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The Hebgen Lake, Montana Earthquake of August 17, 1959

GEOLOGICAL SURVEY PROFESSIONAL PAPER 435

Cooperating Agencies:
Geological Survey, U.S. Department of the Interior
National Park Service, U.S. Department of the Interior
Coast and Geodetic Survey, U.S. Department of Commerce
Forest Service, U.S. Department of Agriculture

HEBGEN LAKE EARTHQUAKE OF AUGUST 17, 1959

MADISON SLIDE AND EARTHQUAKE LAKE, AUGUST 21, 1959

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FOREWORD

Late on the night of August 17, 1959, southwestern Montana and adjacent areas were shaken by a major earthquake, the fourth within 35 years and the strongest so far recorded in the State. The disturbance centered near Hebgen Lake on the Madison River a few miles northwest of Yellowstone National Park. The magnitude of the earthquake was 7.1 (Pasadena); the shock was felt throughout an area of 600,000 square miles and as far as 350 miles from the epicenter.

Ground displacement accompanying the earthquake in the epicentral area was large and, because of the presence of the lake with more than 50 miles of shoreline, it was measurable to an extent not often possible in studies of earthquake deformation. Eighteen miles of major fault scarps were produced, mainly on two faults along one side of the lake, and several miles of smaller and less continuous scarps were formed in unconsolidated deposits around the lake. The entire lake basin subsided, locally as much as 22 feet, and its shape was changed, causing violent water waves during the earthquake and a permanent displacement of the water surface relative to the former shore. Precision level lines run by the U.S. Coast and Geodetic Survey shortly after the earthquake indicate that measurable changes in ground altitude occurred throughout an area of at least 200 square miles.

In addition to ground deformation, the main shock touched off dozens of landslides throughout the meizoseismal area. Most were rockfalls and rock avalanches from steep rocky slopes; others were landslides into Hebgen Lake. One, in the canyon of the Madison River a few miles northwest of the lake, carried 37 million cubic yards of broken rock into the canyon, interrupting the flow of the river there for 3 weeks, and forming a lake 6 miles long and 190 feet deep at its deepest point. This was one of the three largest rapid landslides that have occurred in North America within historic time, and the only one of these resulting from an earthquake.

Because these events took place in a popular vacation area during the height of the tourist season, they were witnessed by an unusually large number of people. Many eye-witness accounts were published in the newspapers in the days immediately following the earthquake, and illustrated descriptions were published in the periodicals Life (August 31, 1959), Geotimes (October 1959), The Geographical Review (January 1960), and the National Geographic Magazine (March 1960). Many people were camping out or occupying cabins along the lake and in the canyon downstream and experienced the full force of the earthquake. A few, unfortunately, were in the path of the great Madison Slide; 7 persons are known to have perished, and 19 are still missing and are presumed dead. Two more people were killed by rockfalls 15 miles west of Hebgen Lake.

Two field projects of the Geological Survey were in progress in the region: one in the Hebgen Lake area itself, and the other near Ennis, Mont., 50 miles to the northwest. Because of the unusual opportunity as well as the need to learn the meaning of the earthquake in terms of the geologic history of the region, a detailed investigation, organized by Charles B. Hunt, of the geologic aspects of the earthquake was undertaken by the Survey.

First on the scene were I. J. Witkind and J. B. Epstein, who were engaged in the second season of geologic mapping and stratigraphic and structural studies, and whose field camp was only a quarter of a mile away from an active fault. They were soon joined by J. B. Hadley, W. B. Myers, Warren Hamilton, W. H. Hays, W. D. Quinlivan, W. H. Nelson, G. D. Fraser, K. B. Ketner, and G. M. Richmond, geologists; J. R. Stacy, photographer; and W. H. Jackson, geophysicist. Witkind and Epstein traced and studied the newly formed scarps northeast of

Hebgen Lake and continued their previous mapping in the faulted area. Myers, Hays, and Hamilton studied the lake basin; and Hamilton and Richmond mapped in reconnaissance the unmapped southern part of the earthquake area. Hadley, assisted by W. D. Long, studied the Madison Slide and other landslides and, together with Nelson, Fraser, and Ketner, mapped in reconnaissance the northwestern part of the earthquake area. Stacy spent a week with the geologists taking many of the photographs used in this report. Jackson, assisted by Hays and others, made a survey of the bottom of Hebgen Lake by means of acoustic depth soundings. These investigations are reported in parts 1 and 3 of this Professional Paper.

At the same time, changes in the behavior and quality of ground water and surface water related to the earthquake were studied by geologists and hydrologists of the Geological Survey. After the earthquake, the Madison River and its tributaries in the immediate area, as well as many more remote streams, flowed for some time with increased volume. Frank Stermitz collected and studied records of these changes, made at previously established gaging stations of the Survey and at some temporary ones established for the earthquake study. T. F. Hanly investigated the closely related subject of increased sediment load in these streams, and F. A. Swenson investigated the remarkable effects of the earthquake on springs, sand spouts, and other ground-water phenomena. Seismic waves from the main earthquake and its aftershocks caused fluctuations in water wells throughout a large part of the conterminous United States and as far away as Alaska and Hawaii; data on these fluctuations are reported here by J. A. da Costa. Because many springs in the earthquake area became turbid, it was suspected that the water might have been affected chemically; therefore, R. C. Scott investigated the chemical composition of the water from springs in the vicinity of Hebgen Lake and from hot springs in nearby parts of Yellowstone National Park. G. D. Marler of the National Park Service studied notable earthquake-induced changes in the temperature and discharge of the hot springs and geysers of Yellowstone National Park. Reports of these hydrologic studies constitute part 2 of the present publication.

The U.S. Coast and Geodetic Survey, as part of its normal program of monitoring earthquakes in the United States, made seismic records and preliminary determinations of epicenters for the main shock and the principal aftershocks and calculated the distribution of surface intensity. A preliminary report of these data is included as chapter C of the present report. In addition, a leveling party working in a nearby area at the time of the earthquake was detailed to run precise level lines in the Hebgen Lake area a few weeks after the earthquake. The results of this survey were of great value in the study of the structural effects and have been extensively used in chapters I and J.

Major contributions to the earthquake investigation were also made by the U.S. Forest Service in furnishing excellent new aerial photographs of the area within a week after the earthquake and in preparing a detailed photogrammetric map of the Madison Slide within 2 weeks. Several helicopter flights by Survey geologists were also provided through cooperation with the Forest Service.

These studies of the Hebgen Lake earthquake and of the attendant phenomena afford a rare glimpse into the dynamic processes that have been shaping the rock structure and landscape of southwestern Montana and adjacent areas.

Thomas B. Nolan

Director

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Events on the Night of August 17, 1959—the Human Story

By IRVING J. WITKIND

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

GEOLOGICAL SURVEY PROFESSIONAL PAPER 435-A

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EVENTS ON THE NIGHT OF AUGUST 17, 1959—THE HUMAN STORY

By Irving J. Witkind

ABSTRACT

The Hebgen Lake area and the nearby canyon of the Madison River are popular tourist localities, and on the night of August 17, 1959, almost every campsite was occupied. At 11:37 p.m. an earthquake jarred most of the campers awake. A few panicked, and in trying to escape from one danger, faced another. Most of the campers, however, showed unusual courage and calmness. With few exceptions, the people near the lake spent the rest of the night either around campfires or in their cars. Those in the confines of the canyon downstream from Hebgen Dam moved to high ground to await rescue. Despite serious damage, the dam remained standing and averted greater disaster. Nevertheless, 28 persons were dead or missing when the earthquake was over.

ACCOUNT OF EVENTS

The Hebgen Lake area, an attractive tourist site near Yellowstone National Park, is crowded with visitors during the spring and summer months. Many summer homes have been built in wooded patches along the south shore of the lake; and motels, dude ranches, campgrounds, and trailer parks line Montana State Route 499 (former State Highway 287) along the north shore. Below Hebgen Dam, where the highway follows the Madison River through a narrow canyon with steep towering walls, the U.S. Forest Service has built and maintained a series of campgrounds, among them the Rock Creek Campground.

During the early and middle parts of August 1959 the weather was nearly perfect—cool sunny days and clear calm nights. On the night of August 17 the moon was full. Tourists crowded the area, so much so that late travelers through the Madison River canyon could find no available site at any of the campgrounds and reluctantly continued on.

The first earthquake shock came at 11:37 p.m., after most of the residents and tourists were abed. It is uncertain just how long the earth trembled—estimates range from about 5 seconds to as much as several minutes—but those who experienced the violent shaking will certainly never forget it. Some nearly panicked; others showed great courage.

Our Geological Survey camp consisted of two house trailers parked on a small hill near the Blarneystone Ranch. Mr. and Mrs. J. B. Epstein occupied one trailer; I occupied the other. A day after the earthquake, Mrs. Epstein wrote this to her friends:

> All of a sudden the trailer began to shake violently up and down and back and forth. I thought at first that Jack was fooling around and shaking the trailer, but in a split second I looked around and saw:
>
> 1. Water pouring out of the wash basin.
> 2. All dishes, groceries, and clothes falling out of the cabinets.
> 3. The gasoline lantern hanging from the ceiling, swinging in a 2-foot circle, and looking as if it would fall any minute. If it did it would have set the whole trailer on fire.
>
> There were fantastic rumblings. The farthest thing from my mind was an earthquake. In this same split second I thought that the 100-pound propane tank outside the trailer was starting to explode and that's what caused the noise and shaking. In pure horror and fright I dashed out the door and screamed for everyone to follow and run as far away from the trailer as possible. Jack was still in the trailer, trying to stop the lantern. He got beaned on the head with it, gave up, and came charging out. He had realized from the first that it was a quake. My complete horror came after I hit the ground and found that it was no better than in the trailer. The solid earth, "terra firma," was like a glob of jelly. I was frantic—there was nowhere to get away from the fantastic sensation. Jack screamed not to run near the woods because trees were toppling all over. We could hear loud rumblings due to rockslides and landslides in the mountains.

At the time of the earthquake I was asleep in my trailer. As I later wrote to my wife,

> "I went to sleep about 9:30 p.m. and was awakened by the frenzied jiggling of the trailer. Things were falling from shelves all over the place. I thought that the trailer had somehow come off its jacks, jumped the chocks, and was rolling down the hill. I scrambled out the front door determined to stop the trailer, no matter what, although I had no idea as to how I would go about it. When I got outside, the trailer was in place, but the trees were whipping back and forth and the leaves were rustling as if moved by a strong wind—but there was no wind. I knew right then that it was an earthquake. I could hear avalanches in the canyons behind me, and could see huge clouds of dust billow out of the

canyon mouths. Jack Epstein, who was awake at the time of the earthquake, says he heard a deep rumbling in the earth. I noted the time of the first shock, and kept track of the major aftershocks for about 20 minutes before I decided to visit the Blarneystone Ranch and see if they needed help. I drove down the hill toward the ranch. About a quarter of a mile from camp I came upon a large new fault scarp that cut across and displaced my access road.

The Blarneystone Ranch was severely damaged (fig. 6). When John Russell discovered that the doors of his apartment there were jammed shut and that he and his family were trapped, he broke a window and used it as an exit. Others in the main dwelling made their way to safety through once-orderly rooms that were now a scene of displaced and overturned furniture, fallen pictures, and plaster.

The Parade Rest Ranch, about half a mile to the south, was also damaged, although not as severely. Mr. Wells Morris, Jr., the owner, was suddenly awakened by the distinct sensation that his bed was falling from beneath him in short spasmodic jerks.

Also awakened by the quake, a family in a motel on U.S. Highway 191 hastened into their car and raced southward, trying to escape, only to drop off a new fault scarp which crossed the highway about 500 feet from the motel. The car turned over and was demolished; the family returned to the motel, unharmed.

A rancher hurrying to help his neighbor drove off a new scarp about 100 feet from his house. His car remained on end all night.

Shortly after the earthquake, the occupants of a small house trailer parked near the north shore of Hebgen Lake discovered to their horror that the lake which had been 50 feet away was now swirling around the trailer, and rising. The couple made their way out, waded through the water to higher ground, and then watched the trailer float away. The next morning only the open hatch in the rooftop could be seen, some 100 feet offshore.

At Hilgard Lodge, about a mile southeast of Hebgen Dam, the earthquake and the accompanying surges of the lake wreaked havoc (fig. 9). The main dwelling and a group of connected motel units were built on a broad sloping alluvial cone at the northeast edge of the lake—a sector of ground that dropped about 10 feet during the earthquake and was broken by many small scarps and gaping fissures. All the buildings were tilted and knocked askew, then lifted and dropped by a wave of water that moved toward the dam. Mrs. Miller, the owner, was asleep at the time of the quake. Suddenly aware that her house was sliding into the lake, she scrambled out with great difficulty and, dazed and shocked, stumbled across scarps and fissures in a rough sagebrush pasture to the Kirkwood Ranch, more than a mile away.

George Hungerford, foreman at Hebgen Dam, and Lester Caraway, his assistant, were awakened by the major shock and within moments recognized it as an earthquake. With their wives, they hurried to a water gage downstream from the dam to see if the river flow showed that the dam was leaking. As they neared the gage, Hungerford heard a roar. He glanced up to see a wave of water about 4 feet high moving down the river. Fearing that this meant the collapse of the dam, he returned to his house on the highway above the gage and tried to telephone a warning, but the line was dead. The two couples then drove toward the high ground near the dam and arrived there at about 11:55 p.m.

The moon was obscured by dust, and it was very dark. The water had withdrawn from sight, but they noticed that the downstream side of the dam was wet. Then, before they could see it, they heard water again; it was coming down the lake. They climbed out of the way and watched the water rise, overtopping the dam by about 3 feet. After 5 or 10 minutes it receded, then disappeared from sight. "All we could see down the dam was darkness again," Hungerford recalls.

The crest of the dam was again submerged in 10 or 15 minutes, but this time by less water, and the water receded sooner. In all there were four surges over the crest. Between them, Hungerford and Caraway could see no water on the upstream side, even once when they ran out onto the dam. The water in Hebgen Lake had been sloshed about like water in a bathtub, and it continued to oscillate, though less violently, for at least 12 hours after the quake (Myers and Hamilton, chapter I).

Many of the campers downstream from the dam were slow to realize what had happened. They woke and looked about them, confused, but the aftershocks alerted most of them to the danger. Fearing that the dam would break, they abandoned their trailers and fled by car to higher ground.

Among them was the Lewis Smith family of Greeley, Colo., who were sleeping in a small house trailer in the Beaver Creek Campground, about 2 miles downstream from Hebgen Dam. Awakened by the violent shaking of the trailer, Mr. Smith called to a neighbor and asked what had happened. An earthquake, he was told—and the dam was not far upstream. Smith decided to evacuate his family and leave the trailer behind.

When he reached State Highway 499 he had two choices. He could turn northeast and drive toward the dam for about 2 miles to the high ground on which the dam was built, or he could turn southwest and drive down the Madison River canyon for about 5 miles and then leave the confines of the canyon for the open country west of the Madison Range. Believing that the dam would break at any moment and unwilling to risk driving toward it, Smith turned southwestward, down the canyon. Three and a half miles from the campground a huge boulder blocked the road, but Smith was still unwilling to drive the other way. The family left their car and climbed to higher ground, where they spent a cheerless night. Boulders were crashing down the mountainside, but not until daybreak did they realize that the route was blocked by a great landslide.

Early the next morning, Smith returned to his car and listened to a radio broadcast in which the collapse of Hebgen Dam was declared imminent. He drove his car part way up a small hill, then rejoined his family. The Smiths were rescued by helicopter that afternoon; their car was gradually inundated as the river, penned up behind the Madison Slide, rose to form Earthquake Lake (frontispiece).

For some travelers the night of August 17 was more tragic. Mr. and Mrs. F. R. Bennett and their four children, en route from their home in Coeur d'Alene, Idaho, to Yellowstone National Park, camped that night in the Madison River canyon near Rock Creek Campground, which was later partly buried by the Madison Slide. The parents were sleeping in their small house trailer, and the children were in bedrolls on the ground nearby when Mrs. Bennett was awakened by a loud noise. "Some time later" she heard a great roar and, alarmed, went with Mr. Bennett to check on the children. Just as they left the trailer a tremendous blast of air struck them. Mrs. Bennett saw her husband grasp a tree for support, then saw him lifted off his feet by the air blast and strung out "like a flag" before he let go. Before she lost consciousness she saw one of her children blown past her and a car tumbling over and over. Her son Phillip, 16 years old, was buffeted about by the air blast and immersed in water, but somehow, with a broken left leg, he managed to crawl into a clump of trees, where he burrowed into the mud for warmth. He and his mother, sole survivors of the family, were rescued the next morning.

Rev. E. H. Ost and his family were among the survivors at Rock Creek Campground. Awakened by the first tremor, the family left their tent. As they stood in the bright moonlight, about 20 seconds after being aroused, they heard a tremendous grinding noise and, with it, the sound of water. No wave of water was moving downstream; instead, Rev. Ost saw water moving upstream. He shouted to his family to hang onto trees. His daughters, alerted by the call, scrambled upslope. Water swirled through the campground, rolling and tumbling their car about 50 feet upstream, but Rev. and Mrs. Ost held firm to the trees and soon were able to pick their way out of the debris. With other survivors, they climbed to higher ground. Dawn revealed the east edge of the Madison Slide about 100 yards from the Osts' former campsite. By 6:00 a.m. all the cars in the campground were submerged beneath Earthquake Lake.

The house trailer of an elderly couple, Mr. and Mrs. Grover Mault, of Temple City, Calif., was carried some 200 feet upstream in the wave created by the Madison Slide. The water rose rapidly as the Maults tried to escape from the trailer, so rapidly that by the time they made their way to the door the trailer was almost completely submerged. They managed to get out and climb first onto its roof and then, as the trailer was inundated, into a nearby pine tree. The ever-rising water forced them to climb higher. Several times, when the boughs broke under their weight, they fell into the lake. Each time they climbed back. Just as daylight was breaking, they were rescued, after 5 hours in the tree.

At Ennis, Mont., about 50 miles downstream from Hebgen Dam, the major tremor aroused many of the residents. A few went outside to see what had happened; most merely went back to sleep. At about 2:45 a.m. a message reached Ennis that the dam had failed or was about to fail, and the authorities immediately began to alert the residents. Sirens sounded, and people went from house to house warning their neighbors and friends. The word swept through town. All residents were urged to move to higher ground as soon as possible. Most took a few prized possessions and drove to a high terrace at the west edge of town. By 4:00 a.m. Ennis was nearly deserted, and as the night passed many residents moved to still higher ground farther west. By dawn a large field in this new area was covered with cars.

The threat of flood still hung over Ennis on August 18, and by midmorning roadblocks had been established on State Highway 499 leading to the Madison River canyon. As the threat lessened, many residents wished to return to their homes, but the town officials, fearful that Hebgen Dam still might collapse, issued an official evacuation order to prevent their return. Most of the populace spent the rest of the day and that night camped out. The order was retracted on the afternoon of the 19th, and the residents returned to Ennis on the basis of a standing alert. The warning signal was to be the sounding of the siren. This was never used, for Hebgen Dam held firm.

At West Yellowstone, which was much closer to the epicentral area and therefore more intensely affected by the major shock than Ennis, the fear of additional severe shocks prevailed. Most of the residents left their dwellings at the time of the major shock and got into their cars. An amateur radio operator radioed: "The pavement looks like it is coming toward me in waves a foot high." The aftershocks kept people away from buildings, and many spent the night in their cars. In the early morning large numbers of tourists left the area via U.S. Highway 20 to Idaho—the only route not blocked.

The earthquake was felt in most of the surrounding communities as a moderate rolling and pitching motion. At Bozeman, Mont., about 60 miles north of the epicentral area, the major tremor jostled the community awake, and some residents heard a low rumbling roar in the ground. A few of the more curious went outside; most of the others, satisfied that their lights and telephones were still working, went back to sleep. At Butte, Mont., about 100 miles northwest of Hebgen Lake, the motion reminded one reporter of "negotiating a rough patch of water in a small boat." The tremor was first felt as a slight shaking which worsened and lasted about 30 seconds. At Ashton, Idaho, about 55 miles south of Hebgen Lake, most of the populace was shaken awake but no one was injured, nor was there any serious structural damage.

Structural Damage in the Hebgen Lake-West Yellowstone Area

By IRVING J. WITKIND

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

STRUCTURAL DAMAGE IN THE HEBGEN LAKE-WEST YELLOWSTONE AREA

By Irving J. Witkind

ABSTRACT

The area of maximum destruction during the Hebgen Lake earthquake was along the northeast shore of Hebgen Lake near the reactivated Red Canyon and Hebgen faults. Such destruction is commonly attributed to intensities of IX and X (Modified Mercalli scale). Damage diminished progressively away from this intensely shaken area, and West Yellowstone, 10 miles away, was probably on either the VI or VII isoseismal.

In the epicentral area every dwelling was damaged. Masonry buildings were cracked or destroyed. Chimneys fell, plaster was cracked and broken, and some large plateglass windows were broken. Highways were disrupted, and segments slid into the lake; bridges were knocked askew. Cisterns and water lines were broken and clogged. A few streams were displaced into new channels. Much pastureland was inundated and, on some alluvial flats, numerous fissures made the ground almost useless for grazing.

INTRODUCTION

The area of maximum destruction in the earthquake of August 17, 1959 (the meizoseismal area), was confined to the northeast shore of Hebgen Lake, adjacent to two major fault structures in the bedrock, the Red Canyon and Hebgen faults (chapter G). This was also the area of maximum subsidence (chapter I). Most of the north shore was marred by destruction commonly attributed to intensities between IX and X on the Modified Mercalli (Wood-Neumann) scale (chapter F). In this area, every manmade structure was damaged. Frame buildings were thrown out of plumb or shifted off their foundations. Masonry buildings were damaged and partly collapsed. Heavy furniture was overturned. The ground was cracked, and buried pipes were broken. Highways were disrupted and bridges knocked askew. Fences were bent, broken, and twisted; springs were clogged and cisterns broken, and the water-supply pipelines on many ranches parted in innumerable places. Some pastureland was inundated; other excellent pasture was made useless by the presence of fissures, which were avoided fearfully by all grazing animals.

Away from this intensely shaken area, the damage diminished progressively. Along the south shore of the lake, the major damage was to chimneys, which were shaken down. West Yellowstone, about 10 miles away from the reactivated faults, was slightly to moderately damaged and probably was on either the VI or VII isoseismal. Here the earthquake was felt by all, and most people ran outdoors. Many chimneys were broken and collapsed, plaster fell, some large plateglass windows broke, but in general the damage was slight.

Considering the magnitude of the earthquake (7.1—Pasadena), the amount of damage, although extensive, does not seem as great as it could have been. This is directly attributable to the sparsity of dwellings. If the major shaking had occurred in a crowded residential area, the damage would have been devastating.

HIGHWAYS

Highways in the epicentral area were damaged extensively, chiefly due to compaction of the roadbed with resultant breakage and offset of the asphalt surface (fig. 1). All roads were broken by fractures, many of which were normal to the roadway. Offset along these fractures was mainly vertical and measurable in inches, but in several localities the offset of the paving due to slump of the roadbed created an impression of minor horizontal displacement. Near their junction, U.S. Highway 191 and State Highway 499 (SE¼ sec. 16, T. 12 S., R. 5 E.) were displaced vertically as much as 5 feet by fault scarps. At three localities—(a) E½ sec. 31, T. 11 S., R. 4 E.; (b) W½ sec. 25, T. 11 S., R. 3 E.; and (c) center sec. 23, T. 11 S., R. 3 E.—landslips carried parts of State Highway 499 into Hebgen Lake (fig. 2). Near Kirkwood Ranch the road was inundated for about 100 yards (SE¼ sec. 25, T. 11 S., R. 3 E.). Earthquake Lake, which formed behind the Madison Slide, submerged 3 miles of State Highway 499, and an additional mile of the route is buried by the slide (frontispiece; fig. 54).

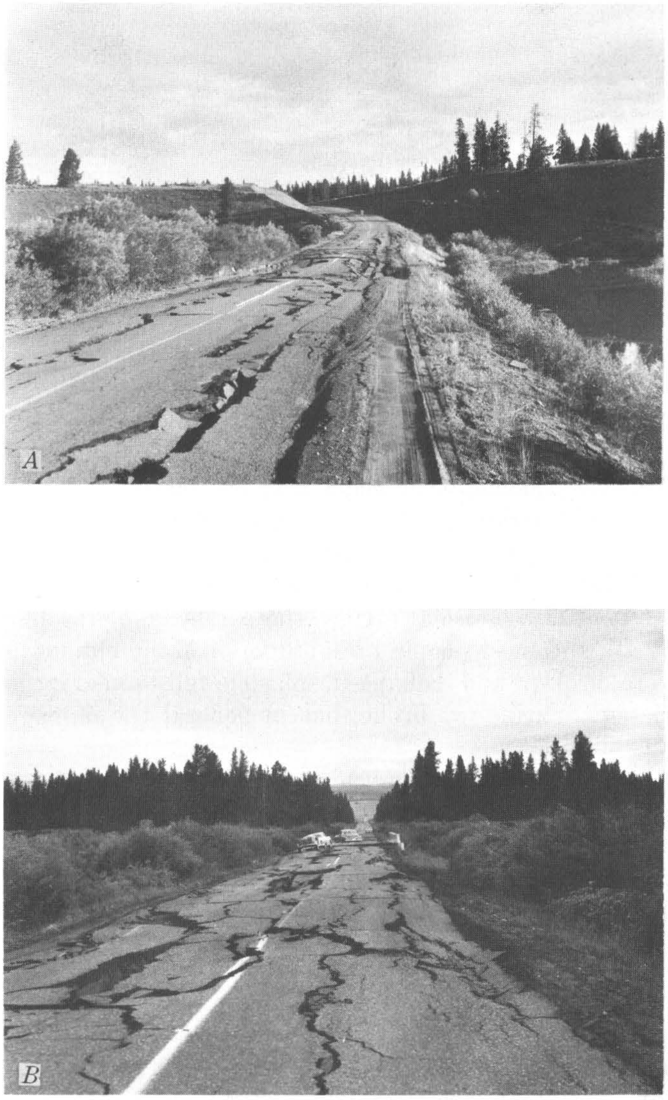

FIGURE 1.—Damage to highways in the epicentral area. *A*, Shattered remnants of asphalt-surfaced U.S. Highway 191 where it crosses the alluvial fill of the Madison River; *B*, disruption of U.S. Highway 191 near Cougar Creek.

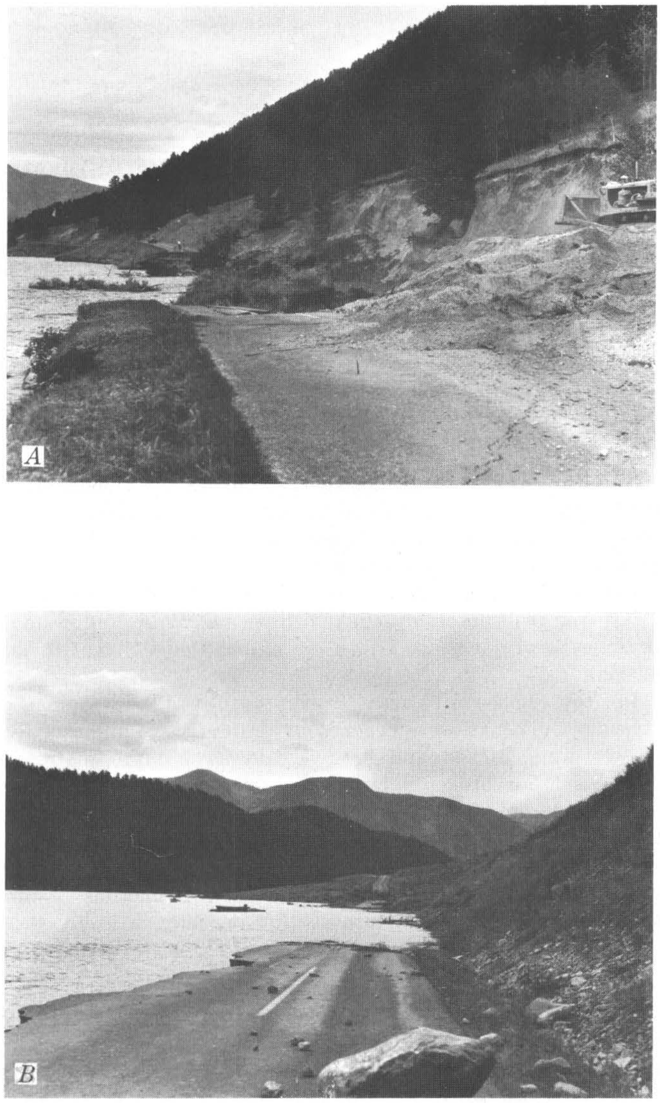

FIGURE 2.—Landslips of State Highway 499 along north shore of Hebgen Lake. *A*, Landslip at E½ sec. 31, T. 11 S., R. 4 E.; *B*, small landslip and submergence at W½ sec. 25, T 11 S., R. 3 E.

Slump of an adjacent hillside caused a segment of Highway 499 to buckle and break in the E½ sec. 12, T. 12 S., R. 4 E. (fig. 77). About 1½ miles to the southeast (NW¼ sec. 17, T. 12 S., R. 5 E.), slight southward movement of a huge ancient earthflow bowed the centerline of Highway 499 as much as 2 feet out of line.

BRIDGES

Bridges remained intact, but concrete guardrails on many were broken and separated, chiefly between adjacent segments (fig. 3). Gaps of 2 to 3 inches appeared in the floors of some bridges along expansion joints. Abutments shifted slightly or stood free of adjacent fill, which was compacted by lateral oscillation of abutments during the earthquake. Most bridges in the epicentral area were 6 inches to 2 feet above the former road surface, presumably the result of unequal settling of bridge and road during the earthquake (fig. 3).

TRAILS

After the earthquake, trails in the epicentral area were passable only with great difficulty. Several were cut and displaced by the Red Canyon or Hebgen fault scarps (chapter G). Most were covered locally by extensive rockslides, and the Kirkwood Creek Trail (Trail 210) was partly destroyed during movement of a large earthflow (chapter K, fig. 70). All trails were cut by gaping fissures, and wherever the trails extended along ridgetops they were broken by fractured

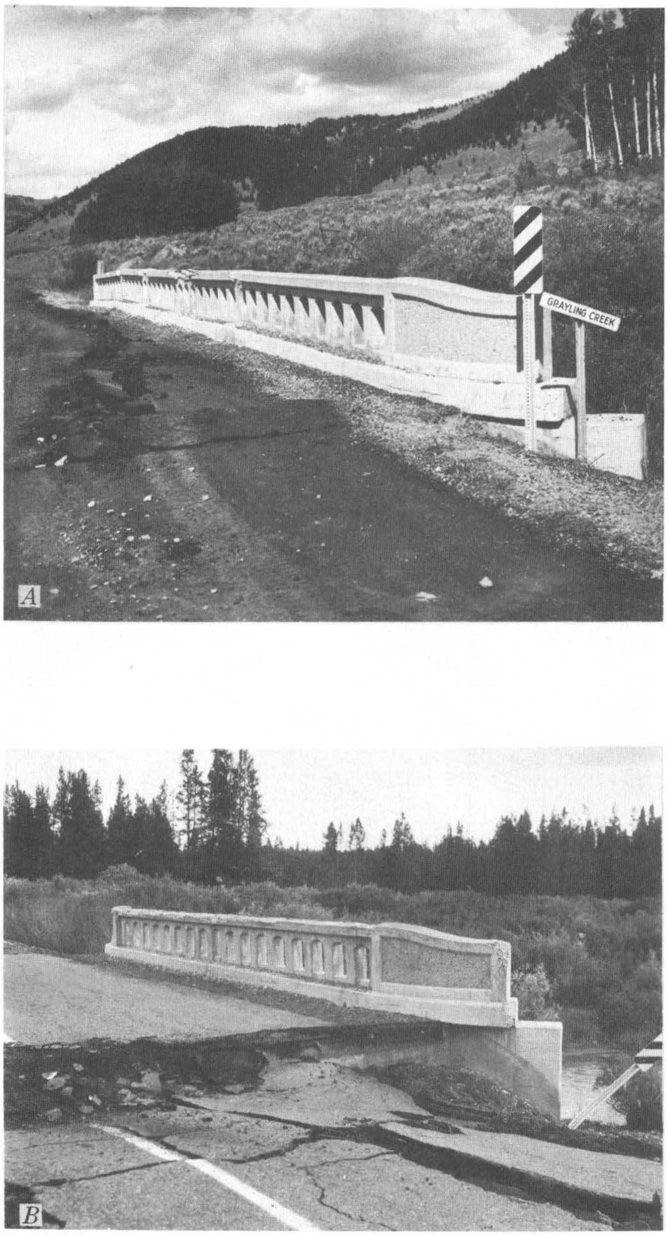

FIGURE 3.—Damage to bridges in epicentral area. *A*, Buckling of floor and separation of concrete guardrail on bridge across Grayling Creek; *B*, damage to road and bridge across Madison River.

ground (chapter G). Where the trails passed through wooded tracts, downed trees were a major obstacle.

BUILDINGS

All buildings in the Hebgen Lake area were damaged, some of them so drastically as to make them unfit for further occupancy. Most buildings were constructed on surficial material, either glacial outwash as in West Yellowstone, or alluvial fill as along the north shore of Hebgen Lake. Only a few buildings near the dam were on bedrock.

FIGURE 4.—Collapse of chimney at the Union Pacific Dining Lodge, West Yellowstone, Mont.

In general, wooden structures were not as severely damaged as those built of masonry. The more rigid masonry buildings fractured, broke, or collapsed. Much of the damage to hotels, depots, and dwellings in both West Yellowstone and Yellowstone National Park resulted from falling chimneys. At Old Faithful Inn the main chimney fell through the roof, and the same thing happened at the Union Pacific Dining Lodge at West Yellowstone (fig. 4). Apparently many masonry walls fell because they were not adequately tied to interior walls of buildings. At the West Yellowstone School, the brick veneer along one entire wall fell, and so did the decorative stone wall flanking the main entryway (fig. 5).

FIGURE 5.—Damage to elementary school at West Yellowstone, Mont. Decorative stone entryway has been shaken down.

Plate-glass windows on several stores in West Yellowstone were shattered, and plaster was broken in all homes. Dishes, pottery, lamps, and bric-a-brac fell. Canned goods and bottles stacked on the shelves of several markets tumbled to the floor.

Blarneystone Ranch, in the SW¼ sec. 9, T. 12 S., R. 5 E., sustained the most severe damage in the epicentral area (fig. 6). The Red Canyon fault scarp cuts across the ranch and passes beneath the northern part of the main group of buildings (fig. 6*A*). The domestic quarters and storage sheds, which were astride the scarp, were completely demolished (fig. 6*C*); other buildings in the group were dropped 10 to 12 feet and consequently were severely shaken and shattered (fig. 6*B*). Their foundations cracked and were locally displaced or caved (fig. 6*D*). Chimneys fell, and masonry supports for the entryway were cracked and tilted. The entire main dwelling was tilted northward toward the scarp, and no room inside was free from damage. Large pieces of furniture moved about or toppled. Pictures fell, as did large quantities of stored canned goods. Most doors were jammed, but surprisingly many of the large panes of window glass were undamaged.

Most of the stables, corrals, storage sheds, and garages were on the downthrown block. Large parts of the corrals were flooded by Grayling Creek, and fences were splintered, displaced, and broken. At the

FIGURE 6.—Damage to Blarneystone Ranch (SW¼ sec. 9, T. 12 S., R. 5 E.). *A*, Red Canyon fault scarp; block of ground on left (south) has dropped; collapsed culvert crosses Grayling Creek. *B*, The main group of buildings on the downdropped block; view is from the top of the fault scarp. *C*, Buildings astride the Red Canyon fault scarp; scarp extends from lower left edge of photograph toward center; left side dropped 10 to 12 feet. *D*, Damage to foundation of main dwelling.

time of the earthquake all livestock was on the downthrown block. They ran amuck, and many received bad gashes as they ran into barbed wire fences. All barbed wire, which had been tautly drawn between steel fenceposts, was loosened and sagged. In places the wire appeared to have been stretched; elsewhere the steel fenceposts had been pulled from the ground. Access roads to the ranch were broken, and the culvert across Grayling Creek collapsed (fig. 6*A*).

Parade Rest Ranch, about a quarter of a mile southwest of Blarneystone Ranch, was jostled severely, and damage was extensive. Several well-constructed log buildings were tilted; one was knocked askew on its foundation. A water tower about 75 feet high near the ranch was surprisingly stable. It consists of a wooden trestle on which rests a large water tank of about 500-gallon capacity. Although the tower must have swayed precariously, the water tank was not dislodged, nor was the tower damaged.

Duck Creek Motel at the junction of U.S. Highway 191 and State Highway 499 (NW$\frac{1}{4}$ sec. 22, T. 12 S., R. 5 E.) is near the Red Canyon fault scarp but was only moderately damaged. Most of the damage was confined to the main residence; its chimney fell on the concrete roof of an attached garage.

The residence of Mr. Robert Hurless, in the NE$\frac{1}{4}$ sec. 11, T. 12 S., R. 4 E., on the downdropped block, was one of the few homes that was but slightly damaged. The house is well constructed of wooden logs on alluvial fill. Part of the masonry chimney was cracked, and the earthfill about the foundation was compacted, forming a gap of about a quarter of an inch between it and the foundation. The foundation was slightly cracked in one place. In the house, a television set and other large pieces of furniture were thrown over, lamps fell, and dishes were broken.

Motels and resorts along the north shore of Hebgen Lake were severely damaged. Much of the shoreline was inundated, and, consequently, buildings were flooded and docks submerged (fig. 7). The ground nearby slumped in places and was broken by gaping fractures (fig. 8). Some motel units were tilted toward the lake. Terrain was modified slightly, and shallow sags appeared where none had existed before.

Mr. and Mrs. Frank Jans, owners of The Narrows Motel in the E$\frac{1}{2}$ sec. 10, T. 12 S., R. 4 E., agree that the land slopes lakeward more steeply than before, and Mr. Jans estimates that the drop was about 10 inches in 20 feet. He reports that long gaping cracks as much as 3 inches wide appeared in the soil the night of the earthquake and closed the next morning; also, that a flowerbed beside the house was compacted from a width of about 18 inches to about 8 inches.

