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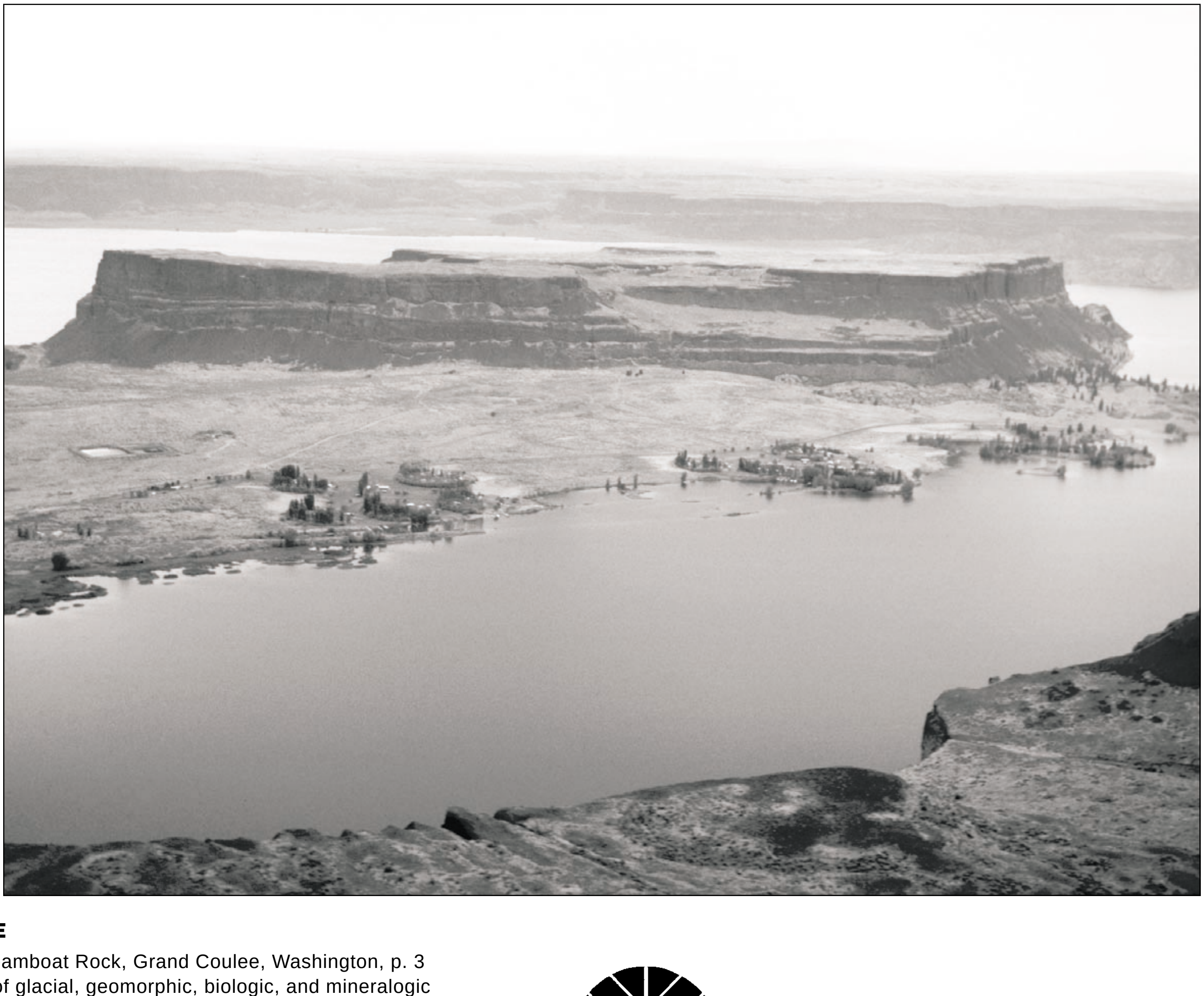

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WASHINGTON STATE DEPARTMENT OF
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The Future of Washington Geology

We apologize for our sporadic publication schedule over the last two years. Budget cuts and understaffing have taken their toll. Starting with the year 2000, we plan to issue *Washington Geology* three times a year—in March, July, and November. Issues will be smaller, but we plan to keep the same high standards. We will still be looking for articles on various facets of the geology of Washington that appeal to both the professional geologist and the interested amateur. We do not charge for publication, but neither do we pay the author. Instead, authors receive 20 complimentary copies of the issue containing their contribution. Keep in mind that *Washington Geology* is not copyrighted.

Advice to Prospective Authors

To submit a manuscript to *Washington Geology* (or a more technical manuscript for publication in some other format), send a paper printout and an electronic file (preferably in WordPerfect for the PC) to the editor at Washington Division of Geology and Earth Resources; PO Box 47007; Olympia, WA 98504-7007. Figure numbers and annotations for all illustrations should be indicated on a photocopy, not on the photo itself. Photos should be unmounted and included separately. Captions should be at the end of the text file.

Our style is similar to that of U.S. Geological Survey publications. (See *Suggestions to Authors of the Reports of the United States Geological Survey*, 7th ed., 1991, and recent issues of *Washington Geology.)* Authors are responsible for the accuracy of their manuscript. It should be reviewed by several experts in the subject before it is sent to us. Include the names of the reviewers in the acknowledgments. The bibliography should be limited to references cited. If you are not sure how to cite a reference, send us a photocopy of the cover and title page of the book or a copy of the article.

Photos may be color or black and white, slides or prints. We can accept e-files if they are scanned in black and white at 300 dpi at 7 inches wide, 600 dpi for color. We prefer photos to be of good quality, properly exposed, with no spots, etc., but we can do a certain amount of electronic manipulation to improve the image, if necessary. We will still take diagrams on paper, but we prefer them in electronic form as vector (draw) files. Sketches can be 300 to 600 dpi bitmaps. Electronic files may be attached and e-mailed to jari.roloff@wadnr.gov.

We can also use stand-alone photos of interesting geologic sites in the state if they are accompanied by an explanatory caption. If they are used, we will credit you.

We welcome letters or comments responding to articles in *Washington Geology*. Notices of meetings (far enough in advance) or websites that may be of interest to our readers should be e-mailed to jari.roloff@ wadnr.gov. ■

Cover Photo: Oblique aerial photo of Steamboat Rock from approximately the southeast corner of the study area map (Fig. 1) on p. 3. Photo shows Steamboat Rock's position in Banks Lake with the walls of upper Grand Coulee in the foreground and background.

USGS/UW/DNR URBAN GEOLOGIC MAPPING

For information about the USGS/UW/DNR urban geologic mapping efforts, see http://www.washington.edu/newsroom/news/1999archive/02-99archive/k021999.html
http://www.geophys.washington.edu/SEIS/
Info line: 206-543-7010

Geology of Steamboat Rock, Grand Coulee, Washington

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INTRODUCTION

Steamboat Rock is a large monolith located in the center of the Grand Coulee in north central Washington (cover, Fig. 1). The rock is triangular in shape, 2.1 mi (3.3 km) long and 0.75 mi (1.2 km) wide. At 2,312 ft (628 m) elevation, it stands 898 ft (274 m) above Banks Lake, a reservoir that currently occupies the floor of the northern portion of Grand Coulee. At the base of Steamboat Rock, a major pre-Miocene unconformity is visible. Below this nonconformity are Jurassic, Cretaceous, and early Tertiary granitic and metamorphic rocks. Most of the monolith is composed of the Grande Ronde and Wanapum Basalts of the Columbia River Basalt Group. The top of Steamboat Rock is traversed from east to west by a coulee approximately 100 ft (30.5 m) deep. A thin veneer of regolith (unconsolidated rock material) covers the top of the monolith, through which basalt bedrock protrudes in places. Large granitic boulders or *erratics* litter the surface of Steamboat Rock and can also be found at its base. Sinuous piles of sediment that cross the northern portion of the rock are interpreted as moraines deposited by the Okanogan lobe of the Cordilleran ice sheet.

In this paper, we examine Steamboat Rock in order to establish a sequence of events during the late Pleistocene glaciation of north central Washington. Of particular interest are floods

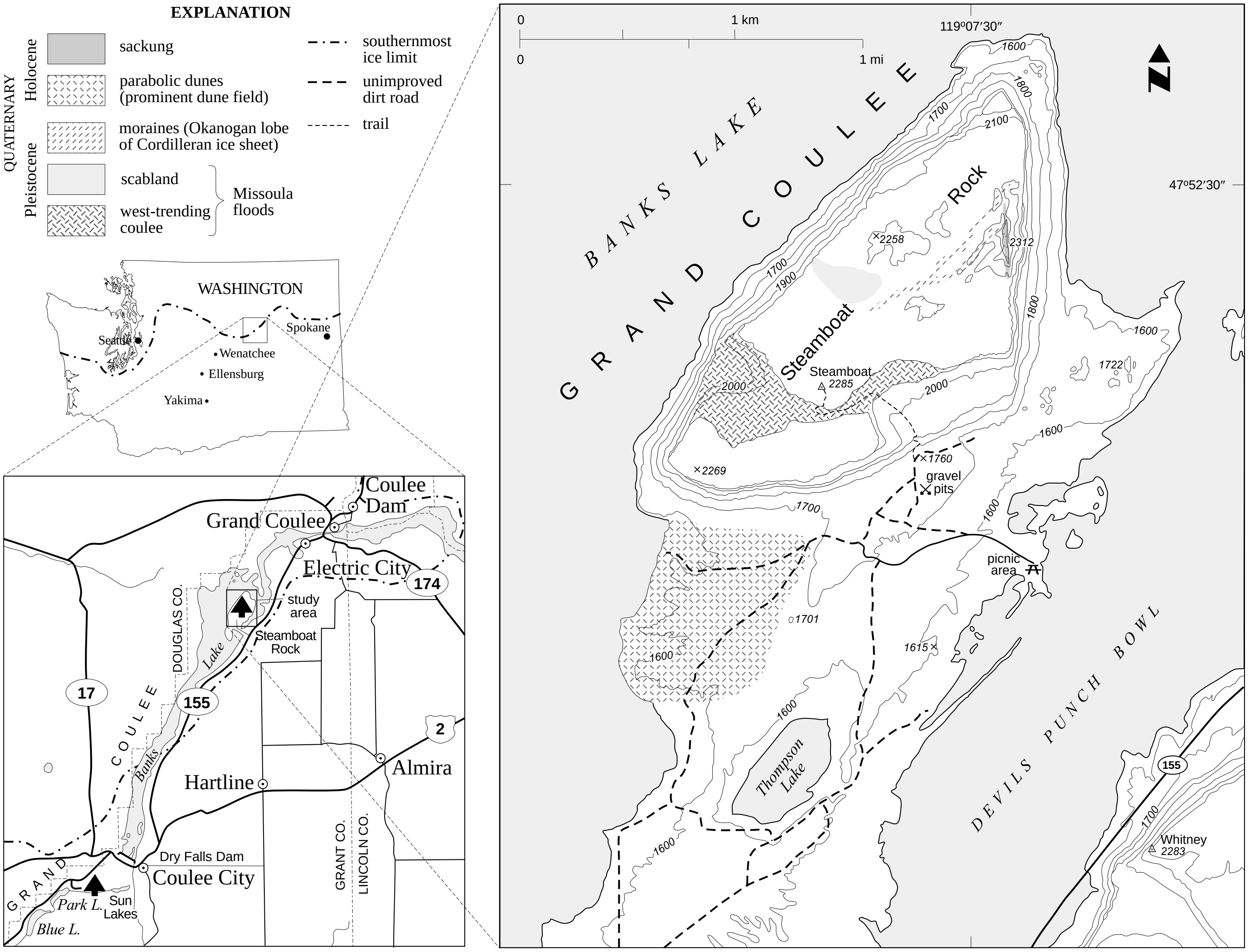

Figure 1. Location maps and detailed map of the study area showing geomorphic features mentioned in text.

from glacial Lake Missoula and advances of the Okanogan lobe of the Cordilleran ice sheet.

BEDROCK AND STRUCTURAL GEOLOGY

Two principal rock types can be distinguished in the vicinity of Steamboat Rock. The most visible of these is the Miocene Columbia River Basalt Group. Steamboat Rock and the rest of the Grand Coulee region are underlain by horizontal to slightly folded flows of the Grande Ronde Basalt (N_2 magnetostratigraphic unit) and the Wanapum Basalt (Priest Rapids and Roza Members) of the Columbia River Basalt Group (Gulick and Korosec, 1990). Within the upper portion of Grand Coulee, the Columbia River basalts unconformably overlie Jurassic, Cretaceous, and early Tertiary crystalline rocks.

These plutonic, hypabyssal, and high-rank metamorphic rocks are exposed directly north and east of Grand Coulee. The nonconformity between the crystalline rocks and the extrusive basalt is an erosion surface that displays considerable relief and is representative of regional topography prior to the eruption of the Columbia River basalts (Hanson, 1970).

The nonconformity between the crystalline rocks and the overlying basalt flows is exposed near the shoreline of Banks Lake. It is visible at the northeastern end of Steamboat Rock (Fig. 2), as well as in the west-facing wall of Grand Coulee opposite the monolith (Fig. 3). Granite is also exposed as a number of small islands in Banks Lake.

The basalt flows of the Columbia River Basalt Group are generally horizontal in the region of Steamboat Rock. The exception to this is the Coulee monocline located 9 mi west of the town of Coulee City (Fig. 4). The monocline has several hundred feet of down-to-the-southeast relief (Hanson, 1970). The Grand Coulee is cut into the basalt flows at the base of the Coulee monocline, probably as a result of structural weakness caused by the folding (Weis, 1982).

Figure 2. Nonconformity at the base of the northeastern end of Steamboat Rock. Below the nonconformity are Jurassic, Cretaceous, and early Tertiary crystalline rocks; above the nonconformity is Miocene Columbia River basalt.

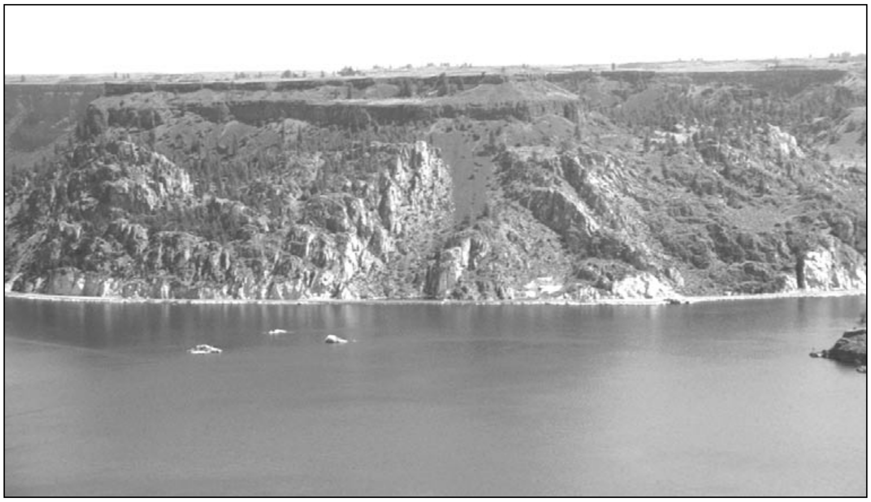

Figure 3. Nonconformity exposed in the west-facing wall of the Grand Coulee, opposite Steamboat Rock. Miocene Columbia River basalt overlies Jurassic, Cretaceous, and early Tertiary crystalline rocks.

QUATERNARY GEOLOGY

Cordilleran Ice Sheet

Richmond and others (1965) state that the Cordilleran ice sheet advanced into north central Washington at least twice during the late Pleistocene. They determined that the Okanogan lobe of ice traveled southeast across the Waterville Plateau, reaching its southern limit near U.S. Highway 2, west of Coulee City. Mapping by Gulick and Korosec (1990) suggests that the ice reached its maximum southeastern limit at the eastern edge of the Grand Coulee, just beyond Steamboat Rock.

Missoula Floods

The Purcell lobe of the Cordilleran ice sheet advanced south into Idaho down the Purcell Trench, damming the Clark Fork River (Fig. 4). Glacial Lake Missoula developed behind the ice dam. Periodically, the ice dam failed releasing vast quantities of water westward in glacial outburst floods or *jökulhlaups*. This process of glacial damming followed by catastrophic flooding occurred repeatedly, with evidence for at least 40 separate jökulhlaups between 15,300 and 12,700 yr B.P. (Waitt, 1980; Atwater, 1984). The jökulhlaups raced across northern Idaho and then west and south across Washington. One path of the floodwaters was along the base of the Coulee monocline (Weis, 1982). Geologic evidence indicates that enormous floods swept across this area in at least two earlier times, about 200,000 yr B.P. and more than 790,000 yr B.P. (Baker, 1978; Reidel and Fecht, 1994; O'Conner and Waitt, 1995a,b). These jökulhlaups formed eastern Washington's Channeled Scabland.

Evidence for Glacial Advance and Jökulhlaups at Steamboat Rock

Freeman (1937) recognized evidence for glaciation of the top of Steamboat Rock. He described the large granitic boulders as being "brought for many miles and left by the ice when it

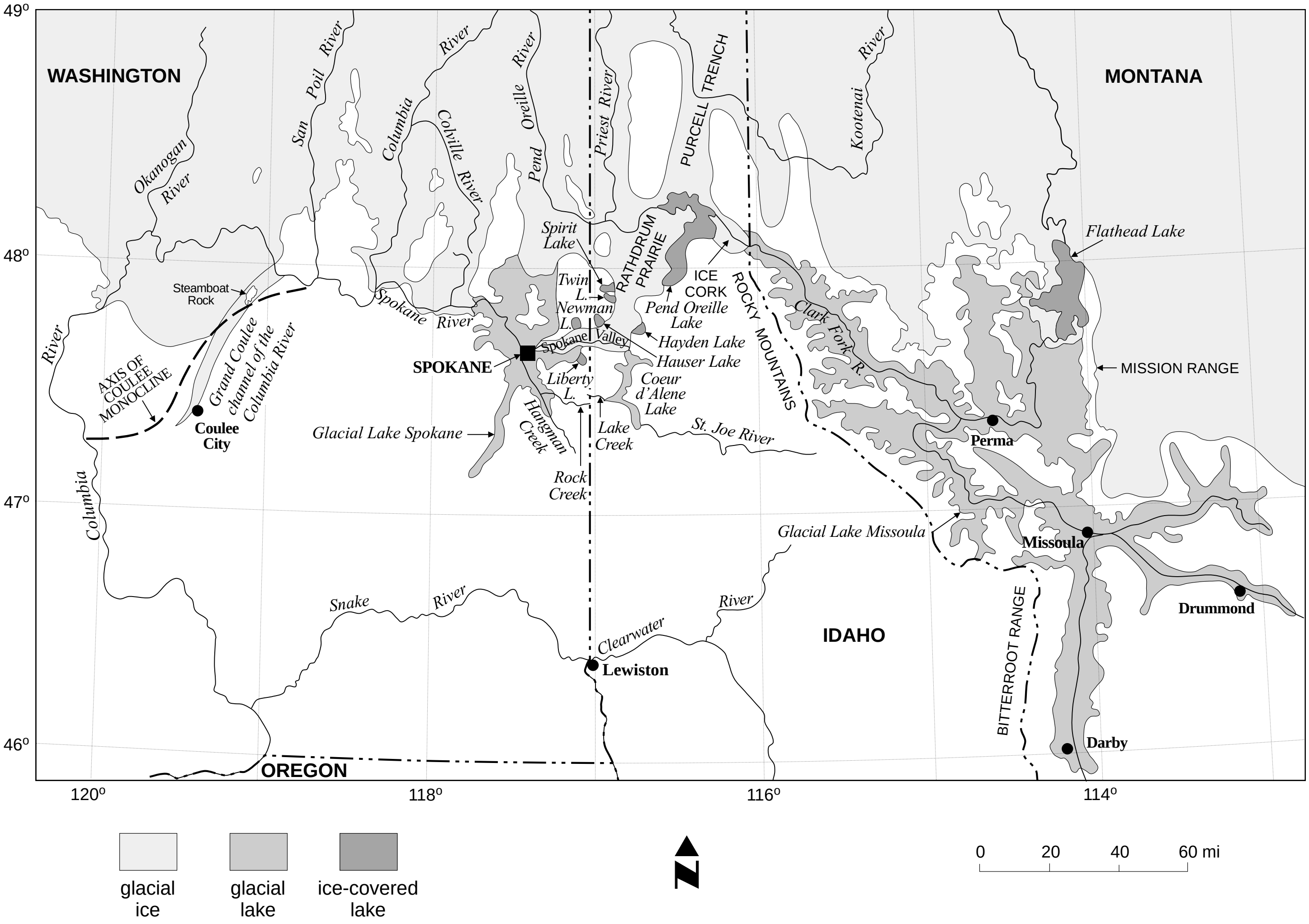

Figure 4. The advance of glacial ice and the corking of the Clark Fork River. Note that the formation of glacial Lake Missoula required the existence of a large river system located in deep mountain valleys lying almost entirely to the south of the glacial ice margins. *(Redrawn from Weis and Newman, 1989.)*

melted”. Freeman also noted the presence of moraines across the northern third of the rock.

