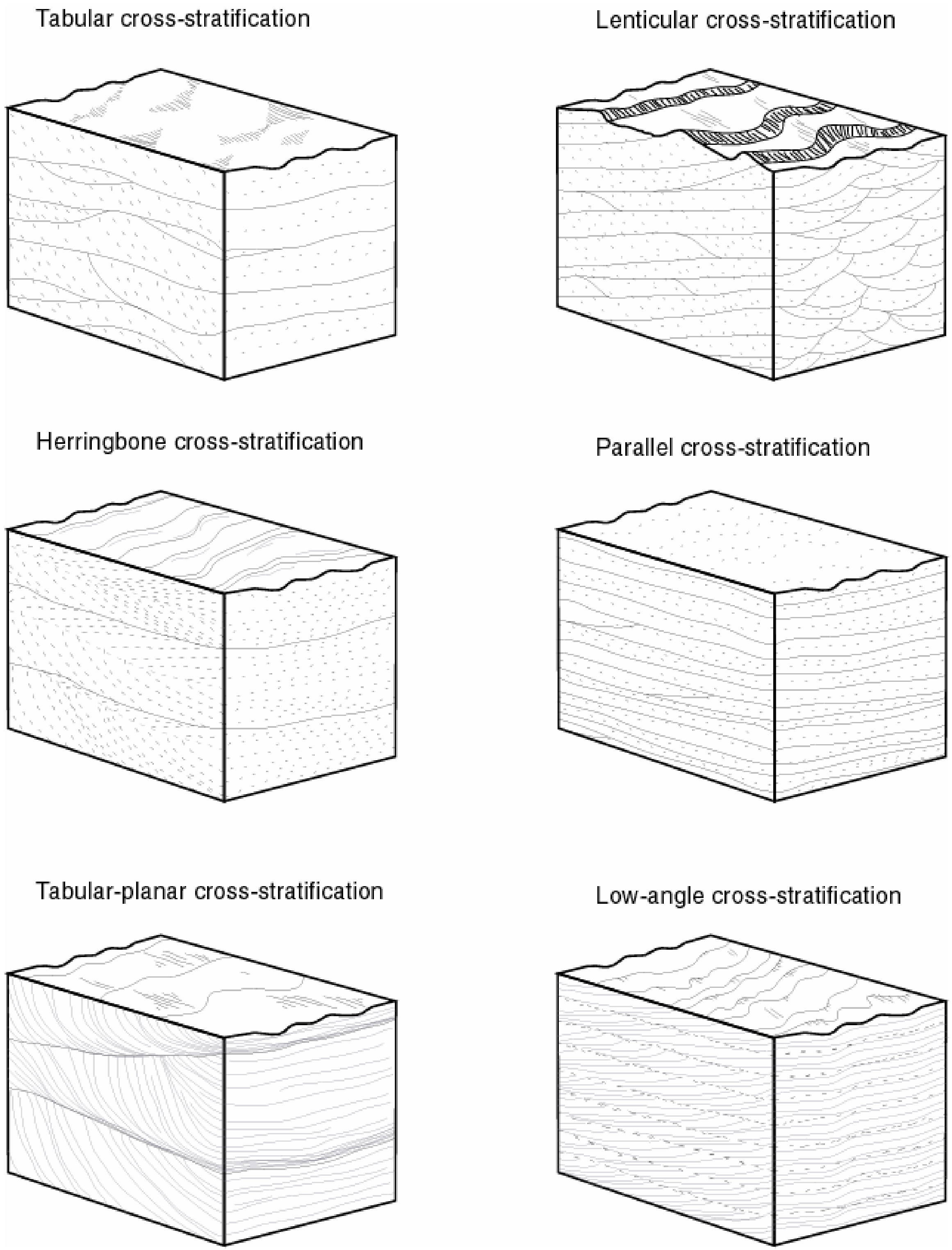


Figure10: Different kinds of sedimentary cross-stratification

Unit 2 (81 – 123 m, 265 - 405 ft) contains an abundance of different types of sediments: medium- to coarse-grained sand, very fine- to fine-grained sand, flat-laminated silt, desiccated mud, silt to clay beds, and microlamina. However stacked beds forming a westward thinning wedge of medium- to coarse-grained, mudchip-rich sand (intraformational conglomerate) with interbeds of the desiccated mud, dominate this unit and occur in sequences up to 10 ft thick (Kuhn 1987; Link 1997). The highest thick bed of medium- to coarse- grained sand marks the top of Unit 2. Although very fine- to fine-grained sand is generally more thinly bedded (2 – 10 cm) than medium- to coarse-grained sand, both sizes occur in sets capped by straight-crested to sinuous-crested symmetric ripples containing unidirectional cross-beds. These are mostly capped by desiccated mud drapes. Abundant mudchips are more common on medium- to coarse-grained sand than in the finer sand in Unit 2 (figure 11). Accreted mudballs (analogous to ooids where layers form concentrically around a nucleus, in this case, a mud chip) were absent from sands in this unit (Kuhn 1987).

Figure 11: Mudchip conglomerate in the Grinnell Formation. Note edge of knife for scale. Photo from Dawes and Dawes 2001, for more information, see: http://wvcweb.ctc.edu/rdawes/VirtualFieldSites/GrinnellGlacier/VFSGrinnell.html

Unit 3 (123 – 268 m, 405 - 880 ft) is easily distinguished as a bright red argillite characterized by desiccated mud sediment. Silt to clay beds and microlaminae are relatively rare in this unit. Thin lenses and indistinct stringers of coarse- to very coarse sand occur in the desiccated mud as mud-supported grains (contained in a muddy matrix) and, many are discernable by brownish-orange iron reduction spots along the outcrop (Link 1997). Beds of angular to rounded mudchips, mudcracks and fluid-escape structures are common throughout Unit 3. Flat-laminated silt contains load casts and climbing ripples and occurs in beds up to 15 cm (5.9 in)thick in the lower portion of this unit (123 – 174 m, 405 – 572 ft), but is rare in the upper portion (174 – 268 m, 572 - 880 ft), where desiccated mud predominates (Kuhn 1987).

Unit 4 (244 – 390 m, 800-1280 ft) is characterized by abundant beds of the green flat-laminated siltite (Link 1997). The lateral continuity, and general thickness (30 – 90 cm, 11.8 - 39.4 in) of abundant silt beds, separates this interval as a distinct unit. Interbedded with the silt layering is mostly desiccated mud, microlaminae, and silt to clay beds (figure 12). Load structures and climbing ripple are common in beds of the flat-laminated silt sediment, and beds are sharply separated by mudcracked clay drapes or thin beds of the desiccated mud. Occasional thin beds of the medium- to coarse-grained sand increase upward in the section. Thin (1 – 10 cm, 0.4-3.9 in) beds of very fine- to fine-grained sand are also common in this unit and are capped by straight-crested, symmetric unidirectional ripples with clay drapes (Kuhn 1987).

Figure 12: Desiccation features in mud layer of the Grinnell Formation. Note knife for scale. Photo from Dawes and Dawes 2001.

Unit 5 (390 – 555 m, 1280 – 1822 ft) contains a variety of sediments, but medium- to coarse-grained sand characterizes it as a distinctly separate unit. It contains mainly white sandstone (up to 20 percent of the section, locally) and red and green argillite and siltite (Whipple et al. 1984; Link 1997). Stacked, mudchip-rich, medium- to coarse-grained sand beds are similar in character to those of Unit 2. However, mudchips and accreted mudballs (up to 10 cm (3.9 in) in diameter) increase noticeably in abundance in unit 5. Individual sand beds are thicker (up to 60 cm, 23.4 in) and occur in more thickly stacked intervals than those in Unit 2. The uniform coarseness of some mudchip-rich sand beds obscures internal stratification, causing some beds to appear massive (lacking sedimentary structures). However, flat-laminated, mudchip-rich sets commonly pass upward into abundant tabular-planar, tangential and trough cross-bedded sets (see figure 10). Foresets average 25 – 35 degrees, and are easily distinguished by heavy mineral laminae and by oxidation or reduction of mudchips and thin clay drapes. Reactivation surfaces and bimodal cross-beds occur in some cosets (Kuhn 1987). Most ripples are symmetrical (Link 1997).

Empire Formation. The Empire Formation consists primarily of interlaminated grayish green argillite and siltite arranged commonly as fining-upward sequences in less than 3 cm (1 in) thick beds (Whipple et al. 1985). Subordinate maroon argillite, buff and green, locally dolomitic siltite, and light-gray quartz arenite are also present locally (Whipple et al. 1984; Winston and Lyons 1997). Carbonate cement is present in the uppermost part, and thin quartz arenite beds are common in the lower part of the formation (Whipple et al. 1985). The formation is 122-158 m (400-518 ft) thick in Glacier National Park (Winston and Lyons 1997).

In much of northern Glacier National Park, the base of the Empire is placed at a 20 to 50 cm (7.9 to 19.7 in) orange-weathering, cross-bedded, medium-grained quartz arenite bed interbedded with plane-laminated silt and clay above the distinguishing red argillite of the Grinnell Formation. This arenite bed contains distinctive sulfide ooids. The ooids contain concentric layers of pyrite and chalcopyrite (iron sulfide minerals) (Winston and Lyons 1997).

The lower 18 m (60 ft) of the Empire formation is dominated by beds of white to buff quartz arenite that range in thickness form 13 cm to 3.5 m (5 in to 11 ft) and contain minor carbonate cement and pyrite. In this interval, most arenite beds are well sorted, but between beds, grain size ranges from fine to coarse (Whipple et al. 1984). These are arranged in crudely fining-upward siliciclastic (sediments almost entirely composed of silica bearing minerals) cycles. Some arenite beds at the bases of cycles have well-developed cross-bedding, load structures and asymmetrical ripple marks (Winston and Lyons 1997). Arenite beds decrease in number and thickness from bottom to top of the formation (Whipple et al. 1984).

The upper two-thirds of the formation contains primarily olive-green and a few purplish-red beds of argillite a few cm to 1.5 m (1 in to 5 ft) thick. Thin interbeds of dolomite are present near the middle of the formation and increase in number and thickness upward in the section (Whipple et al. 1984). This upper part of the Empire contains mixed siliciclastic-to-carbonate-cycles 1 to 2 m (3 to 6 ft) thick, that start with a coarse sand bed, and become finer grained and more carbonate-rich upward, to laminated argillite and finally molar-tooth calcite structures (described just below) and intraclastic beds of dolomite (Winston and Lyons 1997). The upper contact of the Empire Formation is placed on top of a roughly 6 ft thick interval of green argillite that is overlain by the thick dolomite beds of the Helena Formation (carbonate) (Whipple et al. 1984).

> "Origin of the strange calcitic "molar-tooth" structure has been queried ever since Bauerman (1885) first likened the crenulated ribbons to "markings in the molar tooth of an elephant". More recent interpretations have ranged from biotic to replacement of evaporite minerals to filled open-space structures such as desiccation cracks, syneresis cracks, gas bubble and fluid voids to earthquake dewatering structures. Molar-tooth structure ranges from vertical ribbons, horizontal ribbons, blobs and pods. The most accepted recent theory for molar-tooth structure formation involves a process whereby gas generated in the sediments below the depositional interface rises through the still wet sediment and is sometimes trapped to form irregular cavities that are then filled with calcite precipitate. The gas may have been H_2S, generated by sulfate reducing bacteria or methane formed by organic decay, or probably both." Winston and Lyons (Belt III)

Helena Formation. In Glacier National Park, the Helena Formation makes up the Middle Belt carbonate. Formerly, the strata of the Empire and Helena Formations in Glacier National Park were referred to collectively as the Siyeh Limestone. The Helena Formation consists of the middle and upper Siyeh Limestones of Horodyski (1983) (Winston and Lyons 1997). The Helena Formation in Glacier National Park is about 750 to 1030 m (2460 - 3380 ft) thick and contains dolomite, limestone and minor quartz arenite (Whipple et al. 1985). The base of the Helena Formation is placed at the first bed of dolomite with molar-tooth calcite structure above a 1.8 m (6 ft) interval of green argillite in the Empire Formation (Winston and Lyons 1997).

The Helena Formation can be divided into seven informal lithologic units based on carbonate content, presence or absence of "molar-tooth" structure (described above), presence or absence of oolitic limestone and the type of stromatolites present (Horodyski

1983). Near Going-to-the-Sun Road, the Helena Formation is divided into just three distinct parts. For simplicity, these parts will be described below. The lower part is 179.8 m (590 ft) thick and consists of interbedded quartz arenite and thin-bedded dolomite near the base; thin beds of horizontally laminated and "molar tooth" dolomite in the middle; and thick smoky-gray limestone beds near the top. "Molar tooth" structures were defined by Bauerman (1885) as vertical to subhorizontal, wrinkled segregations of massive calcite that weather to a greater degree than the host rock, and when viewed on weathered surfaces, they display a microtopographic expression that resembles the molar teeth of elephants (Whipple et al. 1984).

The middle part of the Helena, which is 548.6 m (1800 ft) thick, is dominated by dolomitic molar-tooth beds, some as much as 30.5 m (100 ft) thick. A few thin beds of quartz arenite and stromatolitic limestone are present in the middle part of the Helena. The upper part consists primarily of interbedded stromatolitic limestone, dolomite, oolitic limestone, and quartz arenite. At the base of this upper part is an interval of stromatolitic limestone about 30.5 (100 ft) thick, known as the Conophyton zone, which is composed of Baicalia-Conophyton (separate species) stromatolite cycles. The massive character of the Conophyton zone caused it to stand in relief in most sections of the Helena formation in the Park. Because of its significance as a unique marker bed, the Conophyton zone will be described in more detail below. A diorite sill (igneous rock of intermediate quartz or silica content), 39.6 m (130 ft) thick, intrudes the Helena in this part of the measured section just above the Conophyton zone (figure 13). This sill, which is present throughout the Park, varies widely in stratigraphic position, from near the base of the Helena in the southeast part of the Park, to the lower part of the Snowslip in the northernmost part of the Park (Whipple et al. 1984).

Figure 13: Photograph of prominent diorite sill pervasive in the Helena Formation of Glacier National Park. Location is within the Grinnell Glacier cirque. Modified photo by Dawes and Dawes 2001. For more information see: http://wvcweb.ctc.edu/rdawes/VirtualFieldSites/GrinnellGlacier/VFSGrinnell.html

Helena Formation – Conophyton-Baicalia Stromatolite Cycles. Of all the Belt strata in Glacier National Park, stromatolites are most abundant in the Helena Formation (Horodyski 1983). The Conophyton – Baicalia cycles present in the Helena Formation are unique to the Belt Supergroup of the Glacier National Park area, but are similar to parts of stromatolite cycles that are much more extensive and repetitious in the Siberian platform in Russia. It comprises a cliff-forming unit which forms a distinct biostratigraphic marker bed (Raup 1997). This stromatolitic unit is by far the most prominent stromatolitic unit in the Belt Supergroup, ranging from 24 to 32 m (79 - 105 ft) thick in Glacier National Park and consisting of a variety of branched columnar and conical stromatolites. In the central part of Glacier National Park the Baicalia-Conophyton stromatolite cycles can be divided into six units (Figure 14) (Horodyski 1983; Whipple et al. 1985).

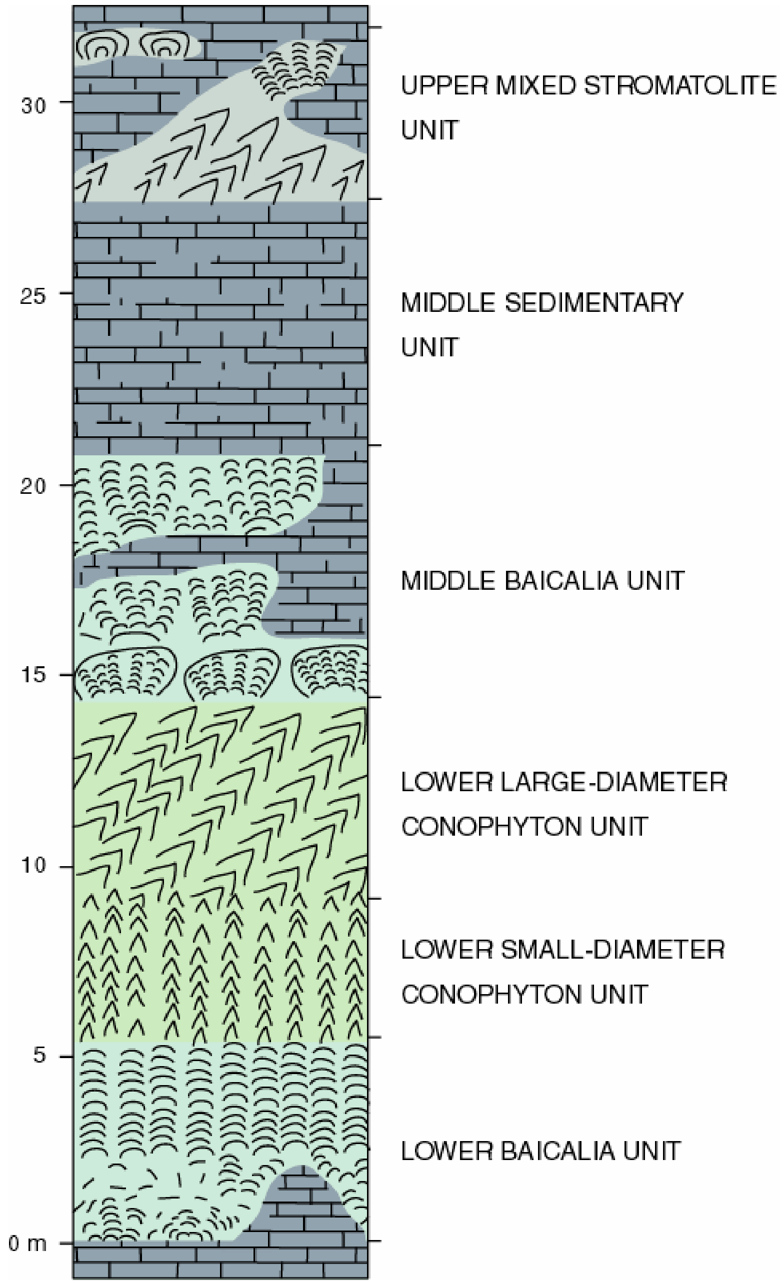


Figure 14: Diagrammatic representation of the Conophyton-Baicalia stromatolite cycles in central Glacier National Park, adapted from Horodyski (1985).

A conspicuous interval, 0.9 - 1.8 m (3 - 6 ft) thick, of pyritic, blackish-green mudstone at the base of the Conophyton Zone occurs nearly everywhere the stromatolite zone is exposed in Glacier National Park. The fine-grained mudstone is not noticeably laminated and breaks along conchoidal fractures (scalloped shaped breaks analogous to a windshield chip). Geochemical investigations imply that the interval has a volcanic component and may be mostly a metatuff or the metamorphic equivalent of a volcanic ash deposit (Raup 1997). Overlying this basal interval, the first unit, or lower Baicalia unit consists of 3 to 15 cm (1.2 - 5.9 in) diameter, branched columnar stromatolites of the group Baicalia in the lower part of the unit and 10 to 30 cm (3.9 - 11.8 in) diameter, parallel branched and unbranched columnar stromatolites in the upper part of the unit (Horodyski 1985; Whipple et al. 1985). Interbedded sediments, including silty and sandy dolomitic limestone, micritic or very fine-grained limestone, and eroded stromatolite laminae comprise 20 to 60 percent of the lower part of this unit and 5 to 10 percent of the upper part of the unit (Horodyski 1983). Poorly developed Conophyton are locally present in the upper part of the unit and grade laterally into convexly laminated columnar stromatolites (Whipple et al. 1985).

The second unit, herein called the lower small-diameter Conophyton unit, consists of 3 to 15 cm (1.2 - 5.9 in) diameter, conically laminated stromatolites of the group Conophyton (Horodyski 1985). They typically are oriented almost vertically, and rarely, are inclined up to 30 degrees. These stromatolites also commonly form 5 to 20 cm (2 - 7.9 in) diameter bioherms, which are separated by 1 to 2 m (3.3 - 6.6 ft). Although fragments of stromatolitic laminae occur in the second unit, they are not abundant and do not form the stromatolitic debris beds that are common in other parts of these cycles, such as in the first unit (Whipple et al. 1985).

The third unit, or the lower large-diameter Conophyton unit, has a gradational contact with the underlying small-diameter Conophyton unit, and consists of 10 to 60 cm (3.9 - 23.4 in) diameter Conophyton, which typically are inclined 30 to 60 degrees from the vertical. These columns have a uniform orientation throughout Glacier National Park (Horodyski 1983; Whipple et al. 1985). The axes of the columns trend southwest to south-southwest and plunge to the southwest. Some of the columns have a lanceolate shape, with the convex projection typically occurring on the lower portion of the column. Other columns range from circular to irregularly stellate or star-shaped in transverse section. These columns form bioherms (mounds of biologically derived rock material contained within non organic rock) 3 to 20 m (9.8 - 65.6 ft) across, which are separated by 1 to 2 m (3.3 - 6.6 ft) and occasionally as much as 10 m (33.3 ft) (Whipple et al. 1985).

The middle Baicalia fourth unit is primarily composed of branched columnar stromatolites and associated sediments. At the base of this unit are 1 to 1.5 m (3.3 - 4.9 ft) high and 2 to 8 m (6.6 - 24.2 ft) wide bioherms composed of very closely spaced, generally unbranched or parallel-branched columns. These bioherms are particularly distinctive within the unit and can be correlated for more than 80 km (48 mi) along strike. Black mudstone and calcareous mudstone, 10 to 50 cm (3.9 - 5.9 in) thick, underlies these unique bioherms. The remainder of the middle Baicalia unit consists of branched columnar stromatolites of the group Baicalia, mound-shaped stromatolites, and interbedded sediment (Whipple et al. 1985).

The next, fifth unit, called the middle sedimentary unit, consists of silty and sandy dolomitic limestone and micritic limestone. "Molar-tooth" calcite structure is common in the lower part of this unit, and small, mound-shaped stromatolites are of rare occurrence. The last unit of the Baicalia – Conophyton cycles in the Helena Formation is the upper mixed stromatolite unit (Horodyski 1985). It consists of Conophyton, Baicalia, a variety of mound-shaped stromatolites, and interbedded sedimentary layers. Conophyton occurs as columns 5 to 50 cm (2 – 19.7 in) in diameter, generally forming bioherms 2 to 20 m (6.6 - 66.6 ft) across. The larger diameter Conophyton typically are inclined 45 to 75 degrees from the vertical (Horodyski 1983). Branched columnar stromatolites of the group Baicalia and mound-shaped stromatolites are common in the uppermost part of this unit. Silty and sandy dolomitic limestone and micritic limestone are also common in this last unit. Eroded stromatolite debris is commonly associated with Baicalia and mound-shaped stromatolite beds, but is rarely associated with the Conophyton bioherms (Whipple et al. 1985). Reasons for this have yet to be stipulated.

Snowslip Formation. The Snowslip Formation overlies the Helena Formation in Glacier National Park and in the southern part of the Whitefish Range to the west (Ackman 1988). The contact with the underlying Helena Formation is represented by a sharp break from the carbonate shoals of the Helena to the siliciclastic mudflats of the Snowslip Formation. The Helena-Snowslip contact marks the boundary between the middle Belt carbonate and the Missoula Group of the Belt Supergroup (Whipple et al. 1997). Lithofacies or units of the Snowslip Formation correlate reasonably well between the east and west stratigraphic sequences in Glacier National Park (see figure 9). The formation is dominantly terrigenous sequences of green and red argillite, dolomitic argillite, and muddy sandstone (Horodyski 1983). The Snowslip varies in thickness from 489.5 m (1606 ft) at the type section at Snowslip Mountain in the southern part of the Park to 357.2 m (1172 ft) at Hole-in-the-Wall, in the north, near the International Boundary (Whipple et al. 1984). In the northern part of the Park, the upper part of the Snowslip encloses the Purcell Lava, a series of thin mafic (or low silica) volcanic flow units, which due to its uniqueness to the area, will be described in detail below. The Snowslip is subdivided into six informal lithostratigraphic units, not including the Purcell Lava. From bottom to top, oldest to youngest, the units are labeled 1 to 6 (Whipple and Johnson 1988).

Interbedded calcareous siltite and argillite are the principal lithologies of member 1. The lower contact is conformably placed between the uppermost bed of pink to gray oolitic limestone of the Helena Formation and the lowermost bed of poorly sorted calcareous cross-laminated arenite of member 1 (Whipple et al. 1984; Whipple and Johnson 1988). They are typically arranged as wavy nonparallel, fining-upward layers less than 10 mm (0.4 in) thick, and they vary in color from reddish hues to grayish-green. The member is characterized by thin (less than 3 cm (1.2 in) thick) lenticular beds of poorly sorted, fine- to coarse-grained calcareous arenite. Clasts are dominantly rock fragments, quartz, and oolites, many of which are partially replaced and coated by hematite, lending an orange hue to the beds (Whipple and Johnson 1988). Arenite beds are commonly light gray and weather to tan. Mottling is common among the red and green strata to the extent that rock color changes along strike. Sedimentary structures include syneresis and desiccation cracks, numerous thin laminae of mud-chip breccia, fluid-escape structures, some flaser lamination (streaky, parallel layers wrapping around larger clasts), and some small-scale cross-lamination in oolite-rich arenite beds. Member 1 is 24 to 94 m (80 to 300 ft) thick (average 25.91 m thick) (Whipple et al. 1984; Whipple and Johnson 1988).

Member 2 is mostly composed of greenish-gray siltite and olive green argillite beds (Whipple et al. 1984). The siltite fraction is commonly two-thirds the bed thickness. Individual siltite-argillite sets are generally 10-20 mm (0.4 - 0.8 in) thick. Outcrops, which generally weather recessively, are poorly exposed and are yellowish gray or brown due to local calcite and dolomite cements (Whipple and Johnson 1988). Interbedded in member 2 are layers of pink, red, and cream-colored stromatolitic limestone as much as 15 cm (5.9 in) thick, particularly in the lower portion of the member (Horodyski 1983; Whipple et al. 1985). The distinct reddish color is the result of several laminae that contain fine-grained hematite. Commonly associated with the stromatolites are beds of pyritic flat-pebble conglomerate and poorly sorted lithic arenite, some as thick as 10 cm (4 in) (Whipple et al. 1984). Clasts are mostly rock fragments (including limestone), quartz, and minor feldspar that are cemented by calcite and dolomite. Ripple cross-lamination occurs in most beds of arenite. Member 2 is 69.85 m (229 ft)thick. The lower contact is placed at the base of a stromatolitic limestone, which is underlain by reddish strata of member 1 (Whipple and Johnson 1988).

Member 3 consists of interbeds of arenite, siltite, and argillite that are arranged locally as fining-upward sequences as much as 3 m (10 ft) thick (Whipple et al. 1997). Arenite beds are white to greenish-gray, less than 10 cm (4 in) thick, lenticular, and have irregular bases. Arenite grains are fine to medium sized, moderately sorted, and mostly composed of quartz, feldspar, and some rock fragments. Siltite and argillite occur in fining-upward sequences. Beds of siltite and argillite are greenish-gray and locally dolomitic, similar to beds in member 2. Mud-chip breccias and mud chip fragments aligned on foresets of cross-laminated coarser-grained arenite beds are present locally, particularly at the base of the siltite and argillite fining-upward sequences. Mud-draped ripples (usually formed when mud settles out of suspension over preexisting ripples), shrinkage cracks, and fluid-escape structures are also common locally. Member 3 is 16.6 m (54.5 ft) thick. The lower contact is placed between relatively calcareous, greenish-gray rocks of member 2 and beds of quartz and subfeldspathic arenite if member 3 (Whipple and Johnson 1988).

Member 4 is similar to member 2 in all respects (Whipple et al. 1997). The dominant lithologies of member 4, as in member 2, are greenish-gray siltite and argillite that are arranged as thin, wavy fining-upward sequences as much as 25 cm (5.9 in) thick. Argillite composes 80 percent of the interbedded and interlaminated lithologies in the lower part of the member but decreases to about 25 percent near the top (Whipple and Johnson 198). Stromatolitic limestone, pyrite, carbonate cement, and iron limonitic stain occur locally (Horodyski 1983). In the upper part of the member, thin lenticular beds of very fine-grained, calcareous arenite and gray dolomite are present. Syneresis cracks, fluid-escape structures, and mud-chip breccias are abundant locally. Arenite beds are typically ripple cross-laminated. Member 4 is 79.6 m (261 ft) thick. The lower contact is placed on top of the uppermost reddish argillite of member 3 (Whipple and Johnson 1988).

Member 5 is the most distinct subdivision of the Snowslip. It is resistant to weathering and forms prominent outcrops (Whipple and Johnson 1988). Beds of white, pink, and greenish-gray arenite are overlain successively by reddish beds of interlaminated arenite and siltite, siltite and argillite, and finally argillite; this arrangement of lithologies forms distinct fining-upward successions that range in thickness from a few centimeters to as much as 3.5 m (11.5 ft). The base of each succession is irregular (Whipple et al. 1997). Arenite beds are locally calcareous and commonly lenticular, and, in order of abundance, are composed of quartz, rock fragments, and feldspar. Siltite beds are grayish-red; argillite beds are reddish-purple and have very thin, wavy discontinuous laminae. Sedimentary structures present in member 5 include a wide variety of ripple marks, raindrop impressions, wrinkle marks, desiccation cracks, mud-draped ripples, mud-chip breccias, and some fluid escape structures. Member 5 is 39.82 m (130.6 ft) thick. The lower contact between member 5 and member 4 of the Snowslip Formation is placed at the base of the lowest bed of pinkish arenite where that bed and overlying reddish strata form a mixed fining-upward sequence (Whipple and Johnson 1988).

The uppermost member, 6, encloses facies of the Purcell Lava (described below) and is compositionally similar to members 2 and 4 (Whipple and Johnson 1988). It mainly consists of greenish-gray siltite and minor argillite couplets that are less than 1 cm thick as well as pale-maroon arenite (Whipple et al. 1997). Thin lenticular beds of arenite (less than 10 cm (3.9 in) thick), stromatolitic limestone, and thinly laminated dark-green argillite are present below the lava. Similar beds occur above the lava, but many are grayish-red to reddish-purple (Horodyski 1983; Whipple et al. 1997). One laterally continuous bed of red siliceous arenite just below a minor sill contains pyrite and chalcopyrite, though not in any significant quantity. Sedimentary structures in the member include ripple cross-lamination in arenite beds, syneresis cracks, fluid-escape structures, and mud-chip breccias (Whipple and Johnson 1988). Although the lower contact of the Purcell Lava within member 6 is very irregular (probably due to loading by the deposition of the basalt), no evidence of an unconformity was observed. Where the Purcell is present, thin discontinuous beds of pink and gray stromatolitic limestone occur in the lower part of member 6 (Horodyski 1983; Whipple et al. 1984; Whipple et al. 1985). Strata of member 6 appear to have been deposited conformably onto inherent surface irregularities of the lava, and so is interpreted to enclose the lava conformably (Whipple et al. 1997; Whipple and Johnson 1988).

Upper and lower contacts of member 6 are conformable. The contact of member 6 with the Shepard Formation is placed at the top of the uppermost bed of argillite, where overlying beds of the Shepard consist of yellowish-gray weathering beds of calcareous siltite and argillite, ripple cross-laminated dolarenite, and limestone. The total measured thickness of member 6, the Purcell Lava (99.6 m, 326.8 ft) and the diabase sill is 125.72 m (401.4 ft) at the Hole-in-the-Wall reference section in northern Glacier National Park (Whipple and Johnson 1988).

Snowslip Formation – The Purcell Lava. The Purcell Lava is a sequence of mafic lava flows that forms a persistent and important marker bed in the lower Missoula Group of the Belt Supergroup in both the northern United States and Canada (Whipple et al. 1984; Whipple et al. 1997). In Glacier National Park, it is within member six of the Snowslip Formation and is best exposed near the crests of the Livingston and Lewis Ranges about 15.5 m (50 ft) below the top of the formation (Whipple et al. 1984). The Purcell Lava thickness in Glacier National Park ranges from 77 m (253 ft) near the Canadian border to 15 m (50 ft) at the most southern locality in the Apgar Mountains, above Lake McDonald (Whipple et al. 1997). In general, the flow thickens to the north and west.

In Glacier National Park, the Purcell Lava consists generally of fine-grained, vesicular, medium bluish-gray to greenish-gray altered basalt. Vesicles or bubbles in the rock, commonly contain fillings of quartz, chlorite (a water, iron and magnesium rich alteration mineral), calcite, or various combinations of the three. On the basis of field characteristics and occurrences, the Purcell can be subdivided into four units: (1) subaqueous pillow lava, (2) compound pahoehoe flow, analogous to the lava “rivers” in Hawaii, (3) vent facies rocks, deposited at or near the lava source, and (4) a hypabyssal (meaning too deep to be extrusive and too shallowly emplaced to be considered plutonic) diabase sill (figure 15) (Whipple et al. 1984; McGimsey 1985).

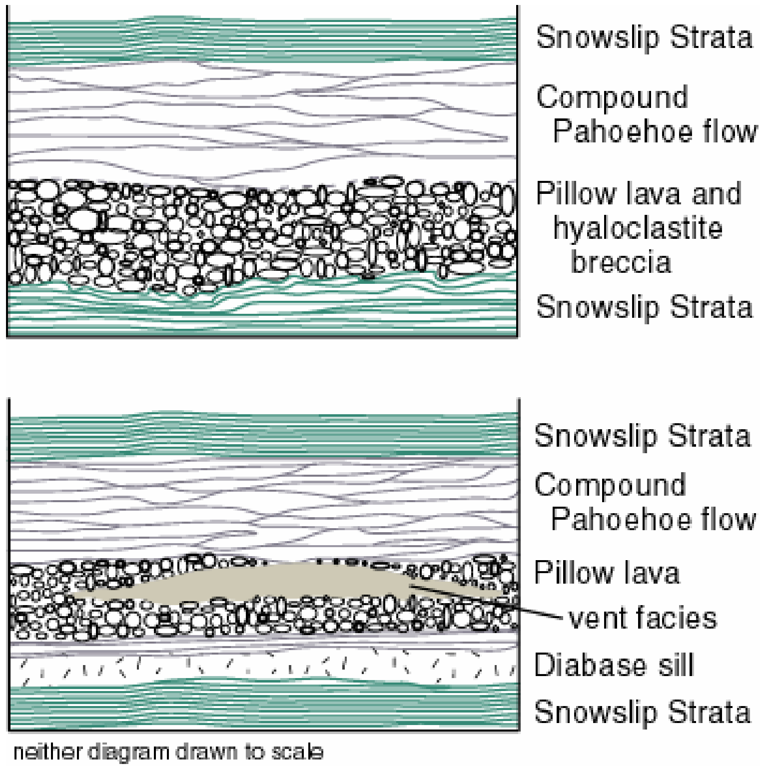


Figure 15: Subaqueous (pillow layers) and subaerial (flows) phases of Purcell Lava enclosed by Snowslip Formation strata in Glacier National Park. The lower diagram depicts the vent facies found within the Purcell Lava locally. Modified from Raup et al. (1997).

The basal 9 – 15 m (30 to 50 ft) of the succession ubiquitously contains well-developed, variably sized pillows (0.9 to 1.5 m, 3 to 5 ft in diameter) in a matrix of hyaloclastite (textural term indicating a small size difference between the clasts and matrix material) breccia; it is the only volcanic unit that can be correlated throughout the Park (Whipple et al. 1997; Whipple et al. 1984; McGimsey 1985). Subunits include isolated and broken-pillow breccias, as well as coalesced pillows at the upper contact (Whipple et al. 1984). The total thickness of the pillowed section probably represents the water depth at the time of emplacement (McGimsey 1985; Whipple and Johnson 1988). Above the pillowed section lies a series of compound subaerial pahoehoe flow units 0 to 54 m (0 to 177 ft) thick (Whipple et al. 1984; Whipple et al. 1997). Individual flow units range in thickness from about 10 cm to 6 m (4 in to 20 ft), and typically cannot be traced laterally for more than a few hundred meters. Three distinct lava types – massive pahoehoe, scale pahoehoe, and ropy lava – are recognized using the scheme of Jones (1943) (Whipple et al. 1984).

The vent facies unit occurs only in the northern exposures and is a 9 to 12.5 m (30 to 40 ft) thick chaotic breccia composed of angular and subrounded cognate or inclusive blocks, bombs (larger than 64 mm in diameter), and lapilli (volcanic ejecta ranging from 2 – 64 mm in diameter) intermixed with accidental sedimentary country rock blocks (as large as 2.5 m, 8 ft) in a fine tuffaceous matrix (Whipple et al. 1984; McGimsey 1985). The vent facies is in the upper part of the pillow unit, linearly distributed along a Northwest-southeast trend, and is overlain by pahoehoe flow-units. Transverse cross sections are lens-shaped. The unit appears to represent explosive fissure eruption processes, , possibly analogous to the famed Pu`u `O`o vent in Hawaii (McGimsey 1985). The hypabyssal diabase (dry volcanic rock type consisting of feldspar and pyroxene) sill is coextensive with the vent facies, both present only in the northern exposures. The sill is generally 18.4 to 21.4 m (60 to 70 ft) thick and located typically about 5 m (15 ft) beneath the base of the pillow unit. Apophyses or branches and upward projecting cupolas of the sill protrude into and through an overlying sedimentary interval and terminate against the base of the pillow unit (Whipple et al. 1984).

Primary flow structures and features in the Purcell Lava are remarkably well preserved, however, the primary mineralogy and chemical composition has been extensively altered by diagenesis related to burial. The original composition was probably that of continental alkaline or sodium-potassium rich basalt (McGimsey 1985). The age of the Purcell has not been directly determined, primarily because no unaltered minerals suitable for dating have been discovered. However, Hunt (1962) dated the hornfelsed (metamorphosed via contact with a molten igneous rock) strata beneath the lowest flow at 1075 Ma by the K-Ar method using biotite. Thus, the Purcell Lava may be part of the widespread 1100 Ma anorogenic or non-mountain building magmatic event that affected much of the North American craton (Whipple et al. 1997). Alternatively, Link and others (1993) give an age range for the Belt Supergroup, which includes the Purcell Lava, of 1450 to 1250 Ma.

The Purcell Lava and the diorite sill and dike in the Helena Formation were inferred to be related (Daly 1912; Ross 1959). However, field relations and geochemical differences indicate that the Purcell Lava is neither cogenetic nor coeval with a large gabbroic sill that is conspicuous throughout the Park and is commonly referred to as the "Purcell Sill." (McGimsey 1985). Trace element compositions of the sill and prominent diorite dike are virtually identical but significantly different from that of the Purcell Lava, which further supports cogenesis of the sill and dike, but not the Purcell Lava (Whipple et al. 1984; Whipple et al. 1997).

Shepard Formation. The Shepard Formation, named by Willis (1902) for the strata near Shepard Glacier, is exposed throughout Glacier National Park and the Whitefish Range to the west. The Shepard in Glacier National Park ranges considerably in thickness; it thins from 472 m (1550 ft) at the southern edge of the Park to 168 m (550 ft) at the northern edge and to about 244 m (800 ft) in the Apgar Mountains, father west (Whipple et al. 1984).

In part, the carbonate rocks of the Shepard closely resemble those of the Altyn and Helena Formations and the uppermost part of the overlying Mount Shields Formation (see below). It consists of primarily of yellowish-gray to greenish-gray dolomitic and pyritic siltite and argillite and minor thin beds of coarse-grained calcarenite, quartz arenite, limestone, muddy dolostone and dolomite (Horodyski 1983; Whipple et al 1984). Irregular cycles one to several meters thick are locally developed in the Shepard Formation. These "cycles", often poorly developed, are best defined 70 – 150 m (230 - 492 ft) above the base of the Shepard Formation where they consist of an irregular, erosional base with a few centimeters of relief, a lower fine-grained sandstone or dolomitic sandy argillite, muddy dolostone and sandy dolostone. "Molar-tooth" structure is most abundant near the top of these cycles, and molar-tooth debris lines beds a few centimeters to decimeters thick, which locally cap the cycles and represent erosional lag or remnant deposits (Horodyski 1983).

Thin beds of simple mound-shaped stromatolites in limestone are common in the southern exposures but rare in the northern part of the Park (Horodyski 1983). Lamination is generally wavy, nonparallel, and composed of fining-upward couplets (a term used for pairs of sediments or structures that commonly occur together in a rock column). Fluid-escape structures, ripple marks, miniature calcite molar-tooth structures and mudchip breccias are common in the Shepard Formation (Whipple et al. 1984). Because of the carbonate and pyrite content of the strata, most exposures weather tan to dusky orange (Whipple et al. 1985). At places, the Shepard rests unconformably on the Purcell Lava; elsewhere, the lower part of the Shepard is transitional with member 6 of the Snowslip Formation (Whipple et al. 1984).

Mt. Shields Formation. The Mount Shields Formation is exposed in parts of Glacier National Park and the Whitefish Range to the west. The type locality for the Mount Shields Formation is near Mount Shields at the southern edge of Glacier National Park. Here, the formation is about 777 m (2550 ft) thick, but to the north and west in the Park, much of the formation has been removed by erosion (Whipple et al. 1984). At the type section, the formation is informally subdivided into five members designated 1 through 5 in ascending order; it has a maximum thickness of about 853 m (2800 ft) (Whipple et al. 1997).