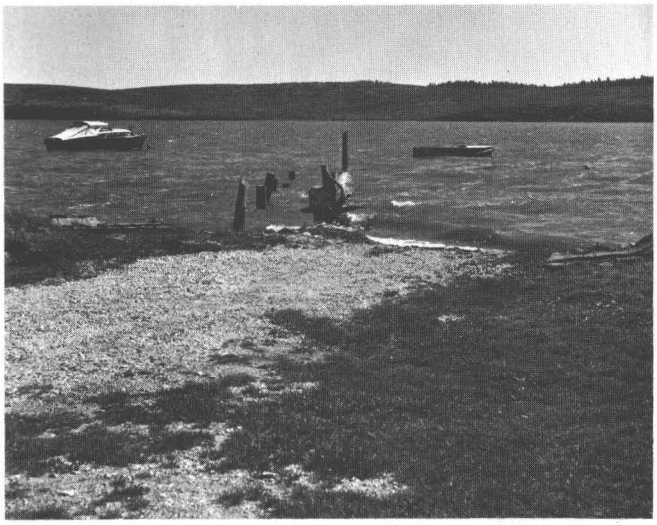

FIGURE 7.—Submerged dock at The Narrows Motel showing inundation of north shore of Hebgen Lake.

FIGURE 8.—Slump of ground toward lake with resultant tilting of foundation pillars, Lakeview Motel.

Oscillatory movement of the foundation of the main residence severely compacted fill next to it and left gaps of 4 to 5 inches. Inside the building, deep freezers and other large pieces of furniture were overturned.

At the Lakeview Motel, about 1$\frac{1}{2}$ miles to the west (E$\frac{1}{2}$ sec. 9, T. 12 S., R. 4 E.), sectors of ground near the new lakeshore slumped toward the lake with a consequent shifting of the supports for the motel units (fig. 8). The ground was intensely fractured, and gaping fissures as much as 6 inches wide and partly filled with water appeared near all the foundations. Ground in this fractured zone was so unstable that it trembled and quivered when walked on. A basement,

FIGURE 9.—Aerial view of Hilgard Lodge showing intense fracturing of ground and partial inundation of units. Main dwelling has floated away. Photograph by John R. Stacy.

a fireplace, and an enclosed porch had been added to the westernmost cabin during the spring and summer of 1959. The fracturing of the ground plus the northward displacement of the lake weakened the ground around the cabin so drastically that the cabin had to be lifted from its foundation and moved to higher ground lest it slide into the lake, which approached to within 2 or 3 feet of it.

A third lakeshore motel, Hilgard Lodge, in the N½ sec. 26, T. 11 S., R. 3 E., was completely destroyed by earthquake motion, landsliding and the surges of the lake (fig. 9). Part of the access road slid into the lake, and the ground was broken by small scarps and gaping fissures. Every building was tilted and knocked askew on its foundation, and many were broken in two. The main residence slid into the lake and floated away.

Kirkwood Ranch, along the north lakeshore in the center of sec. 25, T. 11 S., R. 3 E., is built almost completely of masonry. Damage was extensive. Chimneys fell, part of the brick veneer on the east side of the residence toppled, cracks appeared elsewhere in the veneer, and plaster in the motel units cracked and fell. All waterlines to the resort from springs in Kirkwood Canyon were broken, as were the waterlines that serviced the motel units. Alluvial fill on which the motel was built was unevenly compacted, and, consequently, foundations were weakened. During repair work, 2 to 3 inches of space was discovered beneath the cracked and collapsed basement and garage floors, which had once rested on soil—a result of compaction of the underlying material.

DAMAGE TO NATURAL FEATURES

During the major tremor a great many limestone and sandstone ledges forming ridge crests were severely shaken. As a result, large rock avalanches occurred; and huge boulders, many of them angular, were dislodged from the ledges and bounced down the steep wood-covered slopes (fig. 10). Each time a boulder struck the ground a pit was formed, some of them large enough to accommodate a seated man. Boulders were deflected from side to side by trees, and in many places they cleared a path, broke off branches, and bruised the upslope sides of trees. Boulder pits and rocks strewn along the valley floors suggest that some of the steep slopes at the time of the earthquake resembled bowling alleys during tournament time.

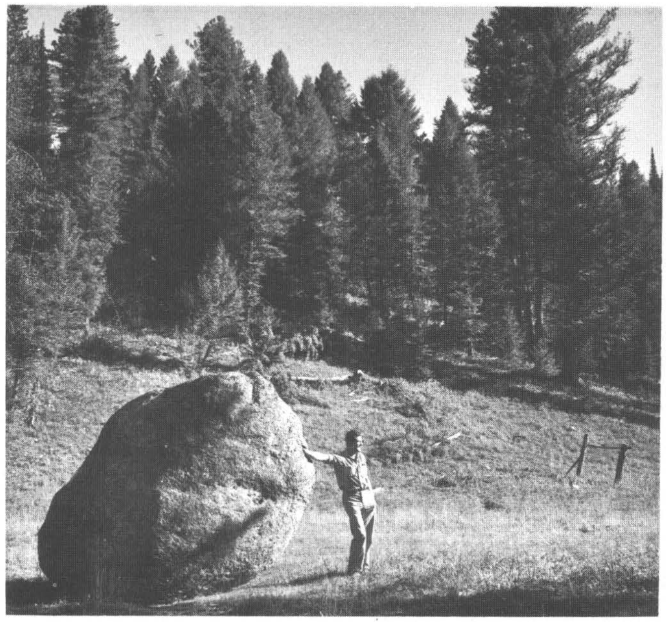

FIGURE 10.—One of the large boulders near Red Creek, displaced by the earthquake.

Seismological Investigations of the Hebgen Lake Earthquake

By LEONARD M. MURPHY *and* RUTLAGE J. BRAZEE

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

SEISMOLOGICAL INVESTIGATIONS OF THE HEBGEN LAKE EARTHQUAKE

By LEONARD M. MURPHY and RUTLAGE J. BRAZEE[1]

ABSTRACT

The earthquake of August 17, 1959, had a magnitude of 7.1 according to Pasadena and was felt over an area of 600,000 square miles. The epicenter is well located. Hypocenters seem to be at depths of 10 to 12 km. More than 1,300 aftershocks were recorded through October 15, 1959. About half the epicenters that have been located can be associated with the fault system that includes the Réd Canyon and Madison Range faults.

Macroseismic effects of the main shock were exceptionally severe in the epicentral region, where the number of known dead was 9, and 19 persons were reported missing. On the Modified Mercalli Intensity (Damage) Scale, the maximum intensity indicated by the faulting was X, and that by structural damage was VIII. Property damage exceeded $11 million.

The U.S. Coast and Geodetic Survey obtained strong-motion seismograph results a few days after the earthquake and good-quality aerial photographs 5 weeks later. First-order releveling of 1934 bench marks indicated a maximum settlement of 19 feet at a point about 4 miles southeast of Hebgen Dam.

INTRODUCTION

On August 17, 1959, the first major earthquake in this region since November 23, 1947, occurred at 23:37:15.0 MST, at 44°50′ N., 111°05′ W., very near the eastern terminus of the Red Canyon fault just northeast of Hebgen Lake. The shock had a magnitude of 7.1 according to Pasadena and was felt over an area of 600,000 square miles. It was felt from Utah and Nevada to British Columbia and from the Pacific coast to western North Dakota. The epicenter is well located with good agreement among the data used in its computation. An analysis of the seismograms from College, Alaska; Kipapa, Oahu, Hawaii; and San Juan, Puerto Rico, shows that although there is considerable variation in the character of the recordings, the hypocenters seem to be at a depth of 10 to 12 km.

There were no foreshocks, but the number of instrument-recorded aftershocks was probably more than 1,300 through October 15. An actual tabulation for the first 2 weeks gave a total of 880, and the rate in mid-October was still about two per day. Most of these were not perceptible and were only registered by the instruments of intermediate sensitivity at Bozeman and Butte. Immediately following the primary earthquake, the activity was so constant, and the large aftershocks so frequent, that no tabulation of the actual number felt or of the damage could be made. However, 35 to 40 separate shocks were reported from mid-September through October 15. Some of these were felt only slightly, although others did minor damage. From these data it can be assumed that the number of distinct tremors perceptible to man exceeded 200.

The macroseismic effects of the main shock were exceptionally severe in the epicentral region, where the number of known dead was 9, and 19 persons were reported missing. Most of the casualties were caused by landslides.

GEOGRAPHICAL DISTRIBUTION

There is no evidence of a geographical sequence of the epicenters that have been located. The open circles on the epicenter map (fig. 11) indicate aftershocks for which sufficient data were available to allow the computation of an epicenter. The crosses represent epicenters that had to be determined graphically because of a paucity of readings.

The epicenters listed in table 1 to the nearest minute of latitude and longtitude are the ones for which the data are in good agreement (table 2). Those listed to the nearest tenth of a degree are well located but have some scatter in the data. Where the agreement is poor for computed epicenters and where the location was graphically determined, the epicenters are shown to the nearest one-half degree. They are listed in this way to indicate the accuracy of location; however, in order to obtain a more meaningful pattern, each earthquake is plotted at the best estimated position.

[1] U.S. Coast and Geodetic Survey.

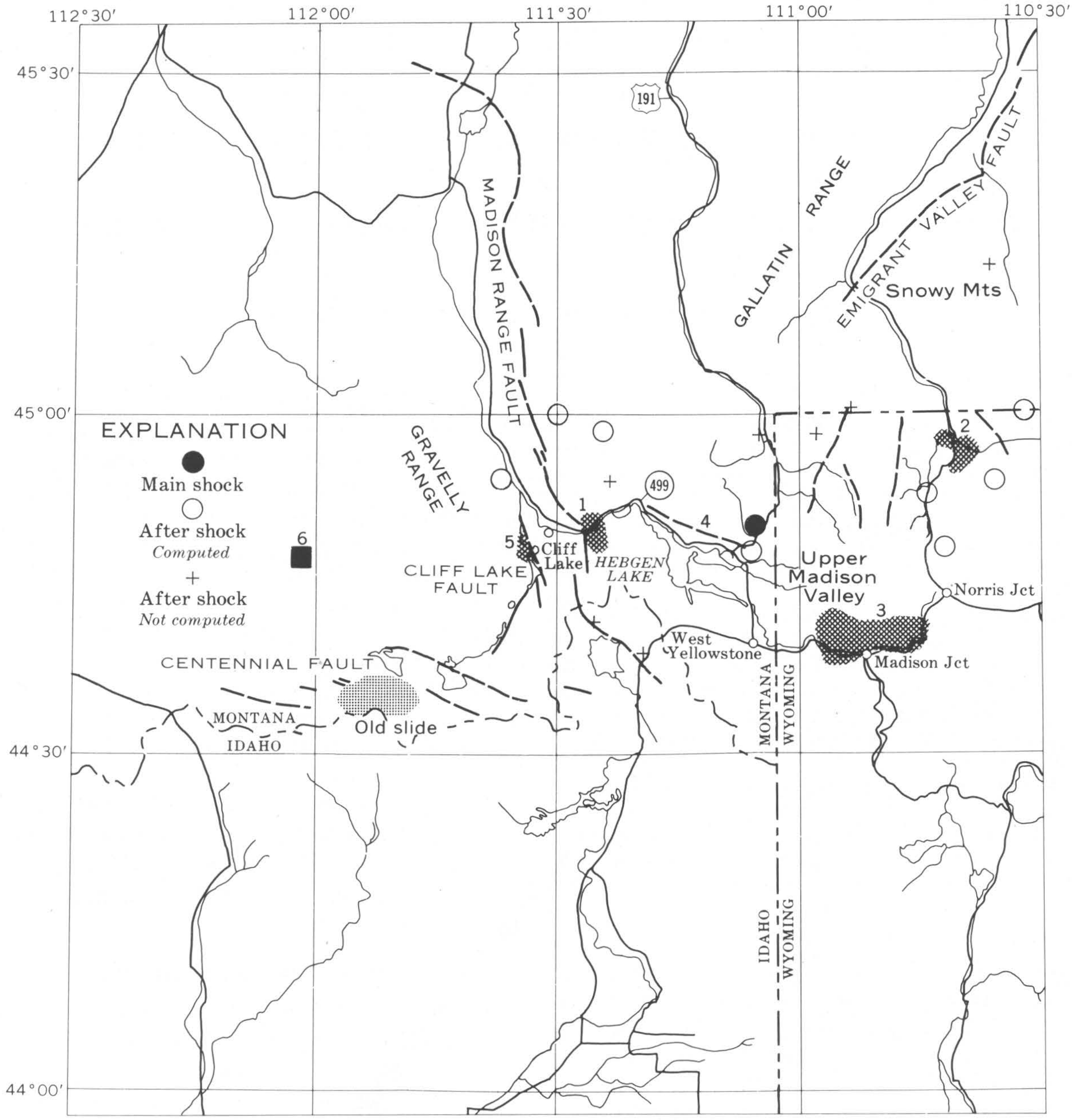

Figure 11.—Map of parts of Montana, Idaho, and Wyoming showing epicenters of the Hebgen Lake earthquakes. 1, Madison Slide; 2, Mammoth—slides and fissures, damage to pavement and walls; 3, Madison Junction—heavy slide activity, roads blocked and extensive fissuring; 4, Red Canyon fault; 5, Cliff Lake slide area; 6, epicenter of November 23, 1947, shock.

About half the epicenters can be associated with the fault system which includes the Red Canyon and Madison Range faults and trends northwest from the main shock area. The main shock and four of the seven aftershocks in this group originated along the Red Canyon fault and the northeastern limits of this system. The other shocks in this group were centered along a line to the southwest along the Madison Range fault. This line closely follows the scarp at the western base of the Madison Range. The scarp is about 40 miles long and showed a vertical displacement of 10 to 30 feet prior to the quake. It is of Recent origin and marks the line of a fault which has had a total relative displacement of several thousand feet since late Tertiary time.

The epicenters of the remaining nine aftershocks trend northeast from that of the main shock in two lines parallel to, but possibly not associated with, the

Emigrant Valley fault. One of these lines could be extended to include the focus of the main shock, whereas the other line, comprising four epicenters, lies to the southeast in the vicinity of Norris Junction. Fault evidence in this area is meager and is generally obscured by overlying volcanics.

TABLE 1.—*Teleseismic data and results*

Date (1959)	Time (GCT)[1]	Location		Magnitude
		Lat N.	Long W.	
Aug. 18	06 37 15.0	44°50′	111°05′	7.1 (Pasadena).
	07 54 32	45°	111°	
	07 56 18	45°	110½°	6½ (Berkeley).
	08 41 50	44.8°	110.7°	6 (Berkeley).
	11 03 52	44.8°	111.1°	5½–5¾ (Berkeley).
	15 26 06.5	44°53′	110°44′	6½ (Pasadena).
Aug. 19	04 04 03.0	44°54′	111°38′	6 (Berkeley).
	19 06 29	45.0°	111.4°	
	19 43 47.5	45°	110½°	
	21 45 57	45°	111½°	
Aug. 20	10 59 11	45°	110½°	
	19 11 27	45°	111°	
Sept. 8	07 09 48	45°	111°	
Sept. 14	09 34 52	44½°	111½°	
Sept. 28	08 05 42	45°	111°	
Oct. 5	11 33 14	44½°	111°	
Oct. 6	11 37 21	44½°	111°	

[1] Greenwich civil time.

TABLE 2.—*Instrument-recorded data*

Station	Phase	Station	Phase
Aug. 18, 1959, 06:37:15.0 GCT, 44° 50′ N., 111° 05′ W.			
Barrett, Calif.	eP 06 40 22	King Ranch, Calif.	iP 06 40 05
Berkeley, Calif.	eP 06 39 54	Logan, Utah	iP 06 38 05.7
Boulder City, Nev.	iP 06 39 33	Lubbock, Tex.	eP 06 40 17
Bozeman, Mont.	iP 06 37 50.3	Mount Hamilton, Calif.	eP 06 39 57
Butte, Mont.	iP 06 37 43	Pasadena, Calif.	iP 06 40 10
		Rapid City, S. Dak.	iP 06 38 42
China Lake, Calif.	iP 06 39 47	Riverside, Calif.	iP 06 40 09
College, Alaska	iP 06 43 13	Saint Louis, Mo.	iP 06 41 08
Dallas, Tex.	P 06 41 04	Salt Lake City, Utah	iP 06 38 20
Eureka, Nev.	eP 06 38 50	Seattle, Wash.	eP 06 39 16.5
Haiwee, Calif.	iP 06 39 49	Shasta, Calif.	iP 06 39 33.8
		Tinemaha, Calif.	iP 06 39 36
Hayfield, Wash.	eP 06 40 07	Tucson, Ariz.	iP 06 40 18
Hungry Horse, Mont.	iP 06 38 18	Woody, Calif.	iP 06 39 54
Isabella, Calif.	iP 06 39 54		
Aug. 18, 1959, 15:26:06.5 GCT, 44° 53′ N., 110° 44′ W.			
Barrett, Calif.	eP 15 29 16	Mount Hamilton, Calif.	iP 15 28 51
Berkeley, Calif.	eP 15 28 52	Palomar, Calif.	iP 15 29 08
Boulder City, Nev.	iP 15 28 26.6	Rapid City, S. Dak.	iP 15 27 30
China Lake, Calif.	iP 15 28 40	Thule, Greenland	eP 15 33 14
Fayetteville, Ark.	iP 15 29 41	Tinemaha, Calif.	iP 15 28 31
Hungry Horse, Mont.	iP 15 27 12	Tucson, Ariz.	iP 15 29 06
King Ranch, Calif.	eP 15 28 59	Woody, Calif.	iP 15 28 49
Lubbock, Tex.	iP 15 29 10		
Aug. 19, 1959, 04:04:03.0 GCT, 44° 54′ N., 111° 38′ W.			
Berkeley, Calif.	eP 04 06 43	Mount Hamilton, Calif.	eP 04 06 40
Boulder City, Nev.	iP 04 06 18.4	Pasadena, Calif.	iP 04 06 56
China Lake, Calif.	iP 04 06 31	Rapid City, S. Dak.	eP 04 05 35
Dallas, Tex.	eP 04 07 56	Tinemaha, Calif.	iP 04 06 21
King Ranch, Calif.	eP 04 06 48	Tucson, Ariz.	iP 04 07 05
Logan, Utah	iP 04 04 53	Woody, Calif.	iP 04 06 38
Lubbock, Tex.	eP 04 07 22		

SEISMIC FIELD SURVEYS AND QUESTIONNAIRE CANVASS

Upon receiving preliminary radio reports of the damaging Hebgen Lake earthquake, geophysicists from the Seismological Field Survey, U.S. Coast and Geodetic Survey, San Francisco, Calif., were immediately sent to the epicentral area. At West Yellowstone they joined Stephen W. Nile, professor of physics at Montana School of Mines and state collaborator in seismology. He had been near the epicenter when the major shock occurred and had already made a preliminary investigation. His early findings are in the Preliminary Report, Hebgen Lake, Montana Earthquakes, August 1959, issued by the Coast and Geodetic Survey on September 10, 1959.

Information obtained by field investigation and a questionnaire canvas is summarized below.

The main earthquake, which was felt over an area of approximately 600,000 square miles, caused 9 known deaths and 19 presumed deaths. The limit of the area where the quake was felt extended from Merritt in southern British Columbia, Canada, northeast to Banff in Alberta; southeast to Regina in Saskatchewan; and south to Estevan, Saskatchewan, near the United States–Canada border; and south into the conterminous United States. There, the limit of the area where the quake was felt extended from a number of communities in western North and South Dakota southwest through central Wyoming to Salt Lake City and Provo, Utah; thence westward to Gerlach, Nev.; northwestward to Portland, Oreg., and Seattle and Everett, Wash.; and back to Merritt, B.C. Although the shock was felt at Portland, Oreg. (122°45′ W.), it was not generally felt farther west than 119°30′ W.

As rated on the Modified Mercalli Intensity (Damage) Scale, the maximum intensity indicated by the faulting was X and that by structural damage was VIII. Property damage was in excess of $11 million, distributed as follows: $5 million suffered by forests and forest roads, $2,575,000 for Yellowstone Park roads, and $2,650,000 for U.S. highways.

STRUCTURAL DAMAGE TO HEBGEN DAM AND NEAR HEBGEN LAKE

Hebgen Dam (fig. 79) is an earthfill structure 700 feet long with a concrete core wall. The core wall is on solid rock from its southwest end to a point about 100 feet from its northeast end, where the foundation is colluvium. All preliminary inspections of the dam indicate that the concrete core was not seriously damaged. At a point where the core is not on bedrock, the fractures amounted to a few inches near the core top. The extent of these cannot be determined until the water level of the lake is lowered. The settlement of the earthfill section of the dam varied from approximately 4 feet on the east side of the core wall near the spillway to a negligible amount on the west end (fig. 81).

Around Hebgen Lake many summer homes and cabins were shifted from their foundations and otherwise considerably damaged. The greatest structural damage in the area probably occurred on the Blarneystone Ranch, just north of State Highway 499, 3½ miles east of Red Canyon, where practically all buildings were severely damaged (fig. 6).

A huge earthslide, comprising between 35 and 50 million cubic yards of debris, blocked the Madison River gorge completely, and all flow into the Madison Valley stopped for about 3 weeks (frontispiece; chapter K). The slide, about 7 miles below Hebgen Dam, formed a barrier which is nearly a mile long and has a maximum height of 400 feet on the north side and of about 200 feet on the opposite side, with a saddle of 175 feet.

DAMAGE IN OTHER AREAS

Damage to buildings and other structures in Montana and adjacent States, as rated according to the Modified Mercalli Intensity Scale, is summarized as follows:

INTENSITY VIII IN MONTANA

Cameron (25 miles south of).—Plaster cracked; chimneys fell; and damage was considerable to brick, masonry, and concrete.

Duck Creek Wye (8 miles north of West Yellowstone).—Chimneys fell, and a concrete garage was badly damaged. At Duck Bay a home was twisted out of shape, and the floors were separated from the walls.

Elliott's Resort (on Wade Lake).—Plaster, windows, walls, chimneys, and foundations cracked; plaster, walls, and chimneys fell. Damage was great to wood, masonry, and concrete. Bridge shook off foundation, shearing anchor bolts.

Halford's Resort (about 1½ miles southwest of dam).—Operators of this resort reported that all 29 resort cabins were damaged, many beyond repair.

Kirkwood Ranch (on north side of State Highway 499, at Kirkwood Creek).—Plaster, windows, walls, and chimneys fell. Damage was great.

Romsett Resort Area (south shore of Hebgen Lake).—Chimney crashed through roof of cabin, and all windows were broken.

West Yellowstone.—Many buildings were damaged to some extent. Many chimneys fell; plaster cracked; and service-station pumps were toppled at one place. Union Pacific Depot and Lodge were severely damaged; large stone pillars were knocked out of line; and a large stone chimney fell through the roof of the lodge building (fig. 4).

Additional damage (intensity VII) was reported at Cameron (Hutchins Bridge locality), Ennis, Harrison, Lionhead Ski area, and Monida (7L Ranch). Slight plaster and chimney cracks and miscellaneous minor damage were reported by 75 other communities. Practically the entire State felt the shock.

INTENSITY VII IN WYOMING

At Old Faithful Inn, Yellowstone National Park, one chimney fell, two small chimneys were damaged to such an extent that they had to be pulled down, and a water main in a building was broken. Other slight damage (intensity VI) occurred at Mammoth Hot Springs, Mount Holmes Lookout, and Northeast Entrance.

INTENSITY VII IN IDAHO

In Henrys Lake area there was considerable damage to chimneys, plaster, and so on. A concrete and rock dock 7 feet thick was broken by three large cracks, a barn was leveled, and, to the southeast, the top 12 feet of a fireplace fell. Similar damage to chimneys occurred in the Island Park and Macks Inn areas.

Many other communities throughout the State experienced some minor damage. Springs became muddy, and natural-gas fires broke out along the mountainside near Howe.

INTENSITY IN OTHER STATES

In eastern Washington, four places, including Farmington and Garfield, reported intensity VI, with slight cracking of plaster and chimneys. Fourteen other communities felt the shock.

In Oregon, chimneys cracked at Richland.

In Utah, several people felt a house shake as if it would slip from its foundation. The shock was moderately felt in other sections of the State, but south of lat 41°30′ only a few people felt the quake.

SPECIAL SURVEYS BY THE COAST AND GEODETIC SURVEY

Strong-motion seismograph results.—Seismograms from permanent stations at Bozeman, Butte, and Helena provided acceleration and period data for the main shock (table 3). Strong-motion seismographs were temporarily installed at West Yellowstone and at Hebgen Dam a few days after the main shock occurred. Seismograms from the permanent stations and these instruments yielded data on one large aftershock. Additional seismograms, too weak to be scaled,

TABLE 3.—*Summary of strong-motion seismograph results, Aug. 17, 1959, at 23:37 MST*

Station and component	Approximate period (sec)	Maximum acceleration (g)
Bozeman, Mont.		
Vertical	0.17	0.027
North	.25	.059
East	.25	.043
Butte, Mont.		
Vertical	.20	.021
North	.12	.032
East	.27	.044
Helena, Mont.[1]		
Vertical	{ .3	.01
	6	.006
North	{ .3	.02
	6	.002
East	.3	.02
Hungry Horse, Mont.		
Vertical	{ .5	.03
	6	.001
North	.2–.3	.0008
East	{ .3	.002
	5	.001

[1] During the first 12 hours after the main shock, six aftershocks were recorded by the Helena, Mont., accelerograph. The maximum acceleration ranged from 0.012 g to about 0.001 g.

An aftershock on August 27 at 16:37 MST, recorded at Hebgen Dam, had a maximum acceleration of 0.07 g in a regular train of 0.3-second waves.

recorded a number of the larger aftershocks. The scalings in table 3 are considered preliminary.

Level survey results.—First-order releveling from West Yellowstone to Sappington, Mont., was undertaken by a geodetic field party in September and October 1959. By comparing the 1934 levels with those of 1959, relative stability was indicated at West Yellowstone and at a location about 20 miles west of Hebgen Dam. The maximum change was a settlement of 19 feet about 4 miles southeast of Hebgen Dam. The settling increased gradually from the locations of stability to those of maximum change. Two anomalous conditions were noted, however, between West Yellowstone and the place of maximum movement. About 10 miles from West Yellowstone there was one bench mark which had remained at the same level. This permanency of level was indicated to a lesser degree at several other points in the same area. Two other bench marks about 12 miles from West Yellowstone changed in altitude by roughly 7 feet relative to others nearby.

REFERENCE CITED

U.S. Coast and Geodetic Survey, 1959, Preliminary report Hebgen Lake, Montana, earthquakes, August 1959: U.S. Dept. Commerce, 15 p.

Some Aftershocks of the Hebgen Lake Earthquake

By S. W. STEWART, R. B. HOFMANN, *and* W. H. DIMENT

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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SOME AFTERSHOCKS OF THE HEBGEN LAKE EARTHQUAKE

By S. W. STEWART, R. B. HOFMANN, and W. H. DIMENT

ABSTRACT

Two portable three-component seismograph systems were operated intermittently at three locations from 3.5 to 6.1 days after the main shock. More than 600 aftershocks ranging in magnitude from 0.4 to 3.7 were recorded at one station. Epicenters of 30 aftershocks are given. Only six of the epicenters are within the region of 1 foot or more of surface depression. Twenty of the epicenters form a marked alinement that strikes about N. 18°E. along the east margin of the seismically active area. This direction of alinement has no known relation to surface geologic features and is not evident in the pattern of surface faulting associated with the main shock.

INTRODUCTION

Two portable three-component seismograph systems were operated intermittently at three locations from 3.5 to 6.1 days after the main shock. Moving-coil seismometers with a damped natural frequency of about 3 cycles per second and calibrated in the frequency range 2 to 20 cycles per second were used. WWV timing signals or chronometer-relay closure marks were recorded on the seismograms. The location and approximate times of operation of the seismograph stations are summarized in table 4.

The Keg Spring and Victor stations were respectively about 55 kilometers southwest and 135 kilometers south of the reported epicenter of the main shock (Seismological Society of America, 1959). Seismic waves recorded at the three stations were generally in the frequency range of 2.5 to 10 cycles per second.

At the Keg Spring station, more than 600 aftershocks were recorded during 22 hours of operation. Magnitudes of these aftershocks, computed by the method of Richter (1958, p. 338–345), ranged from 0.4 to 3.7, with a mean of 2.0 (Stewart and others, 1960).

TABLE 4.—*Location and times of operation of seismograph stations*

Station	Latitude	Longitude	Altitude (ft)	Times of operation after main shock (GCT)[1]				Remarks
				From		To		
				Date	Time	Date	Time	
Keg Spring, Idaho	44°31.21′N	111°37.14′W	7620	8/22	0130	8/22	0530	In Centennial Mountains, on sedimentary rock. Approximately 55 km southwest of reported epicenter of main shock.
				8/22	1400	8/22	1600	
				8/22	1700	8/23	0200	
				8/23	0400	8/23	1000	
				8/24	0300	8/24	0900	
Victor 1 (Victor, Idaho)	43°36.64′N	111°5.36′W	6250	8/21	1800	8/22	0100	At east side of Teton Valley, Idaho, on outcrop of Paleozoic sedimentary rock. Approximately 135 km south of reported epicenter of main shock.
				8/22	0200	8/23	0030	
				8/23	0200	8/23	1000	
Victor 2 (Victor, Idaho)	43°38.41′N	111°10.25′W	6000	8/24	0042	8/24	0800	West of center of Teton Valley, near Teton River, on Quaternary alluvium. Approximately 135 km south of reported epicenter of main shock.

[1] These times of operation include the time used for changing recording paper, making minor repairs, and calibration of the amplifier-recorder units. The actual times of recording, therefore, are somewhat less.

ACKNOWLEDGMENTS

The assistance of W. H. Jackson, C. H. Miller, F. M. Valentine, and R. E. Warrick is gratefully acknowledged. W. B. Myers and G. D. Fraser provided geologic information and furnished the data on the location of the fault zones and regions of at least 1 foot of surface deformation shown in figure 12.

REDUCTION OF DATA

During times of simultaneous recording at the Keg Spring station and one or the other of the two seismograph stations near Victor, 30 aftershocks were sufficiently well recorded to permit an approximate location of their epicenters. Because it is not possible to determine uniquely epicentral locations from the records of only two seismic stations, the following assumptions were made: The foci of the aftershocks were at a depth of 10 kilometers and to the northeast of the line connecting the Keg Spring and Victor 1 stations; the velocity of compressional waves in the crust is 6 kilometers per second; and Poisson's ratio for crustal material is ¼.

Epicentral locations were determined from the time interval between the onset of the P and S waves observed at each station (Byerly, 1942, p. 221–222) and from the differences in arrival times of the direct P wave at both stations (Richter, 1958, p. 230). In the former method two intersecting circles are constructed on a map, and the epicenter is approximately at one of the points of intersection; in the latter method a hyperbola of constant time difference is constructed. The combined constructions give "triangles of closure" which represent, for the assumptions stated above, an estimate of the error in locating the epicenters. For the epicentral locations presented in this report (fig. 12), the longest sides of the triangles range in length from zero to 13 kilometers and have an average length of about 5 kilometers. The centers of the circles in figure 12 are approximately at the centers of gravity of the triangles of closure.

Seismograms representative of those used for locating aftershocks are illustrated in figure 13. The data from which the epicentral locations were determined are summarized in table 5.

TABLE 5.—*Arrival time and magnitude data for epicentral locations in figure 12*

Epicenter	P arrival time (GCT)		S-P interval (sec)		Date (GCT)	Magnitude
	Victor	Keg Spring	Victor	Keg Spring		
	h m s	h m s				
1	03:36:57. 77	03:36:45. 82	15. 66	6. 85	8-22-59	2. 9
2	03:40:27. 40	03:40:15. 70	16. 00	7. 38	8-22-59	3. 2
3	03:42:14. 30	03:42:02. 30	16. 33	7. 20	8-22-59	2. 6
4	04:41:50. 70	04:41:36. 05	16. 00	4. 65	8-22-59	3. 4
5	05:03:37. 23	05:03:25. 67	16. 29	7. 33	8-22-59	2. 7
6	14:09:35. 59	14:09:23. 60	13. 41	6. 03	8-22-59	3. 0
7	16:07:00. 77	16:06:53. 35	10. 81	5. 40	8-22-59	2. 7
8	19:08:21. 60	19:08:08. 70	15. 60	6. 57	8-22-59	3. 5
9	19:15:32. 75	19:15:15. 82	17. 25	4. 40	8-22-59	3. 2
10	23:19:15. 26	23:19:07. 17	11. 60	5. 27	8-22-59	3. 1
11	23:20:46. 22	23:20:38. 17	10. 87	5. 25	8-22-59	3. 3
12	00:38:45. 70	00:38:29. 69	16. 47	4. 99	8-23-59	2. 7
13	04:14:26. 37	04:14:09. 75	17. 33	5. 39	8-23-59	2. 4
14	04:35:02. 72	04:34:51. 89	13. 85	5. 55	8-23-59	3. 4
15	05:02:39. 87	05:02:28. 72	15. 53	7. 10	8-23-59	3. 2
16	05:11:32. 44	05:11:24. 00	10. 99	4. 70	8-23-59	2. 1
17	05:12:06. 29	05:11:54. 20	15. 84	6. 83	8-23-59	2. 4
18	05:15:02. 21	05:14:50. 43	16. 56	7. 40	8-23-59	2. 7
19	06:32:03. 40	06:31:54. 60	11. 58	5. 10	8-23-59	3. 1
20	07:00:12. 10	06:59:55. 40	19. 72	7. 31	8-23-59	2. 7
21	07:23:11. 00	07:22:55. 11	15. 73	5. 46	8-23-59	2. 5
22	08:02:44. 17	08:02:32. 29	17. 61	9. 65	8-23-59	2. 4
23	08:03:42. 50	08:03:32. 46	13. 06	4. 92	8-23-59	2. 7
24	08:05:44. 11	08:05:26. 89	15. 83	4. 14	8-23-59	2. 2
25	08:15:46. 96	08:15:37. 12	13. 54	5. 91	8-23-59	2. 8
26	08:29:23. 20	08:29:09. 28	15. 50	4. 37	8-23-59	3. 2
27	09:05:22. 19	09:05:13. 65	12. 58	6. 22	8-23-59	1. 8
28	09:07:04. 48	09:06:47. 88	13. 82	3. 04	8-23-59	2. 6
29	09:39:51. 43	09:39:33. 71	15. 54	3. 17	8-23-59	1. 9
30	03:36:25. 00	03:36:14. 24	14. 70	5. 79	8-24-59	3. 2

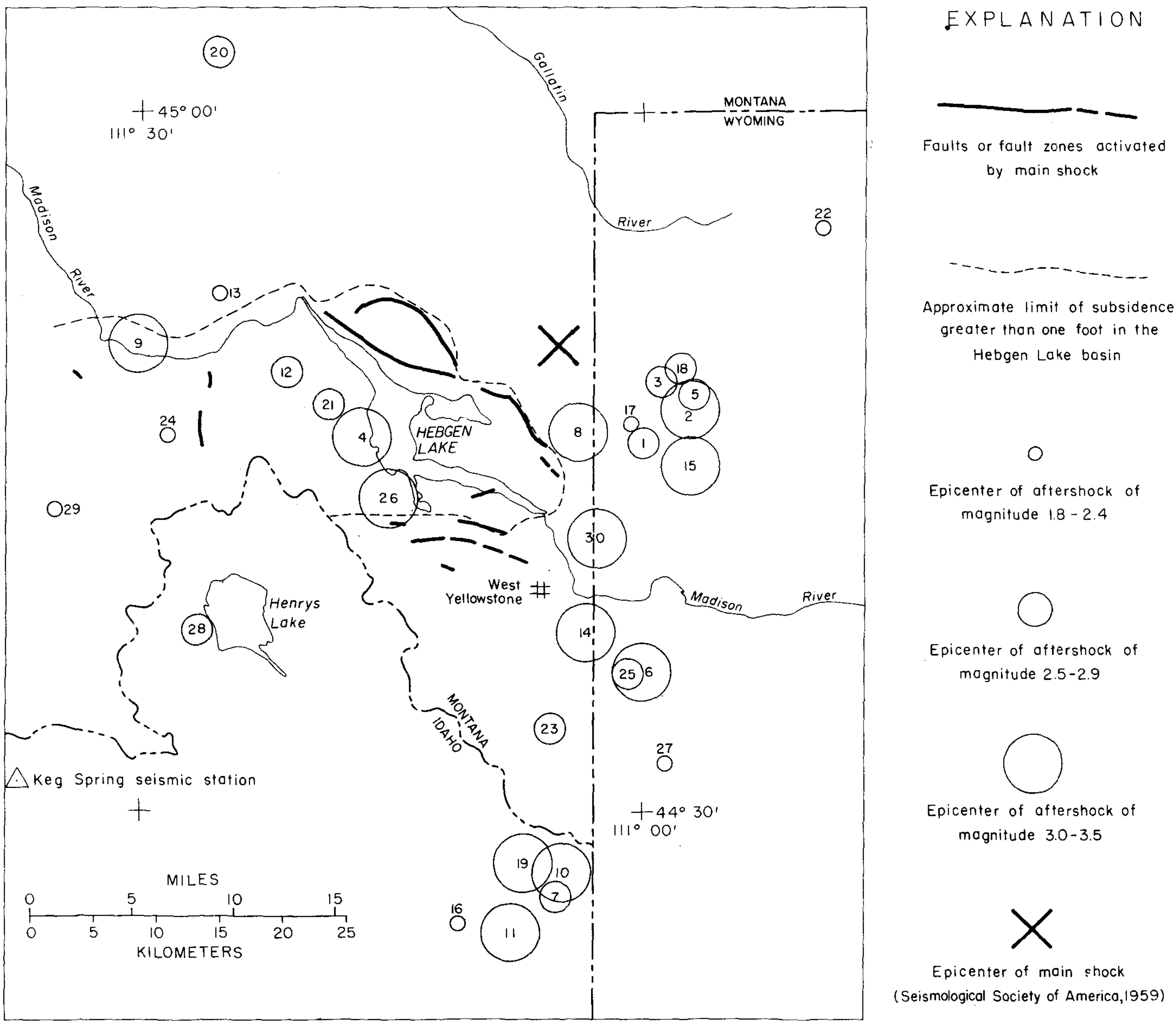

FIGURE 12.—Approximate location and magnitude of 30 aftershocks.