Our examination of Steamboat Rock revealed evidence supporting Freeman’s theories. Granitic boulders noted by Freeman (Fig. 5) are distributed across much of the summit of the monolith. This suggests that the boulders were glacially transported as opposed to being ice-rafted or deposited by jökulhlaups. The two moraine-like features are composed of unsorted, unstratified sediment ranging from silt to boulders, interpreted herein as till. The sinuous shape and hummocky topography of these features confirm they are moraines. The morphology of the moraines indicates that they were deposited during the most recent glaciation, a fact that is supported by the lack of weathering rinds on basalt cobbles within these landforms. Based on the presence of moraines and ice-transported boulders on top of the monolith, we conclude that Steamboat Rock was covered by ice at least once during the late Pleistocene.

Figure 5. A large granitic erratic. Many boulders of this size litter the top of Steamboat Rock.

Most of the top of Steamboat Rock clearly reveals evidence for jökulhlaups. The scabland topography and lack of till on the southern two-thirds of the monolith provide proof that water flowed over top of the monolith with enough velocity to perform significant erosion.

Two hypotheses have been advanced for the origin of the 100-ft (30.5-m)-deep west-trending valley (Fig. 6) that traverses Steamboat Rock. Freeman (1937) proposed that the valley was cut by a tributary of the river that flowed through the area now occupied by the Grand Coulee. In other words, it is a remnant of the pre-glacial and pre-flood land surface.

An alternative hypothesis suggests that the west-trending valley was formed by floodwaters eroding a weakness in the jointed surface of the Columbia River basalt. Therefore, the valley could be considered a coulee. It is likely that the valley was formed by a combination of both processes—it was probably occupied by a small stream previous to any flood event and enlarged by the jökulhlaups.

The relation of flooding to the advance(s) of the Okanogan lobe has not been well documented. At least one of the probable ice-transported boulders stands atop a basalt pedestal that appears to have been cut by floodwaters (Fig. 7). This indicates that ice of the Okanogan lobe of the Cordilleran ice sheet advanced onto Steamboat Rock before catastrophic flooding. Moraines left on top of channeled scabland topography, however, seem to suggest that ice advanced onto Steamboat Rock after significant flooding (Fig. 8). In addition, the moraines appear to be unaltered, indicating that they were not touched by fast-moving floodwaters. Finally, granitic boulders and till are found in the west-trending coulee thought to have been modified by Missoula floodwaters, further pointing to glacial advance after extensive flooding.

Figure 6. The west-trending coulee that traverses the top of Steamboat Rock.

Figure 7. A granitic boulder resting atop a flood-scoured basalt pedestal suggests glaciation to deposit the erratic, followed by flooding that eroded the basalt except under the boulder.

Proposed Sequence of Quaternary Events at Steamboat Rock

Close examination of the evidence seems to indicate that there were at least two separate glacial advances onto Steamboat Rock and at least three separate series of jökulhlaups (Fig. 9).

The first event in the formation of the monolith was a series of catastrophic floods long ago (before 15,300 yr B.P.) that likely shaped and possibly even created Steamboat Rock. This flood was also responsible for carving the west-trending coulee across the top of the monolith. Following this flood, we believe an advance of the Okanogan lobe completely covered Steamboat Rock. Complete glaciation of the monolith is suggested by Gulick and Korosec (1990), who mapped the maximum southeastern extent of the Okanogan lobe at approximately the eastern edge of the Grand Coulee. This glaciation deposited many granitic erratics that litter the surface of the monolith and also left behind lodgment till that is preserved in the west-trending coulee.

Following retreat of the ice, a second series of jökulhlaups scoured the monolith. These floods removed much of the till deposited by the first glaciation, leaving bedrock exposed at the surface. In addition, the floods created the scabland topography that dominates much of the top of Steamboat Rock.

After a hiatus of many thousands of years, the Okanogan lobe of the Cordilleran ice sheet again advanced to Steamboat Rock. This glaciation was only partial however, with ice covering just the northern third of the monolith. The position of the sinuous moraine across Steamboat Rock indicates the maximum extent of the glacier during this advance. In places, the moraines deposited during this advance overlie scabland features formed by earlier flood events (Fig. 8).

Finally, the Missoula floods, the most recent catastrophic flooding event in eastern Washington, rushed through the Grand Coulee. Given that the moraines left by the second glaciation are preserved in an unaltered condition, we believe that the Missoula floods lacked the size necessary to overtop the monolith. These jökulhlaups did, however, remove any glacial landforms from around the base of Steamboat Rock.

Other Landforms at Steamboat Rock

A linear depression at the northeast edge of Steamboat Rock is a ridge-top depression or *sackung* (Fig. 10). Collectively termed *sackungen* (McCalpin and Irvine, 1995), the features are characterized by linear trenches that occur on steep slopes, usually near ridge crests (although here they are on a mesa). Sackungen form as a result of large-scale gravitational spreading on steep-sided ridges (Savage and Varnes, 1987). The sackung at Steamboat Rock probably resulted from the rapid steepening and erosion of the monolith's walls by the Missoula floods. The highly jointed nature of the Columbia River basalt flows provided pre-existing weaknesses. Although it is difficult to determine the age of the sackung, McCalpin and Irvine note that such features are generally early post-glacial in age.


Figure 8. Moraine draped across basalt scabland. This moraine was not eroded by later floods.

At the base of the south side of Steamboat Rock is a large closed depression in Quaternary sediments (Thompson Lake, Fig. 1). It is possible that this feature is a scour depression, formed by the raging Missoula floodwaters as they passed around the monolith. The closed depression could also be a kettle, the final resting place for an iceberg transported by the jökulhlaups.

Also present on the floor of Grand Coulee, near the base of Steamboat Rock are a number of dunes. These dunes are parabolic in shape and are composed of quartz and basalt sand (Hanson, 1970) likely derived from Missoula flood deposits within the Coulee. Small dune complexes such as this one are common along the Columbia River in eastern Washington, and larger examples are the Pot Holes dunes near Moses Lake and the Juniper Dunes northeast of the Tri-Cities.

1. *Cretaceous and early Tertiary*
   Intrusion and metamorphism
2. *Mid-Tertiary*
   Erosion
3. *Miocene*
   Flood basalts of the Columbia River Basalt Group
4. *Pleistocene*
   A. Huge jökulhlaups produce large-scale topography
   B. Complete glaciation of summit of Steamboat Rock
   C. Huge jökulhlaups sweep over Steamboat Rock but leave a few erratics from earlier glaciation
   D. Glaciation of the northern third of Steamboat Rock results in moraines
   E. Moderate jökulhlaups surround but do not overtop Steamboat Rock
4. *Late Pleistocene to Holocene*
   A. Dunes form near base of Steamboat Rock
   B. Talus collects due to rockfall from Steamboat Rock
   C. Sackung develops near summit
5. *Late Holocene*
   Manmade dams result in Banks Lake

Figure 9. Proposed geologic history of Steamboat Rock.

CONCLUSIONS

Glacial advances of the Okanogan lobe of the Cordilleran ice sheet as well as jökulhlaups have dominated the late Cenozoic geologic history of Steamboat Rock. The presence of ice-transported granitic boulders and moraines on the top of Steamboat Rock provides evidence that glaciers rested atop the rock during the late Pleistocene. Scabland topography and a large granitic boulder atop a basalt pedestal formed by erosion, indicate that jökulhlaups overtopped Steamboat Rock with considerable erosive power. Although the sequence of glacial advances and floods is difficult to establish, we believe that the Okanogan lobe of ice covered Steamboat Rock at least twice during the late Pleistocene. In addition, we propose that a minimum of three jökulhlaups also affected the geomorphology of the monolith. The first of the glacial advances completely covered Steamboat Rock and likely occurred after a significant flood event. The second advance probably took place between later flood episodes and was restricted to the northern third of the rock as shown by distinct moraines. These moraines were unaffected by later jökulhlaups of lesser magnitude which did not reach the top of Steamboat Rock, therefore preserving the drift left by the second glacial advance.


Figure 10. Sackung near the east edge of the top of Steamboat Rock.

HIKING STEAMBOAT ROCK

If you are interested in hiking Steamboat Rock, Figure 1 shows the roads and trails. For more information, see "Hiking Washington's Geology" by Babcock and Carson, which will be available in April from The Mountaineers. For more information on the availability of this book, call 1-800-553-4453.

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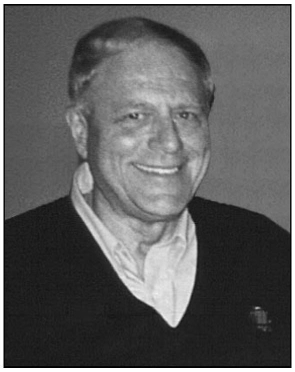
MINER'S POET LAUREATE

The National Mining Hall of Fame and Museum in Leadville, Colo., named Verne Boston of Mabton, Wash., the first place winner in the Miner's Poetry Jamboree and conferred on him the title "1999 Miner's Poet Laureate." The contest attracted 37 poets from 11 states and England and a total of 93 poems.

Boston worked as a raise miner in various camps around the West and served as an inspector for the Mine Safety and Health Administration before retiring in 1986. He has composed poetry all his life and is compiling his first book of verses.

EASTERBROOK FUND ESTABLISHED WITHIN THE GSA FOUNDATION



Don J. Easterbrook

The Don J. and Ellen H. Easterbrook Fund has been established within the Geological Society of America Foundation. The fund will provide a research grant to a distinguished scientist selected by GSA's Quaternary Geology and Geomorphology Division. The first award will be made during the GSA Annual Meeting in Reno in November of 2000. The award will initially be in the range of $15,000 to $30,000, and grants may eventually reach $100,000 or more annually.

The Easterbrook Fund will also provide support for other scientific projects in Quaternary geology and geomorphology, including acquiring, archiving, and disseminating outstanding photographs, satellite and digital elevation images, and various other types of images for the Easterbrook Library of Outstanding Geologic Photos. In addition, as the fund grows, support for publications, education, research, and other programs will be available.

Don Easterbrook, a professor of geology at Western Washington University since 1968, earned his B.S., M.S., and Ph.D. in geology from the University of Washington. He has held offices in national and international professional societies and published many papers on glacial geology, geomorphology, slope stability, and volcanic processes, in addition to several books on surface processes and landforms.

When asked about the genesis of the fund, Don Easterbrook said, "Ellen and I have been thinking for some time about ways to return to our science some of the intellectual benefits that have made our lives enjoyable for more than 40 years. Looking back at my own career and that of others, the one thing there was never enough of was money, to follow and develop all the ideas that we geologists have a propensity to spawn. With the realization that through GSA and the Foundation we could establish a fund that would accomplish what we wanted to do through current gifts now and an endowment later, everything fell into place."

GSA President Gail Ashley said, "Don and Ellen Easterbrook have given us a unique, two-part way, first, to recognize scientists who in their work have made breakthrough discoveries and significantly advanced scientific knowledge, and, second, to reward them generously with research funds that will give them the wherewithal to make even greater contributions. All of us in GSA and earth science are in their debt."

Modified with permission from GSA Today, December 1999

Deadline for nominations for the Don J. Easterbrook Distinguished Scientist Award is April 1, 2000. For more information, see *GSA Today*, January 2000.

Observations of Glacial, Geomorphic, Biologic, and Mineralogic Developments in the Crater of Mount St. Helens, Washington

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INTRODUCTION

Mount St. Helens is an active andesite-dacite volcano that is currently in a semi-dormant state after a catastrophic explosive eruption in May 1980 and subsequent eruptions through 1986. During these eruptions, a dacite dome called Lava Dome has grown over the volcanic vent in the crater. Since the winter of 1982–83, the crater floor has been progressively covered by a layer of snow, firn, and ice mixed with rock debris.

This paper describes firn caves and recent geomorphic, biologic, and mineralogic developments in the crater of Mount St. Helens. The caves are a system of melt passages that have formed in the crater ice body since the mid-1980s. Glaciologists have described geothermal firn and ice caves in other volcanic craters (Kiver and Mumma, 1975; Kiver and Steele, 1975; Le Guern and others, 1999) and have sometimes discussed their origin. No one, however, has yet provided detailed observations of the evolution of such a system.

On Mount St. Helens, we have had a unique opportunity to study the interaction of geothermal energy with the accumulation of alpine snowpack from its inception after a major eruption. The International Glaciospeleological Survey (IGS) began investigative work in the crater in 1981 (Anderson and others, 1998). IGS is made up of fewer than 100 people, who are amateur to seasoned mountaineers and cave explorers. Professionally, they include a mix of scientists, engineers, and non-technical people. Yearly surveys began in 1982 with sketch mapping, description, and photography of cave passages, snow, firn, and ice. This investigation involved reconnaissance mapping and sampling from 1981 through 1998 by members of the IGS with the permission of the U.S. Forest Service and Mount St. Helens National Volcanic Monument.

CRATER SNOW, FIRN, AND ICE

A growing body of firn and ice mixed with rock debris, which we call the 'crater ice body', has accumulated in the crater of Mount St. Helens since 1982 (Figs. 1, 2, and 3). The shade of the steep crater walls to the east, south, and west protects this accumulation. The crater headwall rises to 2550 m (8365 ft) on the south (Fig. 4). The contiguous crater floor ice body extends from a maximum elevation of 2000 m (6560 ft) south of the Lava Dome, downward to the northeast and north around both sides of the dome. The crater floor north of the dome (1800 m or 5900 ft in elevation) hosts only seasonal snow accumulations.

The crater ice body is an incipient glacier that continues to grow. It is not readily apparent from a distance that glacier ice is present in the crater, because it is hidden by snow, firn, and

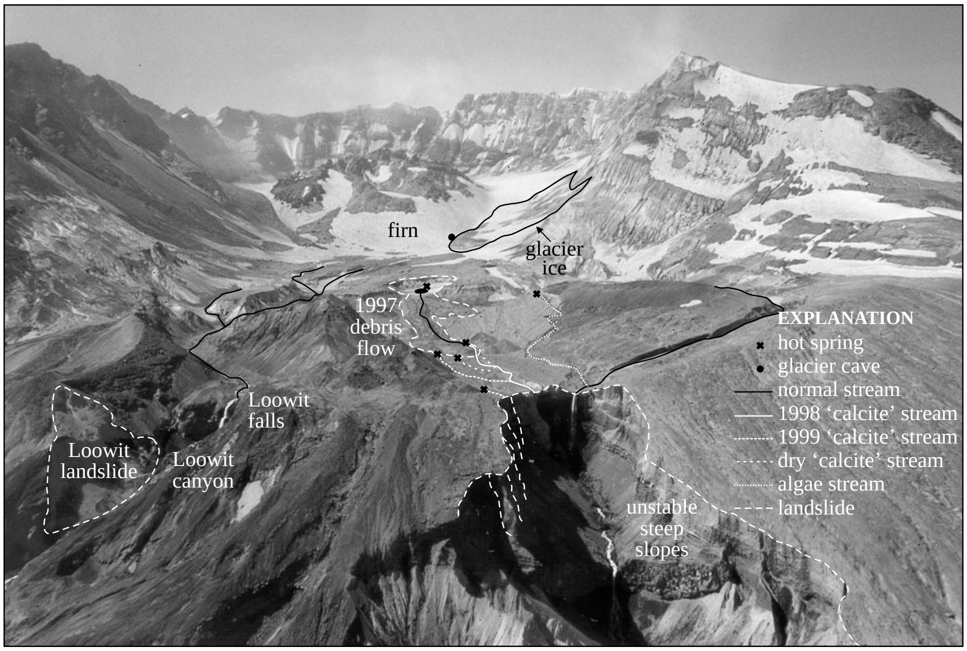

Figure 1. View into the crater from the north in the summer of 1999. Firn and glacier ice reach around both sides of the Lava Dome (center). One patch of ice is clearly visible on the right. Avalanche debris falls from the far crater walls onto the ice body and becomes incorporated into it. Debris flows have formed in the loose, unstable crater floor (center foreground). The August 1997 debris flow and its semicircular scarp are slightly to the left of center. This is the locality of the calcium-rich 'calcite' streams referred to in following figures. Loowit canyon, with its falls and landslide, is on the left side of the photo.

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crevasse – a deep, nearly vertical fissure formed in ice, firn, or snow caused by movement over an uneven surface. A crevasse suggests that movement is taking place (Sharp, 1960).

firn – a material that is transitional between snow and ice, being older and denser than snow but not yet transformed into glacier ice. Snow becomes firn after existing through one summer melt season; firn becomes glacier ice when its permeability to liquid water drops to zero.

glacier ice – a naturally accumulated ice that has reached a bulk density in excess of 0.82 g/cc. It possesses an intergrown crystalline matrix and flows plastically under its own weight.

ice, ice body – an accumulated body of firn and ice in the Mount St. Helens crater, regardless of its density, texture, or fraction of non-ice content (air and rock debris).

rock debris – rock fragments that have fallen from the crater walls after the eruption.

rock debris. Snows stacking higher each year have locally compressed the lower layers (visible in the caves) into dense, crystalline ice. Glacier development is suggested by crevasse formation and the banded texture of alternating higher- and lower-density ice caused by recrystallization under stress (Sharp, 1960). Small areas of ice visible on the south crater wall behind the Lava Dome also exhibit crevasses and flow texture, indicating that a new glacier is forming (Fig. 4).

The crater ice body shows signs of flow (crevasses) around both sides of the Lava Dome toward the north side of the crater. At least two large radial (relative to the crater center) crevasses are present in the ice body, adjacent to the Lava Dome on the east and west. Both crevasses penetrate to the lowest layers of the ice body. We first noticed the crevasse on the northwest side in September 1994, after the roof of an ice cave collapsed.

Figure 2. View from a helicopter looking south and down to the 1997 debris flow. The calcite-covered streambed is visible as a white streak leading down from the nose of one debris-flow tongue in the lower center right. Bacteria and calcite coatings occur in streams running alongside and extending below the debris-flow tongue. A calcite-covered streambed was buried by the east lobe of the debris flow. Warm springs rise near the end of the flow. *(Reproduced from videotape.)*

The crater ice body has been expanding since the winter of 1982–83 (D. A. Swanson, Hawaiian Volcano Observatory, oral commun., 1999). Its volume increased from approximately 28 million m^3 (37 million yd^3) of uncompacted snow and firn in 1988 (Mills and Keating, 1992) to more than 53 million m^3 (69 million yd^3) of snow and compacted firn and ice by 1995 (our estimate). As of late 1998, Anderson and Greninger estimated that the crater contained over 71 million m^3 (92 million yd^3) of snow, firn, and ice[1]. The thickness in places along the crater walls had reached as much as 140 m (460 ft).