Member 1 is about 31 m (100 ft) thick and consists of thinly laminated, maroon to pale-purple argillite, brick-red siltite, and some interbedded arenaceous siltite and thin intervals of greenish-gray siltite and argillite (Whipple et al. 1997). At the northern boundary of the Park, member 1 encloses a bed of basaltic lava that is about 10.7 m (35 ft) thick. Member 2 is about 271 m (890 ft) thick and shows an increase in arenite relative to other members in the Mt. Shields Formation. Thin, fining-upward successions in member 2 are composed of brick-red, very fine-grained arenite and coarse-grained siltite that are capped locally by distinctive dark-red argillite (Whipple et al. 1984; Whipple et al. 1997). Ripple cross-lamination and some even, parallel lamination are common in the lower part of the successions. At the top of member 2 are unique beds of pink to cream limestone that contain oolites and small stromatolite mounded heads; this zone is recognized throughout the northern part of the Belt basin (Horodyski 1983; Whipple et al. 1984).

The base of member 3 is placed on top of the uppermost bed of stromatolitic limestone of member 2, and its lower part contains pale-purple to brick-red, very fine-grained arenite similar to that in member 2 but in more equal proportions of siltite and argillite beds Horodyski 1983; Whipple et al. 1984). Upwards in member 3, arenite beds decrease, argillite beds increase, and salt casts and ripple marks are abundant, but salt casts, or the structures left when a salt crystal dissolves and is filled by other sediment, become less abundant downward as arenite beds increase and argillite beds decrease. Near the top, siltite laminae change color from brick-red to purplish-gray and dark grayish-green, while the argillite laminae remain dark red to purple throughout the member (Whipple et al. 1997). Member 3 is the thickest (450 m, 1475 ft) member of the formation at the type locality in Glacier National Park (Whipple et al. 1984).

Member 4 closely resembles the lithology of the upper part of member 3 of argillite and siltite, including salt casts, but member 4 is grayish-green and contains carbonate generally as cement in siltite (Whipple et al. 1984; Whipple et al. 1997). Member 4 is only about 17 m (55 ft) thick at the type locality but appears to thicken and to contain more carbonate beds northward in the Park. The uppermost member of the Mount Shields, member 5, is characterized by very thinly laminated, blackish-green argillite and some thin, lenticular beds of arenaceous siltite that are more abundant near the top of the member. This distinctive succession of blackish argillite, which is only about 9 m (30 ft) thick, is locally calcareous and overlain sharply by the pale-green, relatively coarse-grained and poorly sorted feldspathic arenite of the Bonner Quartzite (Whipple et al. 1984; Whipple et al. 1997).

Bonner Quartzite. The Bonner Quartzite extends nearly the width of the Belt basin and is exposed locally in Glacier National Park. In the Park, the Bonner was formerly termed the Red Plume Quartzite, (Childers 1963), and in the Whitefish Range, to the west, it was called the Phillips Formation (Johns 1970). The only exposures known in the Park at this time are just east of Mount Shields, where the formation is about 244 m (800 ft) thick, and northwest of the mouth of Coal Creek (Whipple et al. 1984). The Bonner consists primarily of pinkish-gray to pale-red, very fine- to medium-grained feldspathic arenite, which contains large-scale channel deposits, and lesser amounts of interbedded siltite and dark-red argillite that commonly occur as rhythmic, fining-upward sequences as much as 3 m (10 ft) thick. Five informal lithofacies are recognizable in most exposures (Whipple et al. 1997).

The first lithofacies contains cross-bedded arenite, pink to maroon subfeldspathic arenite and fine- to medium-grained feldspathic arenite. Planar cross-bedding sedimentary features and mud-chips are common on foresets and at the base of the first lithofacies.

The second lithofacies is composed of even parallel-laminated arenite, pink to red, fine-grained feldspathic arenite, and very thin mudchips oriented parallel to the laminae (Whipple et al. 1984; Belt III). Faint, wavy, parallel-laminated arenite, and dark red, very fine-grained arenite to coarse siltite, commonly containing buff-colored bleach marks, comprise the third lithofacies of the Bonner Quartzite. The fourth informal lithofacies is composed of ripple cross-laminated arenite and pink to dark red, fine to medium-grained feldspathic arenite, with ripple marks and mudchips common throughout. Thinly laminated argillite, dark-red argillite, and discontinuous lamination distinguish the fifth lithofacies of the Bonner Quartzite, the top of which is irregular and eroded (Whipple et al. 1997).

Fining-upward sequences begin generally with lithofacies 1 and become finer-grained upward in numerical order, but do not comprise all 5 lithofacies. Generally, the cross-bedded arenite lithofacies is the most abundant, particularly in the middle part of the Bonner, where is appears to form broad shallow channels (Whipple et al. 1984; Whipple et al. 1985; Whipple et al. 1997).

McNamara Formation. Rocks that overlie the Bonner Quartzite in Glacier National Park and the Whitefish Range are assigned to the McNamara Formation. The McNamara is the uppermost stratigraphic unit in the Park; only the lower part is exposed and only at two localities: east of Mount Shields (where it is 61 m, 200 ft thick) and northwest of the mouth of Coal Creek (thickness unknown). In the northern part of the Whitefish Range, the McNamara is about 1219 m (4000 ft) thick (Whipple et al. 1984).

Exposures of the lower part of the McNamara in Glacier National Park consist of interlaminated grayish-green siltite and argillite that are commonly arranged as fining-upward couplets, some laminae of calcareous siltite, and thin, lenticular laminae of calcareous arenite and quartz arenite. Thin, interbedded successions of grayish-red siltite and argillite that show green mottled patches are common (Whipple et al. 1984). Silicified mud-chips (chert) are characteristic of the McNamara in many areas of the Belt basin; the Park is no exception, as even limited exposures contain silicified, discontinuous laminae and mud-chips. (Whipple et al. 1997)

Mesozoic Era

The Cretaceous age rocks of Glacier National Park are typically found in scarce exposures beneath the Lewis thrust and a sequence of formations as much as 2134 m (700 ft) thick crops out to very limited extent in the eastern and southeastern part of Glacier National Park and in the adjoining area to the east. This predominantly clastic sequence has a western source area. Willis (1902) first recognized Cretaceous rocks in the Glacier National Park area. They are described briefly below.

Lower Cretaceous – Kootenai Formation. The oldest Mesozoic rock formation exposed in Glacier National Park, the Kootenai Formation of Early Cretaceous age, consists chiefly of variegated mudstone, siltstone and sandstone. In the Park area the thickness of the Kootenai is probably greater than 305 m (1000 ft). The Kootenai is divided into two units.

The lower unit or Cutbank Sandstone Member is 44 m (145 ft) thick. It consists of two distinct sandstones separated by 4 m (12 ft) of olive-green-gray claystone (Whipple et al. 1985). The lower sandstone is about 12 m (35 ft) thick and is composed of a prominent crossbedded chert-pebble conglomerate, which grades upward into very fine sandstone and siltstone at the top. It rests conformably on Upper Jurassic rocks (Rice and Cobban 1977). The upper sandstone is 2 m (8 ft) thick and is graded and crossbedded. The rest of the lower unit is dusky-yellow to olive and dusky-red to very dark purplish-red mudstone. The top of the lower unit of the Cutbank Sandstone Member consists of a conspicuous group of dense light-gray limestone beds as much as 1 m (2 to 3 ft) thick (Rice and Cobban 1977).

The upper unit is composed of gray-green and maroon mudstone and greenish-gray sandstone (Whipple et al. 1985). The sandstone beds are fine- to coarse-grained, crossbedded, lenticular, and as much as 15 m (50 ft) thick. Conglomeratic beds composed of relatively well-rounded pebbles and cobbles of chert, quartzite and silicified carbonate occur locally at the base of sandstone beds locally. Near the top of the formation are distinctive brownish-gray to brown-weathering limestone beds and lenses of coquina that contain nonmarine bivalves and gastropods (Rice and Cobban 1977; Whipple et al. 1985). Specifically, the fossils consist of the bivalves *Protelliptio douglassi*, *P. reesidei*, and *Lampsilis farri*, and the gastropods *Stantonogyra silberlingi* and questionably *Reesidella montanaensis*(Whipple et al. 1985).

Lower Cretaceous – Blackleaf Formation. The Blackleaf Formation conformably overlies the Kootenai and is about 229 – 244 m (750 - 800 ft) thick. A unit consisting of dark-gray fissile shale and quartzose sandstone forms the basal part of the Blackleaf (Whipple et al. 1985). The formation is primarily comprised of alternating beds of light-colored clastics (Rice and Cobban 1977). Two of four assigned members of the Blackleaf formation, called the Flood and Vaughn members, are present in the Glacier National Park area (Rice and Cobban 1977; Whipple et al. 1985).

The Flood Member is composed of dark-gray to black shale and of two quartzose sandstone units. Shale forms the basal part of the member and separates the sandstone units. The lower sandstone is light gray, very fine-grained, and contains many shale partings. Upper surfaces of the sandstone are ripple-marked and contain coal and plant fragments, small, smooth-walled horizontal burrows, and *Arenicolites* (Rice and Cobban 1977). The upper sandstone unit is light gray, is fine- to medium-grained and is transitional to the underlying shale and is overlain by a coal bed. Near East Glacier Park, the Flood Member is 40 m (132 ft) thick (Whipple et al. 1985). It weathers light yellowish brown and forms prominent ledges as much as 11 m (35 ft) thick. This crossbedded sandstone contains abundant carbonaceous fragments in the upper 1.5 m (5 ft), and is substantially burrowed in the lower part (Rice and Cobban 1977).

The Vaughn Member consists of claystone, mudstone, siltstone and sandstone that are largely medium gray to medium greenish-gray but weather to lighter shades. The finer-grained units (claystone and mudstone) are usually bentonitic and locally carbonaceous

(Whipple et al. 1985). Thin beds composed entirely of bentonite, the term for volcanic ash layers altered to clay, are also present. The sandstone beds, which are thicker in the lower part of the Vaughn Member, are characteristically poorly sorted, fine- to coarse-grained and lenticular. The basal part of these sandstone beds commonly contains rip-up mud clasts, coal and plant fragments and, rarely, large logs. Trace fossils and fragments of carbonized plants were observed in the Blackleaf Formation (Rice and Cobban 1977).

Immediately south of Glacier National Park, several ledges of pebble- and cobble-conglomerate as much as 1 m (3 ft) thick occur in the Vaughn. These conglomerates are composed of well-rounded pebbles and cobbles of quartz, chert, quartzite and silicified carbonates, which were derived form strata of Precambrian to Mississippian age.

Upper Cretaceous – Marias River Shale. The Marias River Shale of Late Cretaceous age unconformably overlies the Blackleaf and is about 366 – 396 m (1200 - 1300 ft) thick. Most of the Marias River is dark-gray shale, with local sandy and calcareous beds, and contains calcareous and ferruginous concretions (Rice and Cobban 1977). The Marias River Shale is divided into four members present in the Glacier National Park area, namely, the Floweree, the Cone, the Ferdig and the Kevin Members (Whipple et al. 1985).

The Floweree is the lowermost member of the Marias River Shale; it is 11.6 m (41.5 ft) thick in the Glacier National Park area. It consists of a basal sandstone as much as 2 m (5 ft) thick overlain by dark-gray shale as much as 10.5 m (38.5 ft) thick. The sandstone is very fine-grained, silty and thoroughly bioturbated (disturbed by organisms). The shale sequence is noncalcareous and in the lower beds contains hard, brownish-weathering layers of ripple-marked, relatively coarser-grained siltstone. The Floweree Member overlies the Blackleaf Formation with an angular discordance between the dips of their respective sedimentary layers of as much as 10 degrees (Rice and Cobban 1977).

The Cone Member is 20.2 m (66.5 ft) in the Glacier National Park area. It is predominantly dark-gray calcareous shale, which is fissile and weathers to a distinctive silvery-gray color. Other associated lithologies within the Cone Member are marlstone (an impure limestone), limestone concretions, and bentonite beds, locally. The basal few inches of this unit are thinly bedded and limonitic. Minute white specks, composed largely of coccoliths (calcitic skeletal remains of the microscopic organism, *coccolithophore*) or coccolith detritus, are abundant in the calcareous shale. The widely distributed concretions, which are restricted to the lower 1 m (4 ft) of the member, are septarian (cracked radially), with thick veins of light-brown, coarsely crystalline calcite. The Cone Member contains scant oil reportedly in quantities equivalent to 1 to 2 gallons of oil per ton from distillation tests (Stebinger 1918) (Rice and Cobban 1977).

The Ferdig Member consists of dark-gray, noncalcareous shale and contains many ferruginous concretions. These concretions weather dusky red to dark yellow-orange, and yellowish-brown and impart a distinctive pattern of rusty-brown to red colored patches to the outcrop. A 0.3 m (1 ft) thick conglomerate, consisting of well-rounded cobbles of chert, quartzite and silicified carbonate crops out near the base of the Ferdig

Member. Thin pebbly layers are also characteristic of the Ferdig. The conglomeratic zones in the Glacier National Park area are unique to the member because they occur within a thick, fine-grained shale sequence (Rice and Cobban 1977).

The Kevin Member is the youngest and thickest member of the Marias River Shale. Determination of the thickness of the Kevin Member is difficult near Glacier National Park, due to intense deformation within it. Based on observations to the east on subsurface sections, the Kevin is approximately 229 m (750 ft) thick. The Kevin is primarily composed of dark-gray, noncalcareous shale, and is divided into three units on the basis of the abundance of bentonites and variety of concretions. Many beds of bentonite, calcareous concretions, and concretionary limestone characterize the lowermost unit. In the middle unit of the Kevin Member are a few beds of bentonite and many scattered red-weathering ferruginous concretions and calcareous concretions. The upper unit of the member contains concretionary limestone that patchily weathers to a yellowish-gray and a few thin interlayers of shaly sandstone and bentonite. It is calcareous, contains minute white specks and is sandy locally. Fossils, though relatively rare in the Marias River Shale include *Inoceramus (Mytiloides) labiatus, Ostrea, Watinoceras reesidei, Scaphites nigricollensis*. The limestone concretions in the Kevin Member contain fossils of *Inoceramus deofmis, Baculites mariasnsis*, and *Scaphites preventricosus* (Rice and Cobban 1977).

Upper Cretaceous – Telegraph Creek Formation. Conformably overlying the Marias River Shale is the transitional unit consisting of interbedded shale and sandstone, the

Telegraph Creek Formation. The contact between the Telegraph Creek and underlying Kevin Member of the Marias River Shale is marked by a change in color and lithology; dark-gray shale that contains white specks is overlain by interbedded light- to medium-gray shale, siltstone and sandstone (Rice and Cobban 1977). The Telegraph Creek Formation ranges in thickness form 36 – 52 m (120 to 170 ft) on the Kevin-Sunburst dome east of Glacier National Park (see figure 21) (Rice and Cobban 1977; Whipple et al. 1985).

The lower part of the formation consists of gray silty shale, gray siltstone, and minor interbeds of medium- to light-gray, very fine-grained sandstone. The sandstone content increases upward with a corresponding increase in the thickness of the sandstone layers and a slight increase in sand grain size. Abundant ripple marks, bits of carbonized wood, and horizontal burrows occur along the bedding planes of the sandstone beds. (Rice and Cobban 1977).

Molluscan fossils are scarce in this formation except for a few fragments of inoceramids and oysters exist near the southeastern border of Glacier National Park (Rice and Cobban 1977).

Upper Cretaceous – Virgelle Sandstone. The Virgelle Formation in the Glacier National Park area is a massive, cliff-forming formation. It consists primarily of sand deposits and is about 49 m (160 ft) thick. Titaniferous magnetite (titanium bearing iron oxide) deposits are present in many places at the top of the Virgelle. The heavy minerals in the magnetite deposits indicate the source of the sediment was from Late Proterozoic (750 m.y.) sills that are exposed in and southwest of Glacier National Park (Rice and Cobban 1977; Whipple et al. 1985).

The Virgelle is a light-gray to white, fine- to medium-grained, partly calcareous, arkosic (rich in feldspar) sandstone. It is commonly crossbedded in the upper part and contains calcareous sandstone concretions (Whipple et al. 1985). The contact between the Virgelle and the underlying Telegraph Creek Formation is conformable and placed where the lithology of the sequence changes from sandstone to generally siltstone and shale (Rice and Cobban 1977; Whipple et al. 1985).

Reeside (1927) recorded the ammonite *Desmoscaphites bassleri* near the eastern boundary of Glacier National Park. *Inoceramus lundbreckensis* has also been found in rocks assigned to the Virgelle Sandstone in the Blackfeet Indian Reservation east of Glacier National Park (Whipple et al. 1985).

Upper Cretaceous – Two Medicine Formation. Resting conformably on the Virgelle Sandstone is the Two Medicine Formation, a unit that consists mostly of mudstone with minor lenticular sandstone. The formation is about 610 m (2000 ft) thick in the Park area (Rice and Cobban 1977; Whipple et al. 1985). The contact of the Two Medicine and the underlying Virgelle is distinct. Sandy mudstones at the base of the Two Medicine form a gentle slope above the prominent sandstone cliffs of the Virgelle Sandstone (Rice and Cobban 1977).

The formation primarily consists of silty, locally calcareous mudstone, which weathers pale greenish-gray or gray. The Two Medicine commonly contains hard calcareous nodules and carbonaceous beds in the lowermost 152 m (500 ft). The sandstone is mainly in the basal beds 76 m (250 ft) above the formation base, with only thin interbeds elsewhere in the formation. This lower unit consists of green, gray and purple mudstone and siltstone, interbedded with yellowish-brown weathering sandstone. The grain size of the sandstone varies from fine- to very coarse-grained. It is commonly calcareous and forms massive, lenticular beds. Carbonaceous mudstone and coal occur near the base of this basal unit (Rice and Cobban 1977).

Cobban (1955) described a conglomerate that is about 244 m (800 ft) below the top of the Two Medicine Formation in the Glacier National Park area. It contains clasts of limestone, dolomite, quartzite, contact metamorphism hornfels, welded volcanic tuffs and quartz rich igneous rocks. The Two Medicine Formation contains Campanian fauna, which includes the dinosaur *Barchyceratops montanesis*, scales of ganoid fishes, ostracodes and fresh-water mollusks (Rice and Cobban 1977).

Upper Cretaceous – Bearpaw Shale. A dark-gray marine formation, the Bearpaw Shale, conformably overlies the Two Medicine Formation. The Bearpaw is nearly 122 m (400 ft) thick east of the Park, but thins to only 72 m (235 ft) south of the Park. The lower Bearpaw Shale grades upward into a unit of alternating sandstone, siltstone and shale that is as much as 122 m (400 ft) thick (Rice and Cobban 1977; Whipple et al. 1985). In

Glacier National Park, the Bearpaw Shale is divided into two informal members (Whipple et al. 1985).

The formation is mainly made up of dark-gray shale, comprising the lower member, which contains ferruginous and calcareous concretions, many layers of bentonite, or altered volcanic ash, from 3 to 20 cm (1 to 8 in) thick, and scant, thin beds of sandstone. The Bearpaw is sandier along Blacktail Creek where sandstone beds are fine-grained, crossbedded and as much as 5 m (15 ft) thick (Rice and Cobban 1977).

The upper part of the Bearpaw Shale grades progressively westward and southwestward into a sandy unit that Cobban (1955) designated as the Bearpaw-Horsethief transition. This transition member is as much as 122 m (400 ft) thick and represents a gradation from predominantly shale of the Bearpaw to sandstone of the overlying Horsethief. This unit is generally composed of alternating beds of light-gray sandstone and dark-gray shale. The sandstones are very fine- to fine-grained, commonly laminated to thin-bedded, and seldom exceed 3 m (10 ft) in thickness. The shale beds are silt or sand rich and occur as interbeds a few inches to 21 m (70 ft) thick (Rice and Cobban 1977).

Fossils of *Baculites compressus*, *B. coneatus*, and *B. reesidei* have been found in the Bearpaw Shale on the Blackfeet Indian Reservation east of Glacier National Park. In addition to the baculites, limestone concretions in the Bearpaw commonly contain the ammonites *Hoploscaphites* and *Placenticeras*, and the bivalves *Nucula, Nuculana, Inoceramus, Oxytoma, Cymella* and *Nymphalucina* (Rice and Cobban 1977).

Upper Cretaceous – Horsethief Sandstone. The Horsethief Sandstone is another massive cliff-forming sandstone that was deposited above the Bearpaw Shale. The formation is composed of light-gray, fine- to coarse-grained sandstone, which is commonly crossbedded and contains calcareous concretions. On the Blackfeet Indian Reservation, east of Glacier National Park, the upper part of the Horsethief includes ledge-forming beds of dark-brown sandstone that contain titaniferous magnetite. The formation is about 27 m (90 ft) thick (Rice and Cobban 1977).

The brackish water bivalves *Crassostrea wyomingensis* and *Veloritina occidentalis* and the gastropod *Melania wyomingensis* have been found at the top of the Horsethief Sandstone, whereas a shallow-water marine bivalve, *Tancredia*?, occurs lower in the Horsethief (Rice and Cobban 1977).

Upper Cretaceous – St. Mary River Formation. Conformably overlying the Horsethief Sandstone is the St. Mary River Formation. The St. Mary River Formation is about 305 m (1000 ft) thick near the eastern margin of the Disturbed belt in the Glacier National Park area. It closely resembles the Two Medicine Formation in color and composition. The St. Mary River Formation is composed mostly of greenish-gray mudstone interlayered with lenticular beds of fine- to medium-grained crossbedded sandstone. Red mudstone also occurs in the lower part of the formation. Thin coal beds, at the base and top of the formation were locally mined in the early 1900's. The lower part of the St.

Mary Formation is well exposed along U.S. Highway 89 just north of the Two Medicine River (Rice and Cobban 1977).

The basal beds of the St. Mary River Formation contain the same species of brackish-water fossils that are found at the top of the Horsethief Sandstone. These fossils include nonmarine bivalves, such as *Fusconaia? stantoni*, and locally, fossil leaves. Scattered dinosaur bones occur throughout the formation. An incomplete skeleton of *Montanaceratops* was found on the Blackfeet Indian Reservation, to the east of Glacier National Park (Rice and Cobban 1977).

Upper Cretaceous – Willow Creek Formation. The Willow Creek Formation is composed of about 244 m (800 ft) of variegated clayey rocks, mudstone, and sandstone and commonly contains purplish-gray limestone nodules. The formation is very similar to the underlying St. Mary River Formation, but is distinguished by the dominant red color and the absence of coal beds. The lower part of the Willow Creek formation contains fresh-water mollusks and an occasional dinosaur bone. The top of the Cretaceous stratigraphic column is somewhere within this formation (Rice and Cobban 1977).

Cenozoic Era

Tertiary – Kishenehn Formation. The North Fork Valley is a long trough formed along normal faults during crustal extension. The valley resulted as one block was downdropped relative to another forming a discrete structure known as a graben (the accompanying highland is called a horst) (figure 16). The Kishenehn Formation, a specific local name for the more regionally recognized Renova Formation, filled the basin during its formation in Tertiary time. A notable inclusion in the Kishenehn Formation are several hundred feet of oil shale. In fact, bears would wallow in oil seeps of the North Fork Valley, and their resulting smelly fur was a distinguishable characteristic to early settlers (Alt and Hyndman 1986).

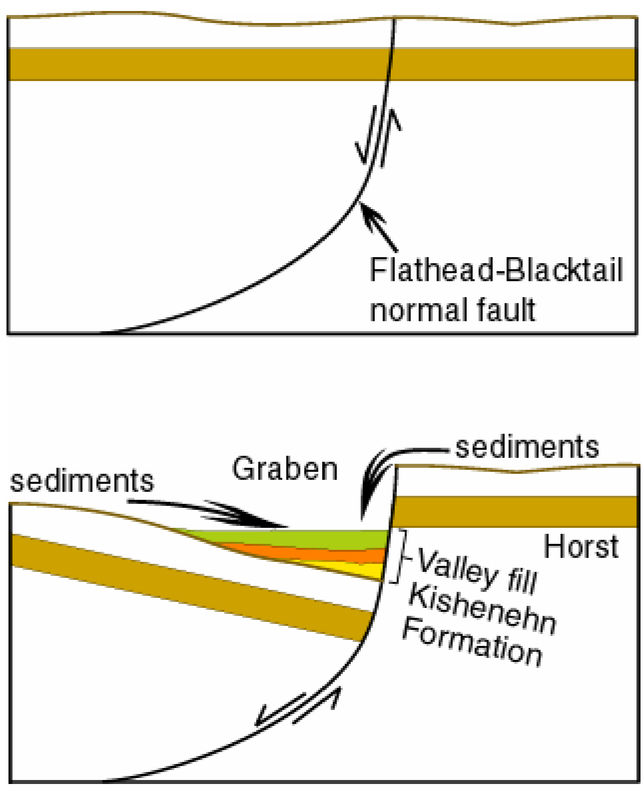


Figure 16: Formation of the Kishenehn Formation as sediments continuously filled the graben valley of the North Fork.

The Kishenehn Formation consists of more than 610 m (2000 ft) of layered gravel, sand, mud, volcanic ash, limestone, and coal. It appears in outcrop as predominantly pale gray and tan rocks that are barely cemented and contain a large percentage of sand and silt. Glacial debris blankets most of the valley making the exposures of the older fill rather hard to find (Alt and Hyndman 1986) . The upper, undivided Kishenehn Formation is separable into two parts. The upper part is a sequences of brick-reddish brown mudstone, sandstone, and coarse conglomerate with interbedded gray, calcareous sandy pebble and cobble conglomerate. The lower part consists of light-gray to grayish green sandstone, siltstone, mudstone, lignite, oil shale, marlstone, and sandy pebble and cobble conglomerate (Whipple 1992).

Below the undivided member lies the conglomerate member of Pinchot Creek of the Kishenehn Formation. It is brownish-red with intercalated mudstone, sandstone, and conglomerate. In specific locations, calcareous sandy mudstone and siltstone grade to muddy sandstone. The muddy sandstone is composed of varicolored, angular to subrounded, sand- and pebble-size clasts. Pebble and boulder conglomerate beds are gray and sandy and contain abundant beds of mudstone and sandstone. The lithic clasts in the pebble and boulder conglomerate beds are rounded and subrounded, consisting entirely of Belt Supergroup fragments, some up to 2.5 m (8.2 ft) in diameter. The maximum thickness of the Conglomerate member of Pinchot Creek is estimated at 700 m (2297 ft) (Whipple 1992).

Below the Conglomerate member of Pinchot Creek of the Kishenehn Formation lies, in a conformable way, the Lacustrine member of Coal Creek of the Kishenehn Formation. It is predominantly a light-gray, heterogeneous assemblage of sandstone, siltstone, mudstone, claystone, coal, oil shale, marlstone and pebble and boulder conglomerate. The Lacustrine member is informally divided into three parts, each bounded by gradational contacts with each other. The total thickness is about 1,150 m (3773 ft). The upper part is typically an interbedded sequence of marlstone, litharenite (barely solidified or lithified), siltstone, conglomerate, mudstone, claystone, and coal. Outcrops of this member have a distinctive pink cast from the presence of variegated maroon, red-brown, and gray-green sandy mudstones. The middle part is interbedded oil shale, marlstone, litharenite and siltstone and some lignite (low grade coal), sapropelic (derived from plant sludge at the bottom of shallow water bodies) coal, tuff, claystone and mudstone. Fission track analysis of zircon (tough mineral, rich in zirconium) from a tuff bed suggest an Eocene age of 43.5 to 4.9 Ma (analysis by Charles Naeser, written communication to Whipple 1992, 1990). The lower part is composed primarily of interbedded carbonaceous siltstone, silty to coarse-grained litharenite that displays climbing-ripple and even, parallel lamination, and light bluish-gray weathering oil shale. Lignite, mudstone, claystone, marlstone, conglomerate and altered tuff beds are present in the lower part locally (Whipple 1992).

The Kishenehn Formation contains abundant petrified wood, including the Dawn Redwood, and fossil leaves (Alt and Hyndman 1986). The upper part contains fossil gastropods, mammals, and palynomorphs. The conglomerate member of Pinchot Creek contains some vertebrate fossils in mudstone layers. The Lacustrine member of Coal Creek contains some fossil gastropods and Eocene-age mammals as well as plants and fragments, fish, insects and mollusks. Leaves of *Macginitea augustiloba* are present in some beds of the lower part (Whipple 1992).

Quaternary – Glacial and Alluvial Sediments. The Quaternary age sediments at Glacier National Park include alluvium, alluvial fill, colluvium, landslide deposits, terrace gravel, and glacial till and outwash deposits. The Holocene and upper Pleistocene glacial till is composed of the jumbled assortment of subrounded to subangular bouldery rubble derived mainly from Belt Supergroup rocks combined with sand, silt and clay. Locally, deposits of till in the form of ground moraines are more than 30 m (98 ft) thick. In mountainous areas these ground moraine deposits are usually 1 – 3 m (3.2 - 9.8 ft) thick. Terminal moraines both in the mountains and valleys range from 3 – 50 m (9.8 - 164 ft) thick (Whipple 1992).

Landslide deposits at Glacier National Park include large slumps, block slides and earth flows. Slides are most common in areas underlain by the Cretaceous age rocks of the eastern side of Glacier Park and by the Kishenehn Formation in the western part of the Park. Some larger landslide deposits in the park exceed 50 m (164 ft) in thickness. Some of the more cohesive blocks in the landslide deposits are composed of till, rock glaciers, talus and colluvium (Whipple 1992).

Colluvium is comprised of locally derived slope deposits. It consists in Glacier of unsorted angular, gravel-size clasts in a sand-silt-clay rich matrix. The unit also locally includes small pockets of till, talus, rock-avalanche, and debris-flow deposits. Colluvium in Glacier National Park is commonly 1 – 5 m (3.2 - 16.4 ft) thick.

With Glacier's abundant rivers and lakes, alluvium is found all over the Park. It is typically comprised of unconsolidated sand and gravel deposits. Locally the alluvium contains lenses of silt. Included in the unit are channel and overbank deposits in the modern floodplains as well as alluvial-fan and terrace deposits. Most of the clasts and grains in the alluvium are derived from the Belt Supergroup rocks. Alluvium in the Park ranges in thickness form 1 – 10 m (3.2 – 32 ft) (Whipple 1992).

Structural Geology and Glacier National Park

Glacier National Park is considered part of the Cordilleran fold and thrust belt of the United States; continuous with that known in the southern Canadian Cordillera (McGimsey 1982). Price (1981) has interpreted the region to be a portion of the complex structures that resulted from intermittent collisional plate interactions beginning nearly 200 m.y. ago. Prominent regional structural features include: 1) the Lewis thrust fault system, 2) the disturbed Belt, a zone of closely spaced imbricate thrust faults to the east, 3) the Rocky Mountain Trench, a zone of closely spaced normal faults to the west, and 4) the Lewis and Clark line to the south, a major intraplate tectonic boundary in the region (figure 17). Deformation in the Cordilleran orogen was initiated during the late Jurassic and is believed to have terminated by late Eocene time (McGimsey 1982). This fold and thrust belt, however, has been undergoing extension since late Tertiary time along a series of major normal faults.

The dominant deformation in Glacier National Park is “thin-skinned”. That is to say, the folds and thrusts occur above a basal decollement (detachment) which lies near the crystalline basement, 25-30 km in depth (figure 18). The deformation is characterized by low-angle thrust faults, such as the Lewis thrust, and listric thrusts (dip shallowing with depth), concentric folds (a fold in which the thickness of the layers is constant), transverse tear faults (like a strike-slip fault), and late normal faults which are commonly listric (sole gently into the low-angle thrusts). Coincident with the Lewis thrust plate are numerous subsidiary thrusts, and near the leading edge on the eastern side of Glacier National Park numerous listric imbricate thrusts occur (figures 18 and 19). The geologic structures of Glacier are part of a system of northwest-southeast trending fold axes and faults of the northern U.S. Rocky Mountains.

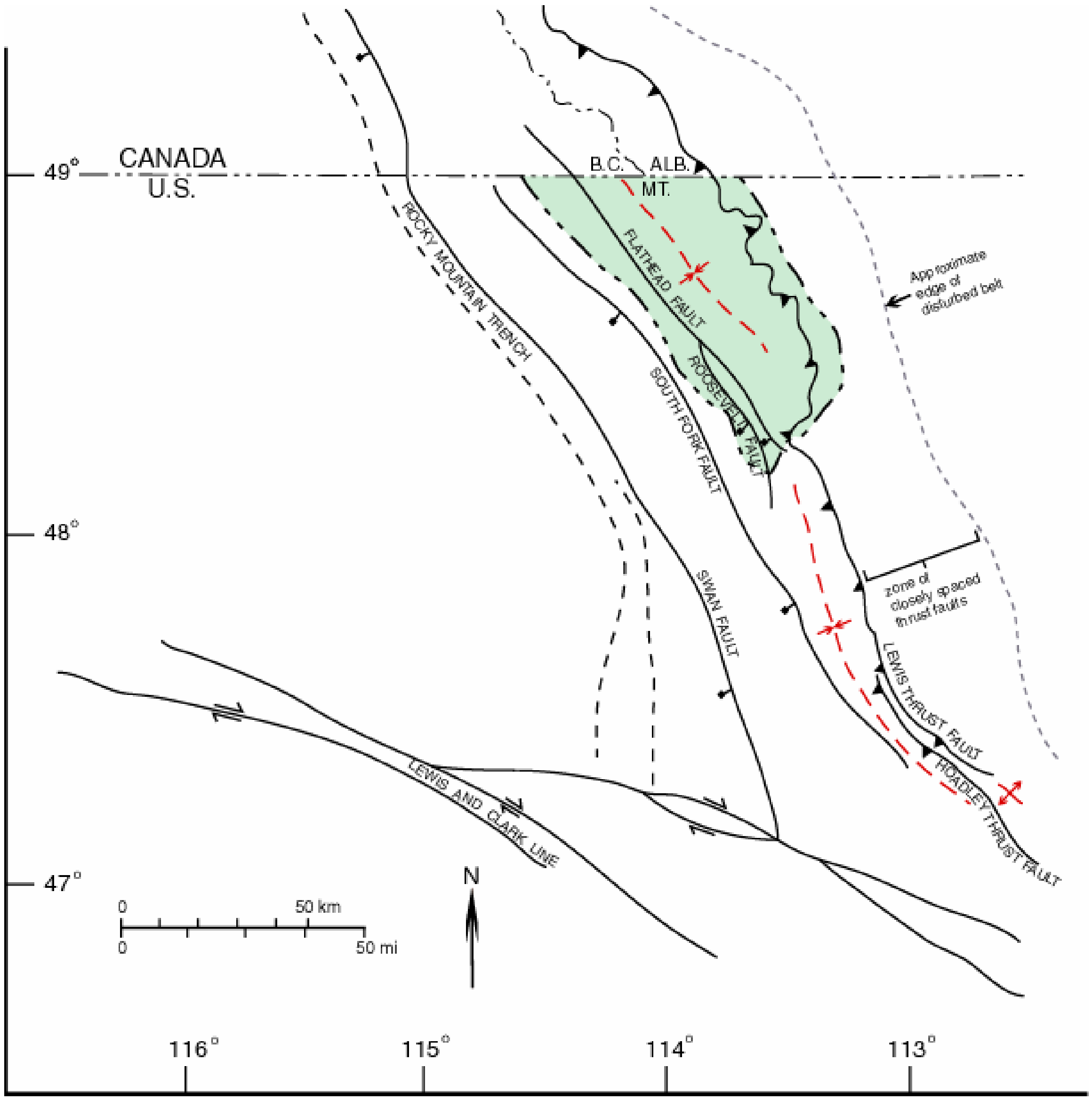


Figure 17: Map of structural features in Glacier National Park area, modified from McGimsey 1982.

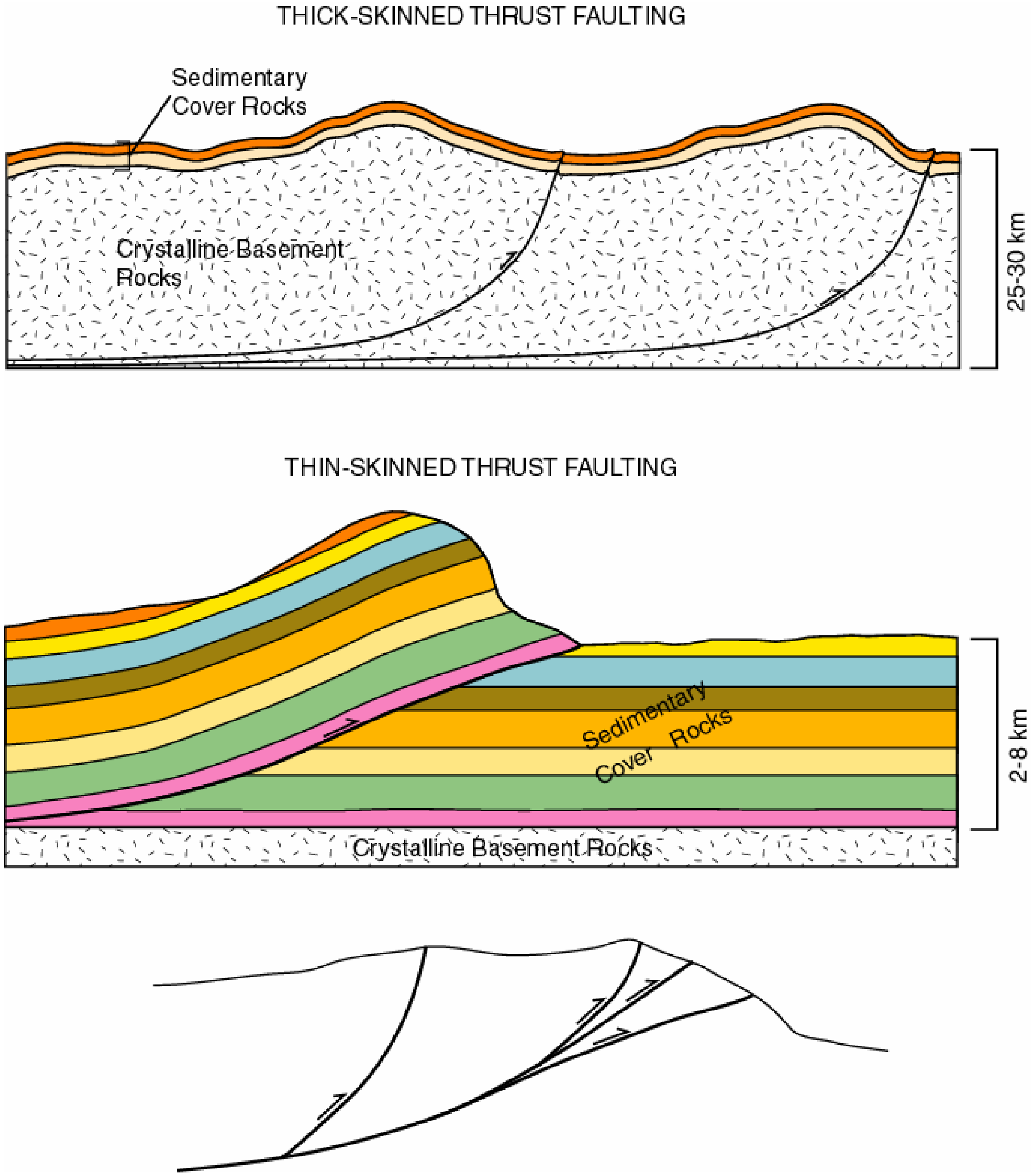


Figure 18: The difference between thin- and thick-skinned thrust faulting. Third diagram shows an imbricate style of faulting.

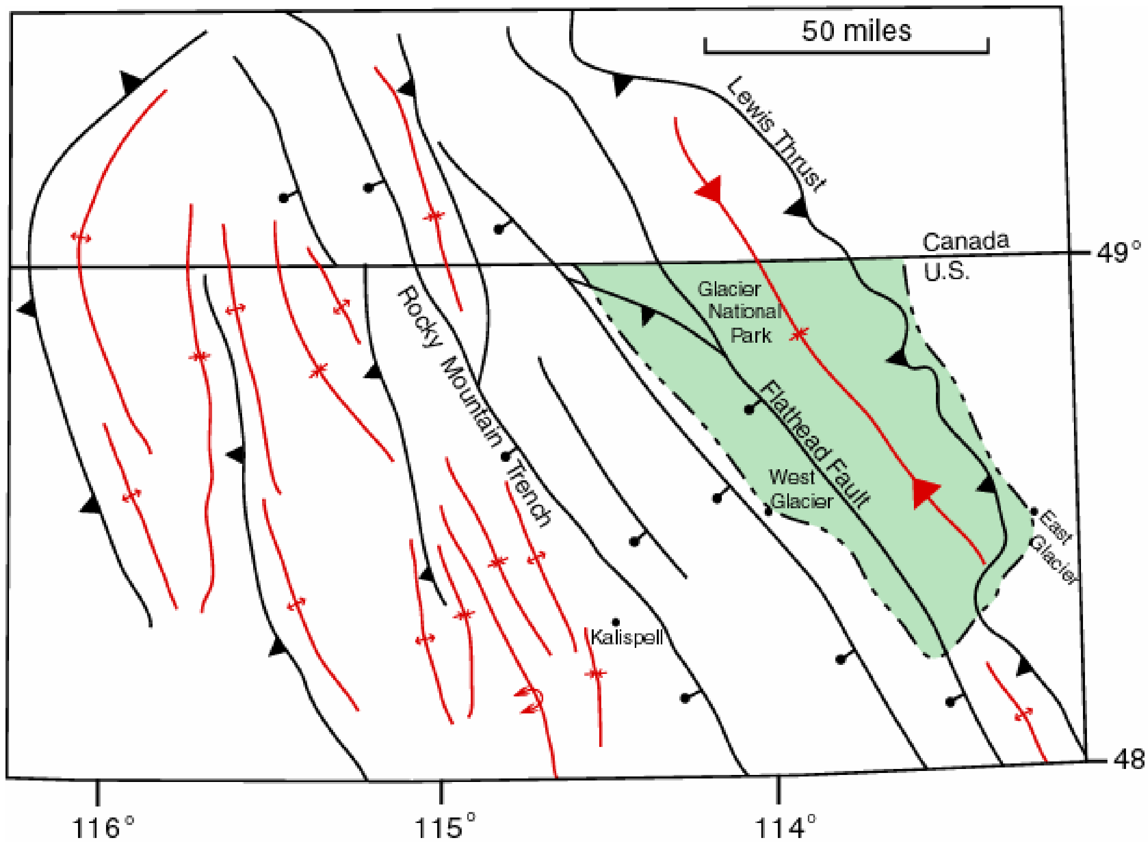


Figure 19: Glacier National Park within the structural setting west of the Park of the northern Rocky Mountains. Note the parallel nature of the anticlines, synclines, normal and thrust faults Adapted from Yin 1991.