In table 6 data are tabulated for the larger aftershocks that were recorded at only one station.

In general, the times of arrival of P and S were picked to about ±0.05 seconds and ±0.1 seconds, respectively, with little difficulty. For events 10V and 153V in table 6, however, two possibilities for the S–P interval are given, which means that there are also two possibilities for the arrival of time of S. The earlier arrival time for S is characterized by a slight increase in amplitude and a definite increase in period over the part immediately preceding it. After about 1 second of the longer period phase, the high amplitude and relatively higher frequency phase that usually characterizes S arrives. In events 10V and 153V it is not possible to say which phase marks the real beginning of S.

Because seismograph gains were varied by factors as much as × 20 during the approximately 22 hours of seismograph operation at the Keg Spring station, it is difficult to estimate the lowest magnitude aftershocks that could have occurred and been recorded sufficiently well to determine its location. For the area shown in figure 12, however, all shocks of magnitude 2.2 or greater were recorded at the Keg Spring station probably during 80 percent of its total operation time, and at the Victor 1 station, during nearly

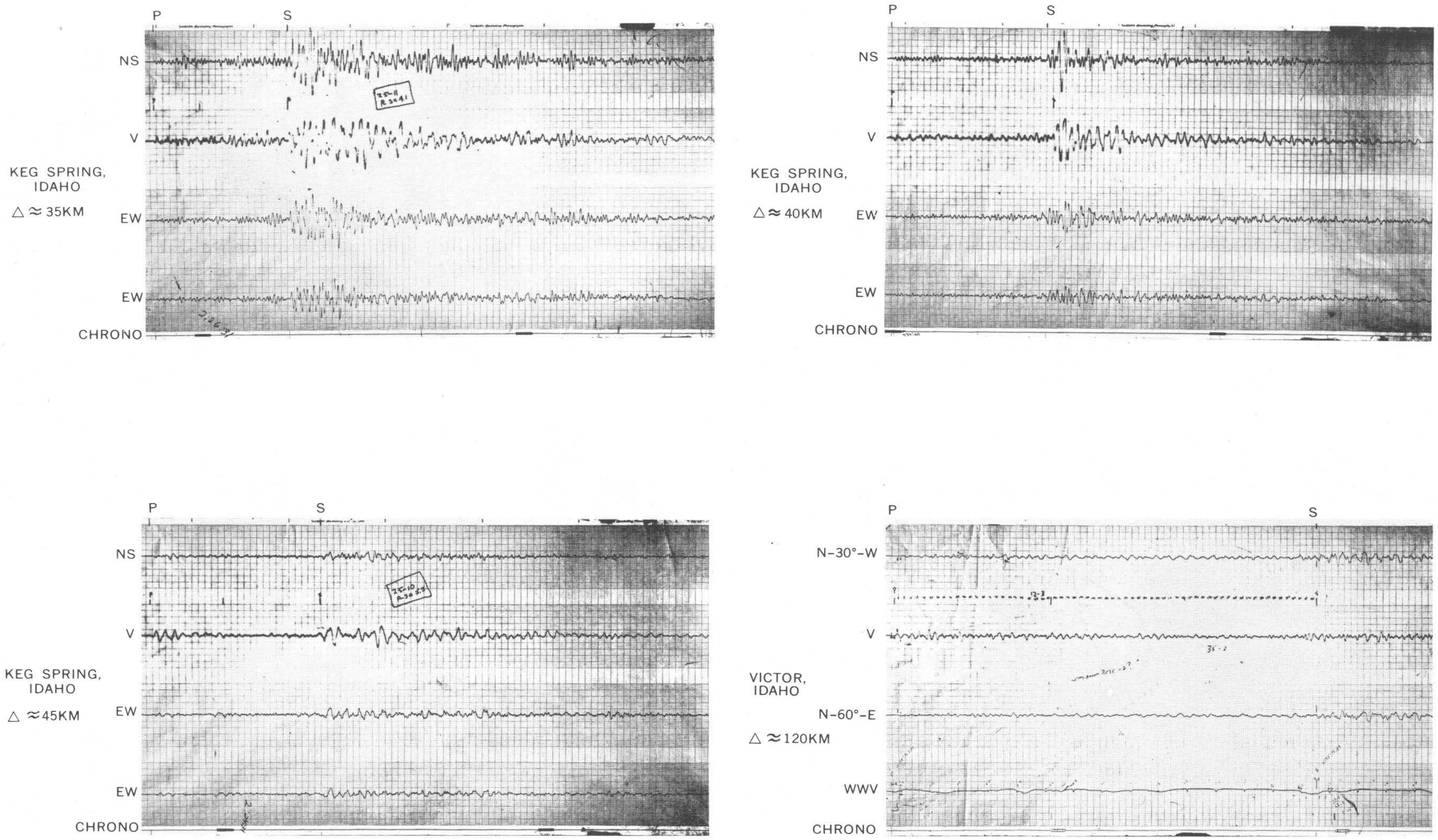

Figure 13.—Representative seismograms of aftershocks of the Hebgen Lake, Montana, earthquake. Chronometer marks at the bottom of each seismogram are 10 seconds apart.

TABLE 6.—*Arrival time and magnitude data for large aftershocks other than those in figure 12*

Event	P arrival time (GCT)[1]	S–P interval (sec)	Date (GCT)	Magnitude
	h m s			
6V[2]	19:11:03.87	12.38	8–21–59	3.5
8V	19:29:32.88	13.84	8–21–59	3.8
10V	19:38:45.90	12.12 (or 11.25?)[3]	8–21–59	3.5
14V	19:54:05.70	16.44	8–21–59	4.0
15V	19:58:05.65	13.75	8–21–59	3.5
71V	06:57:28.47	16.56	8–22–59	3.5
78V	07:34:09.85±0.2	12.19	8–22–59	3.5
96V	08:57:51.72	13.39	8–22–59	3.8
110V	10:07:58.82	11.53	8–22–59	3.5
153V	13:42:59.75±0.1	14.77 (or 15.83?)[3]	8–22–59	3.5
75KS[2]	17:06:05.74±0.1	7.23	8–22–59	3.5
111KS	19:08:20.3±0.2	7.11	8–22–59	3.7
170V	19:26:17.03	17.78	8–22–59	4.0
257V	00:46:56.74	14.46	8–24–59	3.8

[1] P arrival time is accurate to ±0.05 seconds or less unless otherwise indicated.
[2] V or KS following event number means event was recorded at Victor or Keg Spring station, respectively.
[3] Either of two phases could be S.

100 percent of its operation time. Aftershocks of magnitude less than 2.2 that are listed in table 5 were usually recorded during intervals of either low microseismic noise level or higher instrument magnification, or both.

With the exception of epicenter number 30 (table 5) and event number 257V (table 6), the direct P phases for all other aftershocks recorded at the Victor 2 station were too emergent in character, and the ground noise level was too high, to permit determinations of P arrival times and S–P time intervals to an accuracy consistent with that obtained for the other tabulated times. The adverse recording conditions presumably were caused by the Quaternary alluvium on which the seismometers were placed.

DISCUSSION AND CONCLUSIONS

Of the 30 epicentral locations in figure 12, it seems significant that only six lie within the region of 1 foot or more of surface depression. Of the remaining 24 locations, 20 form a marked N. 18° E.-strking alinement along the east margin of the seismically active area. This alinement, which has no known relation to surface geological features, and which is not evidenced in the pattern of surface faulting, is interpreted to be the result of a major regional-stress distribution at depth.

Byerly's fault-plane solution (1955, p. 80–83) for the Montana earthquake of June 28, 1925, indicates that the fault movement was predominantly right-lateral strike slip, and that the fault had a strike of N. 26° E. and a dip of 87° SE. The epicenter given by Byerly is about 180 kilometers (110 miles) north of the Hebgen Lake region. Two possible locations have been suggested for the fault that may have caused the 1925 earthquake. Byerly (1955, p. 81) suggests that the earthquake may have occurred along the Lombard overthrust. Pardee (1926, pl. 4) suggests that it may have occurred along a normal fault of post-Miocene age about 7 miles east of the Lombard overthrust. Both faults strike about N. 10°–20° E. and dip to the northwest. Thus the field evidence in the epicentral region supports Byerly's interpretation of the direction of strike of the fault but otherwise differs from it in that at the surface the faults dip to the northwest and do not show predominantly strike-slip movement. It may be significant that no major surface breakage was observed in the epicentral region of the earlier earthquake to the north (Pardee, 1926, p. 22), and none was observed along the northeast-trending zone of aftershock alinement associated with the Hebgen Lake earthquake.

The possible error in the location of the epicenter for the main Hebgen Lake shock is probably such that this shock could have been associated with either the zone delineated by the northeast-trending alinement of aftershocks or the northwest-trending zone of faulting and subsidence observed at the surface. A third possibility, which goes furthest in explaining the observed effects of the earthquake and its aftershocks, is that the main shock occurred along or near the intersection at depth of these two zones. The greater aftershock activity along the northeast-trending zone suggests that this direction was a primary one for the accumulation and subsequent release of strain energy. The northwest-trending zone of surface faulting and subsidence is then interpreted as representing an older zone of dislocation which was reactivated by the forces that produced the northeasterly trending zone.

If the very similar directions of strike of about N. 26° E. deduced for the fault plane of the Montana earthquake of 1925 (Byerly, 1955, p. 81) and about N. 18° E. for the fault plane of the Hebgen Lake earthquake are not merely coincidental, then this general direction may be significant in regard to tectonic processes now working in the region.

Unfortunately, a fault-plane solution for any of the Hebgen Lake shocks that utilizes only directions of motion from the compressional phases would give a choice of two possible, mutually perpendicular, fault planes. These fault planes would probably correspond closely in strike to the predominant strike of the surface fracturing and deformation and the strike alinement of aftershocks discussed above. Perhaps the ambiguity could be resolved by considering the relation of the dips and relative motions of the seismically determined fault planes to the geologic data, but quite possibly the difficult problem of determining directions of first motion of the shear arrivals would also have to be considered. It is not possible to interpret further the seismograms obtained for this study in terms of depths of foci of the aftershocks, or to place limits upon the nature of the focal mechanism of the aftershocks, although such data would be valuable in helping to eliminate the ambiguity of a fault-plane solution.

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Regional Seismicity and Brief History of Montana Earthquakes

By CLYDE P. ROSS *and* WILLIS H. NELSON

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CONTENTS

REGIONAL SEISMICITY AND BRIEF HISTORY OF MONTANA EARTHQUAKES

By Clyde P. Ross and Willis H. Nelson

ABSTRACT

The earthquake at Hebgen Lake, Montana, occurred within a seismically active zone that extends across western Montana and Wyoming, southeastern Idaho, and Utah. The zone generally parallels the structural grain of the region, which is of complex origin and reflects intermittent deformation through much of geologic time.

Southwestern Montana has suffered 54 major recorded earthquakes. The earliest supported by evidence was in about 1770, but most records are for quakes in the present century. Earthquake hazards should be considered in planning future structures in the region.

REGIONAL SEISMICITY

The earthquake of August 17, 1959, at Hebgen Lake, Montana, occurred within a seismically active zone that forms an arc through western Montana, northwestern Wyoming, southeastern Idaho, and Utah. Richter (1959) and Woollard (1958) note the presence of this zone. Figure 14, which shows the distribution of historic earthquakes in this region, was compiled from U.S. Coast and Geodetic Survey publications (Bodle and Murphy, 1948; Heck, 1947; Murphy, 1950; Murphy and Ulrich, 1951, 1952; Murphy and Cloud, 1953, 1954, 1955, 1956). Some of the earthquakes were identified and located by means of seismograms, but most were reported by persons in the areas involved. The data collected south of lat 42° N. may show a somewhat biased distribution pattern of earthquakes because the population there is concentrated at the foot of the mountains along the line that is probably most active seismically. North of lat 42° N. there is no such linear concentration of population.

Earthquakes of both large and small intensity are concentrated within the seismic zone. In Montana, six earthquakes with intensities of M.M. VIII (Modified Mercalli scale) or greater have occurred within this zone since 1868, and three such earthquakes have occurred within the zone in Utah during the same period. (In chapter F, Fraser discusses the various scales by which the intensities of earthquakes are rated.)

The trend of the seismic zone generally parallels the structural grain of the region through which it passes. The structural grain is of diverse origin; it consists of fold axes and surface traces of overthrust faults that formed near the end of the Mesozoic era (during the Laramide orogeny), and scarps of high-angle faults of Cenozoic age. In south and central Utah the zone encloses the faults which lie on the boundary between the Colorado Plateau on the east and the Basin and Range province on the west. The Hurricane fault in southwestern Utah is typical of these faults. It has been active at least since Miocene time (Gardner, 1941) and probably since Eocene time (J. Hoover Mackin, oral communication, 1959), and it follows an older structure of late Mesozoic age, the Kanarra fold of Gregory and Williams (1947).

In northern Utah, southeastern Idaho, and western Wyoming, the zone is about parallel to the surface traces of thrust faults, with axes of folds of late Mesozoic age, and with normal faults, some of which have been active throughout Cenozoic time.

In the vicinity of Yellowstone Park, bedrock structures are hidden beneath volcanic rocks of Quaternary age.

Between Yellowstone National Park and Helena, Mont., where most of the strong historic earthquakes in Montana (including the Hebgen Lake earthquake) have occurred, the structural grain is heterogeneous. Locally, crystalline rocks of Precambrian age, with older but, for the most part, unknown, structural grain are exposed. In the same general area the trends of other fold axes, such as those in the Bridger Range, the Big Belt Mountains, the Elkhorn Mountains, and locally in the Madison Range, are about parallel to the seismic zone. Near Three Forks, Mont., and north of Hebgen Lake, however, fold axes and surface traces of thrust faults of late Mesozoic age trend at high angles to the seismic zone. (See the Geologic Map of the State of Montana, Ross, Andrews, and Witkind, 1955.) Here also many faults of Cenozoic age, such

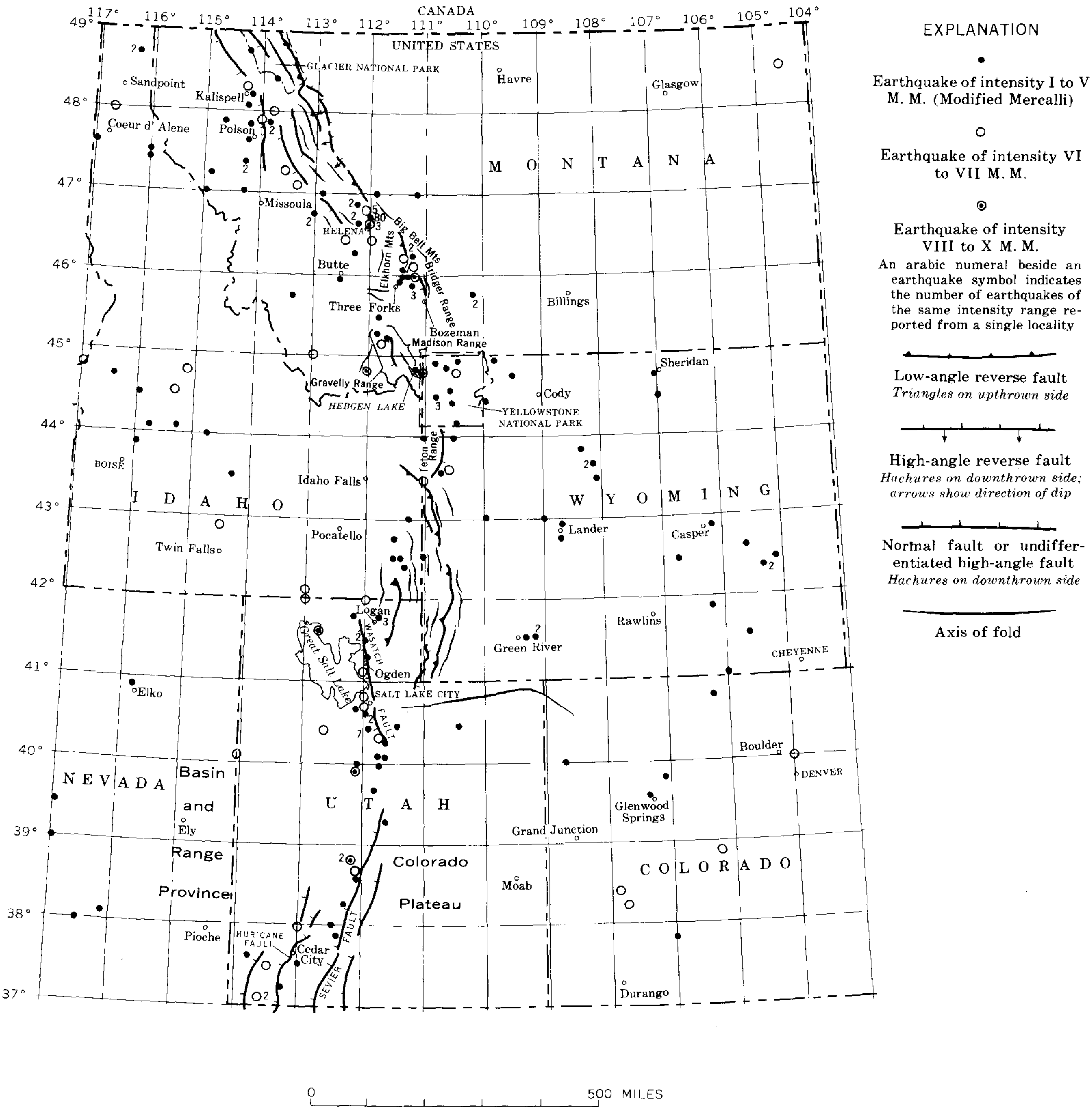

FIGURE 14.—Map showing earthquakes reported from the Rocky Mountain region and adjacent areas from 1869 through 1954, including the Hebgen Lake earthquake of August 17, 1959. Seismic data from publications of the U.S. Coast and Geodetic Survey (see text for references). Structural data modified from the tectonic map of the United States (King and others, 1944). Structures are indicated only in the zone of greatest seismic activity.

as those along the west side of the Madison Range, Bridger Range, and Big Belt Mountains, are more or less parallel to the trend of the seismic zone.

North of Helena, Mont., the seismic zone trends northwestward parallel to known folds of late Mesozoic age and parallel to faults of Cenozoic age (Pardee, 1950).

The trend of the faults of Cenozoic age is undoubtedly influenced in many places by the grain of the older structures of late Mesozoic age, but the fact that most of these faults of Cenozoic age trend parallel to the seismic zone suggests that the seismic zone probably has, in part at least, controlled their trend.

The general area covered by the seismic zone seems to have been structurally significant even before the end of Mesozoic time. Eardley (1951, pls. 2, 3, 5, 6, 15, and 16) shows the borders of subsiding troughs at or near the location of the present seismic zone throughout much of Paleozoic and Mesozoic time.

The preceding observations suggest that the present seismic zone may reflect the existence of a fundamental and probably persistent tectonic phenomenon. This fundamental phenomenon must in some way result from differential movement of material at depth on either side of a line or narrow zone. Earthquakes can result from the sudden release of accumulated strain, either along the fundamental structure, or along faults at shallower levels in the earth's crust that were strained by slow movements at depth. If the earthquakes are due to movement on faults above the fundamental structure, the trend of such faults may be guided by some inherited grain in the bedrock they cut rather than by the trend of the deep fundamental structure.

From the above discussion it is apparent that the surface faults that developed during the Hebgen Lake earthquake may be either continuous with, and directly related to, a plane of movement at depth, either primary or secondary, or they may be features which resulted from the earthquake and need not continue down to the plane along which the movement that caused the earthquake occurred. Data that would definitely indicate which of these situations occurred during the Hebgen Lake earthquake are not available.

HISTORY OF EARTHQUAKES IN SOUTHWESTERN MONTANA AND VICINITY

Pertinent historical data about southwestern Montana are summarized below. The records of the U.S. Coast and Geodetic Survey (Bodle and Murphy, 1948; Heck, 1947; Murphy, 1950; Murphy and Ulrich, 1951, 1952; Murphy and Cloud, 1953, 1954, 1955, 1956) for an area bounded by latitudes 43° and 47° N. and longitudes 108° and 114° W. are abstracted in table 7 and shown graphically in figure 15. Only disturbances with intensities of IV or greater are listed in the table and figure. In both, the intensities shown for disturbances before 1946 are those of the Rossi-Forel scale (labeled R.F.); those since 1946 are given according to the Modified Mercalli scale (labeled M.M.). The differences between the two scales are not great. Where several earthquakes occurred in a single year, only the largest is plotted in figure 15. Many earthquakes of minor intensity are on record.

Table 7 lists 54 earthquakes. Of these, 19 centered around Helena, and 8 of those at Helena occurred in 1935. Most of the records were made during the pres-

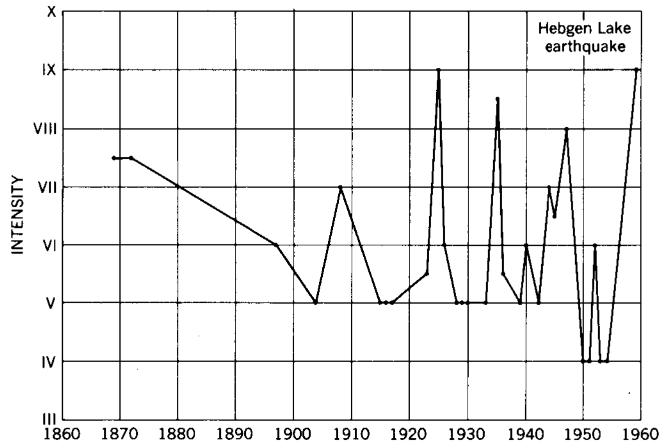

FIGURE 15.—Graph showing major earthquakes recorded in southwestern Montana.

ent century, and those of intensities greater than R.F. VII were mostly obtained since 1920. These facts correspond to the relative recency of settlement in the region rather than to any change in the character of seismic activity there. Clearly the period for which accurate data are available is too short for prognostications as to future earthquake activity. One may venture the suggestion that earthquakes strong enough to attract marked attention and to do some damage are to be expected to recur in and near southwestern Montana sufficiently often so that they should be taken into account in planning houses and other structures in the region. Pardee's remarks (1926, p. 22–23) to this same effect, made long ago, might well be borne in mind in connection with any future construction. He called attention to the Madison Range in particular.

The geology of the region under consideration is complex and varied, and one of its major features is the variety and abundance of faults of late Cenozoic age. Any of the young faults might become the site of a modern earthquake. So far as Montana is concerned, Pardee (1950) furnishes an adequate summary. In the present connection, his remark (p. 373) that the Recent fault scarp at the base of the Madison Range has existed in its present form since about 1770 is of particular interest. The estimate is based on the stump of an old tree found near the top of the scarp slope which represents the minimum age of the scarp rather than the date of the faulting. Alden (1953, p. 188-189) also comments on the presence of old trees on the fresh-looking face of the fault scarp. Pardee (1950, p. 369–370) states that the scarp is about 40 miles long and has a vertical displacement of between 10 and 30 feet or more. The total relative displacement on the fault since late Tertiary time is

TABLE 7.—*Major earthquakes in and near southwestern Montana*

Date	Time (hr., min., sec.)	Locality	North Latitude (degrees)	West Longitude (degrees)	Area (sq. mi.)	Intensity [3]
May 22, 1869	a.m.	Helena, Mont.	46.6	112.0		
Dec. 10, 1872	4:30 p.m.	Montana	46.4	112.5		R.F. VII–VIII
Nov. 4, 1897	2:29 a.m.	do	45.0	113.0		R.F. VI
Aug. 3, 1904	8:00 p.m.	do	45.5	111.8	(1)	R.F. V
Dec. 20, 1908	4:30 p.m.	do	45.2	111.8	(1)	R.F. VII
May 8, 1915	9:10 a.m.	Wyoming	44.9	110.7	10,000	R.F. V
Sept. 9, 1916	7:57 p.m.	Idaho	43.5	114.3	(1)	R.F. V
Apr. 19, 1917	9:30 p.m.	do	44.0	114.8	(1)	R.F. V
Mar. 23, 1923	9:00 p.m.	Wyoming	43.6	110.6	1,500	R.F. V–VI
June 27, 1925	6:21 p.m.	Montana	46.0	111.2	310,000	R.F. IX
Nov. 17, 1925	6:50 p.m.	Wyoming	44.6	107.0	3,000	R.F. V
May 31, 1926	5:25 a.m.	Montana	46.0	111.4	1,000	R.F. V
Dec. 12, 1926	5:44 p.m.	do	46.1	111.2	30,000	R.F. V–VI
Feb. 13, 1928	7:00 a.m.	Central Wyoming	43.5	108.2	3,000	R.F. V
Feb. 15, 1929	8:00 p.m.	Montana	46.1	111.3	40,000	R.F. V
Mar. 16, 1930	6:00 a.m.	Helena, Mont.	46.5	112.0	(1)	R.F. V
July 9, 1930	6:00 p.m.	West Montana	47.0	115.0	[2] Large	R.F. V
Aug. 24–Dec. 22, 1930	Various	Yellowstone National Park	44.0	111.0		R.F. IV–V
June 10, 1933	10:59 p.m.	Helena, Mont.	46.5	112.0	(1)	R.F. V
Aug. 19, 1933	3:13 a.m.	Logan, Mont.	45.8	111.3	(1)	R.F. V
Nov. 29, 1933	10:00 a.m.	Virginia City, Mont.	45.3	111.8	200	R.F. V
Nov. 23, 1934	4:40 p.m.	Lander, Wyo.	43.0	109.0	8,000	R.F. V
Oct. 3, 1935	7:47 p.m.	Helena, Mont.	46.5	112.0	(1)	R.F. V
Oct. 7, 1935	12:30 p.m.	Craig, Mont.	47.0	111.9	(1)	R.F. V
Oct. 12, 1935	12:51 a.m.	Helena, Mont.	46.6	112.0	70,000	R.F. VIII
Oct. 18, 1935	9:48 p.m.	do	46.6	112.0	230,000	R.F. VIII–IX
Oct. 27, 1935	12:00 p.m.	do	46.6	112.0	(1)	R.F. VI
Oct. 31, 1935	11:38 a.m.	do	46.6	112.0	140,000	R.F. VIII
Nov. 21, 1935	8:58 p.m.	do	46.6	112.0	13,000	R.F. VI
Nov. 28, 1935	7:42 a.m.	do	46.6	112.0	90,000	R.F. VI–VII
Jan. 14, 1936	9:40 p.m.	Yellowstone National Park	44.0	110.5	1,200	R.F. V
Feb. 13, 1936	4:55 p.m.	Helena, Mont.	46.6	112.0		R.F. V–VI
March, 1936	Various	Marysville, Mont.	46.8	112.3		R.F. V
June 11, 1936	4:13 p.m.	Helena, Mont.	46.6	112.0	(1)	R.F. V
May 11, 1939	5:00 p.m.	Trident, Mont.			(1)	R.F. V
Dec. 23, 1940	2:50 p.m.	Helena, Mont.			7,000	R.F. VI
Aug. 5, 1942	3:34 p.m.	Yellowstone Park, Wyo.			(1)	R.F. V
July, 12 1944	1:30 p.m.	Seafoam, Idaho	44.7	115.2	70,000	R.F. VII
June 1, 1945	10:56 a.m.	Helena, Mont.			6,000	R.F. VI
Sept. 23, 1945	2:58 a.m.	Western Montana			5,000	R.F. V
Nov. 23, 1947	2:46:05 a.m.	Madison County	44.8	112.03	150,000	M.M. VIII
Feb. 10, 1950	7:10 p.m.	Superior, Mont.				M.M. IV
June 27, 1950	9:31:04 p.m.	West Yellowstone	44.75	110.5		
Aug. 18, 1950	12:22 a.m.	Helena, Mont.				M.M. IV
Apr. 23, 1951	1:57 p.m.	do				M.M. IV
Nov. 14, 1951	5:14 a.m.	do				M.M. IV
Jan. 3, 1952	7:34 p.m.	do				M.M. IV
Apr. 15, 1952	7:20 a.m.	Whitepine, Mont.				M.M. IV
Apr. 22, 1952	9:54:42.5 a.m.	Townsend, Mont.	46.2	111.4	1,500	M.M. VI
May 29, 1952	8:15 p.m.	Greyson Creek and Townsend, Mont.				M.M. V
March 8, 1953	6:39:34 p.m.	Helena, Mont.				M.M. IV
Aug. 15, 1953	12:08:35 a.m.	Lombard, Mont.				M.M. IV
July 4, 1954	12:40 a.m.	Yellowstone National Park	44.9	110.8		M.M. IV
Sept. 10, 1954	12:50:21 p.m.	do				M.M. IV

[1] Local.
[2] Uncertain.
[3] Rossi-Forel scale, R.F.; Modified Mercalli scale, M.M.

several thousand feet. There is no written record of the earthquake at the time the scarp was formed, but presumably the earthquake was more powerful at the ground surface than any that have followed it with the possible exception of the 1959 Hebgen Lake earthquake.

Noises noticed by members of the Lewis and Clark expedition when camped near the Great Falls of the Missouri River in 1805 (Coues, 1893, p. 402) have been attributed to an earthquake (Pardee, 1926, p. 22–23; Scott, 1936, p. 2). The members of the expedition did not report any earth movement accompanying the noises, however. Whether the center of this disturbance was within the region discussed or whether it was actually an earthquake is not known.

About 100 years after the youngest probable date for the formation of the scarp along the base of the Madison Range, other earthquakes were recorded. One of these was at Helena on the morning of May 22, 1869 (Scott, 1936, p. 2), and was strong enough to shake a house and upset dishes and furniture. Another, much closer to the site of the recent earthquake (Hayden, 1872, p. 82), was perceptible on the northeast side of Yellowstone Lake. Hayden remarked that earthquake shocks were sufficiently frequent so that the Indians tended to avoid the region. A rather severe shock in 1872 is listed in table 7. This shock centered near Emery and shook buildings and furniture, but did little permanent damage.

After 1872 no shock worthy of remark was recorded until 1879. Since then, as shown in table 7 and figure 15, shocks have been recorded every few years, and some were spaced less than a year apart. One of the most severe occurred on June 27, 1925 (Pardee, 1926). This earthquake, whose epicenter was near Lombard, caused considerable damage within an area of 600 square miles and was felt over an area of 310,000 square miles. The maximum intensity was R.F. X (Pardee, 1926, pl. 3). Buildings were damaged, cracks appeared in the ground, and a landslide blocked the track of the Chicago, Milwaukee, St. Paul and Pacific Railroad. The slide caused a lake to form, and 2 weeks were required for the construction of a temporary track around the slide so that railroad traffic could be resumed. Only two people were injured as a result of this earthquake. Aftershocks were recorded at intervals until September 3, 1925.

In 1935 (Scott, 1936) a large number of shocks occurred near Helena. The most severe of these was on October 18 and had an intensity of R.F. IX. More than 1,200 shocks were recorded in a period of 80 days. Before the main earthquake, 62 tremors occurred. Damage in Helena resulting from the earthquake of October 18 was about $3.5 million (Scott, 1936, p. 10–15); but some of the other tremors in the series also caused damage, so the total may have reached $4 million. Many buildings were damaged, and cemetery monuments were moved. Minor ground cracks formed in various places within a distance of 15 miles from the center of disturbance. Four people were killed, and a few additional deaths have been attributed to nervous shock related to the earthquake. About 50 persons were injured.

On November 23, 1947 (Murphy, 1950, p. 8–14), an earthquake near Lima reached an intensity of M.M. VIII. This was felt over an area of 150,000 square miles but seems to have attracted less publicity and investigation than similarly strong earthquakes in the region. Buildings were damaged, rocks rolled down mountain slopes, and there were some effects on streams and springs. No injuries to people are recorded.

The next major earthquake was the Hebgen Lake earthquake, the largest recorded in Montana history. It is compared with previous Montana earthquakes in the following chapter.

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Intensity, Magnitude and Ground Breakage

By GEORGE D. FRASER

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

INTENSITY, MAGNITUDE, AND GROUND BREAKAGE

By George D. Fraser

ABSTRACT

The Hebgen Lake earthquake has been assigned a greater maximum intensity and a greater magnitude than any earthquake previously studied in Montana. Further, the Hebgen Lake earthquake was the only one accompanied by surface faulting. The threshold magnitude for ground breakage in this part of Montana seems to be 7 (Richter scale). The empirical method of Tocher (1958) for estimating energy release from ground breakage yields a value consistent with magnitude, although there are several reasons why the surface-fault method of earthquake evaluation might not work in Montana. The fact that the method does yield acceptable results suggests that displacement on the surface faults reactivated during the earthquake is quantitatively controlled by the earthquake-generating fault(s) at depth.

INTRODUCTION

Three of the many ways of comparing or rating earthquakes are used here—intensity, magnitude, and the degree of surface-fault breakage (Tocher, 1958, p. 147). This last method is relatively new and untried outside the region of northern California and Nevada; but it offers, for the first time, a quantitative geological method of evaluating large shallow earthquakes in some areas. Table 8, abstracted from Richter (1958) and Tocher (1958), gives a very approximate correlation of the three methods. Correlation between methods can never be precise, and the usefulness of such a table probably diminishes rapidly if areas other than California, Nevada, and southwestern Montana are considered. Especial caution must be used with the surface-fault method of comparing earthquakes.

INTENSITY

The Modified Mercalli scale of intensity (abbreviation M.M.) has been in general use since 1931. Before adoption of this scale, and for some purposes even today, the similar Rossi-Forel scale (abbreviation R.F.) was applied throughout the world. When isoseismals are drawn, it is on a basis of one or the other of these intensity scales. Either scale, therefore, rates damage in a specific area. Any scheme of this type is bound to be subjective—conditioned by the kind of ground in the area rated, the type and number of structures damaged or available to be damaged, the reliability of personal reports, the focal depth of the earthquake, and the changing emphasis on different criteria. Richter (1958) gives a good discussion of the problems. In spite of all its drawbacks, this method remains a useful one.

Without going into details of just how each earthquake was rated, it seems that the 1925 shock which centered about 20 miles northeast of Three Forks near lat 46° N., long 111.2° W., was the largest recorded earthquake in Montana before the one at Hebgen Lake (table 7). It has been assigned a maximum R.F. intensity of IX. Converted to the Modified Mercalli scale, this would be slightly less than IX. Two later shocks have also been assigned a maximum M.M.

Table 8.—*Approximate correlation of magnitude, intensity, and surface faulting*

[After Richter, 1958, p. 353 and 651; Tocher, 1958, p. 149]

Magnitude	2	3	4	5	6	7	8
Maximum intensity:							
Modified Mercalli	I–II	III	V	VI–VIII	VII–VIII	IX–X	XI
Rossi-Forel	I–II	III	V–VI	VI–VIII(–)	VIII(–)–IX	IX(+)–X	X
lD for surface faulting (Tocher)	Not applicable below threshold magnitude.				$<10^4$	$>10^4$, $<10^5$	$>10^5$

intensity of VIII—the Helena shock of October 18, 1935, and the Virginia City shock of November 23, 1947. For the Hebgen Lake shock, the maximum M.M. intensity in a small area north of Hebgen Lake was close to X; a larger area would be enclosed by the IX isoseismal. As a further indication of intensity, the limit of perceptibility for the 1925 shock was about 400 miles, whereas the limit for Hebgen Lake shock was at least 550 miles (felt in Seattle, Wash., by several persons). Perceptibility limits are quite unreliable as measures of maximum intensity except in comparison of earthquakes originating in the same area at the same focal depth (Richter, 1958, p. 353). For this reason Richter's California values are not included in table 8.

The Hebgen Lake earthquake had a maximum local intensity and a limit of perceptibility greater than any other recorded shock in Montana.

MAGNITUDE

The most generally accepted method of comparison is that of instrumentally determined magnitude. A logarithmic-magnitude scale is objective and precise (reproducible), but even it has drawbacks and is constantly being revised for certain classes of earthquakes (Richter, 1958, p. 338–371). Magnitudes have been determined for the three previous Montana shocks mentioned above. These are: 6¾ for the 1925 shock, and 6¼ for the other two (Gutenberg and Richter, 1954, p. 223, 245). Richter's correlation of magnitude and intensity, reproduced in table 8, shows in these instances that magnitude and maximum intensity are fairly comparable measures of earthquake size.

The Hebgen Lake earthquake has been assigned a preliminary magnitude of 7.1 (Pasadena). This agrees with a maximum M.M. intensity of IX–X in the meizoseismal area north of Hebgen Lake and again confirms Richter's approximate correlation of magnitude and maximum intensity which lists the Hebgen Lake shock as Montana's largest.

SURFACE FAULTING

The recently proposed method of Tocher (1958) opens new possibilities for geological evaluation of large shallow earthquakes. One of Tocher's conclusions is that for each area a threshold magnitude must be exceeded before surface-bedrock faulting can be expected, but above this magnitude nearly all shallow shocks will be accompanied by surface faults. In the California-Nevada area the threshold magnitude is about 6. For Montana, with very few earthquakes to consider, the threshold appears to be 7. Among recorded shocks, only the one at Hebgen Lake has broken surface bedrock. Along the west side of the Madison Range, fault scarps at least as old as 1770 (p. 27) attest to another large earthquake in the immediate vicinity, but there is no way of knowing whether all scarps or all movement on the scarps resulted from one earthquake. Minor reactivation of a portion of that fault zone during the Hebgen Lake earthquake (p. 78) suggests that many events are represented.

Tocher gives an empirical relationship between surface-fault breakage and energy released by certain earthquakes. Because magnitude and energy are also related systematically, this offers an indirect way of determining approximate magnitude or of comparing earthquakes by field criteria alone. The method uses an lD factor in the energy equation

$$E = 3.4 \times 10^{17} \, (lD)$$

where l, in kilometers, is the maximum length of the area showing surface-fault breakage; and D, in centimeters, is the maximum displacement (net slip) on a surface fault.

