

Figure 14. Cross section through the area of the gazebo at Cascade Springs showing localities to be examined, and breccia pipes, believed to be former conduits for the springs, extending up from dissolution cavities in the Minnelusa Formation at depth. Fractured Minnekahta shown in figure 16 lies about 100 feet above the gazebo at locality B. Green shale of the Stockade Beaver Member of the Sundance Formation caps the Spearfish west of S.D. highway 71.

The other “pipe”, about 150 feet to the west as noted by Hayes (1999), is characterized by a gypsum bed that is downwarped a few feet, and beds above and below are not affected.

The many veinlets represent a zone where much gypsum has been removed by solution and/or have been intruded into the surrounding rock from the parent bed by processes not fully understood. Broken beds of gypsum several feet thick merge laterally into gypsum-red bed breccia (fig. 15C) and veinlets. The impression is that the original bedded gypsum at this locality has been modified by solution removal, injection into veins, contortion by expansion, and brecciation. It is possible that much of the original mass of gypsum has been removed. Also, anhydrite, which may have been the original form of calcium sulfate, when converted to gypsum, may have expanded considerably to create the force for vein injection and bed crumpling. However, several beds at the northernmost end of the exposure along SD 71 were X-rayed and no anhydrite was found in veins or beds, only gypsum (John Johnson, USGS, pers. comm.).

B. Fractures (“pull apart”) in the Minnekahta Limestone

Walk several hundred feet to the west along the highway and pass through an open gate before the parking lot at Cascade Springs and climb up the slope. Bedding in the Minnekahta wobbles a bit, but the dip averages about 20° to the southwest. Many large fractures (fig. 16) are found in a zone between 70 and 100 feet vertically above the base of the slope in an area about 150 feet long. The fractures are more than 10 feet deep in places, and probably extend the entire 40-foot thickness of the Minnekahta. They are as much as 10 feet wide and have various orientations, including N. 35° E., E.-W., N. 5° E., and N. 70° E., following prominent joint directions. There are three possible origins that might be considered for these structures: (1) subsidence due to solution of gypsum below, (2) gravity sliding on the soft sediments of the Opeche Shale, and (3) a combination of sliding and weakening of material below by solution. An initial impression is that these fractures are caused by tension due to downhill sliding. An interesting comparison with similar fractures in the Moenkopi Formation related to dissolution of salt at depth in the Holbrook basin (Epstein and Johnson, 2003) will be made. There are many small-scale structures, folds and faults, such as those described at mileage 4.5, that have been attributed to gravity sliding, believed to have occurred after erosion had exposed the surface of the Minnekahta (Epstein, 1958; Brobst and Epstein, 1963).