The Lewis thrust fault, discovered and named by Willis in 1902, juxtaposes Proterozoic rocks over highly folded and faulted Mesozoic rocks in Glacier National Park (McGimsey 1982; Yin 1988). Ross (1959) described the average strike, or compass direction of a horizontal line of intersection of a plane with Earth's surface, of the thrust fault in this area as N30ºW and the dip of the gently folded fault surface as generally less than 10º SW. From a point 225 km (140 mi) north of the International boundary, the fault extends southward for 452 km (281 mi) (McGimsey 1982). Maximum translation, or horizontal distance the fault has moved, of the Lewis thrust is approximately 65 km

(40 mi), this measurement was inferred at a location along the southern edge of the Glacier.

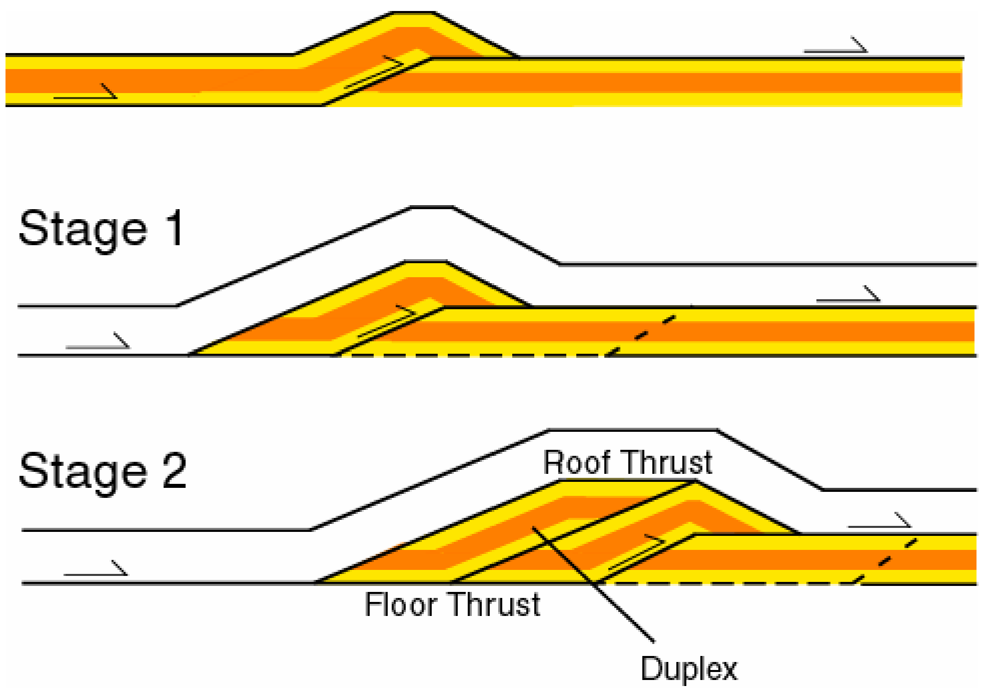


Figure 20: Diagrammatical view of duplex formation. The primary thrust breaks through another column of rock forming a fault bounded pocket of rock called a duplex. Modified from Hudec and Davis 1989.

The Lewis thrust has long been considered the “classic” example of an overthrust. However, the model of the planar fault moving a cohesive chunk of earth is not very realistic. The thrust plans usually follow bedding in the more incompetent, weaker, shaly rock layers and cut upward in the direction of transport through the more competent, usually carbonate or sandstone, layers. Fault planes are folded in some places due to movement on adjacent faults, or due to the effects of climbing upsection in the rock column (McGimsey 1982). The architecture of the Lewis thrust is extremely complex. It consists of symmetric and asymmetric concentric folds, high-angle and low-angle contractional and extensional faults, zones of complex structures, and imbricate thrust systems (see figure 20) (Yin1988).

The Lewis thrust fault can be divided into three informal segments. The northern segment extends from Kidd Mountain to North Kootenay Pass. The central segments lies between North Kootenay Pass and Marias Pass at the southern edge of Glacier National Park. The southern segments of the Lewis thrust is from Marias Pass to Steamboat Mountain (Yin 1988).

The Lewis thrust sheet (column of rock above the fault surface) in the northern and central segments is highly deformed within a zone approximately 156 – 468 m (500-1500 ft) thick immediately above the Lewis thrust. This zone of complex structures consists of imbricate thrusts, low-angle thrust faults, duplex structures, and low-angle faults along which younger strata overlie older strata. Above this zone of complex structures is the main mass of the Lewis thrust sheet several kilometers thick, is characterized by broad open folds in relatively undeformed rocks. The southern segments of the Lewis thrust fault sheet and the structures in its upper plates from Marias Pass to Sun River show broad folds and little faulting. In contrast, structures in the lower plate (rock below the fault surface) of the Lewis thrust fault are very complicated consisting of closely-spaced imbricate thrust faults and folds, some of which are locally overturned (Yin 1988).

The Lewis thrust fault juxtaposes mid-Proterozoic Belt Supergroup rocks in its upper plate with the late Cretaceous age Marias River Shale in its lower plate. The surface of the fault is usually covered by talus and other slope debris which makes it difficult to measure the surface directly, however, the attitude of the Lewis thrust fault from a direct measurement at the western end of Forty One Mile Creek is N20ºW 60ºSW. The fault is located at an elevation of about 1829 m (6000 ft) along the eastern edge of Glacier National Park, and dips down to less than 1372 m (4500 ft) in elevation in western areas of the Park (Yin 1988). The fault overlies the Cretaceous age rocks of the disturbed belt; a narrow north-south trending strip of land to the east of Glacier National Park (figure 21). Individual faults within the disturbed belt are difficult to distinguish given the extreme, pervasive deformation. East of the disturbed belt lies the Sweetgrass Arch and associated Kevin Sunburst Dome. These features are likely the results of flexural folding of the crust in response to the weight of the thick rock column to the east (Lewis thrust sheet).

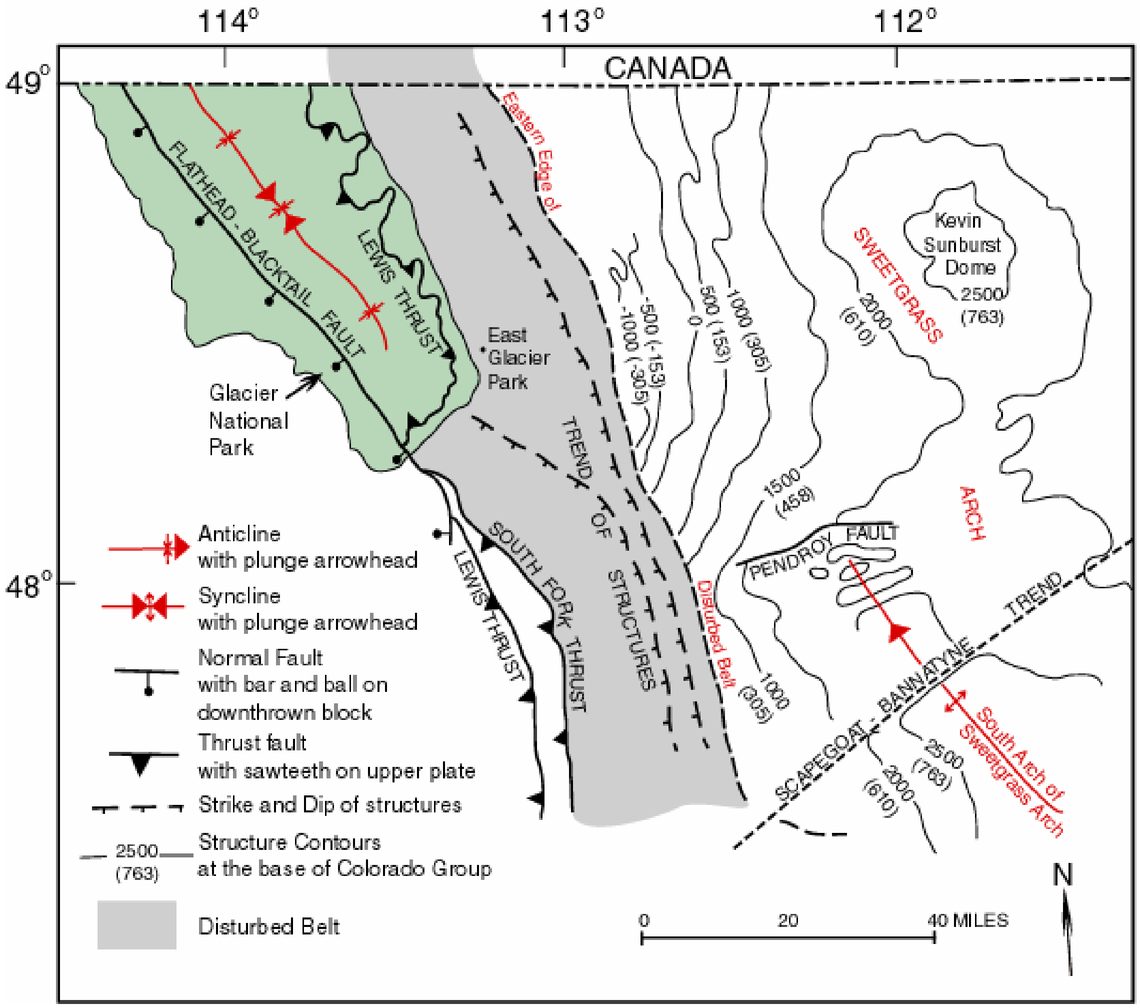


Figure 21: Structural features east of Glacier National Park. Note the structure contour lines recording the elevation below the surface of the top of the Cretaceous age Colorado Group rocks. The contours record the subsurface depression the Park is situated in.

The prominent syncline, the Akamina Syncline, running up the axis of the park is a broad open structure which lies totally within the upper plate of the Lewis thrust (figure 22). The trend of the syncline is roughly parallel to the strike of the thrust and numerous local normal faults in the Glacier National Park area (McGimsey 1982). The Akamina syncline is a doubly plunging structure (a fold that reverses its direction of plunge within an area, resembles the surface of a upturned bread loaf) that occupies the entire Lewis

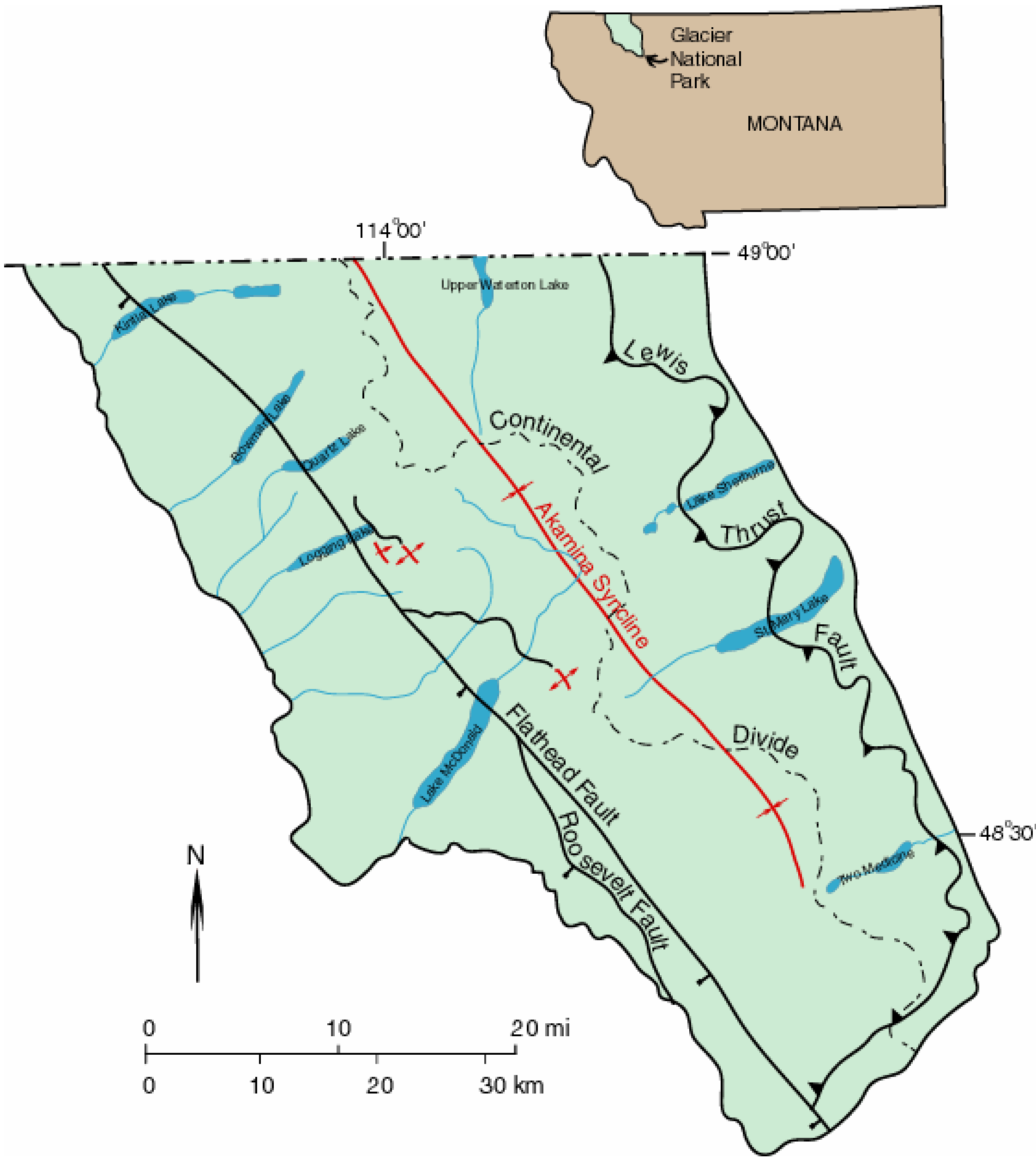


Figure 22: Prominent structural features inside Glacier National Park. Modified from McGimsey 1982

salient (portion of the thrust sheet which projects further than surrounding areas) from North Kootenay Pass to Marias Pass. It records the structural low, bowl shaped depression, in which the Park lies (figure 21). Regionally, the Akamina syncline is bounded by the Lewis thrust to the east, and the Flathead-Blacktail normal fault system to the west. Within Glacier National Park, the hinge or axis of the syncline follows the continental divide and ends at the south-central edge of the Park (Yin 1988).

The Lewis thrust sheet is offset by the Flathead-Blacktail normal fault zone in the North Kootenay Pass and Marias Pass areas (Yin 1988). The Flathead fault is a family of major listric normal faults with an average high angle (40º or greater westward dip), but seismic data indicates the fault flattens with depth and may either sole into or just offset the Lewis thrust at depth. Displacement along the Flathead fault is estimated to be at least 13 km (8 mi) (McGimsey 1982; Yin 1988).

Running along the western edge of Glacier National Park is the Kishenehn Basin. This basin formed on the down-dropped side of the Flathead-Blacktail normal fault system. As the Roosevelt and Flathead-Blacktail normal fault systems sole into and locally offset structures in the Lewis thrust sheet, the Roosevelt and Flathead-Blacktail fault postdate the Lewis thrust fault. Thus, the Tertiary age of the Kishenehn basin presumably provides an upper limit for the movement along the Lewis thrust (Yin 1988). The basin is very deep, more than 600 m (1969 ft). These faults are among a number of roughly parallel faults west of Glacier National Park which sole into the Lewis Thrust fault. They record the extensional tectonic regime which followed the orogenic compressional events in the Rocky Mountains.

Depositional and Tectonic History of Glacier National Park

With regards to geology, Glacier National Park is mysterious. Why such magnificent mountains exist miles from a continental boundary remains the subject of many studies. Its rugged vistas inspire visitors with a sense of wonder. Yet, GLAC is more than a scenic attraction; Glacier National Park is part of the rich geological history of the Rocky Mountains. A brief synopsis is presented here, drawn on various localities in the region, to illustrate the interconnectedness of GLAC with the evolution of a continent.

Surface Geologic History

The Precambrian. The Precambrian Era extends from 570 million years ago (Ma) to about 4,500 Ma. That makes the Precambrian almost *4 billion years long.* In contrast, the Cenozoic is only about 66 million years long (0.066 billion years) and the Paleozoic, Mesozoic, and Cenozoic Eras combined only cover approximately 570 million years. Trying to read the story of the Precambrian is like trying to read *War and Peace* – only in Russian with whole chapters torn out. In other words, geologists know very little of what happened in the Precambrian, and we have very little data with which to minimize the uncertainty (Graham et al. 2002). Glacier National Park contains some of the finest Proterozoic age sedimentary rocks on earth. These exposures provide a unique insight into the regional history during the Precambrian (Table 1). About 88% of Earth's history is contained in the Precambrian. With regards to the western United States including Glacier National Park, our Precambrian story begins in the Late Archean.

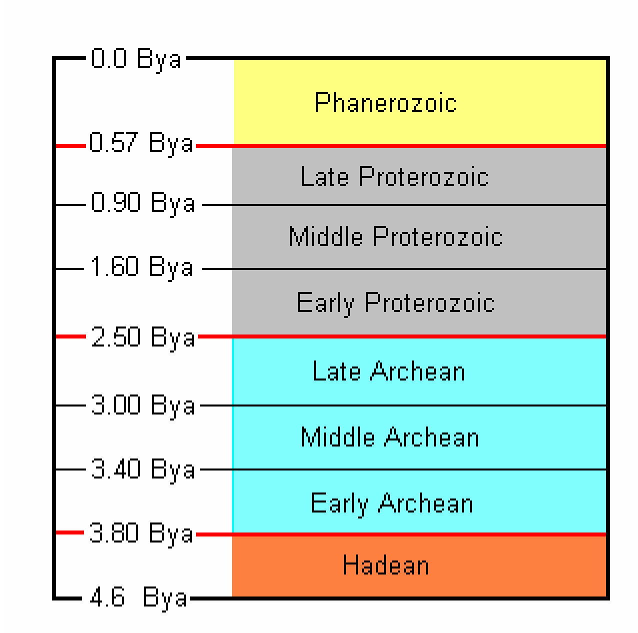


Table 1: Subdivision of the Precambrian Bya: Billion years ago.

The southern border of what is now Wyoming formed the southern margin of the North American craton during the Archean, and the metasedimentary rocks were deposited in a broad continental shelf that extended to the south and west. As the Archean drew to a close, however, the North American craton suddenly experienced a growth spurt. Observations of the folds and faults in the Red Creek Quartzite rock unit below the angular unconformity with the overlying Uinta Mountain Group rocks in the Uinta Mountains of Utah and Dinosaur National Monument (Utah and Colorado) indicate that a collisional event began to affect the area about 2.3 billion years ago (Hanson, 1975; Graham et al. 2002).

The Proterozoic – Belt Supergroup. Recorded Precambrian history began in the Glacier National Park area in the Proterozoic. The middle Proterozoic Belt Supergroup is a sequence of argillite, quartzite, and carbonate rocks that unconformably overlie the Archean Wyoming province to the east and thicken southwestward to a maximum of 20 km in western Montana. It was deposited in either (1) a large intracratonic rift basin, or (2) a passive margin (Moe et al. 1996). The Belt Supergroup is generally regarded to be of Middle Proterozoic age, but its exact span within the Middle Proterozoic is poorly constrained. The base of the Belt rocks dates approximately to 1500 Ma, although Armstrong and others (1987) prefer 1450 Ma. To the east in the vicinity of Neihart, Montana, Belt rocks rest on Archean age rocks of the Wyoming province that bear radiometric dates of about 1.9 Ga. On the present day western margin of the basin Belt rocks rest on the Priest River core complex with metamorphic U-Pb dates of 1.67 Ga from zircons. The upper limit of the Belt is contested and estimates range from about 1350 Ma to 900 Ma, despite these difficulties, the sequence within the Belt itself is relatively clear (Winston 1989).

New, sensitive high resolution ion microprobe (SHRIMP) U-Pb zircon analyses from two tuffs and a felsic (silica rich magma) flow in the middle and upper Belt Supergroup of northwestern Montana indicate that, in ascending stratigraphic order, 1) a tuff from the upper part of the Helena Formation at Logan Pass dates to 1454 +/- 9 Ma; 2) a regionally restricted porphyritic rhyolite of the Purcell Lava in the Snowslip Formation dates to 1443 +/- 7 Ma; and (3) the volcanic tuff in a transition zone between the Bonner Quartzite and the Libby Formation, located west of Glacier National Park, near Libby, Montana dates to 1401 +/- 6 Ma. These results indicate that all but the uppermost Belt strata (about 1700 m) were deposited over a relatively short period of geologic time of about 70 million years (Evans et al. 2000).

The original western and southwestern limits of Belt deposition are problematic. To the southwest in central Idaho lie the Middle Proterozoic Yellowjacket Formation and Lemhi Group, which contain rock types similar to those of the Belt, and may in fact, represent the southwest extent of the Belt basin. The western lip of the Belt terrane is marked by combinations of: 1) Late Proterozoic Windermere Supergroup unconformably overlying the Belt Supergroup, 2) Mesozoic accreted terranes (chunks of land smeared onto the continental margin during subduction), 3) Mesozoic age granitic complexes, 4) Eocene age extensional core complexes containing Early Proterozoic metamorphic rocks cut by Phanerozoic age granite, and 5) overlapping Miocene age Columbia Plateau flow basalt. Thus the Proterozoic western margin of the Belt basin cannot be identified definitively and was probably removed by latest Proterozoic rifting (Winston 1989).

There has been considerable discussion about whether the Belt Supergroup rocks were deposited in a basin that was marine (open ocean) or lacustrine (constrained, fresh water). Winston and Lyons (Belt III) describe hummocky cross-stratification in the deepest water facies as evidence of a broad shallow body of water. Kuhn (1987) describe herringbone crossbeds, lenticular and wavy bedding, desiccation cracks, stromatolites, and oscillation ripples as compelling evidence for marine deposition of the Grinnell Formation. Ackman (1988) contends that the debate itself is exemplified by the desiccation features complete with fining-upward layers (tidal deposits) versus the nearby sheetflood deposition analogous to modern playa mudflat environments. In other words, evidence points to either possibility in the Belt basin. Winston (1986) proposed a lacustrine intracratonic basin for the deposition of the Belt Supergroup, however, Hoffman (1988) suggested an episutural basin (a basin which persists between two converging land masses) underlain by oceanic crust trapped in the North America continent. Whether marine or lacustrine, what all researchers agree upon is that the Belt Supergroup was deposited in a largely shallow basin with a variety of nearshore and subaerially exposed deposition environments recorded in the rocks of Glacier National Park. For a general look at how different rock types present in the Belt Supergroup are interpreted, see figure 23.

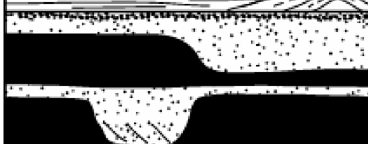
Sketch of sediment layer	Description of sediment	Depositional environment (interpreted)
	Even, mudcracked, graded, fine sand and silt-to-mud layers 0.3 to 3 cm thick.	Sheetflood flow across exposed mudflats followed by deceleration, suspension settleout and desiccation.
	Oscillation-rippled fine sand and silt lenses, capped by clay laminae, cut by mudcracks.	Wave transport of fine sand and silt, followed by clay settleout and desiccation.
	Graded silt-to-clay or silt-to-dolomitic clay layers 3 to 10 cm thick.	Episodic storm deposition followed by suspension settleout.
	Even, uncracked graded silt-to-clay couplets 0.3 to 3 cm thick.	Episodic suspension transport and settleout.
	Oscillation-rippled fine sand and silt lenses, capped by clay laminae, cracked and uncracked.	Wave transport of fine sand and silt, followed by suspension settleout.
	Interlayered or graded silt and clay laminations less than 0.3 cm thick.	Alternating silt and clay suspension settleout.
	Graded fine sand to dark mud layers with undulating scoured and loaded bases >3 cm thick.	Episodic transport of fine sand and mud by turbidity flows or storms and deposition on scoured or loaded mud surfaces.
	Graded fine sand to dark mud layers with undulating scoured and loaded bases, 0.3 to 3 cm thick.	Episodic transport of fine sand or mud by turbidity flows or storms and deposition on scoured or loaded mud surfaces.
	Micrite and dolomicrite without detectable siliciclastic laminations.	Aragonite or calcite precipitation, in places followed by dolomitization.
	Coarse- to fine-grained, sand and flat clasts, crossbedded and imbrecated at various angles.	Transport of coarse sand grains and scoured clasts by breaking waves.

Figure 23: Description of types of Belt Supergroup sediments with general depositional environment interpretations. Modified from Winston 1989.

The Middle Proterozoic – Prichard, Altyn, and Appekunny Formations. The Belt basin formed in response to high angle faulting and subsidence of large continental basement rock blocks, veneered by a discontinuous blanket of quartz sand. Dark quartzite, argillite, and carbonate of lower Belt formations including the Neihart, Prichard, Altyn and Appekunny mark the first and most extensive spread of the great Belt lake over basement crystalline rocks (Winston V/P 1989). As presently defined, the lower Belt has at its base the Neihart Quartzite (not exposed at Glacier National Park), which crops out above pre-Belt crystalline rocks and below the Prichard Formation in isolated patches westward across the basin. The Neihart is crossbedded probably reflecting fluvial deposition, however, ventifacts (any stone polished smooth by wind) and superbly rounded coarse quartz sand grains indicate some eolian activity. The Neihart represents a period of relative tectonic stability that preceded severe block faulting and initiation of the Belt basin. It also served later as a source of distinct quartz sand deposited in the following formations of the Belt basin (Winston V/P 1989).

Perhaps the most striking tectonic event during lower Belt deposition was block faulting along the Perry line, which uplifted the Dillon crustal block of the Archean Wyoming province to the south and downdropped crustal rocks to the north, forming the southern margin of the basin. The resulting eastern indentation is called the Helena embayment and comprises an arm of the Belt basin. The Prichard, Altyn and Appekunny Formations filled the main part of the basin that extended across western Montana, northern Idaho, and into eastern Washington (Figure 24) (Winston V/P 1989).

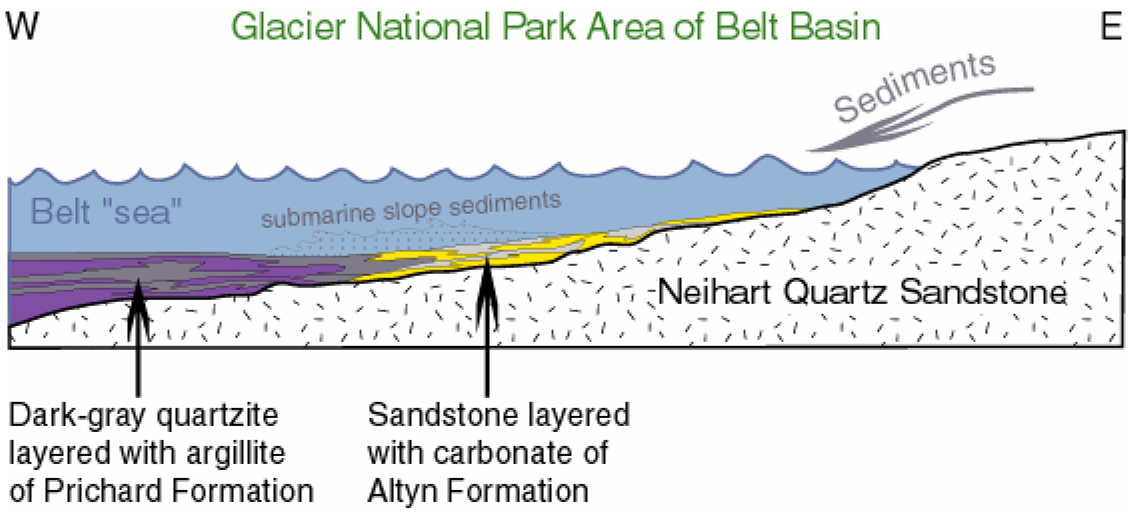


Figure 24: Depositional environment of the Prichard-Altyn Formation atop the Neihart Quartzite which acts as a sediment source.

The western side of Glacier National Park has the dark gray to nearly black quartzite and argillite as well as graded arenite beds containing pyrrhotite (iron sulfide), carbon and metamorphic biotite (flaky silicate mineral rich in potassium, magnesium and iron) of the Prichard Formation (Winston 1989; Winston V/P 1989). The Prichard is contemporaneous with the Altyn Formation to the east. Most of the Prichard formation is composed of subaqueous, sub-wave base, pelagic and turbidite deposits that pass to the deltaic facies in the west, reflecting sediment influx from the west of the Belt basin (Winston 1989). Interlayered carbon-poor and carbon-rich sediments indicate alternating oxygenated and anoxic, stagnant bottom conditions for the Belt basin. Paleocurrent data from western Prichard exposures indicate that the basin deepened eastward to the vicinity of the Purcell anticlorium, west of Glacier National Park, near Libby, Montana, where currents were deflected northwestward (Winston V/P 1989).

Simultaneously, sediments entering the eastern, more tectonically stable side of the basin, appear to be limited to coarse quartz sand, most likely reworked from the Neihart sediments. The coarse sand was mixed with carbonate mud forming the micritic limestone - sandstone mixture of the Altyn Formation, now present along the eastern side of Glacier National Park (Winston 1989; Winston V/P 1989). Micrite deposition along the eastern and northeastern margin of the basin probably reflects tectonic stability of the Archean craton to the east. This tectonic stability contrasts sharply with the southern boundary of the Belt basin, where the continued uplift of the Dillon block of the Wyoming province shed coarse the conglomerate of the LaHood Formation (present south of Glacier National Park) into the lower Belt basin and the western boundary, where uplift of the inferred continent bordering the basin on the west contribute most of the lower Belt sediments (Winston 1989).

The Altyn is overlain by black, green and red argillite and coarse-grained arenite of the Appekunny Formation. Deposition of the siliciclastic Appekunny Formation over the Altyn Formation may reflect the filling of the Purcell trough (sediments of the Prichard Formation) and spreading of suspended mud from the west across the eastern part of the basin (Winston V/P 1989).

The Middle Proterozoic – Ravalli Group (Grinnell and Empire Formations). From the relatively quiet, subaqueous, evenly layered deposits of the lower Belt formations, the upward transition into the ripplemarked, mudcracked redbeds of the Ravalli Group indicate progradation of playas and alluvial aprons from the west across the Belt basin

(Winston V/P 1989). In other words, the advance of continental environments, as opposed to marine, broadly across the basin (figure 25). The dominantly redbed Ravalli Group is represented in Glacier National Park by the Grinnell Formation, and the green argillite unit, the Empire Formation.

The well-exposed Grinnell Formation records the intermittent emergence of subaerial shoreline in the Belt basin (figure 26). The lowermost units of the Grinnell record subaqueous accumulation of sediments settling from suspension, and being partly reworked by wave oscillation. Upsection, desiccated mud, indicates exposure on extensive flats. Thin, rare beds of medium- to coarse-grained sand record flooding across these mud flats (Kuhn 1987). Unit 2 of the Grinnell is characterized by beds of coarser grained sands, with sparse mud and silt interbeds. This unit represents the distal reach of ephemeral floods carrying coarser sand in traction and silt and mud in suspension onto an expansive subaerially exposed sand and mudflat (Kuhn 1987). The next unit is dominated by desiccated mud sediments. Periodic floods, evidenced by climbing-ripple cross-stratification, ripped up mudcracked polygons and moved them as mudchips to be deposited as thin conglomeratic beds. The lowermost layers of unit 4 record laterally continuous silt beds, indicating sheetlike deposition over broad, flat surfaces, with the occasional microlaminated beds recording submergence. The increasing amounts of coarser grained sand upsection in unit 4 signal the return to a prograding sandflat, dominating unit 5 of the Grinnell Formation (Kuhn 1987). Westward thinning of the sand layers within the Grinnell indicates an eastern source terrain for medium- to coarse-grained sand in the Ravalli Group (Winston 1989).

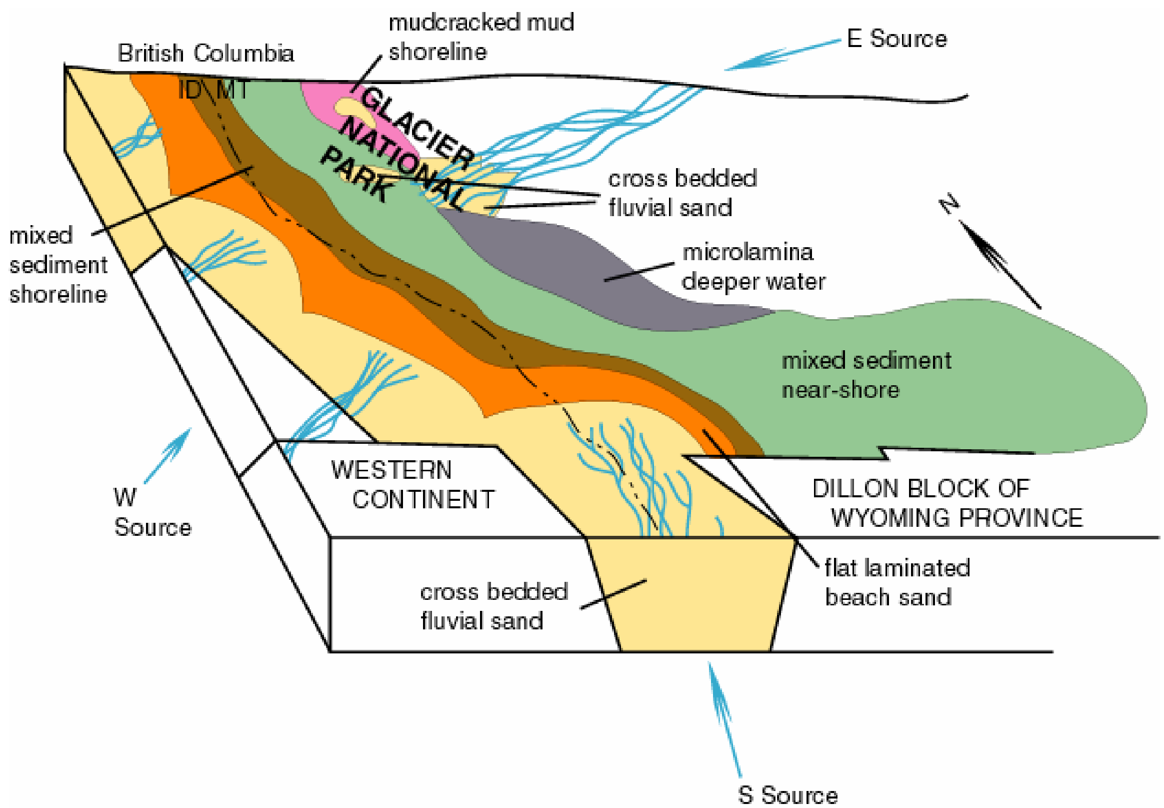


Figure 25: Conceptual depositional environment of the Ravalli Group, showing basinward progression from the western continent to the near shore and deeper water settings. Adapted from Winston 1989.

The Empire Formation is dominated by shallow water rippled and mudcracked rocks (Winston and Lyons Belt III 1993). With deeper water facies increasing gradually upsection. The green argillite of the Empire Formation on the eastern side of the Belt basin becomes calcareous and tan weathering upsection. Authigenic (formed in place) carbonate mud mixed with clay is concentrated in the upper units of the Empire. Carbonate diagenesis formed layers of oblong carbonate pods that also characterize this interval (Winston V/P 1989). The upsection change from redbeds of the Grinnell Formation into greenbeds at the base of the Empire Formation signals submergence and the onset of the Middle Belt carbonate progression across the Belt basin (Winston V/P 1989). The first change recorded in the Empire is water deepening and a shift from oxidizing to reducing conditions. The second dramatic change recorded in the Empire is the decline of the supply of terrigenous clastic material, from both the eastern and southwestern sources. This clastic supply was replaced by carbonate mud (Link Belt III 1993).

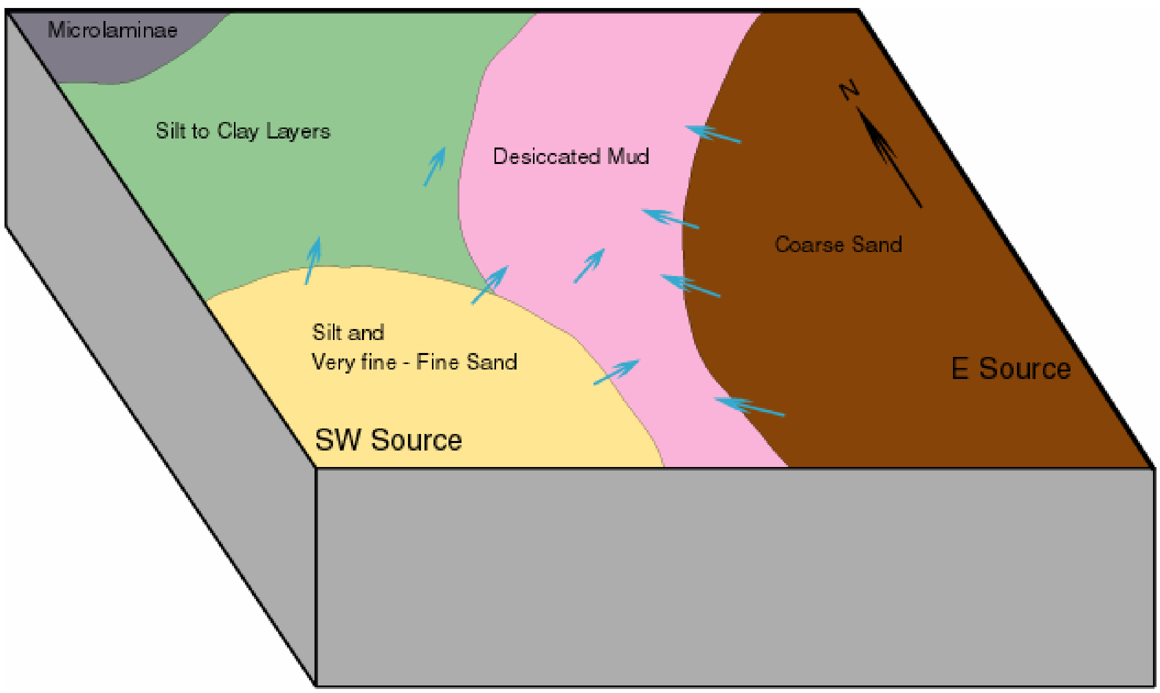


Figure 26: Generalized sediment type distribution during Grinnell Formation deposition. Arrows indicate basinward flow across alluvial sand and mudflats into ponded water. Modified from Kuhn 1987.

The Middle Proterozoic – Middle Belt Carbonate (Helena Formation). The Helena Formation, consisting of a thick sequence of interlayered siliciclastic and carbonate cycles, represents the eastern facies of the Middle Belt carbonate. The Helena contains deeper water sediments than the underlying Empire Formation, recording an overall transgression of the sea onto the shoreline. The sediments fine and thin eastward. This is interpreted to record periodic turbidite underflows and interflows from the west that spread out above the sub wave base floor of the Belt basin (Winston V/P 1989) (figure 27). The sediment may have come mostly from floods that drained the western side of the basin. Along the eastern side of the basin, authigenic carbonate mud precipitation overwhelmed siliciclastic influx forming the carbonate mud sediments there. The occasional siliciclastic beds in the Helena Formation record the episodic expansion of the Belt "sea" across subjacent exposed surfaces in response to wetter climatic conditions. The water was at these times undersaturated with respect to calcium carbonate, whereas a shift to a more arid climate, with accompanying evaporation, would change the water chemistry to supersaturation of calcium carbonate and thus induce precipitation of carbonate (Winston V/P 1989). Erosion scoured the carbonate flats exposed as the shoreline further regressed seaward, shifting the carbonate deposition westward.