Because of the limited quantity of bulk density data we collected for the crater ice body, the mean bulk density, and therefore the total mass of ice, can only be approximated. We measured the bulk density of ice at the base of a crevasse (Fig. 2) as 0.85 g/cc in September of 1994. We measured the bulk density of ice in the lowest cave passage as 0.86 g/cc in September of 1996. We obtained bulk density measurements by cutting samples with a cylindrical saw and weighing and measuring them in the field.

Our estimates and maps of the crater ice body are based on visual observations and local surface surveys. The areal distribution of snow and firn varies throughout each year. It is greatest in the spring when the winter snowfall first starts to melt. It decreases through the summer to a low in fall as winter snowfall returns. Our maps for 1997 and 1998 are *not* correlated to the same point in the season. Figure 3 is based on photos from September; Figure 4 is based on early season photos. The cave surveys date from September, but the snowpack areal limits do not. This doesn't make much difference near the dome, but is substantial on the crater wall.

CRATER FIRN CAVES

Bodies of firn and ice exposed to conditions above freezing tend to develop internal systems of water drainage. Flow of warm air subsequently expands these conduits, forming interconnected cave networks. The well-known ice caves of Mount Rainier occur in stagnant ice bodies such as Paradise Glacier (Anderson and others, 1994; Schmoe, 1926), the summit crater ice body (Kiver and Steele, 1975), and active glaciers such as the Carbon River Glacier (Halliday and Anderson, 1970).

The firn caves on Mount St. Helens are in the crater ice body next to the Lava Dome (Fig. 5). Cave passages form above fumaroles and fractures in and adjacent to the dome. The passages form a circumferential pattern around the dome, with their entrances on the dome flanks. Subglacial fumaroles and

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[1] Modified from Mills and Keating (1992). We used new thickness data collected by IGS members. Our computations also included the volume of rock debris derived from the crater walls, which was estimated from changes in our topographic maps (1988 base map of Mills and Keating) and more recent photographs. We also used altimeter readings at several locations around the Lava Dome and crater walls. We have not tried to keep a numerically defensible account of ice volumes as Mills and Keating did. They analyzed topographic maps that captured crater wall changes based on a series of aerial surveys. We have only been watching the changes in general snowpack level through the years and seasons near the dome perimeter, mainly because it directly affects our cave surveys. We used control points on the dome, placed by USGS personnel.

Our figures for snowpack volume are subject to the same correlation problem mentioned above. There are two reasons for this: (1) volume decreases, even if the mass remains the same, because of the steady metamorphism of snow to firn to glacier ice, increasing average density, and (2) melting reduces the mass (and therefore the volume even more) through a season. We used the same topographic basemap for each year. Therefore, our maps are not suitable for volumetric computations. Our maps are more records of our interpretation of the extent of the snowpack and the location of the caves.

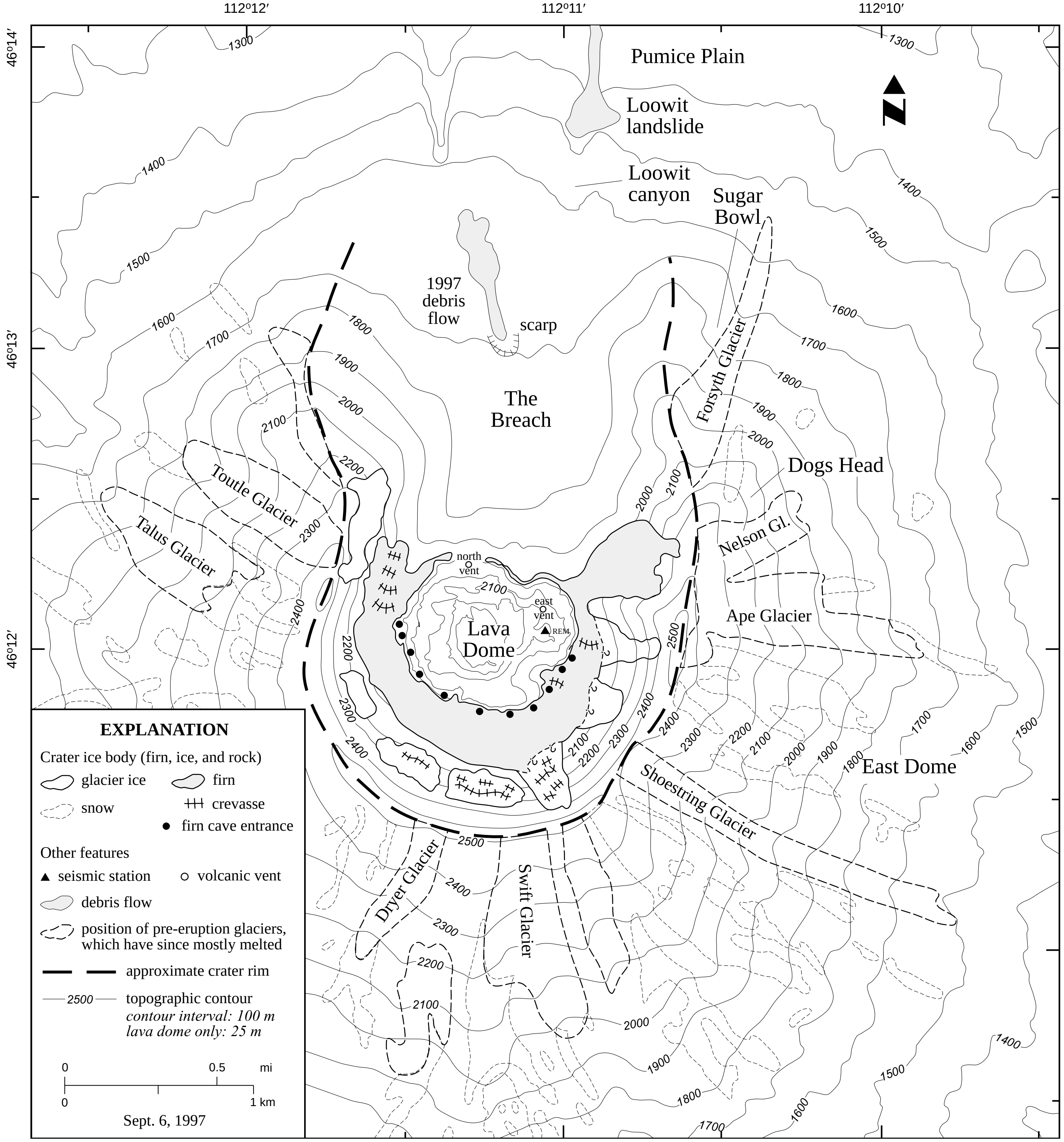

Figure 3. Sketch map on a simplified topographic base showing the cone of Mount St. Helens, the crater ice body (delineating firn and ice areas), the August 1997 debris flow, and other features in and around the crater. Note the Loowit landslide in Loowit canyon, which is located inside The Breach to the east of the debris flow (Fig. 1).

relatively warm air currents form and maintain the passages. To date, we have found more than 2415 m (7925 ft) of cave passages in the crater ice body.

The cave system is dynamic, responding to ice body growth and decay processes. Ablation, caused by outside air circulation, gradually enlarges cave passages. Basal melting of the whole ice body tends to diminish the caves. Increases in geothermal activity in the crater are expressed by the rapid enlargement of ‘steam cups’, dome-shaped melt pockets localized near fumaroles (Kiver and Steele, 1975). Air circulation converts these into the typical scalloped ceiling and wall forms seen in ice caves (Anderson and others, 1994) (Fig. 6). We believe the Mount St. Helens caves to be approximately in balance with the present geothermal heat release, because they have reached an overall stable morphology. Individual passages were observed to change over time, but the system as a whole remains much the same. Changes in the geothermal activity or climate would be expected to affect the dimensions and location of these caves, as well as ceiling, wall, and ablation features.

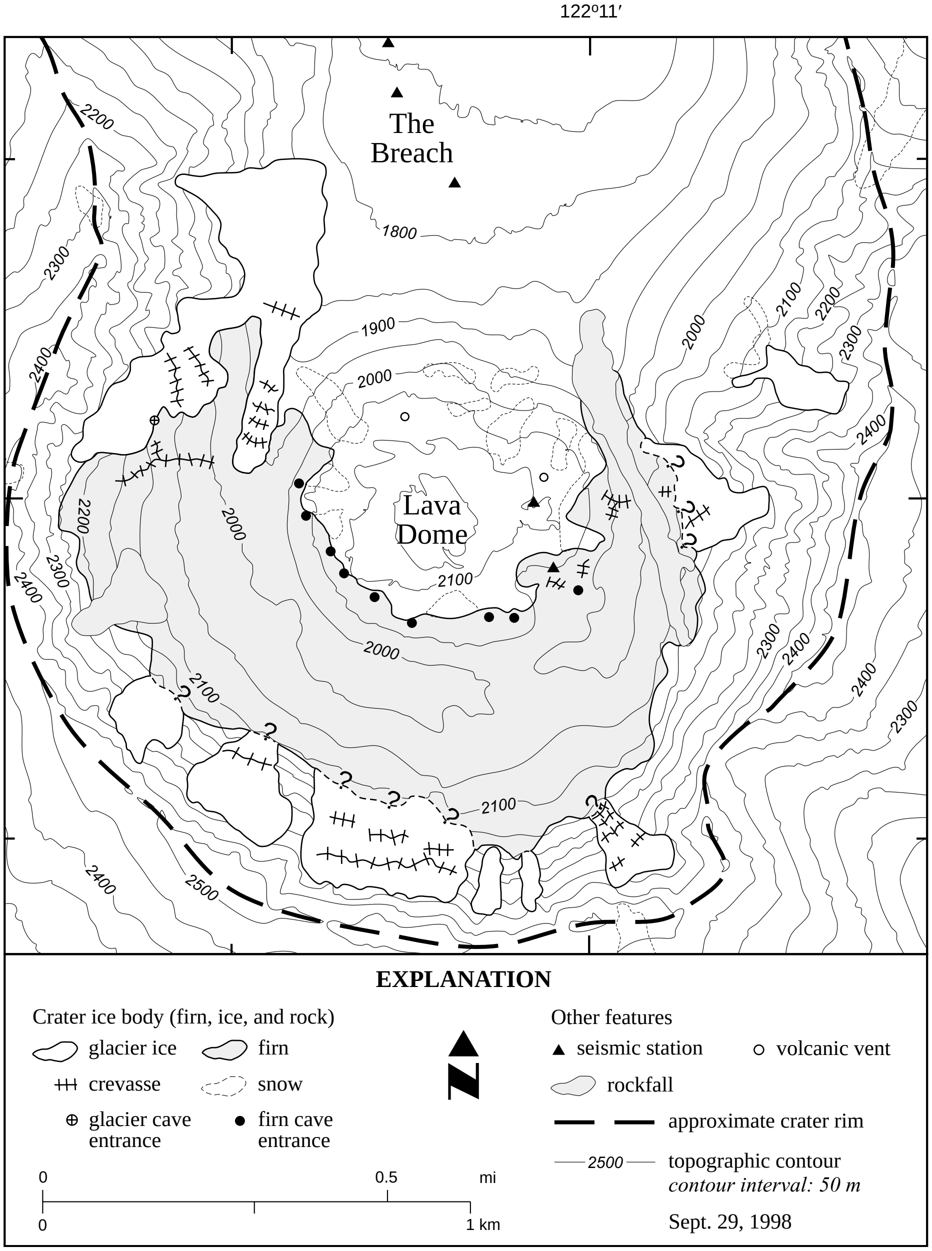

Figure 4. Sketch map on a topographic base showing the location of firn cave entrances, crevasses, rockfalls, and the surface extent of snow, firn, and glacier ice in the crater ice body on Mount St. Helens in September of 1998. The queried dotted lines between glacier ice and firn indicate that an ice front is probably concealed under the firn.

Cave Description

We mapped the Mount St. Helens caves by compass and steel tape survey. All gear was carried on foot. We recorded our observations on the surface and inside the caves with videotape and still camera. We visually estimated the physical dimensions of rooms and cave features.

We found entrances to and mapped 15 firn caves around the perimeter of the Lava Dome from 1996 through 1998 (Fig. 5). Some have spectacular large rooms. Most have small rooms and crawlways. Cave features include scalloped ceilings and walls (Fig. 6), moulins in the ceiling, multiple domes connected by crawlways, and skylights. In winter, short-lived ice stalactites, stalagmites, and helictites form inside the caves from water dripping from protrusions on the cave ceiling (Fig. 7). Cave floors are formed by the crater floor and, in places, the dome flanks. Room sizes range from 4.6 by 4.6 by 2.4 m (15 by 15 by 8 ft) high to 12 by 24 by 6 m (40 by 80 by 20 ft) high. Most caves occur in the presence of fumaroles. Other caves form adjacent to the dome where melt water undermines the ice body.

Six main entrances and numerous smaller ones lead down the 40-degree slope of the dome flank (Figs. 8 and 9). Passages paralleling the slope contours are surprisingly horizontal. Without geothermal control, passage patterns would be dendritic and follow the crater slope.

Descending passages have vertical sides and ceilings that are convex upward. Passages paralleling the slope contours are often shaped like right triangles with the 90-degree angle located at the junction of the downslope ice wall and the ice ceiling. Floors are composed of mud with up to boulder-size volcanic rubble and slope about 30 degrees. Against the Lava Dome flanks, the slope may exceed 40 degrees.

Ridge-like accumulations of rock debris from the Lava Dome form in many places on the floor of cave passages. They are composed of unsorted, unstratified mud and rock debris derived from the upslope portion of the cave floor. In some places, these ridges are in contact with the downslope ice wall and, in others, they occur toward the middle of the passage. The ridges probably started out as rock debris caught against the passage wall. Passage walls appear to retreat in response to the production of warm geothermal gas emanations. As the walls retreat, the ridges are stranded closer to the middle of the passage.

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crawlway – a cave passage that can be navigated only by crawling.

moulin – a circular, nearly vertical hole or shaft in the ice of a glacier, formed by percolating surface water and enhanced by air circulation.

skylight – an opening to outside light in the ceiling of a cave.

stalactite – a cylindrical or conical dripstone deposit that hangs from the ceiling of a cave.

stalagmite – a cylindrical or conical dripstone deposit that rises from the floor of a cave.

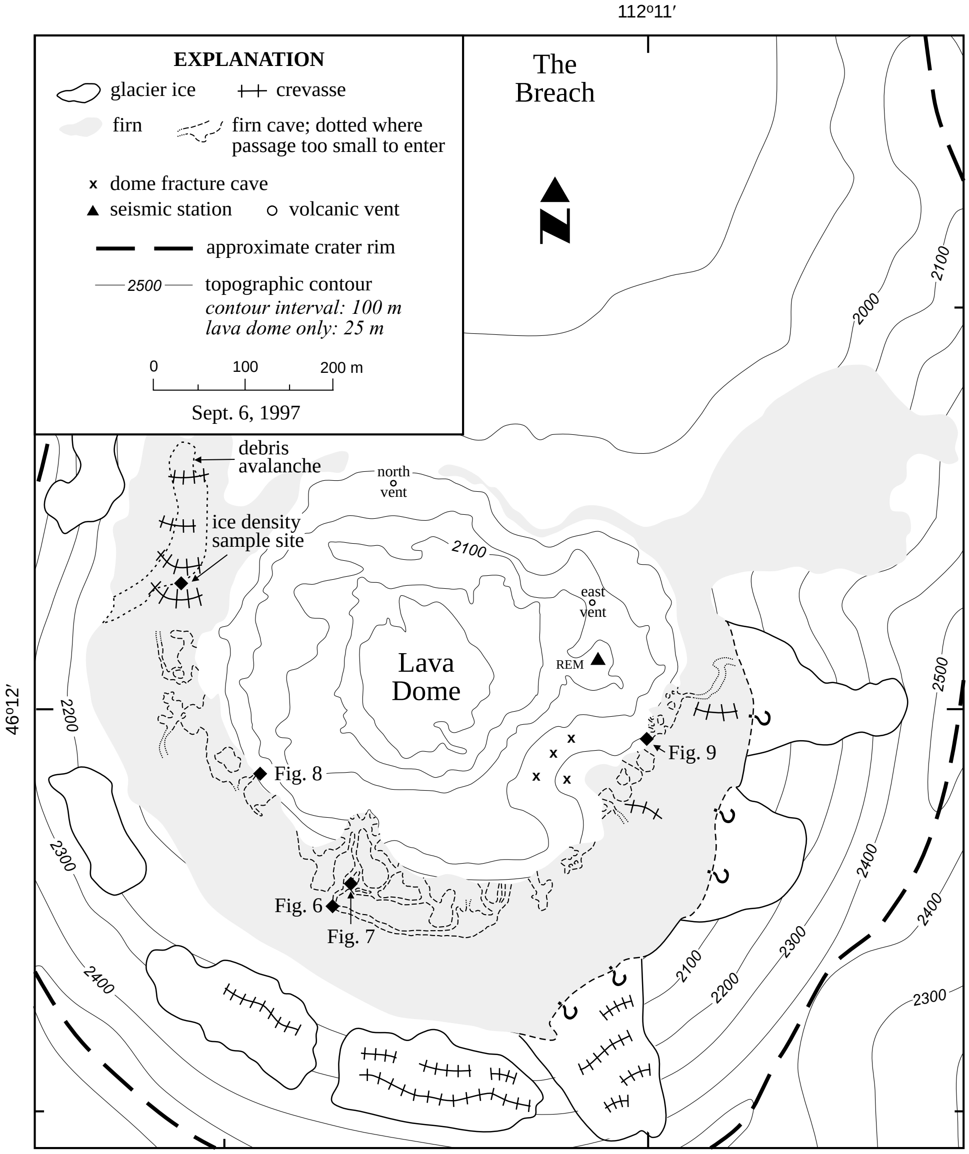


Figure 5. The Mount St. Helens crater firn cave system as mapped in 1997. This figure is based on tape and compass survey, which is inherently prone to distortion over long distances without an opportunity to close loops (as is the case with caves distributed around the Lava Dome). Passage shapes and relative sizes are fairly accurate here, but positions relative to the dome are not—we approximated the cave positions around the Lava Dome. The ice and rock debris avalanche shown is now covered by snow and ice. Diamonds indicate locations where photos and the ice-density sample were taken.

Progressive Recrystallization of Crater Ice

Generally in ice caves, older firn is distinguished from recrystallized recent snow by textural differences and stratigraphic relationships. Winter snowpacks from multiple years persist and provide the pressure increase necessary to convert snowfall into a permanent ice body. As recrystallization continues, individual ice crystals in the deepest layers grow together to form a rigid fabric with limited permeability (glacier ice).

From 1986 to the present, we observed the gradual change from snow to firn to glacier ice in cave passages (ice bulk densities were not measured systematically). An abrupt decrease in percolating water occurred in the final stage of the transition.

An incipient glacier has developed and grown on the Mount St. Helens crater floor. Through the heavy winter snowfalls and mild summers of the 1980s and 1990s, a continued sequence of yearly net snow accumulation enabled the ice body to persist.

Geothermal Activity in the Caves

The Mount St. Helens Lava Dome is the locus of the active volcanic vent and a source of volcanic gas emanations. The caves are primarily a result of the concentration of heat. They are localized at active fumaroles and form as conduits of venting for the heated gases. They are further enhanced by the drainage of heated surface water from the dome directly into the ice body.

Hundreds of small fumaroles emit considerable quantities of steam that frequently impair visibility in the firn caves and make mapping, photography, and other observations difficult. Some of these fumaroles make audible hissing and gurgling noises. Although the rising heat and steam cause the ice walls and ceilings to drip constantly, we have not observed appreciable quantities of standing or flowing water in the caves, perhaps because the permeability of the crater floor allows seepage. Changes in passage dimensions and location (from periodic observations and resurveys of the caves) indicate changes in heat-flow and the location of volcanic emanations.