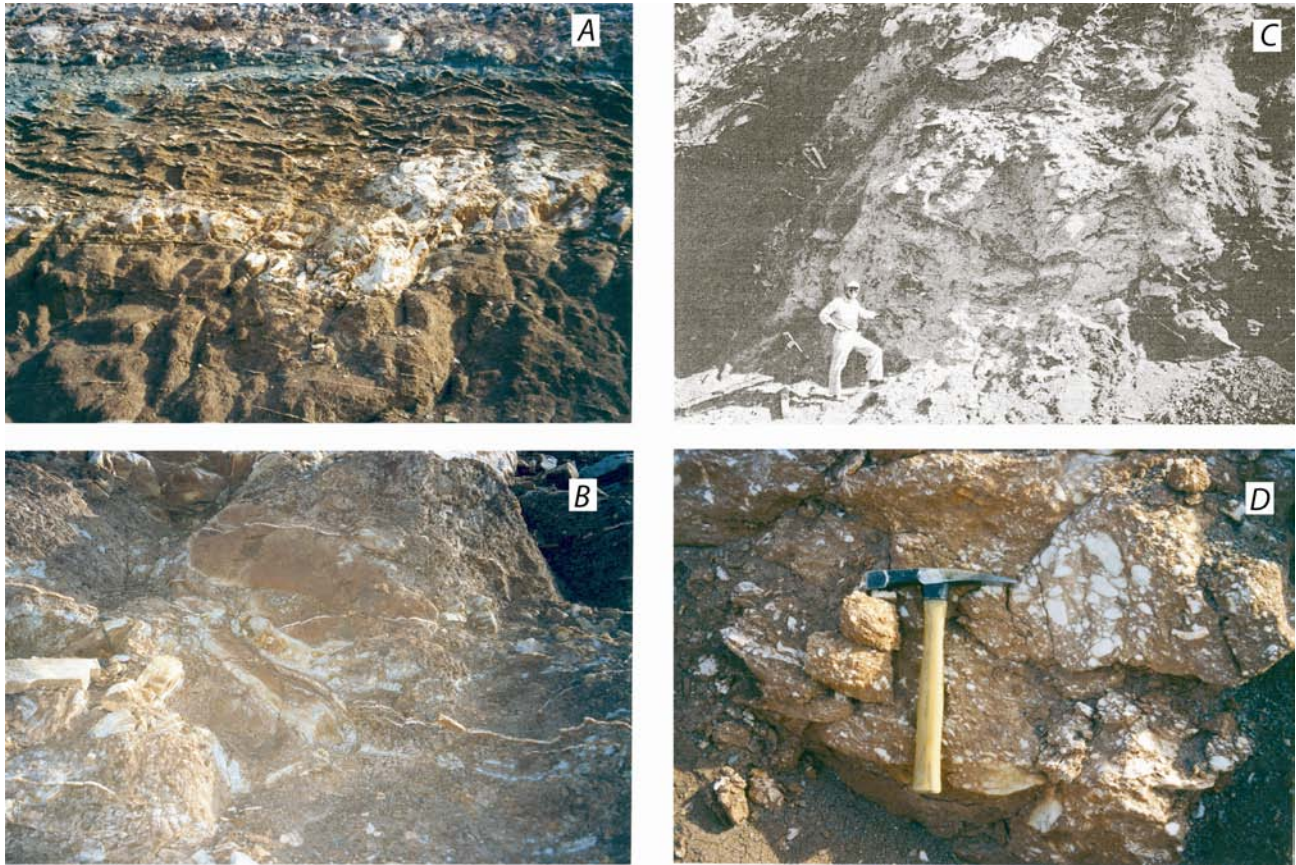


Figure 15. Structures in gypsum in the Spearfish Formation at locality A.

A, Downwarped gypsum bed interpreted to be part of a breccia pipe by Post (1967) and Hayes (1999). Note abundant thin veins extending into surrounding red beds. Shale and siltstone at base of outcrop is minutely fractured.

B, Breccia just below the "pipe".

C, Breccia pipe figured by Post (1967, fig. 87) at same spot as structure in fig. 15A. Note lack of bedding in the bleached pipe as compared to stringers sub-parallel to bedding in the surrounding rock.

D. Brecciated gypsum believed not to be sedimentary because of limited horizontal extent.



Figure 16. Fractures in Minnekahta Limestone, apparently extending down through the entire thickness of the formation, locality B, Cascade Springs, Stop 4.

C₁. Hydrology of Cascade Springs

Cascade Springs is a group of warm ($\sim 20^{\circ}\text{C}$) artesian springs (fig. 13) with collective flow of about $19.6\text{ ft}^3/\text{s}$ and a surface drainage area of 0.47 square miles. The drainage area is not large enough to supply this large flow rate to the spring indicating that its source water is not local. The large springflow rate suggests that a large contribution of flow probably is supplied by the highly permeable Madison aquifer, more than 800 feet below. Hydraulic head in the Madison aquifer is higher than that of the overlying Minnelusa aquifer in the vicinity of the springs, which would result in upward flow from the Madison to the Minnelusa aquifer if adequate vertical permeability exists. Back and others (1983) examined geochemical and hydrologic data of the springs and concluded that Madison aquifer water, which is recharged in the western and southwestern Black Hills, sweeps eastward around the southern end of the Black Hills and supplies most of Cascade Springs, but they are also partially supplied from the Minnelusa aquifer. Stable isotopes of oxygen confirm the interpretation that the source of spring water is from recharge in the Black Hills many miles to the northwest of Cascade Springs (Naus and others, 2001). A spring-water sample had a $\delta^{18}\text{O}$ value of -15.4 parts per thousand, which is considerably lighter than the local recharge but is similar to water recharged on outcrops of Paleozoic rocks near the Pennington-Custer County line. Naus and others (2001) concluded that the low tritium concentrations in spring water indicate that a large proportion of greater than 40 year old ground water discharges from the springs. This long residence time is consistent with the conclusion that the recharge area is many miles from the springs. Moreover, the combination of geochemical information generally excludes substantial contribution from regional flow outside of the Black Hills (Naus and others, 2001).

A long-term response in flow from Cascade Springs may reveal some interesting aspects about karst aquifer storage. Although dry conditions and declining ground-water levels have prevailed in the Black

Hills during recent years, an anomalous increase in springflow of from 1.5 to 2.0 ft³/s over the normally steady flow rates occurred during the spring of 2005. This could be explained by a long-term lag time in the low-frequency response component of springflow (or hydraulic head), which can occur in karst aquifers because of the unique storage properties of karst. If this is the case, the recent increase in springflow would be in response to wet conditions during the mid 1990s. An alternate explanation might be that declining water levels caused flowpaths to shift, possibly focusing more flow to Cascade Springs.

C₂. The Origin of Cascade Spring

(Summarized from Hayes, 1996 and 1999)

Beginning on February 28, 1992, a large discharge of red suspended sediment was observed from two of the six known discharge points. Similar events during 1906-07 and 1969 were documented by local residents and newspaper accounts, and a resident reported a similar reddening of the water during 2003, although not as prolific as in previous years.. This periodic sediment discharge at Cascade Springs probably results from episodic collapse that is caused by subsurface dissolution of anhydrite beds and cement in the middle and upper Minnelusa Formation, accompanied by replacement of dolomite by calcite. Mineralogic and grain size analyses support a hypothesis that many breccia pipes now exposed in outcrops of the Black Hills were, at one time, throats of artesian springs. Some of these springs have been abandoned as water levels have declined over geologic time due to the lowering of the Black Hills by erosion. The flow paths for Cascade Springs is believed to be along breccia pipes that also will someday be abandoned. Hence, the locations of artesian spring discharge points probably have been shifting outward from the center of the Black Hills uplift, essentially keeping pace with regional erosion over geologic time (Epstein, 2001, 2003). Furthermore, artesian springflow probably is a factor in controlling water levels in the Madison and Minnelusa aquifers, with hydraulic head declining over geologic time in response to development of new discharge points.