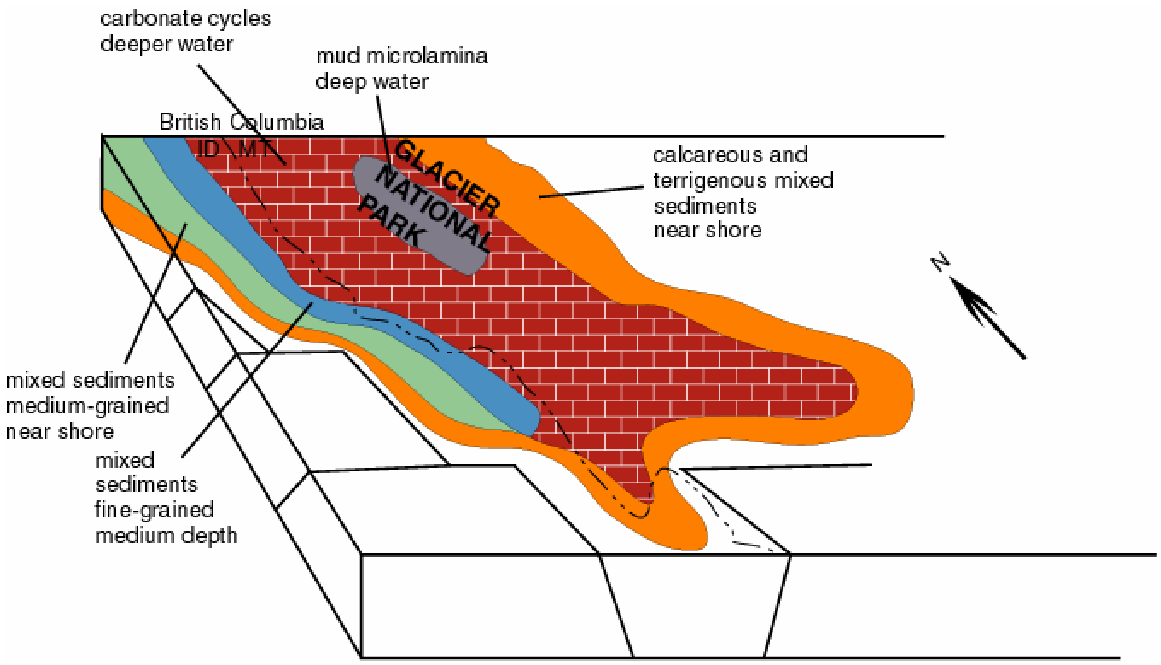


Figure 27: Middle Belt Carbonate depositional facies. Modified from Winston 1989.

The outcrop extent of the Helena Formation is immense. Today is extends from eastern Washington to Glacier National Park, which shows that, during its maximum spread, the Belt "sea" had a fetch of more than 300 km, and this is not including the western edge of the basin, now lost as described earlier. This inferred basin is comparable in scale to the northern part of the modern Caspian Sea (Winston and Lyons Belt III 1993). Hummocky cross stratification in the Helena Formation sediments, inferred to be from the deepest water deposits, indicate that even during the greatest sea extent, the bottom was within reach of occasional storms. Again, this is analogous to the broad, shallow modern day Caspian Sea. The very small dips of the individual beds of the Helena Formation indicate a very level bottom, without any indication of a carbonate shelf fringed by a reef.

The Middle Proterozoic – Missoula Group (Snowslip, Shepard, and Mt. Shields Formations, Bonner Quartzite, and McNamara Formation). Above the Helena Formation is a thick sequence dominated by red argillite, pink arenite and green argillite, containing some carbonate and black argillite intervals. These rocks comprise the formations of the Missoula Group, which are from bottom to top: the Snowslip Formation, the Shepard Formation, the Mount Shields Formation, the Bonner Quartzite, and the McNamara Formation and several others not exposed at Glacier National Park (the Garnet Range Formation, and the Pilcher Formation). Major transgression and regression of the Belt "sea" during Missoula Group deposition resulted in four progradational sequences separated by transgressive (highstand) sequences (figure 28). The Snowslip and Mount Shields (member 2) Formations, the Bonner Quartzite and upper McNamara Formation comprise the major progradational sequences whereas the remaining rocks comprise the transgressive (deeper water) sequences. The progradational deposits probably reflect periods of tectonic uplift in the Belt source area and a corresponding downdrop of the basin (Ackman 1988) (figure 29).

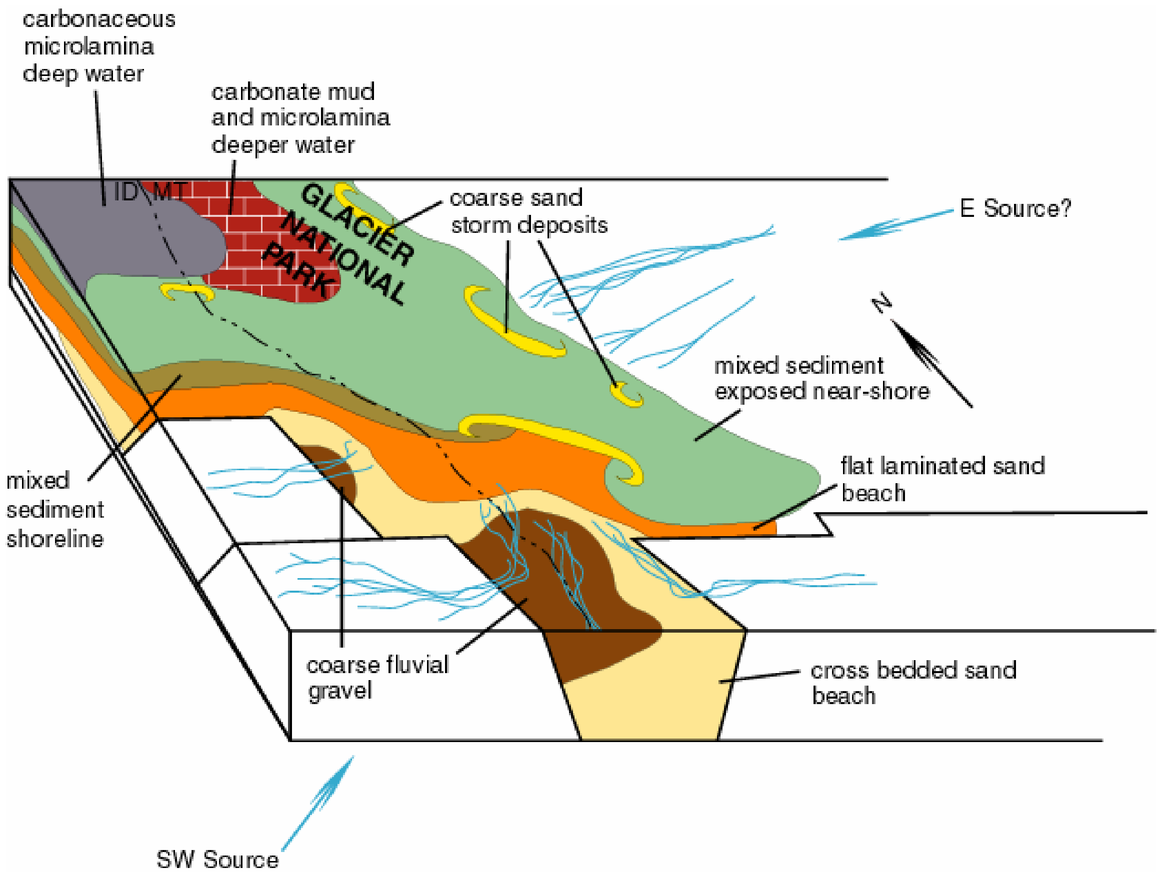


Figure 28: Conceptual depositional environment of the Missoula Group up to the Garnet Range. Deeper water facies to the northwest. Adapted from Winston 1989.

The red argillite of the Snowslip that characterizes the base of the Missoula Group passes progressively westward to green argillite, thus the westward correlative of the Snowslip becomes green and black argillite, overlain by black argillite and carbonate of the Shepard Formation correlative unit (Winston 1989). This indicates deeper water environments progressively westward in the Belt basin at the time of the initial Missoula Group deposition. The red argillite also records the regressional environment that signaled the end of Helena Formation deposition as the mudflats were exposed. Alluvial aprons of great areal extent bordered the uplifted continent southwest of the Belt basin and sloped gently into the Belt "sea". During Missoula Group deposition, a northwest trending depositional axis, or high point, resulted in pronounced thinning and fining of the sediments northwestward (Ackman 1988). Missoula Group rocks record the changing of "sea" level along this alluvial apron. During lowstand conditions, the variety of shoreline environments including: 1) braided flood channels filled with cross bedded sand, 2) shallow sheetflood tracks, 3) shallow ponds with mud layers and small-scale ripplemarks, 4) exposed, desiccated mudflats, vulnerable to rip-ups during occasional floods, 5) broad sandflats, and 6) beaches, all moved progressively basinward whereas in transgressive situations, these environments retreated from the basin towards the craton.

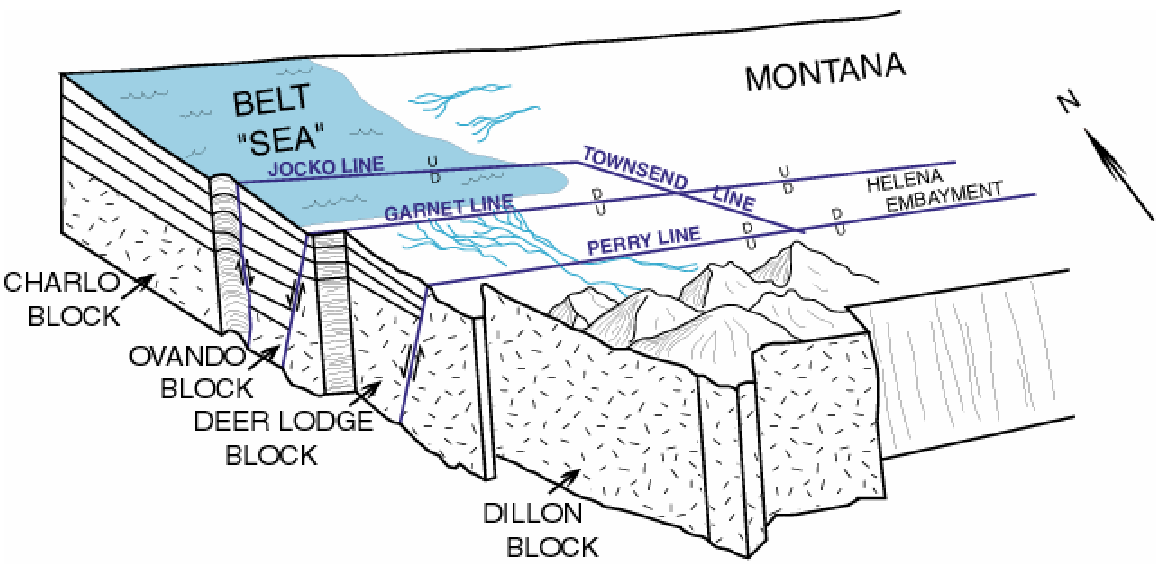


Figure 29: Showing inferred Proterozoic structure of the Belt basin during Missoula Group deposition. Note numerous faults providing means for uplift south of the basin (sediment source area). "U" indicates the uplifted side of the fault, "D", the downdropped side. Modified from Ackman 1988.

The Snowslip Formation is representative in its record of the variety of environments present during Missoula Group Formation. Members 1, 2, 4, and 6 generally record deposition on distal sandflats, exposed mudflats and submerged mudflats during shoreline transgression, whereas members 3 and 5 represent the basinward progradation of the alluvial aprons (Ackman 1988; Winston 1989). Members 3 and 5 contain laminated sand, sheetflood deposits, desiccation cracks and abundant mudchips, ripped up during occasional floods over the increasingly exposed apron (Ackman 1988). Pillow basalts of the Purcell Lava poured out into the waters of the Belt "sea" and were succeeded by a regressive sequence of red microlamina beds capped by lenses of coarse sand and larger clasts and stromatolites (Winston 1989).

The base of the Shepard Formation is marked by thin beds of coarse sand and larger clast sediments and stromatolites. These deposits were overlain by dark carbonaceous beds, signaling deeper water and a transgression of the great Belt "sea" (Winston 1989). In the sea itself, occasional carbonate layers, such as those present in the Shepard Formation were incorporated into mud-rich layers when the water chemistry was supersaturated with respect to calcium carbonate, often prevalent during lowstand conditions (Winston 1989). The middle and upper parts of the Shepard are composed of calcareous varieties of sediments. During Missoula Group deposition, when terrigenous sediment was not included in carbonate deposits, this is interpreted to indicate quiet, protected environments of the Belt "sea" at this time, not necessarily deeper water conditions (Ackman 1988).

Carbonate mud in the northwestern part of the basin was succeeded by widespread development of microlaminated, fine-grained sediment record expansion, transgression and freshening of the Belt "sea" high in the Mount Shields Formation. The pink crossbedded, coarse-grained sand of the Bonner Quartzite records yet another progradational advance of the alluvial apron complex from the southwest across Montana, north into Canada. The argillite of the overlying McNamara Formation records another transgressive drowning of the alluvial aprons and advance of playa flats. A wedge of sand sediment high in the McNamara marks progradation of a final alluvial apron complex in the southern part of the basin from that direction (Winston 1989).

Regional and Local Geologic History

The Paleozoic Era - Cretaceous Period. What happened geologically in the time between the Mid to Late Proterozoic and the Cretaceous in Glacier National Park remains a mystery because no rocks from that time are present. However, rocks in the region surrounding Glacier National Park record a fascinating series of events which will be described in a broad regional context briefly, below.

Late Proterozoic rifting created a new continental margin along western North America. During the Late Precambrian through the Cambrian, thousands of feet of shallow-water, marine sediments accumulated along a passive plate-tectonic margin on the western side of the Transcontinental Arch, an upland that stretched from northern Minnesota southwestward across Nebraska, Colorado and northwestern New Mexico (figure 30) (Speed, 1983; Sloss, 1988; Graham et al. 2002).

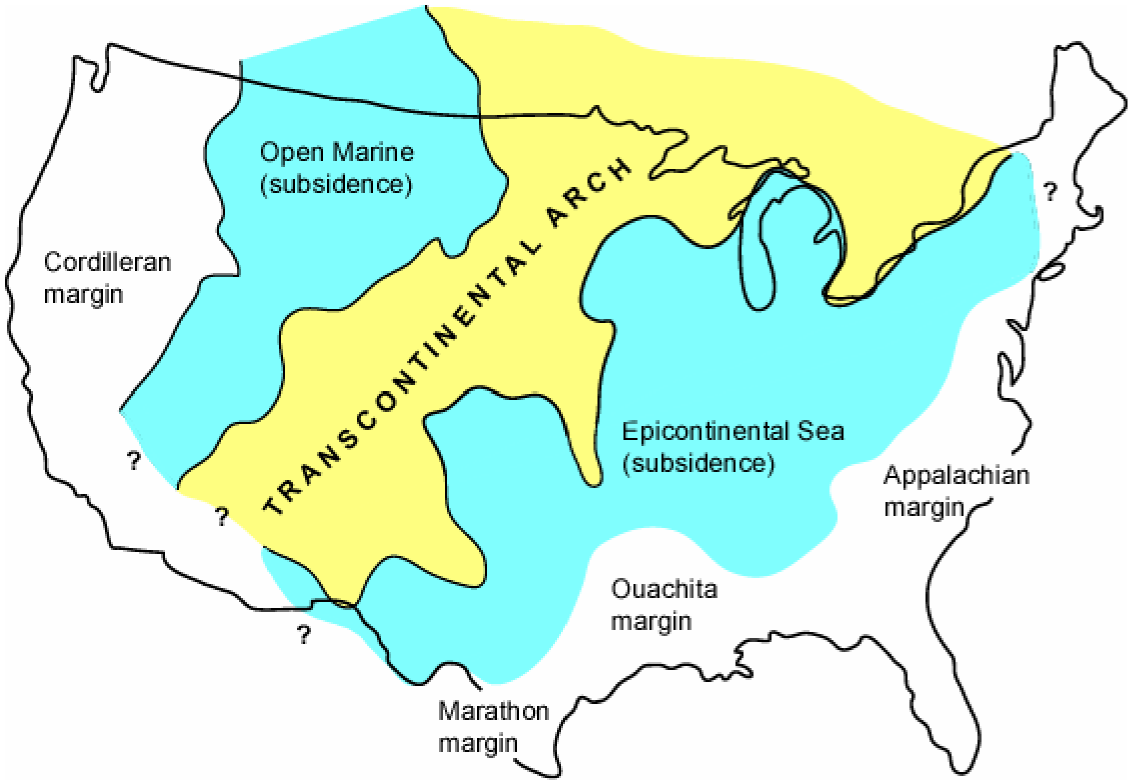


Figure 30: General paleogeographic map of North America at the time of the Sauk Sequence (latest Proterozoic to Early Ordovician Periods). Most, if not all, of the Transcontinental Arch's surface (yellow shading) was exposed at this time. Relatively shallow, epicontinental seas (blue shading) flooded onto the eastern margin of the continent, and an offshore, submarine ramp extended westward from the western margin of the Arch. Later orogenies (Appalachian, Ouachita, Marathon, and Cordilleran) obscured the eastern and western extent of the seas. Modified from Sloss, 1988.

Throughout North America, medium- to coarse-grained quartz sandstones dominate basal Cambrian units. Pebbles and cobbles of local origin are also common. These sandstones represent a combination of dune and beach processes subjected to unimpeded wind erosion since land plants did not exist (Dott et al., 1986). Subsequent shallow marine planation was insufficient to erase all the characteristics of an eolian landscape. From

Wisconsin to what is now Nevada, sand was eroded from these isolated highlands and deposited into these Cambrian-age seas. With burial, the sand was cemented into sandstone and orthoquartzite (silica-cemented quartz sand) (Graham et al. 2002) The Cambrian age Flathead sandstone, in the areas surrounding Glacier National Park, was formed in the shallow marine environments west of the isolated highlands.

During the Sauk Sequence (Cambrian age), a structural inversion transformed the region into one of compression rather than extension, and for the rest of the Paleozoic and all the way into the Early Tertiary (from about 570 Ma to 35 Ma), Western North America was influenced by compressional tectonics (Figure 31)(Stone, 1986). The transgression of the Sauk Sea onto the craton was coincident with southwest-northeast directed compressive stress fields.

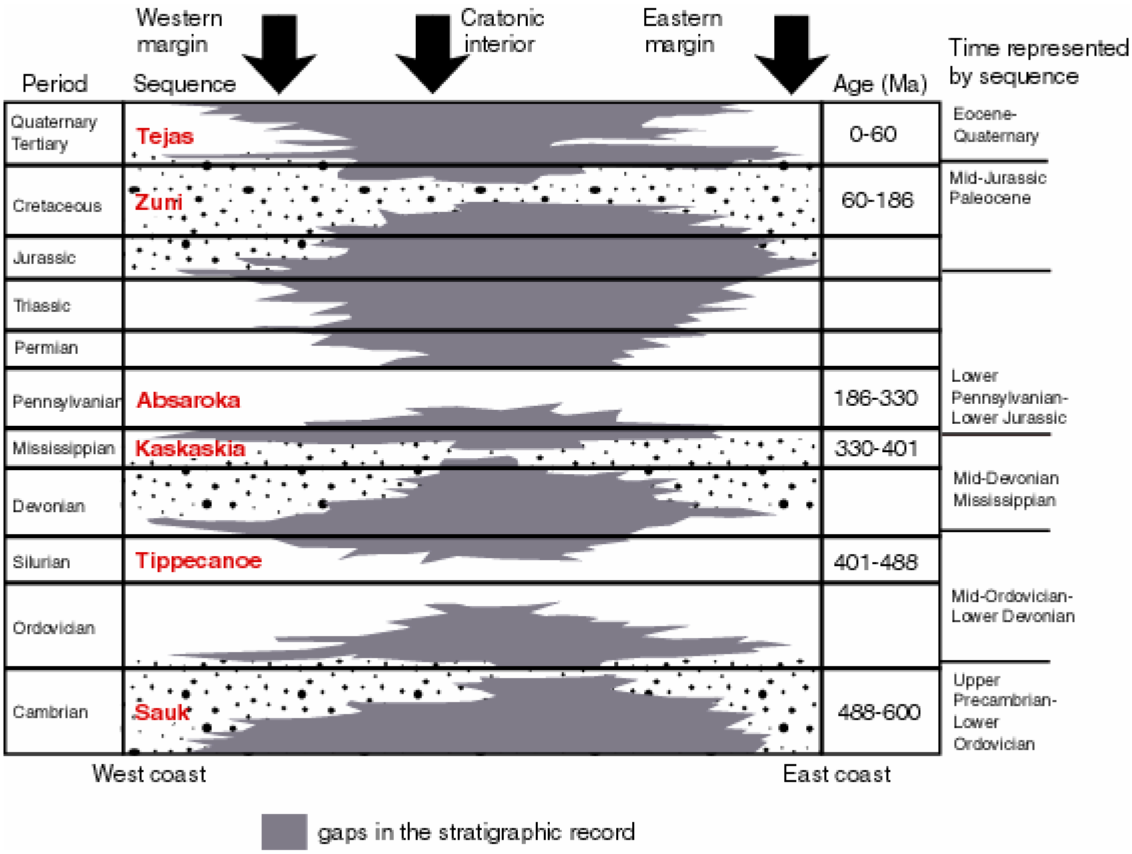


Figure 31: Major cratonic sequences defined by disconformities and their relationship to geologic time. Times of maximum "gaps" in the stratigraphic record are times of maximum regression of the shoreline off the craton while times when there are no gaps in the stratigraphic record are times of maximum transgression when marine water inundated all, or part, of the North American craton. Ma: Millions of years ago. Modified form Sloss, 1988.

After the craton-wide regression at the end of the Sauk Sequence, sea level began to inundate the cratonic margins once again at the beginning of the Tippecanoe sequence (Silurian age). Coincident with the craton-wide transgression, the east coast of North America experienced compressional tectonics as the Taconic Orogeny impacted New England (figure 32). The Taconic Orogeny eventually stretched southward and blended into the Appalachian Orogeny. At the beginning of the Ordovician Period, dolomite (a carbonate with a high magnesium content) was deposited in vast supratidal to subtidal carbonate mud flats. Scattered islands on the Transcontinental Arch protruded from the mud flats (Ross and Tweto, 1980). As the shoreline receded from the Lower Ordovician carbonate platform, the limestone was buried by subtidal to intertidal and supratidal carbonate facies migrating westward (Poole et al., 1992; Graham et al. 2002).

Eon	Era	Period	Epoch			Life Forms	N. American Tectonics
Phanerozoic (Phaneros = "evident"; zoic = "life")	Cenozoic	Quaternary	Recent, or Holocene		Age of Mammals	Modern man	Cascade volcanoes
			Pleistocene	1.6		Extinction of large mammals and birds	Worldwide glaciation
		Tertiary	Pliocene	5.3		Large carnivores	Uplift of Sierra Nevada
			Miocene	23.7		Whales and apes	Linking of N. & S. America
			Oligocene	36.6			Basin-and-Range Extension
			Eocene	57.8		Early primates	Laramide orogeny ends (West)
			Plaeocene	66.4			
	Mesozoic	Cretaceous		144	Age of Dinosaurs	**Mass extinctions** Placental mammals Early flowering plants	Laramide orogeny (West) Sevier orogeny (West) Nevadan orogeny (West)
		Jurassic		208		First mammals Flying reptiles	Elko orogeny (West) Breakup of Pangea begins
		Triassic		245		First dinosaurs	Sonoma orogeny (West)
	Paleozoic	Permian		286	Age of Amphibians	**Mass extinctions** Coal-forming forests diminish	Super continent Pangea intact Ouachita orogeny (South) Alleghenian (Appalachian) orogeny (East)
		Pennsylvanian		320		Coal-forming swamps Sharks abundant	Ancestral Rocky Mts. (West)
		Mississippian		360		Variety of insects First amphibians	
		Devonian		408	Fishes	First reptiles **Mass extinctions** First forests (evergreens)	Antler orogeny (West) Acadian orogeny (East-NE)
		Silurian		438		First land plants	
		Ordovician		505	Marine Invertebrates	**Mass extinctions** First primitive fish Trilobite maximum Rise of corals	Taconic orogeny (NE)
		Cambrian		570		Early shelled organisms	Avalonian orogeny (NE) Extensive oceans cover most of N.America
Proterozoic ("Early life")		Precambrian		2500		1st multicelled organisms Jellyfish fossil (670Ma)	Formation of early supercontinent First iron deposits Abundant carbonate rocks
Archean ("Ancient")				~3800		Early bacteria & algae	Oldest known Earth rocks (~3.93 billion years ago)
Hadean ("Beneath the Earth")				4600		Origin of life? Formation of the Earth	Oldest moon rocks (4-4.6 billion years ago) Earth's crust being formed

Figure 32: Geologic time scale. Red lines indicate major unconformities between eras. Absolute ages shown are in millions of years and are from the United States Geological Survey (USGS) time scale found at http://geology.wr.usgs.gov/docs/usgsnps/gtime/timescale.html.

As regression continued, Middle Ordovician carbonates and sandstones were stripped from the exposed continent and an erosional unconformity developed between the eroded units and the overlying Upper Ordovician rocks. Throughout the west, thick Ordovician sandstone beds overlie shallow-marine dolomite units (Ross and Tweto, 1980). The Late Ordovician lasted about 20 million years, and during this time, the sea advanced onto the continent once again. The continent became submerged to a greater extent than in any previous Paleozoic time (Ross and Tweto, 1980; Poole et al., 1992). A shallow sea inundated the source areas of quartz sand and re-established carbonate production (Graham et al. 2002).

The end of the Ordovician Period (438 Ma) is marked by one of the five most extensive mass extinctions of all time. The other four occurred at the end of the Devonian Period, the end of the Permian Period, at the close of the Triassic, and of course, at the Cretaceous – Tertiary, or "K–T", boundary (figure 32). Reasons for these mass extinction events are still being debated and the cause of one event isn't necessarily repeated for the others. For example, not all mass extinctions are marked by extraterrestrial impact craters and a high level of iridium as is found at the Cretaceous – Tertiary boundary. Meteor impacts, glaciation, sea level rise, sea level fall, and global warming have all been offered as mechanisms leading to extinction, but for the Ordovician Period, at least, no definitive explanation has been accepted. Sporadic episodes of regression began to be felt on the craton by the end of the Tippecanoe sequence (Silurian age) (figure 31) (Sloss, 1988). Regression reached a climax in the earliest Devonian time when the entire craton interior emerged from the Tippecanoe Sea (Graham et al. 2002).

By the beginning of the Devonian Period (end of Tippecanoe sequence), the seas that had covered most of the continent had receded, or regressed, and the shoreline was far to the west. However, this quiet scene suddenly changed. With the initiation of the Kaskaskia Sequence in the Middle Devonian Period, the first pulses of the Antler Orogeny in the west and the Acadian Orogeny in the east (part of the Appalachian Orogeny) (figure 32, figure 31) began to be felt as landmasses accreted onto both the western and eastern borders of North America. To the west of Montana, a subduction zone formed. As lithospheric plates collided against one another, their rocks were bent, folded, and thrust-faulted into a north-south trending mountain range stretching from Nevada to Canada. The Roberts Mountains Thrust marks the easternmost thrust sheets generated by the Antler Orogeny. By the end of the Devonian, great inland seas again covered the continent, and a sea inundated the area east of the mountains (figure 33) (Johnson et al., 1991; Graham et al. 2002). In the areas around Glacier National Park, the Devonian age is recorded in the Three Forks, Jefferson and Maywood Formations .

The sea that transgressed onto the continent in the Late Devonian Period did not have a good circulation system. We know this because black, organic-rich shales were deposited over immense areas in the eastern, southern, western, and northwestern parts of the United States (Sloss, 1988). When ocean circulation provides well-mixed and oxygenated seawater, a variety of bottom feeders and bacteria degrade most, if not all, organic matter that drifts down through the water column, Devonian shales would later generate much of the oil that would be discovered in this country (Graham et al. 2002).

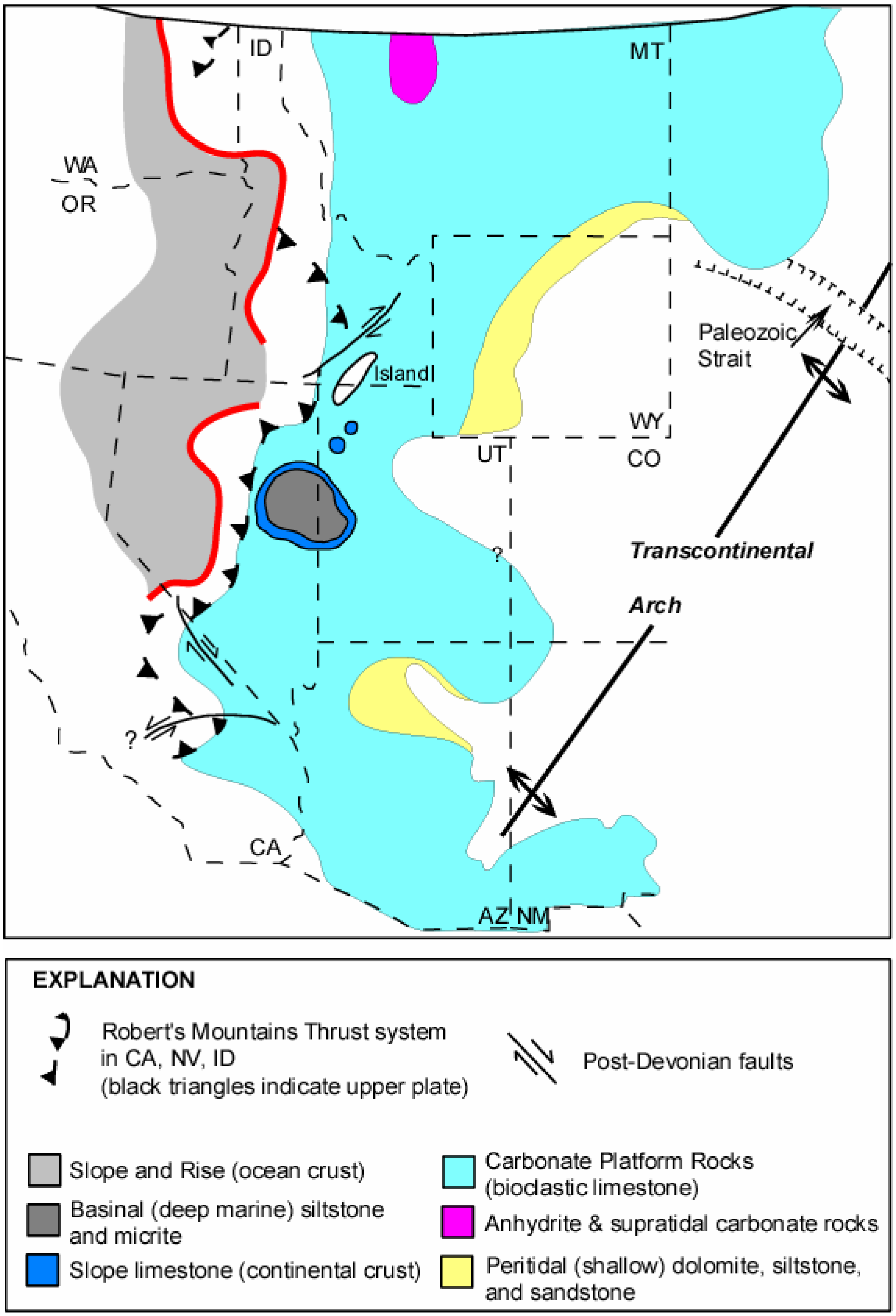


Figure 33: Distribution of lithofacies during the Upper Devonian, Frasnian stage (*jamieae* conodont biostratigraphic zone) of western United States. The red line is the strontium isotope line wherein $^{87}Sr/^{86}Sr = 0.706$ and is interpreted to represent the break between continental and oceanic crust. Modified from Johnson et al., 1991.

The Late Devonian was accompanied by one of the five major extinction events in the geologic record (figure 32). Unlike the Ordovician, a large impact crater has been found that correlates to the Devonian extinction at 367 Ma. The crater, discovered in Siljan, Sweden, measured 52 km (31 mi) in diameter and struck Earth 368 Ma (Raup, 1991). Like the Ordovician, however, the Late Devonian was also a time of world-wide transgression that perhaps destroyed specific habitats and thus, led to extinction. As with the extinction at the end of the Ordovician Period, the cause or causes for the mass extinction during the Late Devonian are not clear (Graham et al. 2002).

As the sea became shallower during the regression that followed the Antler Orogeny and the Kaskaskia Sequence came to an end, habitats dwindled and animal life became more restricted. The shoreline again receded from the area so that by the end of the Kaskaskia Sequence about 320 million years ago (early Pennsylvanian time), soils were beginning to form in low lands and the higher areas were being worn down by erosion (Graham et al. 2002). The Rocky Formation, present in the northern Whitefish Range, west of Glacier, are Pennsylvanian- Permian in age.

Throughout the Paleozoic Era, Europe, Africa, and South America were approaching North America as the two great landmasses, Laurasia and Gondwana, collided. The ancient continent of Gondwana included Australia, Antarctica, Africa, South America, and India south of the Ganges River, plus smaller islands. Laurasia, located in the northern hemisphere, is the hypothetical continent that contained the present northern continents. Aggressive tectonism in the Pennsylvanian Period built mountains in the Western U.S. with as much as 3,000 m (10,000 ft) of relief, herein called the Ancestral Rocky Mountains (De Voto, 1980B). The Ancestral Rocky Mountains were less extensive than the Rocky Mountains we know today (Graham et al. 2002). They did not extend as far north as the Glacier National Park area.

As the final suturing of the continents took place at the end of Pennsylvanian time and the beginning of the Permian Period (about 286 Ma), tectonic shockwaves rippled into the North American interior and renewed uplift in the central Rocky Mountain area. The trend of the Transcontinental Arch was superimposed on the Ancestral Rockies structures in the late Permian (Peterson and Smith, 1986; Gregson, 1992). With re-emergence of the Transcontinental Arch, erosion created a widespread unconformity between the Permian age formations and the Triassic age formations (Graham et al. 2002).

On the western margin of the continent, oceanic rocks were thrusted eastward over the continental margin during the *Sonoma Orogeny* at the end of the Paleozoic Era about 245 Ma (figure 32) (Silberling and Roberts, 1962). Subduction of the oceanic crust beneath the continental crust caused melting and magma generation and formed a volcanic island chain on the western margin of the continent in the Triassic. The close of the Permian Period also brought the third, and most severe, major mass extinction of geologic time (figure 32). Although not as well known as the extinction event that exterminated the dinosaurs at the end of the Mesozoic Era, the Permian extinction was much more extensive. The most recent hypothesis on the Permian event suggests that a comet, about four to eight miles in diameter, slammed into Earth (Becker et al., 2001). Such an impact may have triggered vast volcanic eruptions that spread lava over an area two-thirds the size of the United States (Graham et al. 2002).

Powerful updrafts would have carried dust and grit swirling into the upper atmosphere. The particulate matter would have reflected and scattered sunlight, resulting in years of global cooling with freezing temperatures even during summertime. A recent example of this type of global cooling occurred in 1816, "the year without a summer," following a volcanic explosion in Tambora, Indonesia, in 1815. The sulfuric emissions from the volcanoes would have mixed with atmospheric water to produce downpours of acid rain. The combination of these variables could have caused the thousands of species of insects, reptiles, and amphibians to die on land, and in the oceans, coral formations to vanish, as well as snails, urchins, sea lilies, some fish, and the once-prolific trilobites. The catastrophe wiped out 300 million years of living Earth history! Five million years later, at the dawn of the Mesozoic Era, the oceans began to evolve into the chemistry of the modern oceans and on land, there would soon be a new order, one that the world had never seen before and will never see again (Graham et al. 2002).

During the Triassic, the supercontinent Pangea reached its greatest size (figure 34). All the continents had come together to form a single landmass that was located symmetrically about the equator (Dubiel, 1994). To the west, explosive volcanoes arose from the sea and formed a north-south trending arc of islands along the border of what is now California and Nevada (Christiansen, et al., 1994; Dubiel, 1994; Lawton, 1994). Eruptions from these volcanoes were similar to the eruptions from Mt. Vesuvius that destroyed Pompeii in 79 A.D., only more violent (Christiansen et al., 1994). Along with lava, stratovolcanoes eject pyroclastic material ranging in size from volcanic *bombs* (rocks cobble size and greater) to volcanic ash (Graham et al. 2002).

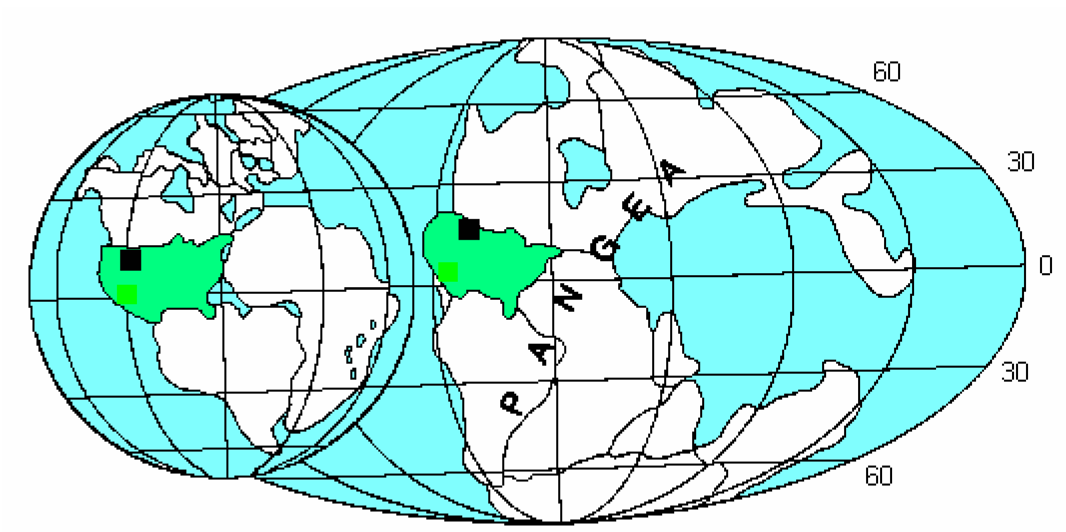


Figure 34: Schematic reconstruction of Pangea showing the approximate Permian location of Glacier National Park (black box) and the United States (green). The rest of the land mass is shown in white. Round inset: Partial reconstruction of Pangea depicting different orientation and thus rotation of the area. From Dubiel et al., 1996.

Feeding the central chamber of these volcanoes was liquid magma that would be either extruded onto the surface or would solidify beneath this volcanic mountain range in large igneous masses, called plutons. The Sierra Nevada mountain range in California exposes a remnant of these large plutons (collectively called the Sierra Nevada batholith) that were once buried deep within Earth but are now exposed at the surface. The magma originated as continental crust melted above an east-dipping slab of oceanic crust that subducted beneath the continent in a collision between the North America part of Pangea, traveling west, and the oceanic Farallon lithospheric plate, traveling east. Magma, being more buoyant than the surrounding rocks, rose to the surface, melting the overlying rock layers, and turning groundwater to the steam that accelerates volcanic explosions (Graham et al. 2002).

During the Late Permian (about 245 to 250 Ma), a marine basin formed in a trough stretching from Canada to New Mexico and from western Utah to central Nebraska, called the Western Interior Basin. Following the Permian, the Triassic Period opened onto a desolate scene. The atmosphere and the oceans were slowly recovering from the catastrophic extinction event. Pig-sized mammalian reptiles, called *Lystrosaurus*, began to rule Earth, but the Mesozoic Era would soon become the "age of dinosaurs" (Dubiel 1994; Graham et al. 2002).

In the Early Triassic (240 to 245 Ma), volcanic activity decreased on the western margin of the supercontinent (Christiansen et al., 1994). The depositional environments in the Early Triassic represent a transition from marine and marginal marine environments in east of the continental margin. Upper Triassic rocks tell a different story from the Lower Triassic strata. Continental rocks of the Western Interior form a complex assemblage of alluvial (stream), marsh, lacustrine (lake), playa (dried lake), and eolian (wind) deposits (Stewart et al., 1972B). Throughout the region, layers of bentonite, montmorillinite clay formed from the alteration of volcanic ash, are interlayered with clastic sediments. The bentonite layers indicate a period of renewed volcanism to the west and can be radiometrically dated to provide absolute ages for the adjacent rocks (Christiansen et al., 1994; Graham et al. 2002).

Increased volcanic debris in the Upper Triassic rocks marks the first appearance of abundant Mesozoic volcanic ash in the continental interior (Christiansen et al., 1994). Granitic plutons in southern California, central Nevada, on the border of Idaho and Oregon, and in central Washington, along with volcanic rocks of this age, are remnants of a magmatic arc that developed above an east-dipping subduction zone. Uplift associated with the evolution of the magmatic arc along the west coast may be related to drainage reversals and extensive lacustrine and marsh deposits in Utah and Nevada (Dubiel, 1994). As highlands rose above sea level and the continental crust was deformed, rivers that once flowed northwesterly now reversed their direction and flowed to the south and west. Lakes formed at the base of the rising highlands (Dubiel, 1994). The monsoon precipitation that characterized the Early Triassic depositional systems disappeared in the Late Triassic. Pangaea began to break apart in the Late Triassic and Early Jurassic (about 195 to 216 Ma), and the monsoon climate changed as the western North America interior slowly rotated into a position farther north of the equator. The area would soon become a Sahara (Graham et al. 2002).

The western interior during the Jurassic was a time of extensive eolian sand seas-- called *ergs*--similar to the Sahara/Sahel regions today (figure 35). The region was located about 18^{o} north latitude at the beginning of the Jurassic (about 208 Ma) and about 30-35^{o} north latitude at the end of the Jurassic (about 144 Ma) (Kocurek and Dott, 1983; Peterson, 1994). This is the latitude of the northeast trade wind belt where cool, dry air descends from the upper atmosphere and sweeps back to the equator in a northeast to southwest direction. The cool, dry air becomes warm, dry air that can pick up additional moisture. This is the latitude of intense evaporation. Most modern hot deserts of the world occur within the trade wind belt (Graham et al. 2002).