Sulfurous fumes occur locally in the caves. Gases from the numerous fumaroles and circulating surface air mix throughout the cave passages. The presence of breathable air in the known cave system indicates that volcanic gases are rapidly mixed with fresh air and removed from the caves. Earlier workers occasionally observed minor carbon dioxide accumulations (D. A. Swanson, Hawaiian Volcano Observatory, oral commun., 1999). Although we have not come across any passages with bad air, we carry portable hydrogen sulfide and carbon monoxide detectors as a routine safety precaution.

Cave Ablation

Within the caves, evaporation, sublimation, and heat conduction are the major ablative processes (Anderson and others, 1994). Since the caves are sheltered from sunlight, radiation from the sun has no direct influence on cave ablation, but energy from heated ground and fumaroles has an appreciable effect. The main control of cave ablation is the amount of air flow against the cave walls. In cave networks possessing substantial vertical relief, trunk passages tend to form as major meltwater conduits and remain dominant because air circulation is enhanced by convection.

As cave ablation and surface ablation continue through a summer season, it is normal for the cave ceiling to approach and intersect the ice surface progressively over time. If the ice is fractured, or perhaps after winter snow adds weight to the ceiling, a cave passage may experience ceiling failure. In either case, the cave system suddenly gains a vent to outside air. The effect of venting in summer is to allow cold cave air out and warm outside air in. The effect in winter is reversed. The importance of ablation vents is exaggerated when there is any superimposed restriction in the system, such as winter snow or a rockfall blocking other entrances. In this case, the vent entrance becomes the major means of communication with outside air. When all vents to the surface are closed, the ordinary glacier cave becomes dormant. In a cave that has internal heat sources, the ablation process can continue by convection, even when all external openings are blocked. This type of system is therefore less seasonally dependent and may evolve faster than an ordinary glacier cave.

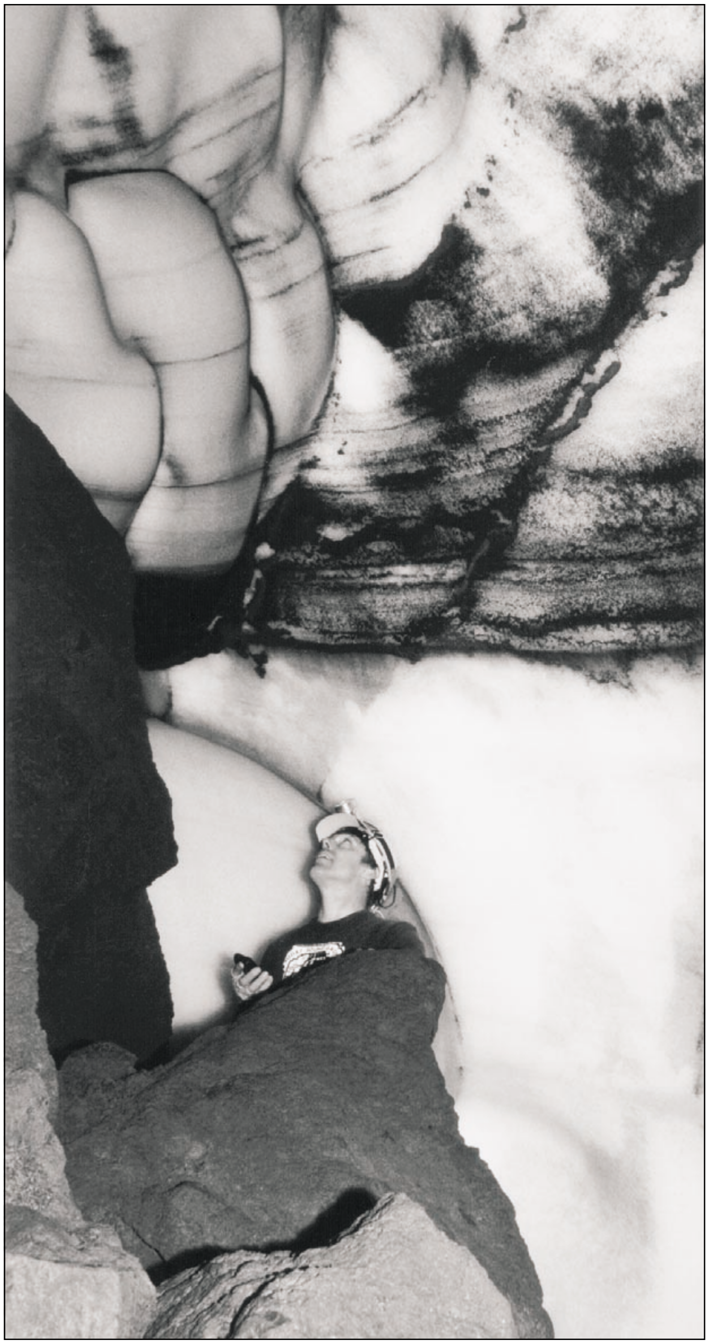


Figure 6. A typical cave passage in Mount St. Helens crater firn adjacent to the flanks of the Lava Dome (Fig. 5). Dacite boulder debris forms talus at the angle of repose, about 30 degrees. The scalloped ceiling and walls continually drip cold water during the summer, but ice stalactites form at these points during the winter. Bill Greninger, IGS team member, is looking up at scallops on the cave walls. Photo taken July 28,1997.

Figure 7. Ice stalagmite in the lowest cave passage (Fig. 5).

FAUNA OF THE CRATER AND CRATER CAVES

There is little direct evidence of animals inhabiting the crater floor, with one exception—mice were reported on the crater floor north of the dome in 1982 (D. A. Swanson, Hawaiian Volcano Observatory, written commun., 1999). Deer have visited the lower part of the crater on occasion, leaving only tracks for the careful observer to notice. We have seen insects, including honeybees, ladybird beetles, and carpenter ants, in the crater environs, presumably blown in by winds. We also found a mountain beaver skull, probably left by a predatory bird. Fauna observed during ice cave exploration include insects and ice worms that are presently inhabiting the cave and snowfield environment. Similar species are known from ice caves at Mount Rainier (Anderson and Halliday, 1969; Anderson and others, 1994).

Biologists have long sought the primitive, cold-adapted beetles of genus *Grylloblatta* in the glaciers and craters of Mount Rainier, Mount Baker, Mount Hood, and Mount St. Helens. We observed an unidentified species of *Grylloblatta* in September of 1997 on the ice surface on the northwest side of the Lava Dome. Grylloblattids are also known from the Paradise and Stevens glacier caves of Mount Rainier (Halliday and Anderson, 1970).

Mountain climbers have observed ice worms (Oligochaeta: Plesiopora Enchytraeidae) of the species *Mesenchytraeus solifugus rainierensis* in snowfields of several Cascade mountains, especially Mount Rainier (Rod Crawford, Burke Museum, oral commun., 1998). In August of 1996, we collected a living specimen from approximately 1 cm (0.4 in.) beneath the surface of an ice wall in the largest of the Mount St. Helens firn caves. These worms are thought to migrate through the ice in a diurnal cycle, taking advantage of pore spaces between ice crystals to move about.

We collected nymph and adult stoneflies (Plecoptera: Perlodidae) of the species *Rickera sorpta* on the surface of the ice body and in cave interiors (Rod Crawford, Burke Museum, oral commun., 1998), which are also found in the Paradise and Stevens Glacier caves of Mount Rainier (Anderson and others, 1994). Stonefly nymphs are aquatic. The near-mature state of specimens collected at Mount St. Helens indicates that they had crawled out of water for the molt to adulthood. The dark coloration of nymphs makes them almost invisible against the dark bottom of a cave pool. Nymphs are extremely sensitive to warmth—one collected specimen expired after approximately fifteen seconds of exposure to human body heat.

Figure 8. A typical cave entrance adjacent to the Lava Dome, looking in. IGS team members are climbing down the flank of the Lava Dome to the entrance to the lowest passage that is parallel to the slope contours. Photo taken Oct. 5, 1997.

Figure 9. The northwesternmost cave entrance as seen from inside the cave (Fig. 5). An IGS member climbs with the Lava Dome in the background. Photo taken September 1997.

GEOMORPHIC CONDITIONS IN THE CRATER

Crater Floor Environment

The present crater floor is underlain by loose, porous, and permeable debris from the landslide caused by the collapse of the upper third of the volcano during the 1980 eruption. The bulk of the debris avalanche flowed downward and to the north, filling in parts of the Spirit Lake basin and upper valley of the North Fork Toutle River. Subsequent eruptions, including the later part of the May 1980 eruption, covered the landslide surface with juvenile pumice and tephra deposits, smoothing the landslide topography and creating what is known today as the Pumice Plain. The first lava domes formed at the top of the volcanic conduit were wholly or partially destroyed by explosions (Holcomb and Colony, 1995). After the October 1980 eruption, dome growth gradually covered the fringe areas of crater-filling rockfall talus cones (Mills, 1992). These cones are intercalated with accumulating snow. The whole body was insulated and compacted by its own mass. Later tephra eruptions have added only minor amounts to the sediment pile. The most volumetrically significant addition to the post-1986 crater floor environment, therefore, is accumulated ice and rock debris. Through 1988, the rock debris fraction of post-1980 crater fill gradually dropped from 100 percent to about 65 percent of the total (Mills, 1992).

The most active surface processes taking place in the crater are (1) continued landslides from the steep crater walls, (2) fluvial downcutting in the stream courses that have established themselves across the crater floor, and (3) debris flows developing from slope failure on the north crater floor. Perhaps the most significant subsurface process acting on the crater floor contents is percolation of meteoric water and consequent alteration and leaching of the volcanic minerals.

Several small surface streams flow intermittently from the crater ice body. Snowmelt and rain percolating through fractures in the Lava Dome and through the permeable crater fill, rise in geothermal springs that feed the crater streams.

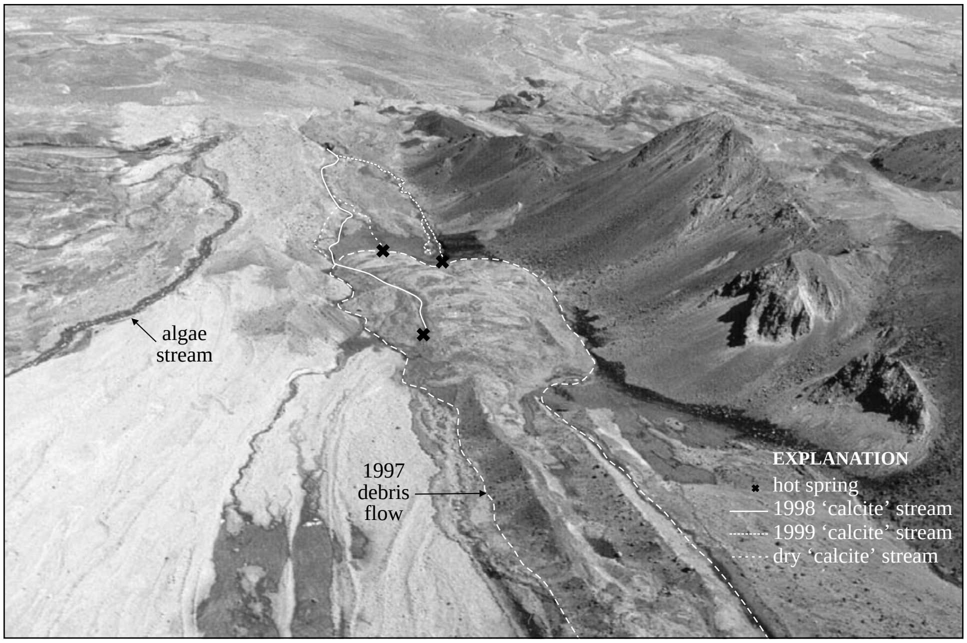
Degradation of the Crater Floor

Nearly two decades of precipitation and runoff have eroded and leached material from the thick, unconsolidated mass of volcanic debris on the crater floor. Streams draining the crater have cut through this material and formed steep-walled canyons with unstable slopes (for example, Loowit canyon on the northeast flank of the crater, Fig. 1; Shevenell and Goff, 1995). These canyons are too dangerous to be used as conduits for crater access (Anderson and others, 1998). Workers in the crater have observed repeated slope failures and small slides.

In the spring of 1997, an ice and rock debris avalanche from the crater walls formed a tongue about 25 m (83 ft) in height, 150 m (500 ft) in length, and 15 m (50 ft) in width on the ice surface near the southwest side of the dome (Fig. 5). We estimated that the deposit was about 40 percent rock debris. The tongue froze and lasted through the summer of 1997. Dark-colored rock debris around the tongue speeded surface ablation of the ice with heat collected from solar radiation.

In August of 1997, a debris flow was triggered by the failure of a mass of saturated crater-floor material at The Breach (Figs. 1, 2, 3, and 10). The semicircular, steep-walled scarp was originally 20 m (65 ft) deep and about 150 m (500 ft) wide. The deposit extended about 700 m (2300 ft) downslope, from the 1700 m (5580 ft) elevation at the scarp brink to 1550 m (5085 ft) at the lowest point. The scarp cuts across the bed of a geothermal stream that now rises from the scarp floor and feeds clear heated pools that appear to be free of living matter. Two streams exit the scarp mouth and flow through the debris flow deposit in recently excavated gullies. One tongue of the debris flow followed the original stream and filled that stream course. Post-debris flow seepage was diverted around and through the deposit, producing additional springs and seeps throughout its length. We measured water temperatures as high as 80°C (175°F) in pools in the scarp and temperatures of 50°C (120°F) or greater downstream of the debris flow.

Another slide occurred in September 1997 in the east part of The Breach, passing down the Loowit drainage (Figs. 1 and 3). Water-saturated loose volcanic material collapsed to form a lahar that roared out of the crater and reached past Loowit Trail below on the Pumice Plain. The trail was temporarily closed for rebuilding after the slide, and only recently reopened. Similar slides must be expected in the future from the over-



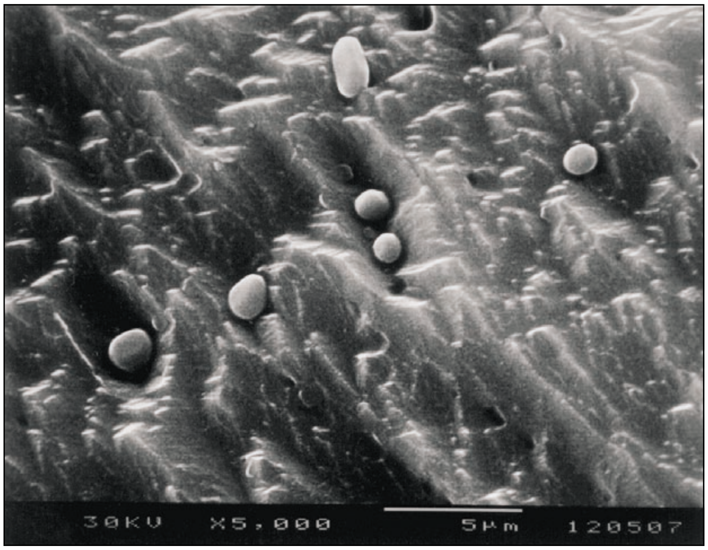


Figure 10. *(top)* Deposits from the August 1997 debris flow in The Breach (Fig. 3). This view is looking downslope (north) along the path of the debris flow, the reverse of that shown in Figure 1.

Figure 11. *(middle)* Samples of calcite deposits formed in a calcium-rich geothermal stream below the debris flow tongues on the crater floor of Mount St. Helens. These samples were taken from stalactites that form on rock projections in the stream. The water temperature of the stream at the sample site was 49°C (120°F).

Figure 12. *(bottom)* SEM photomicrograph of calcite encrusting bacterial strands in a sample taken from a geothermal streambed. The rounded objects are sulfur bacteria. Note rhombohedral crystal terminations. Photo courtesy of Robert Folk, University of Texas at Austin.

steepened canyon walls of The Breach area. (Note steep, unstable walls in Fig. 1).

Calcite and Bacterial Growth in Geothermal Streams

Calcite ($CaCO_3$) is actively precipitating from solution in the stream water that rises from the scarp floor mentioned above. It has formed deposits of travertine and tufa as flowstone, dripstone, helictites (cored by bacterial filament aggregates), and cave pearls. These coatings have formed on the streambed and hang from steps and waterfalls. Samples of the calcite coating (Fig. 11) exhibit compact, fan-shaped aggregates of acicular (needlelike) to bladed crystals as much as 1 mm in cross section. These appear to be pseudomorphs after aragonite bundles. Figure 12 is a scanning electron microscope (SEM) photomicrograph of a flowstone surface from a waterfall overhang. We previously (1996 and 1997) observed and filmed calcite growth in thermal streambeds now covered by the debris flow.

Water percolating through freshly exposed loose material in the debris flow supplies nutrients and mineral components to the streams. Red (sulfur), orange (iron), and minor green (chlorophyllic) bacterial slime coats the streambed and accumulates in streambed pockets (Folk, 1993). We observed (summer 1998) flourishing bacterial growths in the presence of abundant water seeping from gully walls. Downstream of the debris flow for about 0.5 km, heavy coatings of calcite had grown on streambed rocks and encapsulated bacterial growths. These encrustations actively grow in flowing water and in the splash zone along the stream banks. Helictites grow as thin calcite coatings on strands of red bacteria that hang from rocks in the streambed. Calcite coatings continue to grow on and engulf the bacterial colonies (Fig. 13A,B). Remains of the bacterial growths can be found inside hollow flowstone crusts. SEM microscopy indicates the presence of bacteria and

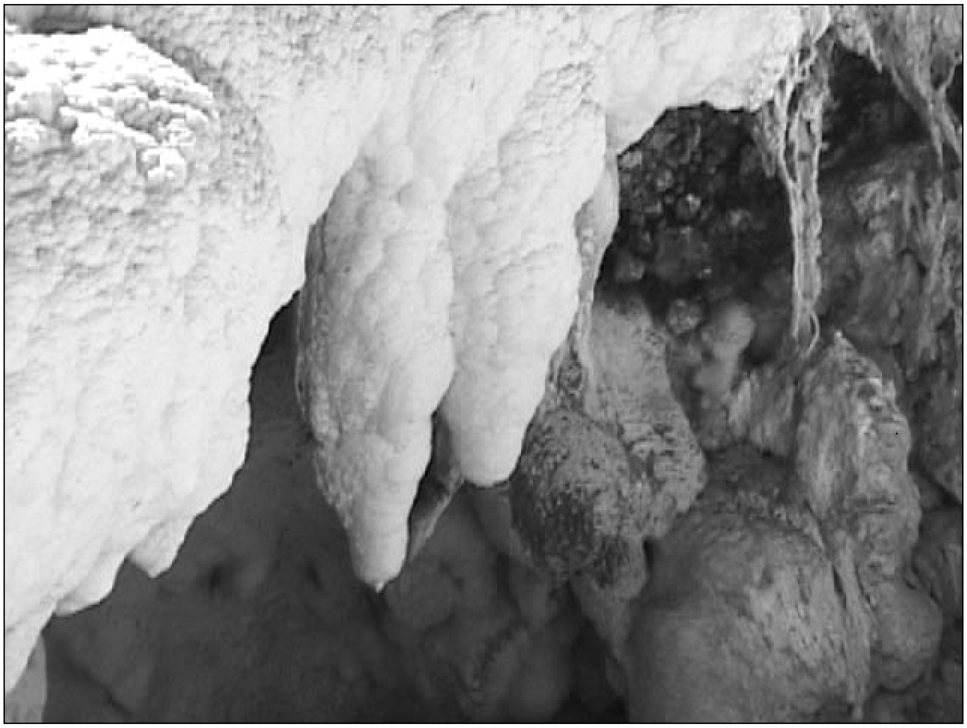


Figure 13A. Travertine dripstone growths at a streambed overhang. These calcite growths continue to expand forming travertine stalactites. *(Reproduced from videotape.)*