A suitable hydrologic and geochemical model for Cascade Springs involves dissolution of anhydrite accompanied by dedolomitization in the upper Minnelusa Formation, which is caused by upward leakage of relatively fresh water from the Madison aquifer (fig. 17). The anhydrite dissolution and dedolomitization account for the net removal of minerals, creating cavities that would lead to breccia pipe formation by gravitational collapse. Networks of interconnected breccia layers and breccia dikes are also common in the upper Minnelusa. These breccia structures, along with vertical fractures and faults, are likely pathways for transmitting upward flow from the Madison to the Minnelusa aquifer.

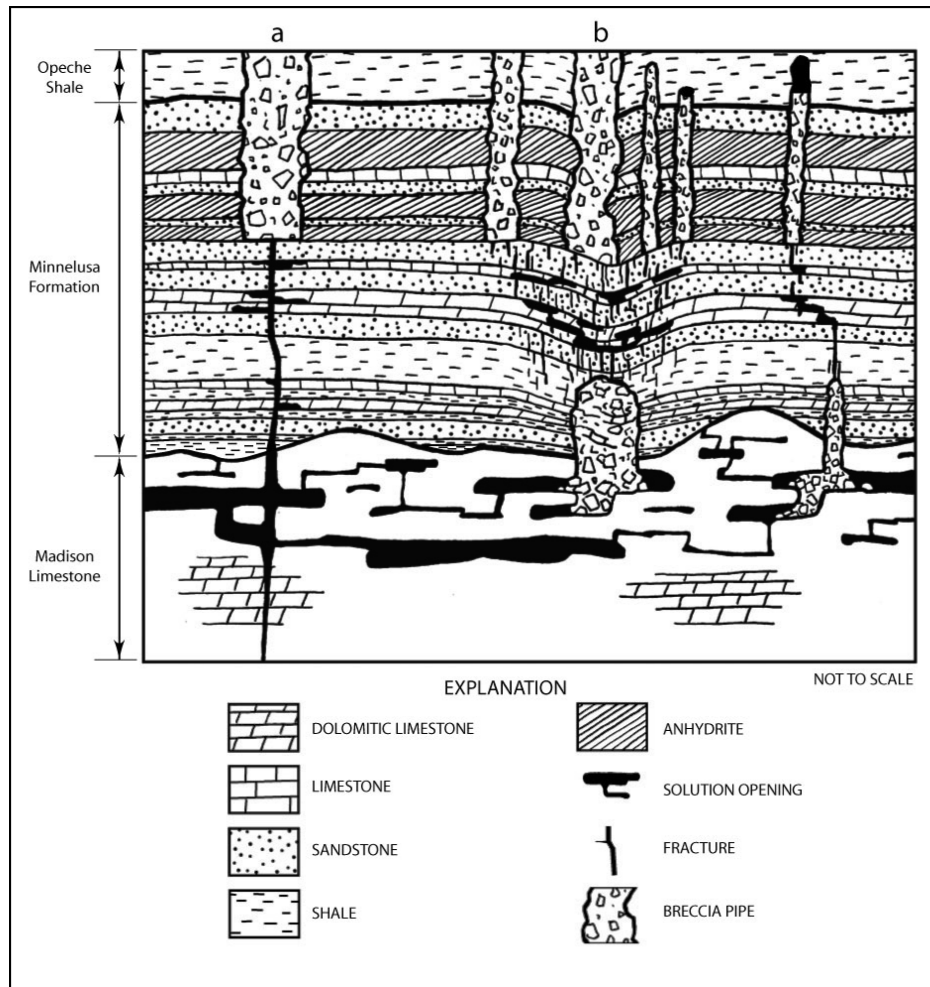


Figure 17. Geologic features that could enhance vertical hydraulic conductivity including (a) breccia pipe that formed along a fracture and (b) breccia pipe initiated by collapse of Minnelusa Formation into Pahasapa Limestone cave.

C₃. Biology of Cascade Spring

(Summarized from various materials provided by The Nature Conservancy, Black Hills Ecoregion, 8100 Sheridan Lake Road, Rapid City, SD 57702, Phone: (605) 342-4040, Fax: (605) 348-6060
<http://www.nature.org/wherewework/northamerica/states/southdakota/preserves/art9147.html>)

The springs support a unique warm riverine system that includes four rare plant species – tulip gentian, beaked spikerush, southern maidenhead fern, and stream orchid found nowhere else in the Black Hills or the surrounding Great Plains. Their presence in the Black Hills is due in large part to the warm water of Cascade Springs, which allows for their survival during severe winters. Tulip gentian (*eustoma grandiflorum*) requires a fairly high water table in moist open fields and meadows underlain by sandy alluvial soils. Its habitat has been reduced to the point that it is now rare over much of its former range. Beaked spikerush (*eleocharis rostellata*) is an obligate wetland species that occurs in many types of alkaline wetlands, including hot spring edges. It typically occurs on sand bars and along stream edges in saturated soil. Also, beaked spikerush can occur in marl beds, which are formed from calcium-carbonate precipitates. Southern maidenhead fern (*adiantum capillus-veneris*) owes its localized occurrence along Cascade Creek to the warm, limey waters of Cascade Springs. Stream orchid (*epipactis gigantea*) grows on calcareous, porous substrates or thin, partially decomposed, wet organic substrates and is more common in the open than in forests.