The lD factors for different earthquakes can be directly compared and used as a measure of the size of an earthquake (table 8). If we use $l = 25$ and $D = 610$ for the Hebgen Lake shock (Witkind, chapter G), this earthquake also falls within the limits of values for lD in table 8. It is the smallest of the magnitude 7–8 earthquakes so far studied by this method, but it is larger than any of the magnitude 6–7 earthquakes compared. Tocher (1958, fig. 3) shows that this method is not absolutely comparable to magnitude but gives a good approximation. If energies for earthquakes above magnitude 6¾ in these areas are calculated and compared with energies derived by other methods the results are likewise convincing. Energies computed directly from magnitude (Richter, 1958, p. 366, equation 10) are slightly higher than those computed by the lD method. Energies computed by another method (Byerly and DeNoyer, 1958) agree almost exactly with those derived by the surface-fault method, but only two earthquakes can be checked.

LIMITS ON THE SURFACE-FAULT METHOD

One does not expect an empirical method to apply beyond the limits of the data; and it is not surprising, therefore, that the surface-fault method does not work in some areas. Any attempt to use it for Assam (1897) gives a grossly inadequate energy, and the method fails completely for the Bihar-Nepal shock (1934) which resulted in major slumping without conspicuous surface faulting. If, however, the method is applied to the great Yakutat Bay shock in Alaska

(1899), a large and apparently adequate energy is indicated. The remainder of this chapter attempts to show why, in some cases, the surface-fault method may fail to yield adequate energy values, and how this information can shed some light on the geology and geophysics of an area.

Three major assumptions were used or implied in construction of the surface-fault method. The basic assumption is "that the energy of an earthquake is dependent mainly on the volume within which it has been stored" (Tocher, 1958, p. 151). This implies that the strength of the rock is constant from place to place. The equation empirically relates ground breakage to energy, but fundamentally it depicts the energy-volume relationship of the basic assumption. This follows from Tocher's discussions which require that l be indicative of the length of a strained volume of rock, that D be roughly proportional to width, and that depth (thickness) be constant. When these dimensions are assigned to a strained volume of rock the second major assumption (constant depth of strain) is apparent. This is necessary if changes in the lD factor are going to measure the entire change in energy.

Bullen (1955) states that strength may vary from place to place. He asserts that strength may be considerably greater where very large earthquakes occur. To the extent that strength does vary, the basic relationship between energy and volume must be modified. Depth of strain likewise may vary, as demonstrated by Byerly and DeNoyer (1958) and as seems probable from a consideration of known variations in crustal thickness and the focal depths of earthquakes. Absence of large shallow earthquakes in ocean basins known to have very thin crusts suggests that crustal thickness may be a dominant factor in determining the maximum possible size of an earthquake. Whether or not this is true, strength and depth of strain are probably interrelated so that separation of the individual effects on the lD factor may be difficult. Empirical lumping of all such effects, and also of the geological factors discussed below, seems a rational approach at present; but in areas remote from the original study, Tocher's constant (3.4×10^{17}) may not be valid.

The third major assumption is that the surface fault is identical with (or bears a constant relationship to) the fundamental fault required by the elastic rebound theory. This last assumption is most difficult for geologists to accept because layers of conspicuous structural contrast are seen in almost every area. In the Hebgen Lake area, for example, profound structural differences are known between Precambrian metamorphic and Paleozoic sedimentary rocks, Paleozoic and Mesozoic rocks, and between any of these and Cenozoic deposits. Doubtless there are vertical and horizontal differences both structural and lithologic within the Precambrian crystalline rocks as well. The degree to which these contrasting units can deflect, damp out, or exaggerate the fundamental fault is, of course, unknown; but one can assume that surface faults are not all mirrored at depth, and also that deep strong faults may be qualitatively and quantitatively modified when passing upward through successive units of contrasting rock. (See Witkind, chapter G; Slemmons, 1957.)

For many years geologists have known about vertical and horizontal changes in faults. As many of these faults represent "fossil" earthquakes, a study of them may reveal something about the surface expression of present-day earthquake faults. In mines some faults are seen to stop before they reach the surface, or to stop, change direction, or divide into countless smaller faults when passing from one kind of rock into another (Knopf, 1929; Newhouse, 1942). Furthermore, there is an intimate relationship between faults and folds, and faults are characteristically different in different parts of a fold. Preexisting faults, folds, joints, and bedding planes are known to change the direction and type of later faults. Evidence is also accumulating, to show a common transition between tear and thrust faults. Finally, where thick surficial blankets, strong lithologic differences, or poor bonding between layers exist, deep tear faults may be represented at the surface as thrust faults in some places and as echelon folds, normal faults, or fissures in other places. (See Richter, 1958, p. 180, fig. 13–7; Hills, 1953, p. 39, 133; Fath, 1920.) For all these geological reasons, it seems inevitable that the surface indication of earthquake-producing faults sometimes will be imperfect or nonexistent. If faults can change so dramatically, then energy release determined from them must likewise change.

The elastic-rebound theory demands a fault for every tetonic earthquake, but we know from the threshold concept introduced earlier that only those earthquake faults associated with large shocks extend to the surface. Apparently preearthquake strain attenuates upward away from the focus. Where elastic strain preceding an earthquake accumulates all the way to the surface (as has been demonstrated for the San Andreas fault), or extends nearly that far, it is possible for Tocher's surface-fault method of energy evaluation to give reasonable values. In other instances preearthquake elastic strain may stop short of the surface, and any deformation which ultimately shows up may be a derived thing subject to all the known vagaries of faults.

Two recently proposed alternatives to elastic rebound (Orowan, 1960; Griggs and Handin, 1960) seem to yield the same results—displacement by faulting, and sudden release of stored energy in the form of elastic waves radiating from the focus. Either the "creep instability" of Orowan or the "shear melting" of Griggs and Handin culminates in sudden movement concentrated in thin zones. By any geometric definition these zones are faults. The quadrantal distribution of initial compression and dilatation observed instrumentally (Richter, 1958, p. 197, 304) confirms the belief that any earthquake mechanism must include faulting.

The new hypotheses stress rapid propagation of faults away from the focus by planar creep or melting without fracture in the ordinary sense, but it is obvious that faults which appear at the surface during earthquakes are not propagated all the way by shear melting or even by creep instability. Rather they seem to be brittle fractures brought about by a very rapid transmittal of stresses occasioned by sudden fault displacement at depth.

The importance of preexistent planes of weakness in guiding or permitting surface faults cannot be overemphasized. Some faults, like the San Andreas, remain active for one or more geologic eras, and many are active for millions of years—long enough to generate mountain fronts by countless separate movements along persistent fault zones. Such long-lived structures seem to argue against the new hypotheses. However, if the new hypotheses are valid for focal depths, then an upward change in the nature of faulting is required. This does not rule out a direct or indirect mechanical linkage, even a geometric continuity between the fault at the focus and one which may appear at the surface. Possibly the semipermanent flaws in the crust act as imperfections in the confining jacket of a high-pressure experiment and therefore help to localize instability in the deep seismogenic zones.

In recent years a seismological method has been devised for determining the direction of first motion on earthquake faults at the focus (Byerly, 1955). One striking result of this work is that nearly all large earthquakes seem to result from faults with a strong strike-slip component. Geologists, on the other hand, do not find a corresponding preponderance of tear (transcurrent, wrench, strike-slip) faults in the course of routine mapping. Several explanations are available, and one of these may be that deep faults sometimes change character, as described above, before we see them. This concept of different but associated structures at different levels is an old one. Many reasons can be found for believing that normal faults should be more abundant high in the crust, and an attractive explanation of thrust faults (Hubbert and Rubey, 1959) now gives an excellent reason for supposing that thrust faults also are favored in a near-surface environment. (See Gilluly and others, 1951, p. 176, for a live example of a slow surface thrust without earthquakes.) Deep thrusts or deep normal faults are by no means ruled out, but it would seem either that they are less abundant than deep tear faults or that, when present, they break too easily to cause a major earthquake.

At Hebgen Lake, nearly pure dip-slip movement is expressed at the surface in earthquake faults. A suggestion of left-hand shear is found in a few places but is certainly not conclusive (Seismological Society of America, 1959, p. 419). Pardee (1950) shows two echelon fault zones near Hebgen Lake, one of which, the Madison Range fault, was reactivated slightly by this earthquake. Echelon systems of normal faults have been interpreted elsewhere as surface results of deep strike-slip movement (Fath, 1920). The geologist, therefore, should not be alarmed if seismological fault-plane solutions show a strike-slip component miles below at the focus. On the other hand there is no rule which says a magnitude 7.1 earthquake in Montana has to result from strike-slip movement, and there is abundant evidence for late Cenozoic block faulting, only a small part of which appears to be echelon, all over the region (Pardee, 1950). Tocher's method did work at Hebgen Lake, and this permits the alternative hypothesis that the faults we see, though somewhat frayed and distorted at the surface, may be deep and continuous structures which rather precisely depict the fundamental fault or faults required by most theories of earthquake mechanism. Without a seismologically determined direction of faulting at depth there is no way of estimating the kind of lost motion in the geologic section above the focus.

CONCLUSIONS

By any of the three methods of rating earthquakes used here the Hebgen Lake shock is larger than any previously recorded shock in Montana, and all three widely divergent methods give comparable ratings. Because of the many factors which conceivably could invalidate Tocher's method, one is impressed that it works so well at Hebgen Lake. A study of why, in a given area, energy indicated by the surface-fault method does or does not compare with energy indicated by magnitude should yield new information on the geology and geophysics of that area.

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Reactivated Faults North of Hebgen Lake

By IRVING J. WITKIND

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THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

REACTIVATED FAULTS NORTH OF HEBGEN LAKE

By Irving J. Witkind

ABSTRACT

Major normal faults along the northeast shore of Hebgen Lake were reactivated during the earthquake of August 17, 1959. Fissures, and new fault scarps which dip valleyward (southward), appeared in surficial material along or near old fault traces. The dominant motion was dip slip with the blocks southwest of the reactivated faults thrown down.

The Red Canyon and Hebgen fault scarps are spectacular features that extend unbroken across hills and valleys. The Red Canyon fault scarp, well exposed along the south flank of Kirkwood Ridge, is about 14 miles long and as much as 14 feet high near its midpoint. The Hebgen fault scarp parallels the north shore of Hebgen Lake and is about 7½ miles long and as much as 20 feet high near its midpoint.

New fault scarps are mainly in colluvium, and commonly a small prism of colluvium is left attached to the bedrock above the scarp. In places along the base of the fault scarp, the gap between hanging wall and footwall is large enough to accommodate a man. Where the material of the hanging wall has subsided into the gap, a "subsidence zone" in colluvium parallels the fault scarp. Locally, subsidiary fractures parallel to the major fault scarp cut the material of the hanging wall or footwall, and a wedge of colluvium drops into and fills the gap. Where the colluvium of the hanging wall is cut and so offset, a "gravity graben" is formed between facing scarps. Where the colluvium of the footwall is cut and offset, the resultant scarp faces valleyward and a "longitudinal step fault" results. Each type of fault scarp merges with the others. These offsets in surficial material mask the true throw on the reactivated faults.

INTRODUCTION

On the night of August 17, 1959, a series of faults were reactivated, and a major earthquake was felt in the Hebgen Lake area. This report includes a discussion of the main fresh fault scarps which appeared north of Hebgen Lake along or near the traces of the old faults and related features. The term "Fault", as for example "Red Canyon fault", is here used to refer to a fault known to have moved repeatedly prior to August 17, 1959. "Fault scarp" as used here refers to a scarp that formed during the earthquake and that locally shares the trace of an older fault. For example, the "Red Canyon fault scarp" coincides with, or is close to, the "Red Canyon fault."

At least four normal faults north of Hebgen Lake were reactivated on August 17, 1959. Two of these, the Red Canyon and Hebgen faults, are major structures that markedly influence the topography. The other two, the Kirkwood and West Fork faults, are small and inconspicuous, both structurally and topographically. The two major faults are described in a companion paper on the structural geology of the area (chapter R), and all four faults are shown on the geologic map (pl. 5). The character and extent of the new scarps are shown in plate 2.

The reactivated faults are expressed both as spectacular scarps as much as 20 feet high in unconsolidated material (fig. 16) and as gaping clefts which have little or no displacement. Commonly each scarp is paralleled by many smaller scarps. Most of these are in the unconsolidated material of the hanging wall, although a few cut the footwall.

FAULT SCARPS

The main fault scarps face the valleys and dip valleyward, which indicates that the faults are normal. This implies either that the ground south of the fault was dropped or that the ground north of the fault was raised. The topography and geologic pattern of the area suggest that the first of these implications is correct—the ground south of the fault dropped. This implication is confirmed by geodetic surveys made before and repeated after the earthquake by the U.S. Coast and Geodetic Survey (chapters C, I).

The scarps vary in height from place to place; but, in general, they are lowest near the ends and highest near the midpoint. This implies that the blocks of ground south of the scarps were warped as they were dropped, as measurements around the lake basin have confirmed (chapter I).

The dominant motion, as indicated by markings on

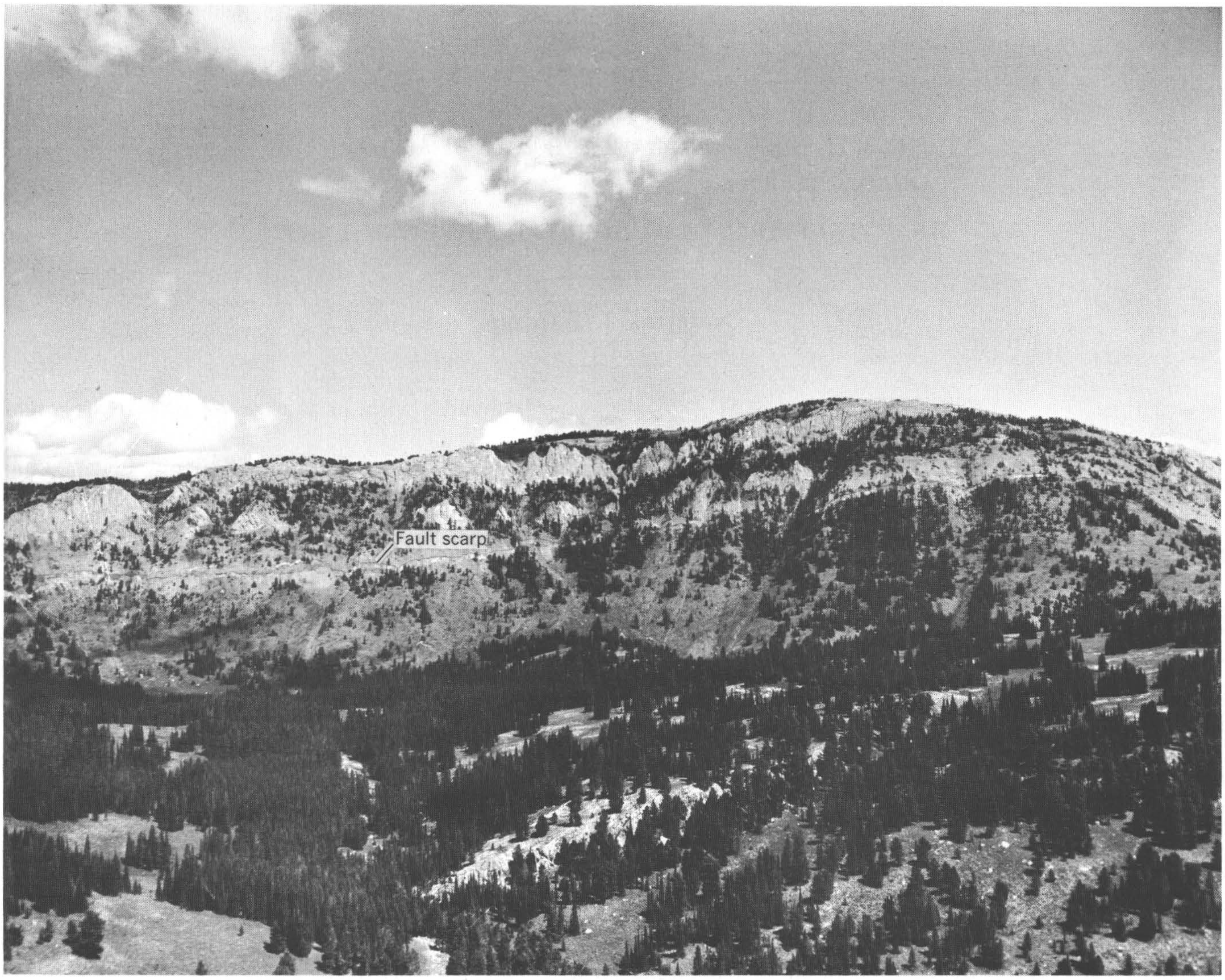

FIGURE 16.—The Red Canyon fault scarp as exposed along the south flank of Kirkwood Ridge. Several longitudinal step faults are shown near the left edge of the photograph.

the scarps along all four reactivated faults, was vertical (dip slip) with little or no horizontal (strike slip) component.

RED CANYON FAULT SCARP

After the earthquake, the trace of the Red Canyon fault was reflected by fissures and fresh scarps that extend for about 14 miles from the center of sec. 22, T. 12 S., R. 5 E. to the SW¼ sec. 19, T. 11 S., R. 4 E. (pl. 2). Near its southeast end the reactivated fault appears as a series of small gaping discontinuous fractures. Traced northwestward, these fractures pass into small local scarps about 2 feet high which follow low embankments composed of outwash and are either vertical or dip southward toward the downthrown south block. The embankments are tentatively interpreted as the eroded scarps of older movements. The scarps end to the northwest, and, near the junction of Highways 499 and 191, two scarps displace the roads; the southwestern scarp ends several hundred feet west of Highway 499, whereas the northeastern scarp continues to the northwest across a broad alluvial flat. The latter scarp increases in height gradually until it is about 14 feet high near Blarneystone Ranch (SW¼ sec. 9, T. 12 S., R. 5 E.), where it dips southward, the south side is downthrown, and the rake of the slickensides on the scarp face is 90° (fig. 17).

Directly east of Blarneystone Ranch (secs. 9 and 16, T. 12 S., R. 5 E.), a fresh scarp follows small preexisting arcuate escarpments that face valleyward, that is, southward (figs. 36, 37). This fault scarp

FIGURE 17.—Red Canyon fault scarp as exposed east of Blarneystone Ranch (SW¼ sec. 9, T. 12 S., R. 5 E.), showing general nature of the unconsolidated material cut and offset. Rake of slickensides is 90° along the scarp in this locality. Area in center of picture is broad subsidence zone tilted toward fault scarp. Scarp height is about 14 feet; corrected surface displacement is about 10 feet.

seems to be on a buried extension of the Red Canyon fault and is therefore discussed as part of the Red Canyon fault scarp. The arcuate escarpments rim broad gently sloping benches which are covered by surficial deposits, chiefly sand and gravel, and are underlain, at least locally, by volcanic rocks. As the new scarp outlines the base of these escarpments, there is a strong implication that the escarpments are controlled by older faults.

The fault scarp passes beneath the domestic quarters and storage sheds of the Blarneystone Ranch (fig. 6*C*). Buildings north of the fault scarp were crushed. Most of the house south of the scarp still stands although severely damaged.

Just west of the domestic quarters on the Blarneystone Ranch the scarp gives way to small fissures which end against a cliff composed of volcanic rocks. The volcanic rocks are overlain by surficial deposits, and these are broken by fissures of minor displacement that continue the northwest trend. These fractures and minor scarps cross the hillsides northeast of Grayling Arm of Hebgen Lake and locally appear as opposed facing scarps outlining shallow grabens which trend westward. In a few places, the scarps face valleyward. A zone of these fractures about 1,000 feet wide extends to the mouth of Red Canyon. Beyond that point the zone can be traced along the east side of Red Canyon, where it dwindles to a width of about 50 feet and consists of three or four low similar scarps. None can be designated as the main fracture.

Near the N½ sec. 1, T. 12 S., R. 4 E., the zone gives way to a well-defined scarp about 6 feet high that dips southwest or valleyward at about 55°. The scarp increases in height to about 10 feet and extends unbroken to the northwest across Red Canyon Creek and along the south flank of Kirkwood Ridge (pl. 2). Wherever exposed, the Red Canyon fault scarp is normal and dips southwestward, or valleyward, from 50° to 85°. The valley side is downthrown. The scarp maintains an average height of about 10 feet, although locally it increases to about 20 feet. Near its west end it decreases in height to about 2 feet.

The scarp either coincides with, or is parallel to, the Red Canyon fault, which follows the broad convex northward curve of Kirkwood Ridge (pl. 5) and joins the Kirkwood fault near the northwest end of the ridge (SE¼ sec. 19, T. 11 S., R. 4 E.). At the junction a new scarp about 2 feet high splinters off the Red Canyon fault scarp and extends eastward for about 1,000 feet along the trace of the Kirkwood fault (pl. 2). The new scarp faces and dips southward but does not extend the full length of the Kirkwood fault.

HEBGEN FAULT SCARP

The Hebgen fault scarp parallels the northeast shore of Hebgen Lake, and appears both as clefts in the ground and as a well-defined scarp. The scarp extends for about 6 miles northwest from the W½ sec. 4, T. 12 S., R. 4 E., to the center of sec. 15, T. 11 S., R. 3 E (pl. 2). Near its southeast end the fault scarp passes into a series of small fissures which are near and parallel to the contact of colluvium and bedrock. These form a zone about 100 feet wide that extends eastward for about 1½ miles and dies out northeast of Lakeview (N½ sec. 10, T. 12 S., R. 4 E.).

The height of the scarp varies. Near its ends the scarp is about 2 feet high, but this increases gradually toward the center, and for most of its length the scarp is about 10 feet high. In a few places, especially near the center of sec. 23, T. 11 S., R. 3 E., the scarp height increases to about 20 feet.

The scarp faces valleyward, and generally dips 60° to 85° southward; in a few places it is vertical. The southwest side is downthrown.

Near the SE¼ sec. 23, T. 11 S., R. 3 E., the fault scarp splits to form two or three small splinter segments in both walls.

In at least two localities, one in the NE¼ sec. 31, T. 11 S., R. 4 E., and the other near Hebgen Dam (NW¼ sec. 23, T. 11 S., R. 3 E.), the scarp is deflected upslope and around thick deposits of unconsolidated material on steep slopes (p. 41–42). In both places the scarp more or less clearly follows the margins of the surficial deposit. The fault itself is concealed but probably continues as a linear feature beneath these deposits (pl. 5). By contrast, where the fault crosses the thick but gently sloping alluvial cone at the mouth of Kirkwood Creek (NE¼ sec. 25, T. 11 S., R. 3 E.) the fault scarp is not deflected but persists as a series of low scarps which face valleyward.

KIRKWOOD FAULT SCARP

The Kirkwood normal fault has been traced about 2 miles westward from the east edge of sec. 20, T. 11 S., R. 4 E., to the center of sec. 19, T. 11 S., R. 4 E., where it joins the Red Canyon fault (pl. 5). Of this length only about 1,000 feet at the west end has been reactivated to form a south-facing scarp about 2 feet high. The new scarp coincides with the trace of the established fault.

WEST FORK FAULT SCARP

Northwest of the Kirkwood scarp, a new scarp extends for about 2 miles from the W½ sec. 24, T. 11 S., R. 3 E. to the E½ sec. 18, T. 11 S., R. 4 E. (pl. 2). The scarp, which follows the West Fork fault, trends about N. 65° E., parallel to and about 1 mile northwest of the western part of the Red Canyon fault scarp. Near its west end the West Fork fault scarp is about 4 feet high and faces valleyward. To the northeast, the scarp decreases in height and finally passes into a series of short fissures.

RELATION OF FAULT SCARPS TO SURFICIAL COVER AND TOPOGRAPHY

The scarps and fissures displace unconsolidated or weakly coherent material, chiefly colluvium, which has been omitted from the geologic map (pl. 5) except where thick. No clearly cut and offset exposures of bedrock were found. Where the fault scarp abuts bedrock, the bedrock has parted along the fault trace; but the breaking seems to be along preexisting joint surfaces or bedding planes, and the general impression is that the rocks have fallen apart rather than been torn apart.

Commonly both the scarps and the clefts in the ground are a short distance downslope from the contact of the surficial cover with the bedrock. A small prism of surficial debris is left, therefore, attached to the underlying bedrock.

The local dip of each scarp is related to the kind of surficial material cut. Scarps in soil are vertical; in colluvium they are less steep and dip 60° to 85° (fig. 18).

Colluvium in the hanging wall is more fractured than identical material in the footwall. Along a typical segment the hanging wall may be cut by as many as 20 small fissures, whereas a similar width of the opposite wall may be cut by only 2 or 3 small fractures (fig. 26).

In a few places the scarp is at the contact between the surficial debris and the bedrock which forms the footwall; at such places the bedrock surface is striated (fig. 19). Locally, where near bedrock, the fault scarp parallels the contact of colluvium and bedrock, and, consequently, this contact deviates from the regional linear trace of the scarp.

FIGURE 18.—Typical fault scarp. Dip is vertical where fault scarp cuts soil. In the underlying colluvial and alluvial debris the dip is less steep and ranges from 60° to 85°.

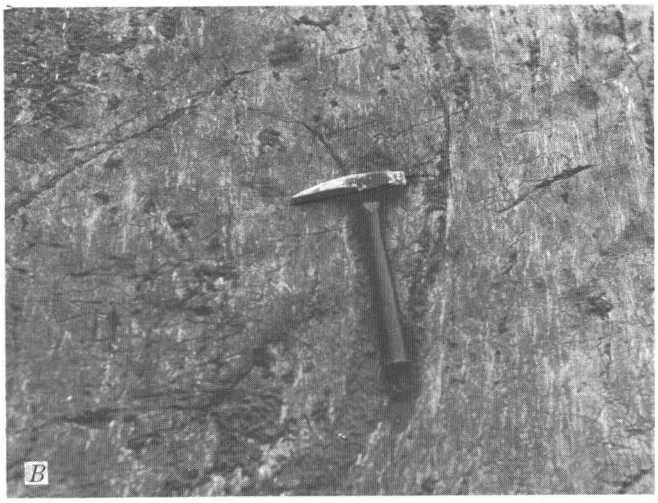

FIGURE 19.—Red Canyon fault scarp at contact between bedrock and colluvium. *A*, Footwall locally formed by steeply dipping strata; elsewhere footwall is at base of small prism of colluvium still attached to bedrock. *B*, Face of bedrock footwall striated by downward movement of the colluvium that forms the hanging wall.

In general, where the colluvial cover is thin and the bedrock is fairly well exposed, the fault scarp coincides with, or is near, the trace of the old fault. Good examples of this are along the Red Canyon fault near the west end of Kirkwood Ridge (W½ sec. 20, T. 11 S., R. 4 E.), along the Kirkwood fault (E½ sec. 19, T. 11 S., R. 4 E.), and along the West Fork fault (N½ sec. 24, T. 11 S., R. 3 E.). Where the surficial cover is thick, however, and the bedrock concealed, the position of the new scarp in relation to the projected trace of the reactivated fault seems to be a function of both the angle of slope and the thickness of the surficial deposits. For example, the scarp does not deviate from its established trace where it crosses thick surficial deposits that lie at low angles of repose. Thus, the Hebgen fault scarp is not deflected where it crosses the alluvial cone at Kirkwood Creek (sec. 8, T. 12 S., R. 5 E.) (p. 40). By contrast, where the slope is steep and the surficial detritus is thick, the scarp is deflected around and outlines the detritus. Where the colluvial debris is outlined, it may have been both thick enough and inclined steeply enough to act as a coherent unit. The contact of the colluvium with the underlying bedrock may have acted as a plane of easy gliding. At the time of the earthquake the mass of colluvium probably shifted slightly downslope on this plane to take up the displacement on the concealed bedrock fault.

Excellent examples of scarps outlining surficial debris are widespread—one of the best is just northeast of Hebgen Dam (fig. 20). Here the scarp is fairly straight until it reaches the large colluvial cone on which the northeast abutment of the dam rests. At this point the scarp curves abruptly upslope along the east edge of the cone, then cuts across the neck of the cone, and

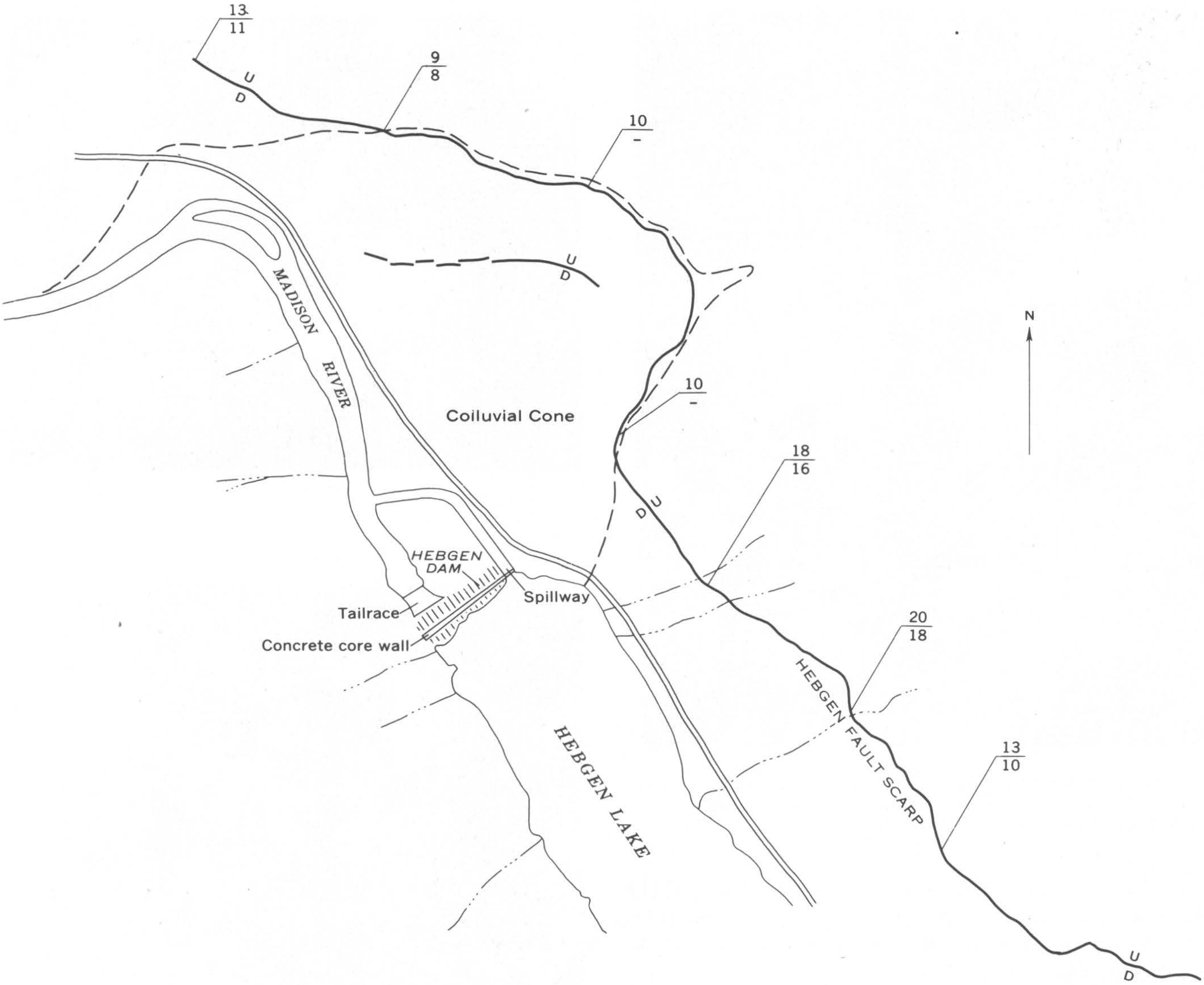

FIGURE 20.—Deflection of Hebgen fault scarp around colluvial cone near damsite. Numerals above the platform indicate approximate scarp height; numerals below platform indicate corrected surface displacement. Missing numerals below platform indicate that scarp height was influenced by factors other than those used in determining corrected surface displacement. (See text, p. 49.)

ultimately descends on the west side. From this point it continues northwestward as a straight scarp.

The rake of slickensides on the fault scarps indicates the local direction in which the colluvium has slumped. Along both the ridge tops and the valley bottoms the slickensides are vertical, whereas along ridge flanks they are inclined in the same direction as the ground surface. Hence, the rake of the slickensides is eastward along the east side of a ridge and westward along the west side. Further, the angle of inclination of the slickensides is related to the slopes of the ridge flanks. The slickensides thus are steeply inclined where the scarp cuts across a steep valley side and are but moderately inclined where the scarp cuts across the ridge flank of moderate slope.

Locally, a steep scarp is adjacent to steeply dipping consolidated strata, whereas near gently inclined strata the scarp gives way to small scarps and fissures. These relations are by no means universal, but they have been found repeatedly in different localities and along different fault scarps. Just east of Hebgen Lake Lodge (W½ sec. 4, T. 12 S., R. 4 E.), for example, the clefts in the ground end abruptly, and a well-defined scarp begins, near the point where moderately inclined beds give way to overturned steeply dipping strata (pls. 2, 5). The change from fissures to scarp is striking and cannot be related to any change in the nature of the surficial deposits or the topography. The colluvium may have responded as a fairly coherent unit in those places where it overlies steeply inclined beds. In this view the colluvium slumped as an unbroken mass to take up the displacement along the concealed bedrock fault. As a result a single fault scarp was formed. By contrast, where the beds are gently inclined the bedrock displacement was reflected in the colluvium as a series of parallel fractures. As the colluvial segments dropped unevenly, fissures and small scarps were formed.

RELATION OF FAULT SCARPS TO BEDROCK STRUCTURES

The attitude and location of the concealed parts of the faults are not precisely known, but in certain localities along all four faults the fresh fault scarps coincide with exposed bedrock faults. Elsewhere, the fault scarps are offset upslope from the projected trace of the reactivated fault, and here they parallel the concealed faults (fig. 21). The repeated coincidence between the fault scarps and the faults is so striking, that it seems clear that the scarps were formed as a result of movement along the concealed faults.

The reactivation of these main faults north of Hebgen Lake stems from the dropping and tilting of two large preexisting fault blocks within the Cabin Creek zone of faulted and folded sedimentary rocks (Witkind, Hadley, and Nelson, chapter R, p. 205; fig. 22). One of these fault blocks, the Red Canyon block, is north of Hebgen Lake, has a crude oblong shape, elongate northwestward, and is about 7 miles long and 4 miles wide. It lies between the Red Canyon and Hebgen faults (fig. 22). The other fault block is vastly larger and is called the Hebgen Lake block (fig. 22). Its northern boundary is defined in part by the Hebgen fault; its western, southern, and eastern limits are unknown.

SCARP MORPHOLOGY AND CLASSIFICATION

Gilbert (1890, p. 354), in his discussion of the general features of fault scarps, offers four diagrams to illustrate how fault scarps are formed in surficial deposits. These diagrams (fig. 23), originally proposed as hypotheses, are valuable aids in the study of recent fault scarps. Slemmons (1957, p. 367–368) proposes specific names for each type of fault scarp, and his names are used here.

Simple fault scarp.—The dip of the fault steepens as it passes from bedrock into surficial material and a prism of surficial material is left attached to the bedrock to form the footwall. The fault scarp commonly has a gaping fissure at its base.

Subsidence zone.—The gap is closed solely by differential movement of the unconsolidated material which forms the hanging wall.

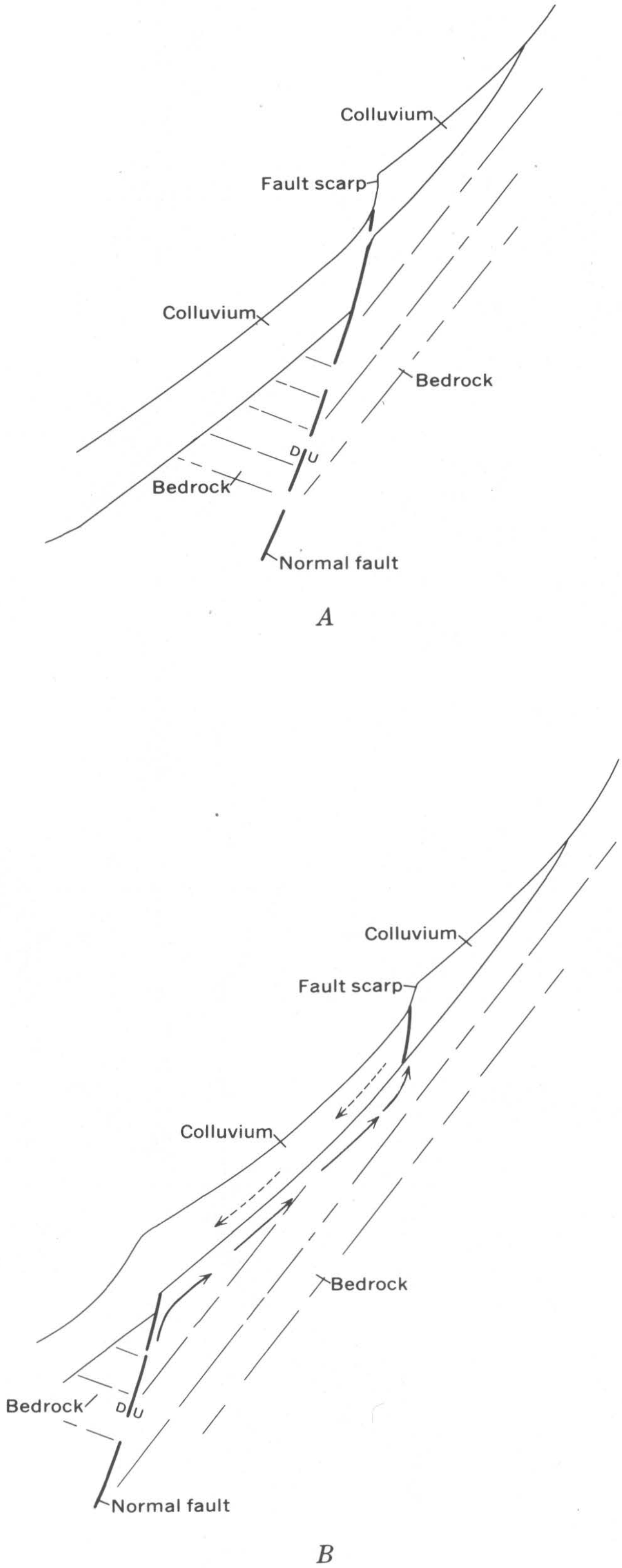

FIGURE 21.—Diagrammatic sketches illustrating possible relations between fault scarps and concealed faults. *A*, Fault scarp directly overlies and coincides with concealed fault; *B*, fault scarp is offset upslope from concealed fault. Dashed arrows indicate direction of slump of colluvium.