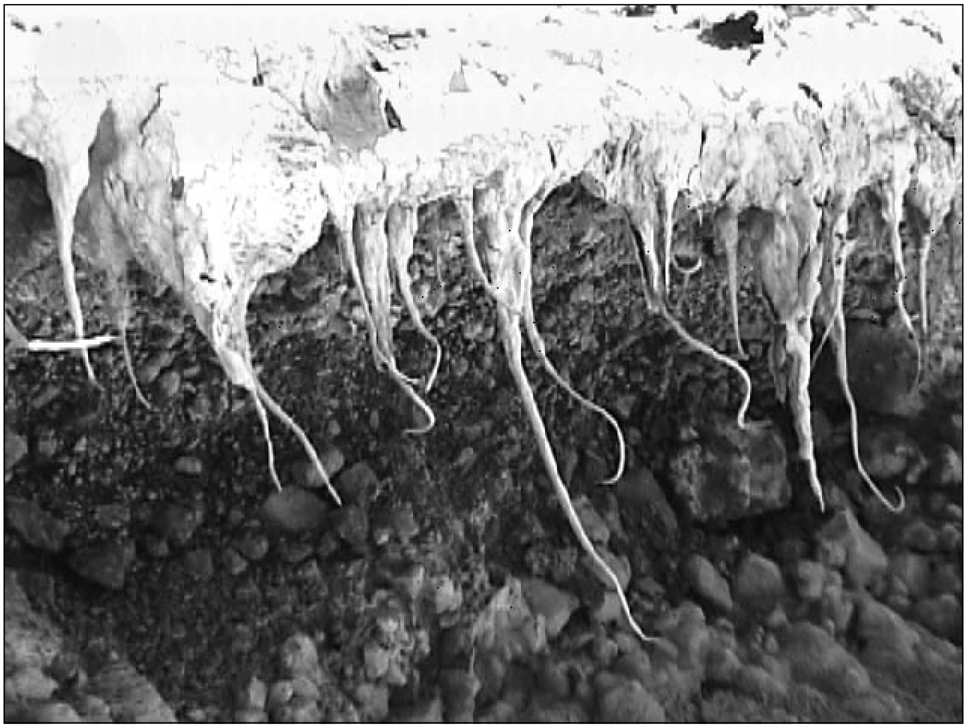


Figure 13B. Calcite coatings growing on and engulfing red sulfur bacteria strands. *(Reproduced from videotape.)*

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cave pearl – an unattached, subspherical to spherical calcite concretion formed in splashing or dripping water, usually deposited on a sand particle or rock fragment nucleus.

dripstone, flowstone – mineral coatings (usually calcite, but may be other minerals or ice) deposited by precipitation from water flowing over an exposed surface, usually found in caves. The distinction indicates the nature of water flow during growth: dripstone forms free-hanging or free-standing deposits; flowstone forms as a wall or floor coating.

helictite – a curved, angular, or dendritic twig-like growth from a flowstone or dripstone surface.

pseudomorph – a mineral whose outward crystal form is that of another mineral species from which it has been changed by alteration, substitution, or some other process.

travertine, tufa – a dense, finely crystalline massive or concretionary limestone of white, tan, or cream color, commonly having a fibrous or concentric structure and splintery fracture; formed by rapid chemical precipitation of calcium carbonate from solution in surface or ground water, as by agitation of stream water or by evaporation. The spongy or less compact variety is called tufa.

nannobacteria, similar to those described by Folk (1993), in the growths. Only incipient, very thin calcite coatings grew in the scarp pools and streams leading out of the scarp mouth.

Calcite deposition in streams of The Breach area has been rapid and continuous (Fig. 14). From September 1997 through August 1998, at least two episodes of calcite deposition took place in a gully cut into the 1997 debris flow deposits. Older calcite-coated terraces are preserved on the walls of the gully 1 to 2 m above the present calcite-coated streambed, indicating that the newest coatings developed after the most recent gully-deepening erosion. Within a one-year period, calcite stalactites and stalagmites (Fig. 15) grew to a maximum size of 27 cm (11 in.) in diameter and 30 cm (12 in.) in length, and calcite cave pearls grew to 3.6 cm (1.4 in.) in diameter.

Figure 14. Pre-August-1997 geothermal 'calcite' stream issuing from Mount St. Helens crater. The white calcite coating highlights the streambed. This view is upstream of the white streambed visible in Fig. 2. This section of the stream was buried by the August 1997 debris flow.

We believe the supply of calcium to thermal streams derives from the leaching of fresh, porous dacite in the crater by percolating meteoric water. The chief process affecting the chemistry of crater runoff has evolved from degassing of newly injected magma (waning to insignificance about 1985) to passage of meteoric water through the crater floor deposits in a manner too fast to attain equilibrium (Shevenell and Goff, 1995). Such undersaturated ground-water conditions could leach mobile components from a large volume of crater deposits. High rainfall produces a high flux of water through the dome area and out the crater mouth. Heated ground water resurges where the local unconfined water table intersects the crater floor. Farther downstream, calcite precipitates in the rapidly cooling surface streams. Nutrients derived from decomposition of volcanic material appear to support the bacterial population of crater streams.

The presence of red sulfur bacteria indicates that sulfur is an active component in the aqueous chemistry of the crater environment and a prominent source of acidity in the water that acts to digest crater rocks. Elemental sulfur from magmatic emanations interacts with oxygen-rich meteoric water to produce an acidic ground-water system in the dome area. At the hot springs (where nothing is growing and no calcite is present), the pH is about 6.5 and the temperature is 55°C (130°F). At the sampling locations where red bacterial colonies are in contact with actively growing calcite deposits, the water is somewhat alkaline (pH about 8) and the temperature is about 35° to 40°C (95–105°F).

CONCLUSIONS

A growing body of firn and ice mixed with rock debris, which we call the 'crater ice body', has accumulated in the crater of Mount St. Helens since 1982. Its mean bulk density is increasing with each passing year, and the transition from snow to firn to glacier ice (with active crevasses) is presently taking place. Net ice mass budget balances have been positive in the crater since 1986, when the snowpack was first recognized to be growing.

Figure 15. The geothermal stream is coated by a thick mantle of travertine up to 15 cm in thickness. This entire growth of this sample occurred during a single summer season. The tape is graduated in inches.

Ice caves form above fumaroles that are located along fractures in the Lava Dome and the surrounding crater floor. Cave passages are gradually enlarged by ablation caused by geothermal sources beneath the ice and by outside air circulation. Passages grow laterally and vertically toward the surface, leading to ceiling collapse. The network of fumaroles has produced a ring of relatively horizontal passages that are connected to the surface by a number of ascending entrance passages.

Changes in geothermal activity in the crater of Mount St. Helens have become noticeable through cave passage observation and remapping. Calcite precipitated from geothermal streams on the crater floor produces coatings as thick as 15 cm (6 in.) thick in a single year. Chlorophyllic and later sulfur and iron bacteria are associated with these streams. In the summer of 1997, a small debris flow developed in the crater north of the Lava Dome, and later the same year, another flow occurred in Loowit canyon.

Increased thermal activity could mobilize crater ice to produce debris flows that could affect the discharge and sediment load in Toutle River. Our mapping and investigations of the crater environment could furnish additional indicators of geothermal activity and incipient geomorphic changes that could augment information provided by remote surveys.

ACCESS TO THE CRATER

The crater of Mount St. Helens can be a dangerous place, particularly because of snow and rock avalanches. Other hazards include invisible snow caves and unstable slopes. The potential also exists for pockets of 'dead' (oxygen-depleted) air and unexpected explosions and discharges of volcanic ash. The U.S. Forest Service strictly regulates access to the crater of Mount St. Helens. The area is part of the Mount St. Helens National Volcanic Monument, and special permits are required for any activities other than visitation of public facilities. The authors have a crater access permit for the purpose of scientific study. No one should attempt to approach Mount St. Helens by foot or by air without written clearance from the Forest Service.

ACKNOWLEDGMENTS

The authors are grateful to International Glaciospeleological Survey members for assistance with mapping and to staff of the U.S. Forest Service at Mount St. Helens National Volcanic Monument for logistical assistance and advice in conjunction with permits and crater entry. From the Washington Division of Geology and Earth Resources, we thank Wendy Gerstel and Patrick Pringle for critical review and Jari Roloff for editorial and graphic assistance in preparing this paper. Robert Folk of the University of Texas at Austin, Texas, provided SEM photos and identified bacterial components. Rod Crawford of Burke Museum, University of Washington, identified our insect and worm specimens.

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The latest information on Mount St. Helens is reported at the U.S. Geological Survey's Cascades Volcano Observatory website at http://vulcan.wr.usgs.gov.

Augite Crystals from Doty Hills, Lewis County, Washington

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Mineral collectors in southwest Washington have been aware of the L-1000 Road augite crystal locality in the Doty Hills, Lewis County, Washington, for many years. The locality is an excellent example of an augite-plagioclase porphyry lithic tuff. Augite crystals are abundant and provide a unique opportunity to the rockhound and mineralogist to collect and study these wonderful crystals in Washington State.

Norman Johnson of the Washington Agate and Mineral Society, Olympia, donated a jar full of loose crystals to the Washington Division of Geology and Earth Resources (DGER) back in 1961. Twenty years later, the logging road cut had sloughed in and become overgrown with thick alder trees, making the locality unrecognizable. Then, in 1996, a landslide a few hundred yards west of the original augite locality destroyed the road. The resulting reconstruction and new road cuts exposed unweathered rocks full of augite crystals (Fig. 1).

Figure 1. Augite-plagioclase porphyry lithic tuff outcrop.

Location

The augite locality is a series of road cuts along logging road L-1000 in NE1/4, sec. 15, T14N, R5W. Road L-1000 can be accessed from the paved Lincoln Creek Road. The gate to road L-1000 is generally open as it is the main road to lands managed by the Department of Natural Resources (Fig. 2). For more detailed information on how to get to the locality from Exit 77 of Interstate 5 and State Route 6, see the U.S. Geological Survey (USGS) Doty and Rainbow Falls 7.5-minute quadrangles.

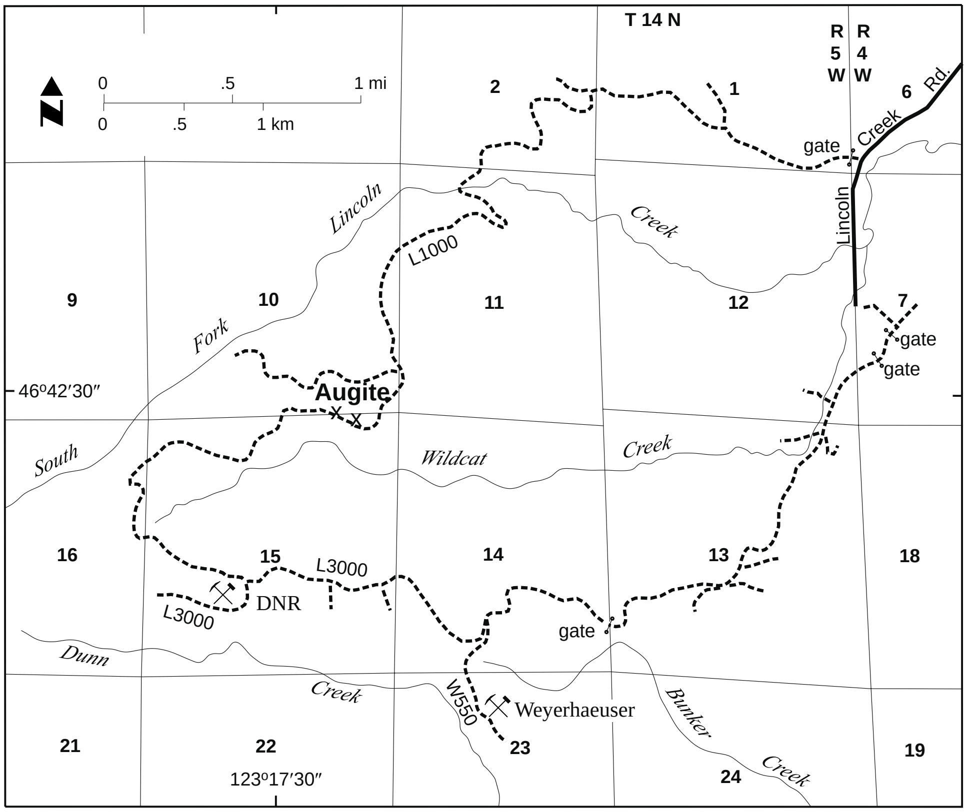

Figure 2. Sketch map of logging road network showing location of augite crystals and nearby rock pits.

Geology

The augite crystals are scattered through a greenish black augite-plagioclase porphyry lithic tuff. The felsic lithic fragments are completely overgrown by green Mg-chlorite. Thin sections show the matrix to consist of brown glass filled with radial growth

Figure 3. Augite crystals from Road L-1000 cut, NE 1/4, sec. 15, T14N, R5W, Doty Hills, Lewis County, Wash.

Figure 4. Augite crystals from logging road, center sec. 10, T1N, R8W, south slope Cedar Butte, Tillamook County, Ore.

crystallites. Pale green prehnite-pumpellyite replaces some of the lithic fragments, and late calcite lined fractures cut the rock (J. Dragovich, Washington Division of Geology and Earth Resources, written commun., 1997).

Whole-rock and trace element content of the porphyritic lithic tuff is shown in Table 1. The x-ray fluorescence spectroscopy (XRF) analysis was performed by Diane Johnson at the Washington State University GeoAnalytical Lab in 1997. The whole-rock composition plots close to that of augite (Anthony and others, 1995).

The massive, poorly sorted, porphyritic lithic tuff has a northerly strike, dips at 20 degrees to the east, and is a distinct unit near the base of the McIntosh Formation as mapped by Pease and Hoover (1957). Abundant foraminifera from a sample of basaltic mudstone along strike from the porphyritic lithic tuff gave an age of late Middle Eocene, Narizian Stage (K. McDougall, USGS, written commun., 1997). The fossiliferous sample was collected above a thick basalt unit exposed in Weyerhaeuser's quarry in sec. 23 (Fig. 2).

A very similar geologic unit is exposed on Cedar Butte in the Tillamook Highlands, Coast Range, Tillamook County, Oregon. There, an augite crystal-vitric tuff overlies palagonitic pillow breccia and lensoidal masses of basalt (Nelson and Shearer, 1969). It consists of submarine basalt tuff with abundant euhedral augite and (or) plagioclase crystals and is part of the Tillamook Volcanics of late Middle Eocene age, Narizian Stage (Wells and others, 1994). My personal observation is that this unit of the Tillamook Volcanics is indistinguishable from the porphyritic lithic tuff exposed along Road L-1000 in Doty Hills, Lewis County, Washington.

Mineralogy

The name *augite* is derived from the Greek *auge*, meaning sunlight, referring to the luster of augite along cleavage planes. Augite is a silicate (double chain structure) and one of the pyroxene group of minerals. It is usually black, greenish black, or dark green in color and occurs as an essential constituent in many basic igneous and some metamorphic rocks. The chemical composition of augite is $(Ca,Na)(Mg,Fe,Al,Ti)(Si,Al)_2O_6$.

In the Doty Hills, the augite occurs as perfectly formed euhedral phenocrysts up to 1.8 cm long and 1.3 cm wide (Fig. 3). Augite belongs to the monoclinic crystal system. For similar augite crystals collected from Cedar Butte, Oregon, see Figure 4. Many of the augite crystals show beautiful twinning. The morphology is nicely illustrated using SHAPE software from crystals collected by R. Peter Richards (Fig. 5).

Acknowledgments

Thanks go to Ray Wells of the USGS for information on the Tillamook Volcanics, to R. Peter Richards of Morphogenesis, Inc., for days in the field and crystal drawings, and to Division staff Joe Dragovich for thin-section work, Tim Walsh for mineral photography, and Keith Ikerd for the location map.

Table 1. Whole-rock and trace element XRF analysis of augite-plagioclase porphyry lithic tuff, Road L-1000 cut, Doty Hills, Lewis County, WA. Note: major elements are normalized on a volatile-free basis, with total Fe expressed as FeO

Normalized results (weight %)		Trace elements (ppm)			
SiO_2	47.79	Ni	733	Cu	100
Al_2O_3	7.61	Cr	1471	Zn	80
TiO_2	1.663	Sc	29	Pb	0
FeO	11.68	V	228	La	8
MnO	0.175	Ba	104	Ce	20
CaO	9.97	Rb	4	Th	2
MgO	20.81	Sr	125		
K_2O	0.05	Zr	94		
Na_2O	0.06	Y	17		
P_2O_5	0.182	Nb	17.4		
Total	99.99	Ga	14		

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porphyry – an igneous rock of any composition that contains conspicuous large crystals (phenocrysts) in a fine-grained matrix. The rock name descriptive of the matrix composition usually precedes the term, for example, augite-plagioclase porphyry.

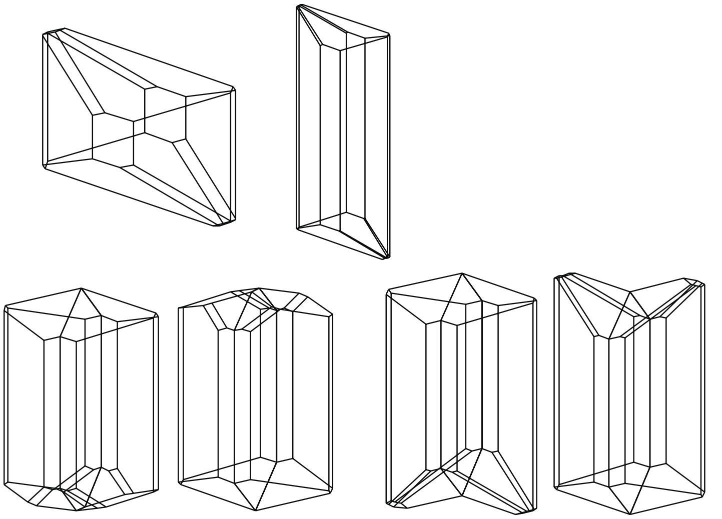
crystallite – a minute body of unknown mineralogic composition or crystal form that does not polarize light. Crystallites represent the initial stage of crystallization of a magma or glass.

radial growth – a growth pattern in which crystal blades grow out from a point, forming a ball shape.

thin section – a piece of rock or mineral mechanically ground thin enough to be transparent or translucent and then mounted between glass slides for microscopic viewing.

lithic tuff – a hardened deposit of volcanic ash that contains abundant fragments composed of previously formed rocks (lithic fragments), for example, accidental particles of sedimentary rock, accessory pieces of earlier lavas in the same cone, or small bits of new lava that first solidify in the vent and then are blown out.

Figure 5. Augite crystal morphology from Road L-1000 cut, Doty Hills. Top row: two untwinned crystals showing the range of habits, looking obliquely onto Miller Indicies {010}; bottom row, two crystals twinned on {100}, views with each end facing up, showing the variations in habit at the notched end of the twinned crystal; R. Peter Richards specimens (by permission of R. Peter Richards, Morphogenesis, Inc., 154 Morgan St., Oberlin, OH 44074).