This area contains prime mountain lion habitat, with a nearby active den, and it includes an unusual Townsend's big-eared bat nursery roost. Because Cascade Creek remains ice-free throughout the year, it's a valuable winter fishery and provides an important water source for wildlife - especially birds. This unique riverine system has received The Nature Conservancy's second-highest biodiversity ranking, making it an outstanding example of an extremely rare natural community.

Resume driving. Leave parking lot, turn left on Route 71 back towards Hot Springs.

- 81.0 7.9** Intersection with Truck Route US 18 East, turn right.
- 81.7 0.7** Turn left on US 385 North into Hot Springs.
- 82.6 0.9** Stop light, intersection with US 16, continue straight following US 385.
- 83.4 0.8** Three-way stop sign. Turn right, continuing on US 385 North.
- 83.5 0.1** Gypsum in Spearfish Formation to left.
- 83.6 0.1** Fire tower atop Battle Mountain to right.

HISTORIC MARKER

The historic sign states: "According to tradition, American Indians were stricken with an epidemic known as "fell disease" about the middle of the 16th century that threatened to obliterate the tribes. A Messenger arrived from the Great West with news of a wonderful water which, he said, had been touched by the finger of the Great Spirit and would cure all manner of diseases. Indians came to these springs by the thousands.

After a lapse of more than 200 years, the Cheyenne took possession of the springs and built an immense tipi city covering hundreds of acres.

In the following years, the Sioux migrated west and disputed the ownership of the springs. This culminated in a fierce conflict in about 1869, the memory of which is preserved in the name of the eminence to the east, Battle Mountain, where the besieged Cheyenne established fortifications. The Sioux won the battle and possession of the springs which they called wi-wi-la-kah-to (Springs - hot). They called the area Minnekahta (Water - hot) and termed the Black Hills a great "Medicine Home".

After the Battle Mountain fight, tradition says the Sioux and Cheyenne agreed to allow the springs to be a health sanctuary to give their sick and lame the benefit of the healing waters. Around 1880, pioneers began to settle the area".

- 84.1 0.5** Quarry located in the Minnekahta Limestone to left. The Minnekahta is the most important source for aggregate in the Black Hills. We are riding in the Red Valley comprising weak beds of the Spearfish Formation, straddled between the Dakota hogback on the right and hills of the Minnelusa on the left.
- 86.2 2.1** Entering Custer County.
- 88.8 2.6** Minnelusa Formation in gulley to right.
- 89.5 0.7** Entering Wind Cave National Park. Keep an eye out for bison, coyotes, and prairie dogs.

89.6 0.1 Undulating Minnekahta Limestone at entrance to park. Poorly exposed, progressively older rocks are encountered traveling north through the park, including the Minnelusa Formation, Pahasapa Limestone, and Deadwood Formation, and Precambrian rocks.

92.5 2.9 Turn left to Park Visitor Center.

93.2 0.7 Park at Visitor Center.

STOP 5. WIND CAVE NATIONAL PARK: CAVE AND KARST VULNERABILITY

LEADERS: Rod Horrocks and Marc Ohms

Significance of Wind Cave:

- Most boxwork of any known cave
- The most complex rectilinear maze cave known in the world
- 4th longest surveyed cave in world (over 117 miles currently surveyed)
- One of best paleokarst exposures known anywhere
- One of the strongest blowing barometric wind caves
- Diverse mineralogical assemblage
- Diverse and unusual speleothem types
- Intersects the water table

Cave/Karst Vulnerability:

The National Park Service built the infrastructure to facilitate visitors viewing Wind Cave within an erosional window through the Minnelusa Formation and on top of the Pahasapa Limestone and adjacent to the natural entrance of the cave. This is, arguably, the worst place such structures could have been built. Since the Park can not remove these historic Civilian Conservation Corps (CCC)-era structures, it is now attempting to mitigate the impact from these structures on sensitive cave resources.

Sewage systems, roads, parking lots, fire suppression, and buildings change the quality, direction, and amount of water entering Wind Cave. To better protect Wind Cave, the old leaking sewer lines above the cave were replaced with dual-contained HDPE sewer lines in 2001. The inner, primary line was surrounded by an outer, secondary line, which captures and contains any leaks that may arise from the primary line. Visual inspection ports were built into the system, allowing the park to quickly and easily monitor the line for leaks and fix them before any spillage occurs.

To stop contaminated runoff from the parking lot from entering Wind Cave, the park replaced the asphalt lot with concrete in 2004. The concrete has mitigated the effect of dripping gasoline and antifreeze “melting” the asphalt and releasing hydrocarbons which were washed into the cave in as little as 6 hours. It is also no longer necessary to conduct annual chip sealing of cracks in the asphalt, a major hydrocarbon source. In addition, the new concrete lot captures all runoff and funnels it through an oil and grease separator before releasing it into Wind Cave Canyon.