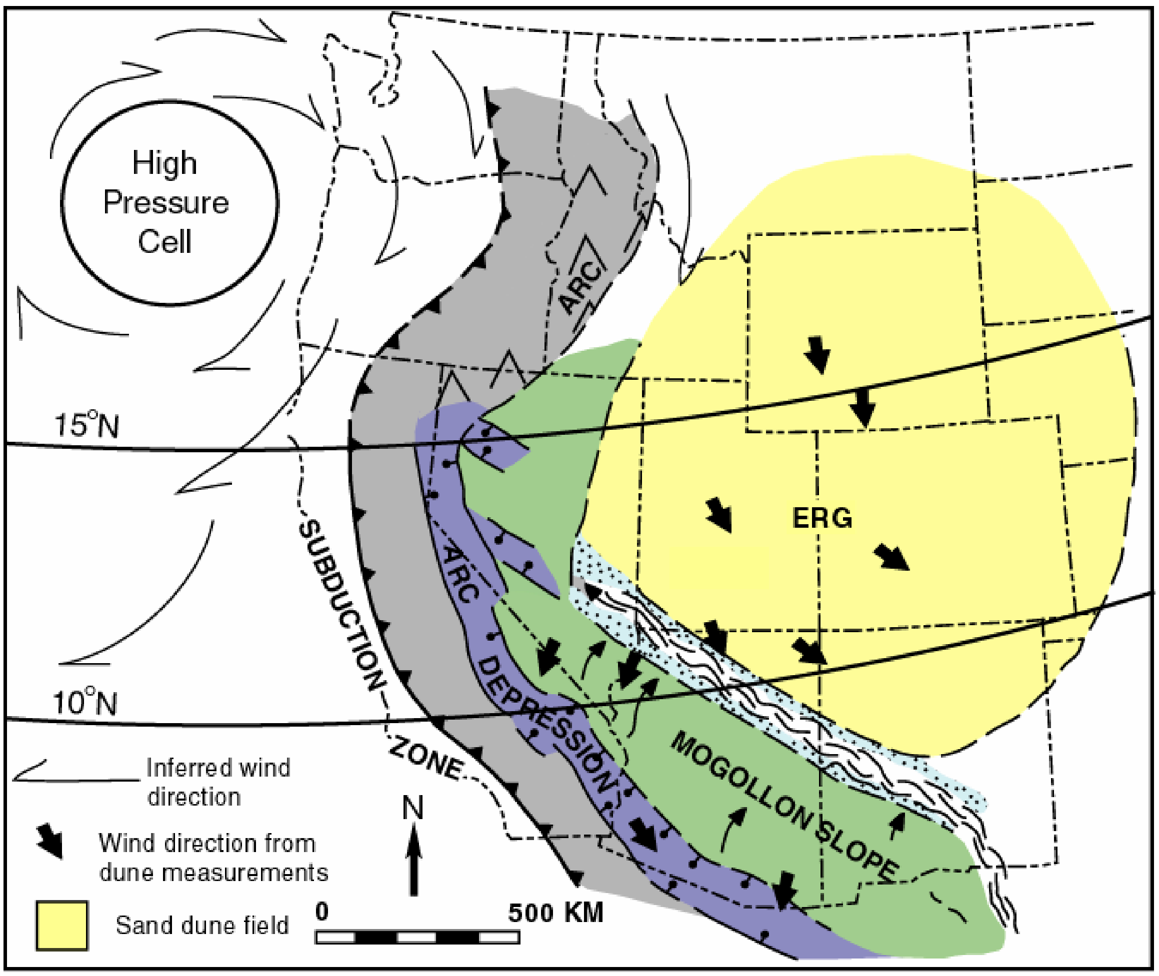


Figure 35: Lower Jurassic paleogeography of the western U.S. Thick arrows indicate eolian transport of sand. Thin arrows on Mogollon Slope indicate fluvial transport of sediments. Gray, shaded area indicates the location of the volcanic arc. Solid saw teeth indicate the location of the subduction zone with the teeth on the overriding, upper lithospheric plate. Glacier National Park is just north of this sand sea. Modified from Lawton, 1994.

In the Sahara, the world's largest desert, only 10% of the surface is sand-covered. The Arabian Desert, Earth's sandiest desert, is only 30% sand-covered. The Jurassic deserts that inundated the Colorado Plateau for roughly 40 million years (not counting the time represented by erosion) contained sand dunes that may be the largest recorded in the history of Earth (Kocurek and Dott, 1983). These were ergs that formed on a coastal and inland dune field affecting the present areas of southern Montana, eastern Utah, westernmost Colorado, southwest Colorado, northeastern Arizona, and northwestern New Mexico (Kocurek and Dott, 1983; Peterson, 1994). The volume of sand in these systems was enormous. Where did it come from? And how did it get there? Wind directions, measured from cross-bedding, indicate that sand migrated from the north to the south, but the trade winds blew to the west-southwest. Why did the sand migrate to the south rather than to the west? Answers to these questions lie with the plate tectonic setting, the climate, and the potential sediment sources (Graham et al. 2002).

During the 100 million years of the Jurassic Period, episodic incursions from the north brought shallow seas flooding into Wyoming, Montana, and a northeast-southwest trending trough on the Utah/Idaho border. The Jurassic western margin of North America was associated with an "Andean-type" margin where the eastward subduction of the seafloor gave rise to volcanism similar to that found in today's Andes of South America. Volcanoes formed an arcuate north-south chain of mountains off the coast of western Pangea in what is now central Nevada and belched ash that blew onto the Colorado Plateau and much of the west (figure 32). Much of this activity happened in the

Middle Jurassic (178 to 157 Ma) and was caused by renewed subduction during the Elko Orogeny (figure 19) (Christiansen et al., 1994). To the south, the landmass that would become South America split away from the Texas coast just as Africa and Great Britain rifted away from the present East Coast and opened up the Atlantic Ocean. The Ouachita Mountains of southeastern Oklahoma and Texas, formed when South America collided with North America, remained a significant highland to the southeast, and rivers from the highland flowed to the northwest, towards the desert. The Ancestral Rocky Mountains remained high in the Jurassic and contributed sediment, also (Graham et al. 2002).

As the Absaroka Sequence drew to a close in the Early Jurassic, the craton emerged above sea level, and the shoreline migrated away from the continental margins as sea level fell (figure 31) (Sloss, 1988). During the Early Jurassic, plate collisions were not so severe and the northern sea did not encroach onto the continent. Consequently, erosion of Triassic and Upper Paleozoic sandstones from as far north as Montana and Alberta provided abundant quantities of sand to be transported by wind to desert area to the south (Kocurek and Dott, 1983). Yet, it seems illogical that westerly winds would transport Alberta sand to Colorado unless those winds were diverted to the south, perhaps by a rising mountain range to the west. Indeed, this is what may have happened. The volcanic arc may have acted as a barrier to block the wind and channel it to the south (Kocurek and Dott, 1983). While the volcanic arc diverted the wind, the volcanoes did not provide sediments to the system, however. Sediments from the volcanic arc province to the west are missing from the dune sand. An additional source of sand could have been the ancestral Rockies. The perpetually dry, arid climate may have also helped by keeping the groundwater table depressed and thus, the sand particles were always available to the wind (Kocurek and Dott, 1983; Graham et al. 2002).

When the pace of west coast collision increased during the Elko Orogeny in the Middle Jurassic (about 236 to 240 Ma), the rock layers on the continental side of the collision, from southwestern Montana to Colorado, deformed in response to the collision to the west. Like a ripple effect on water, only here in rocks, the layers folded upward and over millions of years, this ripple fold migrated eastward. As the strata bowed upward, weathering and erosion stripped away the rocks and the time represented by those rocks to create unconformities. As plate tectonic activity increased to the west again, the sea began to onlap the continent from the north and signaled the initiation of the Zuni Sequence (figure 31)(Graham et al. 2002).

Middle Jurassic strata represent a complex interfingering of marine and nonmarine environments just east of the western orogenic margin. Broad tidal flats formed marginal to a shallow sea that lay to the west (Wright et al., 1962). The sea encroached into west-central Utah from the north and lay in the Utah-Idaho trough bordered to the west by the Elko Highlands. Flat-bedded sandstones, siltstones, and limestones filled in depressions left in the underlying eroded strata (Doelling, 2000; Graham et al. 2002).

As plate tectonic activity increased at the end of the Middle Jurassic and beginning of the Late Jurassic (about 157 Ma), a major transgression of the inland seaway forever destroyed the vast eolian sand seas that once covered the western interior (Kocurek and

Dott, 1983). Tidal flats covered the area as marine environments pushed south. Two additional marine transgression/regression couplets occurred in the Late Jurassic before the seas finally receded and the extensive Upper Jurassic, Morrison Formation was deposited above an erosional unconformity across the continental western United States (figure 36). The Morrison Formation is world renown for both its dinosaur bones and its uranium deposits. Jurassic dinosaur bones from the Morrison Formation grace many museums worldwide. About 50% of the uranium resources of the United States are found in the Morrison Formation (Peterson, 1994; Graham et al. 2002). The Morrison and Fernie Formations are exposed just southeast of Glacier National Park.

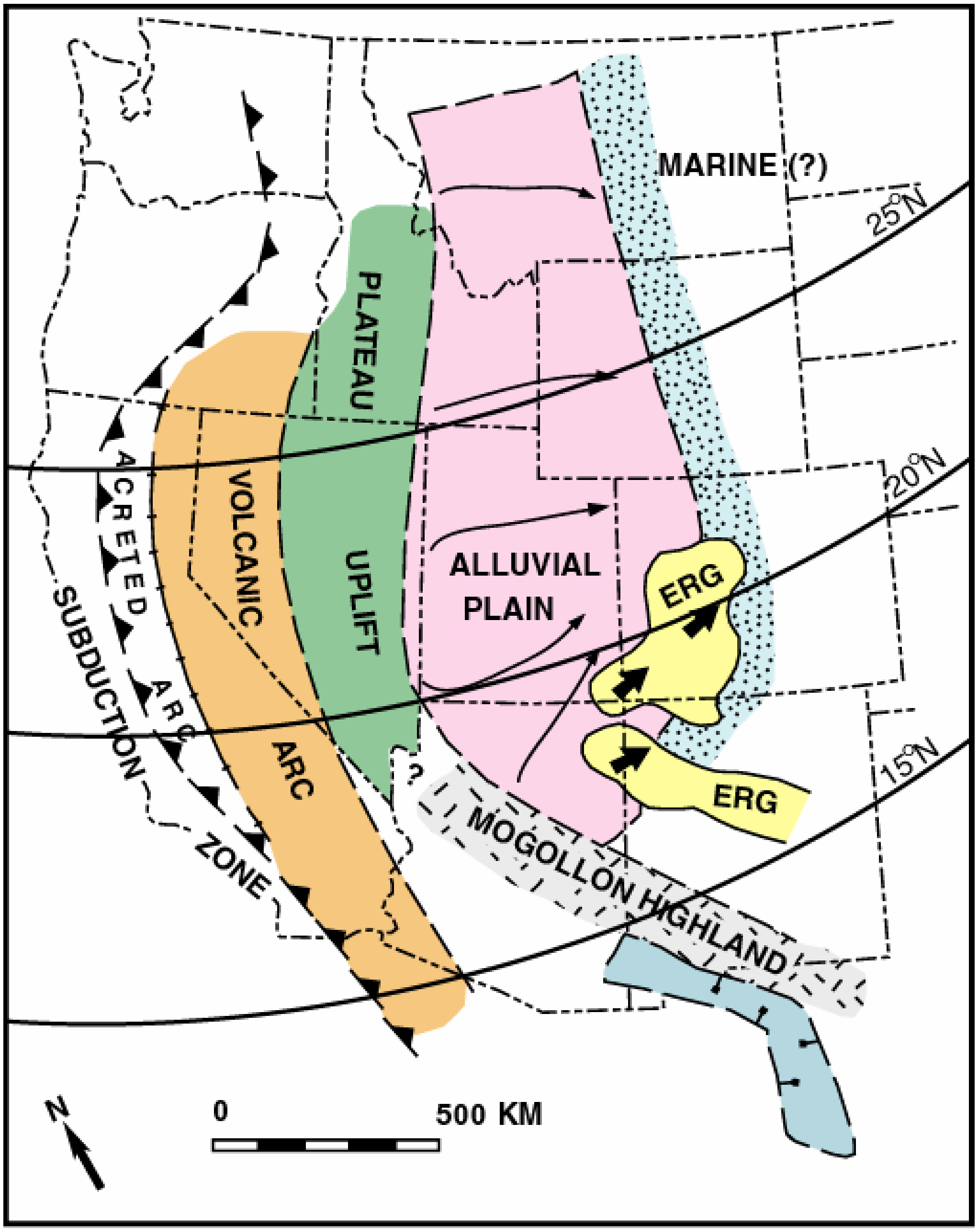


Figure 36: Upper Jurassic Period environments. Thin arrows indicate fluvial dispersal. Thick arrows indicate wind directions. Saw teeth indicate the location of the subduction zone with the teeth on the overriding, upper lithospheric plate. Note the possible marine environment to the east where continental environments were previously established. The alluvial plain expanded to the east with time. Modified from Lawton, 1994.

The Mesozoic Era - Early Cretaceous Period. The Cretaceous age rock exposures in Glacier National Park are scarce as they are only present below the Lewis thrust fault system and east of the Park which is largely vegetated. This section will rely on rocks in the surrounding areas as well as those exposed in the Park.

During the late Jurassic, the subducting oceanic slab (Farallon plate) that was sliding eastward beneath the continental lithosphere is hypothesized to have changed its angle of descent to become steeper. This change caused the volcanic arc, at the continental margin, to develop farther to the west near the present-day border of California and Nevada. The late Jurassic Period and earliest Cretaceous Period magmatic activity associated with the volcanic arc is called the Nevadan Orogeny. The Nevadan Orogeny evolved into the widespread Sevier Orogeny as the rate of convergent lithospheric plate movement increased (figure 32)(Graham et al. 2002).

The Sevier Orogeny formed a roughly north-south trending thrust belt that is well defined in present-day southern Nevada, central Utah, and western Montana (figures 6 and 32). A series of eastward-directed overthrusts carried upper Precambrian and lower Paleozoic sedimentary rocks over upper Paleozoic and lower Mesozoic rocks (Stewart, 1980). Collision caused deeply buried rocks to the west to be thrusted over younger rocks in Montana and Wyoming and to be stacked piggyback style on top of one another (Graham et al. 2002). The Wasatch Range in Utah, the Wind River Range in Wyoming, and the Lewis Range in Montana today record the easternmost extent of the Sevier Orogeny.

As the mountains rose in the west and a roughly north-south interior trough expanded, the Gulf of Mexico separating North and South America continued to rift open in the south, and marine water began to spill into the basin. At the same time, marine water began to transgress from the Arctic region. As the shallow sea advanced onto the continent, the currents redistributed the sediments deposited from river systems in much the same way sediments are redistributed along the shorelines of North America today. With the sediments redistributed, more space was available for depositing river sediments, and the basin continued to subside under the weight of the continuous influx (Graham et al. 2002).

The sea advanced, retreated, and readvanced many times during the Cretaceous Period until the most extensive interior seaway ever to cover the continent drowned much of western North America (figure 37). The Western Interior Seaway was an elongate basin that extended from today's Gulf of Mexico to the Arctic Ocean, a distance of about 4,827 km (3,000 mi) (Kauffman, 1977). During periods of maximum transgression, the width of the basin was 1600 km (1,000 mi). The basin was relatively unrestricted at either terminus (Kauffman, 1977). The western margin of the seaway coincided with the active Cretaceous Sevier orogenic belt, but the eastern margin was part of the stable platform or ramp that gradually sloped down to the west in Nebraska and Kansas. Consequently, sedimentation into the basin from the rising mountains on the western margin was rapid compared to the slow sedimentation from the craton onto the eastern margin of the basin (Graham et al. 2002).

Oxygen isotope analyses of well-preserved foraminifera (unicellular marine animals with shells made from calcium carbonate or cemented sedimentary grains) show a greenhouse global climatic pattern in the Late Cretaceous that ranged from a warm greenhouse state about 101 to 94 Ma to a hot greenhouse phase about 94 to 75 Ma and then to cool greenhouse conditions during the latest Cretaceous about 75 to 65 Ma (Huber et al., 2002). Oxygen isotopes ($\delta^{18}O$) were used in these analyses because they reflect the isotopic composition of seawater at the time when the calcitic foraminifera shell formed. Assuming a mean isotopic composition of seawater in a nonglacial world, a paleotemperature can then be calculated for a specific time period. Foraminifera are excellent fossils to use for age-dating because their species changed relatively rapidly through the Cretaceous Period and they dispersed rapidly throughout the Cretaceous global marine environment (Graham et al. 2002).

Although the seaway was not physically restricted at either end, water circulation appears to have been periodically disrupted. Paleontologic and isotopic evidence support a restricted circulation and the subsequent formation of a brackish surface layer except during times of maximum transgression when warm waters from the Gulf flooded into the seaway (Kauffman, 1977). When warm water from the gulf entered the seaway, cold water biota were pushed farther north whereas during regressions, the water temperature cooled and cold water biota migrated back south (Kauffman, 1977). The hydrologic regime would be disrupted, as well, as relative sea level rose or fell. More or less quantities of fresh water drained from surrounding highlands and entered the basin. For example, during the times when the sea retreated (regression), delta systems, coal swamps, and beaches followed the shoreline. Rivers incised across the previous shoreface and old delta systems as sea level fell. When the sea level rose again (transgression) and the shoreline migrated landward, the river valleys were drowned and became estuaries with their mix of fresh, brackish, and marine water species. Restricted coal swamps formed in the quiet backwaters of estuaries, and sand deposited in beach and near-shore environments was reworked and re-deposited farther inland. The finer grained sediments such as clay and silt drifted offshore to bury the environments that had formed during the previous low stand of sea level. Over time, these changes would be reflected in the vertical section of rocks and fossils (Graham et al. 2002). The spectacular specimens found near Choteau, Montana, just southeast of Glacier National Park record these diverse environments.

At the beginning of Late Cretaceous time, the interior seaway had advanced to cover the a large portion of the continent (figure 37). The advance of the sea, however, was not continuous. The seaway grew in fits and starts, through transgressive and regressive pulses caused by different rates of plate movements, different rates of subsidence, and/or different rates of sedimentation into the basin. Eventually, at times of maximum transgression, the shoreline of the interior seaway was next to the mountains in western Utah (Griffitts, 1990; Graham et al. 2002).

These marine cycles of advance and retreat are reflected in the Cretaceous age rocks east of the Belt Terrane and beneath the Lewis thrust sheet in Glacier National Park. The Kootenai Formation is largely lacustrine, probably formed when the sea had regressed and local lakes and ponds formed on the former shoreline. The Blackleaf Formation records coal swamps and river deltaic environments, possibly indicating a drainage system from the west into the seaway. The Marias River Shale signals a return to deeper water conditions with fine-grained, calcareous rocks. The Telegraph Creek Formation and Virgelle Sandstone record the gradual inundation of the area by the marine Cretaceous Seaway. The Sandstone of the Virgelle is largely beach deposited. Then, the Two Medicine Formation again records a change to deltaic, subaerial deposition, followed by the marine Bearpaw Shale and the Horsethief Sandstone deposited in a marine regression event. Following this regression, the St. Mary River Formation contains brackish water fossils and thin coal beds recording a deltaic marshy environment. The Willow Creek Formation resembles the nonmarine rocks of the St. Mary River Formation, but is lacking coal, possibly indicating either a drier climate or a more active drainage pattern (Rice and Cobban 1977; Whipple et al. 1985).

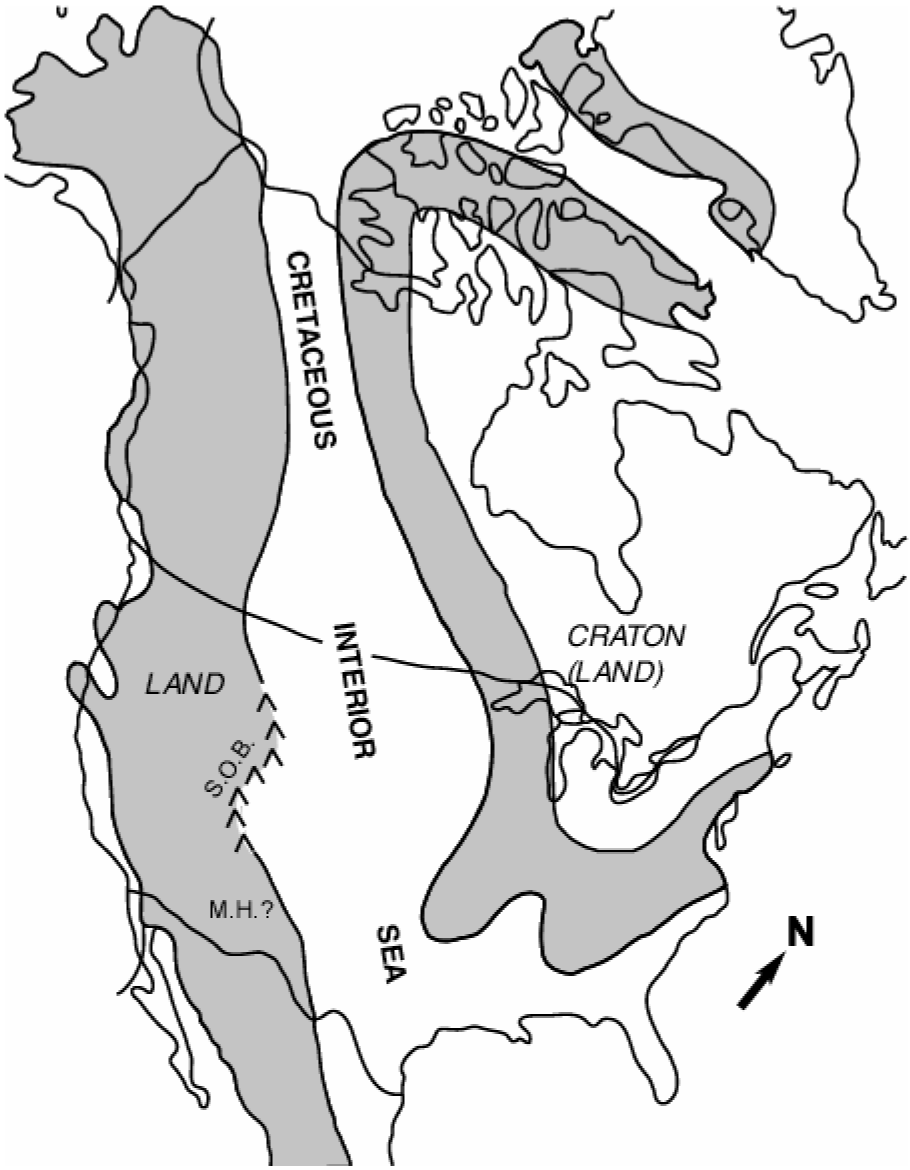


Figure 37: Location of the Cretaceous seaway. Shaded areas indicate land above sea level. “S.O.B.” and inverted “V”s indicate the Sevier Orogenic belt. “M.H.?” is the Mogollon Highland in southwestern Arizona. North indicates the Cretaceous north. Modified from Rice and Shurr (1983).

As the Late Cretaceous interior seaway shrank away, the shoreline regressed to the northeast. Eventually, the shoreline migrated to the north and east and a broad, coastal plain formed were the sea once was. Sluggish streams meandered across this nearly flat coastal plain. Broad shallow swamps developed in the floodplains behind the levees, on the deltas, and behind the low-relief shoreline. The climate was warm and wet and enough plant material accumulated in the swamps to develop into coal beds. Late Cretaceous coal beds of economic importance form a trend from the San Juan Basin in northwestern New Mexico, through southeast and central Utah, and into southwestern Wyoming (Roberts and Kirschbaum, 1995; Graham et al. 2002).

Vigorous volcanic activity in the Late Cretaceous was the surface expression of thick accumulations of magma that was generated above the subduction zone on the western margin. Below the volcanoes, plutons were actively emplaced along the Sierra Nevada of California and Nevada as well as in southwestern Arizona (Christiansen et al., 1994).

A globally high sea level, high rates of thrusting in the Sevier orogenic belt, and rapid subsidence of the associated foreland basin created an extensive sediment accommodation zone along the western margin of the Cretaceous Interior Seaway.

During the Cretaceous-age Sevier Orogeny (about 105 to 75 Ma), great sheets of sedimentary rocks that covered miles of terrane were thrust westward from what is now western Nevada into central Utah, a distance of roughly 500 km (300 mi). The emplacement of these thrust sheets can be seen today in the mountains of Arizona, Utah, Wyoming, Montana, and Idaho. The Sevier Orogeny showed an example of *thin-skinned* thrust faulting wherein just the upper sedimentary strata of Earth's crust were transported on laterally extensive thrust planes that dip at a low angle, generally 10-15 degrees, from the horizontal surface of Earth. In contrast, thrust faults associated with the Late

Cretaceous-Early Tertiary Laramide Orogeny are *thick-skinned*, that is, they are faults with nearly vertical fault planes at the surface of Earth that flatten and sole out in Precambrian basement crystalline rock at depths up to 9,000 m (30,000 ft) below sea level (Gries, 1983; Erslev, 1993). During the Laramide Orogeny, tectonic forces folded and faulted the entire geologic column, from Precambrian to Cretaceous age rocks, into the north-south trending Rocky Mountains and adjoining basins (Graham et al. 2002).

At one time, the Proterozoic Belt Supergroup and overlying rocks were buried beneath thick deposits of Cretaceous strata. Then the Laramide orogeny came to Montana. The Lewis thrust fault, though having moved intermittently for some 200 m.y., became the primary plane of movement for the massive rock column containing the Belt Terrane. This column moved eastward some 10's of miles. Changing the Glacier area, once underwater during the tenure of the interior seaway into a mountainous highland. The Laramide event transformed the extensive basin of the Cretaceous Interior Seaway into smaller interior basins bordered by high arches (anticlines and synclines on the scale of miles) (Ehrlich 1999). The unique depositional patterns originating from Laramide uplifts mark the initiation of the Tejas Sequence on the western margin of North America (figure 31) (Sloss, 1988). The Late Cretaceous-early Tertiary Laramide Orogeny (about 75 to 35 Ma) is one of the more perplexing episodes in the structural history of the Rocky Mountains. The orogeny affected rocks hundreds of miles inland from the continental margin. Modern geologists have struggled with models to predict these results (Graham et al. 2002).

In 1978 William Dickinson and Walter Snyder proposed a plate tectonic model for the Laramide Orogeny that explained both the shutdown of volcanism and the eastward migration of deformation. They used the Andes Mountain chain as a modern equivalent to the Laramide Orogeny, and the model still remains the best explanation for the Laramide event. In their model, the east-dipping, oceanic Farallon plate was subducting beneath the overriding, continental North American plate at a steep angle during the Sevier Orogeny. The oceanic crust would melt when it reached a relatively deep level below the North American plate. Molten material would creep upward, melting some of the overlying continental rocks, eventually to explode at the surface as volcanoes or slowly cool beneath the surface and crystallize into igneous plutons. Plutons were emplaced above the subduction zone and are now exposed in the Sierra Nevada mountains of eastern California. Volcanic ash from the volcanic island chain spread across the continent. Ash deposits suggest that two periods of volcanism occurred in the Late Cretaceous, one peaking about 95 Ma and one about 75 Ma (Christiansen et al., 1994; Graham et al. 2002).

Near the end of the Cretaceous, the subducting Farallon plate may have changed its angle of dip from steep to relatively flat (termed *flat-slab* subduction) or pieces of the slab may have been "rafted" beneath the continent (figure 38) (Dickinson and Snyder, 1978; Livaccari, 1991; Fillmore, 2000). Subducting at a flatter angle, the oceanic plate would not melt and thus, the source for volcanic activity was cut off. Tremendous shear stresses were generated between the two slabs as the subducting oceanic plate pushed far inboard. Stresses at the base of the thick continental crust were transmitted upward in the form of compression, thrusting great wedges of basement rock skyward to form the impressive Laramide Rocky Mountains. Whatever the mechanisms, the North American continent was compressed in an average northeast/southwest direction to produce the Laramide Rocky Mountains. Complex interactions among local crustal blocks can explain most of the arches and basins that formed during the Laramide Orogeny (Joe Gregson, NPS, personal communication, 2001; Graham et al. 2002).

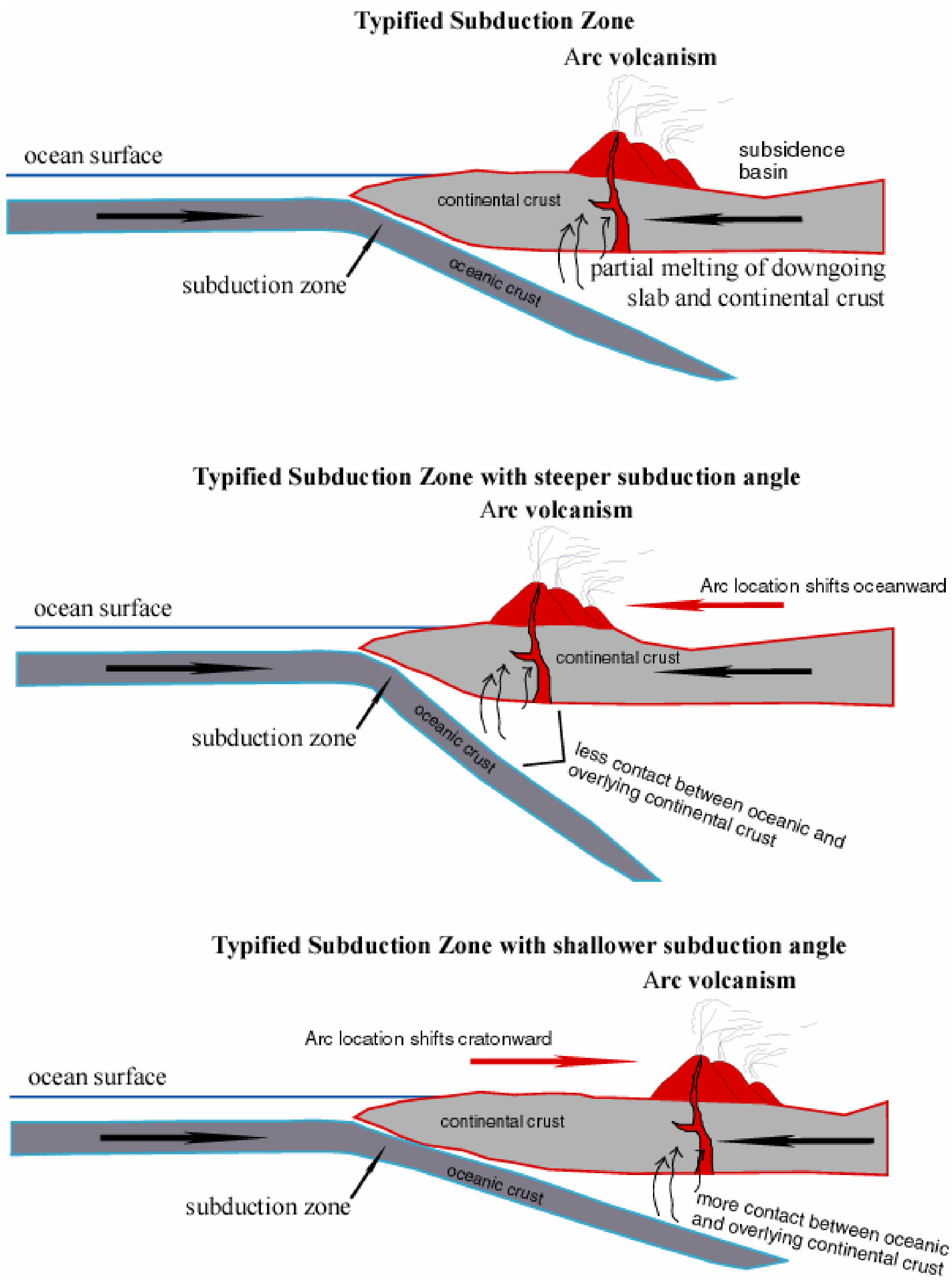


Figure 38: Schematic illustrating "flat-slab subduction". The angle of subduction in the upper drawing decreases, or flattens out, as indicated in the lower diagram. As a result, the influence of subduction is felt farther inland. Graphic designed by Trista L. Thornberry.

Plate tectonics can be called upon to explain the abrupt end of the Laramide Orogeny. Today, the North American plate converges obliquely on the Pacific Plate, but in the

Cretaceous, the Farallon plate was being subducted beneath the North American plate. A mid-ocean spreading ridge separated the Farallon plate from the Pacific plate that lay farther to the west. A spreading ridge is a linear feature, but the ridge is laterally offset by *transform* faults (a type of strike-slip fault). The rate of convergence in the Cretaceous-Tertiary was greater than the rate of spreading so that, eventually, after most of the Farallon plate disappeared, the mid-ocean ridge was subducted beneath the North American continental plate. When this happened, the subduction process ceased and the Laramide Orogeny was brought to a grinding halt. The dense oceanic slab that remained beneath the continental crust slowly foundered and sank into the mantle and hot mantle material rose to take its place (Graham et al. 2002).

By 20 Ma, only remnants of the Farallon plate remained. The San Andreas strike-slip fault system between the Pacific plate and the North American plate began to grow and as it lengthened, the southwestern margin of North America began to undergo extensional deformation. As the crust was extended about 15 Ma, the surface began to be broken into the basin-and-range topography we see today in western Utah, Nevada, Arizona, and the Rio Grande Rift in New Mexico. In summary, from the Late Cretaceous to the Late Tertiary, the tectonic regime on the western margin of the North American continent changed from a steeply-dipping subduction zone to flat-plate subduction to extension caused by a growing transform fault system (Graham et al. 2002).

This extensional tectonic setting in Tertiary time is recorded in the array of normal faults in Glacier National Park. These include the Roosevelt, Flathead and Blacktail faults as well as many parallel faults west of the park. When one block of rock falls relative to another, the resulting graben valley forms a substantial sediment sink. Such was the case for the North Fork valley and the resulting sedimentary fill, the Tertiary Kishenehn Formation. The Kishenehn is characterized by its variety of sediments ranging from sandstone, mudstone, limestone, and coal to volcanic ash deposits. This assortment indicates the tectono-climatic variations during the Tertiary. The volcanic ash records volcanic activity along the western margin of North America, mostly in the Cascade Mountains of Oregon and Washington. The limestone and coal seams imply the presence intermittent ponds and restricted lakes that formed in the graben. These occur in isolated patches probably indicating a relatively dry climate. However, the abundant petrified wood specimens in the Kishenehn Formation indicate that the climate was not desert like, but wet enough to support abundant plant life.

The Cenozoic Era - Quaternary History. The Quaternary Period is subdivided into two epochs: 1) the Pleistocene, which ranges from about 1.6 Ma to 10,000 years before present (B.P.), and 2) the younger Holocene Epoch that extends from 10,000 years B.P. to the present. The Pleistocene Epoch is known as the Ice Age and is marked by multiple episodes of continental and alpine glaciation. Great continental glaciers, thousands of feet thick, advanced and retreated over approximately 100,000-year cycles. Huge volumes of water were stored in the glaciers during glacial periods so that sea level dropped as much as 300 feet (Fillmore, 2000). When sea level lowered, land bridges emerged such as the Bering Land Bridge that linked North America and the Eurasian continents. During interglacial periods, Earth warmed and the glaciers retreated toward the polar regions. Sea level rose and the great land masses were once again isolated (Graham et al. 2002).

Glaciers played a huge role in forming the landscape now present in Glacier National Park. This role is manifested in the glacial features and deposits now omnipresent in the Park. These deposits include huge terminal and lateral moraines which record the greatest glacial extent. Many of these deposits dam glacial valleys resulting in the narrow lakes so characteristic of the Park. Jumbled glacial till, esker, and outwash deposits are strewn along every valley traversed by a glacier in the Park. These impacts are described in detail in the following section.

The Holocene, of course, is the Age of Humans and our impact on our global ecosystem is complex. With the retreat of the glaciers and the end of widespread glaciation about 12,000 years ago, the climate continued to warm and global sea level rose. In some local areas (i.e., the coast of Maine and the Great Lakes region), however, relative sea level lowered as the land rebounded from the weight of the glaciers. Local tectonism, sediment input, global warming, and global cooling are some of the factors affecting global sea level and their relative importance, and humans' influence on them, continues to be debated today (Graham et al. 2002). Climate change is readily observable in the retreat of the Park's remaining glaciers. The response time of a glacier to climate change is rapid and the recent melting rate serves to substantiate the greenhouse effect hypothesis.

The Glacier Park area is experiencing significant growth in population. The Park boundaries are being inundated with the effects of development. Man-made bridges, roads and dams, such as the Hungry Horse Dam in Hungry Horse, Montana change the course of rivers and sediment deposition.

Glaciology and Glacier National Park

Glacier National Park is on of only a few Parks in the Rocky Mountain of the conterminous U.S. with a climate suitable for maintaining substantial glaciers since the end of the last glaciation some 10,000 years ago. This climate exists for two reasons. First is the geographical situation of the Park. It is far enough north and has mountains that are high enough to keep relatively cool in summer. The other reason is that the mountains in the Park capture significant precipitation from moist Pacific air moving inland. The impressive glaciers in northwestern Montana at the turn of the twentieth century, along with the impressive rugged mountain scenery and abundant wildlife, led to the creation of the Park in 1910. When the Park was founded , the glaciers still present there were in the process of thinning and melting back from their maximum extent which occurred around 1850, when they were bigger than they had been since the Pinedale Glaciation (Elias 1996). The ice retreat since this "Little Ice Age", a term now used to describe the period from about 1500 to the mid-19th century glacial maximum, is a dramatic example of the dynamic nature of glaciers.

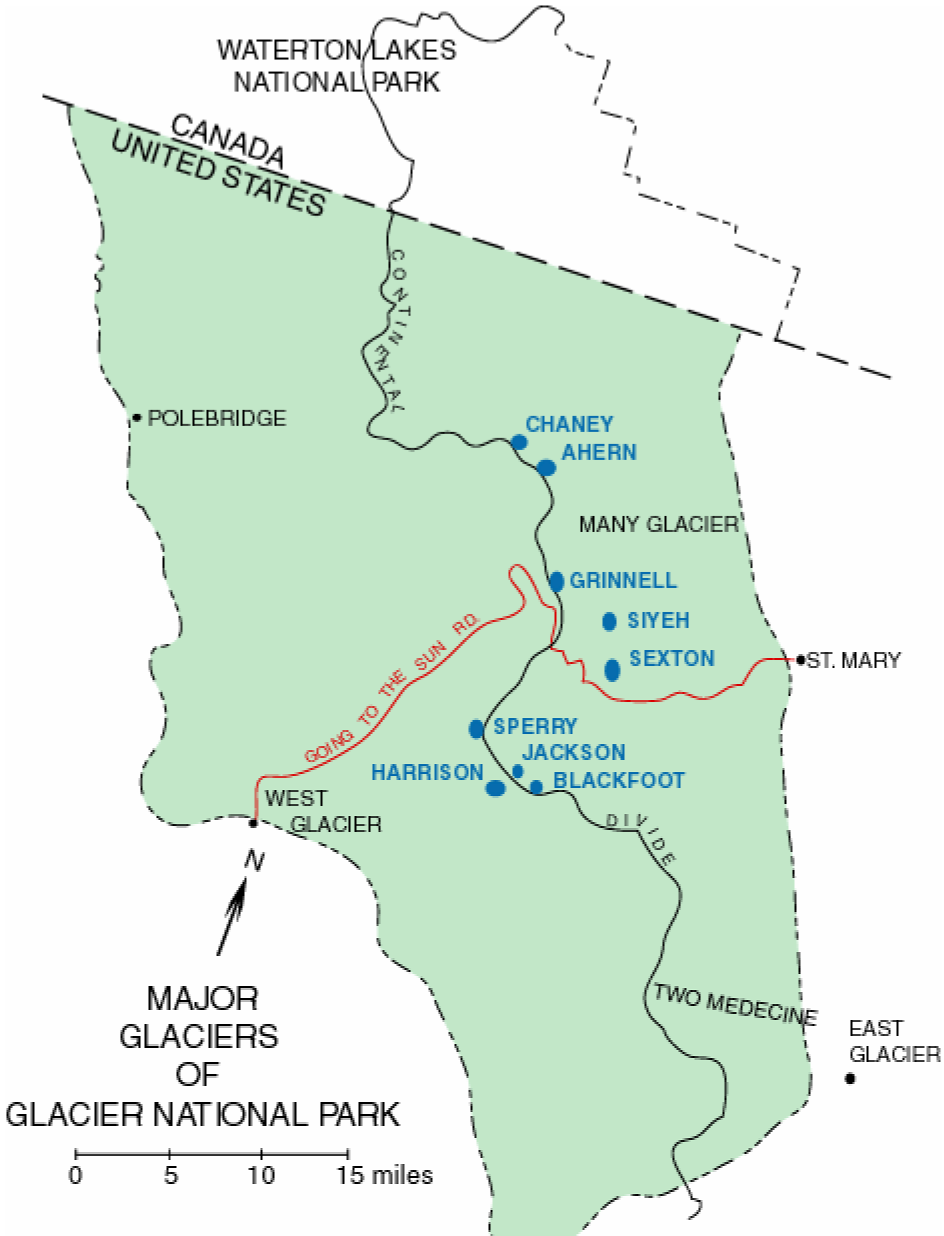


Figure 39: Map of Glacier National Park showing locations of the largest glaciers present today. Modified from Dyson 1966.