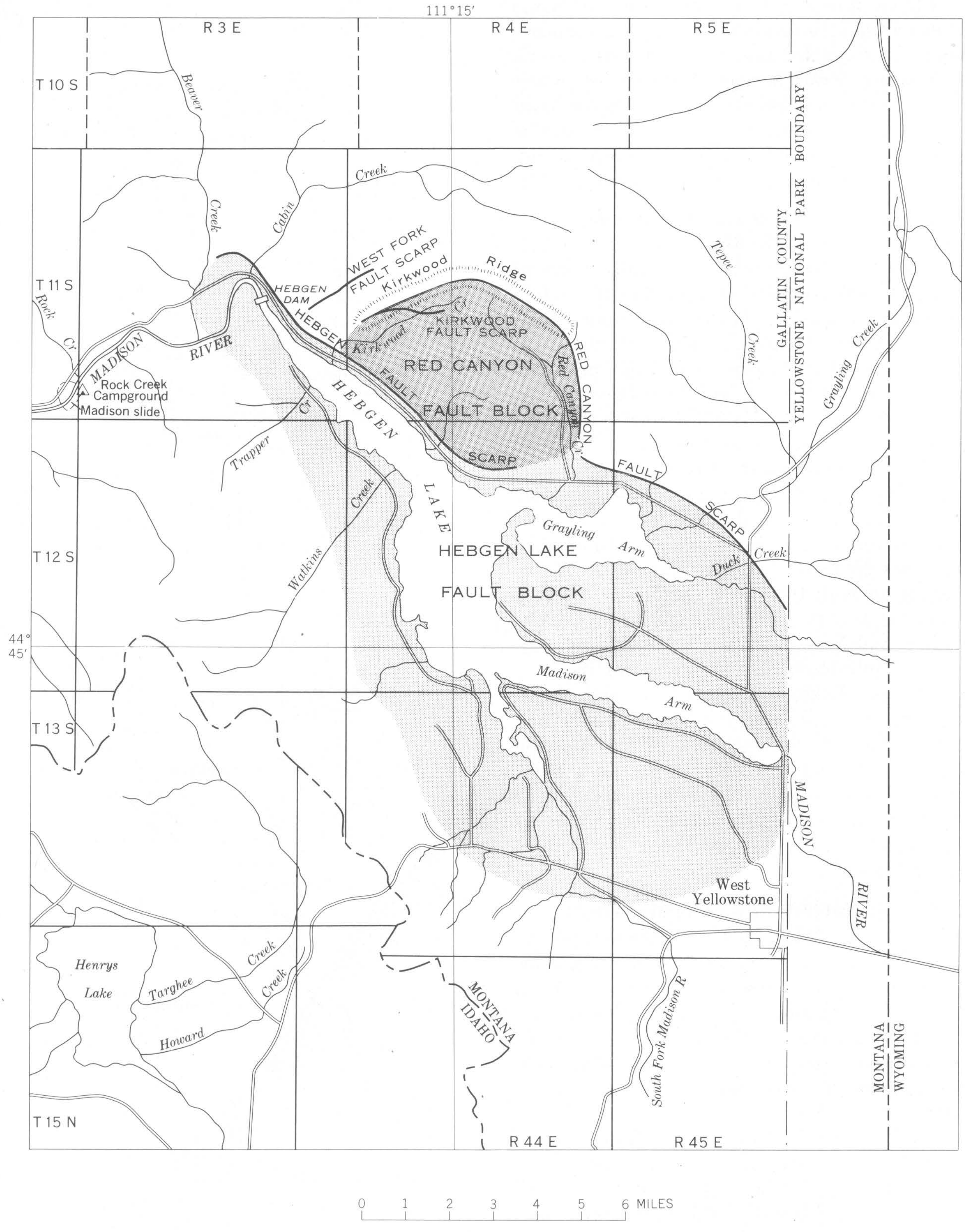

FIGURE 22.—Index map of the Hebgen Lake area showing the trace of the major fresh fault scarps and the location of the Red Canyon and Hebgen Lake fault blocks.

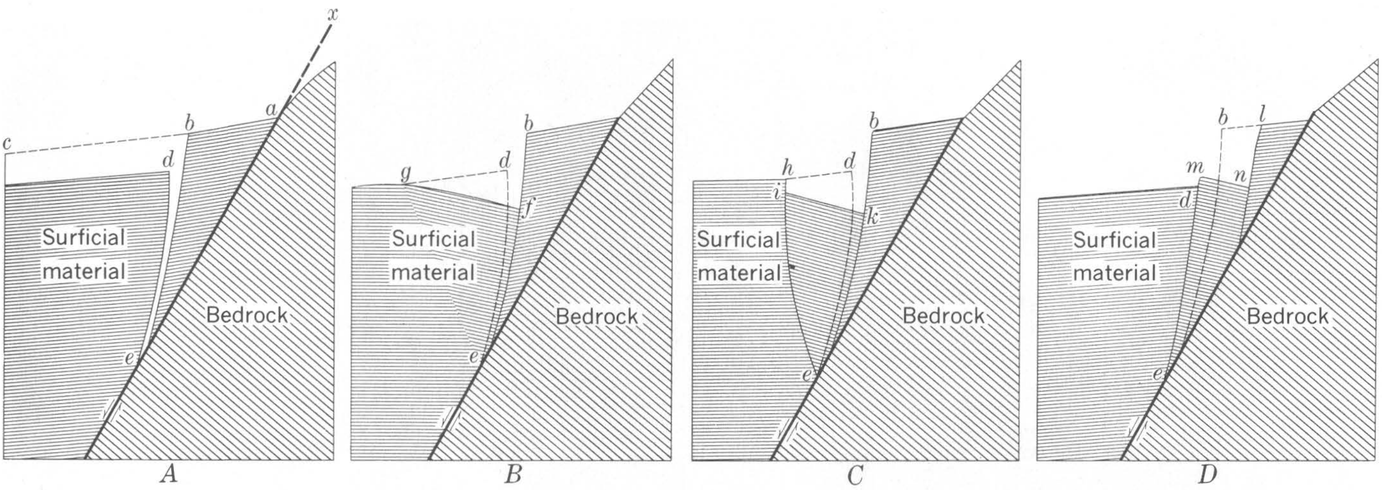

FIGURE 23.—Gilbert's four diagrams. *A*, Simple fault scarp; *B*, subsidence zone; *C*, gravity graben; *D*, longitudinal step faults. Diagrams *B*, *C*, and *D* illustrate how the crevice (*b*, *e*, *d*) is closed by the resultant slump of the weak surficial material. The corrected surface displacement (p. 49), possibly the same as true throw, is approximated by the interval *b*, *d*. All examples indicate general relations between scarp height and corrected surface displacement.

Gravity graben.—A subsidiary fracture cuts through part of the hanging wall resulting in the settling and spreading of a triangular prism (*i*, *e*, *k*) of surficial detritus. A small graben is formed between two scarps facing in opposite directions.

Longitudinal step faults.—Similar in mode of formation to the gravity graben but different in that the subsidiary fracture cuts through the footwall rather than the hanging wall.

Relations similar to those shown in figure 23 are described by Slemmons (1957) in a study of the scarps formed during the Dixie Valley-Fairview Peak, Nevada, earthquakes of December 16, 1954.

The dip of the fault steepens as it passes from bedrock into surficial material, and, as a simple fault scarp is formed, a prism of colluvium is left attached to the bedrock (*b*, *e*, *a* of fig. 23*A*). Both the Red Canyon and Hebgen fault scarps are expressed for much of their lengths as simple fault scarps.

Locally a gap (*d*, *e*, *b*, of fig. 23*A*) large enough to accommodate a man is at the base of the simple fault scarp (fig. 24), but commonly the gap is closed by debris which has fallen from the footwall (fig. 25). This gap may be as much as 10 feet deep and 5 feet wide, but generally it is about 5 feet deep and some 3 feet wide. Where exposed, it can be traced for hundreds of feet along the scarp before it disappears beneath detritus.

In places the gap is closed by subsidence of the unconsolidated material in the footwall, and a subsidence zone is formed (fig. 23*B*; see also fig. 25). Subsidence zones were noted chiefly where the fault scarps cross broad gently inclined flats. In any one subsidence zone the surficial material of the hanging wall commonly is broken by irregular tension fissures (fig. 26) which parallel the main scarp. Several cabins in a subsidence zone at the Hebgen Lake Lodge were tilted 2½° northward toward the footwall.

Elsewhere the gap is closed by a subsidiary scarp that cuts and offsets the hanging wall. As a result, collapse structures (*i*, *e*, *k* of fig. 23*C*) known as gravity graben, are formed and are marked by inward-facing scarps. These graben range in width from a few yards to as much as 100 yards.

Graben are common near the southeast end of the Red Canyon fault along both the hillsides and alluviated flats. Normally the dropped block of surficial material is broken by grouped fractures that are more or less parallel to the flanking scarps. Part of the Culligan house at the Blarneystone Ranch is astride a gravity graben (fig. 26).

Even as some subsidiary fractures cut the hanging wall, others cut the footwall. As shown in figure 23*D*, the resultant block (*m*, *e*, *n*) of the footwall has slid into and closed the gap. The scarps so formed face valleyward and are parallel. Downthrown segments form large elongate steps, and consequently, the fault group is known as a longitudinal step fault. A good example of a longitudinal step fault is along the south flank of Kirkwood Ridge in the N½ sec. 21, T. 11 S., R. 4 E. (fig. 16).

In a few localities, another fault group is formed similar to the longitudinal step fault. Here, however, a fracture or series of fractures cut the detritus that forms the hanging wall (fig. 27). Because the slopes are steep, each splinter segment slides down-

FIGURE 24.—Gap in a simple fault scarp as exposed along the trace of the Hebgen fault near Kirkwood Canyon.

hill slightly, and a series of scarps, all of which face valleyward, are formed (fig. 28). Commonly three or four minor scarps are below the main scarp, which can be recognized because of its continuity and height. Each splinter segment between the subsidiary scarps is about 50 feet wide, and each is broken without displacement by smaller fractures which parallel the flanking scarps. The whole fault group, although resembling the longitudinal step fault group, results from displacement of the hanging wall rather than the footwall. Locally, where the slump is not as marked, the splinter segments will not move as far, and gravity graben are formed between the facing scarps.

These steplike structures in the hanging wall are concentrated on the steeper slopes below the Red Canyon, Hebgen, and West Fork fault scarps. Good examples are along the hillsides that parallel the Grayling Arm of Hebgen Lake (secs. 1 and 12, T. 12 S., R. 4 E., and sec. 7, T. 12 S., R. 5 E.), as well

FIGURE 25.—Red Canyon fault scarp where it crosses Grayling Creek east of Blarneystone Ranch (SW¼ sec. 9, T. 12 S., R. 5 E.). Rake of slickensides is vertical. Debris from footwall conceals scarp and partly fills gaps. Note inclination of subsidence zone behind man who is standing on downthrown part of road. Scarp height about 14 feet; corrected surface displacement about 10 feet.

as farther northwest along the steep slopes near the site of the former Hilgard Lodge (sec. 23, T. 11 S., R. 3 E.).

In a few places, a new subsidiary scarp that faces valleyward is upslope from the trace of the major scarp (fig. 16). This is the result of a subsidiary fracture which cuts and displaces part of the colluvial prism that forms the footwall and is classified as a longitudinal step fault. Elsewhere, a single scarp or a series of scarps, all facing valleyward, are down-

FIGURE 26.—Storage shed and domestic quarters of the Blarneystone Ranch astride a gravity graben. View is to southeast from position on footwall. Brow of main scarp trends obliquely across center of photograph; subsidiary scarp about 2 feet high and parallel to main scarp is shown by arrow. Note group of small parallel fractures cutting colluvium of footwall.

slope from the trace of the main scarp. These new scarps are the result of subsidiary fractures which cut and displace the colluvium in the hanging wall.

Each type of fault scarp merges with the others. The surficial material in the hanging wall may slump into the gap at the base of the scarp and form a subsidence zone. Elsewhere, the simple fault scarp locally becomes a gravity graben through the addition of a subsidiary fracture that cuts the unconsolidated material in the hanging wall. In places the simple fault scarp splits into subsidiary fractures valleyward from the main scarp; the segments of the hanging wall thus formed slump on steep hillsides, and a series of step faults are formed. Still elsewhere,

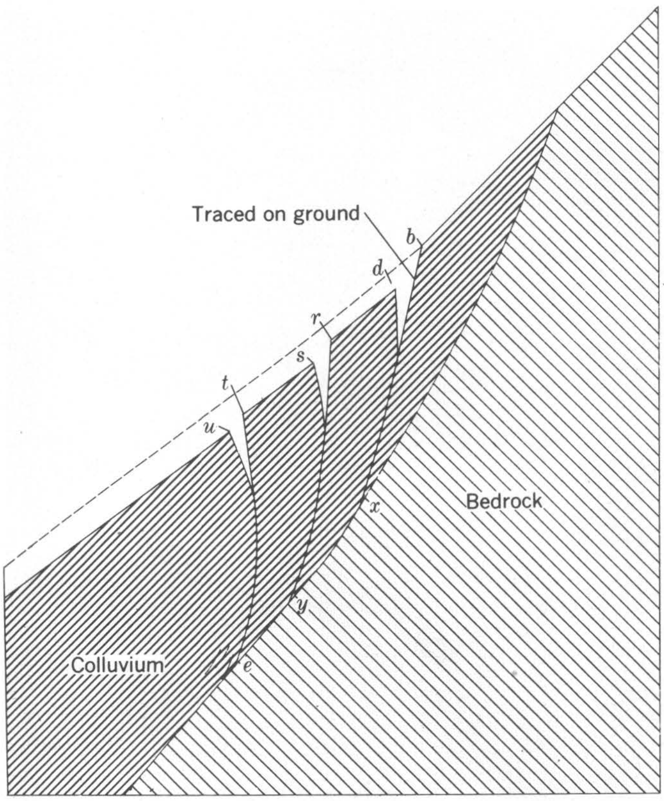

FIGURE 27.—Suggested mode of formation of step faults in the Hebgen Lake area. The major fault scarp as traced on the ground is along the plane *x*, *b*; subsidiary fractures (*y*, *r*, and *e*, *t*) cut the material in the hanging wall and slump of the splinter segments (*t*, *e*, *y*, *s*, and *r*, *y*, *x*, *d*) results in a series of scarps all of which face valleyward.

the footwall may be cut by subsidiary fractures to form longitudinal step faults.

RELATIONS BETWEEN SCARP HEIGHT, CORRECTED SURFACE DISPLACEMENT, AND GEODETIC DISPLACEMENT IN THE EPICENTRAL AREA

Plate 2 shows the relations between scarp height, corrected suface displacement,[2] and geodetic displacement of the epicentral area. Shortly after the earthquake the main scarps north of Hebgen Lake were traced on the ground, and their heights were measured at regular intervals of about 2,000 feet. These data, if corrected, might locally approximate the amount of fault movement and reflect the geodetic displacement at that locality. The determination of absolute geodetic displacement, however, must await the evaluation of data from geodetic surveys just completed by the U.S. Coast and Geodetic Survey.

In several places the new scarp is upslope from a concealed fault (p. 41–42) and therefore has but casual relation to the displacement on the bedrock fault. In the example cited of the fault scarp that outlines the colluvial cone near the damsite (p. 41–42), it is clear that the position and height of the scarp are dependent upon factors other than the actual bedrock displacement. Conversely, in many places the scarp coincides with the reactivated fault, and these localities offer the best possibilities for relating corrected surface displacement to actual fault movement. Consequently, only such localities have been considered. On plate 2 paired figures are shown along the fault scarps. The numerals above the line represent the scarp height, in feet, at that locality. The numerals below the line represent corrected surface displacement. Where the scarp height (upper figure) is given but the corrected surface displacement (lower figure) is omitted, it is assumed that the scarp is offset from the fault, and that any data determined for corrected surface displacement are meaningless.

By reference to figure 23, it is obvious that the scarp height differts markedly from the corrected surface displacement (Gilbert's true throw). In all four illustrations the corrected surface displacement is shown by the interval *b–d*. In both the subsidence zone (*B*) and the gravity graben (*C*) the scarp height (*b*, *f*, and *b*, *k*, respectively) is much greater than the corrected surface displacement; in the longitudinal step fault (*D*) the scarp height is about equal to the corrected surface displacement.

Slemmons (1957, p. 375), in his discussion of the fault scarps formed during the Dixie Valley-Fairview Peak, Nevada, earthquake of December 16, 1954, indicates that the stratigraphic and geodetic displacements were nearly coincident. By contrast, the geodetic surveys completed in the Hebgen Lake area indicate that only locally is there coincidence between corrected surface displacement and geodetic displacement.

In general, the best agreement between corrected surface displacement and geodetic displacement is along the Hebgen and Red Canyon fault scarps northwest and southeast of the Red Canyon block (fig. 22). The greatest deviation is along the southwest edge of the block. Northwest and southeast of the Red Canyon block, the ground north of the scarp was probably raised slightly or remained stable, whereas the ground south of the scarp was downthrown. The corrected surface displacement thus approximates the geodetic displacement. By contrast, along the southwest edge

[2] The term "corrected surface displacement" seems preferable to "stratigraphic displacement" or "true throw" in property describing the results of the present work. True throw implies a degree of accuracy that is not likely to be achieved by the relatively crude geological methods used. Many of the data gathered are tenuous and subject to errors of visual estimation. Moreover, corrections made as suggested by figure 23 are subject to such large errors of judgment that the results may only fortuitously coincide with true throw.

FIGURE 28.—Step faults along the Red Canyon fault scarp where it follows the south flank of Kirkwood Ridge.

of the Red Canyon block the ground on both sides of the scarp went down, the ground south of the scarp more than the ground north of the scarp. The whole block, therefore, went down unevenly, the north side more than the south side. Consequently, there is little relation in this sector between corrected surface displacement and geodetic displacement.

FISSURES

Many fissures are along the trace of, or adjacent and parallel to, the fault scarps and evidently relate to the faults. Others, widely separated on the hillsides and valley floors, clearly are superficial, formed chiefly by slump.

Most fissures related to the faults are short and fairly straight, with jagged and serrated brows. A typical fissure is about 100 feet long, ranges in depth from a few inches to about 3 feet, and has a maximum gap of about 6 inches at the surface. A few are much larger; they gape about 2 to 3 feet, and are as much as 10 feet deep and 500 feet long. All are approximately parallel, and generally in echelon; they form a rift zone as much as 1,000 feet wide.

One possible explanation for a clear fault scarp in one place and fissures in another involves the mode of failure of the underlying rocks; that is, whether the rocks broke as a result of slippage along an established fault plane or flexed downward sharply instead with little or no faulting. At one locality the dropping and tilting of a fault block may have caused the strained rocks to break along an old fault and a fresh scarp resulted. Fissures elsewhere were formed in response to a similar movement of the fault blocks, possibly because the rocks adjusted by warping. Here, the frictional resistance on the fault plane was not overcome.

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THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CHANGES IN THE FLOOR OF HEBGEN LAKE

By Wayne H. Jackson

ABSTRACT

Depth soundings of Hebgen Lake after the earthquake of August 17, 1959, detected a reversal in gradient of the prelake Madison River channel near Hebgen Dam. Evidence of tilting of the lake floor was observed in parts of the lake, but the soundings gave no indication of major faulting of the lake bottom.

INTRODUCTION

Changes in the land surface bordering Hebgen Lake are described by Myers and Hamilton (chapter I). In summary, the major surficial changes accompanying the earthquake were: Tilting and warping of the Hebgen Lake basin; reactivation of the Hebgen fault and exposure of new scarps as much as 20 feet high; and the collapse, extending above the shoreline for several hundred feet, of the banks of the lake in many places. Because of the probability that similar changes occurred in the submerged banks and floor of Hebgen Lake, a detailed survey of the lake bottom from Hebgen Dam to the shallow parts of Grayling and Madison Arms was made in September 1959. The lake was traversed from shore to shore; and the water depths, measured with an acoustic echo sounder, provided continuous profiles of the lake floor along the traverse lines.

The area of greatest change in the lake bottom is within 2 miles of Hebgen Dam where at least 16 feet of landslide debris was deposited in the deepest part of the lake. Subsidence of the floor occurred in this part of the lake, and evidence of tilting was found in the Grayling Arm. The soundings, however, gave no indication of major faulting of the lake bottom.

INTERPRETATION OF PROFILES

The only available maps of the Hebgen Lake basin as it existed before flooding by the lake in 1915 are 1:24,000 planimetric maps based on a predam survey by the Montana Power Co. in 1906. Because of the close agreement between the location of features such as the projected shoreline and parts of the Madison River channel with their known position from latter surveys, the maps are believed to be reliable and have been used to interpret, in part, the sounding data.

Before the formation of Hebgen Lake, the Madison River channel (pl. 1) had developed extensive meanders in the area now called the Grayling and Madison Arms and also in the area joining the two. In the present northwestern part of the lake, where the river was more confined, the channel was relatively straight. After the area was inundated, sedimentation probably took place in the upper part of the Grayling and Madison Arms and in limited parts of the lake affected by drainage from small streams. The lake level has been lowered from time to time, and this has resulted in the development of new stream channels in the sediment-covered areas. Most of the lake bottom, however, seems to have been relatively free of sediment, and the location of the channel by depth sounding agrees very well with that of the original Madison River channel.

In both the Madison and Grayling Arms the profiles show numerous meander scars (pl. 1), and with the exception of profiles 43 and 45, the location of the deeper scars agrees well with the intersecting points of the traverses and the prelake Madison River channel. The disagreement at traverse 45 is probably the result of a slight error in navigation; the traverse crossed the lake at an angle and intersected the channel on a tight meander. Traverse 43 also intersects the prelake river channel at an angle; however, it would require a considerable error in navigation to produce the large channel offset. It is more likely that the river changed its course here during a time when the lake level was lowered. The shape of the scour on profile 43 further substantiates this possibility, because the channel at this point curves to the south, probably to rejoin the main channel.

The west side of the central part of the lake bottom between Grayling and Madison Arms is exten-

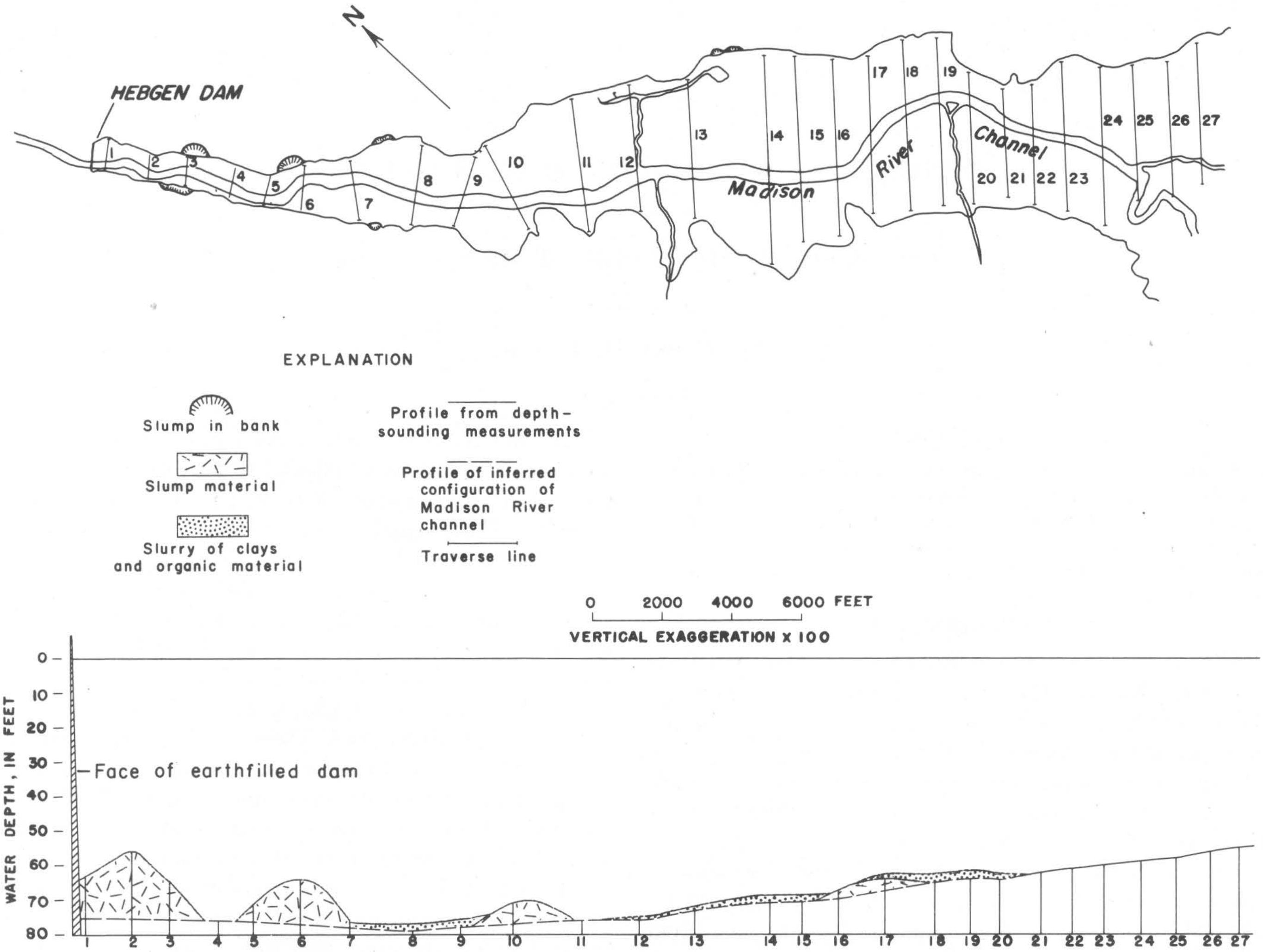

FIGURE 29.—Map and profile of the northwest arm of Hebgen Lake after warping. *A*, Map showing location of former Madison River channel and of depth-sounding traverses made in September 1959; *B*, profile of the lake floor along the channel.

sively terraced, although steepr slopes occur on the east side (profiles 34 and 35). Numerous meander scours cut the relatively level floor between traverses 33 and 35.

Terraces occur in the northwest arm as far as traverse 7; however, meandering of the river seems to have been reduced beyond traverse 24. Two large alluvial fans are evident from the profiles. The larger one, the Watkins Creek fan, is expressed by a gradual slope of the southwest side of the lake bottom on profiles 15 through 24; the smaller fan is evident on the northeast side on profiles 9, 10, and 11.

All the profiles have irregularities that, when taken individually, could be indicative of faults; none, however, has the continuity between traverses that would be expected along a major fault. Many of the changes in slope parallel the river channel and can be assumed to be terrace steps or other erosional features. All the changes in slope are much exaggerated on plate 1 as a result of the 20 to 1 vertical exaggeration, and local fault offsets of only a few feet would be difficult to distinguish from other irregularities in the lake floor.

Three large landslides that changed the shoreline of the lake within 1½ miles of Hebgen Dam (fig. 29) deposited a large amount of debris on the underwater banks of the lake and in the prelake river channel (pl. 1, profiles 2, 3, 5 and 6). Landslide debris may also cover the northeast bank at traverse 1, as suggested by the irregular appearance of profile 1; however, no slides were observed at the surface near this traverse. Smaller slides altered the shoreline between traverses 7 and 8 and between 13 and 14 (fig. 29), but there is no indication of any major change in the lake bottom.

Abrupt changes in the gradient of the Madison River channel could be indicative of tilting or warping of the lake floor. In the Madison Arm, the gradient of the river channel between traverses 37 and 47 is a relatively uniform 5 feet per mile, but between traverses 37 and 35 the gradient increases to about 6 feet per mile. In the central part of the lake joining Madison and Grayling Arms the gradient reaches a maximum of about 8 feet per mile between traverses 37 and 33 and decreases to about 6 feet per mile in the northwest arm of the lake between traverses 11 and 24. The 3-foot per mile increase in slope between traverses 37 and 33 must be the result of the tilting of the lake floor that accompanied the earthquake. In this area the river channel is approximately parallel to the direction of tilt (see Myers and Hamilton, chapter I), whereas between traverses 37 and 47 the channel is nearly perpendicular to the direction of tilt and shows no increase in gradient. Myers' measurements in the area between traverses 37 and 33 indicate a northward tilt of slightly less than 3 feet per mile.

Similar tilting of the unusually smooth floor of the Grayling Arm is suggested by profile 31, where a gentle slope to the north is noticeable. The slope is approximately 3 feet per mile, which is in agreement, within the limits of accuracy of the measurements, with the tilt determined by Myers in this area.

Warping of the northwest arm of Hebgen Lake reported by Myers and Hamilton (chapter I) is confirmed by depth-sounding measurements. A vertical profile of the river channel (fig. 29) shows that the gradient between traverses 27 and 12 is uniform at about 6 feet per mile and decreases gradually between traverses 12 and 8. From traverse 8 to the dam the channel has been partly covered with landslide debris, and control points cannot be so readily determined. The position of the old river channel, however, shown in construction drawings of the dam, is now at least 4 feet higher than the channel in traverse 8; and the present gradient, determined by points at the dam and in traverses 4 and 8, is upstream at about 2 feet per mile. Assuming a prelake gradient of 3 feet per mile downstream in this part of the channel, the relative vertical movement of the dam with respect to lowest point in the channel is about 10 feet, which agrees approximately with Myers' data on the deformation of the lake basin.

Assuming a uniform channel gradient between traverse 8 and the dam (fig. 29), the approximate depth of debris covering the channel can be determined. About 16 feet of material has been deposited at traverse 2, 12 feet at traverse 1, 10 feet at traverse 3, 5 feet at traverse 6, and 6 feet at traverse 10. The material covering the channel between traverses 1 and 7 is undoubtedly part of the several hundred thousand cubic yards of debris resulting from the landslides along the banks described by Hadley (chapter K, p. 121–122).

EQUIPMENT AND MEASUREMENTS

A sonic depth recorder, Edo model 255C, was used for the depth soundings. This instrument, on loan from the U.S. Bureau of Reclamation Hydrology Branch where it is used regularly for reservoir surveys, measures and permanently records water depths of from 2 to 1,400 feet on calibrated paper.

The equipment is mounted in a small boat powered by an outboard motor. A pulse of electrical energy is converted periodically to sound and transmitted downward from a transducer mounted on the side of the boat. When the energy strikes the bottom or any other object having acoustic properties different from those of water, part of it is reflected back to the transducer as an echo. This energy is reconverted into electrical energy for presentation on the electrosensitive chart paper as a dot or a short line. Because the speed of sound in water is nearly constant, the time between transmission of a pulse and reception of its echo is a measure of the distance traversed, which is, in this case, depth.

High lateral resolving power is accomplished by using a transducer that emits a narrow beam of sound pulses of very short duration and high frequency (37.5 kc). Because of the sharp attenuation of the high frequency energy in consolidated sediments, the records would not be expected to reveal stratification of subbottom structures, and the penetration of the sound pulse would be limited to only a few feet of unconsolidated material such as a slurry of fine clay.

The accuracy of the instrument, determined primarily by the timing mechanism and its stability, is rated at 0.5 percent. With correct adjustment and proper operating procedures, the greatest source of error in a series of soundings would be caused by the change in the velocity of sound in water. Water temperature is a factor in the velocity of sound. Temperature measurements were taken twice daily, and the greatest difference measured was one-half degree, which would have negligible effect on the accuracy of the soundings.

The lake level was used as a reference so any large variations from the assumed altitude would result in erroneous soundings. According to records of the

U.S. Army Corps of Engineers, the lake level changed less than 0.1 foot while the measurements were being made, a variation which would have little effect on the accuracy of the soundings. Repeated measurements of complete traverses and comparison of sonically recorded depths with those measured by means of a sounding line indicate that a sonic sounding in the deepest part of the lake is accurate within 1 foot.

Continuous profiles of the lake floor were obtained along 47 traverses having a total length of approximately 35 miles. Two methods of position control were used. In running traverses 14 through 27, the boat was sighted by a transit operator on shore; directions for keeping the boat on course were transmitted by radio and by flag signals. A second transit (theodolite) operator on shore, upon a flag signal from the instrument operator in the boat, noted the azimuth of the boat position along the traverse; the intersection of the azimuth and traverse lines defined the position of the boat at that instant. Azimuth sights were taken at the start of each traverse, at 1-minute intervals during a run, and at the end of a traverse. At the instant he signalled for an azimuth sight, the instrument operator marked the depth recording, thereby establishing a correlation between the depth sounding and its position on the lake.

Correlation points established in this way were approximately 500 feet apart, the exact separation depending upon the speed of the boat. With the exception of a few traverses run during brief windy periods, the regular spacing of the correlation points along the traverses indicated that the boat speed was reasonably constant. No strong currents were noted in making the runs, and there was little difficulty in maintaining a straight course during calm weather; therefore, a more rapid positioning method was adopted for the remaining traverses. Using postquake aerial photographs, prominent landmarks were chosen for the initial and terminal points of the traverse, and the course run between these points was assumed to be a straight line.

The speed of the boat was maintained at about 500 feet per minute; but when greater horizontal detail was desired, higher speeds were used.

Deformation Accompanying the Hebgen Lake Earthquake of August 17, 1959

By W. BRADLEY MYERS *and* WARREN HAMILTON

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

DEFORMATION ACCOMPANYING THE HEBGEN LAKE EARTHQUAKE OF AUGUST 17, 1959 [3]

By W. Bradley Myers and Warren Hamilton

ABSTRACT

The gently sloping obsidian sand plain that forms the surface of most of the West Yellowstone basin south of Hebgen Lake is the south flank of a broad basin of new subsidence produced during the 1959 earthquake. An area 43 miles long and 14 miles wide subsided measurably; there was a measured maximum elevation above previous levels of 1.7 feet. The maximum subsidence, 22 feet, was in part of Hebgen Lake basin where a tract of about 60 square miles dropped more than 10 feet. Subsidence was determined by releveling of bench marks and road profiles and by measurements of lakeshore changes and of heights of new fault scarps. Contours of subsidence can be drawn with high reliability and precision over a considerable part of the affected area. Extensions of the contours across areas of few or less reliable data seem to indicate a broad basin that plunges gently eastward, directly across Madison Valley and the Madison Range, to the West Yellowstone basin and Hebgen Lake. Lessened subsidence continued eastward to the Gibbon River.

The obsidian sand plain of the West Yellowstone basin is broken by scarps 0.1 foot to 3 feet high along faults that were reactivated in the 1959 earthquake. Older displacements along these same structures, produced by normal faulting in the basement rocks, had formed monoclines and faults in upper Pleistocene surficial deposits, with a maximum structural relief of 50 feet. Some new surficial structures that appear on casual inspection to indicate compressional deformation actually are due to refraction in the unconsolidated fill of reactivated normal faults in the basement rocks.

The basin of new subsidence ends obliquely and abruptly on the northeast against reactivated faults with scarps 5 to 20 feet high that trend southeastward near the northeast side of Hebgen Lake. Subsidence is nearly constant for many miles along the zone of major scarps. Two of the three major scarps occur in areas where bedrock is exposed and are limited to places where bedding in the Paleozoic rocks dips steeply toward the basin and is thus oriented favorably to permit sliding down the dip; where the bedding attitude flattens, the scarps dwindle and die out. The subsidence is accounted for almost entirely by new scarps where the bedded rocks have optimum attitudes. It was accomplished by a combination of warping and faulting where attitudes are less favorable, and by warping alone in unfaulted areas where bedded rocks would otherwise have been broken at high angles to directions of easy faulting. The bedding that controls these two faults at the surface can hardly extend to a depth of more than a mile or two, so these scarps cannot directly portray the pattern of deformation in deep basement rocks.

Releveling of bench marks and comparison of the postquake road profile with the prequake profile show that the lower part of the Madison River canyon and the adjacent part of Madison Valley subsided evenly about 7 to 8 feet and that there was no new warping at the Madison Range front. East of Madison Slide and Earthquake Lake, which buried prequake reference points, bench marks and road profiles prove that the upper part of the Madison River canyon subsided 6 to 14 feet. The subsidence of Madison Valley decreased to a small fraction of a foot in a few miles northwest of the mouth of the Madison River canyon. South of the canyon a segment of the Madison Range front fault was reactivated, but the amount of absolute subsidence is not known. Observed shoreline changes at Cliff Lake, west of the upper Madison Valley, indicate that the lake was tilted. A belt 10 miles wide north of the subsided tract was elevated by from 0.1 to at least 1.7 feet.

North-trending Madison Valley was much deformed during late Quaternary time by east-trending faults and warps that extend from Centennial Valley and the Centennial Mountains; other east-trending structures have been superimposed across the older structures of the Madison Range, West Yellowstone basin, and the northwestern part of the Yellowstone Plateau. The 1959 earthquake seems to be part of a relatively new structural system in which older north-trending blocks are being complexly segmented and deformed by the eastward extension of structures from the Centennial region.

INTRODUCTION

The violent earthquake of August 17, 1959, which centered near Hebgen Lake in southwestern Montana, was accompanied by the formation of several large fault scarps and many smaller ones, and by the abrupt subsidence and warping of a very large area. The high fault scarps, many miles long, that formed northeast of Hebgen Lake are described in detail by Witkind (chapter G). Other aspects of the deformation, as revealed by detailed studies of the area south of the lake and of the lake basin are described in this paper. Also considered is the whole deformation pat-

[3] August 17 by local date; August 18 by Greenwich date.

tern in relation to older structures and to the complex structural evolution of the region. The geology of part of the area discussed is shown on plate 5 and described in chapters R, S, and T. Scarps and other features formed during the earthquake and contours illustrating the amount of accompanying subsidence are shown on plate 2. A brief note that summarizes some of the deformation data and our interpretation of them has been published (Myers and Hamilton, 1961).

ACKNOWLEDGMENTS

Incorporated in this report are the results of studies by a number of men both within and without the Geological Survey. Releveling of bench marks northwestward from West Yellowstone, which provided the critical information regarding absolute changes of elevation, was made possible by the cooperation of Capt. J. H. Brittain, Chief of the Geodesy Division, U.S. Coast and Geodetic Survey, and was accomplished by a field party under the supervision of W. N. Grabler. W. K. Cloud, of the Seismological Field Survey of the Coast and Geodetic Survey, performed many services of liaison between that organization and the Geological Survey.