Due to decades of fire suppression and planting of trees by the CCC during the 30’s, the area above Wind Cave contains many more trees than it historically did. This second overgrowth has reduced the amount of water entering the cave. The park has initiated an active prescribed fire program to reintroduce fire back into the ecosystem.

Artificial entrances to Wind Cave have impacted the cave's climate by allowing increased or unnatural airflow. When warm summer air enters the cave, it cools and water condenses on the walls of the cave.

When cool, dry winter air enters the cave it warms up and evaporates water. When the cave is expelling air due to barometric pressure changes, as much as 16.1 gallons of water is lost per hour out the natural entrance (Nepstad, 1986). The most dramatic effect of climate change on the cave were three collapses that occurred near the walk-in entrance that were caused by frost wedging. To help control these changes, a revolving door was added to the Walk-In Entrance and airlocks were added to the two elevator landings. We are currently monitoring and studying the cave climatology to help us better understand the whole system and to determine if further actions are necessary to restore natural climatological conditions.

The developed tour route portion of Wind Cave is impacted from construction-generated dust, dust from tours traveling over unpaved trails, dust tracked into caves on shoes or brought in on clothes, lint accumulations, and other materials shed from humans. Human visitors shed a plethora of materials into caves including lint, hair, dandruff, mites, microbes, shoe rubber, and pet animal fur (Jablonsky 1994). Lint and dust removal projects are conducted in Wind Cave to restore natural conditions, prevent unnatural speleothem dissolution, remove artificial food sources, eliminate unnatural odors, and to restore visual scenes. Many of these concerns, although seemingly unrelated, are part of a bigger picture. The relationships between animals, plants, fire, water, the cave and people are all interconnected. No one part of the ecosystem is separate from another. We try to understand, conserve, and interpret all aspects of Wind Cave National Park as this is the mission of the National Park Service.

Barometric Breathing Cave:

The strong barometric wind coming from or blowing into the natural entrance of Wind Cave is one of its defining characteristics and the reason for the name of the cave. This wind is a reaction to changing atmospheric pressure conditions. When there is a high pressure outside, air rushes inside towards the lower pressure. This exchange happens because the cave is fairly well sealed by the overlying Minnelusa Formation and because of the small number of entrances (2) and blowholes (6). Wind speeds up to 25 miles an hour have been reported through the natural entrance. Airflow has been documented flowing in the same direction (out or in) for up to 32 hours, thus indicating a very large volume for the cave (Pflitsch, 2002).

The barometric airflow through the natural entrance provides an opportunity to determine the approximate volume of Wind Cave. In the mid 1960's, Herb Conn built an instrument to measure the wind flow through the Natural Entrance. He calculated that Wind Cave had a volume of 56,000,000 m³ (Conn, 1966). The current surveyed portion of the cave has a volume of 1,400,000 m³, a little over 2% of his estimate (However, it should be pointed out that a significant percent of this volume may be in cracks too tight for humans to enter or in-between breakdown blocks). Conn's research was repeated by Daniels in 2001, this time taking into account the Snake Pit entrance and more accurate and precise instrumentation.

There are numerous potential sources of error in any measurements of cave airflow. There are at least six documented blowholes near Wind Cave that are likely to be connected to the cave. The revolving door at the Natural Entrance leaks air, as does the elevator shaft and Snakepit Entrance cover. Cavers have observed airflow in high dead-end domes near the surface, possibly indicating diffuse airflow through the bedrock and overburden. All of these were not accounted for in Daniels' study, and will lead to an underestimate of the airflow. However, Daniels estimated the total volume of Wind Cave to be between 6,000,000 m³ – 10,000,000 m³, significantly lower than Conn's estimate. However, it is reasonable to conclude from both Conn's 1966 research and Daniels' 2001 research, that airflow indicates that there is a significant amount of cave that has not yet been discovered in Wind Cave.

Cave Potential:

A very active exploration and mapping program, which began in the 1950's, is still on-going within Wind Cave. Currently, 3-4 miles are added to the surveyed length of the cave each year. It is not uncommon to hear cavers participating in the current survey effort remark about the "endless" potential of the cave. A famous diary quote from 1891 by an early Wind Cave explorer, Alvin McDonald, said, "Have given up the idea of finding the end of Wind Cave" (McDonald 1891). This is as true today as it was then.

Many people have speculated on the potential extent of Wind Cave, even to the possible connection with the world's second longest cave, Jewel Cave, located 18 miles to the NW. Although, theoretically possible, this seems highly unlikely based on how both caves react to surface changes in barometric pressure, how network maze caves form, and upon their geologic setting. By calculating passage density within the current boundaries of Wind Cave, a minimum potential length of humanly accessible passages in the cave was determined (Horrocks and Szukalski, 2002). Based on the total density in each of five distinct regions of the cave, a minimum length of 250 miles for Wind Cave was estimated. The current 117 miles of survey represents no more than 47% of that minimum predicted length. By examining the geologic factors, a likely potential areal extent was also identified. It was determined that the current cave boundaries cover 1/8 of the likely extent of the cave. Based on the known passage density, the length of the Wind Cave survey could be as much as 1,100 miles. The current mapped 117 miles represents about 10% of that maximum potential length. The final mapped length of the cave will depend on human caving capabilities. Whether 10% or 46% of the cave has been surveyed, it is obvious that a tremendous amount of surveyable passage remains within the cave system. Based on current mapping rates, it seems probable that Wind Cave will soon become the third longest known cave in the world. It is doubtful that it will ever attain the first or second position.