Some 37 named glaciers (50-60 total) currently exist in various cirque, niche, ice apron, group and remnant forms in Glacier National Park (Key et al. 1996). Only two have surface areas of nearly one-half square mile, and not more than seven others exceed one-fourth square mile in area (Dyson 1966) (Figure 39). However, the real glacial story of the Park begins with what you don't see; that is, what was left behind when the huge glaciers of the past melted away. The Wisconsin Glaciation (the most recent of the 4 major Pleistocene Glaciations, from oldest to youngest: Nebraskan, Kansan, Illinoian, and Wisconsin), is called the Pinedale Glaciation in the Rocky Mountain region (after terminal moraines near the town of Pinedale, Wyoming). The Pinedale Glaciation began after the last (Sangamon) Interglaciation, perhaps 110,000 years before present (B.P.), and included at least two major ice advances and retreats. In the west, the buildup of glacial ice in the mountains of British Columbia to the north of Glacier National Park formed what is called the Cordilleran Ice Sheet, which butted up against the Laurentide Ice Sheet (Continental Ice Sheet) in western Canada and flowed south in Montana to the site of present-day Flathead Lake. The two ice sheets covered more than 16 million km^2 (6 million square miles) of North America, stretching from coast to coast. Glacier National Park was uniquely situated, hemmed in by the two great ice sheets (Elias 1996) (figure 40).

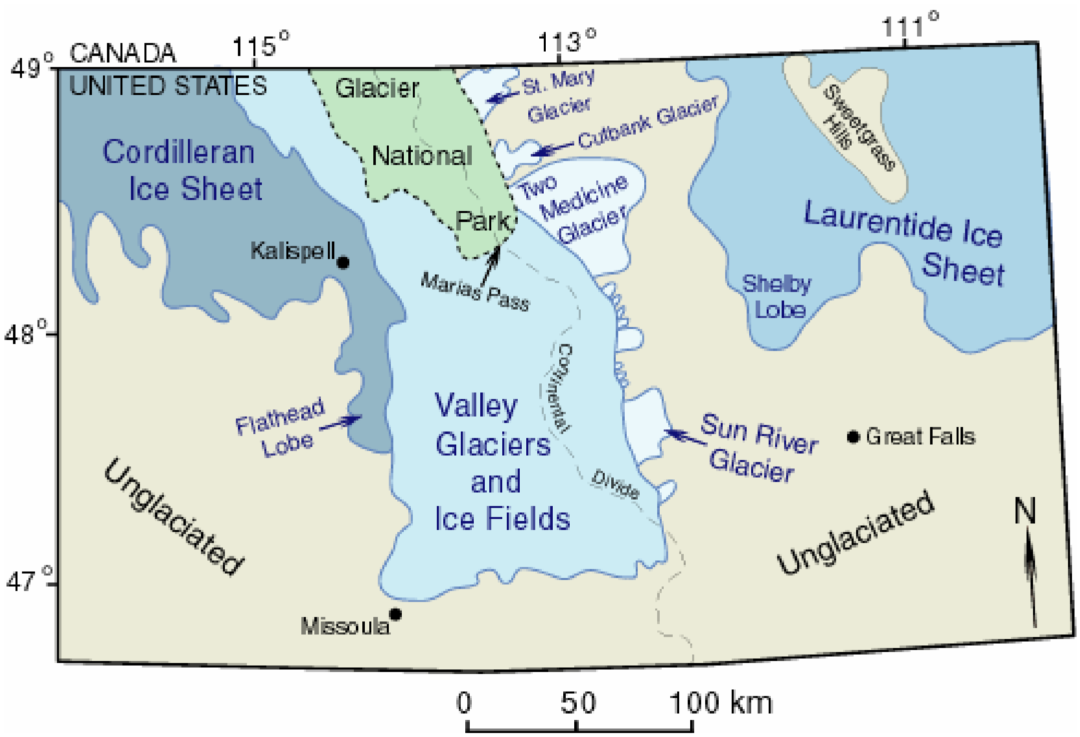


Figure 40: Map showing location of Glacier National Park in relation to glacier and ice sheets present during the Pinedale Glaciation's maximum extent. Modified from Carrara 1986.

At the height of the Pinedale Glaciation, about 20,000 years ago, only the highest ridges and peaks in Glacier National Park were free of ice. These ice-free regions are typically called nunataks. All valleys of Glacier National Park were filled with valley glaciers (Dyson 1966). Mountain glaciers that formed on the western slope of the Continental Divide in the Livingstone Range flowed/carved down the valley of the North Fork of the present-day Flathead River. These glaciers joined with ice coming from the eastern slope of the Whitefish Range, and the combined glaciers flowed south, ploughing over the lower Apgar Mountains and contributing ice to the "Flathead Lobe" of the large Cordilleran Ice Sheet. The Flathead Lobe gouged out a deep depression, and the terminal moraine left by this glacial lobe upon glacial retreat dammed the Flathead River about 20,000 years B.P., creating Flathead Lake (Dyson 1966; Elias 1996).

Glaciers that flowed from the western side of the Lewis Range south of Lake McDonald merged with ice from the southeastern flank of the Flathead Range. This ice flow, combined with other mountain glaciers southeast of the Park, formed a large body of ice around the southwestern corner of the Park. This body pushed up and over the low divide at Marias Pass and ended up in the large piedmont glacier called Two Medicine Glacier, east of the Continental Divide. This glacier was also fed by ice flowing from the southern end of the east slope of the Lewis Range. Farther north along the eastern slope of the Lewis range, the Cut Bank and St. Mary's glaciers were formed from other mountain glaciers flowing out onto the plains. Ice from the northwestern flank of the Lewis Range and the northeastern flank of the Livingstone Range flowed north into Canada to join the combined Cordilleran and Laurentide Ice Sheet (Elias 1996).

No one knows exactly how many times glaciers moved down the Park valleys during the million or more years of the Pleistocene period, but geologists have found evidence for at least eight distinct advances in the area. It is difficult to determine just when the first advance took place, as evidence of early advances is often destroyed by later advances, but it may have been very early in the period. Evidence of the several distinct glacial advances is yielded by the moraines; a term for deposits of rock debris left in various configurations by the melting ice (Dyson 1966).

During the Pinedale Glaciation, active volcanoes of the Cascade Range in Washington and Oregon were erupting and depositing ash layers in the Glacier Park Region. These ash layers can be accurately identified, dated, and correlated in the large moraines left from Pinedale Glaciers. They provide useful dates about the timing of glacial recession in the area. Pinedale deglaciation was fairly rapid and by 11,200 years B.P. the ice had retreated to local mountain valleys in the Park, and by 11,300 and 10,000 ^{14}C years B.P. (using radiocarbon dates, tree-ring data, macrofossils, and the presence of the Mazama, Glacier Peak G and St. Helens J volcanic ash beds) it had withdrawn to the same high-mountain cirques and shaded niches where small glaciers persist today (Carrara and Wilcox 1984; Elias 1996; Osborn and Gerloff 1997). The present day ice is probably not the remnants of Pinedale Glaciation. About 6,000 years ago all glacial ice probably disappeared from the mountains. After this there was a warm, dry climatic period during which it is probable that no glaciers were present (Carrara 1986). Then about 4,000 years ago the small glaciers present today were formed (Dyson 1966).

What was left behind when the huge Pinedale Glaciers retreated? Essentially, the incredible rugged landscape of Glacier National Park is due to glacial processes. A glacier is an extremely powerful agent of erosion, capable of profoundly altering the landscape over which is passes. Glaciers erode primarily by two distinct processes, plucking and abrasion (Dyson 1966). The first is more active near the head of the glacier, at its source, but may take place anywhere throughout its course; abrasion or scouring is most effective along the base of the glacier, particularly where the ice moves in a well-defined channel. In plucking, the glacier actually quarries out distinct masses of rock, incorporates them within the ice and carries them along. At the head of the glacier this is accomplished principally by water, either meteorologic or melt, which trickles into crevices and freezes. The expansion upon freezing liberates blocks of rock to be pulled out by the glaciers. The weight of the glacier also aids in plucking by actually squeezing ice into the cracks in the rocks. As the glacier moves forward these blocks of ice are dragged or carried along with it (Dyson 1966). Usually a large crevasse, the bergschrund, develops in the ice at the head of a glacier as a result of gravitational stresses pulling the glacier away from the headwall.

The bergschrund of most active glaciers in Glacier National Park consists of an opening, usually 10 to 20 feet wide at the top and as much as 50 feet deep, between the head of the glacier and the mountain wall. It is at this site that plucking is most active and dominant because water enters by day and freezes in the rock crevices at night. By quarrying headward and downward the glacier finally carves the formation of a steep-sided, bowl-shaped basin called a cirque or glacial amphitheatre. The cirque is the first place that ice forms and the place from which it last disappears, thus it is subjected to glacial erosion longer than any other part of the valley. If sufficient erosion has occurred, a body of water known as a cirque lake forms in the depression after the glacier disappears (Dyson 1966). Iceberg Lake, for example, lies in one of the most magnificent cirques in the Park (figure 41).

Figure 41: Iceberg cirque, Glacier National Park, NPS photo.

Rock fragments of various sizes frozen into the bottom and sides of the glacier form a huge file or rasp which abrades or wears away the bottom and sides of the valley course down which the glacier flows. The valley thus attains a characteristic U-shaped cross section, with steep sides and a broad bottom. Practically all the valleys of the Park, especially the major ones, possess this distinct U-shaped cross section. Splendid examples are the Swiftcurrent Valley, St. Mary Valley, and the Belly River Valley. The floors of many of the Park's major U-shaped valleys are marked by several steep drops or "steps", between which the valley floor has a comparatively gentle slope. Such a valley floor is called a glacial stairway (Dyson 1966). These features result from the differences of erosional resistance between different rock types of the of the underlying formations. A weaker rock such as a shale will erode more easily than a stronger rock such as a sandstone. Thus a glacier will scour more deeply into a weaker rock forming a "tread" in contrast to the cliffs or "risers" formed by the erosion of stronger rocks. Resistant layers in the lower portion of the Altyn Formation, the upper part of the Appekunny Formation, and the upper part of the Grinnell Formation normally create risers in Glacier National Park.

The "tributaries" of glacial valleys, filled with smaller glaciers which feed into the larger ones, are also noteworthy in Glacier Park, known as hanging valleys. They form from a difference in erosional capacity between the smaller glacier and the valley glacier. The thicker a stream of ice, the more erosion it is capable of; consequently, the main valley becomes greatly deepened, whereas the smaller glacier in the tributary valley does not cut down so rapidly, leaving its valley hanging high above the floor of the major valley once the ice melts (figure 42) (Dyson 1966). The valleys of Virginia and Florence Creeks, tributary to St. Mary Valley are excellent examples of hanging valleys.

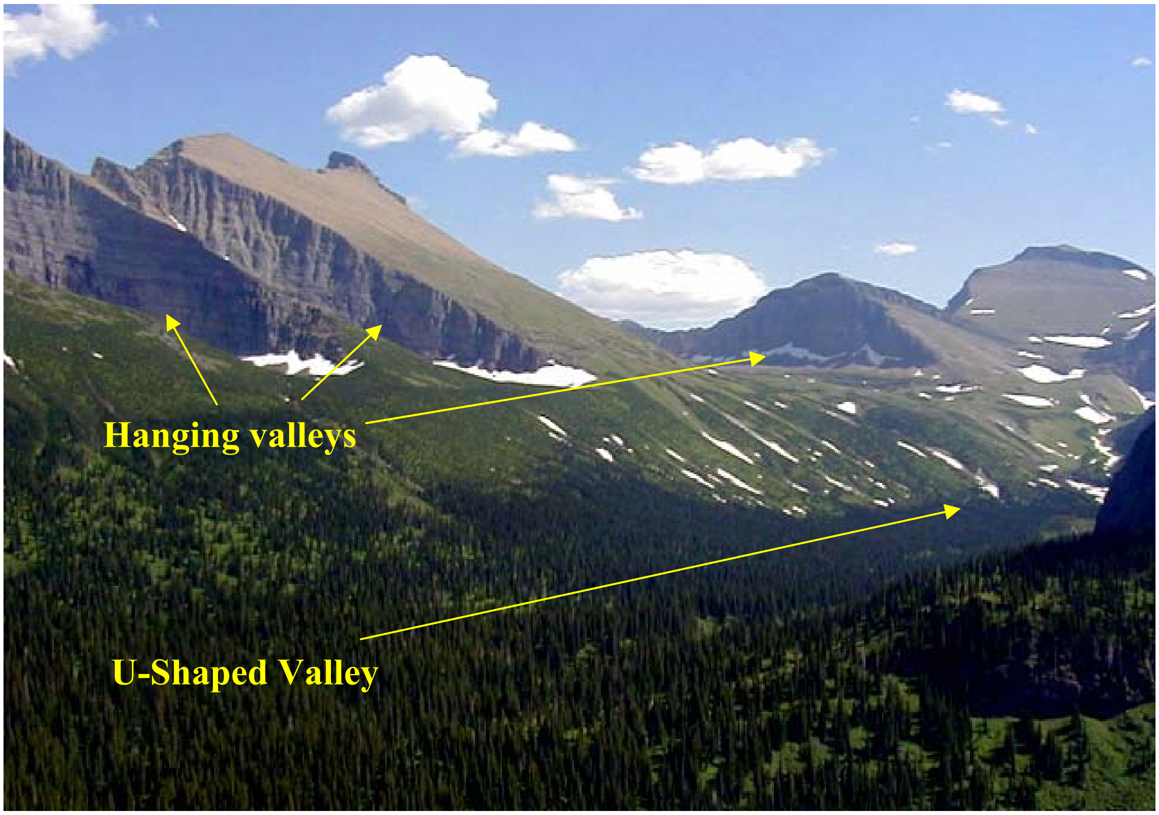


Figure 42: Three hanging valleys left as remnants of former glaciers flowing into the larger that once occupied the U-shaped valley running up the axis of the image. Modified from photo by Dawes and Dawes 2001. For more information see: http://wvcweb.ctc.edu/rdawes/VirtualFieldSites/GrinnellGlacier/VFSGrinnell.html

Conspicuous throughout the Park are the long, knife-sharp ridges which form most of the backbone of the Lewis and Livingston Ranges. These features, of which the Garden Wall is one of the most renowned, are known as arêtes and also owe their origin to glacial processes (figure 43). As the long valley glaciers enlarged their cirques by cutting farther in toward the axis of the mountain range, the latter finally was reduced to a very narrow steep-sided ridge, the arête. In certain places glaciers on opposite sides of the arête essentially cut through the ridge creating a low place known as a col, usually called a pass (Dyson 1966). Gunsight, Logan, and Red Eagle passes are a few examples. At locations where three or more glaciers plucked their way back toward a common point, they left at their heads a sharp-pointed peak known as a horn. Reynolds, Bearhat and Clements Mountains are excellent examples of horns (see figure 1).

Figure 43: View of the Garden Wall arête, Glacier National Park, NPS photo.

Another feature of the Park which must be attributed, at least in large part, to glaciation is the myriad of spectacular waterfalls. There are two principal types of falls at Glacier, one which occurs in the bottom of the main valleys and one at the mouth of the hanging tributary valleys (Dyson 1966). The former exemplified by Trick Falls of the Two Medicine River, is located where the stream drops over the risers of the glacial stairway (see description above). Examples of the hanging tributary type of fall, which is due directly to the activity of the glaciers, are Bird Woman and Grinnell Falls. Most lakes throughout the Park also owe their existence directly or indirectly to glaciers. They may be divided into five main types, depending on their origin: 1) cirque lakes, 2) other rock-basin lakes, 3) lakes held in by outwash, 4) lakes held by alluvial fans, and 5) moraine lakes (Dyson 1966).

Cirque lakes fill the depression plucked out of solid rock by a glacier at its source. Other rock-basin lakes fill basins created where glaciers moved over areas of comparatively weak rock. In all cases of cirque or rock-basin lakes, the water is contained by a bedrock dam. A typical example of this type of feature is Swiftcurrent Lake. Lakes held in by glacial outwash are dammed by stratified gravel which was washed out from former glaciers when they extended down into the lower parts of the valleys. Lake McDonald exemplifies this type of lake (figure 44). Lakes held by alluvial fans differ from the previous lake type in that they may have started as rock-basin lakes, but at a relatively recent date streams entering the lake valley have completely blocked the valley with deposits of gravel; thus creating a lake or raising the level of the one already present. St. Mary and Lower St. Mary lakes probably were joined originally, but the alluvial fan of Divide Creek, entering the basin from the south created a dam which cut the original lake body into two separate bodies. Finally, moraine lakes are formed when a moraine deposit blocks a stream outlet. Josephine Lake is a prominent example of a moraine lake. Another type of unique moraine lake has a glacier for part of its shoreline. In Glacier National Park, there are two of these lakes at Sperry Glacier and one at Grinnell Glacier.

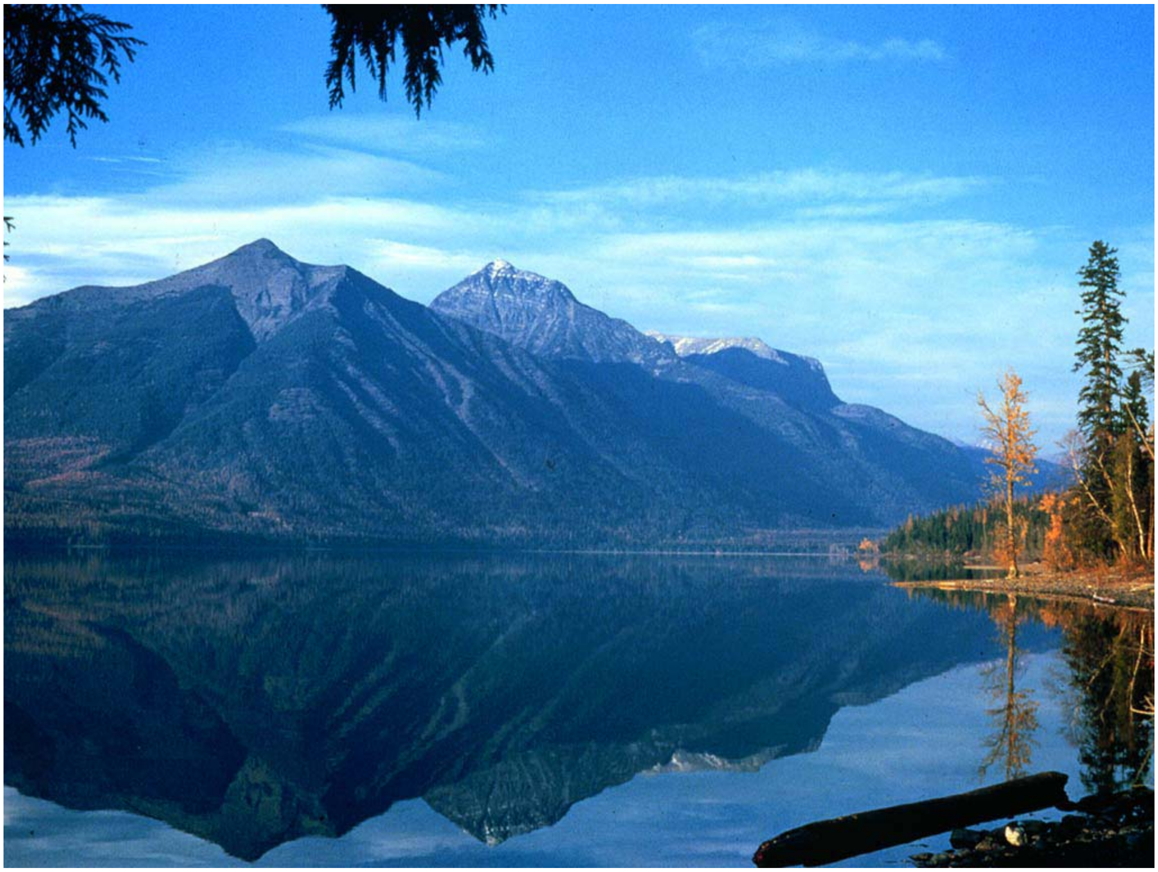


Figure 44: View of Lake McDonald from the moraine which dams it, Glacier National Park, NPS photo.

As the glacial ice moves it continually breaks rock fragments lose. Some of these are ground into a type of powder as they move against each other and against the bedrock under the glacier. Most types of rock, especially the limestones and shales on which the glaciers rest in Glacier National Park, yield a gray powder when finely ground. All melt-water streams issuing from present-day glaciers are cloudy or milky from their load of this finely ground “rock flour”. Much of this silt is deposited in lakes giving them a unique, characteristic milky, turquoise color.

Although the former large glaciers of the Ice Age transported huge amounts of rock debris down the valleys of the Park, the moraines which they deposited are, as a rule, not conspicuous features of the landscape. They are susceptible to intense erosion immediately following deposition as well as the obscuring effects of a vegetative cover of over 10,000 years The Going-to-the-Sun Road transverses a number of moraines along the shore of Lake McDonald. Because of the large proportions of rock flour (clay) in these accumulations, the material continually slumps, sometimes sliding onto the road surface. Knowledge of the location of such deposits is critical in Park resource management decisions.

All the present-day bodies of ice in Glacier National Park lie at the heads of valleys with high steep headwalls, on the east and north sides of high ridges at elevations between 6,000 and 9,000 feet, in all cases well below the snowline (the elevation above which more snow falls in winter than can be melted or evaporated during the summer, about 10,000 feet in Glacier). Consequently, these glaciers owe their origin and existence almost entirely to wind-drifted snow (Dyson 1966; Allen et al. 1995). Ice within these glaciers moves slowly. The average rate in the smallest ones may be as low as 6 to 8 feet a year, and in the largest glaciers probably 25 to 30 feet a year. There is no time of year when the glaciers are motionless, although movement is somewhat slower in winter than in summer. Despite its slow speed, over a period of years, the glacial ice transports large quantities of rock material ultimately to the glacier's end where it is piled up in the form of a moraine. The largest glacier in the Park is Grinnell Glacier. In 1960 it had a surface area of 315 acres. Sperry Glacier is the second largest glacier in the Park. Its surface in

1960 was 287 acres (Figure 45). Both Grinnell and Sperry Glaciers probably have maximum thicknesses of 400 to 500 feet (Dryson 1966; Key et al. 1996).

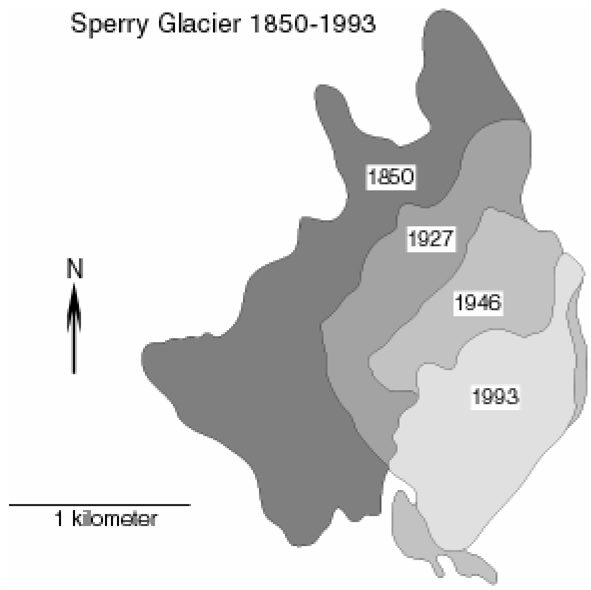


Figure 45: The areal extent of Sperry Glacier reflecting steady glacial retreat from 1850-1993. Modified from Key et al. 1996

Other important Park glaciers, although much smaller than the first two mentioned, are Harrison, Chaney, Sexton, Jackson, Blackfoot, Siyeh, and Ahern Glaciers (see figure 39; figure 46). Several others approach some of these in size, but because of isolated locations they are seldom measured. There are people who visit Glacier National Park without seeing a single glacier, while others, although they actually see glaciers, leave the park without realizing they have seen them. This is because the roadways afford only distant views of the glaciers, which from a distance appear as mere accumulations of snow.

Figure 46: View of Jackson Glacier from the Jackson Glacier Overlook in Glacier National Park, photo from Bad Rock Country B & B.

There is a myriad of interesting, short-lived surface features which can be seen at times on any glacier. These include crevasses, moulins (glacier wells), debris cones, and glacier tables (figure 47). Crevasses are cracks which occur in the ice of all glaciers due to tensions caused by differences in ice velocity throughout the body of the glacier. Debris cones result from the insulating effect of rock debris, usually deposited by a stream running over the glacier's surface, which protects the ice underneath from the sun's rays. As the surface of the glacier is lowered by melting, cones or mounds form beneath the rock-insulated area and grow gradually higher until the debris slides from them. They are seldom higher than 3 or 4 feet. A glacier table is a mound of ice which is capped and insulated by a large boulder. Its history is similar to that of the debris cone or mound (Dyson 1966). Snow which fills crevasses and wells during the winter often melts out from below leaving thin snowbridges over the cracks in the early part of the summer. These pose a very real danger to those traveling on a glacier because of their inherent weakness and instability.

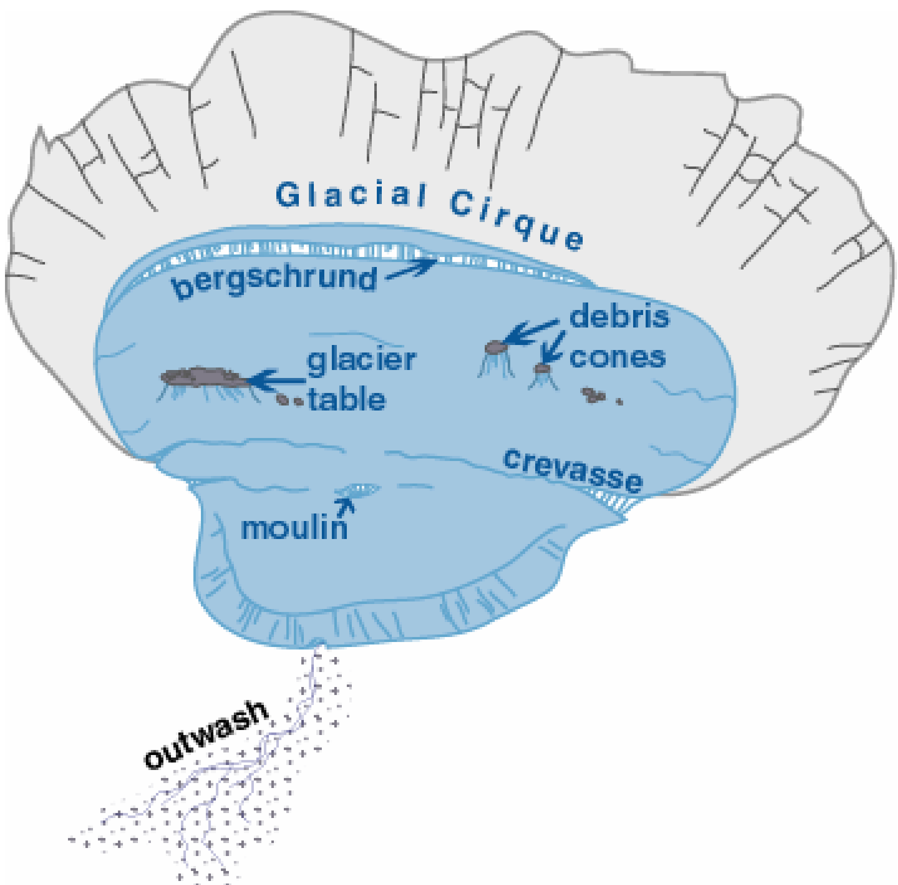


Figure 47: Short-lived surface features on glaciers.

Prior to the beginning of the 20^{th} century all glaciers in the Park, some 150 in number, began to shrink in response to a slight change in climate, probably involving both a temperature rise and a decrease in annual precipitation. From about 1900 to 1945 glacial retreat was very rapid (Dyson 1966). Over a period of several years such shrinkage is apparent to the eye of an observer and is manifested by a lowering of the glacier's surface elevation and a reduction of surface area. When the yearly snow accumulation decreases, the ice front of the glacier seems to retreat or move back, whereas the mass of the glacier is merely decreasing by melting on top and along the edges, analogous to an ice cube melting on a kitchen counter.

Evidence for a general, regional recession is not new, but was recognized early this century (Dyson 1940). Thorough glacier surveys began with mapping the first USGS 30-minute quadrangles (1904-1914), and the work of Alden (1914). Only two glaciers, Sperry and Grinnell, have repeated assessments of surface profiles and movement. An additional nine glaciers have been mapped at varying intervals, extending back to the mid-1800's. Since then, retreat has been consistent, with previous studies referencing conditions up to the late 1970's (Key et al. 1996). The National Park Service initiated observations on glacial variations in 1931. At first the work consisted only of the determination of the year by year changes in the ice front of each of the several glaciers. From 1937 to 1939, the program was expanded to include the detailed mapping of Grinnell, Sperry, and Jackson Glaciers to serve as a basis for comparisons in future years. Aerial photographs were obtained of all the known Park glaciers in 1950 and 1952 and again in 1960. Since 1945, the glacier observations have been carried on in cooperation with the U.S. Geological Survey. The work has included the periodic measurement of profiles to determine changes occurring in the surface elevation of Grinnell and Sperry Glaciers and also the determination of the rate of annual movement (Dyson 1966).

From six glaciers mapped from 1993 aerial photos, and two mapped through 1979, retreat from 1850 maxima ranged from 818-1440 m and averaged 1244 m. Overall retreat rates varied between 6-17 m/yr. Those glaciers were reduced in area by 62-80%, for an average shrinkage to 27% of the estimated area in 1850. Retreat rates were never constant over time on any single glacier, but roughly correlative with warmer climate trends. Pulses of recession occurred during the 1920's through the mid-1940's, and seem to be recurring now, as evidenced by dramatic change in glacier size since 1979. This inconsistency also means there have been periodic glacial advances throughout recorded history in Glacier National Park. Between 1966 and 1979, several of the larger glaciers in the Mount Jackson area advanced as much as 100 m (Key et al. 1996). This means that glacial response to climatic trends is very rapid, and could be relevant to the recent worries about the greenhouse effect and global warming.

Recent moraines of two different age groups have been identified fronting the present-day glaciers and snowfields in Glacier National Park, Montana, ranging in size from a few feet to more than 200 (Dyson 1966; Carrara 1987). The subdued, vegetated moraines (sparse willows and other forms of dwarf vegetation) of the older group have been found at 25 sites, mainly in the central part of the Lewis Range. These older moraines are in places overlain by volcanic ash from Mount Mazama in Oregon, and therefore predate

6,800 years B.P. (Carrara and Wilcox 1984; Carrara 1986; Osborn and Gerloff 1997). The younger set of moraines, which has accumulated during the last several hundred years, consists of fresh bouldery rubble on which only small pioneer plants and lichens have begun to establish themselves. They are common throughout Glacier Park (Dyson 1966; Carrara 1986). Tree-ring analyses indicate that some of these younger moraines were deposited by advances that culminated during the mid-19th century (Carrara and Wilcox 1984; Carrara 1986). The moraines are particularly striking at Grinnell, Sperry, Blackfoot, Agassiz and Sexton Glaciers. Because of recent glacial retreat most, if not all, of the glaciers are no longer in contact with these newer moraines. In some cases a quarter of a mile or bare rock surface intervenes between the moraine and the glacier terminus (figure 48) (Dyson 1966).

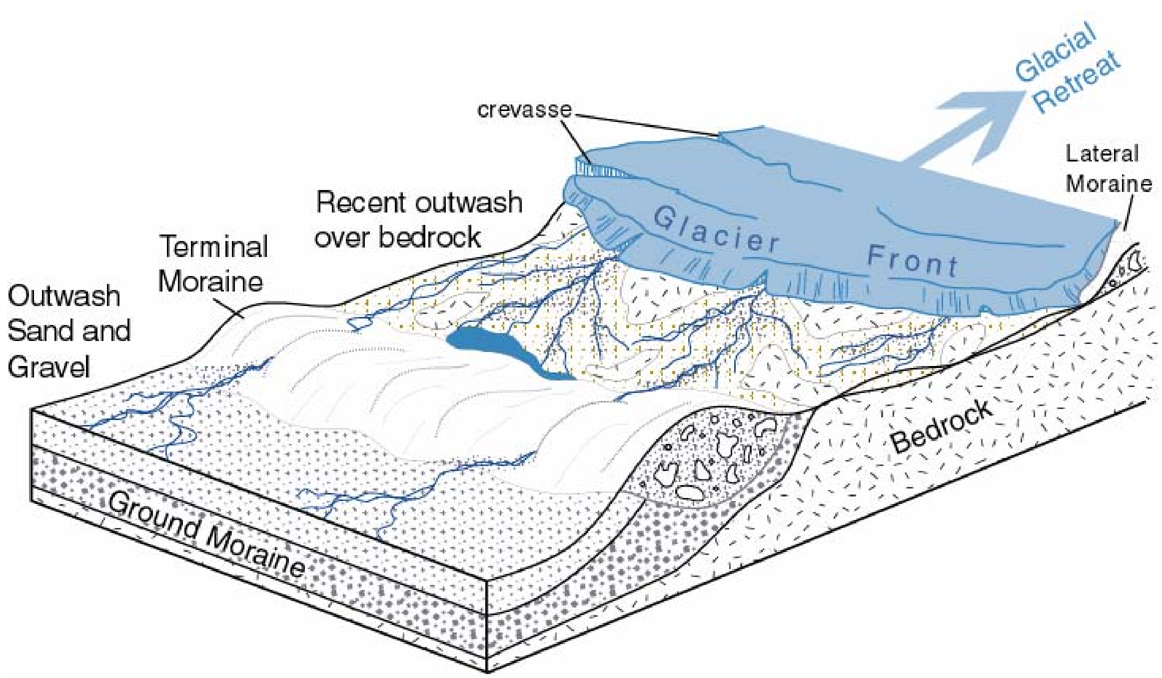


Figure 48: Diagrammatic view of a retreating glacier with a recent terminal moraine and a broad space between the moraine and the glacier front. Note the variety of deposits associated with glacial melt.

A few glaciers have disappeared within recent years, but their distinctive moraines remain as evidence of former glacier activity. One of the most notable examples is the former Clements Glacier, a small body of ice which existed until about 1938 in the shadow of Clements Mountain at Logan Pass. The trail to Hidden Lake skirts the outside edge of the moraine. Between the mountain and moraine lies an obvious, bare expanse of rock where the Clements Glacier once existed (Dyson 1966).

Hydrogeology and Glacier National Park

Water seems to be everywhere in Glacier from sparking lakes to breathtaking waterfalls. Glacier's clear rivers and lakes inspire visitors with a sense of ecological harmony. Because of the dynamic nature of the interaction between water and geologic processes at Glacier National Park the role of hydrogeology must be taken into consideration when making management decisions.

Water is a prevalent presence and preserving its pristine state is crucial. The potential for pollution from human waste facilities, campgrounds, watercraft, and vehicular use requires intense regulation of the human impact in the Park. Groundwater, either from precipitation or snow melt moves quickly through the fractured bedrock and down steep hydraulic gradients in the Park.

The interaction between water and geology is especially obvious in a mountainous environment like Glacier which receives significant snowfall each year. The glacial features described above lend a unique aspect into the situation. The moraines left by glaciers commonly act as dams to the drainages in front of the glaciers. These can be found in any scale in Glacier from Lake McDonald, to a small alpine tarn lake. If these dams should fail, the potential flash flood would have a tremendous impact on the valley below the lake. Similarly, avalanches or ice rafts from glaciers can form temporary dams across waterways. One dramatic example of this was the damming of the Middle Fork of the Flathead River by an avalanche, destroying a bridge, in 1979. When the dam inevitably melted, the rush of water and debris destroyed many structures along U.S.

Highway 2, and it created a very long detour for motorists and a large mess for clean-up crews and residents below the slide. Just recently, during the winter of 2002-2003, an avalanche destroyed some of the railroad and blocked the highway. These processes leave a large mark on the landscape and the Park should be prepared to adjust accessibility when necessary to protect the resources and public safety.

Water continues to play a critical role in sculpting the present landscape of Glacier National Park. During intense seasonal thunderstorms, rain acts like a sledgehammer on unprotected soil, knocking apart individual soil particles and washing unconsolidated sediment into the valleys. The steep slopes of Glacier's valley walls are susceptible to intense erosion and rock fall during these storms. Water freezing and thawing also plays a large role in changing the landscape. The expansion upon freezing pries rocks apart, crack by crack.

The thin, poorly sorted, glacially-derived alpine soils of glacier are susceptible to slumping and sliding especially when water-saturated. This can be an extremely hazardous situation for roadways and trails which are situated on or near steep slopes and planners must be prepared to predict when the potential hazard is high and adjust accessibility to those areas accordingly. Slide Lake for instance, formed when a landslide blocked Sherburne Creek, diverting forever the trail that once was along the lakeshore.

The North Fork Valley is largely underlain by the Tertiary age Kishenehn Formation. This formation contains clays derived from volcanic sediments including bentonite.

Bentonite is a clay mineral which greatly expands when wet causing the rocks around it to crack and move. This type of shrink-and-swell clay poses an obstacle to road and trail builders. The presence of bentonite and its hydrogeologic properties in the North Fork Valley should be noted in the planning of future roads, facilities, and/or trails.

Paleontology

The Geologic Resources Division of the National Park Service is conducting a separate paleontology inventory of the National Parks and Monuments, so a detailed description of the paleontology and biostratigraphy of Glacier National Park is beyond the scope of this report. The sections that follow are simple lists of the stromatolites from the Mid Proterozoic Belt Supergroup rocks and some fauna and flora associated with the Cretaceous age rocks (exposed beneath the Lewis thrust fault and to the east of Glacier National Park) that have been reported in the literature.

Altyn Formation

- *Baicalia-like* stromatolites are common

Grinnell Formation

- Mound-shaped stromatolites locally

Empire Formation

- Mound-shaped stromatolites locally

Helena Formation

- *Baicalia* and *Conophyton* stromatolites abundant, mound-shaped stromatolites present

Snowslip Formation

- Mound-shaped stromatolites locally abundant

Shepard Formation

- Stromatolites are relative uncommon

Kootenai Formation

- *Protelliptio douglassi, P. reesidei,* and *Lampsilis farri,* and the gastropods *Stantonogyra silberlingi* and questionably *Reesidella montanaensi* and some poorly preserved clams (Rice and Cobban 1977).

Blackleaf Formation

- *Arenicolites*. Some small burrows, and plant remains (Rice and Cobban 1977).

Marias River Shale

- *Inoceramus (Mytiloides) labiatus, Ostrea, Watinoceras reesidei, Scaphites nigricollensis*. The limestone concretions in the Kevin Member contain fossils *of Inoceramus deofmis, Baculites mariasnsis,* and *Scaphites preventricosus.*. Coccoliths (Rice and Cobban 1977).

Telegraph Creek Formation

- Some molluscan fossils, inoceramids and oysters, all rare (Rice and Cobban 1977).

Virgelle Sandstone

- Ammonite *Desmoscaphites bassleri* from a locality near the eastern boundary of Glacier National Park. *Inoceramus lundbreckensis* has also been found in rocks assigned to the Virgelle Sandstone in the Blackfeet Indian Reservation east of Glacier National Park (Rice and Cobban 1977).

Two Medicine Formation

- The dinosaur *Barchyceratops montanesis*, scales of ganoid fishes, ostracodes and fresh-water mollusks (Rice and Cobban 1977).

Bearpaw Shale

- Zones of *Baculites compressus*, *B. coneatus*, and *B. reesidei* have been found in the Bearpaw Shale on the Blackfeet Indian Reservation east of Glacier National Park. In addition to the baculites, limestone concretions in the Bearpaw contain the ammonites *Hoploscaphites* and *Placenticeras*, and the bivalves *Nucula*, *Nuculana*, *Inoceramus*, *Oxytoma*, *Cymella* and *Nymphalucina*. The boundary between the Campanian and Maestrichtian Stages may be near the base of the Baculites reesidei Zone (Rice and Cobban 1977).

Horsethief Sandstone

- The brackish water bivalves *Crassostrea wyomingensis* and *Veloritina occidentalis* and the gastropod *Melania wyomingensis* have been found at the top of the Horsethief Sandstone, whereas a shallow-water marine bivalve, *Tancredia*?, occurs lower in the Horsethief (Rice and Cobban 1977).

St. Mary River Formation

- Nonmarine bivalves, such as *Fusconaia*? *stantoni*, and locally, fossil leaves. Dinosaur bones occur throughout the formation, but articulated skeletons are very rare. An incomplete skeleton of *Montanaceratops* was found on the Blackfeet Indian Reservation (Rice and Cobban 1977).

Willow Creek Formation

- Some freshwater molluscan fossils, dinosaur bones, all rare. The top of the Cretaceous is somewhere within this formation (Rice and Cobban 1977).

Kishenehn Formation

- Some petrified wood, including leaves and wood from the Dawn Redwood, fossil leaves abundant. Gastropods, fossil mammals, fish, insects, mollusks, and leaves from *Macginitea augustiloba* are common.

Unique Geological Features

Understanding the geology of Glacier enhances one's understanding of the unique relationship between geology and their environment. It gives a deeper meaning to the spectacular scenery which astonishes visitors every time they visit Glacier National Park. The special geologic quality of Glacier first rests in the rocks themselves. The preservation of minute sedimentary details in the Belt Supergroup rocks incites interest in anyone who sees them. Features such as mudcracks, rain drop impressions, ripples, storm rip-up layers and stromatolites are all tangible in many places in the park. These are among the finest Proterozoic sedimentary rocks preserved on earth!

Building off the rocks themselves are the structural features on display in Glacier. The Lewis thrust fault, visible along U.S. Highway 2 in grand fashion has long been considered among the classic examples of faulting in the Rocky Mountains. Often, large-scale faults are either buried or so disseminated they are unrecognizable to the untrained eye, yet anyone who looks into the Park from Marias Pass can see clearly the surface of a major thrust fault in addition to the juxtaposition of two clearly different rocks on either side.

Further sculpting the rocks and structures of Glacier National Park are the forces of erosion by water, liquid or frozen. As described in detail earlier, the action of glaciers affords Glacier some incredible scenery. Glaciers are responsible for the grand U-shaped valleys, punctuated by hanging valleys with cascading waterfalls along the sides. Glaciers carve cirques and other depressions for the incredible abundance of lakes.