Personnel of the Montana Power Co. provided much information regarding Hebgen Lake and Hebgen Dam before and after the earthquake. Chief Engineer Ray M. Ball and Engineers James Kreitzberg and Glenn Jones are among those to whom particular thanks are due.

The U.S. Forest Service took low-altitude vertical aerial photographs along the new fault scarps and supplied prints of these photographs for our use less than a week after the earthquake. Helicopter flights provided by the Forest Service permitted inspection of many earthquake features from the air.

Releveling of road profiles in the Madison River canyon by the Bureau of Public Roads during the year after the earthquake demonstrated that local subsidence continued there; we are indebted to Lynn D. Tingey of the Bureau for these data.

DEFORMATION IN WEST YELLOWSTONE BASIN

The surface of most of the West Yellowstone basin, including the southern part of the Hebgen Lake basin as well as the area to the south, is a broad expanse of obsidian sand and fine gravel that slopes gently northwestward (pl. 5). These surficial deposits are 200 feet thick where penetrated by an oil-test hole 2 miles northwest of West Yellowstone and more than 240 feet thick on the south shore of the Madison Arm where a water well of this depth bottomed in sand. Major streams heading in the mountains cross the plain; but due to its high permeability, no streams rise on its surface. Near the widely spaced streams the plain has been cut into steplike erosional terraces, but over most of its extent it is a monotonous constructional surface, essentially unmodified since its deposition after the Bull Lake glaciation. The sediments and the dating of the sand plain are described by G. M. Richmond (chapter T).

The sand plain had been deformed before 1959 by a number of small faults and monoclines with structural relief of as much as 20 feet each. Monoclines predominate at the surface, but they are probably related to normal faults at depth. Almost all these structures were reactivated, with minor surficial faulting and tilting, during the 1959 earthquake, and the entire plain was tilted gently northward. The new tilt steepens gradually northward from West Yellowstone. Most of the reactivated structures are in the least tilted part of the flank of the major structure.

1959 STRUCTURES

MADISON ARM FAULT AND MONOCLINE

The structure with the greatest local relief formed south of Hebgen Lake during the 1959 earthquake lies midway along the south shore of Madison Arm (pl. 2). This Madison Arm structure—partly a monocline, partly a fault, and partly a combination of both—is three-quarters of a mile long and trends east-northeastward. Tilts and displacements during the earthquake were greater here than elsewhere on the sand plain, and much of the interpretation of the structures of the sand plain at depth is based upon knowledge of this well-developed structure.

The 1959 displacement was greatest along the western half of the structure. The most obvious new features are north-facing scarps as high as 27 inches that dip very steeply southward beneath the relatively upthrown block (fig. 30). Although there are generally cracks an inch or two wide along the faults, the downdropped block in places is driven tightly in under the relatively upthrown block so that the latter slightly overhangs; this is clear evidence that this near-shore feature is not a lakeward slump.

Monoclinal warping was also produced along the structure during the 1959 earthquake. This flexure is made obvious, where its dips are more than several degrees, by a zone of correspondingly tilted lodgepole pines (fig. 31), and the monocline can in places be traced most easily by following the tilted trees. The new structural relief is generally 2 to 2½ feet, represented chiefly by faulting at some places, by

FIGURE 30.—New scarp in the sand plain. Dip of fault at surface is 85° back under scarp; opposite sides are tight together so that down-dropped block is slightly overhung by the relatively upthrown block. This is a refracted normal fault, not a reverse fault. (See fig. 39.) South of Madison Arm, 5.3 miles northwest of West Yellowstone.

warping at others, and by a combination of both elsewhere.

Along the western half of the structure, where it affects the main surface of the sand plain, flexure and scarps are superimposed on an older structural rise that is about 2 feet high and as steep as 8°. The total ground-surface tilt is several degrees steeper than the new tilting shown by the trees, and the difference represents tilting before the trees grew.

The eastern part of the Madison Arm structure crosses river terraces cut to several levels into the sand plain. The lowest of these was under shallow water in Hebgen Lake before the emergence of Madison Arm that resulted from warping of the lake bed during the earthquake (see p. 76). Only the earthquake structures affect these terrace surfaces; any pre-1959 warping and faulting in this eastern segment occurred before the cutting of the terraces.

The base of the monocline is commonly broken by a nearly continuous miniature underthrust that at the surface dips gently back under the incline of the larger structure (fig. 32). The small structure is called a mole-track thrust, a name adapted from the term "mole track scarp" which T. W. Dibblee (in Oakeshott, 1955, p. 24, fig. 1) applied to a similar-appearing surface structure. These features do indeed resemble the little ridges made by moles (figs. 31, 32). The feature at Madison Arm is a small welt several inches high and marked by an abrupt upward roll of the ground surface that curves irregularly along the base of the monocline in the upper plate of a surficial thrust fault. Thrust displacement is typically 4 to 6 inches. The mole-track thrust was dug out in a number of places; the gentle surface dip continues for a foot or a foot and a half and then steepens abruptly to 45° or more. Pine-needle duff is the chief com-

FIGURE 31.—New scarp and monocline in the sand plain. Fault (left) with tightly closed fissure gives way to an atypically steep and abrupt monocline (center) which gives way in turn to a gentle monocline. A mole-track thrust runs along the base of the monocline, past the head of the hammer. South of Madison Arm, 5.3 miles northwest of West Yellowstone. Photograph by John R. Stacy.

ponent of the material offset at the surface. The similar mole-track thrusts, described by Dibblee from the area deformed during the Arvin-Tehachapi earthquake in California, have been generally accepted as indicators of compression. Although this may be a valid interpretation in California, at Hebgen Lake they are products of normal tensional faulting, as is brought out later. The original analogy with the track of a mole was made by Kotô (1893, p. 328–333) for a similar but much larger rounded ridge of soil that developed along a major strike-slip fault and "* * * resembles very much the pathway of a gigantic mole."

The eastern half of the Madison Arm structure, developed on stream-cut terraces, is a nearly continuous new fault scarp several inches to 1½ feet high and without apparent associated warping. This scarp cuts across a bay of Hebgen Lake that was emergent after the earthquake, and offset of the prequake shoreline demonstrates that large blocks moved along this small structure during the earthquake. Levels were run with a telescopic alidade along the prequake high-water line for the length of the bay in both directions from the new scarp, a total length of about 300 yards obliquely across the strike of the fault. The prequake shoreline of the southern block was uniformly 13.0 ± 0.2 feet above the level of Hebgen Lake of the moment, and the shoreline of the northern (downdropped) block was 11.5 ± 0.3 feet above the lake. Most of the variations in the readings on the north block probably represent the broad sag typically developed in front of each fault and monocline of the sand plain (see page 62).

FIGURE 32.—Mole-track thrust at base of monocline formed during 1959 earthquake. White tapes indicate horizontal (17 feet) and vertical (1.5 feet) components of new monoclinal slope. South of Madison Arm, 5.3 miles northwest of West Yellowstone. Photograph by John R. Stacy.

FAULTS AT HORSE BUTTE

Numerous new scarplets along the south and southwest sides of Horse Butte formed during the earthquake. The scarps form two sets; most strike west-northwest, and fewer strike east-northeast. The west-northwest scarps lie in a zone a mile wide that extends from Horse Butte to Edwards Island. Along nearly all these faults, the south or southwest side is relatively downdropped.

The zone of northwest-trending scarplets is the site of more than a dozen springs formed at the time of the 1959 earthquake. Most of these springs are small and cold, and their temperature (about 47°F) indicates a shallow ground-water source. One spring has a larger flow and a warmer temperature (55°F) than the others, and this water must have come from a deeper source; presumably it has migrated upward along a fault zone. A. Tanner (oral communication, 1959) found this spring to have a much higher radon content than did the others of the group, which is possibly further evidence of a deeper source.

The shape of Horse Butte is suggestive of a fault-block origin.

GRABEN INVOLVING FAULTED BEDROCK

At only one place south of Hebgen Lake was an area of exposed bedrock seen to be displaced by 1959 scarplets. No actual outcrop was cut by scarplets, but bedrock must be involved at a depth of a few feet. This area is in the western part of a small graben near the Basin Ranger Station (west of the South Fork of the Madison River, and south of Madison Arm, along the south edge of sec. 9, T. 13 S., R. 4 E.).

A parallel-sided block of rhyolite, 300 to 400 feet wide and 1,500 feet long, extends eastward from the bedrock hillside and projects through the Bull Lake moraine. It is apparently a horst uplifted before deposition of the moraine. Along the south boundary of this horst, a scarplet with 4 to 6 inches of displacement formed during the earthquake. West of the horst, an area underlain by rhyolite was cut by several shorter new scarplets that are on the same trend but en echelon, stepping to the left; the vertical displacement on these was also about 4 inches. These occurred along earlier small topographic escarpments that are visible on the ground and on prequake aerial photographs. East of the horst, and again stepped to the left, another new scarplet as much as 10 inches high formed along an obvious old scarp where the moraine had previously been displaced 5 feet.

About 500 feet south of the horst, several other new scarplets formed in 1959. They are downthrown on their north sides by as much as a foot and thus define a graben flanking the horst. The easternmost of these new scarplets follows a prominent old scarp on the surface of the moraine that is as much as 15 feet high. (This scarplet, like another longer one 4 miles to the east, was discovered by field investigation of the site of an old scarp, that was recognized on prequake aerial photographs.) The western scarps follow obscure topographic alinements which may or may not be of structural origin.

Several established springs issue from the moraine at points on projections of the lines of new scarplets. One spring is at the southeast end of the exposed-bedrock part of the rhyolite horst, less than 50 feet beyond the last quake fissure. Another lies at about the south margin of the graben, although at a point where the graben cannot be precisely defined.

The close relationship of earthquake scarplets to earlier topographic features, whose forms alone are suggestive of fault origin, indicates that the old features are fault scarps and that the new scarplets also probably represent movement on bedrock faults.

WARPING OF THE GROUND AT MADISON FORK RANCH

Changes of level associated with movement on several of the small faults in surficial material are well demonstrated by effects at the Madison Fork Ranch, which is just west of the South Fork of the Madison River and south of U.S. Highway 191, 4.3 miles north of west from West Yellowstone. Several low scarps here received new surface-breaking offsets of a few inches to a foot, with the north sides downthrown; these faults are indicated on plate 2. Both old and new scarps are short. About 75 yards east of the end of the northernmost of the faults mapped, a large log building, used as a dude-ranch lodge, was built in three tiers of logs end-to-end across the projection of the fault. The north and south tiers are still nearly level, but the center tier was skewed down to the north so that the north tier is now lower than the south. This effect had partly accumulated over many years before the earthquake—showing a surprising speed of deformation unaccompanied by earthquakes—and was partly an effect of the ground shift accompanying the quake. The inside of the lodge was much damaged.

East of the lodge, and still on strike from the fault, the bed of the South Fork of the Madison River was steepened for a distance of about 100 feet by a gentle monocline, and the water now runs more rapidly than it did before the earthquake. The water now piles up against the bank at the base of the monocline and partly spills over, which it did not do before the earthquake.

West of the lodge and just east of the end of the northern fault, a ditch previously carried water northward through a horse corral 100 yards beyond the faults. After the earthquake, the water, instead of flowing through the corral, reached only to the south end of it, and the ditch spilled out into a newly created broad depression in the sagebrush flat midway between fault and corral. Another new pond formed by tilting of the ground just above the upper fault. Springs in the vicinity of the ranch were drastically changed by the earthquake: springs previously large became small, ones previously small became large, and several new ones were formed.

Because of the presence of buildings, river, and ditch at the Madison Fork Ranch, warping of the ground, with an amplitude of about a foot, was quite obvious. Such relatively slight deformation would be unrecognizable in most settings, but similar warping and changes of level presumably accompanied most of the new displacements on the faults of the West Yellowstone basin. Relatively rapid warping without earthquakes has presumably occurred elsewhere in the basin also.

EN ECHELON FISSURES AND SCARPLETS

During the 1959 earthquake, discontinuous fissures and scarplets formed along the upper slopes of preexisting monoclines throughout their length. Vertical offsets on the new scarps are almost invariably down toward the topographic basin. The offsets amount to as much as 1½ feet but are commonly less than 1 foot and do not exceed half a foot along considerable lengths of the monoclines. New displacement is generally related to the height of the monocline: the highest scarps tend to be on the highest monoclines;

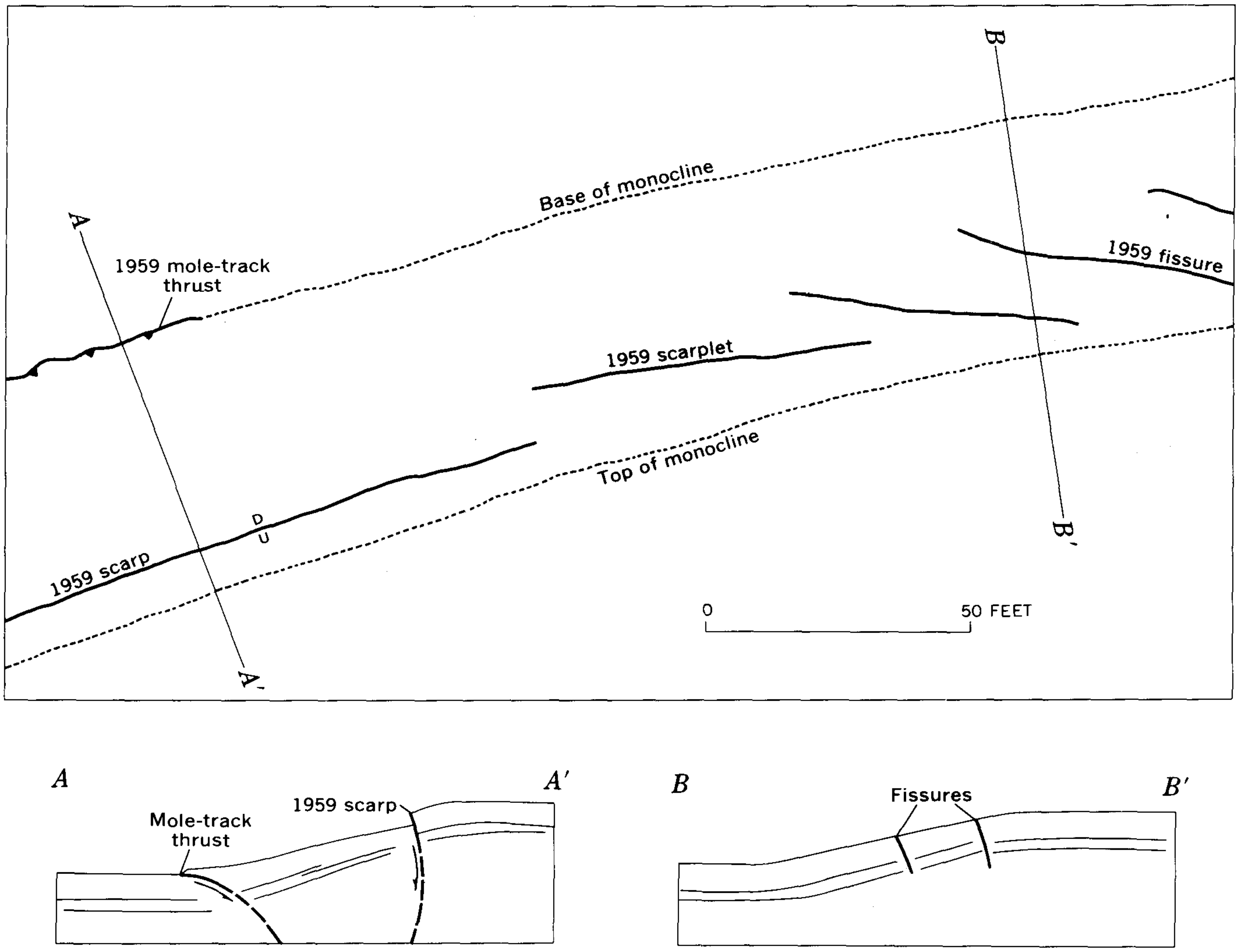


FIGURE 33.—Schematic map and sections showing typical relations between scarps and fissures formed during 1959 earthquake, and earlier monoclines of the sand plain. Heights of structures exaggerated.

and as the rises dwindle, so do the scarps, and both commonly disappear together; but there are conspicuous exceptions. Mole-track thrusts like those along the Madison Arm structure formed at a number of places at the base of monoclines where new scarps were relatively high.

Most of the structures face north. Fissures and faults dip steeply south at the surface, so that the downdropped blocks are in many places overhung slightly by the upraised blocks. Dips tend to be intermediate between vertical and perpendicular to the monoclinal slopes.

The newly developed structures range from markedly oblique fissures a few yards long that show no vertical displacement, through longer and less oblique scarplets of small displacement, to a continuous or nearly continuous low scarp parallel to the monocline (fig. 33). The en echelon structures commonly overlap and are confined to the monoclinal slope.

Almost all the fissures strike west-northwestward from N. 60° to 80° W., but a few swing a few degrees into the southwest quadrant. All en echelon fissures along monoclines diverge in strike in a clockwise direction from the controlling structures: the fissures step to the left. At two places where fissures break level ground undeformed by monoclines, however, the fissures diverge counterclockwise from the controlling structures. One of these exceptions is west of the South Fork of Madison River (fig. 35), and the other is along the west bank of Grayling Creek upstream from the Blarneystone Ranch.

The origin of the fissures is not clear. Their en echelon pattern seems at first glance to be evidence for minor strike-slip fault displacements. The fissures, however, show no preferential strike-slip displacement of opposite walls. Most simply opened directly.

Where one fissure had an inch or two of strike-slip component (as shown, for example, by the offset of sticks embedded in the duff), the next might have the same amount of opposite displacement. More significantly, even the highest of the new scarps showed no more than this same maximum of an inch or two of strike-slip displacement, and that in inconsistent directions. Were there a strike-slip component in the basement faulting, the highest surface scarps should show the most lateral displacement.

A more likely explanation is that the en echelon fissures were produced by the passage of longitudinal-compressional earthquake waves through the unconsolidated sand. All fissures are subparallel to the long axis of the basin of new subsidence, and therefore to the presumed controlling regional structures at depth, and to the earthquake-generating fault as determined by Ryall's (1962) first-motion study. Over most of the basin the passage of waves of alternate compression and rarefaction (perhaps the long-period Rayleigh waves) left no surface trace; but where the obsidian sand was actually being folded during the earthquake, the topmost layers were in tension, and the rarefaction impulses of the passing waves may have opened the surface fissures. A refracting effect of divergent monoclinal trends could have produced the exceptional variations of individual fissures from the general strike.

PREEXISTING STRUCTURES

Structures south of Hebgen Lake that were activated during the earthquake all follow preexisting structures, along at least some of which there has been repeated deformation during late Quaternary time. The nearly universal coincidence of new and old scarps, their occurrence in linear zones consistent with known bedrock faulting peripheral to the basin, and numerous other features described in both the preceding and the following sections demonstrate this. The structures cannot be explained in terms of basinward slumping, for any slumping on the very gentle slope of the sand plain would necessarily have produced nearly horizontal separations. Some local slumping that did occur on low-angle surfaces along the lakeshore is described on later pages.

The prequake faults and monoclines appear as gentle north-facing rises 1 foot to 15 feet high which interrupt the smooth, gentle northward slope of the sand plain. Frontal slopes of the rises commonly range between 5° and 10° and uncommonly reach 15°, rarely 20°. The rises are readily distinguished by their relative straightness, and generally by their gentler slopes, from the irregularly curving fronts of stream terraces. Broad, very shallow sags 100 feet or more wide are developed in front of some of the monoclinal rises. As the margins of the sags are poorly defined, their depths cannot be measured precisely, but they are commonly about a foot. Most of the rises lie in a west-trending arcuate zone, which is gently convex in plan to the north and 6 miles long, that extends from the Madison River to the west edge of the sand plain. The individual structures are separable into two groups that trend about N. 70° W. and N. 75° E., respectively.

A section of one of these old structures in the sand plain is exposed in the east bank of the South Fork of the Madison River, 2.5 miles south of Madison Arm of Hebgen Lake (fig. 34). The ground profile is that of a stream terrace deformed by a combined fault and monocline and slightly modified by erosion. The terrace surface is underlain by about 5 feet of sandy pebble gravel composed of rhyolite reworked from the upper part of the obsidian sand plain, whose original surface lay about 30 feet above the present terrace. Loess covers the terrace, and a weak soil is developed on the loess and the upper part of the gravel. The terrace surface, its soil zones, and the loess and gravel veneers have been warped and faulted into a rise having 7 feet of structural relief. Of this amount, about 15 inches represents offset along a normal fault that dips 75° north, and the remainder is due to monoclinal flexing that produced structural dips in the surficial material as steep as 9° near the fault.

Earlier movement on the same fault had offset the underlying beds of the sand plain. The scarp then formed was planed off by river erosion as the terrace was cut, and the 5-foot thickness of terrace gravel was deposited unconformably above the sand plain sediments. Preterrace movement on the fault was more than 2 feet: the foot of silt exposed between the river and the terrace gravel on the downdropped side of the fault is obviously from a different bed than is any of the sand exposed beneath the terrace gravel on the opposite side of the fault. About 150 yards eastward along the monoclinal rise is the edge of the river terrace, and the rise continues up the river-cut terrace scarp, 20 feet high, onto the main surface of the sand plain. On the main surface the rise is 10 feet high, and this represents the total displacement of the sand plain, or the sum of that accomplished before the formation of the terrace and that after. The relations at the river cut suggest that more than half this amount is postterrace deformation.

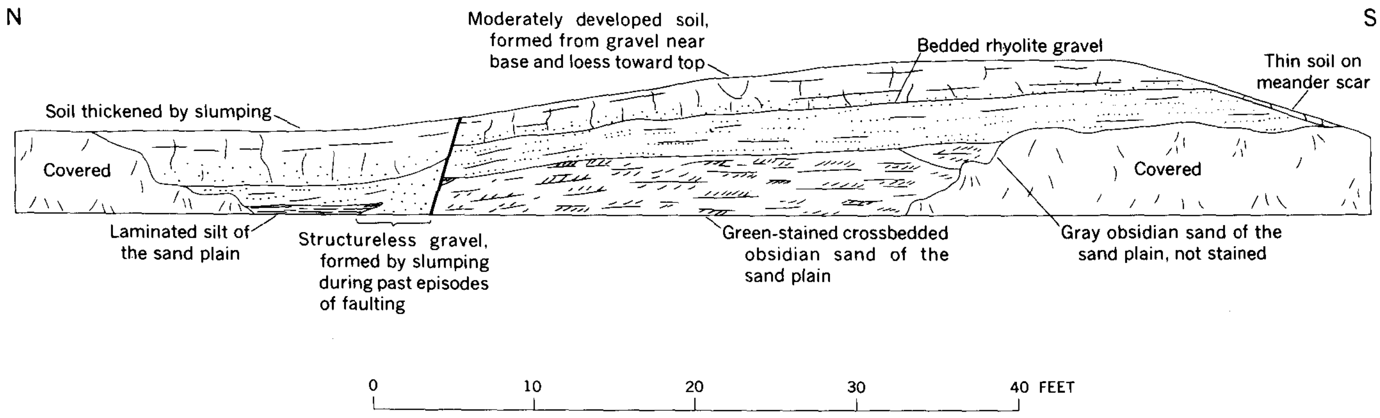


FIGURE 34.—Fault in the sand plain. The interglacial obsidian sand was faulted without warping; offset was more than 2 feet. The scarp was beveled by a river-cut terrace, and a terrace gravel deposited; this was in turn mantled by loess, after which a soil was developed. The fault was then reactivated with another foot of offset, accompanied by the warping of a monocline with a structural relief of an additional 6 feet. This fault was again reactivated along most of its length, though not at this exposure, during the 1959 earthquake. Exposure in bank of South Fork of Madison River, SE¼ sec. 14, T. 13 S., R. 4 E.; drawn from a series of photographs.

Two different structural effects are displayed in the river cut. The deformation of the terrace surface produced a monocline, and this was slightly augmented by faulting. The older preterrace deformation of the sand plain sediments, which occurred when they were buried beneath 30 feet of similar material, produced a steep normal fault without monoclinal warping. This difference in behavior between surface and buried materials is a critical factor in the interpretation of the mechanics of the structures of the sand plain (p. 68). The surface monocline presumably formed above a fault that did not break through to the surface.

No structure, old or new, is visible in the modern swampy flood plain of the river along strike west of the riverbank exposures. Farther west, on the flood-plain deposits that are higher and hence drier, a zone of fissures formed during the earthquake. This zone continues westward and emerges from the alluvium to nearly coincide with a long straight fault scarp in Bull Lake till, overlapped from the north by alluvium (fig. 35). The exposed height of the fault scarp is 18 feet where moraine appears on both sides of it, and this indicates the amount of posttill offset. Above the steep scarp the surface of the till rises more gently for another 30 feet, and this may indicate still older uplift; if so, total structural relief is nearly 50 feet.

This structure of the sand plain has thus undergone at least four episodes of displacement during late Quaternary time. The moraine was offset at least once, and probably twice, before the deposition of the sand plain. The next episode occurred after the sand-plain sediments were deposited, and the next after the sand was cut down by the South Fork to the terrace level. Any of the above episodes might have been composites of multiple separate offsets. Further offset occurred during the 1959 earthquake, as the structure is followed throughout its length by discontinuous new scarplets, small fissures, and mole-track thrusts.

Another old structure of the sand plain deforms an abandoned channel of Cougar Creek 6 miles north of West Yellowstone. The channel is shown on the topographic base map of plates 2 and 5 as the uninterrupted course of an intermittent stream, and its essential continuity is evident on the aerial photographs from which the topographic map was made. At the park boundary, however, a structural rise with a slope of 15° to 20°, apparently a monocline, developed obliquely across this channel and formed a drainage divide with a relief, both topographic and structural, of 10 to 15 feet. A meander loop of the dry channel, cut off on the north side of the rise, is now a flat-floored closed depression 200 feet wide which occasionally contains a shallow pond. Southeast of this, the structure again cuts and dams the channel and has produced another small undrained depression. Farther southeast the topographic rise is as steep as 30° and probably is a slumped fault scarp rather than a monocline. The structure was reactivated with the usual suite of minor structures during the 1959 earthquake.

The sides of the abandoned channel of Cougar Creek are steeper than the slope of the younger structural rise which deforms the channel. The original slope of the rise therefore must have been less steep than that of the cutbank; so the rise could not be a slumped fault scarp, but must be wholly or chiefly a monocline. The height of the rise and of the channel sides is about the same, so that the shorter life-

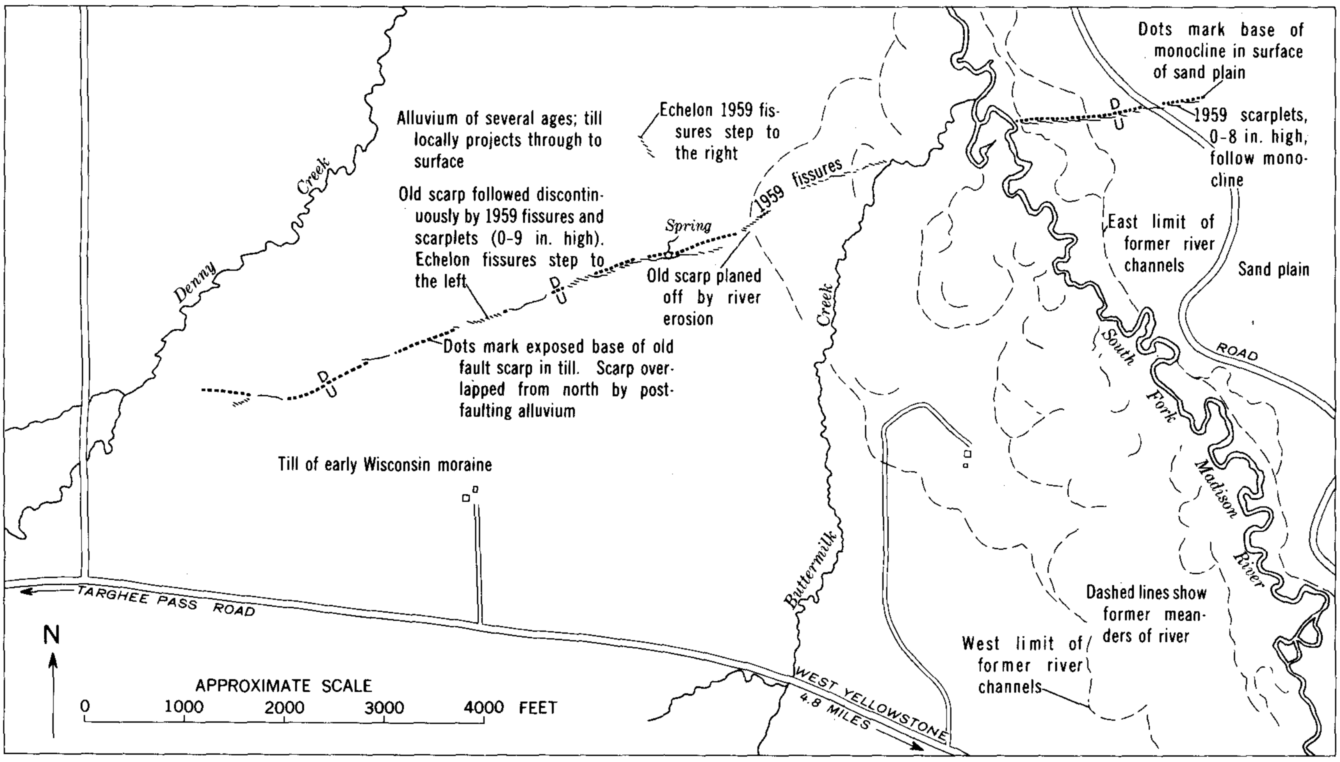


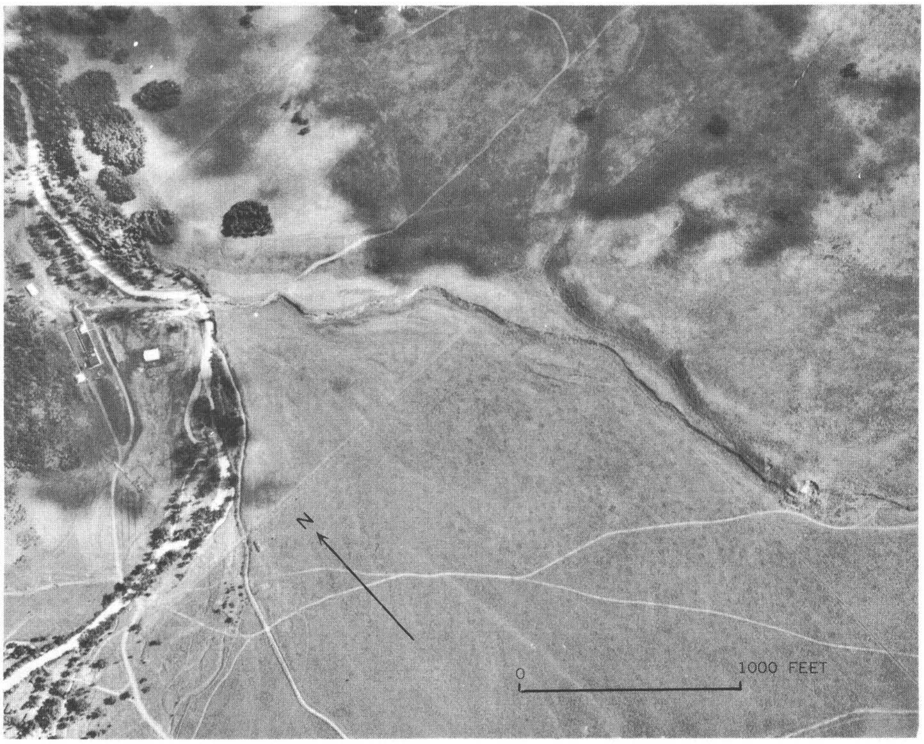
FIGURE 35.—Sketch map of southwest corner of West Yellowstone basin. Earthquake fissures follow earlier fault and monocline in unconsolidated Quaternary deposits. Structural relief of old fault—displacement of Bull Lake moraine—is at least 18 feet between Denny and Buttermilk Creeks. Maximum relief of monocline in sand plain east of South Fork is 10 feet; structure dies out east of road. No new fissures formed at surface in saturated modern flood plain. Figure 34 shows monocline and fault exposed in riverbank. Eastward migration of river is due to tilting of surficial deposits. Map compiled from aerial photographs without photogrammetric corrections.

span of a lower scarp is not a factor. The steepest part of the structural rise—southeast of the channel—is as steep as the cutbank, however, and may be a slumped fault scarp.

The sand plain is crossed by a number of other rises similar to those described above, although most slope more gently than does the one at Cougar Creek. All the rises are followed by the same suites of structures formed during the earthquake. River-cut terrace banks show none of the new structures and are quite different. The rises are commonly distinctive on aerial photographs. The rise now marked by the longest new scarp was first recognized on prequake aerial photographs—the new structure was found by photographic recognition of the older feature. All the rises were walked out for their entire length, and their depiction on plates 2 and 5 is based on continuous field observation. By analogy with the South Fork and the Cougar Creek structures, these other rises are inferred to be also dominantly monoclines rather than slumped fault scarps.

Most of these rises in the sand plain can be dated only as younger than the sand plain; but it is an obvious inference that, like the South Fork structure, they represent repeated deformation. The occurrence of major earthquakes within very late Quaternary and Recent time is shown by the Recent but prehistoric 20- to 40-foot scarp, which may have formed in a single earthquake, along the front of the Madison Range and the important old fault scarps that cut the moraines about the West Yellowstone basin. Probably the sand-plain structures received offsets during some of these earthquakes; and probably, like the structure at Madison Fork Ranch, they also grew by slow deformation without earthquakes.

The new south-facing fault scarp in flat-lying unconsolidated material east of Grayling Creek near the Blarneystone Ranch is on a fault that has had two previous episodes of late Quaternary displacement. It precisely follows an older scarp of about the same height—10 feet (figs. 36, 37). An older and much higher—60 feet—scarp has been eroded back by Grayling Creek; possibly several episodes of faulting contributed to this height. Southeastward along strike the old fault scarps dwindle and give way to a low monocline that is apparently identical with those of the sand plain south of Hebgen Lake. As the monocline becomes lower and more gentle, the 1959 scarp-



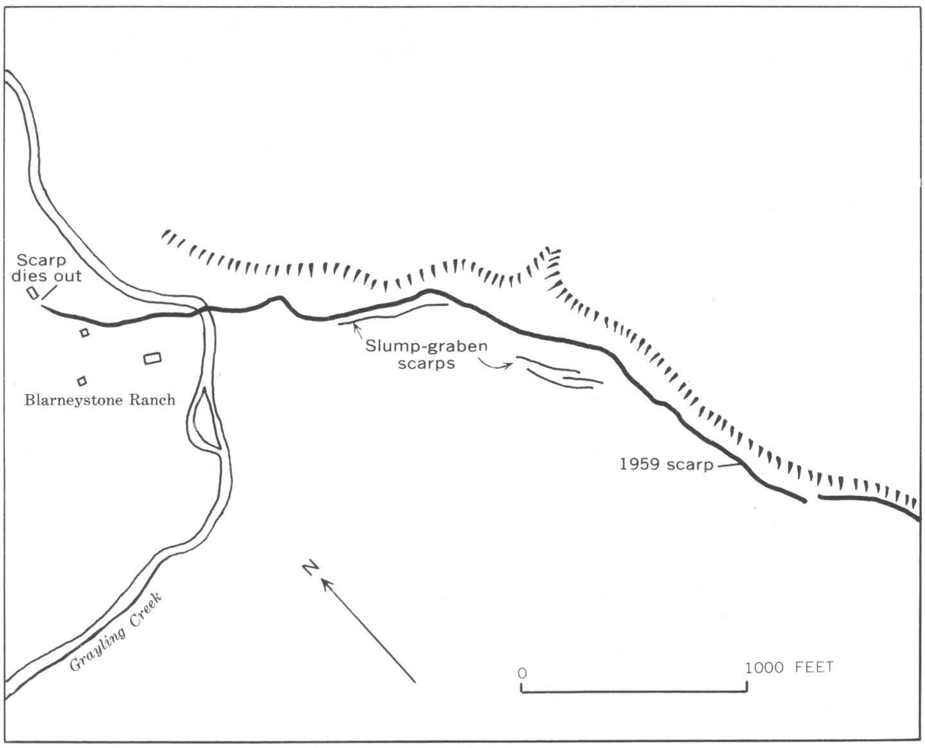


Figure 36.—Reactivated fault scarp at Grayling Creek. The 1959 scarp, about 10 feet high, is precisely superimposed upon an older fault scarp of similar height that shows best just southeast of Grayling Creek. (See also fig. 37.) A still older and greater displacement produced a moat within which Grayling Creek flowed for a time; the stream cut back the scarp then formed to produce the 60-foot erosional escarpment that parallels the younger structural scarps. Dark areas are cloud shadows. Vertical aerial photograph by U.S. Forest Service.

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FIGURE 37.—Reactivated fault scarp at Grayling Creek. The new fault scarp is superimposed upon an older one of similar height. Behind this composite scarp is an escarpment eroded back by Grayling Creek from a larger older fault scarp on the same trace. View looking east across Blarneystone Ranch; see also figure 36. Photograph by John R. Stacy.

lets superimposed on it lessen correspondingly, and the last of the new fissures and the old monocline die out together near the park boundary.

Features associated with the younger of the two prequake scarps are clearly shown on pre-1959 aerial photographs and are strikingly similar to those associated with the 1959 scarp. A slump graben, or moat, marked by low inward-facing scarps in front of the main scarp, is obvious in each case. Along the segment represented in the central third of figure 36 the prequake moat is an undrained depression. On the north side of Grayling Creek—and on a lower and younger stream-graded surface—the extension of the earlier fault scarp is clearly marked. Although the scarp was artificially accentuated by roadway fill behind the ranch buildings before the earthquake, the base of the natural feature there can be traced on figure 37 as an alinement of aspen trees surmounted by a few feet of sagebrush-covered slope, above which the quake scarp rises.

Grayling Creek is deflected southeastward across its modern flood plain, parallel to and just above the prequake fault scarp (see figures 36 and 37), where it apparently flows along a slump moat of a branch of the fault. Tiny en echelon fissures, formed during the earthquake, were traced upstream for more than a quarter of a mile in the surficial deposits on the northwest side of the creek and may mark very slight quake shifting on this branch structure.