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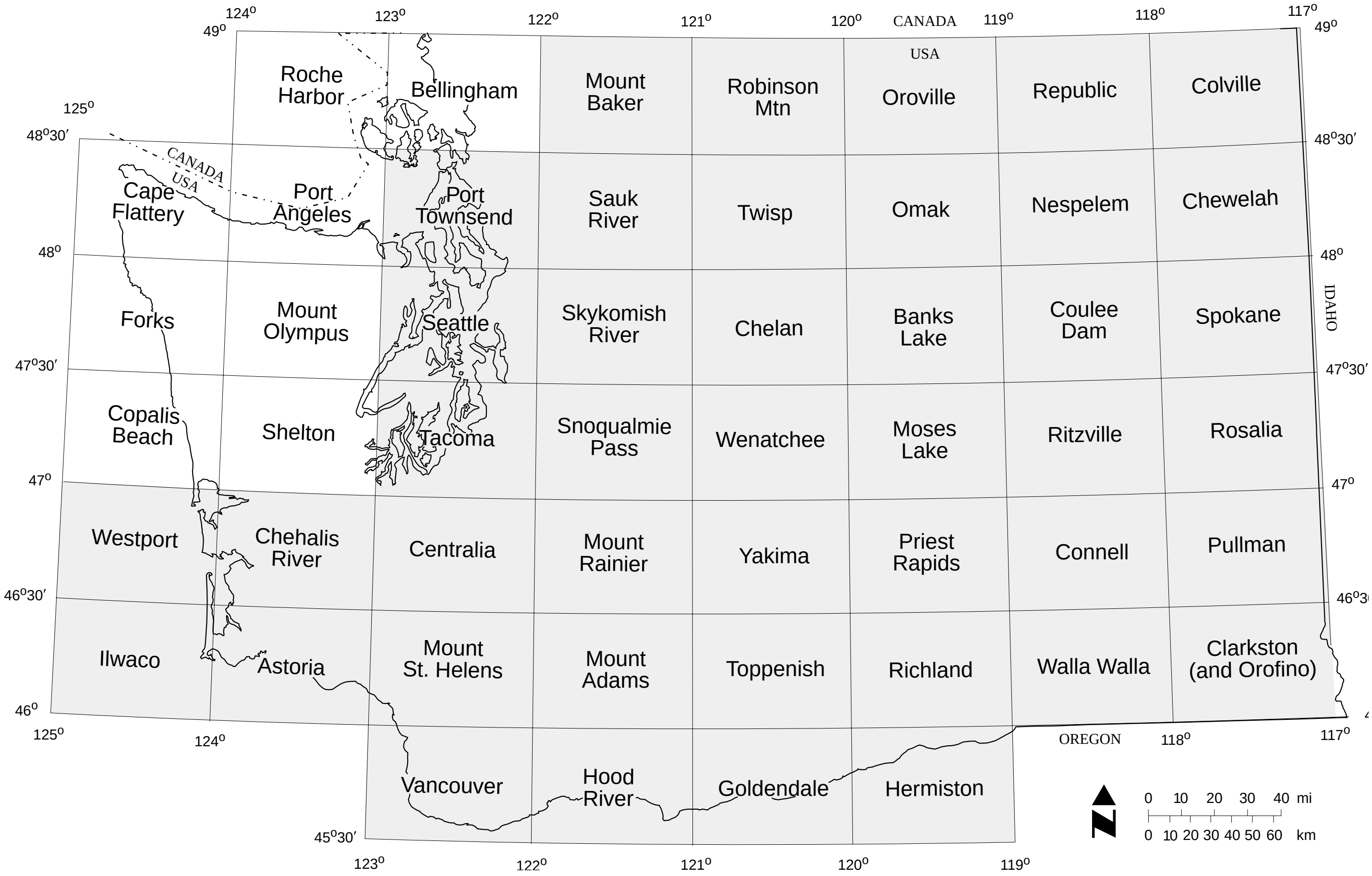
Digital Geologic Maps Available

Digital geologic maps in Arc/Info 7.1.2 (standard cover format or covers bundled into TAR files; stateplane south / NAD 27) are available for the following 1:100,000 quadrangles (shaded): Astoria, Banks Lake, Centralia, Chehalis River, Chelan, Chewelah, Colville, Connell, Coulee Dam, Goldendale, Hermiston, Hood River, Ilwaco, Moses Lake, Mount Adams, Mount Baker, Mount Rainier, Mount St. Helens, Nespelem, Omak, Oroville, Port Townsend, Priest Rapids, Pullman, Republic, Richland, Ritzville, Robinson Mountain, Rosalia, Sauk River, Seattle, Skykomish River, Snoqualmie Pass, Spokane, Tacoma, Toppenish, Twisp, Vancouver, Walla Walla, Wenatchee, Westport, and Yakima as shown below. Mapping is currently in progress for the unshaded quadrangles. We can provide all of these maps on a CD (please send $1.00 to cover shipping and handling) or to your FTP site. We'd appreciate your giving us credit as the source of data. This work was supported by the U.S. Geological Survey STATEMAP program, agreements 1434-HQ-96-AG-01523, 1434-HQ-97-AG-01809, and 1434-HQ-98-AG-2062. To order, contact Chuck Caruthers at (360) 902-1455 or charles.caruthers@wadnr.gov.



Some Notable Finds of Columbian Mammoths from Washington State

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On March 25, 1998, Governor Gary Locke signed House Bill 1088 into law establishing the Columbian mammoth (*Mammuthus columbi*) as the "official fossil [species] of the State of Washington". This legislation marked the culmination of a four-year effort on the part of students in Mrs. Sara Aebly's second grade class at Windsor Elementary School near Spokane (Barton, 1998). Because of the students' remarkable persistence, Washington now joins several other western states, including Alaska (woolly mammoth) and California (sabre-toothed cat), in having designated Ice Age (Pleistocene) mammals as their official state fossils.

Mammoth fossils are particularly common in Washington, with several hundred finds having been reported in various publications or donated to local, regional, and national museums or collections. Where sufficient data exist to assign them to species, the vast majority have proven to be Columbian mammoths (Barton, 1998). Of the 39 counties in Washington, only heavily forested counties on the west side of the Cascade mountains (for example, Skamania and Wahkiakum) and less populated counties on the east side (for example, Ferry and Pend Oreille) have thus far failed to produce mammoth fossils.

Most of the reported remains from Washington are of single skeletal elements, with molars by far the most common. Tusks are also quite common, though rarely well preserved. More notable or significant mammoth finds are less common. These include sites with multiple skeletal elements (bones and/or teeth) found in direct association with one another, sites that can be well dated (either absolutely as in radiocarbon dating or relatively through stratigraphic association), and sites that represent geographic range extremes for this genus within the state (Barton, 1999).

Columbian Mammoths in North America

Columbian mammoths are one of two species endemic to North America, the other being the imperial mammoth (*M. imperator*). The remaining two species of mammoth found in North America, *M. meridionalis* (*M. hayi*) (southern mammoth) and *M. primigenius* (woolly mammoth), both evolved in the Old World and migrated into North America from Asia by way of the Bering land bridge. Columbian mammoths speciated from imperial mammoths roughly 300,000 to 500,000 years ago and quickly became the dominant mammoth throughout North America. Columbian mammoth remains have been found from Alaska to Florida, and from northern Canada to southern Mexico. In Utah and Colorado, *M. columbi* has been found at elevations greater than 2700 m (8858 ft) (Gillette, 1989), while on the continental shelf off the Atlantic coasts of Canada and the U.S., molars from this species have been recovered from depths of at least 120 m (393 ft) (Cooke and others, 1993; Whitmore and others, 1967).

Columbian mammoths were moderate in size, standing roughly 3.4 m (11 ft) at the shoulders. This made them taller than their contemporary cousins, the woolly mammoth, but shorter than their immediate predecessors, the imperial mammoth (Madden, 1981). Based on their more southerly geographic distribution, they seem to have been adapted to warmer temperatures than the woolly mammoth and were probably therefore less hairy than *M. primigenius.* They most likely resembled an overly large Asian elephant (*Elephas maximus*) that we see today, only with smaller ears and carrying more massive tusks.

First and Last Mammoths in Washington

The imperial mammoth teeth that have been found in Washington suggest a long presence for mammoths in this state, exceeding at least 300,000 to 400,000 years (Hay, 1927). Additionally, a *M. meridionalis* was found in southeastern Idaho that would allow for the possibility of mammoths in the Pacific Northwest as far back as 1,700,000 yr B.P. (Malde and Powers, 1962). Unfortunately, most Washington mammoth fossils have been recovered without due consideration of their stratigraphic context, so it is difficult to know precisely when *M. columbi* first arrived in the state.

In eastern Washington, the oldest mammoth fossil may be the one recovered from loess of the Palouse Formation near St. John, Whitman County, in 1962 (see site 14 below; Fryxell, 1962). Other early mammoth remains that were found in pre-Wisconsinan-age loess deposits are from Burr Canyon (site 02) and Cheney (site 03). They could be as old or older than the St. John mammoth. In western Washington, Columbian mammoth molars have reportedly been recovered from Whidbey Formation sediments at Scatchet Head on Whidbey Island (Barton, 1992). All of these finds were in stratigraphic contexts that pre-date the last (Wisconsinan) glaciation and therefore suggest a late middle Pleistocene or early late Pleistocene age if not earlier.

We know more precisely when the last Columbian mammoths roamed Washington because their remains, or associated botanical finds, have been dated by radiocarbon analysis. Based on current data from the Puget Lowland, the last mammoths were gone by 15,000 to 17,000 yr B.P., although most of our well-dated sites from this subprovince date to between 20,000 and 22,000 yr B.P. (see sites 11 and 12 below; Barton, 1992). In eastern Washington, Columbian mammoths were still present as late as 11,000 to 13,000 yr B.P. (see sites 01, 16, 17, and 18 below; Waitt, 1980).

As far as we know, Columbian mammoths were obligate herbivores with a dietary preference for grasses, sedges, sages, mosses, ferns, and aquatic plants (Barton, 1998). In both eastern and western Washington, they seem to have been driven from the state by rapidly changing climatic conditions and deteriorating habitat, rather than having been hunted out by Paleoindians, as was once believed. In the Puget Lowland, mammoths were physically blocked from what had previously been their seasonal grasslands range by rapidly advancing lobes of the Vashon glaciation by 15,000 yr B.P. In eastern Washington,

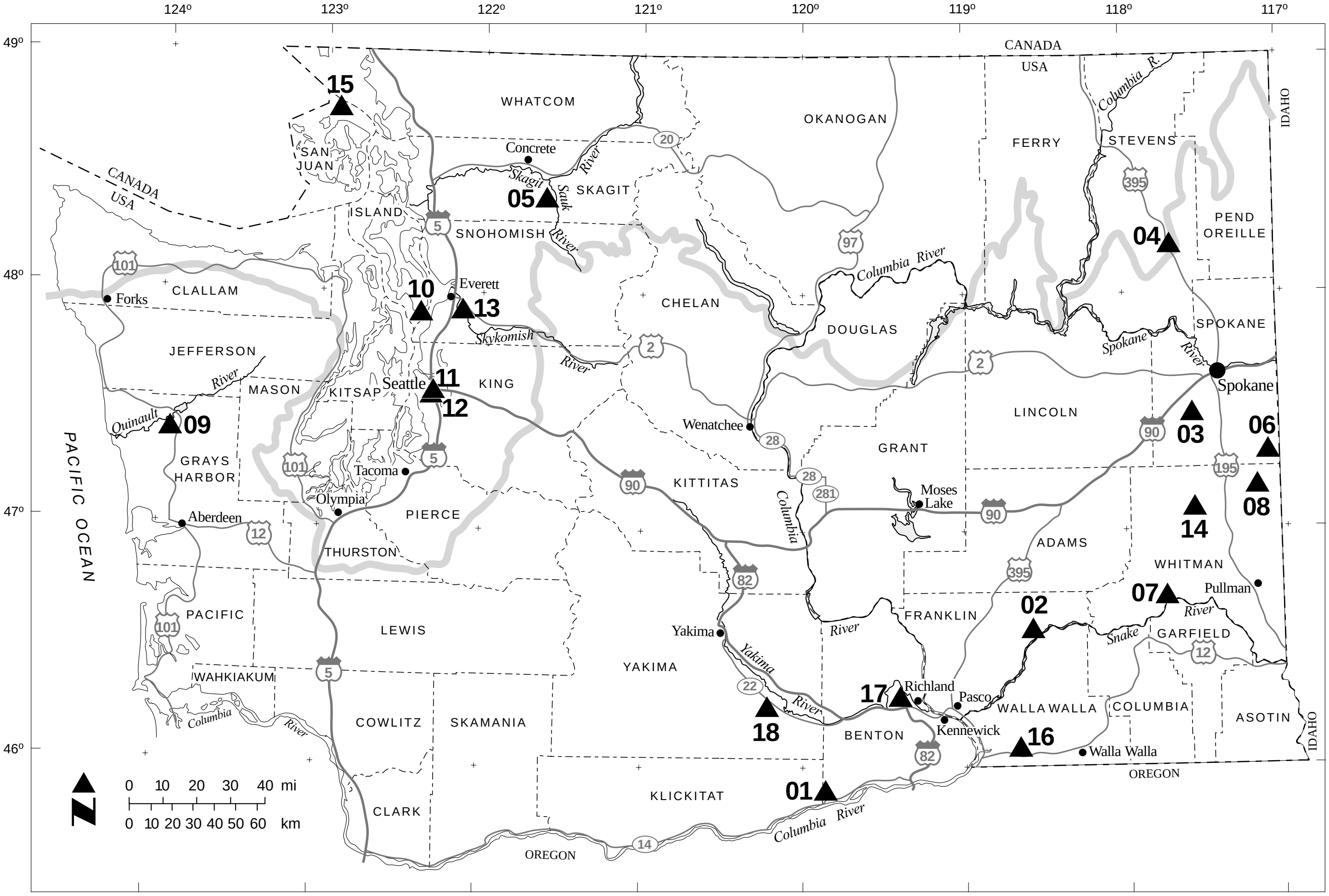

Figure 1. Distribution of Columbian mammoth sites discussed in this article. In westen Washington, mammoths are commonly found in sediments of the Olympia nonglacial interval (20,000–60,000 yr B.P.); in eastern Washington, most mammoths are found in the later part of the Touchet Formation (11,000–20,000 yr B.P.). The shaded line suggests the maximum extent of the Cordilleran ice sheet at the Wisconsinan late glacial maximum (c. 15,000–20,000 yr B.P.). Many finds in western Washington are north of this line; most finds in eastern Washington are well south of the line.

mammoths were eventually driven from the state by the increasing temperatures of the late post-glacial/early Holocene climatic warming at about 11,000 yr B.P.

Some Notable Washington State Columbian Mammoth Sites

The list that follows gives each mammoth find a site number (Fig. 1), a name (based on geographic location), the name of the finder or first reporter in *italics*, the county in which it is located [in brackets], and a brief description of the remains.

01 **Artesian Coulee/Dead Canyon – *Newcomb* [Benton Co.]:** Post-cranial mammoth remains recovered from a blowout within the Touchet Formation. A ^{14}C date on these bones produced an anomalously young date of 4905 ±140 yr B.P. [GX-1457]. They were relatively dated by stratigraphic association to between 11,000 and 13,000 yr B.P. (Newcomb, 1971; Newcomb and Repenning, 1970; Waitt, 1980).

02 **Burr Canyon – *Strahorn/Bryan* [Franklin Co.]:** Most of the skeleton of a very aged Columbian mammoth collected by a soil survey crew of the U.S. Bureau of Soils in 1923 and forwarded to the U.S. National Museum/Smithsonian Institution in Washington, D.C. This mammoth was reportedly recovered from loess deposits in the Palouse Formation, and therefore is probably older (perhaps much older) than 32,000 yr B.P. This find must certainly be older than the mammoths recovered from the flood deposits of the Touchet Formation, which date between 11,000 and 32,000 yr B.P. (Bryan, 1927; Hay, 1927).

03 **Cheney – *Freeman* [Spokane Co.]:** Well-preserved teeth and badly decayed bones of an early Columbian mammoth reportedly found in 1926 by a farmer plowing his fields near Cheney. At least one of the molars, a lower fifth (?M_5), was sent to the University of Chicago collections. The bones were found in an older loess deposit directly above a "well weathered" pre-Wisconsinan-age till, suggesting a relative date of mid- to early late Pleistocene. (Freeman, 1926; Hay, 1927).

04 **Chewelah – *Lewis/Hay* [Stevens Co.]:** A single upper right sixth molar (RM^6) collected in 1920 near Chewelah by workmen of the Magnetite Company. Found at roughly 48°15′N, this molar is currently the northernmost reported mammoth find from eastern Washington. All other reported mammoth fossils from eastern Washington have been recovered from unglaciated lands south of the last glacial Cordilleran ice sheet margins. The 'Chewelah' mammoth may have been found at such a northerly latitude because it pre- or post-dates the last glacial maximum or because it occupied unglaciated lands between the Colville

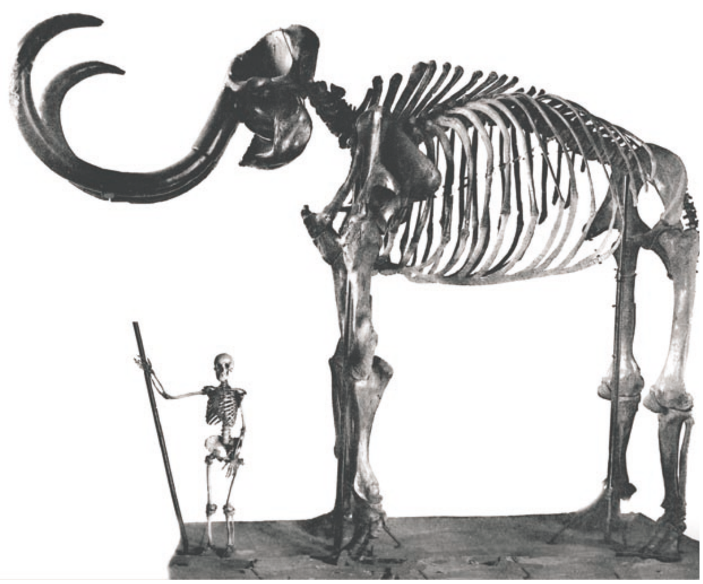

Figure 3. Mounted composite skeleton of a Columbian-type mammoth made from skeletal elements recovered in the 1870s from the 'swamps' at the Copelin Ranch along Latah Creek in Spokane County (site 06). When assembled in 1886 in the Field Museum of Natural History in Chicago, Illinois, this 'mammoth' was considered to be the first fully mounted specimen, albeit a composite from several individuals, of a mammoth in North America. *(Photo from Higley, 1886.)*

and Spokane lobes of the Cordilleran ice sheet at or near its last maximum advance (Hay, 1927).

05 **Concrete/Sauk River – *Thompson* [Skagit Co.]:** Cranium, two tusks, and two upper sixth molars (left and right, LM^6 and RM^6) from sands and gravels exposed above the Sauk River (Fig. 2). Recovered in 1979 by a crew from the Geology Department of Western Washington University. Unlike most mammoth finds in the Puget Lowland, which are generally found within 1 km (0.6 mi) of the marine coastline, this mammoth was located well upvalley (although still at less than 120 m or 394 ft above sea level) and some 60 km (38 mi) east of the nearest marine estuary at Padilla Bay. This site also marks the most easterly known occurrence of a mammoth west of the Cascade Range (Barton, 1992).

06 **Latah Creek/Copelin Ranch – *Higley* [Spokane Co.]:** Bones and teeth of at least six mammoths were taken by the wagon-load from 'swamps' here in the 1870s (Fig. 3). A composite Columbian mammoth skeleton was pieced together from these fossils in 1886 and was displayed in the Field Museum of Natural History in Chicago. At the time, this was believed to be the first fully mounted (assembled) mammoth skeleton in North America. Details of the discovery suggest a post-glacial, post-Missoula floods age for these fossils (Hay, 1927).

07 **Penawawa – *Lewis* [Whitman Co.]:** A large part of the skeleton of a Columbian mammoth was found near here, but was too unstable to be conserved at the time. Nothing is apparently known of the find site or its probable age (Hay, 1927; Madden, 1981).

08 **Pine Creek – *Sternberg* [Whitman Co.]:** Excavations at springs along this creek in the 1870s yielded a considerable number of mammoth bones, some of which were eventually acquired by the American Museum of Natural History in New York. The circumstances of these finds are similar to those at Latah Creek (site 06 above), and their dating is probably roughly contemporaneous with those mammoths (Hay, 1927; Sternberg, 1903).