Glaciers shave mountains into knife-like ridges and horns and inundate lower areas creating cols or passes. The deposits left by glaciers dam rivers to create long, narrow lakes.

Finally, among other dynamic features in Glacier National Park, some 50-60 actual glaciers exist to capture the imagination. Visitors can see glaciers in action, eroding or retreating, creating glacial deposits that litter bedrock surfaces and turn lake water into a vibrant turquoise hue. A glacier is never the same two days in a row. Its surface is always changing, creating vast crevasses, debris tables and cones, as well as snow bridges. Glaciers can not exist anywhere, the right climate and elevation as well as aspect and slope are needed to accumulate the snowfall necessary for glacial upkeep. For this reason the presence of active glaciers is a true gift in Glacier.

Acknowledgments

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Appendices

Appendix A: Description of Geologic Map Units

Quaternary

Qal Alluvium (Holocene and upper *Pleistocene*)

- Sand and gravel deposits locally containing thin lenses of silt. Includes channel and overbank deposits in modern floodplains, as well as alluvial-fan and terrace deposits. Unit consists mainly of rounded and subrounded clasts of Belt Supergroup rocks. Thickness 1-10 m.
- *al (Holocene)* Sand and gravel deposits and minor amounts of silt; in places silt forms lenses. Includes narrow mountain valleys locally includes small areas of colluvium. Unit consists mainly of rounded and subrounded clasts of Belt Supergroup rocks; other rock types, chiefly coarse-grained granites, are also present in minor amounts along the North Fork Flathead River. Thickness 1-5 m.

af Alluvial-fan deposit (Holocene and late *Pleistocene*)

- Fan-shaped deposits of fluvial sand and gravel. In places unit contains thin lenses of silt. Unit consists chiefly of rounded and subrounded clasts of Belt Supergroup rocks. Locally includes debris-flow deposits. Thickness 2-50 m.

Qc Colluvium (Holocene and upper Pleistocene)

- Locally derived slope deposits consisting of unsorted, angular, gravel-size clasts in a matrix of unsorted sand, silt, and clay. Unit locally includes small areas of till, talus, rock-avalanch4e, and debris-flow deposits. Commonly 1-5 m thick.
- *co Colluvial deposit (Holocene and late Pleistocene)* Mainly locally derived slope deposits consisting of unsorted angular grave-size clasts in a matrix of unsorted

sand, silt and clay. Unit locally includes some small areas of bedrock and till as well as talus, rock avalanche, and debris flow deposits. Commonly 1-5 m thick.

ac ***Alluvial and colluvial deposit (Holocene and late*** *Pleistocene*)

- Locally derived deposits of silt and sand on Flattop Mountain. Unit locally includes sandy and silty sheetwash deposits. Thickness 0.5-2 m.

Qls ***Landslide deposits (Holocene and upper*** *Pleistocene*)

- Includes large slumps, block slides, and earth flows. Slumps are common in east side of park in areas underlain by Cretaceous sedimentary rocks. Block slides are present, although not common, in areas underlain by Belt Supergroup rocks. Block slides and earth flows are common in west side of park in areas underlain by sedimentary rocks of the Kishenehn Formation. Some of the larger landslides deposits in park exceed 50 m in thickness and cover several square kilometers. Unit locally includes small areas of till, rock glaciers talus and colluvium.
- *ls (Holocene and late Pleistocene)* Unit includes large rock slumps, slump-earth flows, and rock block slides. The size and the kind of clasts and the grain size of the matrix vary according to the bedrock units involved in the landslide. Rock slumps are common in the eastern side of the park in those areas underlain by Cretaceous sedimentary rocks. Rock block slides, although not common, are present in areas underlain by Belt Supergroup rocks. Rock slumps and slump-earth flows are common in areas in the western side of the park underlain by the soft sedimentary rocks of the late Paleogene Kishenehn Formation. Some of the larger landslides exceed 50 m in thickness and cover several square kilometers. Locally includes small areas of till and colluvium.

rg ***Rock-glacier deposit (Holocene and late*** *Pleistocene*)

- Lobate masses of unsorted angular blocky rubble; interstices filled with unsorted sand, silt, clay, and ice. Unit occurs at the head of some cirques. Thickness 10-30 m.

ta ***Talus deposit (Holocene and late*** *Pleistocene*)

- Unsorted and mainly unvegetated, angular, bouldery rubble in a matrix of sand, silt and clay at bases of steep valley walls or cliffs. Some of the larger deposits exceed 30 m in thickness.

so ***Solifluction and related deposit (Holocene and*** *Pleistocene*)

- Includes solifluction lobes, sorted polygons, and sorted stone stripes. Unit found mainly in unglaciated upland areas. Locally includes other mass-wasting deposits. Thickness 0.5-2 m.

or ***Organic deposit (Holocene and late*** *Pleistocene*)

- Peat and organic muds. Common in the valley of the North Fork Flathead River. Thickness 2-5 m.

Qg ***Till (Holocene and upper Pleistocene)***

- Unsorted, subrounded to subangular bouldery ruble, consisting mainly of Belt Supergroup rocks, and lesser amounts of sand, silt and clay. Striated rocks common. In valleys of the North and Middle Forks Flathead River, unit deposited as a thick (locally >30 m) blanket of ground moraine by large trunk glaciers that filled these valleys. On valley floors in mountainous areas, unit deposited by locally mountain glaciers as ground moraine usually 1-3 m thick. In front of many of the glaciers and snowfields in higher regions of park, unit forms moraines 3-50 m high. On Boulder, Cut Bank, and Swiftcurrent Ridges, unit also includes "pre-Wisconsin glacial drift" of Alden (1912), which in places is as much as 60 m thick. Unit also locally includes small areas of bedrock and colluvium.

- *t1 (late Holocene)* Unsorted subrounded to subangular bouldery rubble, consisting of Belt Supergroup rocks, and minor amounts of sand, silt, and clay. Striated rocks are common. Unit forms steep, rubbly moraines 10-50 m high in front of many of the glaciers and snowfields in the park. Unit is unweathered and supports little vegetation. Many of these moraines were deposited by glacial advances during the mid-19th century.
- *t2 (late Pleistocene)* Unsorted subrounded to subangular bouldery rubble, consisting of Belt Supergroup rocks, and minor amounts of sand, silt, and clay. Unit commonly forms subdued, vegetated moraines 3-10 m high immediately downvalley from t1 deposits. Unit supports thin soil, which in places contains Mazama ash dated at about 6,845 B.P. Unit is though to date from about 10,000 B.P. or slightly earlier.
- *t3 (late Pleistocene)* Unsorted subrounded to subangular bouldery rubble, consisting mainly of Belt Supergroup rocks, and minor amounts of sand, silt and clay. Striated rocks are common. Found on valley floors within mountainous areas where it was deposited as ground moraine by locally mountain glaciers; here, its thickness is usually 1-3 m. also found in the valleys of the North and Middle Forks Flathead River where it was deposited as a thick blanket of ground moraine by the large trunk glaciers that filled these valleys; here, its thickness exceeds 30

m in places. Locally includes small areas of bedrock and colluvium. This unit is in places also overlain by the Glacier Peak G ash.

at ***Ablation till (late*** *Pleistocene*)

- Unsorted subrounded to subangular bouldery rubble, consisting mainly of Belt Supergroup rocks, and minor amounts of sand, silt, and clay. Striated rocks are common. This unit, deposited by stagnating mountain glaciers, forms hummocky, poorly drained deposits in valley tributary to the valley of the North Fork Flathead River. In places this unit overlies t3 deposits, yet it is also older than some t3 deposits that lie upvalley. Thickness exceeds 40 m at some localities. This unit is in places overlain by the Glacier Peak G ash, which has been dated at about 11,200 B.P.

es ***Esker deposit (late*** *Pleistocene*)

- Identified in two areas: (1) along north side of Lake McDonald in the Fish Creek campground area, where deposit consists of well-sorted silty sand and gravel composed of Belt Supergroup rocks; (2) in the Railroad Creek area in southeastern corner of park, where deposits consists of sand, silt, and clay derived from the local bedrock. In both areas these deposits form sinuous ridges 0.5-1 km long and about 10-20 m high..

tg ***Terrace deposit (late*** *Pleistocene*)

- Sand and gravel deposits underlying terraces along the North and Middle Forks Flathead River and the mouths of Ole and Park Creeks. Unit consists mainly of rounded and subrounded clasts of Belt Supergroup rocks. Locally contains thin lenses of silt. Terraces range from 3 to 20 m above present stream levels. Thickness 2-10 m.

di ***Diamicton (early*** *Pleistocene or Pliocene?*)

- Unsorted subrounded to subangular bouldery rubble, consisting of Belt Supergroup rocks, and minor amounts of sand, silt and clay. Unit occurs beneath Boulder, Cut Bank, and Swiftcurrent Ridge. Striated rocks are common. In places, unit is weakly cemented by calcium carbonate. Locally, unit is as much as 60 m thick. This unit is equivalent to the "pre-Wisconsin glacial drift."

Tertiary

Tku ***Kishenehn Formation (Oligocene and*** *Eocene*)

- Generally divisible into two parts.
- Upper part is a sequence of brick-red, red-brown, and vermilion mudstone, sandstone, and conglomerate and interbedded gray, calcareous, sandy pebble and cobble conglomerate. Mudstone beds have yielded fossil gastropods, mammals, and palynomorphs. Maximum thickness about 1,500 m.
- Lower part consists of light-gray to gray-green sandstone, siltstone, mudstone, lignite, oil shale, marlstone, and sandy Pebble and cobble conglomerate. Gastropod fossils prevalent throughout lower part. Maximum thickness at least 3,500 m.

Tkp ***Kishenehn Formation – Conglomerate member of Pinchot Creek (****Eocene*)

- Brick-red, red-brown, and maroon, intercalated mudstone, sandstone, and conglomerate. Locally, calcareous sandy mudstone and siltstone grade to muddy sandstone; muddy sandstone is composed of varicolored, angular to subrounded, sand- and pebble-size clasts. Pebble and boulder conglomerate beds are gray and sandy and contain abundant mudstone and sandstone matrix material; rounded and subrounded lithic clasts in pebble and boulder conglomerate beds consist entirely of angular to subrounded Belt Supergroup rocks that attain a maximum size of 2.5 m. Vertebrate fossils in some mudstone units. Conglomerate member rests conformably on the lacustrine member of Coal Creek. Maximum estimated thickness 700 m.

Tkcc Kishenehn Formation – Lacustrine member of Coal Creek (Oligocene and *Eocene*)

- Predominantly a light-gray, heterogeneous assemblage of sandstone, siltstone, mudstone, claystone, coal, oil shale, marlstone, and pebble and boulder conglomerate. Lacustrine member is informally divided into three parts, each bounded by gradational contacts. Total thickness about 1,150 m.
- Upper part consists typically of an interbedded sequence of marlstone, litharenite, siltstone, conglomerate, mudstone, claystone, and coal. Variegated maroon, red-brown, and gray-green color of sandy mudstone beds gives outcrops of upper part a distinctive pink cast., in contrast to the predominantly light-gray to gray color of lower and middle parts of member. Fossil gastropods and Eocene-age mammals common in mudstone beds. Thickness about 100 m.
- Middle part is interbedded oil shale, marlstone, litharenite and siltstone, and lesser amounts of lignite, sapropelic coal, tuff, claystone, and mudstone. Gastropod fossils extremely abundant; plants and plant fragments, fish, insects, and mollusks

also common in middle part. Fission-track analysis of zircon from a tuff bed suggest and Eocene age of 43.5±4.9 Ma (analysis by Charles Naeser, written communication, 1990). Thickness about 500 m.

- Lower part consists primarily of (1) interbedded carbonaceous siltstone, (23) silty of coarse-grained litharenite that displays climbing ripple and even, parallel lamination, and (3) light-bluish-gray-weathering oil shale. Lignite, mudstone, claystone, marlstone, conglomerate, and devitrified tuff beds present to a lesser extent. Eocene-age fossil leaves of *Macginitea augustiloba* present in beds of litharenite. About 550 m thick.

Cretaceous

***Km Marias River Shale (Upper** Cretaceous*)

- Dark-gray, marine mudstone and lesser amounts of interbedded sandstone, limestone, and arenaceous shale. Ranges from 365 to 395 m thick.

***Kb Blackleaf Formation (Upper? and Lower** Cretaceous*)

- Gray, marine mudstone and interbedded sandstone. About 260 m thick.

***Kk Kootenai Formation (Lower** Cretaceous*)

- Gray-green and maroon, non-marine mudstone and abundant lenticular beds of poorly sorted, greenish-gray sandstone that are locally crossbedded and contain lenticular interbeds of conglomerate. Brown to brownish-gray limestone and thin to thick lenses of coquina containing pelecypods and gastropods near top of formation. Brown, iron-stained limestone nodules common in mudstone beds. Thickness ranges form 198 to more than 305 m.

***Kbk Blackleaf and Kootenai Formations, undivided (Upper? and Lower Cretaceous*)**

- Unit shown where parts of both formations present but too thin to map separately.

Ku Sedimentary rocks, undivided (*Cretaceous*)

- Lithostratigraphic units not identified or mapped separately.

Cretaceous - Jurassic

KJm Mount Pablo Formation (Lower Cretaceous*), Morrison Formation (Upper Jurassic), and Ellis Group (Upper and Middle Jurassic), undivided*

- Units not mapped separately.

Late – Middle Proterozoic

***Zd Diorite, diabase, and gabbro sills and dikes (Late Proterozoic?*)**

- Dark gray to greenish black, metamorphosed, fine to medium grained, equigranular. Abundant chlorite replaces amphibole and pyroxene; commonly pyritic. Laterally continuous sill in Helena Formation cuts up and down section locally, ranges in thickness form 16 to 90 m, and is commonly flanked by bleached zones of hornfels. Sills (0.5 – 65 m thick) also occur locally in Grinnell, Empire, and Snowslip Formations. Dikes intrude the Appekunny and Altyn Formations on east side of Glacier National Park and intrude the Purcell Lava and overlying strata near Granite Park; dikes are 3-6 m thick. Sills and dikes closely resemble Late Proterozoic intrusive rocks in west-central Montana reported to be 750±25Ma (potassium-argon age method).

***Ym McNamara Formation (Middle Proterozoic*)**

- Grayish-green siltite and argillite, commonly interlaminated as wavy, nonparallel fining-upward couplets. Contains abundant beds of mudchip breccia; locally, some breccia clasts and thin discontinuous laminae of argillite are silicified. Calcareous and quartzose arenite beds near base. Some calcareous siltite beds also present in lower part. Ripple marks and subaqueous shrinkage cracks common; salt casts rare. Top for formation not exposed; base of formation places on top of uppermost red feldspathic arenite and siltite of the Bonner Quartzite. Present in southern part of park east of Mount Shields (600 m exposed) and near mouth of Coal Creek (450 m exposed).

***Ybo Bonner Quartzite(Middle Proterozoic*)**

- Pinkish-gray to pale-red, very fine grained to medium-grained feldspathic arenite and lesser amounts of interbedded siltite and dark-red argillite. Rhythmic fining-upward successions as much as 3 m thick. Large-scale channels, crossbedding, and ripple cross-lamination common. Lower contact placed at base of lowermost green feldspathic arenite bed. Exposed east of Mount Shields and near mouth of Coal Creek; thickness 250-280 m.

***Yms Mt. Shields Formation (Middle Proterozoic*)**

- At type section near Mount Shields in southern part of park, the Mount Shields Formation is informally subdivided into five members designated 1 through 5 in ascending order. Members are not mapped separately but can be recognized throughout exposures in park. Maximum thickness about 850 m
- Member 5 is characterized by very thinly laminated, blackish-green argillite and some thin, lenticular beds or arenaceous siltite that are more abundant near top of

member. This distinctive succession of blackish-green argillite is locally calcareous and sharply overlain by pale-green, coarse-grained, poorly sorted feldspathic arenite of the Bonner Quartzite; lower contact places at base of lowermost interval of thinly laminated blackish-green argillite and siltite. Thickness about 9 m.

- Member 4 consists mostly of grayish-green, fining upward couplets of siltite and argillite and contains carbonate mostly as cement in siltite. Salt casts common, particularly in lower part. Lower contact is placed on top of uppermost noncalcareous, pale-purple siltite and argillite interval of member 3. About 17 m thick at type section but appears to thicken and contain more carbonate beds northward in park.
- Member 3 is mostly couplets of siltite and argillite. Siltite laminae in couplets successively change color upward from brick red in lower part of member to purplish gray in middle part to dark grayish green near top; argillite laminae in couplet s remain dark red to pale purple throughout member. Salt casts and ripple marks are common, but salt casts become less abundant downward as arenite beds increase and argillite beds decrease. Lower part contains pale-purple to brick-red, very fine grained arenite similar to that in member 2 but in equal proportion to siltite and argillite. Base of member 3 is placed on top of uppermost bed of stromatolitic limestone of member 2. Member 3 is thickest (450 m) member of formation throughout park.
- Member 2 consists mostly of thin, fining-upward successions of brick-red, very fine grained arenite and coarse-grained siltite capped locally by dark-red argillite;

member 2 contains more arenite than other members. Ripple cross-lamination and some even, parallel lamination are common in lower part of successions. Pink to cream limestone beds at top of member contain oolites and small stromatolite heads; this zone is recognized throughout northern part of Belt basin. Base of member 2 is placed at base of lowermost succession of brick-red arenite and siltite. About 270 m thick.\

- Member 1 consists of thinly laminated, maroon to pale-purple argillite, brick-red siltite, and some interbedded arenaceous siltite and thin intervals of greenish-gray siltite and argillite. Lower contact placed on top of uppermost sequence of dolomitic siltite of Shepard Formation. Near northern boundary of park., member 1 encloses basaltic lava (shown by black symbol), and because the lava is only about 10.5 m thick and near the base of member 1, it is shown at contact between Mount Shields and underlying Shepard Formation. Member 1 is about 30 m thick.

***Ysh Shepard Formation (Middle Proterozoic*)**

- Typically consists of yellowish-gray to greenish-gray dolomitic and pyritic siltite and argillite and a few thin beds of coarse-grained calcarenite, quartz arenite, limestone, and dolomite. A succession of very thinly laminated, olive-green argillite beds about 53 m thick occurs in lower part of formation near Mount Shields at southern edge of park but is not present near U.S.- Canada boundary. Thin beds of stromatolitic limestone are common on southern exposures but are rare in northern parts of park. Lamination is generally wavy nonparallel and composed of fining-upward couplets. Fluid-escape structures, shrinkage cracks, ripple marks., miniature molar-tooth structures, and mud-chip breccias are common. Because of the carbonate and pyrite content of strata, most exposures weather tan to dusky orange. Lower contact placed at base of lowermost bed of dolomitic siltite or dolomite. Thickness about 400 m in southern part of park; thins northward to 165 m at Hole-in-the-Wall and westward to 210 m in Apgar Mountains.

Ypb Purcell Lava (Middle Proterozoic)

- Grayish-green to dark-greenish-gray mafic lava flow(s) (which can be subdivided into three facies) and a hypabyssal sill. In the park, the Purcell occurs within strata of the upper part of the Snowslip Formation as defined here. Because of map scale and position of lava in the uppermost part of the Snowslip, the Purcell is shown on map at contact between the Snowslip and Shepard Formation in northern part of park. Maximum total thickness of the Purcell is 92 m in northernmost exposures at Hole-in-the-Wall; thinks southward to 19 m at Granite Park and pinches out at Huckleberry Mountain.
- Upper facies of subaerially emplaced pahoehoe ranges in thickness from 0 to 54 m, is a compound flow sequence of multiple flow units (0.6 – 6 m thick), and overlies a lower pillow-lava facies; ropy flow structures are common on upper flow surfaces. Lower pillow-lava facies is 9-15 m thick and consists of interconnected pillows, which range in diameter from 20 cm to 2 m, and associated hyaloclastite breccia. Locally, a third facies of vent rock is confined to north-central exposures in park; it forms a lens-shaped, chaotic breccia (maximum thickness 10 m) containing randomly distributed, angular to

subrounded, equidimensional cognate blocks (5-35 cm) and lapilli intermixed with accidental, deformed and undeformed blocks of Snowslip strata (as much as 2 m long) in a devitrified, oxidized matrix. Vent facies is within pillow-lava facies and is overlain by pahoehoe flow units.

- A hypabyssal diabase sill, spatially correlative with vent facies but interpreted to be from a younger igneous event, is 18-21 m thick and generally enclosed by the Snowslip Formation about 5 m below base of pillow-lava facies.

Ysn Snowslip Formation (Middle Proterozoic)

- At type section at Snowslip Mountain in southern part of park, the Snowslip is informally subdivided into six members designated 1 though 6 in ascending order. Members are not mapped separately but can be recognized in exposures throughout park. Contact with underlying Helena Formation is sharp, apparently disconformable, and placed at base of fist occurrence of red lithic arenite that overlies gray limestone or dolomite of the Helena. The Snowslip ranges from about 360 m thick (including 95 m of Purcell Lava) at a reference section on east wall of Hole-in-the-Wall cirque to 635 m thick in Apgar Mountains (including 32.5 m of Purcell Lava); thickness about 490 m at type section.

- Member 6 consists of interbedded noncalcareous, grayish-green and pale-maroon, fine-grained arenite, siltite and minor argillite at type section. In northern part of park, member 6 conformably encloses the Purcell Lava and consists mostly of grayish-green siltite and argillite beneath the lava and alternating beds of pale-maroon and grayish-green, very fine grained arenite, siltite and argillite above the lava. Where the Purcell is present, thin discontinuous beds of pink and gray stromatolitic limestone occur in lower part of member 6. Base of members is placed on top of uppermost red, fining-upward succession of arenite and argillite of member 5. Thickness ranges from 8 to 130 m.

- Member 5 is composed of rhythmic, fining-upward successions as much as 8 m thick, but typically 2-3 m thick, of very fine grained to medium-grained, white to pink quartz arenite and subfeldspathic arenite, fining upward to siltite in middle of succession and dark-red argillite at top. Base of each succession is erosional and forms a sharp contact with dark-red argillite at top of underlying succession. Sedimentary structures include abundant ripple marks, desiccation cracks, mudchip breccias, and fluid-escape structures. Lower contact is placed at base of lowermost fining-upward succession that rest son calcareous strata of member 4. Thickness ranges from 35 to 145 m.
- Member 4 is predominantly wavy, nonparallel-laminated, grayish-green and yellowish-gray calcareous siltite and argillite. A few interbeds of very fine grained arenite and several thin, conspicuous beds of pink stromatolitic limestone are present, particularly in lower part of member. Lower contact is placed at base of lowermost sequence of calcareous grayish-green strata. Thickness ranges from about 85 to 140 m.
- Member 3 is similar to member 5 but rhythmic, fining-upward successions from arenite to argillite are not as regular as in member 5. For example, a succession could have arenite at base and argillite at top, but no siltite in middle, or a succession could be all siltite, coarse grained at base and very fine grained at top. Successions are as much as 3 m thick. At type section, rhythmic successions

show a more regular change in grain size, grading from arenite to argillite and are thicker than elsewhere in park. Thickness ranges from about 15 to 65 m.

- Member 2 is nearly identical to member 4; it includes a few interbeds of arenite and several thin beds of pink stromatolitic limestone, particularly in lower part of member. Thickness ranges from about 70 to 150 m.
- Member 1 is characterized by alternating pale-maroon and grayish-green sequences of calcareous siltite, argillite, and oolitic arenite. Arenite grains are fine to very coarse, moderately to poorly sorted, subrounded to rounded; arenite beds are thin and commonly cross-laminated. Siltite and argillite laminae are commonly arranged as wavy, nonparallel, fining-upward couplets that contain abundant mud-chip intraclast, subaqueous shrinkage cracks, and fluid-escape structures. Thickness ranges from 25 to 90 m.

Yh Helena Formation (Middle Proterozoic)

- Generally subdivisible into three parts at most exposures in park. Lower contact of the Helena is placed on top of a 1.8-m-thick interval of green argillite of the Empire Formation. Thickness ranges from 750 m in most of the park to a maximum of about 1,030 m in southwestern part. The following description is from a measured section along Going-to-the-Sun Road between Logan Pass and The Loop on west side of Park.
- Upper part consists primarily of interbedded stromatolitic limestone, dolomite, oolitic limestone, and quartz arenite. At base of upper part, and interval of stromatolitic limestone about 30 m thick, known as the *Conophyton* zone, is composed of *Baicalia-Conophyton* stromatolite cycles. Massive character of

Conophyton zone causes it to stand in relief in most exposures of the Helena Formation in park. A 40 m thick diorite sill intrudes the Helena in this part of measured section just above *Conophyton* zone. This sill, which is present throughout park, changes stratigraphic position from near the base of the Helena in southeastern part of park to the lower part of the Snowslip in northernmost part of park. Upper part is 235 m thick at measured section.

- Middle part is predominantly dolomitic molar-tooth beds, some as much as 30 m thick. A few thin beds of quartz arenite and stromatolitic limestone are present in the middle part of the Helena. Thickness 360 m at measured section.
- Lower part consists of thick, smoky-gray limestone beds near top, thin beds of horizontally laminated and molar-tooth dolomite in middle, and interbedded quartz arenite and thin-bedded dolomite near base. Thickness 180 m at measured section.

Ye Empire Formation (Middle Proterozoic)

- Consists primarily of argillite, siltite, and lesser amounts of arenite and dolomite. Upper part is composed primarily of olive-green and a few purplish-red argillite and siltite beds that range in thickness from a few centimeters to 1.5 m. Thin dolomite beds are present near middle of the Empire is composed largely of white to buff quartz arenite beds that range in thickness form 13 cm to 3.5 m and contain minor carbonate cement and pyrite; most arenite is well sorted, however, within some beds grain size ranges from fine to coarse. Locally, arenite beds have well-developed crossbedding, load structures, and asymmetrical ripple marks. Arenite beds decrease in number and thickness from bottom to top of the

Empire. Lower contact is placed at base of lowermost bed of white quartz arenite that overlies the uppermost sequence of red argillite of the Grinnell Formation. Thickness ranges from 158 m on Scalplock Mountain to 122 m near Grinnell Glacier.

Ygl Grinnell Formation (Middle Proterozoic)

- Mostly quartz arenite on east side of park and interlaminated siltite and argillite on west side. Contact between the Grinnell and underlying Appekunny Formation is placed where red argillite and siltite of the Grinnell change to green argillite of the Appekunny. Thickness ranges from 530 to 790 m.
- In southeastern most part of park, the Grinnell averages 60 percent quartz arenite, and its upper part is nearly 100 percent quartz arenite or quartz conglomerate. Quartz arenite beds in eastern exposures are typically white, medium to coarse grained, lenticular, ripple marked, and prominently crossbedded, and in general, become more common upward in the Grinnell. Basal scours and red argillite chips, pellets, and cobbles are common. Red or purplish-red laminated siltite, silty argillite, and argillite are commonly interbedded; lamentation ranges from even parallel to wavy nonparallel and locally includes ripple cross-lamination; mud cracks and fluid-escape structures commonly disrupt bedding in these red beds. Greenish-gray siltite and argillite are locally present in the upper and lower transition zones with adjacent formations.
- Interbedded quartz arenite and red argillite in the eastern exposures changes northwestward to a lithofacies that contains less quartz arenite and instead is composed mainly of pale, grayish-green and grayish-purple siltite and argillite. In

northwestern part of park, the Grinnell can be subdivided into two parts. Upper part is 425 m thick and is similar to the lower part except that it contains more lenses of ripple, white quartz arenite (locally as much as 20 percent of the section). Lower part is 365 m thick and is predominantly interbedded blocky siltite and evenly laminated argillite that contains a few thin lenses of fipple-marked, white quartz arenite. Bedding is disrupted by abundant shrinkage cracks, fluid-escape structures, and interlayers of mud-ship breccia.

***Yap Appekunny Formation (Middle Proterozoic*)**

- At Apikuni Mountain in northeastern part of park, the Appekunny is informally subdivided into five members designated 1 through 5 in ascending order; members are not mapped separately. On west side of park only parts of members 5, 4 and 3 are present and are mapped along with the disconformably underlying Prichard Formation as unit Yapp. On east side of park, the Appekunny disconformably overlies the Altyn Formation; the contact between the two is placed on the top of the uppermost dolomite bed of the Altyn and locally shows as much as 2 m of erosional relief. Thickness on east side of park ranges form 530 to 690 m. Member thicknesses are from measured section near Apikuni Mountain.
- Member 5 consists of bright-green argillite and lesser amounts of siltite. Lamination is wavy, nonparallel, fining-upward couplets. Mud0chip breccias, fluid-escape structures, and dolomite-filled subaqueous shrinkage cracks common. About 60 m thick.

- Member 4 contrasts sharply with member 5 and is poorly exposed because outcrops are mostly cleaved and easily weathered. Member consists of thin to very thin laminae of olive siltite and thin lenticular beds of rusty-brown arenite. Commonly stained by iron and manganese oxides. Notably cleaned, folded, and faulted near thrust faults. Lower contact placed at base of lowermost sequence of thinly laminated siltite. About 135 m thick.
- Member 3 is characterized by interlaminated and interbedded grayish-green siltite, yellowish-brown arenite, and lesser amounts of grayish-green argillite; subaqueous shrinkage cracks, load structures, and mud-chip breccia common. Lamination is wavy nonparallel; arenite beds typically contain pyrite. Lower contact placed at base of lowermost bed of pyritic arenite, where pyritic arenite and overlying beds are wavy laminated and contain numerous shallow-water sedimentary structures. About 165 m thick.
- Member 2 consists mostly of interlaminated siltite and some argillite. Thin arenite beds, 2.5-7.5 cm thick, common in lower part. Bed lamination is even parallel to nonparallel and curved nonparallel; some beds show broad, low-angle hummocky cross-lamination and small-scale, sour-and-fill structures. Lower contact is placed on top of uppermost maroon sequence of member 1 and generally coincides with an increase in thickness of siltite laminae in member 2. In areas where maroon beds are absent, contact between members 1 and 2 may be indistinguishable. About 165 m thick.
- Member 1 closely resembles member 2 except for the presence of maroon beds and consists of alternating successions of pale-maroon and grayish-green siltite

and minor argillite. Laminae are generally thinner in member 1 than in member 2. A quartz arenite interval forms a key marker about 55 m above base of member 1. This interval thins gradually northward from about 25 m at Elk Mountain (at south end of park) to 15 m at Bear Mountain (near U.S.-Canada boundary). About 135 m thick.

***Yapp Appekunny and Prichard Formations, undivided (Middle Proterozoic*)**

- Present only on west side of park where unit is divisible into three parts, but parts are not differentiated on map. Upper part is the Appekunny formation; middle and lower parts are subdivisions of the Prichard Formation. Base not exposed. Minimum thickness ranges from 1,608 to 2,165 m.
- Upper part of map unit is partial sections of member 5, 4, or 3, or of all three members of the Appekunny that appear to rest disconformably on the Prichard. Upper part ranges in thickness from 63 m in west-central part of park where it consists only of member 5, to about 500 m near the U.S.-Canada boundary where it consists of members 5, 4, and 3.
- Middle part of map unit is the upper part of the Prichard and consists of wavy, nonparallel laminae of greenish-gray to medium-gray calcareous siltite. Locally, middle part contains thin lenticular beds of white quartz arenite and discontinuous beds of fragmental limestone or breccia and stromatolitic limestone. Equivalent to the transition zone of the Prichard. Thickness ranges from 245 to 365 m.
- Lower part of map unit is characterized by thin, even, parallel laminae of rusty-weathering, blackish-gray argillite and light-gray siltite that contain disseminated pyrite and pyrrhotite. Some small-scale cross-lamination is present in siltite

laminae. Carbonate occurs locally near top of lower part as cement in thin siltite laminae and as pods and nodules of black manganiferous limestone. About 1,300 m thick.

***Ya Altyn Formation (Middle Proterozoic*)**

- Occurs only on east side of park; completely exposed at Yellow Mountain and northward in northeastern pat of park. In exposures south of Yellow Mountain, base is not exposed and formation is truncated by Leis thrust fault. In Yellow Mountain area, the Altyn can be informally subdivided into three members designated 1 through 3 in ascending order; members not mapped separately. Locally where map scale permits, the Altyn and an eastern facies are differentiated and mapped separately. At Yellow Mountain, the Altyn ranges in thickness from 238 to 255 m.
- Member 3 is interbedded and interlaminated, light-gray to brownish-yellow dolomite, dolarenite, and arenite. Dolomite beds are 2-20 cm thick; arenite beds are medium to coarse grained and commonly crossbedded, some herringbone lamination. Stromatolites and stylolites common. Thickness 55-62 m.
- Member 2 is massive, medium- to thick-bedded, white to gray dolomite. Some medium- to coarse-grained, poorly sorted arenite beds in upper part. Stromatolites and dark-orange dolomite blebs occur locally in lower part. Contains black asphaltic veinlets near Lewis thrust fault. Thickness 58-68 m.
- Member 1 is yellow- to orange-weathering, dark-gray to black dolomite in think to thick (2 m) beds, and thin, lenticular interbeds of fine-grained arenite. Stromatolites common in lower part. About 125 m thick.

Yae Eastern facies (Middle Proterozoic)

- Similar to main body of Altyn except middle member (member 2) is mostly thick beds of brownish-weathering, coarse-grained quartzite. Low-angle cross-lamination common. Minor interbedded argillaceous gray dolomite. Partially exposed in thrust fault plates in Divide Mountain area.

Yapa Appekunny and Altyn Formations, undivided (Middle Proterozoic)

- Undivided in areas adjacent to southern part of Lewis thrust fault where map scale and complex structure preclude differentiation.

Ywt Waterton Formation (Middle Proterozoic)

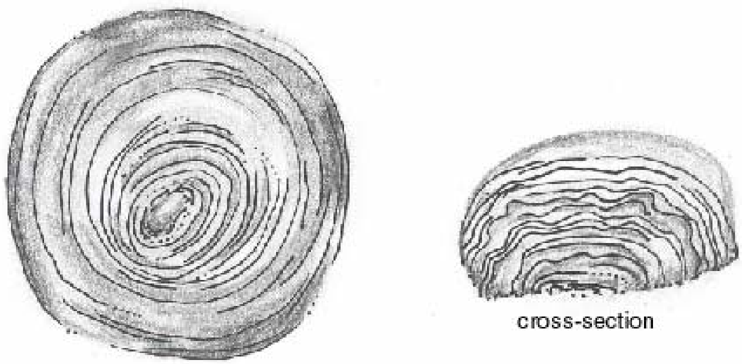
- Present only in northeastern part of park, where base is not exposed and Waterton is truncated by Lewis thrust fault. In Yellow Mountain area, Waterton is informally subdivided into five members designated 1 through 5 in ascending order.; members not mapped separately. At Yellow Mountain, Waterton ranges in thickness from 170 to 229 m; becomes thicker northward as Lewis thrust fault cuts downsection.
- Member 5 consists of silty dolomite, dolomitic siltite, and dolomitic sandstone. Most strata are rusty reddish brown, some are maroon, orange, and green. Bed thickness usually thin to medium but sometimes thick. Crossbedding and stromatolites are common locally. Thickness ranges from 22 to 46 m.
- Member 4 is orangish-yellow and light-brown, fine-grained dolomite. Usually medium bedded but some thin or thick beds. Thickness 18-28 m.
- Member 3 is dark-gray and bluish-gray limestone and minor light-tan dolomite. Bedding is generally thin to medium, thickens downward, and has striped

appearance because of discontinuous 1- to 3-cm-thick mottled dolomite layers within 7- to 10-cm-thick limestone and limestone breccia beds. Breccia composed of rip-up clasts of fine-grained dolomite. Stromatolites, thin cherty layers, pisolites, and soft-sediment deformation are present. Thickness 15-25 m.

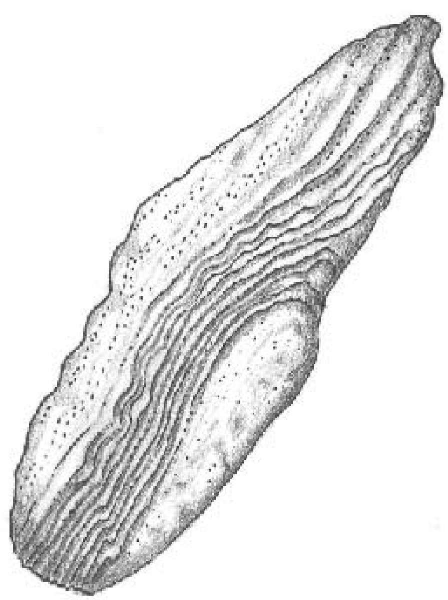
- Member 2 consists of yellowish-gray and light-gray, fine-grained, medium- to thin-bedded dolomite. A 1-m-thick dolomite bed contains chert nodules 5-10 m from top. Thickness 15-25 m.
- Member 1 is mostly dolomite that contains chert nodules and cherty beds. Color is generally tan to gray with a light-yellow tint, but locally some beds are grayish brown and yellowish white. Most beds are 0.3-1 m thick. Bedding is usually weakly defined by discontinuous, black calcareous beds, 2-4 cm thick, that are commonly dolomitic and have concretionary form. Chert nodules, chert “blebs”, and siliceous laminae are common. Chert is black when fresh and weathers rusty orange. Chert nodules are usually less than 5 cm in diameter. Stromatolites as much as 30 cm in diameter are common and sometimes have cherty tops and bottoms. Small-scale crossbedding is present locally. Asphaltic material fills veinlets adjacent to Lewis thrust fault. Member is truncated by Lewis thrust fault but is at least 100 m thick.

Appendix B: Illustrations of some common invertebrate fossils

All drawings are by Trista L. Thornberry (Colorado State University).