Boxwork:

Wind Cave is known for its world-class displays of boxwork. Boxwork consists of thin interconnecting veins of calcite that protrude from all surfaces within Wind Cave, especially in the Middle Level and within layers of dolomite. The formation of these calcite veins pre-date the cave. Therefore, boxwork is a speleogen, a dissolution feature that formed when the surrounding bedrock was dissolved or weathered away. These veins are truncated by paleokarst features and thus predate that event and the formation of the cave. The current theory states that these veins resulted from fractures that formed when anhydrite hydrated to gypsum and expanded, fracturing the surrounding limestone soon after the rocks were deposited. These gypsum veins were later replaced by calcite when fresh water circulated through the limestone late in the Mississippian Period. Subsequent removal of the surrounding limestone left the calcite vein fillings standing in relief. The bedrock was more easily removed because it consists of calcite crystals, which are pseudomorphs after gypsum that are held together by a sparse secondary quartz cement and which became a friable sand upon partial dissolution along the grain boundaries (Palmer and Palmer, 2000).

Paleokarst:

Wind Cave is one of the best spots in the United States to view paleokarst. Between 310-320 million years ago, a karst surface was developed on the Pahasapa Limestone. Sinkholes, vertical pipes, and short horizontal caves developed within this landscape. The caves appear to have been developed under phreatic conditions in zones of freshwater-saltwater mixing. About 310 million years ago, a marine transgression filled in these karst features with layers of red sand, silt and clay, along with fragments of limestone, chert, and sandstone derived from previously overlying sediments that were entirely eroded away. These filled-in features became an important structural control during the development of Wind Cave 40-60 million years ago. Many of these features were partially or wholly excavated when the main passages in Wind Cave formed. These paleofills are readily seen on the Garden of Eden and Fairgrounds Tour Routes in Wind Cave. These features are concentrated in the upper half of the Pahasapa Limestone.

Development of Wind Cave:

(Note: The “stages” referred to below are only for convenience in this outline and include only the major events. They are not formal divisions. – Art and Peg Palmer (personal communication, 1999).

Stage 1: Deposition of the Pahasapa Limestone (Madison) on a shallow sea floor 340-320 million years ago (Mississippian Period). Low areas of the continent were covered by shallow sea water. Some of the major cave-forming limestones and dolomites of North America were deposited at this time. Several distinct lithologic layers formed, from bottom to top: massive dolomite (route to Lakes); bedded dolomite and limestone (major boxwork zones); chert (ceiling of Ice Palace); and massive limestone (Fairgrounds, Garden of Eden). Gypsum (hydrated calcium sulfate) and anhydrite (calcium sulfate) were also deposited within some of the lower and middle layers.

Stage 2: Gypsum and anhydrite are physically and chemically unstable. Soon after the rocks were deposited (about 320 million years ago) they were uplifted slightly above sea level, allowing the following to happen:

Anhydrite hydrated to gypsum, causing expansion that formed many small cracks in the surrounding rocks, especially the dolomite beds in the middle Pahasapa Limestone. Dissolution of dolomite followed by crystallization of calcite, as well as plastic deformation of the sulfates, probably contributed to the fracturing. Subsequent calcite was deposited in the fractures.

The pressure of the overlying rocks forced the gypsum and anhydrite to migrate into fractures in the surrounding rock.

Reduction of gypsum and anhydrite in the deeper layers produced hydrogen sulfide, which migrated upward to oxygen-rich areas, where it was oxidized to sulfuric acid. The reaction of this acid with the surrounding limestone formed the earliest cave openings--generally small pockets and fissures--and the adjacent limestone and dolomite were altered to a weak, crumbly, bleached zone. The basic layout of the cave passages was determined at this time.

Some of the hydrogen sulfide combined with dissolved iron to produce iron sulfide (pyrite, etc.).

Stage 3: The climate became wetter, and considerable amounts of fresh water entered from the surface. Gypsum and anhydrite were replaced by calcite. Oxidation of the iron sulfide around the old hydrogen sulfide zones produced red and yellow zones of iron oxide in and around the cave, and the calcite deposited at this time is orange-brown as a result. This includes the veins that now protrude as boxwork fins. In the upper strata, gypsum was simply dissolved away, leaving a fractured jumbled breccia in the limestone (as in the Garden of Eden).

Stage 4: Eventually the climate became so wet that sinkholes and solutional fissures formed at the surface, and new caves were formed or enlarged forming part of the upper level of the cave, most of which were subsequently filled by sediment during stage 5. Some of the fissures extended below the chert level intersecting the cave network established in Stage 2. Much of the cave enlargement simply followed earlier openings, fractures, and altered rock zones.