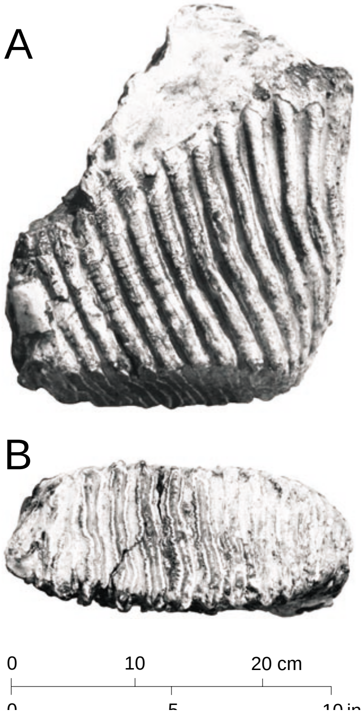

Figure 2. Upper left sixth molar (LM^6) from the Concrete/Sauk River – *Thompson* site (site 05). **A**, lateral (side) view; **B**, occlusal (grinding surface) view. (B. Thompson, private collection.)

09 **Quinault River/Blue Banks – *Geoghegan/Hall* [Grays Harbor Co.]:** A partial skeleton of a Columbian mammoth was recovered here from a thick deposit of "blue" lake clays along the lower Quinault River. Current research into the date of these and similar clay units along the outer Washington coast suggests an Olympia nonglacial interval age (~20,000–60,000 yr B.P.) for this find (Thackray, 1996).

10 **Scatchet Head/Whidbey Island – *Willoughby/Lawson* [Island Co.]:** Various mammoth remains found about 1860 at the foot of a seacliff, reportedly brought down to beach level by a massive landslide. These may be the first reported mammoth finds from the state. They were collected by Capt. Charles Willoughby of the U.S. Coast Survey Brig *R. H. Fauntleroy* and donated some 14 years later [1874] to the California Academy of Science (CAS) by J. S. Lawson. No longer in the CAS collections, these fossils are assumed to have been destroyed in the 1906 San Francisco earthquake and subsequent firestorm (Lawson, 1874).

11 **Seattle/Mercer & Yale – *Stewart/Sharahira* [King Co.]:** Skeletal and dental elements of a single Columbian mammoth, unearthed in 1963 during excavations for a freeway access ramp and reported by the bulldozer crew of Allan Stewart and Don Sharahira (Fig. 4). Recovered from a "blue-green clay" unit within Olympia nonglacial interval sediments and relatively dated by stratigraphic association to late in the Olympia nonglacial interval (15,000–25,000 yr B.P.) (Mullineaux and others, 1964). See site 12 below.

12 **Seattle/6th & Seneca – *Green* [King Co.]:** Skeletal and dental elements of a single Columbian mammoth unearthed in 1963 during excavations at the IBM Building site and reported by the bulldozer operator, Byron Green (Fig. 4). Found in a "blue clay" unit within Olympia nonglacial interval sediments. ^{14}C dating of an associated wood sample yielded a corrected age of 21,836 ±300 yr B.P. [UW-55] (Fairhall and others, 1966).

13 **Snohomish – *Preston/Ludwig* [Snohomish Co.]:** A partial skeleton of a mammoth was recovered near here in 1936, and several of the bones (scapula and ulna) were deposited in the Burke Museum at the University of Washington

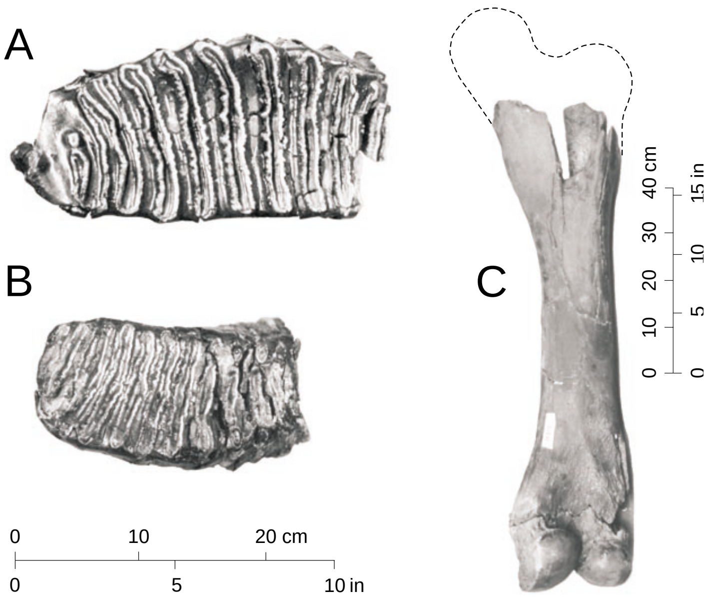

Figure 4. **A**, occlusal view of a lower left sixth molar (LM$_6$) [UWBM Geology no. 19190] found at the Seattle/6th & Seneca – *Green* site (site 12); **B**, occlusal view of a lower right sixth molar (RM$_6$) [UWBM Geology no. 27450] found at the Seattle/Mercer & Yale – *Stewart/Sharahira* site (site 11); **C**, lateral view of a right femur (thighbone) [UWBM Geology no. 18817] found at the Seattle/6th & Seneca – *Green* site (site 12). Dashed line indicates missing portion of bone.

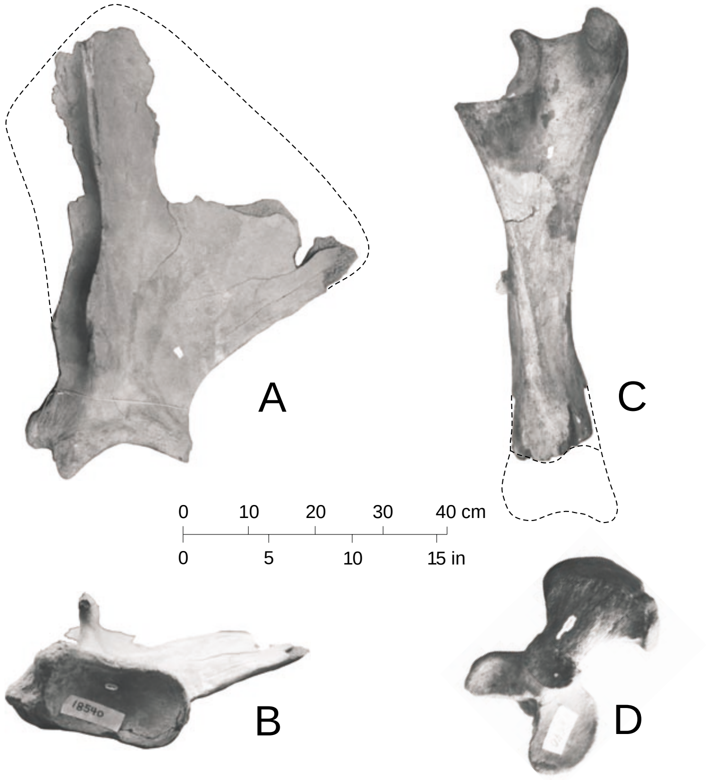

Figure 5. Bones from the Snohomish – *Preston/Ludwig* site (site 13). **A**, lateral view of the left scapula (shoulder blade) [UWBM Geology no. 18540]; **B**, proximal view of the left scapula; **C**, lateral view of the right ulna (distal forearm bone) [UWBM Geology no. 18540d]; **D**, proximal view of the right ulna. Dashed line indicates missing portion of bone.

(Fig. 5). The absolute age of these bones is unknown, and the relative age is uncertain.

14 **St. John – *Fryxell* [Whitman Co.]:** Fragmentary bones and tusks of a mammoth were collected near here in 1962 by a crew from Washington State University. The skeleton was found in slack-water sediments within Palouse Formation loess deposits and was regarded as “not only pre-Wisconsin, but mid-Pleistocene” in age (Fryxell, 1962).

15 **Sucia Island – *Godsall/Newcombe* [San Juan Co.]:** A single lower left fifth molar (LM$_5$), donated by B. Godsall in 1895 to what is now the Royal British Columbia Museum (Victoria, B.C.) (Fig. 6). Found at roughly 48°45′N, this is currently the northernmost reported find of Columbian mammoth in Washington (Hay, 1927).

16 **Walla Walla/Gardena – *Fulgham* [Walla Walla Co.]:** A “fairly complete” Columbian mammoth skeleton was recovered here in 1966 and placed in the geology collections of Whitman College (Fig. 6). Many of the bones however were crumbly, scattered, and fragmentary “suggesting postmortem redistribution”. These bones were found 1.5 m (5 ft) below the surface within the Touchet Formation flood deposits, which suggests an age, estimated by stratigraphic association, of roughly 12,000 to 13,000 yr B.P. (Scott and Clem, 1967; Waitt, 1980).

17 **West Richland – *Jeppson* [Benton Co.]:** A partially articulated skeleton of a Columbian mammoth was excavated here in 1978 by a crew from the Burke Museum at the University of Washington (Fig. 7). The bones were found at a depth of 2.5 m (8 ft) within the Touchet Formation flood deposits and beneath a deposit of Mount St. Helens ‘set S’ tephra, suggesting an age greater than 13,000, but less than 20,000 yr B.P. (Martin and others, 1982; Waitt, 1980).

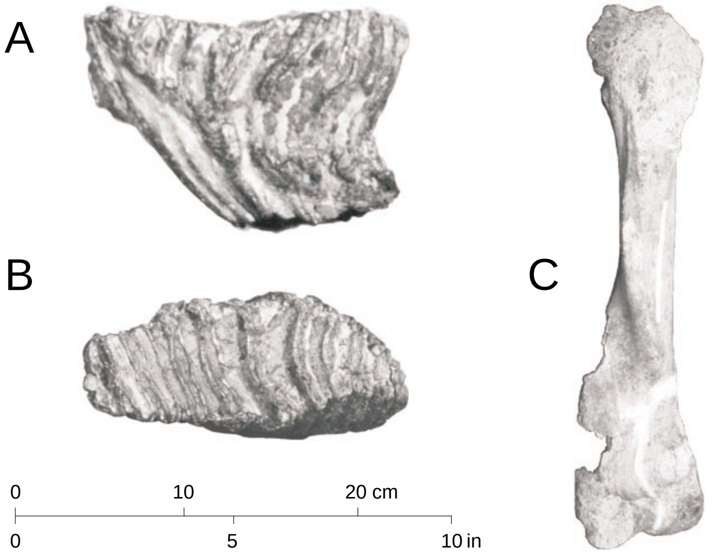

Figure 6. **A**, lateral view of a lower left fifth molar (LM$_5$) [RBCM no. 170] from the Sucia Island – *Godsall/Newcombe* site (site 15); **B**, occlusal view of the same molar; **C**, right humerus (upper forearm bone) from the Walla Walla/Gardena – *Fulgham* site (site 16) (no scale given).

18 **Yakima Valley – *Gustafson* [Yakima Co.]:** A partial mammoth skeleton was found here, some 2.5 m (8 ft) below a tephra deposit within the Touchet Formation flood deposits. Columbian mammoth remains are common finds in the Yakima Valley from Selah southeast to the Columbia River. Like the West Richland mammoth (site 17 above),

Figure 7. Bones from the West Richland – *Jeppson* site (site 17). **A**, mandible (lower jaw) [UW Burke Museum no. 61675]. Note molars in jaw for scale. **B**, left humerus [UW Burke Museum no. 7828].

the age of this find can be estimated by stratigraphic association as between 13,000 and 20,000 yr B.P. (Waitt, 1980).

Acknowledgments

I am grateful to P. T. Pringle and J. M. Roloff for their patience and constructive comments on this article; to T. J. Ayers, R. C. Byersdorf, D. J. Easterbrook, B. Hallet, C. R. Harington, R. H. Hevly, T. M. Oakley, S. C. Porter, R. Scott, T. W. Swanson, and B. Thompson for their support and encouragement, and to C. R. Harington and R. B. Waitt, Jr., for copies of offprints cited above.

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Author's Note: Because of their size and density, mammoth bones/teeth are some of the most commonly found fossils in Washington State. Anyone with such finds is encouraged to contact the author, who will be pleased to examine your specimen(s) for their scientific merit. If you have an interest in mammoths and other Ice Age genera, the following museums have displays featuring such finds: Adam East Museum (Moses Lake); Burke Museum/University of Washington (Seattle); Geology Department collections at Western Washington University (Bellingham) and Whitman College (Walla Walla); Karshner Museum (Puyallup); Sequim Museum (Sequim); and Yakima Valley Museum (Yakima).

July 2, 1999, Satsop Earthquake

A large earthquake shook the Pacific Northwest on July 2, 1999, at 6:44 p.m. (PDT). The quake occurred at a depth of 25 mi (41 km) beneath Satsop, Wash., about 27 mi west of Olympia. The Pacific Northwest Seismograph Network reported a coda magnitude (based on how long the shaking lasts as recorded by seismograms) of 5.1. Although the coda magnitude scale works well for small and moderate size earthquakes, it has not been well calibrated for deep and large earthquakes. Other magnitude scales commonly used include the body wave magnitude (mb), surface wave magnitude (M_s) and moment magnitude (M_w). Other estimates of magnitude for this earthquake were mb=5.5, M_s=5.5, and M_w=5.7 to 5.9. By any measure, this was the largest deep quake to hit the region since the Puget Sound earthquake of 1965. The Satsop earthquake was similar to that and the 1949 Olympia (magnitude 7.1) in that it occurred within the subducting slab of the Juan de Fuca plate. The focal mechanism is normal and is probably down-dip tensional, that is, it probably represents a pulling-apart of the down-going plate.

The only historical earthquakes that have done significant damage were ones that occurred like this one, deep down in the subducted plate. “These quakes can kill people,” Steve Malone, University of Washington seismologist, said.

Gas leaks, toppled chimneys, and power outages were reported all over Grays Harbor County after the earthquake, according to Rob Harper of Washington State Emergency Management, particularly in Hoquiam, Aberdeen, Brady, Satsop, and Montesano. Karin Frinell-Hanrahan of Grays Harbor County Emergency Management reported initial damage estimates at ten million dollars for county buildings alone. The historic Grays Harbor County Courthouse (Fig. 1) accounted for a major portion of it. More than 300 homeowners also reported damage. The County’s 911 operations center reported receiving over 2,700 phone calls the night of the quake. Many callers asked about the danger of a tsunami following the tremor.

In Montesano, Dennis Selberg, Facilities Director for Grays Harbor County, said that the County Courthouse, built in 1910, sustained ”very scary, substantial damage (Fig. 1) and is currently undergoing extensive repairs.” Other damage in the area was reported from a large furniture store, where the ceiling and an exterior wall collapsed (Fig. 2), and from a number of fire stations. The earthquake was widely felt in western Washington and Oregon.

Figure 1. Cupola of the Grays Harbor County Courthouse in Montesano, Wash. This historic building suffered several million dollars’ worth of damage during the July 2 earthquake whose epicenter was about 1 mile away. *Photo and caption courtesy of Grays Harbor County.*

Figure 2. Merchandise inside Moore’s Furniture in Aberdeen, Wash., lies smashed after a 5.9 magnitude earthquake struck the area July 2, 1999. Extensive damage to the Moore’s building was caused by an exterior wall that collapsed inward, crashing through the ceiling of the furniture store. *Photo and caption courtesy of Kevin Hong, The Daily World, Aberdeen, Wash.*

Compiled from the Pacific Northwest Seismic Network website:
http://www.geophys.washington.edu/SEIS/
EQ_Special/WEBDIR_99070301435p/welcome.html
[downloaded Feb. 9, 2000]
and “Washington earthquake deepest since 1965”,
Oregon Geology, v. 61, no. 4, p. 95, July/August 1999

Do We Really Need Another Wake-up Call?

A first-person account of the July 2, 1999, Satsop earthquake. Reprinted with permission from John Hughes' "Letter from the Editor" column in *The Daily World,* Aberdeen, Washington, July 4, 1999, page A4.

Dear Reader: That got my attention. By eerie coincidence, reporter Ryan Teague Beckwith and I were discussing the major natural disasters of the 20th century on Grays Harbor—the Columbus Day Storm of 1963, the blizzard of 1950, the rainfall record of 113.49 inches in 1933—when two tectonic plates did a bump and grind that stopped short of cataclysmic. A Richter here and a Richter there, and we could have had a front row seat for the No. 1 headline of the fast-ebbing old millennium and never lived to write about it.

Although we're joking about the emotional fallout—the brain's way of coping—most of us now have a better understanding of post-traumatic stress. When I think of the what-ifs, I really get scared. And I was scared at 6:43 p.m. Friday, July 2, 1999. So scared that I stood for several seconds in front of a seven-foot-tall bookcase instead of diving under my desk, an antique so substantial that it likely could withstand a direct hit by an ICBM.

(Would I have shared my space with Ryan? He has his whole life ahead of him. I'm 55. I've lived in Bermuda, owned two Porsches and have a personal letter from Annette Funicello. Although he sometimes looks at me with the secret glint of youthful contempt, as if I'm just another worn-out Boomer worried about prostate trouble and glued to a 401(k) hotline, it would have been the right thing to say, "Quick, Ryan! Under here!" But naaaaa! Crawl under your own desk, cheeky twirp!)

Beckwith, given often to ironic understatement, stood frozen in the doorway and declared, "I think this is an earthquake." And I said, "Holy #@$*!" Or words to that effect. The newsroom emptied into the parking lot fronting historic State Street, which sits atop several jillion cubic yards of sawdust spaltz. In fact, this whole end of town was a salmonberry marsh a century ago.

We rode the wave for 40 seconds. It seemed like an eternity. Streetlight poles shook, my Volkswagen Beetle did the Macarena while Dee Anne Shaw's Chrysler coupe was undulating. There were a half-dozen of us looking at one another like deer caught in the headlights of an oncoming car. Then the shaking stopped. Seconds later, the first siren.

I've been in bigger quakes—a lot bigger quakes—but this one lasted longer and felt stronger.

A Rude Awakening

In June of 1992, I was finishing up a month-long stint as acting editor of our company's newspaper at Hemet, Calif., east of L.A., when I endured the longest 30 seconds of my life. Then it happened all over again three hours later.

It was the definitive rude awakening at 4:58 a.m., when the bright-red Mickey Mouse alarm clock my daughter Sarah had loaned me for the trip rocketed off the nightstand. The four heavy drawers in the bureau slid open with a whoosh and everything in the bathroom medicine cabinet crashed onto the tile floor. The room was rolling. I was riding the bed and saying Hail Marys.

It was California's strongest earthquake in 40 years—7.4 on the Richter, infinitely stronger than the 5.5 we experienced Friday night. The aftershocks were relentless. I couldn't get back to sleep, so I actually read the Gideons' Bible. I was brushing my teeth at 8:07, when the second one hit. It was only a 6.5, but the jolt was even stronger—a violent side-to-side motion.

As a rule, I only need one wake-up call, literally and figuratively. I had to go to the bathroom, but the thought crossed my mind that I didn't want to be found dead on the toilet a thousand miles from home, so I threw everything in the suitcase and headed for the stairs.

I waited for an hour in the hotel parking lot, bags at my side. The sky was alive with arcing bolts of light, as transformers exploded for miles around.

Dave Caffoe, who was general manager at *The Daily World* in the early '70s, was the publisher at Hemet. I was there as a favor to him.

He was laughing as he pulled into the portico of the Doubletree and popped the trunk lid on his white Oldsmobile. "I gather you'd like to go home," he said.

I declined his offer of a Bloody Mary with celery stalk, opting for black coffee and a boarding pass.

Not In My Backyard

That was then; this is a more sobering now. This is home. Despite the absolute consensus by scientists that The Big One is coming to the Northwest—not if, but WHEN—I've always kidded my friends in California about their precarious existence.