MECHANICS OF STRUCTURES

Where bedrock faults had large new displacements during the earthquake, they broke through to the surface and formed continuous new scarps like the one east of Graying Creek. The large new scarps, all of which are in surficial material, are unusually steep, dipping about 70° to 80°. The small new scarps, also in surficial material, are steeper still or even dip in the "wrong" direction: they have dips between vertical and perpendicular to the monoclinal slopes upon which they occur. These smaller structures require a complex explanation to account for such features as the anomalous dip and the mole-track thrusts—an explanation in terms of the response of weak unconsolidated material to gravitational stresses that result from movement on bedrock faults.

Because of its lack of strength, the unconsolidated material of the sand plain cannot maintain a surface so oblique to the applied force—gravity—as that formed in the bedrock beneath, and the weak material should fail along a surface more nearly parallel to the vertical shearing stress. As the break in the surficial sediments is steeper than the fault in the bedrock, the surficial deposits tend to pull away across the developing break, but a variety of slump effects prevents any wide gap from forming (fig. 38).

The diverse structures formed during small displacements along the faults and monoclines of the sand plain are due to the combined effects of refraction of faults and of slump. Dips of the major faults in the surficial material are near 75° (fig. 34), and the magnitude of the slump effects indicates that the dips of the same structures in the underlying bedrock are about 10° less, or near 65°.

We visualize that where the movement along a bedrock fault is small, with slip beginning at depth and progressing upward during an appreciable interval of time, the resultant slumping toward the fault of the loose material of the downdropped block causes a sweeping deflection of the developing fault past the vertical to 45° and, right at the surface, almost to horizontal (fig. 39). The "overhanging" part of the footwall is lowered smoothly, and the product at the surface is a monocline with a mole-track thrust along its base. The little thrust is an underthrust produced by slumping in response to tension. New scarps may or may not be produced at the upper side of the monocline, as scarping and flexing alternate at the surface along the trend of the structure. Mole-track thrusts converge in plan with the scarps at places, so the structures must converge at depth also in some fashion like the one shown.

Moats and slump grabens like those along the high new scarp east of Grayling Creek are due to slump effects along faults refracted to steeper dips in thin sections of unconsolidated material. The thinner the surficial fill, the narrower and better defined is the graben.

Where the fill is thick, a broad shallow sag is produced, instead of a narrow moat, in front of the scarp or monocline as the surface expression of the slump of the subsiding block. Structural relief, as measured across the surface monocline, is greater than the vertical displacement of the basement beneath by an amount about equal to the depth of the slump sag.

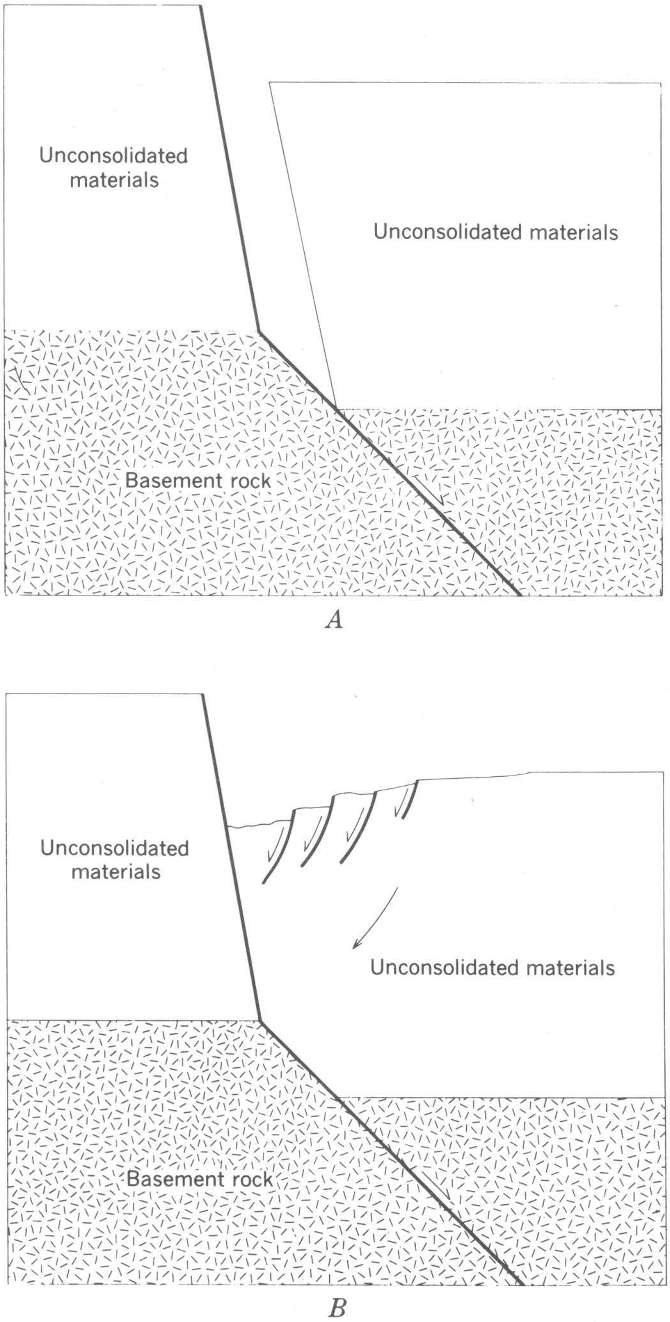

FIGURE 38.—Effect of fault refraction upon unconsolidated material. Fault is steeper (change in dip exaggerated) in unconsolidated material than in bedrock. *A*, Gap in surficial deposits produced by displacement on fault; *B*, fault-produced gap in surficial deposits closed by slump effects.

The width of the sag is presumably related to the depth of basin fill. Depressions wider than 100 feet were not recognized in the field, but a slight sag about 700 feet wide is suggested by the pattern of the 6,600-foot contour where it crosses in front of the long northern scarp of the sand plain 2 miles south of Madison Arm. If this is indeed such a depression,

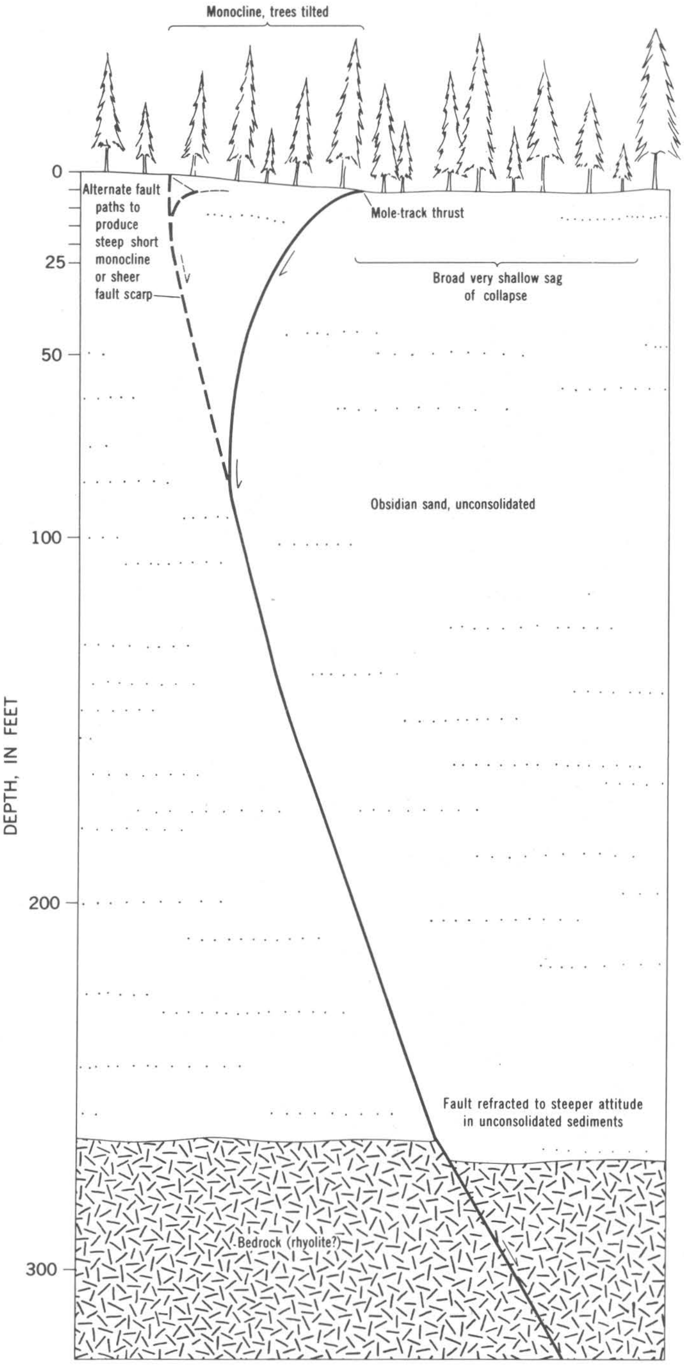

FIGURE 39.—Mechanism of formation of monocline and mole-track thrust south of Madison Arm. See text for explanation.

the unconsolidated fill beneath it might be about 1,200 feet deep.

Most of the new scarps in the sand plain lie parallel to the contours of subsidence along the south edge of the subsided terrane, where warping during the earthquake produced a surface convex upwards. The faults may be shallow products of tension caused by this warping and may have formed as their hanging-wall blocks slumped northward toward the center of the basin.

SLIGHTLY REACTIVATED STRUCTURES

OLD CONICAL DEPRESSIONS

Conical depressions in the sand plain occur in two areas in the eastern part of the main zone of monoclines, north-northwest of West Yellowstone (pl. 2). Several of these structures were reactivated during the earthquake, and their mechanism of formation was revealed. One line of five depressions begins 500 feet east of the east end of the 2-mile-long structure in secs. 17 and 18, T. 13 S., R. 5 E. and continues for 800 feet along strike. The depressions are circular, ranging in diameter from 6 to 80 feet, and in depth from 2 to 12 feet; inward slopes vary from 17° to 33°, the steepest cone being also the smallest. In three of these five depressions, central cylindrical plugs 4 to 7 feet across subsided 1 foot to 2 feet during the earthquake. Two of these plugs dropped by a combination of downwarping and faulting, and the third dropped within a vertical ring fault with 18 inches of displacement.

Two more depressions occur near the west end of the monocline three-quarters of a mile southeast of those described above. One is a conspicuous circular, conical depression (fig. 40) 130 feet in diameter that has a closure of nearly 10 feet and a central plug 18 feet in diameter that subsided a few inches during the earthquake. The second depression is smaller, much shallower, and was not reactivated.

These conical depressions are collapse features that pock an undissected sand plain which slopes only very gently northward. They are interpreted as the result of collapse into the pits of giant sand boils formed during past earthquakes and are similar to those produced during the 1959 earthquake and described by F. A. Swenson (chapter N). The cylindrical central plugs that subsided during the 1959 earthquake probably represent the compaction of the sand in the throats of these short-lived sand-clogged artesian springs. Although no deposits from these postulated sand boils have been recognized, such deposits would be recognizable only by topographic form, as they would consist simply of detritus reworked from the sand plain. The large sand boils developed during the 1959 earthquake produced only thin veneers of sand and gravel despite the considerable size of the pits from which this material came; such veneers, formed of material identical with that beneath them, would not long be recognizable.

Of the two groups of conical depressions, the northern ones are the steeper sided and have the more con-

FIGURE 40.—Old conical depression in the sand plain. Depression is 130 feet across, nearly 10 feet deep, and probably was formed as a sand boil. Two miles north-northwest of West Yellowstone. Photograph by John R. Stacy.

spicuously downdropped central plugs. The southern group may be older, be degraded by slow slopewash, and have cores that were better compacted before 1959 than those of the northern group.

Both groups of conical depressions occur at the ends of monoclines, at or beyond the visible limits of the structures as surface features. Such depressions were seen nowhere else, so a genetic relation to the monoclines is indicated. At the ends of the underlying bedrock faults, relatively intense warping may occur which affects the entire thickness of surficial material (in contrast to the superficial warping that occurs in the monocline), and this deep warping may have squeezed out saturated sands, forming the sand boils.

Similar features, locally called sinkholes, were formed during the New Madrid earthquake and have been described by Lyell and others (Fuller, 1912, p. 87–88). Eyewitnesses observed fountains of mud and water thrown up from some of the "sinkholes." At least some had the original form of an inverted cone, or crater, surrounded by a ring of sand and lignite fragments. Many similar collapse cones formed during Alaskan earthquakes in 1958, and ejected sand was deposited thinly over broad surrounding areas (Davis, 1960).

RING-FAULT DEPRESSIONS

Seventeen small depressions, defined by ring faults formed during the earthquake, are grouped in a belt less than 200 feet wide that stretches for 1,150 feet east-northeastward across the quake-emerged flat northeast of Edwards Island (pl. 2). Each depression is 6 to 25 feet across, a few inches deep, and occurs in modern lake silt that here veneers the sand plain. These features differ from the prequake conical depressions south of the lake in that they are simply depressed plugs unaccompanied by surrounding cones of collapse. The depressions are probably the quake-compacted throats of sand boils that formed along a line of deformation of some past tectonic episode. Although no new fissuring developed along this zone in 1959, two new sand boils were formed within it—the only ones for several miles in any direction—and these suggest that a slight shift did indeed occur. The trend of this zone parallels that of two earthquake scarplets along the south side of Horse Butte, 2 miles to the east.

EFFECTS OF EARTHQUAKE ON HEBGEN LAKE

The deformation that accompanied the 1959 earthquake was particularly pronounced in and around the basin of Hebgen Lake. The basin subsided differentially and the lake bed was thus abruptly warped. This caused extensive slumping of the banks and threw the water into violent surges that alternately drained and flooded parts of the shore and sent large waves over other parts.

SURGES

During the earthquake, the main body of Hebgen Lake subsided more than either the Madison Arm or the dam, and a southeast-trending trough was formed. The water surface was warped so that, just after the earthquake, water in Madison Arm stood 5 feet higher than the dam and 10 to 15 feet higher than water in the main part of the lake. Water from Madison Arm accordingly rushed northward, and its momentum carried it on as a great surge that overtopped the dam. The dam subsided only half as much as the deepest part of the subsidence trough, and therefore the new static water level against the dam, instead of being above the dam as it would have been had the dam subsided as much as the main lake bed, was actually three-fourths of a foot lower than the prequake level.

The first surge of water rose 10 feet above the postquake static water level at the dam. As the crest fell away from the dam and withdrew southward, the momentum of the water again carried it beyond the equilibrium position. Like the sloshing of water in a bathtub, the lake water continued to oscillate back and forth with gradually decreasing intensity for at least 12 hours after the earthquake. Because of the large amplitude of the waves and their general violence, they are here called surges to distinguish them from the much milder but otherwise similar oscillations of closed water bodies called seiches. (For similar usage, see Prins, 1958.)

OVERTOPPING OF THE DAM

Water poured over Hebgen Dam three or four times, as reported by the dam foreman, Mr. Hungerford (chapter A, L). It flowed first through the spillway outlet, and continued to rise until it poured across the entire width of the dam. The concrete spillway was badly damaged, partly by this flood but largely by the preceding earthquake (fig. 81, chapter K). The downstream face of the dam, a fill of densely compacted, calcareous stony alluvium considerably cemented by lime, was carved by gullies 3 to 4 feet deep (fig. 79). The crest of the dam showed little evidence of the overflow, which suggests that the water welled over slowly. The concrete core wall was covered during construction under about half a foot of rubbly fill that was partly cemented by lime; this cover was eroded away by the overflow in some places, but in others only the fines were stripped off its upper surface. Streamers of grass roots, some with soil clods still attached, were locally overturned and alined downstream, proving that the settling of the fill which exposed the core wall took place before the last of the overtopping surges. Evidence of water flow was found a maximum of 1½ feet above the top of the core wall, about 1 foot above the original top of the fill on the core wall, and the preservation down to this limit of delicate soil structures and accumulations of fragile plant litter showed clearly that the water could not have gone higher. Hungerford's estimate that "3 or 4 feet" of water poured over the dam during the second surge is certainly too high.

SURGES ALONG THE LAKESHORE

By studying features along the lakeshore it was possible to determine the height of the highest surge at a number of places. It is a reasonable assumption that the highest surge at any one place represents the first surge. Along the relatively submerged shoreline, particularly along the northeast or lee shore of the northern arm, the surges left swash marks of abundant debris lodged in plants and dumped along the ground. From this sort of evidence it is plain that the surges reached their highest level near the dam.

Dramatic evidence for the highest surge was found at Hilgard Lodge, on the north shore of Hebgen Lake 1 mile from the dam. Subsidence during the earthquake brought the new static lake level up to the doorsteps of two cabins and put the base of a Douglas-fir tree, for example, under 6 feet of water (fig. 9). The surge then rose 9 feet above the new lake level and left a clear high-water line of debris in the sagebrush behind the cabins. As the two cabins floated off their foundations, they tipped, and the high-water marks on the inside walls therefore sloped conspicuously. One cabin was set back on its foundation only a few inches out of place and was little damaged, whereas the other was dumped a yard out of place and was badly skewed.

Five-eighths of a mile southeast from Hilgard Lodge, the surge high-water line was only 6 feet above the postquake lake level; and three-fourths of a mile farther southeast, at the south limit of the Kirkwood Creek fan, the high-water line was only 3½ feet above the postquake lake level. The height of the highest surge thus decreased away from the dam.

Such evidence of the main surge was for the most part limited to the lower 7 miles of the north shore. The southern part of the lake subsided less than the rest and hence became emergent; its shoreline is entirely in porous surficial deposits, largely those of the obsidian sand plain (pl. 5). Although the abundant water weeds growing in the lake must have been oriented by the surges, the rapid lowering of the water level resulted in such an outflow of water from the lake banks that the weed orientations, studied after the earthquake, recorded only the pattern of this outflow.

Clear evidence of strong longshore currents was found at only two of many points examined. On a small low hill of glacial till (Edwards Island in the full prequake lake, but connected to Horse Butte peninsula by a dry sand flat in the postquake lake), long filaments of water weed recorded a strong northwestward current that extended half a foot above the postquake static water level. At one place on the south shore of Madison Arm an orientation of water weed produced by an eastward-directed current (a returning surge) was preserved after the streaming of the water stored in the porous banks.

A witness to the behavior of the surges, Col. H. B. Crowell, was in his cabin at the end of the road along the southwest shore 2 miles from Hebgen Dam. "About 7 to 10 minutes after the quake" he heard the sound of rushing water and found that the lake had risen about 20 feet vertically above its prequake position. About half of this amount represents submergence of the shoreline beneath the postquake static water level, and about half was the additional height of the surge. Within "a minute or two" the water was rushing out again. Boats and docks were washed away and were subsequently found on the northeast shore of the lake, directly north of their original position. Later that night and early the next morning the water level changed gradually by much smaller amounts as the surges became more regular. Crowell also quoted reports by friends at Watkins Creek Ranch who said that shortly after the earthquake, several large waves rolled up on the alluvial fan there.

Another report of waves on the lake was obtained by engineer Glenn Jones of the Montana Power Co. from a couple who were standing by the road near Kirkwood Creek at the time of the earthquake. The man was knocked down by the quake, and when he got up he noted a wave "10 feet high" coming northeast toward him. This was closely followed by a wave "15 feet high" moving directly down the lake (northwest); and this in turn was followed by a wave coming northward, obliquely down the lake, toward him. No times were noted, but the total interval was short.

Ervin S. Armstrong, owner of the ranch at Corey Spring, reported that 15 to 17 minutes after the earthquake, the water below the ranch, where the shoreline is on a gently sloping fan, was about 100 yards out from its prequake position. Then during the night—he did not witness the event—the water rose well above its prequake level. By morning it had essentially stabilized at its new level, several feet above the old.

George Gunnell reported that he ran out of his cabin on the south shore of Madison Arm during the earthquake and saw in the bright moonlight that the lake was covered by vertically bouncing waves. These broke the rope by which his boat was tied to the dock. As the earthquake ended, the water began to move westward, down the lake, and rapidly withdrew from the shore. Like others who witnessed this phenomenon, Gunnell assumed that Hebgen Dam had burst.

Another boat was broken loose by the earthquake from a dock a mile east of the inlet of the South Fork of the Madison River and was found the following day in the sagebrush near Hebgen Lake Lodge, 6 miles away.

PERIOD OF THE SURGES

Different approaches to the problem of determining the interval between successive surges yield contradictory results. No firm, unambiguous data are available.

The dam foreman, Hungerford, and his assistant witnessed two or three surges and know that water flowed over the dam once before that. They did not presume any clear estimation of the times involved. People who had been camped in the Madison River canyon above the Madison Slide left their camps after the earthquake and apparently began to arrive at the dam about the time of the last overtopping, although Hungerford's statement only infers this. From his account of the things he did before the arrival of the campers turned him to other tasks, and from the presumption that the first of the campers did not linger in the canyon before driving up to the dam, it seems that the total sequence of overflowings took about half an hour.

Overtopping of the dam was recorded by a continuously operating stage recorder at the river-gaging station one-fourth mile below the dam. The river height here varied not only as the water overflowed the dam, but also as the water depth, and hence pressure and outflow, varied at the intake tower in the lake. The time scale on the graph of the river-depth recorder is only 0.1 inch = 1 hour, and the instrument is designed so that fast changes in level are damped out and not recorded immediately. The earthquake knocked the time scale out of position, and the first surge made the pen slip to a new depth position. These various factors combine to make the graph ambiguous during the first several hours after the earthquake. The main earthquake occurred at 11:37 p.m. local time. The record before 2 or 3 a.m. is ambiguous, but for 6 hours or so after that the recorder clearly indicated alternate high and low water levels, with a period (high to high) near 15 minutes. Howard Matthai (written communication, 1959) of the Geological Survey interprets the graph for the few hours immediately after the earthquake, including the period when the surges flowed over the dam, to indicate that they arrived at the dam about 17 minutes apart.

The period of uninodal oscillation in a rectangular tank can be calculated from the simple Merian formula

$$t = \frac{2l}{\sqrt{gh}}$$

where t is the period, l the length of the basin, g the acceleration of gravity, and h the water depth (Ruttner, 1953, p. 43). Using 15 miles for the length, the period indicated by this formula is about an hour, varying with the depth assumed. As the ratio of length to depth is large, this variation is not great. Hebgen Lake is of course not a rectangular tank. Jeffreys (as quoted by Hutchinson, 1957, p. 307–8) emphasizes that long narrow lakes, especially those with cuspate ends, have periods markedly shorter than indicated by the simple Merian formula.

This calculated period is much longer than that indicated by either Hungerford's account or the river gage. It seems that either the surge was multinodal or that the great height of the initial surge caused it to move far faster than the ideal formula would suggest. The lake bed was so warped that immediately after the earthquake the water in Madison Arm stood nearly 15 feet above the water in the main part of the lake only 5 miles to the north. The wave resulting from this warping apparently reached the dam about 10 minutes after the earthquake. By several hours later, a complex or multinodal surge with an effective short period was apparently active.

PROFILE OF THE HIGHEST SURGE

The approximate profile of the crest of the highest surge in the lower part of the lake is plotted on figure 41. The warped shoreline, and hence the deformed position of the water surface assuming that no movement of water occurred during the earthquake, is also shown.

Maximum surge height increased rapidly toward the dam, largely due to the funneling effect of the convergent steep valley sides, but the highest point was several hundred yards from the dam. This is only in small part due to draining off of the crest by flow over the dam. Much of this effect was due to refraction of the surge toward the valley sides as the surge was retarded by these sloping shores.

The form of the surge was apparently intermediate between a wave sloshing back and forth along the lake and a seesaw rocking of the water. Modern studies of water waves suggest that there is a continuous variation in waveform and behavior between the seesaw movement of a low uninodal seiche in deep water and turbulent surges where the height is large relative to the water depth (Prins, 1958). In both types, velocity is proportional to the square root of the depth.

BANK SLUMP

The banks of Hebgen Lake slumped extensively during and shortly after the earthquake. Along the steep shores of the lake near Hebgen Dam, large landslides went into the lake, greatly changing its bottom topography. These slides and their effects are described by J. B. Hadley (chapter K) and by W. H. Jackson (chapter H).

In the eastern part of the north shore of Madison Arm, a high steep bank of unconsolidated sand faces bays formed by the flooding of deep-cut meanders in the sand plain. During the 40-year life of the reservoir, a narrow beach was formed at the high-water level; lakeward, the flooded meander flat was covered by a mat of waterlogged timber. During or shortly after the earthquake, this beach slipped away from the bank and left behind shallow moats as wide as 50 feet. The timber mats were jostled to chaotic jumbles of logs at all angles (fig. 42). The slip surfaces are essentially the downslope extensions of the modified meander scars. At several places the slump escarpment is continuous throughout the full extent of a concave meander bay. These features are thus

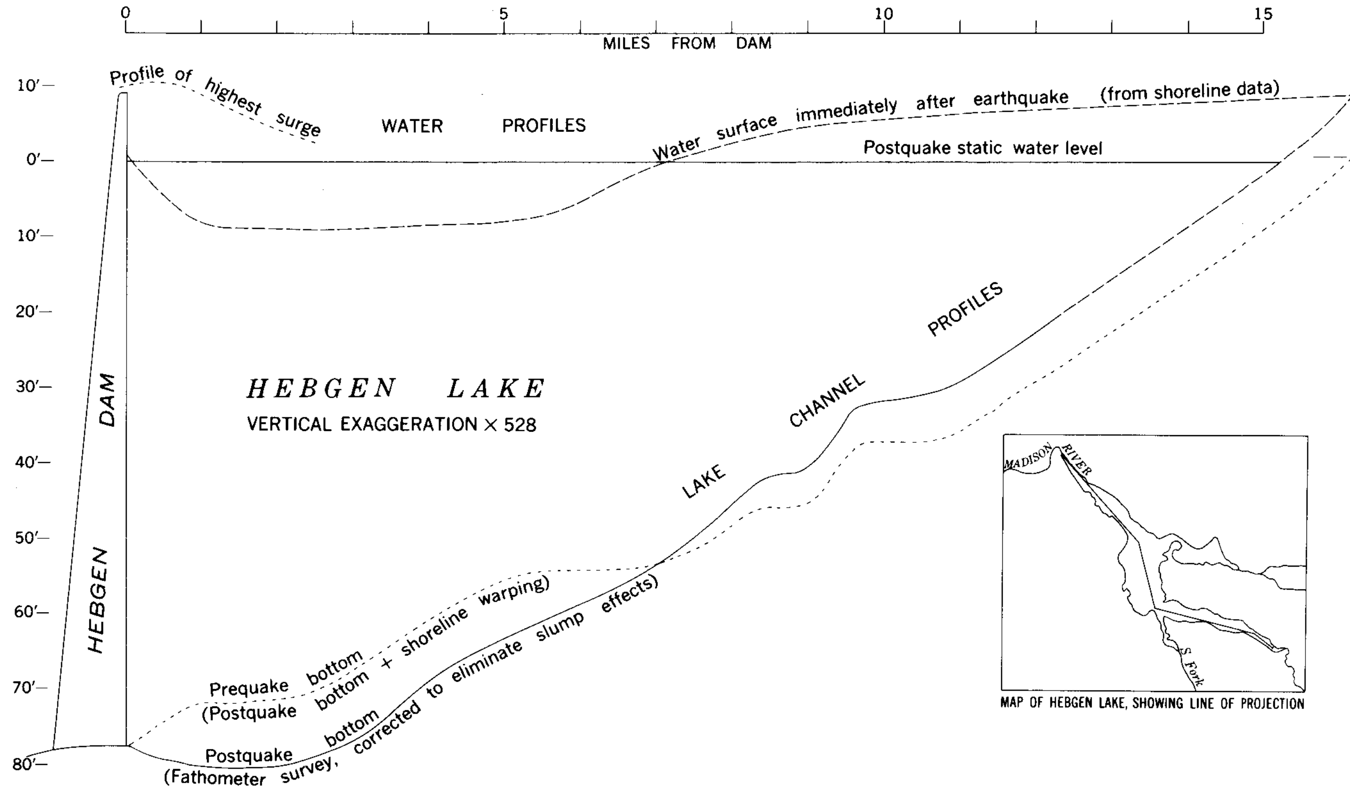

FIGURE 41.—Longitudinal projection of Hebgen Lake, showing deformation due to earthquake.

crescentic; at the intervening headlands the slumps disappear under water.

On the south shore of Madison Arm, a floating dock and its lakeward anchor posts were shifted basinward as the shoreline slumped. The supports were carried out 31 feet yet lost only 2 feet in altitude; this indicates sliding of sediment on a surface sloping only 3.5°. A witness reported that most of the movement took place within a few minutes after the earthquake but that slow creep continued for at least 18 hours. The separate supports, although not connected by any structure, retained their relative positions throughout the sliding.

The above types of slump resulted from the rapid outward flow of ground water from the banks as the lake water retreated. Large quantities of water stored

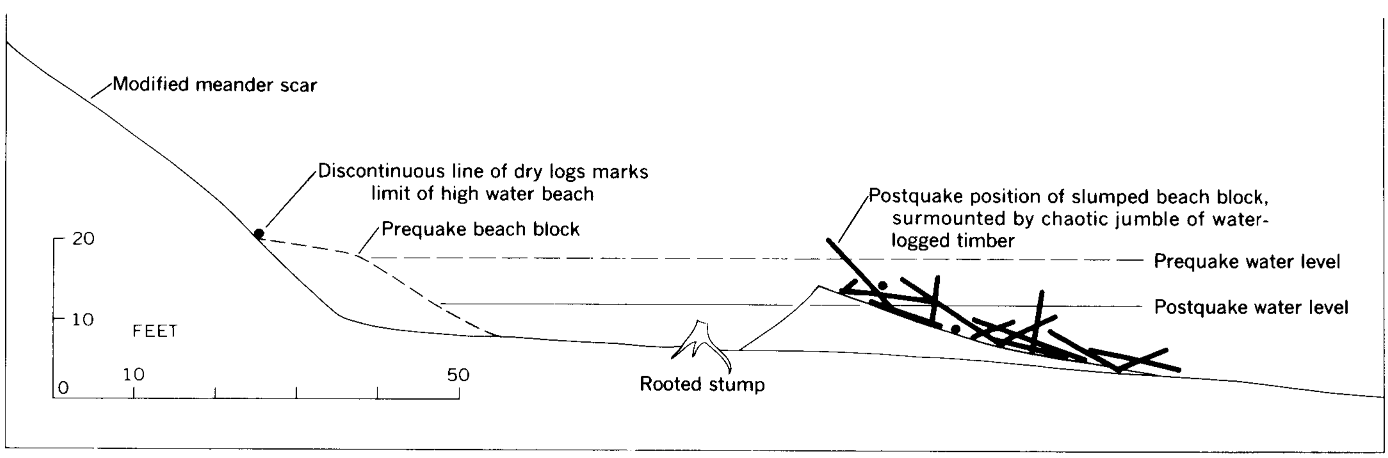

FIGURE 42.—Profile of slumped beach on north side of Madison Arm of Hebgen Lake. Emerged high-water beach slid lakeward, leaving moat behind.

in the highly permeable and porous banks doubtless moved out under differential pressure, thus reducing the cohesion and resistance to flow of the bank materials. The normal slow annual drawdown of the lake has little such slump-causing effect.

Gravity sliding, different from the slumps noted above, affected the lake bank in the west part of Grayling Arm, just east of the Narrows. A sheet nearly a mile wide along the lakeshore pulled away from the hillside and moved outward more than 6 feet while dropping only 1 foot. The east shoreline of the crescentic tip of Horse Butte peninsula is rimmed by almost continuous slump scarps that follow the deeply curved shoreline for more than a mile. Pull-apart scarps and open fissures are steep. The posts of a barbed-wire fence extending directly into the lake across the slide zone still stood vertical landward of the zone but within the slide were rotated, tops shoreward. Most posts were pulled out by the wire, but one post that remained partially imbedded moved at least 6 feet horizontally. The slope of the lake floor here is less than 1°, and it is probable that the water-saturated sediments actually slid out into the lake on glide surfaces of similar gentle slope. Rapid drawdown played no part: this slumped area was slightly submerged as a result of the earthquake deformation and is located where the surge was small.

The Grayling Arm slide may have resulted from a transient excess pore pressure caused by local compaction of surficial sediment during the earthquake. Downdropping of surficial material was followed by abrupt deceleration, and the momentum briefly applied a load throughout the material. As emphasized by Rubey and Hubbert (1959, p. 183), instantaneous loading produces the greatest excess fluid pressure (and hence the tendency toward flotation) at the top of a sedimentary column in compaction equilibrium. If the Grayling Arm gravity slide was rendered effectively frictionless by pore-water pressure, it is of the type illustrated by Rubey and Hubbert (1959, fig. 9).

At the Narrows camp and eastward along the north shore from it, effects of slump and of structurally controlled surface cracks are mingled; features which in detail are clearly related to the hydrographic basin are in a more general way apparently related to structural deformation. A continuous single scarp 1 foot high and facing the lake thus parallels the gently curving shoreline at a distance of only 50 to 100 feet for a mile eastward from the Narrows. This scarp is part of a broad zone of scattered surface breaks that seems from its general pattern to be structurally controlled (pl. 2).

ABSOLUTE SUBSIDENCE

The large fault scarps and many of the other earthquake features around Hebgen Lake are spectacular but localized effects of the wholesale subsidence of a large area. Subsidence has been proved over much of an area nearly 15 miles north-south and almost twice as long east-west. About 60 square miles subsided more than 10 feet, and about 200 square miles subsided more than 1 foot. Only about half a foot of absolute elevation has been proved.

The maximum proved subsidence—22 feet—occurred in the West Yellowstone basin. Part of the Madison Valley near the latitude of Hebgen Lake also subsided, though not as much; and subsidence has been proved through much of the Madison River canyon, which cuts across the intervening Madison Range.

MEASUREMENT OF SUBSIDENCE

The northwest side of the West Yellowstone basin, where subsidence was greatest, is traversed by a precise level line established by the Coast and Geodetic Survey in 1934. Releveling of this line after the earthquake revealed the amount of absolute subsidence in the West Yellowstone basin and in the Madison Valley. We constructed a continuous postquake road profile from the individual level-rod readings and compared this with the prequake profile to measure the absolute subsidence along the parts of the Madison River canyon not buried by Madison Slide and Earthquake Lake. Displacement of the prequake shoreline of Hebgen Lake relative to the new high water line, the absolute altitude of which was determined during the leveling, made it possible to determine absolute subsidence all around the lake. Measurements of displacement along the new scarps and a fathometer survey of the lake floor provided further information on the changes of level that accompanied the earthquake. These data are plotted on plate 2.

RELEVELING

In 1934 the Coast and Geodetic Survey established a line of second-order levels between West Yellowstone and Sappington, Mont., following the highway along the northeast side of Hegben Lake, through the Madison River canyon, and down the Madison Valley. All previous bench marks along the route were tied into the level line at that time. In September and October 1959 this level line was rerun by the Coast and Geodetic Survey, this time with first-order accuracy. Along a 32-mile part of the route, 47 bench marks, all but one of those recovered, subsided more than 0.5 foot. Changes in level of bench marks between the 1934 and 1959 surveys are indicated on plate 2.

It is quite possible that a small part of the subsidence between the 1934 and 1959 surveys took place before the earthquake, but no convincing arguments can yet be made. The data are plotted on plate 2 with the assumption that the total measured subsidence occurred during the earthquake.

Many of the releveled bench marks are in pairs, 132 feet apart, on opposite sides of the highway. Northward from West Yellowstone toward Duck Creek the subsidence indicated by the bench marks increases gradually. Where the total subsidence is small, subsidence of the paired stations tends to be uniform; but where the total subsidence is greater, a general side tilt toward the lake is recorded. Of 10 pairs, 6 were tilted lakeward, 1 was tilted away from the lake, and 3 were not tilted. The three northern pairs, where subsidence is the greatest for this leg of the line, show a side tilt that increases northward from 5.2 to 10.1 feet per mile.

The figures shown on plate 2 for the absolute subsidence of bench marks are the differences between the adjusted 1934 altitudes and the unadjusted 1959 field altitudes, to which however, we applied the orthometric corrections computed by the Coast and Geodetic Survey. These 1959 field altitudes are based on the assumption that bench mark N–33 at the West Yellowstone railroad station was unchanged in altitude, and that this stability is indicated by the nearly unchanged relative altitudes of N–33 and the next two bench marks to the east. The releveling survey was ended at Sappington, 110 traverse miles to the north, where the field altitude based on this assumption indicated a 1934–59 altitude change of –0.1 foot; the minuteness of this figure validates the assumptions within very small limits. The good agreement of 1934 and 1959 altitudes of the bench marks from Sappington south to the subsided area is further validation.

Assuming that some other part of the little-changed relevel line represents absolute stability would change the subsidence figures by a maximum of several tenths of a foot. Such a change has been made by the Coast and Geodetic Survey in a provisional adjustment based on the assumption that the first three bench marks (beginning with Y–138) north of the subsided zone remained stable. This adjustment minimizes the changes between the 1934 and 1959 runs outside of the subsided area and lowers the 1959 unadjusted altitudes by as much as 0.2 foot. The assumed stability of West Yellowstone that is incorporated in the data plotted on plate 2 thus gives maximum values for the uplift outside the subsided area.

HIGHWAY PROFILE DETERMINATION

Bench marks within the Madison River canyon west of Beaver Creek were buried by Madison Slide and Earthquake Lake and so could not be recovered during the Coast and Geodetic Survey releveling. The level line was run along the highway nearly as far as it was exposed within the canyon, however, and we determined the highway profile from the individual level-rod measurements. By comparing this postquake profile with the prequake highway profile, we were able to extend the measure of absolute subsidence well into the canyon from both ends. The prequake profile, provided by the Bureau of Public Roads, was in the form of a profile of turning points of tangents to the road surface. Applying the usual engineering formula, we derived the actual parabolic shapes of the prequake road crests and saddles from these tangents.

The old and new profiles in the Madison River canyon are shown in plate 3. Subsidence is demonstrated throughout. Bench marks, as releveled by the Coast and Geodetic Survey, subsided compatibly with the highway as indicated by the profiles; and points on or near bedrock subsided compatibly with those on surficial material. The profile determination was extended along Hebgen Lake, and it was found everywhere compatible with subsidence indicated by bench marks except in the places where slumping of the highway was made obvious by landslides and surficial scarps. The continuous smooth character of the tectonic subsidence is made clear by the profiles.