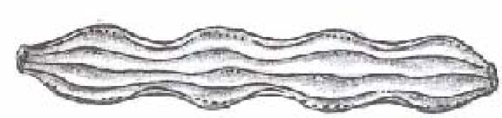


Cryptozoan proliferum
(algal mass - Pre-Cambrian) (8 cm)

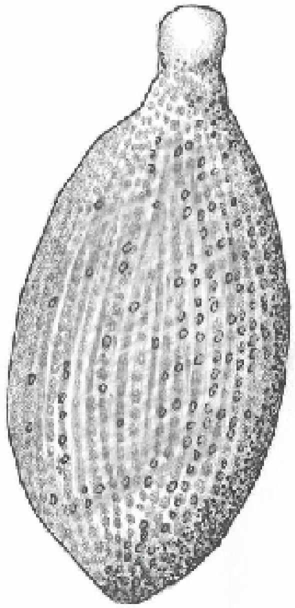


Stromatoporoid
(fossil algae - Mid-Devonian) (x 1)

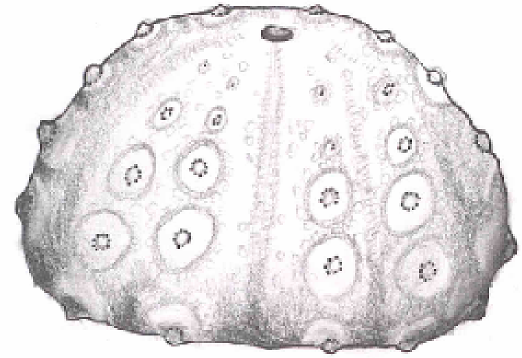
Fusilina "rice rock"
(fusilinid - Pennsylvanian) (x 15)



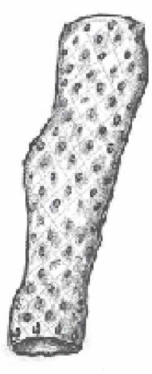
Nodosauria
(foraminifera - Early Mid-Eocene) (x 20)



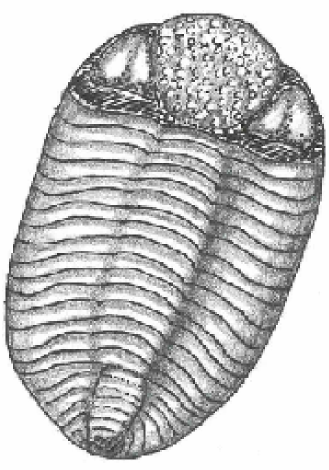
Balanocidaris glandifera
(echinoid - Late Triassic) (32 mm)



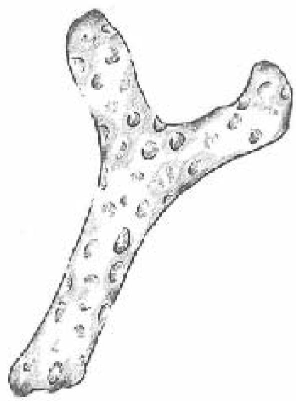
Tylocidaris clavigera
(echinoderm - Turonian) (4 cm)



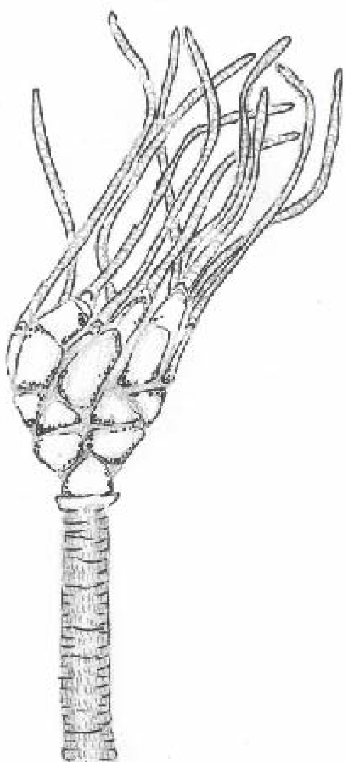
Rhombopora
(bryozoan - Ordovician-Permian) (6 cm)



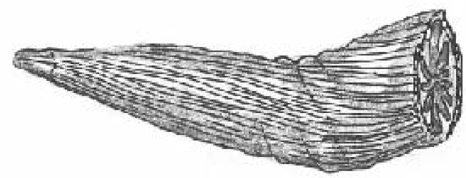
Phacops rana milleri
(trilobite - Mid-Devonian) (4 cm)



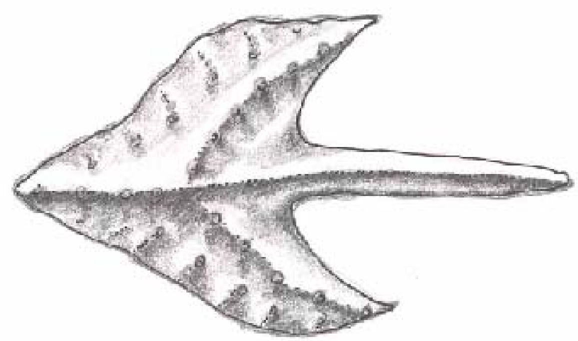
Thamnopora coral
(Mid-Devonian) (x 1)



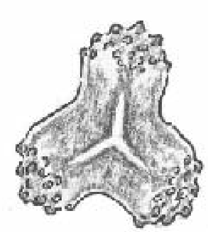
Cyanthocrinites multibranchiatus
(crinoid - Mississippian) (x 1.5)



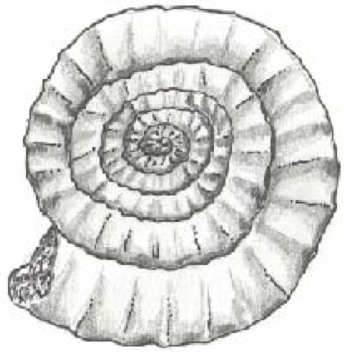
Lophopyllum profundum
(horn coral - Pennsylvanian) (5.5 cm)



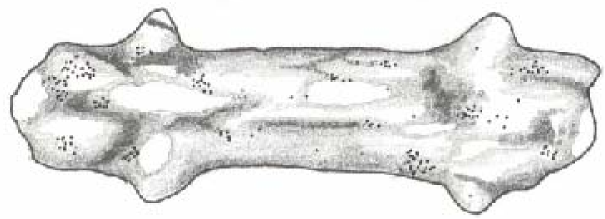
Ancyrodella
(conodont - Late Devonian) (x 55)



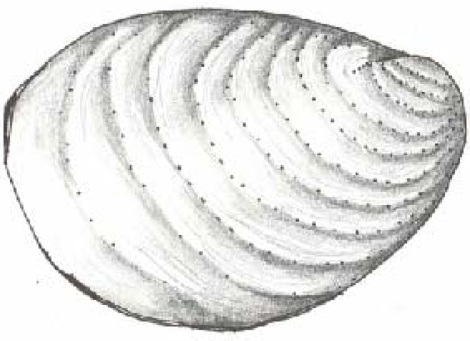
Trilobosporites marylandensis
(fossil spore - Early Cretaceous) (x 1000)



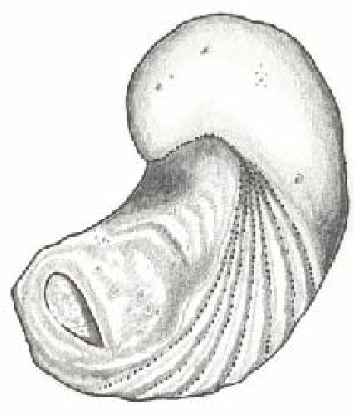
Arnioceras cuneiforme
(ammonite (cephalopod) - Late Jurassic) (x1)



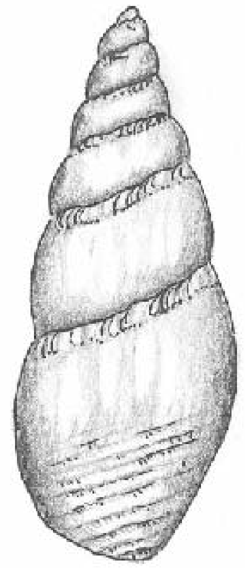
Baculites
(cephalopod - Late Cretaceous) (x1.5)



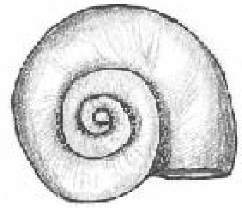
Inoceramus simpsoni
(clam - Late Cretaceous) (x0.5)



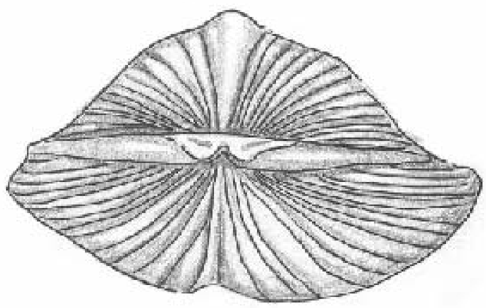
Gryphaea arcuata
(oyster shell - Early Cretaceous) (x1)



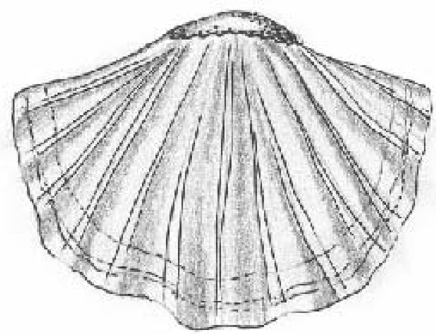
Anomalofusus
(gastropod - Late Cretaceous) (x 2)



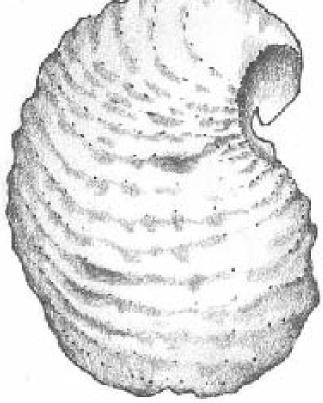
Busycon
(gastropod - Late Miocene) (11 cm)



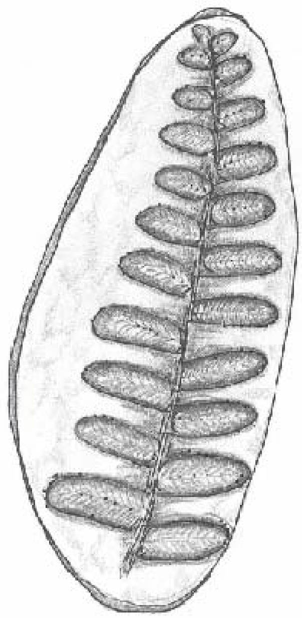
Paraspirifer bownockeri
(bivalve - Mid-Devonian) (x 1)



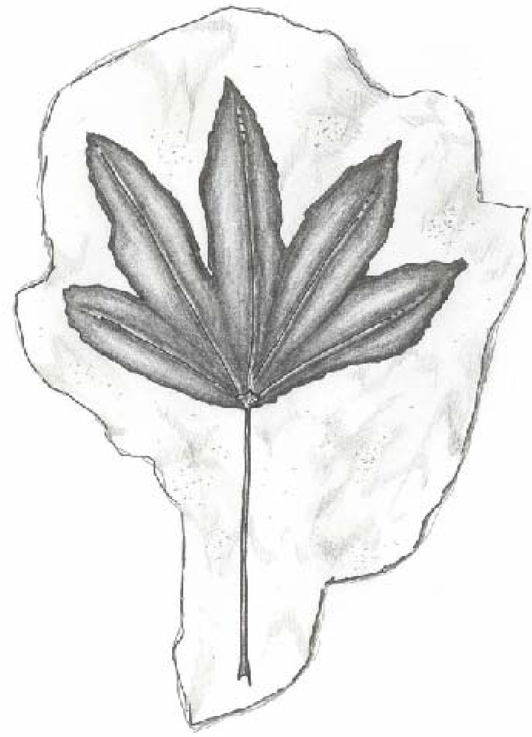
Platystrophia moritura
(bivalve - Late Ordovician) (28 mm)



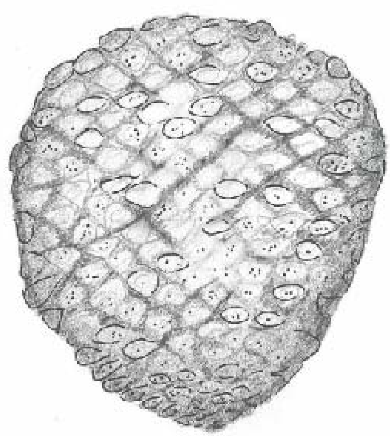
Exogyra cancellata
(pelecypod - Late Cretaceous) (x 0.5)



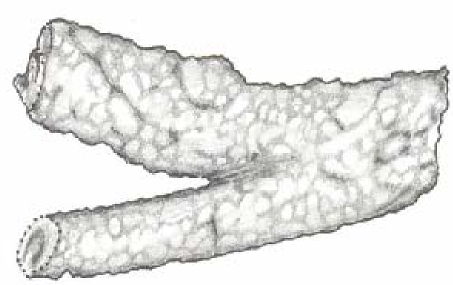
Neuropteris
(fern - Permian) (x1)



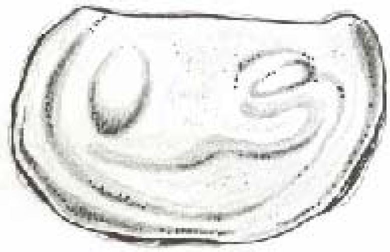
Plantophyllum wyomingenesis
(plane (sycamore) tree - Eocene) (10 cm)



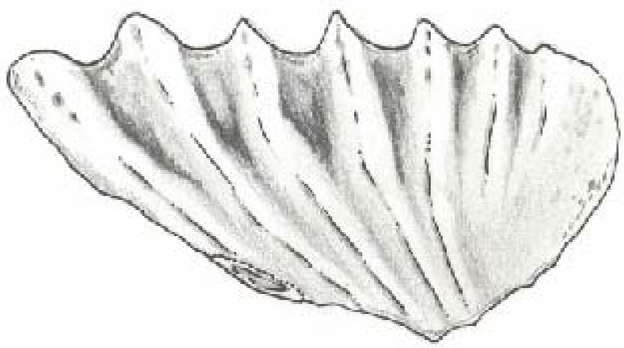
Pine cone
(Triassic) (5.5 cm)



Halymenites major
(burrow cast of *Protocallianassa mortoni* - Late Cretaceous) (x0.5)



Glyptopleurina montifera
(ostracode - Pennsylvanian) (x50)



Sagenodus
(dental plate of a lungfish - Pennsylvanian) (x1.5)

Glossary

For a more complete glossary, visit:

http://www.nature.nps.gov/grd/usgsnps/misc/glossaryAtoC.html

abrasion
: Grinding/rasping erosion by rocks entrained in glacial ice or windblown sand.

Acadian Orogeny
: A widespread deformational, metamorphic, and plutonic event that extended from the late Devonian to the Mississippian and greatly affected the northern Appalachian region.

accretionary prism
: Sediment wedge accumulating at the edge of a continent at a subduction zone.

active margin
: A continental margin where significant volcanic and earthquake activity occurs; commonly a convergent plate margin.

albite
: A light-colored, sodium-rich, end-member silicate mineral of the plagioclase feldspar group [90-100% $NaAlSi_2O_8$] (also see anorthite).

alkali feldspar
: A group of feldspars composed of mineral mixtures or solid solutions with the alkali metals potassium and sodium.

allocthon
: A mass of faulted rocks that have been tectonically transported from their original source or position; also a mass of redeposited sediment (cf., autochthon).

alluvial fan
: A fan-shaped deposit of sediment that accumulates where a high gradient stream flows out of a mountain front into an area of lesser gradient such as a valley.

alluvium
: Stream-deposited sediment that is generally rounded, sorted, and stratified.

Allegheny Orogeny
: A mostly Late Paleozoic deformational event centered in the Valley and Ridge province and Allegheny Plateau of the central and southern Appalachians.

alpine glacier
: A glacier occurring in a mountainous region; also called a valley glacier.

angular unconformity
: An unconformity where the strata above and below are oriented differently; generally caused by structural deformation and erosion prior to deposition of the upper bed.

anorthite
: The calcium-rich, solid solution silicate end member of the plagioclase feldspar group [90-100% $CaAl_2Si_2O_8$] (also see albite).

Antler Orogeny

An extensive deformation event in the western Cordillera of North America during the late Devonian and early Mississippian and characterized by the Roberts Mountain thrust in the Great Basin area of Nevada—broadly contemporaneous with the Acadian Orogeny.

Appalachian Orogeny

A general term applied to the broad Paleozoic deformation, metamorphism, and plutonism throughout the Appalachian region of North America and encompassing the Taconic, Acadian, and Allegheny orogenies.

aquifer

Rock or sediment that are sufficiently porous, permeable, and saturated to be useful as a source of water.

aragonite

A light-colored, often clear or iridescent, polymorph of calcite [CaCO3] that forms the nacreous coating of many shellfish and pearls.

arenaceous

Rocks derived from sand or that contain sand. Also applied to agglutinated Foraminifera, especially those forms that use silt grains for construction of their tests.

argillaceous

A term used to refer to rocks with significant clay or silt content.

arkose

A feldspar-rich (>25%) sandstone or conglomerate.

arkosic

Sedimentary rock that is feldspar-rich (>25%).

ash (volcanic)

Fine pyroclastic material ejected from a volcano (also see tuff).

asthenosphere

Weak layer in the upper mantle below the lithosphere where seismic waves are attenuated.

autochthon

A mass of rock that has remained in its original location, especially during tectonic deformation.

axial plane

A planar axial surface dividing the two limbs of a fold.

axial surface

The surface that passes through successive hingelines in a stack of folded surfaces and divides the two limbs of a fold.

axis (fold)

A straight line approximation that when moved parallel to itself generates the shape of a fold (see and use hingeline)

barchan dune

A crescent-shaped dune with arms or horns of the crescent pointing downwind.. The crescent or barchan type is most characteristic of the inland desert regions.

barrier island

A long, low, narrow island formed by a ridge of sand that parallels the coast.

basalt

A dark-colored, mafic, volcanic rock consisting predominantly of plagioclase, pyroxene, and olivine (also see gabbro)

baseflow

Stream flow supported by groundwater flow from adjacent rock, sediment, or soil.

baselevel

The lowest level to which a stream can erode its channel. The *ultimate* base level for the land surface is sea level, but *temporary* base levels may exist locally.

basement

The undifferentiated rocks, commonly igneous and metamorphic, that underlie the rocks of interest.

basin (structural)

A doubly-plunging syncline in which rocks dip inward from all sides (also see dome).

basin (sedimentary)

Any depression, from continental to local scales, into which sediments are deposited.

batholith

A massive, discordant pluton, greater than 100 square km, often formed from multiple intrusions.

beach

A gently sloping shoreline covered with sediment, often formed by action of waves and tides.

beach face

The section of the beach exposed to direct wave and/or tidal action.

bed

The smallest sedimentary strata unit, commonly ranging in thickness from one centimeter to a meter or two and distinguishable from beds above.

bedding

Depositional layering or stratification of sediments.

bedrock geology

The geology of underlying solid rock as it would appear with the sediment, soil, and vegetative cover stripped away.

bentonite

A sedimentary rock formed from the alteration in place of volcanic ash. Largely composed of the clay mineral montmorillonite.

block (fault)

A crustal unit bounded by faults, either completely or in part.

braided stream

A stream, clogged with sediment that forms multiple channels that divide and rejoin.

breccia

A coarse-grained, generally unsorted, sedimentary rock consisting of cemented angular clasts.

calcareous

A rock or sediment containing calcium carbonate (Cf. carbonaceous).

calcite

A clear to light-colored, calcium carbonate mineral [$CaCO3$] (also see aragonite).

caldera

A large bowl- or cone-shaped summit depression in a volcano formed by explosion or collapse

carbonaceous

A rock or sediment with considerable carbon, esp. organics, hydrocarbons, or coal.

carbonate

A mineral containing the carbon-oxygen radical $(CO3)^{-2}$.

cementation

Chemical precipitation of material into pores between grains that bind the grains into rock.

chalcedony

A cryptocrystalline silica polymorph of quartz with generally radiating fibrous crystal structure; the primary constituent of chert.

chemical sediment

A sediment precipitated directly from solution (also called nonclastic).

chemical weathering

The dissolution or chemical breakdown of minerals at Earth's surface via reaction with water, air, or dissolved substances.

chert

A white-to-gray cryptocrystalline silica rock consisting chiefly of fibrous chalcedony.

clastic

Rock or sediment made of fragments or pre-existing rocks.

clay

Clay minerals or sedimentary fragments the size of clay minerals (<2 cm).

cleavage (mineral)

The tendency of a mineral to break preferentially in certain directions along planes of weaknesses in the crystal structure.

cleavage (rock)

The tendency of rock to break along parallel planes that correspond to the alignment of platy minerals.

compressive stress

A normal stress (pressure) tending to push or squeeze material together.

concordant

Strata with contacts parallel to the attitude of adjacent strata.

condensed section

A facies consisting of thin marine beds that accumulate at a very slow rate.

conglomerate

A coarse-grained sedimentary rock with clasts larger than 2 mm in a fine-grained matrix.

continental crust

The type of crustal rocks underlying the continents and continental shelves; having a thickness of 25-60 km and a density of approximately 2.7 grams per cubic centimeter.

continental drift

The concept that continents have shifted in position over Earth (see and use 'plate tectonics').

continental rise

Gently sloping region from the foot of the continental slope to the abyssal plain.

continental shelf

The shallowly-submerged portion of a continental margin extending from the shoreline to the continental slope with water depths of less than 200 m.

continental shield

A continental block of Earth's crust that has remained relatively stable over a long period of time and has undergone only gentle warping compared to the intense deformation of bordering crust

continental slope

The relative steep slope from the outer edge of the continental shelf down to the more gently sloping ocean depths of the continental rise or abyssal plain.

convergent boundary

A plate boundary where two tectonic plates are moving together (i.e., a zone of subduction or obduction).

Cordillera

A Spanish term for an extensive mountain range that is used in North America to refer to all of the western mountain ranges of the continent.

country rock

Rock into which igneous rock is intruded or emplaced.

craton

The relatively old and geologically stable interior of a continent (also see continental shield).

cross-bedding

Uniform to highly-varied sets of inclined sedimentary beds deposited by wind or water that indicate distinctive flow conditions.

cross section

A graphical interpretation of geology, structure, and/or stratigraphy in the third (vertical) dimension based on mapped and measured geological extents and attitudes depicted in an oriented vertical plane.

crust

The outermost compositional shell of Earth, 10-40 km thick, consisting predominantly of relatively low-density silicate minerals (also see oceanic crust and continental crust).

crystalline

Describes the structure of a regular, orderly, repeating geometric arrangement of atoms

cuesta

An asymmetric landform with one gently-sloping side and one steeply-sloping side due to erosion of gently-dipping rock strata.

debris flow

A rapid and often sudden flow or slide of rock and soil material involving a wide range of types and sizes.

deformation

A general term for the process of faulting, folding, shearing, extension, or compression of rocks as a result of various Earth forces.

delta

A sediment wedge deposited at a stream's mouth where it flows into a lake or sea.

desert

A region with low precipitation (usually less than 25 cm/yr) that supports little vegetation.

diagenesis

Set of all processes affecting sediments after burial between the water table and the depth of incipient low-grade metamorphism.

diapir

A dome or anticlinal fold, the overlying rocks of which have been ruptured by the squeezing out of the plastic core material. In sedimentary strata, diapers usually contain cores of salt or shale.

dike

A tabular, discordant igneous intrusion.

dip

The angle between a structural surface and a horizontal reference plane measured normal to their line of intersection.

disconformity

An unconformity at which the bedding of the strata above and below are parallel.

discordant

Having contacts that cut across or are set an angle to the orientation of adjacent rocks.

divergent boundary

A tectonic plate boundary where the plates are moving apart (e.g., a spreading ridge or continental rift zone).

dolomite

A carbonate mineral [$CaMg(CO_3)_2$] or the chemical sedimentary rock made predominantly of that mineral.

dome

A doubly plunging anticline that dips radially in all directions.

downlap

The situation where an initially inclined layer terminates downdip against an initially horizontal or inclined surface.

drainage basin

The total area from which a stream system receives or drains precipitation runoff.

dune

A low mound or ridge of sediment, usually sand, deposited by wind. Common dune types include the *barchan dune*, *longitudinal dune*, *parabolic dune*, and *transverse dune* (see respective listings).

eolian

Formed, eroded, or deposited by or related to the action of the wind.

ephemeral stream

A stream that flows only in direct response to precipitation.

epicontinental sea

Shallow portions of a sea that are located upon a continental shelf or platform, extending into the interior of a continent.

estuary

The seaward end or tidal mouth of a river where fresh and sea water mix; many estuaries are drowned river valleys caused by sea level rise (transgression) or coastal subsidence.

eustatic

Relates to simultaneous worldwide rise or fall of sea level in Earth's oceans.

evaporite

Chemically precipitated mineral(s) formed by the evaporation of solute-rich water under restricted conditions.

exfoliation

The breakup, spalling, peeling, flaking, etc. of layers or concentric sheets from an exposed rock mass due to differential stresses resulting from thermal changes or pressure unloading.

extrusive

Of or pertaining to the eruption of igneous material onto the surface of Earth (cf: intrusive).

facies (metamorphic)

The pressure-temperature regime that results in a particular, distinctive metamorphic mineralogy (i.e., a suite of index minerals).

facies (sedimentary)

The depositional or environmental conditions reflected in the sedimentary structures, textures, mineralogy, fossils, etc. of a sedimentary rock.

fanglomerate

A sedimentary rock consisting of slightly rounded but generally unsorted lithic fragments deposited in an alluvial fan.

fault

A subplanar break in rock along which relative movement occurs between the two sides.

fault trace

The exposed intersection of a fault with Earth's surface.

fault zone

A fault that is expressed as a zone of numerous small fractures or of fault breccia or gouge.

feldspars

A group of framework silicates, containing aluminum and calcium, sodium, or potassium; feldspars are the most abundant minerals in Earth's crust.

fissile (fissility)

A property of splitting easily along closely spaced parallel planes.

footwall

The mass of rock beneath an inclined fault, orebody, or mine working (cf: hanging wall).

foreland
In a structural sense, the foreland is the region in front of a series of overthrust sheets developing during an orogeny.

foreland basin
A basin or trough that forms in front of the advancing thrust sheets.

formation
Fundamental rock-stratigraphic unit that is mappable and lithologically distinct from adjoining strata and has definable upper and lower contacts.

fracture
Irregular breakage of a mineral (cf: cleavage); also any break in a rock (e.g., crack, joint, fault)

frost wedging
The breakup of rock due to the expansion of water freezing in fractures.

gabbro
A mafic plutonic rock rich in ferromagnesian minerals and plagioclase feldspar (see also basalt).

geology
The study of Earth including its origin, history, physical processes, components, and morphology.

Gondwana
The protocontinent in the southern hemisphere, corresponding to Laurasia in the northern hemisphere that formed the supercontinent Pangea.

graben
A down-dropped structural block bounded by steeply-dipping, normal faults (also see horst).

graded bedding
Vertical upward progression of grain sizes in a sediment layer from coarse to fine.

granite
A plutonic rock rich in quartz and alkali feldspars with minor mica (intrusive equivalent of rhyolite).

Grenville Orogeny
A widely recognized igneous, metamorphic, and deformational event that occurred from about 1000-880 million years ago along the southeastern margin of the Canadian shield (also see Laurentian).

groundmass
The finer-grained matrix of sedimentary or porphyritic igneous rocks.

groundwater (ground water)
Subsurface water in the saturated zone.

Group
A major rock-stratigraphic unit of two or more formations that have significant similar lithologic characteristics.

gypsum
A generally earthy white to translucent crystalline hydrous, calcium sulfate mineral [$CaSO_4 \bullet 2H_2O$].

hanging wall
The overlying side of an orebody, inclined fault, or mine working (cf: footwall).

headward erosion
The erosion of a stream channel upstream toward and into its head.

hiatus
A break or interruption in the continuity of the geologic record such as a lapse in the time interval represented by an unconformity.

hoodoo
Pillars developed by erosion of horizontal strata of varying hardness. Typically found in climatic zones where most rainfall is concentrated during a short period of the year.

hornblende
A dark, generally black, silicate mineral with distinctive 56^{o} and 124^{o} cleavage; the most common mineral of the amphibole group.

horse (fault)
A faulted rock mass caught between and surrounded by adjacent fault branches or splays.

horst
An uplifted structural block bounded by high-angle normal faults.

hydrogeology
The scientific study of subsurface water and its related geology.

hydrology
The science that deals with global water, its properties, circulation, and distribution, on and under Earth's surface and in the atmosphere.

hypsography
The geographic study, observation, measurement, and mapping of Earth surface elevations and topography

hypothesis
A conceptual model or explanation that is proposed for scientific testing.

igneous
Refers to a rock or mineral that originated from molten material; one of the three main classes or rocks: igneous, metamorphic, and sedimentary.

infiltration
The process by which water percolates into the ground; the movement of water or fluids from the source through soil, sediment, or rocks via interstices and/or fractures.

intrusion
A body of igneous rock that invades older rock. The invading rock may be a plastic solid or magma that pushes its way into the older rock.

intrusive
Of or pertaining to intrusion, both the process and rock so formed (cf: extrusive).

island arc
A line or arc of volcanic islands formed over and parallel to a subduction zone.

isostasy

The process by which the crust "floats" at an elevation compatible with the density and thickness of the crustal rocks relative to underlying mantle.

jasper

A generally red, cryptocrystalline, opaque to slightly translucent, variety of chert.

joint

A semi-planar break in rock without relative movement of rocks on either side of the fracture surface.

karst topography

Topography characterized by abundant sinkholes and caverns formed by the dissolution of calcareous rocks.

laccolith

A tack head- to arcuate-shaped, concordant pluton that domed or up-arched the overlying country rocks.

lacustrine

Pertaining to, produced by, or inhabiting a lake or lakes.

lamination

The finest stratification or bedding as seen in shales and siltstones (syn: lamina or laminae) or the formation of lamina.

landslide

Any process or landform resulting from rapid mass movement under relatively dry conditions (cf.: debris flow).

Laramide Orogeny

A time of basement-involved, regional deformation in the eastern Rocky Mountains (esp. Colorado and Wyoming) from the Late Cretaceous to the mid-Cenozoic.

lateritic soil

Extensively leached soil, characteristic of tropical climates.

latitude

The angular distance of a point on Earth's surface measured from and perpendicular to the equator.

Laurasia

The protocontinent in the northern hemisphere, corresponding to Gondwana in the southern hemisphere that formed the supercontinent Pangea

Laurentian

A nondistinct term used for granites and orogenies of Precambrian age in the Canadian shield, originally defined for 1 b.y. old granites in the Laurentian Highlands (also see Grenville Orogeny).

lava

Magma that has been extruded out onto Earth's surface, both molten and solidified.

levees

Raised ridges lining the banks of a stream; may be natural or artificial.

limbs

The two sides of a structural fold on either side of its hingeline.

limestone

A carbonate-rich (predominately calcite) sedimentary rock.

lineament

Any relatively straight surface feature that can be identified via observation, mapping, or remote sensing, often representing tectonic features.

lineation (structural geology)
Any relative straight structure in a rock such as flow lines, slickensides, etc.

listric
The flattening at depth of a high-angle normal fault plane.

lithification
The conversion of sediment into solid rock.

lithology
The description of a rock or rock unit, especially the texture, composition, and structure of sedimentary rocks.

lithosphere
The relatively rigid outmost shell of Earth's structure, 50 to 100 km thick, that encompasses the crust and uppermost mantle.

loess
Silt-sized sediment deposited by wind, generally of glacial origin.

longitude
The angular measure of a meridian on a sphere taken from a reference meridian (i.e., the prime meridian) in a plane of the equator.

longitudinal dunes
Dunes elongated parallel to the direction of wind flow.

longshore current
A current parallel to a coastline caused by waves approaching the shore at an oblique angle.

mafic
A rock, magma, or mineral rich in magnesium and iron.

magma
Molten rock generated within Earth that is the parent of igneous rocks.

mantle
The zone of Earth's interior between crust and core.

matrix
The fine-grained interstitial material between coarse grains in porphyritic igneous rocks and poorly sorted clastic sediments or rocks.

meanders
Sinuous lateral curves or bends in a stream's channel.

mechanical weathering
The physical breakup of rocks without change in composition (syn: physical weathering).

member
A lithostratigraphic unit with definable contacts that subdivides a formation.

mesa
A broad, flat-topped erosional hill or mountain that is bounded by steeply-sloping sides or cliffs.

metamorphic
Pertaining to the process of metamorphism or to its results.

metamorphism

Literally, “change in form”. Metamorphism occurs in rocks with mineral alteration, genesis, and/or recrystallization from increased heat and pressure.

micas

Group of sheet silicates characterized by excellent parallel cleavage between the sheets of silica tetrahedral.

micrite

Carbonate mud or a rock made thereof (also see: sparite).

mid-ocean ridge

The continuous, generally submarine, seismic, median mountain range that marks the divergent tectonic margin)s) in the world’s oceans.

mineral

A naturally-occurring, inorganic crystalline solid with a definite chemical composition or compositional range.

monocline

A one-limbed flexure in strata, which are usually flat-lying except in the flexure itself.

montmorillinite

A group and individual clay mineral name for variable hydrating, dioctahedral, swelling clay minerals of the general formula $R_{0.33}Al_2Si_4O_{10}(OH)_2•nH_2O$ where R = Na, K, Mg, Ca, etc.

mud cracks

Cracks formed in clay, silt, or mud by shrinkage during subaerial dehydration.

mudstone

A very fine-grained clastic sedimentary rock (>50% clay) similar to shale but without parallel splitting.

muscovite

A potassium-bearing phyllosilicate of the mica group [$KAl_2(AlSi_3)O_{10}(OH)_2$].

neck (volcanic)

An eroded, semi-vertical, pipe-like discordant pluton that represents the vent of a volcano.

Nevadan Orogeny

A mountain building event with tectonism, metamorphism, and plutonism in the western Cordillera of North America during the Jurassic and Early Cretaceous.

nonconformity

An erosional surface preserved in strata in which crystalline igneous or metamorphic rocks underlie sedimentary rocks.

normal fault

A dip-slip fault in which the hanging wall moves down relative to the footwall.

obduction

The process by which the crust is thickened by thrust faulting at a convergent margin.

oceanic crust

Earth’s crust formed at spreading ridges that underlies the ocean basins. Oceanic crust is 6-7 km thick and generally of basaltic composition.

oligoclase

A sodic silicate mineral of the plagioclase group with a relative composition ranging from 70-90% albite.

orogeny

A mountain-building event, particularly a well-recognized event in the geological past (e.g. the Laramide orogeny).

orthoclase

A common potassic silicate mineral of the alkali feldspar group ($KalSi_2O_8$).

orthoquartzite

A clastic sedimentary rock of almost pure quartz sand that is almost completely cemented with silica to form a "sedimentary quartzite."

outcrop

Any part of a rock mass or formation that is exposed or "crops out" at Earth's surface.

outwash

Glacial sediment transported and deposited by meltwater streams.

overbank deposits

Alluvium deposited outside a stream channel during flooding.

overburden

Non-economic, often unconsolidated, rock and sediment overlying an ore, fuel, or sedimentary deposit.

overthrust

A nondescript and not recommended term for a large-scale, low-angle, thrust fault.

paleogeography

The study, description, and reconstruction of the physical geography from past geologic periods.

paleontology

The study of the life and chronology of Earth's geologic past based on the phylogeny of fossil organisms.

Pangea (Pangaea)

A theoretical, single supercontinent that existed during the Permian and Triassic Periods (also see Laurasia and Gondwana).

parabolic dunes

Crescent-shaped dunes with horns or arms that point upwind.

parent (rock)

The original rock from which a metamorphic rock or soil was formed.

passive margin

A tectonically quiet continental margin indicated by little volcanic or seismic activity.

pebble

Generally, small, rounded, rock particles from 4 to 64 mm in diameter.

pediment

A gently sloping, erosional bedrock surface at the foot of mountains or plateau escarpments.

pelagic sediments

Fine-grained, deep-sea sediments consisting of clays and organic ooze.

peneplain

A geomorphic term for a broad area of low topographic relief resulting from long-term, extensive erosion.

permeability

A measure of the ease or rate that fluids move through rocks or sediments.

plagioclase

A group of sodic to calcic feldspars, ranging from albite (Na) to anorthite (Ca) end members, that is among the most common rock-forming minerals.

plateau

A broad, flat-topped topographic high of great extent and elevation above the surrounding plains, canyons, or valleys (both land and marine landforms).

plate tectonics

The theory that the lithosphere is broken up into a series of rigid plates that move over Earth's surface above a more fluid aesthenosphere.

pluton

A body of intrusive igneous rock.

plutonic

Describes igneous rock intruded and crystallized at some depth in Earth.

pluvial lakes

Lakes formed during earlier times of more abundant precipitation.

point bar

A sand and gravel ridge deposited in a stream channel on the inside of a meander where flow velocity slows.

porosity

The proportion of void space (cracks, interstices) in a volume of a rock or sediment.

potassium feldspar

An alkali feldspar rich in potassium (e.g., orthoclase, microcline, sanidine, adularia).

Principal of Original Horizontality

The tenet that sediments are originally deposited in horizontal layers and that deviations from the horizontal indicate post-depositional deformation.

Principle of Superposition

The concept that sediments are deposited in layers, one atop another, i.e., the rocks on the bottom are oldest with the overlying rocks progressively younger toward the top.

prodelta

The part of a delta below the level of wave erosion.

progradation

The seaward building of land area due to sedimentary deposition.

provenance

A place of origin. The area from which the constituent materials of a sedimentary rock were derived.

pyroxene

A group of single-chain structure silicates that is mostly ferromagnesian in composition (e.g., diopside, augite, hedenbergite, enstatite, hypersthene, acmite/aegerine, jadeite, spodumene).

quartz

The simplest framework silicate (SiO_2).

quartzite

Metamorphosed (metaquartzite) or completely silica-cemented (orthoquartzite), quartz-rich sandstone.

radioactivity

The spontaneous decay or breakdown of unstable atomic nuclei.

radiometric age

An age in years determined from radioisotopes and their decay products.

ravinement surface

An erosion surface produced during marine transgression of a formerly subaerial environment.

recharge

Infiltration processes that replenish groundwater.

recurrence interval

The computed length of time between floods of a certain stage on a given stream.

red beds

Sedimentary strata composed largely of sandstone, siltstone, and shale that are predominantly red due to the presence of ferric oxide (hematite) coating individual grains.

regression

A long-term seaward retreat of the shoreline or relative fall of sea level.

relative dating

Determining the chronological placement of rocks, events, fossils, etc. from geological evidence.

reservoir rock

Adequately porous and permeable rock in which petroleum deposits are found.

reverse fault

A contractional, high angle (>45^o), dip-slip fault in which the hanging wall moves up relative to the footwall (also see thrust fault).

rift valley

A depression formed by grabens along the crest of an oceanic spreading ridge or in a continental rift zone.

ripple marks

The undulating, subparallel, usually small-scale, ridge pattern formed on sediment by the flow of wind or water.

rouche moutonnee

An elongate, eroded ridge or knob of bedrock carved by a glacier parallel to the direction of motion with gentle upstream and steep downstream surfaces.

rock

A solid, cohesive aggregate of one or more minerals or mineraloids.

roundness

The relative amount of curvature of the "corners" of a sediment grain, especially with respect to the maximum radius of curvature of the particle.

sabkha

A coastal environment in an arid climate where evaporation rates are high.

sandstone

Clastic sedimentary rock of predominantly sand-sized grains.

sapping

The undercutting of a cliff by erosion of softer and stratigraphically lower rock layers.

scarp

A steep cliff or topographic step resulting from vertical displacement on a fault or by mass movement.

seafloor spreading

The process in which tectonic plates diverge and new lithosphere is created at oceanic ridges.

secondary porosity

Porosity formed in a rock after deposition due to dissolution or fracturing.

sediment

An eroded and deposited, unconsolidated accumulation of lithic and mineral fragments.

sedimentary rock

A consolidated and lithified rock consisting of detrital and/or chemical sediment(s).

sequence

A major informal rock-stratigraphic unit that is traceable over large areas and defined by a major sea level transgression-regression sediment package.

Sevier Orogeny

The Jurassic to Cretaceous deformation event in the central North American cordillera characterized by thin-skinned thrusting in Arizona, Utah, Idaho, Wyoming, Montana, and Canada.

shale

A clastic sedimentary rock made of clay-sized particles that exhibit parallel splitting properties.

shoreline

The line along which the land and water surfaces meet at a lake or sea.

sierra

An often used Spanish term for a rugged mountain range.

sill

A tabular, igneous intrusion that is concordant with the country rock.

silt

Clastic sedimentary material intermediate in size between fine-grained sand and coarse clay (1/256-1/16 mm).

siltstone

A variable-lithified sedimentary rock with silt-sized grains.

slickenside

A smoothly polished and often striated surface representing deformation of a fault plane.

slope

The inclined surface of any geomorphic feature or rational measurement thereof (syn: gradient).

slump

A generally large, coherent mass movement with a concave-up failure surface and subsequent backward rotation relative to the slope.

sodium feldspar

A sodium-rich, alkali feldspar [$NaAlSi_3O_8$], specifically albite.

soil

Surface accumulation of weathered rock and organic matter capable of supporting plant growth and often overlying the parent rock from which it formed.

sparite

Clear, crystalline forms of carbonate minerals in sediment or cements (also see: micrite).

spring

A site where water flows out at the surface due to the water table intersecting the ground surface.

stock

An igneous intrusion exposed less than 40 square miles at the surface.

strata

Tabular or sheetlike masses or distinct layers (e.g., of rock).

stratigraphy

The geologic study of the origin, occurrence, distribution, classification, correlation, age, etc. of rock layers, especially sedimentary rocks.

stream

Any body of water moving under gravity flow and confined within a channel.

stream piracy

The process by which active headward stream erosion breaches a drainage divide and intercepts part of an adjacent drainage basin.

strike

The compass direction of the line of intersection that an inclined surface makes with a horizontal plane.

strike-slip fault

A fault with measurable offset where the relative movement is parallel to the strike of the fault.

structural geology

The static and dynamic study of the geometry, fabric, relationships, and deformational processes of rocks in the upper layers of Earth's crust (also see tectonics).

subduction zone

A convergent plate boundary where oceanic lithosphere descends beneath a continental or oceanic plate and is carried down into the mantle.

subsidence

The gradual sinking or depression of part of Earth's surface.

suture

The linear zone where two continental landmasses become joined due to obduction.

system (stratigraphy)

The group of rocks formed during a period of geologic time.

Taconic Orogeny

A mostly Ordovician deformation and intrusive event in the northeastern North American continent (an early part of the Appalachian Orogeny).

tectonic

Relating to large-scale movement and deformation of Earth's crust.

tectonics

The geological study of the broad structural architecture and deformational processes of the lithosphere and aesthenosphere (also see structural geology).

terraces (stream)

Step-like benches surrounding the present floodplain of a stream due to dissection of previous flood plain(s), stream bed(s), and/or valley floor(s).

terrane

A region or group of rocks with similar geology, age, or structural style.

terrestrial

Relating to Earth or Earth's dry land.

terrigenous sediment

Clastic sediment eroded from a continent's land surface.

texture

The geometric relationships, size, shape, and orientation of grains or crystals within a rock, sediment, or soil.

theory

A hypothesis that has been rigorously tested against further observations or experiments to become a generally-accepted tenet of science.

thrust fault

A contractional, dip-slip fault with a shallowly dipping fault surface ($<45^o$) where the hanging wall moves up and over relative to the footwall.

tongue (stratigraphy)

A member of a formation that extends and wedges out away from the main body of a formation (also see formation, member, and bed).

topography

The general morphology of Earth's surface including relief and location of natural and anthropogenic features (also see hypsography).

trace (fault)

The exposed intersection of a fault with Earth's surface.

trace fossils

Sedimentary structures, such as tracks, trails, burrows, etc., that preserve evidence of organisms' life activities, rather than the organisms themselves.

transgression

Landward migration of the sea due to a relative rise in sea level.

transverse dunes

Dunes elongated perpendicular to the prevailing wind direction. The leeward slope stands at or near the angle of repose of sand whereas the windward slope is comparatively gentle.

travertine

A limestone deposit or crust, often banded, formed from precipitation of calcium carbonate from saturated waters, especially near hot springs and in caves (also see tufa).

traps

Sites of localization or concentration of migrating petroleum resources.

trend

The direction or azimuth of elongation or a linear geological feature.

tufa

A chemical sedimentary rock of spongy or porous calcium carbonate formed by evaporation adjacent to springs and streams (also see travertine).

tuff

Generally fine-grained, igneous rock formed of consolidated volcanic ash.

type locality

The geographic location where a stratigraphic unit is well displayed, is formally defined as a typical section, and derives its name.

unconformity

A surface within sedimentary strata that marks a prolonged period of nondeposition or erosion.

uplift

A structurally high area in the crust, produced by movement that raises the rocks.

vent

An opening at the surface of Earth where volcanic materials emerge.

volcanic

Related to volcanoes; describes igneous rock crystallized at or near Earth's surface (e.g., lava).

volcanoclastic

Pertaining to a clastic rock with volcanic material.

water table

The upper surface of the saturated (phreatic) zone.

weathering

The set of physical, chemical, and biological processes by which rock is broken down in place.

well-rounded

A sedimentary grain texture characterized by smooth curvature of all grain surfaces without angular or flat areas.