Stage 5: About 300 million years ago, during the Pennsylvanian Period, a rise in sea level caused the lower part of the Minnelusa Formation (mainly sandstone) to be deposited, filling in the sinkholes, fissures, and most early caves. Much of the sandstone was deposited by rivers along the paleo-shoreline. The red sand and clay deposits in the Beauty Parlor and Garden of Eden are derived from the lowest layers of this formation.

Stage 6: Continued deposition of sediments buried the Pahasapa Limestone to a depth of at least one mile during the Pennsylvanian through Cretaceous Periods (300-70 million years ago). A layer of white calcite (dogtooth spar) was deposited on the walls of earlier cave openings.

Stage 7: The Black Hills and Rocky Mountains began to rise about 70 million years ago. Mobilization of deep fluids early in this stage caused deposition of hydrothermal minerals in some of the early caves and pockets, including quartz crystals (Crown Jewels, etc.).

Stage 8: As the Black Hills continued to rise, the sedimentary rocks were stripped off by erosion, exposing very old (Precambrian) igneous and metamorphic rock at the center of the uplift (Harney Peak, Mt. Rushmore, etc.). The eroded edges of the sedimentary rocks, including the Pahasapa Limestone, were exposed around the perimeter of the Black Hills. Groundwater moved through the rocks in considerable volume, and most cave enlargement took place during this time. Again, the enlargement was concentrated along the zones of older cave development and alteration. Except in a few places the cave does not extend to the top or bottom of the Pahasapa Limestone, nor does it extend far below the water table. Evidently the cave is not the product of simple artesian groundwater flow, infiltration from the surface, or hydrothermal water rising from depth. If it were, the cave would be largest where the water first entered the limestone. It is clearly the result of mixing between two or more of these water sources, which produced a zone of solutionally aggressive water. Its main solutional phase was about 60-40 million years ago.

Stage 9: As the water table dropped, weathering of the limestone walls took place and continues today. Most important, the crumbly, altered dolomite of Stage 2 has decomposed into a powdery sand that formed files of sediment on the cave floors and allowed the thin calcite veins (mainly the orange-brown veins of Stage 2) to protrude as boxwork. Bedrock walls have developed a thin weathering rind of fluffy powder, stained red, yellow, and black from the oxidation of minerals in the rock (such as the pyrite from Stage 2). Moist air rising from the lower levels produces condensation on the walls of the cooler upper levels and condensation erosion. This dissolution produces domes and chutes in the ceilings of upper level passages. The condensation moisture becomes saturated with dissolved calcite and seeps through the bedrock to the water table, although some of it evaporates in the lower levels to produce aragonite frostwork and popcorn. Other deposits are also formed by water seeping from the surface. Active drips fed by infiltration from surface water deposit flowstone, stalactites, etc. In zones of ponding, such as the lakes, a wall crust of calcite has formed, and calcite rafts form at the surface. This water is supersaturated with calcite and cannot dissolve limestone. Older crusts higher in the cave, including dogtooth spar in larger openings (Stage 6), have been shaved off by condensation corrosion and weathering.

Major time indicators:

Orange-brown calcite (boxwork fins); older than 320 million years. It is cut by the red paleofill and never occurs in the paleofill, except as eroded fragments.

Red sediment fill (sand and clay paleofill); about 300 million years old. This marks the well-known unconformity that separates the Mississippian and Pennsylvanian rocks throughout the western states. Be careful--many of the dolomite beds weather to red colors too, as in the Post Office. Much of the paleofill in the cave has subsided into lower levels as the cave enlarged, but this can be easily recognized by the lack of (or disruption of) the white calcite coatings and vein fillings (described below).

White calcite veins and including dogtooth spar ; between 300 and 70 million years ago, probably toward the younger end. This fills cracks and coats pockets in the red paleofill, so it is definitely younger. It is cut by the present fills and overlain by deposits such as wall crusts.

Quartz crystals ; about 100-70 million years old. In places they coat the dogtooth spar, especially along faults.

The cave itself has an origin that spans the entire period from about 320 million years ago to the present. The major solutional phase was about 60-40 million years ago, during which time the present topography developed. However, the present cave follows the patterns of the early gypsum and anhydrite zones, as shown by the fact that the orange calcite (originally gypsum) is concentrated only around the present caves. In large breakdown or blasted areas the density of calcite veins can be seen to diminish away from the cave. Therefore, the cave pattern predates the uplift of the Black Hills. The pattern seems well adjusted to the Black Hills uplift, with fractures radiating away from the center of the hills. However, the Black Hills dome has long been an area of uplift and weakness in the earth's crust, and fractures tend to maintain the same patterns and are repeatedly reactivated.

Resume driving. Leave Visitor Center returning to US 385.