Sure, it could happen here, I thought, but it probably won't. It's gonna be Seattle or, better yet, Bellevue. Not in my backyard. I've been in denial. You too?

I lost a lovely Tiffany-style lamp Friday night. A thousand-dollar lamp that I got for a song 30 years ago. It tumbled off the rolltop desk in the hallway. There's plaster damage in the kitchen and dining room, and a beam in the garage is askew.

But I'm counting my blessings. The lamp, with its heavy leaded-glass shade, could have hit Sarah, who was scrambling for cover. If those tectonic plates had shifted just a little bit more, the ground could have turned to goo and swallowed my family—maybe yours too. Forget the lamp. The tsunami that followed could have killed thousands.

As I made a quick reconnaissance of the area around the newspaper, I imagined the center span of the Chehalis River Bridge upright in the water, like the arm from the Statue of Liberty in the climactic scene of "Planet of the Apes."

I imagined the Becker Building a pile of smoking rubble and the parking lot of Wal-Mart as one giant field hospital.

I saw the remains of Community Hospital halfway down the hill. Dee Anne's house, with husband John and 9-month-old Gordon, the cutest baby in the world, is just below the hospital.

Driving home to Hoquiam through the pitch-black along Sumner Avenue at 2:30 a.m. Saturday after the presses rolled, I imagined no lights anywhere, no water, fires out of control in a hundred homes and businesses, gas lines ruptured. Chaos.

Survival could require a blend of luck, pluck, and smarts. I, for one, as the letter writers always say, am going to start paying attention to those emergency checklists of do's and don'ts.

And if you think the best thing to do in an earthquake is call 911, you might as well hang up and kiss your silly derriere goodbye. Be prepared. There will be another earthquake. Earthquakes. One is bound to be bigger. Maybe a whole lot bigger. I don't need another wake-up call, but if you can repair leaded glass I'd like to hear from you.

John Hughes can be reached
at 360-532-4000, ext. 112,
or editor@thedailyworld.com

EARTH CONNECTIONS

Resources For Teaching Earth Science

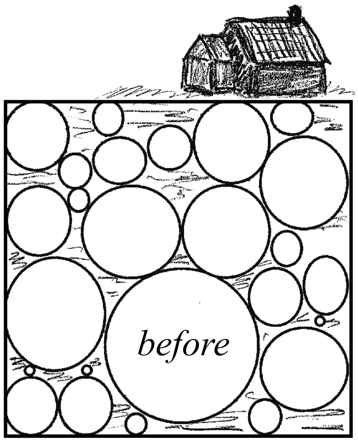
SIMPLE HOME EXPERIMENTS FOR BRINGING GEOLOGY TO LIFE

EXPERIMENT 1: SHAKE, RATTLE, AND LIQUEFY

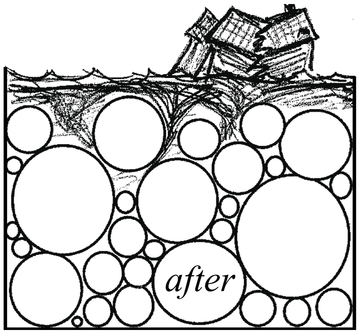
BACKGROUND: When sediments liquefy, they lose their structure and strength. During earthquake shaking, the individual grains of sand within a deposit collapse on each other. Anything built on them can sink or collapse. Picture a container of balls of slightly different sizes—baseballs, golfballs, marbles. If they were transported by water into the container and then deposited, they would settle with spaces between them. Some of the spaces would be filled with water, some with air. When you shake the container, the balls settle against each other, and the water and air are forced to the surface. That is exactly what happens in a sediment-filled valley. The valley is a large ‘container’ holding gazillions of ‘balls’ or grains of sand. Shaking the container simulates an earthquake.

EQUIPMENT NEEDED:

✓ Transparent (glass) baking pan
✓ Enough dry sand to fill your pan 1 to 2 inches
✓ A few toy houses or wooden blocks
✓ Water



PURPOSE: We know that flat river valley bottoms are prone to flooding, but we often think of them as being geologically stable. This experiment will teach you what happens to sandy soils when they liquefy. It will show you how to create a ‘model’ river valley, then watch how and why houses get damaged or collapse during an earthquake in a seemingly stable geologic environment.



PROCEDURE:

1 Evenly pour the dry sand into the baking pan.
2 Mark the level of the sand on the side of the pan.
3 Place the houses or blocks gently on the surface.
4 Slowly add water until about two-thirds of the thickness of the sand is saturated.
5 Gently start shaking the table on which you have placed your baking pan (or the pan itself).

OBSERVATIONS: You should see the following:

✓ The water will work its way to the surface, flooding the area around the houses,
✓ The houses will start leaning over and sinking into the sand, and
✓ The volume of the sand should decrease by a small amount.

EXPANDED ACTIVITY: Now be creative. Try the experiment using clay or gravel to separate sand layers and represent different types of sedimentary layers. Watch what happens to the water and the surface of your model of a river valley.

Wendy Gerstel, Geologist *geology@wadnr.gov*
Washington Division of Geology and Earth Resources
PO Box 47007; Olympia, WA 98504-7007

ESSENTIAL SCIENCE LEARNING BENCHMARKS

1.3 Students will understand that interactions within and among systems cause changes in matter and energy.

2.2 Students will also apply science knowledge and skills to solve problems or meet challenges.

GRADE LEVELS

6th–10th grades

SUBJECTS

Earth science
Liquefaction
Earthquakes

CONCEPT

How different soils react during an earthquake.

SKILLS

Observing, comparing and contrasting, and identifying relationships.

OBJECTIVE

Students will test soils’ reaction to ground shaking representing an earthquake.

TIME NEEDED

60 minutes or less

DISCUSSION

1. Compare what happens to the water when using different soils; describe what happens between water and soil for each type tested.
2. How would these differing soils affect human development, such as structures or houses?
3. How can people plan for earthquakes when considering a new building location?
4. How can people plan for earthquakes if their houses are already in hazardous places?
5. How can people find out if their houses are in hazardous places? (Contact the Washington Division of Geology and Earth Resources.)

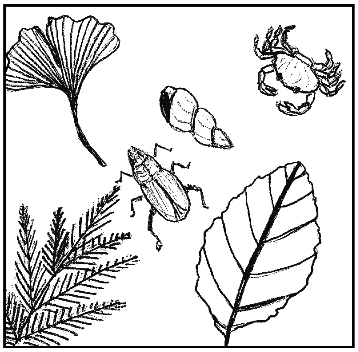
Earth Connections No. 2

EARTH CONNECTIONS

Resources For Teaching Earth Science

SIMPLE HOME EXPERIMENTS FOR BRINGING GEOLOGY TO LIFE

EXPERIMENT 2: CONDENSING GEOLOGIC TIME OR THE ART AND SCIENCE OF MAKING FOSSILS



BACKGROUND: Finding plants, animals, and even early humans buried in the geologic strata gives us clues to what our planet Earth was like in the past.

EQUIPMENT NEEDED:

✓ Small oven-proof dish or pan
✓ Clay, local, natural source if you're lucky, otherwise play-dough or modeling clay will work (No oil-base clays! They will burn in the oven.)
✓ Leaves, *empty* shells, dead bugs, etc.
✓ Sand

PURPOSE: This experiment will teach you about the process of fossil burial, preservation, and discovery. It will give you the opportunity to think about the types of things (or specimens) one finds buried in sediment, about the sediments and processes that preserve these specimens as fossils, and about TIME.

PROCEDURE:

1 Layer the bottom of your dish with about one-half inch of the clay.
2 Explore your backyard or a nearby beach and find things that might become fossilized if they were to be buried for a few million years, making sure that whatever you pick up is no longer alive!
3 Next, press your finds gently into the clay.
4 Then, cover this layer of fossils-to-be with a thin layer of sand. This is so your clay layers will part easily after you 'bake' your fossils.
5 Carefully add another layer (or geologic stratum) of clay to your sample. You are now ready to dry your sediments with the buried 'fossils'.
6 *MAKE SURE YOU WORK WITH AN ADULT FOR THIS NEXT STEP*. Put the dish in an oven on very low heat. You want to dry your sample slowly so it doesn't crack. This might take an hour or more depending on how wet the clay was.
7 When the sample looks dry, VERY GENTLY remove it from the dish and pry it apart at the sand layer.

OBSERVATIONS: You should be able to see:

✓ Your 'fossil' specimens,
✓ The impressions made in the upper and lower clay surfaces, and
✓ How the sample broke along the sand layer.

EXPANDED ACTIVITY: Find a book about fossils at your library and look up the difference between 'casts' and 'molds' and see if you can identify each in your sample.

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ESSENTIAL SCIENCE LEARNING BENCHMARKS

1.1 Students will use properties to identify, describe, and categorize substances, materials, and objects, and use characteristics to categorize living things.

GRADE LEVELS

4th–8th grades

SUBJECTS

Earth science
Geologic time
Fossils

CONCEPT

What geologic conditions lead to the formation of fossils.

SKILLS

Observing, analyzing, classifying, and identifying relationships and patterns.

OBJECTIVE

Students will simulate the making of fossils.

TIME NEEDED

90 minutes

DISCUSSION

1. Why was it important to have the sand layer (sand stratum) between the layers of clay?
2. What are some processes that create fossils?
3. What sorts of fossils have you seen in a museum or collected yourself?
4. Think about how long it took you to make your 'fossils'. How does that relate to GEOLOGIC TIME.

> ALL OUR SCIENCE, MEASURED AGAINST REALITY, IS PRIMITIVE AND CHILDLIKE—AND YET IT IS THE MOST PRECIOUS THING WE HAVE.
>
> *Albert Einstein (1879–1955)*

Earth Connections No. 2

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Natural Resources Youth Camp

Natural Resources Youth Camp is a week of fun and adventure for ages 12–15 exploring the natural environment. Camp instructors are career professionals in fields such as wildlife and fisheries biology, resource management, soil science, and forestry. Young people thrive in this natural environment rich in opportunity for personal growth and group interaction as they actively learn about natural resources through a hands-on approach. Topics covered include aquatics, fish, wildlife, forestry, geology, human impact, ecosystems, stewardship, and careers. Camping, field trips, and other kinds of camp fun are also included.

The camp is held at Cispus Learning Center, 10 mi south of Randle, Wash., on the west side of the Cascades near Mount St. Helens. The facility is used year-round and includes heated dormitories, an education building, gymnasium, dispensary, leisure room, campfire amphitheater, ropes course, dining hall, and extensive trails system.

Camp runs from June 25–July 1. Cost for the week is $300; financial assistance is available. For more information, contact John Bergvall, 360-902-1027, john.bergvall@wadnr.gov, or Kathleen Rankin, 360-754-3588, ext. 114, kathleenrankin@juno.com.

Campers at the forestry site learn how to measure the size and volume of trees.

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Washington Bibliography Available on CD-ROM

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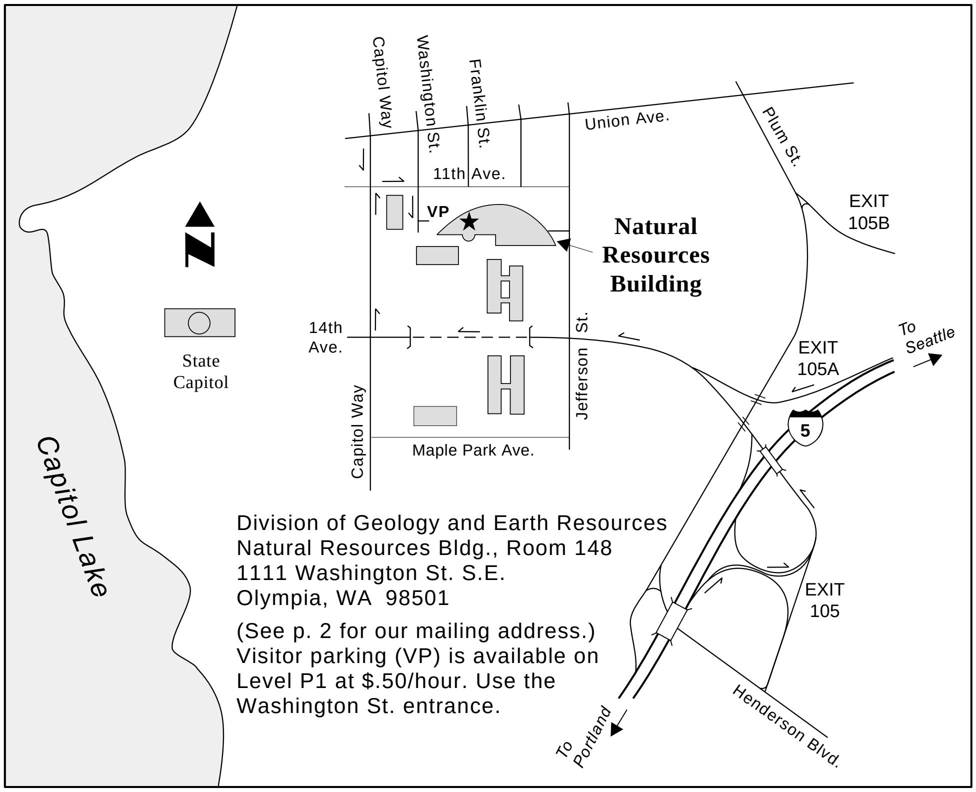

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(See p. 2 for our mailing address.) Visitor parking (VP) is available on Level P1 at $.50/hour. Use the Washington St. entrance.

STAFF NOTES *(continued from p. 40)*

Office Assistants

Kathy Loes came to work for us in August of 1999 after 15.5 years at Rite Aid in Tacoma. As personnel have changed within our support staff, she has moved into the Office Assistant Senior position.

Chandra Thomas is working half-time in Geology and half-time in the Agricultural Resources Division. She was previously with the Department of Transportation and the Olympia Chamber of Commerce.

Rice Museum of Rocks and Minerals Opens New Gallery

Rice Northwest Museum of Rocks and Minerals in Hillsboro, Ore., has opened a new gallery featuring specimens from Oregon, Washington, Idaho, British Columbia, Alberta, Northwest Territories, Yukon Territory, and Alaska. No other museum in the world features an exhibit specifically of the fine minerals found in the northwest corner of the North American continent.

Washington specimens showcased in the new gallery include superb specimens of molybdenite from Crown Point mine, Chelan Co.; autunite from Daybreak mine, Spokane Co.; quartz pseudomorphs after aragonite with calcite from Cadman quarry, Snohomish Co.; palygorskite and calcite with palygorskite from Pend Oreille mine, Pend Oreille Co.; staurolite from Snohomish Co.; laumontite from Lewis River, Lewis Co.; diopside from Vesper Peak, Snohomish Co.; boulangerite from Stevens Co.; amethyst from Skamania Co.; and natrolite from Weyerhaeuser Lincoln quarry, Lewis Co. The exhibit also includes a gold collection of 53 specimens, 37 of them from the northwest region, including several from Washington State.

For more information, see http://www.ricenwmuseum.org or contact Sharleen Harvey at Rice Museum; 26385 NW Groveland Road; Hillsboro, OR 97124 (503-647-2418).

STAFF NOTES

State Geologist **Ray Lasmanis** has been elected to the board of the Rice Northwest Museum of Rocks and Minerals in Hillsboro, Ore.

New Geologists

Karen D. Meyers is our new Geologist 2/Editor. She received her B.S. in geological and environmental sciences from Stanford University in 1994 and attended graduate school at the University of Arizona. Since then she has worked for the University of Arizona Library preparing documents for the Web; for Resource Science, Inc., Tucson, Ariz., as a Geographic Information Systems (GIS) technician and webmaster; for BHP Copper, Florence, Ariz., as a geologist/editor; for the Arizona Geological Society, Tucson, Ariz., as a web page designer; and for the National Geophysical Data Center, Boulder, Colo., as a research assistant. She will be starting with us full time in May.

Karl W. Wegmann is the new Geologist 2 with our environmental geology section. He received his B.A. in geology from Whitman College in 1996, where he did his honors thesis on the Precambrian geology of the Tobacco Root Mountains, Mont. That summer, he worked with Brian Atwater on Holocene paleoseismic research on the Washington Coast and Puget Sound. He got his M.S. in geology at the University of New Mexico in 1999 with thesis research on late Quaternary fluvial geomorphology and active tectonics of the Clearwater River Basin, Olympic Peninsula, western Washington. Prior to being chosen for this position, he was working with Hank Schasse on the geologic map of the Carlsborg 7.5-minute quadrangle in Clallam County.

Geology Interns

Andrew B. Dunn completed a B.S. in geology from Western Washington University in 1997. He is currently enrolled at the New Mexico Institute of Mining and Technology for an M.S. in groundwater hydrology. Andy is working with Bill Lingley to help define significant gravel and rock resources in the state. They are also working on the geology of the Shelton 1:100,000 quadrangle.

Brian D. Evans received his B.S. in geological science from the University of Washington in 1991. After graduation, he worked as a staff geologist at an environmental consulting firm, performing well drilling, soil and water sampling, and report writing/data analysis. At DGER, he is working with Steve Palmer producing liquefaction susceptibility maps for the greater Eastside (Seattle area) and Tacoma.

Thomas J. Lapen received his B.S. from Central Washington University in 1995. In 1998, he received an M.S. in structural geology and petrology from Western Washington University. That summer, Tom volunteered with DGER and worked with Joe Dragovich mapping the Skagit River valley. Since August, he has been compiling and mapping geology for the Bellingham 1:100,000 quadrangle.

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DIVISION PUBLICATIONS

New Releases

Geologic Map of the Sedro-Woolley North and Lyman 7.5-minute Quadrangles, Western Skagit County, Washington, Open File Report 99-3, by Joe D. Dragovich, David K. Norman, Tom Lapen, and Garth Anderson. 37 p. text, 4 plates (maps, cross sections, correlation diagram), scale 1:48,000. $4.17 + .33 tax (Wash. residents only) = $4.50.

Geologic Map of the Easton Area, Kittitas County, Washington, Open File Report 99-4, by Eric S. Cheney. 11 p. text, 1 plate, scale 1:24,000. $1.39 + .11 tax (Wash. residents only) = $1.50.

Preliminary Bibliography and Index of the Geology and Mineral Resources of Washington, 1998, Open File Report 99-5, compiled by Connie J. Manson. This 110 p. report lists 422 items issued in 1998 and 413 items issued prior to 1998 that were not included in earlier compilations. $3.71 + .29 tax (Wash. residents only) = $4.00.

Preliminary Geologic Map of the Spokane NE and SE 7.5-Minute Quadrangles, Spokane County, Washington, Open File Report 99-6, by Robert E. Derkey, Michael M. Hamilton, Dale F. Stradling, and Eugene P. Kiver. 2 plates (maps, cross sections, correlation diagram), scale 1:48,000. $1.85 + .15 tax (Wash. residents only) = $2.00.

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New Report by DGER's Weldon Rau and Sam Johnson of the U.S. Geologic Survey

Well Stratigraphy and Correlations, Western Washington and Northwestern Oregon, U.S. Geological Survey Geologic Investigations Series I-2621, by Weldon W. Rau and Samuel Y. Johnson. This excellent report is a unique summary of the stratigraphy encountered in many important, deep, oil-and-gas exploration wells drilled in the Puget–Willamette trough and Grays Harbor basin. It contains ranges charts for microfauna (2 plates), a stratigraphic cross section with lithologic descriptions and wireline log correlations (1 plate), and a 31 p. text that will undoubtedly become a key reference for the subsurface geology of western Washington and Oregon. Rau's interpretations incorporate more than 50 years of petrologic and biostratigraphic experience in Washington, and Johnson has added valuable interpretations of wireline logs and other information. This report, which is particularly useful for tectonic studies, deep hydrologic investigations, and natural gas exploration, is available from the USGS Information Services; Box 25286, Denver Federal Center; Denver, CO 80225; 303-202-4700.

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