- 93.9 0.7** Turn left on US 385 towards Custer.
- 94.7 0.8** Sewage lagoon on right. Minnelusa exposed on cliff face.
- 95.0 0.3** Poorly exposed Minnelusa littered with cobbles from the overlying White River Formation.
- 96.1 1.1** Prairie dog mounds on right. Coyotes may be seen occasionally examining these morsels.
- 96.4 0.3** Junction with Route 87 north. Continue straight on US 385.
- 96.7 0.3** Sandstone outcrops of the Deadwood Formation in creek to right. For the next 50 miles we will be riding on a variety of Precambrian metamorphic and igneous rocks making up the core of the Black Hills. Lovely scenery, no karst, enjoy the views!
- 102.9 6.2** Junction with Route 89, bear right on US 385 and US 89 north towards Custer.
- 106.2 3.3** Mine tailings to right from feldspar workings in pegmatite.
- 114.2 8.0** Traffic light. Intersection with US 16. Turn right on US 16/US 385/Rte 89 into Custer.
- 114.6 0.4** Traffic light in Custer. Make left turn onto US 16 towards Chief Crazy Horse Mountain and Mount Rushmore.
- 116.8 2.2** The hills to the right are held up by the Harney Peak granite, intruded into mica schist. Chief Crazy Horse Mountain carving at 2 o'clock.
- 117.2 0.4** Chief Crazy Horse at 1 o'clock.
- 119.0 1.8** Entrance to Chief Crazy Horse Mountain (figure 18). Chief Crazy Horse and Chief Sitting Bull were responsible for the defeat of Custer at the Little Big Horn, following conflicts with the White Man after gold was discovered in the Black Hills in 1874. The monument, carved in pegmatitic granite of Thunderhead Mountain, is the largest in the world, as tall as the Washington Monument and taller than the pyramids at Giza. Sculptor Korczak Ziolkowski was invited by Lakota Indian Chiefs to carve Crazy Horse in the Black Hills in 1939 and he began blasting in 1948. When completed the mountain carving will be 641 feet long by 563 feet high. The head of Crazy Horse is 87 feet tall or 22 stories high, and equals in volume all the heads at Mt Rushmore. His outstretched arm is nearly as long as a football field and points to

“my lands are where my dead lie buried”, a quote reportedly given in response to a taunting question asked of a white man after the defeat of the Indians.



Figure 18. Chief Crazy Horse Memorial and model.

- 121.6 2.6 Cross Pennington County line.
- 124.9 3.3 Route 89/87 to right, the Needles Highway to Sylvan Lake, continue straight ahead.
- 125.0 0.1 Turn right on Route 244 East.
- 132.1 7.1 Harney Peak Granite country.
- 132.3 0.2 Enter Mt. Rushmore National Memorial (figure 19).

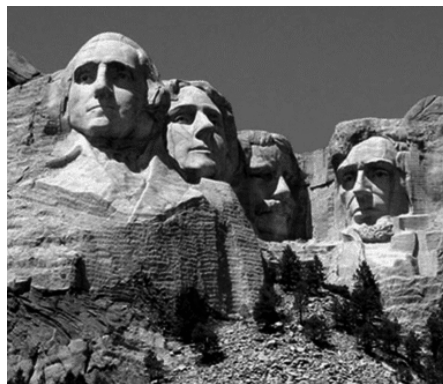


Figure 19. Mt. Rushmore National Memorial.

“The birth of our nation was guided by the vision and courage of George Washington. Thomas Jefferson always had dreams of a greater, more perfect nation, first in the words of the Declaration of Independence and later in the expansion of our nation through the Louisiana Purchase. Preservation of the union was paramount to Abraham Lincoln, a nation where all men were free and equal. At the turn of the Twentieth Century Theodore Roosevelt envisioned a great nation, a leader on the world stage, our nation was changing from a rural republic to a world power. The ideals of these presidents laid a foundation for the United States of America as solid as the rock from which their figures were carved”. (Mt. Rushmore National Park service Web Site).

133.8 1.5 Traffic light; main entrance to Mt. Rushmore national Memorial, continue straight.

Mount Rushmore National Memorial is carved into the Harney Peak Granite that intruded older Precambrian schist about 1.7 billion years ago. Sixty-foot-high heads of four presidents, representing the first 150 years of the Nation's history, are carved in bold relief: George Washington, Thomas Jefferson, Abraham Lincoln and Theodore Roosevelt. The sculptor, Gutzon Borglum, selected Mt. Rushmore because of the consistency of the granite, dominating height, and the southern exposure. Work began in 1927, funded by the Federal government, and completed 14 years later.

134.2 0.4 Nice view of Mt. Rushmore in your rearview mirror.

135.3 1.1 Junction with US 16A. Continue straight on US 16A towards Keystone.

135.7 0.4 Keystone.

138.7 3.0 Bear right on US 16 towards Rapid City.

143.7 5.0 Rockerville, a town straddled by the four-lane highway.

145.7 2.0 Entering *Touristville*.

147.1 3.4 Minnelusa on left.

148.2 1.1 Minnekahta Limestone atop Opeche red shales.

148.4 0.2 Undulating Minnekahta Limestone.

148.9 0.5 Crossing the Red Valley in the Spearfish Formation. Note gypsum ledges.

150.2 1.3 Sandstones in the Sundance formation cap hill on left.

150.8 0.6 Flat uplands on top of the Dakota hogback underlain by sandstones of the Inyan Kara Group. Great Plains in distance to the right and mountains with Precambrian rocks to left.

153.8 3.0 Entering Rapid City while descending through the sandstones of the Inyan Kara Group.

155.6 1.8 Omaha Street, turn right.

156.9 1.3 Fifth Street, turn left.

157.1 0.2 Turn left on New York Street and immediate right into Holiday Inn Parking lot. End of trip.

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