Small uncertainties in the profile comparison result from differences in height of the shoulders and center of the highway, height of the level instrument, depth to which the level-rod pin is set, and ambiguity as to exact position of ridge crests and valley floors. The uniformity of subsidence shown by comparison of the two profiles, however, makes it clear that such uncertainties generally total less than 1 foot and are far smaller than the 6 to 22 feet of absolute subsidence.

SHORELINE MEASUREMENT

About 160 observations were made around the lake of the vertical difference between the prequake high-water mark and the water level of the moment. The survey was conducted from an outboard-powered boat loaned by the U.S. Army Corps of Engineers. The entire shoreline was examined, and points were chosen where measurements could be made most reliably.

The lake level is measured daily at Hebgen Dam by the dam foreman, so that all shoreline measurements can be related to the Coast and Geodetic Survey level line and thus define the amounts of absolute subsidence. The shoreline measurements were made several weeks after the earthquake, by which time the lake level had been allowed to drop about 5½ feet below the high-water mark at the dam. The deformation of the lake bed was such that the relative water level at the dam after the earthquake was three-quarters of a foot lower than it was before the quake. Shoreline observations were recorded as positive and negative readings from the pool level of the moment and were corrected subsequently for changes in lake level, although during the time of shoreline measurements and of the fathometer survey the lake level varied little more than 0.2 foot.

Where the new lake level is lower than the old, the height of the high-water mark above the present level was measured. The lake is kept full for several months each summer, whereas other levels are inconstant: the lake is drawn down more or less steadily in the fall and winter and filled steadily in the spring. The summer high-water mark is in most places easily recognizable by staining of rocks and posts, by grounded driftwood, by the upper limit of beach sand, and by nicks cut in unconsolidated material. Repeated independent determinations by different geologists showed that the uncertainty in specifying the high-water mark was commonly only about 0.1 foot. Most of the determinations were made with Abney level and stadia board and were reproducible in the immediate vicinity of any observation to within 0.2 foot. Along parts of the south shore of Madison Arm, old and new shorelines were widely separated, and determinations were made with planetable and telescopic alidade.

Measurements were necessarily less accurate where the old shoreline was submerged beneath the new lake level. Suspended organic matter and clay generally limited visibility to less than a foot, so the old shoreline could seldom be seen beneath the water. The submerged shoreline was determined mostly on the basis of its submerged land plants and trees (fig. 43). The depth on the lakeward side of submerged trees, shrubs, and plants was probed with a pole marked in feet and tenths. Willows and herbaceous plants grow down to the high-water mark of emergent sectors, so readings based on them were used without correction. For conifers along the submerged shoreline northwest of the Watkins Creek fan, an arbitrary correction factor of 0.8 foot, based upon inspection of the emergent shore, was added to the observed submergence of the base of the trees. This method can only yield minimum values; in some places probing suggested that the high-water beach was as much as a foot or even two feet below the reported level.

The entire lake basin subsided in absolute altitude, but the lake-level changes relative to the postquake full water level were such that the prequake shoreline of Madison Arm emerged about 5 feet. Grayling Arm was little changed, the shore of the main body of the lake was submerged 5 to 10 feet, and there was little change at the dam. The emergent part of the lake subsided less than did the submergent part. The amount of absolute subsidence varied from about 5 feet in Madison Arm and 12 feet in Grayling Arm to 22 feet in the main part of the lake, but only 9 feet at the dam.

SCARP HEIGHTS

At each new fault scarp, the amount of displacement of the ground surface changes abruptly. The heights used on plate 2 for the major scarps northeast of Hegben Lake are those reported by Witkind (chapter G); he describes methods of determining scarp heights and corrected displacements on the ground surface. Heights of lesser scarps east and south of the lake were measured or estimated by the writers and others.

FATHOMETER SURVEY

A contour map of the lake floor after the earthquake was made by Wayne Jackson and is incorporated in his companion paper, chapter H. No comparable survey was made before the earthquake, but deformation during the earthquake is clearly indicated by the data. As the fathometer-indicated deformation cannot be measured precisely, it was not specifically incorporated in the deformation map, but the isobases of plate 2 are consistent with it.

CHARACTER OF SUBSIDENCE

Most of the points whose subsidence was measured are on unconsolidated material, but some points on both sides of Hebgen Lake are on bedrock (pl. 2). The subsidence is smoothly continuous from bedrock to surficial deposits, so obviously the bedrock beneath the cover sank also. The long uniform new scarps, notably those northeast of Hebgen Lake, clearly reflect faulting in buried bedrock. In the large areas where only surficial deposits are seen, the uniform pattern of subsidence makes it clear that here also the bedrock beneath shared in the deformation. Sliding and compaction are quite inadequate to explain more than local details of the changes. Most of the subsidence was tectonic.

Modification of the tectonic subsidence pattern by slumping and possibly compaction of unconsolidated deposits occurred also. Important examples of this are described by J. B. Hadley (chapter K). Slump-

FIGURE 43.—Submerged land shrubs and trees in Hebgen Lake. Northeast shore of lake, near Kirkwood Creek. U.S. Forest Service photograph.

ing was generally made obvious by features such as landslide scars and fissures and by its local and irregular character. Critical analysis of releveling data in wet bottomlands shows that purely surficial subsidence there cannot have exceeded 1 foot and generally was less than a few tenths of a foot.

DEFORMATION WEST OF MADISON RANGE

The spectacular effects of the earthquake are near Hebgen Lake, in the West Yellowstone basin and adjacent parts of the Madison Range, but substantial deformation extended many miles to the west. Part of the Madison Valley near the latitude of Hebgen Lake also subsided, although not as much as the lake basin. Just west of the valley, subsidence and tilting are both indicated by lake features. East of the valley, a few miles south of the mouth of the Madison River canyon, a segment of the fault along the western base of the Madison Range was reactivated and formed a new scarp a few feet high.

MADISON VALLEY

Releveling of bench marks by the Coast and Geodetic Survey demonstrated that the floor of the Madison Valley subsided 7 feet (pl. 2) near the mouth of the Madison River canyon. The line of levels trends obliquely into the topographic trough of the valley, then follows the trough northward. Five miles below the mouth of the canyon the subsidence was little more than a foot, and at Kirby's Ranch, about 3½ miles farther, it was only a few inches. No subsidence was recorded at bench marks still farther north along the line.

The greatest proved subsidence was thus at the southernmost bench mark surveyed. Subsidence of the Madison Valley was probably near its maximum there, as the bench mark is close to the axis of a major syncline formed by repeated late Quaternary warping (p. 89–97).

Other data, relative rather than absolute, gathered both east and west of the Madison Valley indicate a complex pattern of subsidence for the valley. For example, ditches carrying irrigation water around the alluvial fan at Sheep Creek (pl. 2) were tilted, and some became useless as a result.

CLIFF LAKE AND WADE LAKE

Cliff Lake, which is more than 3 miles long, within the mountains just southwest of the Missouri Flats part of the Madison Valley, was tilted during the earthquake. The lake was visited by J. B. Hadley and District Forest Ranger Neil J. Howarth 5 weeks after the earthquake, and by Hadley again in the summer of 1960. They (written communication, 1960) found the prequake shoreline of the northern part of the lake to be submerged beneath the postquake lake surface. The rise in water level attained a maximum of 1½–2 feet at north end of the lake, as shown by such features as the partial submergence of a parking area. By contrast, in all three arms at the south end of the lake, old and new shorelines coincided. The bed of the northern part of Cliff Lake was thus tilted northward; the bed of the entire lake was not tilted eastward.

The water level of Wade Lake, which is just north of Cliff Lake, was reported by Howarth to be higher after the quake than previously, without obvious difference in relative level from one end to the other. The lake is dammed by an old landslide, and its deepening may have been due to increased inflow or to tightening of the permeable dam rather than to deformation.

The isobase map by Fraser, Witkind, and Nelson (chapter J) of subsidence during the earthquake does not account for the tilting of Cliff Lake.

MADISON RANGE FAULT

A continuous Recent but prehistoric fault scarp runs along the west foot of the Madison Range from north of the Madison River canyon to Mile Creek (pl. 5). At Mile Creek this scarp leaves the front of the range and trends along a canyon oblique to the crest of the mountains. The scarp is most impressive: with a height of 20 to 40 feet and a slope of 40°, it cuts all unconsolidated material but exposes no bedrock. The scarp appears to have formed in a single movement. Although trees several feet in diameter grow on it, such sizable streams as Sheep and Little Mile Creeks still cascade down the scarp.

In 1959 a new, smaller scarp formed on this prehistoric scarp. Between Little Mile Creek and Sheep Creek, a distance of 1.5 miles, discontinuous near-vertical new scarps facing valleyward with up to 3 feet of offset were formed along the old scarp. At the mouth of the unnamed canyon next north of Little Mile Creek the prehistoric scarp was formed in extremely coarse morainal debris containing abundant blocks many feet across. No new scarp formed across the canyon during the earthquake, apparently because of the extreme coarseness of this till. Instead, the blocks were jostled about, and the old scarp became heightened. This is shown graphically by the stretching of a wire fence that goes down the scarp. The fence was stretched so tight that its supporting weighted tripods—no posts could be sunk in this material—were pulled clear of the ground to make a straight line from the top of the old scarp to a tripod about 25 yards beyond the base of the scarp, thus illustrating an increase in scarp height of several feet. A nearly continuous new scarp 1 to 3 feet high extends northward 0.6 mile from this canyon along the old scarp. At Sheep Creek, 1.5 miles farther north, there are new cracks along the old scarp, but vertical displacement on these was limited to an inch or two.

The prehistoric scarp makes an abrupt bend at Sheep Creek, from which it trends northwestward in one direction, and a little west of south in the other; reactivation during the earthquake was limited to the south-trending leg.

DEFORMATION IN MADISON RANGE

The Madison River canyon, entrenched directly through the Madison Range, subsided fully as much as did the adjacent parts of Madison Valley and the West Yellowstone basin. This is shown primarily by comparison of the prequake and postquake highway profiles (pl. 3). All bench marks from Beaver Creek westward to the mouth of the canyon were buried under Madison Slide and Earthquake Lake, but the highway-profile level data extended the measure of proved subsidence nearly as far as the road was exposed—eastward to the landslide from the mouth of the canyon, and westward to Earthquake Lake from the head of the canyon.

At the mouth of the canyon, the last recovered bench mark lies at the Stagger Ranch, just west of the Madison Range frontal fault; this bench mark subsided 7 feet. There was no reactivation of the frontal fault nor any apparent steepening of the bed of the Madison River or other evidence of abrupt warping at the range front. The highway profiles from the Stagger Ranch toward the landslide (pl. 2) show that the mouth of the canyon and the adjacent part of Madison Valley subsided evenly; most values range from 7 to 8 feet. The Madison River runs near bedrock at the mouth of the canyon. Although the exposed part of the highway rests on colluvium, no slump scarps or other evidence for sliding of this material were seen, and the striking uniformity of subsidence from Stagger Ranch to the landslide is convincing evidence for tectonic subsidence.

The interpretation of the 1959 deformation pattern by Fraser, Witkind, and Nelson (chapter J) and by Witkind (1961) illustrates the hypothesis that the Madison Range front was warped during the earthquake and that the subsidence shown at the Stagger Ranch was tectonic whereas that within the canyon was due to slump. In the light of the road-level data, this hypothesis requires that slump and tectonic downwarping were exactly balanced in such a way that the thick fill of Madison Valley was unaffected by settling whereas the very thin fill of the adjacent canyon subsided by the same amount entirely by surficial settling. We regard the evenness of the subsidence of the highway (pl. 3) across the range front as conclusive evidence against this hypothesis.

Subsidence of the upper part of the Madison River canyon is related systematically to the preexisting structural topography. For most of its course between Cabin Creek and Beaver Creek the highway runs along or near a bedrock slope and subsided as little as 6 feet (pl. 3). At Beaver Creek the highway enters a broad moraine-filled basin for which a structural origin seems required; subsidence is 8 feet at the bench mark on Beaver Creek bridge and increases smoothly to 14 feet at the point where the highway disappears beneath Earthquake Lake. Deformation during the earthquake increased the depth of the basin within the Madison River canyon at Beaver Creek.

The Beaver Creek basin is surfaced largely by a crescentic Bull Lake moraine that trends obliquely toward the northwestern canyon wall. The moraine is broken by a narrow trough at the foot of that wall, and on the north side of the trough a remnant of what appears from both photo interpretation and field reconnaissance to be the same moraine occurs 150 feet above the highest part of the moraine to the south and 200 feet above the nearby crest of the moraine. We interpret the valley wall as a fault scarp on which there has been about 200 feet of post-Bull Lake offset, and the topographic trough at its foot as a collapse moat along this fault. (The narrow trough of new subsidence along the Hebgen fault scarp at the lower end of Hebgen Lake might be due to similar moat collapse into a fissure in bedrock at depth.)

In the western part of the Beaver Creek basin, the highway passes less than 100 yards from a bedrock hill that projects through the valley fill; no evidence for slump of the fill away from the bedrock was found during a search. The lack of such slump and the evenness of subsidence of the highway contradict the hypothesis of Fraser, Witkind, and Nelson (chapter J) that the subsidence represents compaction of surficial sediment rather than tectonic change.

The Bureau of Public Roads releveled the highway profile in the Madison River canyon in the spring of 1960 (Lynn D. Tingey, oral and written communications, 1961). This new highway profile agrees very closely with the Coast and Geodetic Survey level line run in September 1959, shortly after the earthquake, except that local postquake subsidence between the times of these surveys is shown. Between September 1959 and the spring of 1960, a section of the highway only 300 feet long, centered 2,900 feet southwest of Beaver Creek, subsided up to 9 feet in addition to the 11 feet of subsidence there during the earthquake, and trees were tilted along the newly dropped section. This narrow new sag is coincident with the topographic trough, noted above, which crosses the road obliquely at this point at the northwest edge of the Beaver Creek basin. The new sag is so narrow that a shallow cause is indicated. Possibly this cause is in slump of surficial material toward the new Earthquake Lake; but the stability of the coarse till, the distance to the lake, and the gentle slopes involved argue against slump. The coincidence with the probable structural trough leads us to infer a structural cause instead. We suggest that the Beaver Creek basin was pulled slightly away from the fault bounding it on the north during the earthquake, opening a near-surface fissure in the Precambrian rocks, and that the surficial materials have slowly collapsed into this fissure since the earthquake.

The Bureau of Public Roads survey ran along the colluvial north shore of Earthquake Lake, along the route of a new road since built. A resurvey in September 1960 showed that between spring and fall of that year a section of this route about 2,000 feet long subsided 1.0 to 1.8 feet. This new subsidence might be due either to slump of saturated surficial material into Earthquake Lake or to moat-type slump, of the sort inferred near Beaver Creek, toward the buried fault which may trend along the canyon wall.

DEFORMATION EAST OF MADISON RANGE

Deformation extended east of the Madison Range into the northwestern part of the Yellowstone Plateau. During September 1960, the Coast and Geodetic Survey releveled bench marks along the highway from West Yellowstone to Livingston, Mont., via Madison Junction, Norris Junction, and Mammoth Hot Springs. Data from this releveling have not been rigorously adjusted and are not shown on plate 2, but application of approximate orthometric corrections to the preliminary figures indicates that the sector between 4 miles east of Madison Junction and Norris Junction subsided 0.3 to 0.7 foot, the maximum having occurred in the vicinity of Gibbon River Rapids at the axis of a youthful syncline that trends N. 80° W. to merge with the West Yellowstone basis. Subsidence continued with decreasing amount for 9 miles north of Norris Junction, beyond which there was uplift—the peripheral upbowing, recognized north of the subsided terrane of the West Yellowstone basin and Madison Valley also—by an apparent maximum of about 0.3 foot, decreasing to zero at Mammoth Hot Springs. Madison Junction, south of the newly deepened syncline, rose 0.1 or 0.2 foot to define the southern marginal bulge. No consistent change was detected between Madison Junction and West Yellowstone.

By far the greatest damage to roads within Yellowstone National Park occurred from rockslides along the Madison and Gibbon Rivers, within and just south of the subsided tract.

The fire lookout on Mount Holmes, in the Gallatin Range between Hebgen Lake and the Gibbon River, reported that a crack opened across the ridge between Mount Holmes and Trilobite Point. Whether this was a fault scarp of small vertical displacement or merely a slump feature in unconsolidated materials was not determined.

GEOMETRY OF DEFORMATION

SUBSIDENCE PATTERN

The absolute deformation during the earthquake can be defined with precision along the line of repeated levels and new highway-profile determination and around Hebgen Lake. Other data, relative rather than absolute, are provided by scarp heights and by such features as the tilting of Cliff Lake. From these diverse data, a deformation-contour map was prepared (pl. 2), the contours representing the changes in altitude that accompanied the earthquake. The map is thus a special type of structural-contour map in which the datum is the irregular prequake ground surface rather than a stratigraphic horizon. It may also be pictured as a structural-contour map of an imaginary surface, horizontal before the earthquake, and deformed during the earthquake into the pattern shown. Such contours were named "isobases" by DeGeer (1892).

Releveling showed much absolute subsidence but has proved a maximum of only 1.7 feet of absolute elevation. The subsidence proved by this surveying is quantitatively adequate to account for all of the observed features, and in preparing the map it was assumed, accordingly, that practically all the deformation during the earthquake was by subsidence. It was further assumed that the change in altitude across new scarps continued far beyond the faults—that is, that large areas moved relative to one another across the faults. This assumption was proved valid wherever it could be tested by level or shoreline data.

The isobase map is tightly controlled in the vicinity of Hebgen Lake and fairly well controlled throughout the West Yellowstone basin and in the Madison River canyon; only minor adjustments could reasonably be considered in these places. Elsewhere, several interpretations could be made. The greatest uncertainties lie in the possible extensions of the zone of deformation to the east and to the west, and the biggest problem is in the behavior of the Madison Range southwest of Hebgen Lake and south of the Madison River canyon.

The West Yellowstone basin and Madison Valley depressions of new subsidence are connected via the continuous zone of subsidence along the Madison River canyon. As the canyon cuts directly across the Madison Range, the nearby part of the range must have subsided also during the earthquake. Whether the Madison Range some miles to the south of the canyon subsided more or less than did the nearby parts of West Yellowstone basin, Madison Valley, and the Madison River canyon cannot be proved; but for reasons outlined below, we infer (pl. 2) that the crest of the range did subside, although less than did the basin.

The existence of the Madison River canyon might itself be taken as evidence for a structural depression across the Madison Range, and the structural basin at the mouth of Beaver Creek may be only a part of a bigger Madison River canyon depression. The general axis of late Quaternary depression of Madison Valley trends eastward toward the mouth of the canyon, as is discussed on subsequent pages. The southwest shore of the lower part of Hebgen Lake is a straight, steep slope for which a fault origin is probable, suggesting that the Madison Range has there tended to stand while the Beaver Creek and West Yellowstone basins have subsided in the past; but no evidence for fault reactivation along this slope was found despite intensive search. The isobases of subsidence trend into this slope at high angles, and surveys showed that Hebgen Dam was not tilted crossways. The releveling data prove continuous subsidence between Hebgen Lake and the Beaver Creek basin.

The greatest subsidence of the West Yellowstone basin occurred well to the south of the Madison River canyon, and most of the isobases trend westward at high angles into the Madison Range. The basin of 1959 subsidence is continuous via the Madison River canyon with the depression of Madison Valley. Absolute subsidence of Madison Valley was determined only northwest of the mouth of the canyon, but the northern part of Cliff Lake is known to have been tilted northward. We infer that Cliff Lake marks the southwest limit of subsidence, that Wade Lake lies along an isobase, and that the axis of 1959 subsidence follows the general axis of late Quaternary warping across Madison Valley. Whether the subsidence basins of Madison Valley and West Yellowstone basin are connected by a broad zone of depression of the intervening Madison Range or only by the proved depression along the Madison River canyon through the range cannot now be demonstrated. In either instance, a continuous basin of new subsidence formed that was elongate east-west rather than parallel to the major new northwest-trending fault scarps.

Synthesis of the data (pl. 2) suggests a broad compound basin of subsidence that plunges gently eastward across the upper Madison Valley, the southern part of the Madison Range, and the West Yellowstone basin. It is markedly asymmetric, with the north flank dipping much more steeply than the south. In the West Yellowstone basin the south flank is a gently dipping platform whereas the north flank is terminated obliquely and abruptly by the large new scarps of the Hebgen, Red Canyon, and other faults. Decreasing subsidence continues eastward into the Yellowstone Plateau at least as far as Gibbon River Rapids. The basin of proved subsidence is 43 miles long, from Norris Junction to Cliff Lake, and trends N. 78° W.

Releveling by the Coast and Geodetic Survey demonstrates that a belt 9 to 19 miles wide flanking the subsided terrane on the north was elevated, mostly only 0.1 to 0.3 feet, above previous altitudes between surveys, presumably during the earthquake. Measured uplift reached 1.7 feet along Grayling Creek, about 1.5 miles north of the major scarp near Duck Creek. Elevation in Madison Valley and in Yellowstone National Park between Obsidian Cliff and Mammoth Hot Springs was by a maximum of several tenths of a foot.

RELATION BETWEEN FAULTING AND WARPING

Various combinations of warping and faulting by which the subsidence during the earthquake was accomplished are illustrated by the profiles in figure 44. Profile *A–A′* illustrates the situation in which the near-maximum subsidence of 18 feet took place by displacement along a single fault, the Hebgen fault. The net displacement of the ground surface by the fault (the height of the new scarp, 20 feet, less a 2-foot correction for small backfaults) is 18 feet, essentially equal to the absolute subsidence proved by the highway profiles and the warping of the bed of Hebgen Lake immediately to the south. The downwarped area was not simply tilted, as its profile of subsidence is concave upward. Northwestward along the Hebgen fault the height of the 1959 scarp decreases and the subsidence measured by displacement of nearby bench marks on the downdropped block decreases correspondingly. The northeast margin of the deformed area is thus the scarp of the Hebgen fault northwestward from profile *A–A′*, and in this short interval the total subsidence is effectively represented by the height of the new scarp.

The pattern is quite different southeastward from profile *A–A′*. The new scarp on the Hebgen fault becomes lower so rapidly that it is only 2 feet high 0.6 mile from *A–A′*. As the subsidence of the highway, which is only a few hundred yards southwest of the fault, remains constant at 22 feet in this distance, the dwindling of the scarp must record a downwarp toward the south of the high-standing block northeast of the fault. The Hebgen fault, which forms the margin of the subsidence basin at profile *A–A′*, changes southeastward to become an element of relatively minor displacement that extends for 7 miles within the deeply subsided tract. The northeast margin of the subsided basin has shifted northeastward to the scarps and associated warps of the West Fork and Red Canyon faults.

The orientation of the downwarp that diverges from the Hebgen fault cannot be firmly established, for the only control by which deformation can be determined northeast of the Hebgen fault is the height of fault scarps. However, as the new scarp of the West Fork fault trends southwest in the footwall of the Hebgen fault, directly toward the segment of the Hebgen fault where the abrupt downwarp is recorded, it is a fair assumption, that the downwarp and the West Fork fault are parallel, interdependent features. This interpretation is shown on plate 2.

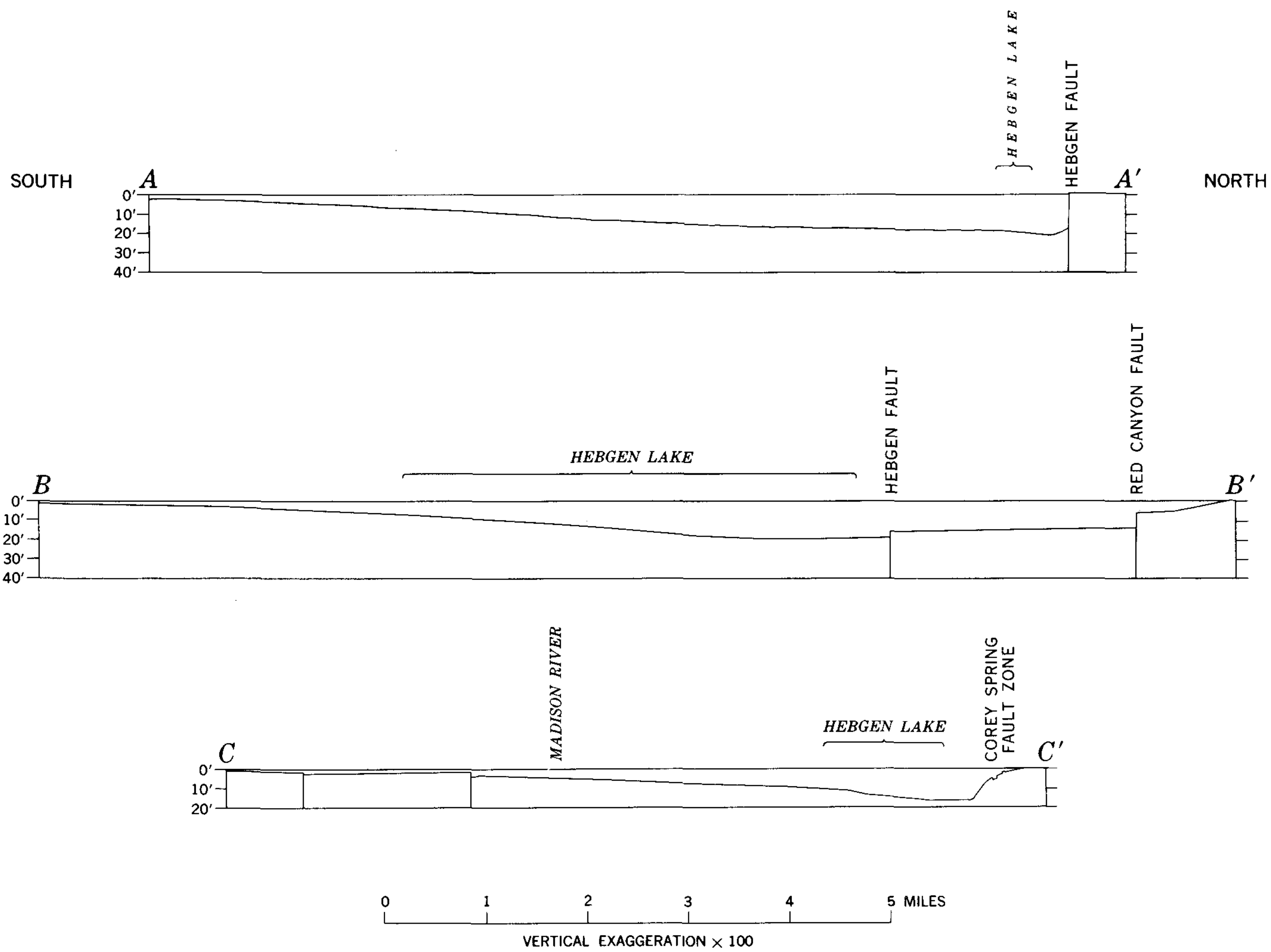

FIGURE 44.—Profiles of subsidence of horizontal plane during earthquake of August 17, 1959. Lines of profiles shown on plate 2.

The new Red Canyon fault scarp, the longest formed during the 1959 earthquake, begins a mile east of the point at which the West Fork downwarp diverges from the Hebgen fault. For the first 2½ miles of its course the fault trends northeastward parallel to the scarp of the West Fork fault, but it then curves through 90° and trends southeastward. New displacement on the Red Canyon fault increases to a maximum of 12 feet in the big curve. This increasing displacement records the eastward transfer of subsidence from the West Fork structures to the larger scarp.

Followed southeastward from the 18-foot maximum height at profile *A–A′*, the Hebgen scarp dwindles rapidly to 2 feet as described above, then increases to 9 feet in about half a mile, and diminishes again to 5 feet in a like distance. An essentially constant subsidence of the lakeshore over this distance indicates not only the presence of the West Fork downwarp but suggests a low basinward rise as well. The scarp maxima that suggest this rise are at the apparent intersection of the Hebgen fault and the axis of the arcuate anticlinal fold which controls the Red Canyon fault (pl. 5). The isobases incorporate the implication that the old fold influenced the subsidence.

Subsidence profile *B–B′* illustrates the effect of the displacements of the Red Canyon and Hebgen faults and of the West Fork downwarp upon the pattern of deformation. The basin of subsidence is broad and asymmetric. The deepest point lies more than a mile south of the Hebgen fault, not against the scarp. The remarkably smooth profile of subsidence south of the scarp is controlled by 34 instrumentally determined points, most of which are measurements of the warping of the prequake shoreline; all are within 1,800 feet of the line of profile and within 5½ miles south of the Hebgen fault.

The sinuous Red Canyon fault varies irregularly in height along its main southeastward trend. By analogy with the known behavior of the subsided Hebgen Lake block and with the proved abrupt warping below the faults at Corey Spring (discussed later), it is presumed that the irregularities record variations in the amount of warping accompanying the faulting. Maximum net displacement across the new scarp is 15 feet, a value found at two localities. These two places are interpreted as points where the total displacement of the evenly subsided Red Canyon fault block was by movement on a single fault. We assume that elsewhere along the fault, where the scarp varies in net height to as little as 8 feet, the sum of fault displacement plus warping is from 13 to 15 feet.

The Red Canyon fault continues with its irregular southeast trend across Red Canyon Creek; then, a mile and a half north of Grayling, it curves southward, rapidly dwindles in height, and disappears as a continuous feature but gives way to a curving complex of many small scarps distributed irregularly over a broad zone. Net displacement on each small fault is no more than 2 or 3 feet. (In chapter G, Witkind retains the name "Red Canyon fault scarp" for this zone and for the new scarp southeast of Grayling Creek into which it trends.)

The broad zone of disconnected fissures and small scarps curves from a southward trend along the east side of the valley of Red Canyon Creek to an eastward trend along the hillside north of Hebgen Lake. One scarp swings southwestward instead of southeastward as this zone of general curvature is entered. The main zone of small faults, here referred to as the Corey Spring fault zone, continues eastward north of the lake and past Corey Spring to another area of curvature where it swings again to a southeast trend. The Corey Spring fault zone continues as a broad zone of disconnected, subparallel faults almost to Grayling Creek, where it gives way to a single high, impressive new scarp.

The displacement on many of the small faults of the Corey Spring zone is down on the uphill side and is thus directed opposite to the general sense of displacement. Some of the scarps must represent movement of purely superficial material and be part of the pattern of downhill mass movement of talus and glacial till, some results of which are described by Hamilton in a section of the companion paper by Hadley (chapter K).

The low and discontinuous character of the scarps of the Corey Spring fault zone might suggest that the basin to the south subsided relatively little along this interval, but this is not so. Displaced bench marks along the highway immediately south of the zone prove more than 15 feet of subsidence. This subsidence is taken up against the fault zone by a relatively steep downwarp. At Corey Spring, the lakeshore approaches to within a third of a mile of the fault zone in a deeply indented bay. The tilting of the shoreline of this bay and the changes in altitude of nearby bench marks prove that warping here, without recognized faulting, accomplished a change in altitude of 9 feet in a horizontal distance of about 650 feet; this is equivalent to a slope of ¾ of 1° (fig. 45).

The prequake shoreline of the bay at Corey Spring was warped smoothly downward, and the least subsidence was at the head of the bay. The strike of the warp is N. 73° W., a direction clearly defined by the intersection of the postquake water surface and the deformed prequake shoreline on opposite sides of the bay. From this line, at the time of measurement, the old shoreline rose smoothly northward to a height of 5.3 feet above the water level at the head of the bay, 350 feet distant. Several hundred feet northeast of the head of the bay, a pair of bench marks on opposite sides of the highway record the continuation of the warp. The southward slope of the downwarp is slightly convex upward.

These data prove that most of the subsidence in the vicinity of Corey Spring was accomplished by relatively abrupt warping (profile *C–C′* fig. 44). The amount of subsidence accounted for by this warping is essentially the same as that due to faulting alone in some places (profile *A–A′*) or to the combination of faulting and warping in others (*B–B′*). The northeast boundary of the subsided basin at Corey Spring is an abrupt downwarp, broken near its upper limit by a zone of minor discontinuous faults.

Discontinuous, low scarps continue eastward from Corey Spring almost to Grayling Creek, where a single high scarp appears abruptly. The scarp is 1 or 2 feet high 50 yards uphill from the edge of the valley fill, yet 8 feet high at the base of the hill and 14 feet high a short distance southeast of Grayling Creek (pl. 2, and figs. 36 and 37). This abrupt increase in the scarp height represents the change in the boundary of the basin from a downwarp, as at Corey Spring, to a single fault scarp. The monocline of the warp must trend into and terminate against this fault.

The fault trends southeastward from Grayling Creek, although the single scarp gives way to a series of discontinuous scarps within a narrow zone. Aggregate displacement on these scarps decreases irregularly, and where the fault curves to the east displacement is only about a foot. These low scarps dwindle

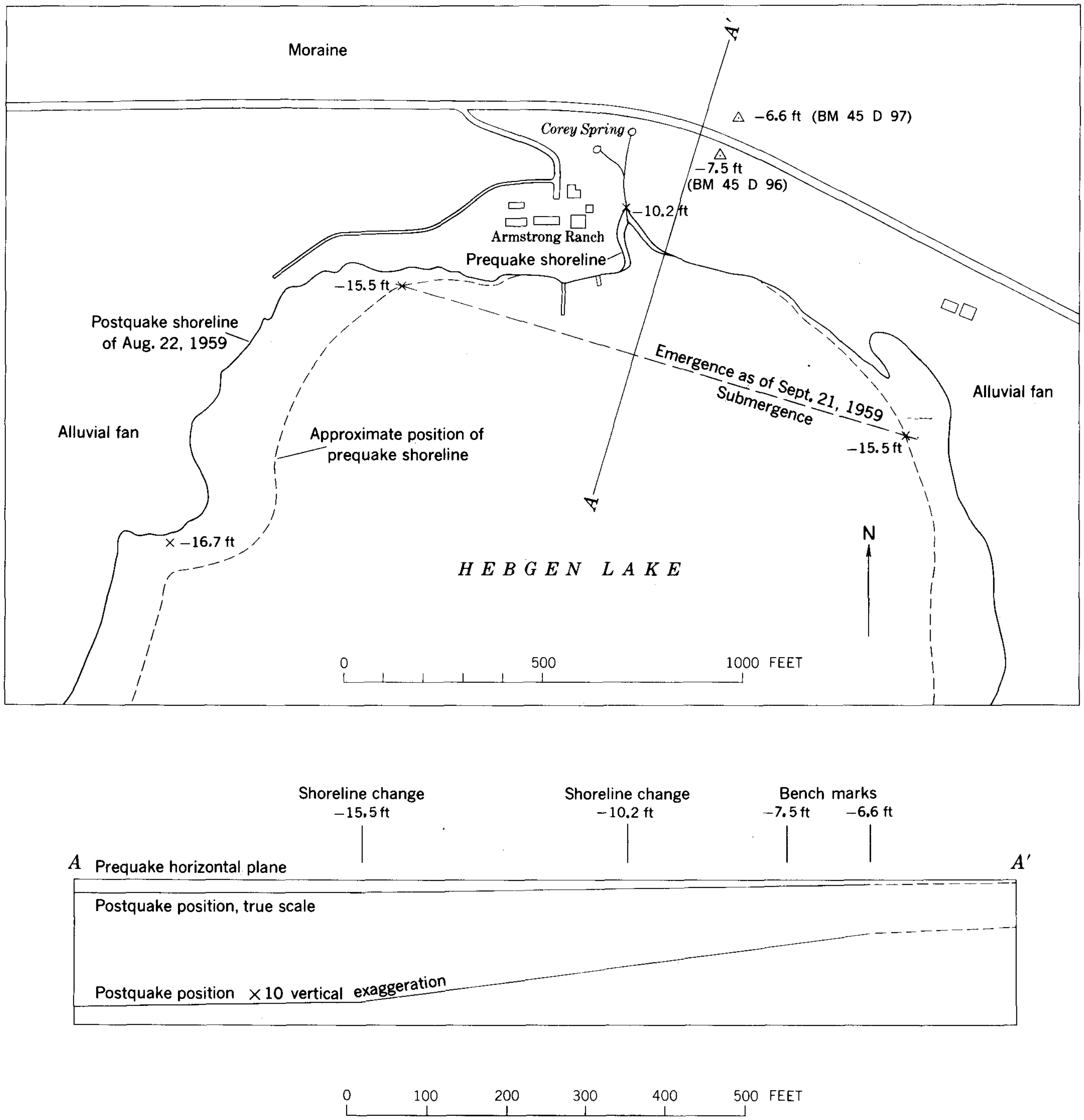

FIGURE 45.—Map and profile showing warping south of Corey Spring fault zone. Numbers give absolute subsidence, in feet, as determined by releveling of bench marks, by relations of old and new shorelines, and by submergence of land plants. Map enlarged from aerial photograph.

in turn and give way to a zone of small disconnected scarplets with only an inch or two of vertical displacement; at the park boundary only a few small cracks without vertical offset appear. This decrease in scarp height is due to the decrease of absolute subsidence southward, as is shown clearly by the releveling of the bench marks.

RELATION TO OLDER STRUCTURES

STRUCTURES NORTHEAST OF HEBGEN LAKE

The basin of new subsidence ends obliquely and abruptly northeast of Hebgen Lake against the Red Canyon and Hebgen fault scarps. That these large new scarps and the major warps associated with them are related systematically to older structures can be seen readily by comparing the geologic map (pl. 5) with the "earthquake" map (pl. 2).

The broad pattern of structures produced during the Laramide orogeny, which culminated in very early Tertiary time, is one of northwest-trending folds, many of them overturned northeastward. Dips in the upright limbs are low, commonly less than 20°, and in various directions, whereas the overturned limbs generally dip steeply to moderately southwestward. The folds are broken by southwest-dipping thrust faults that are roughly parallel to the fold trends. These Laramide structures were deformed during later Cenozoic time by normal faulting and associated warping.

RELATION OF MODERN FAULTING TO CENOZOIC FAULTING

As the 1959 scarps northeast of Hebgen Lake are commonly high on hillsides but below the lowest outcrops, with only talus slopes beneath, there is no direct way to prove the interpretation, shown by plates 2 and 5, that these new scarps are coextensive with older major normal faults in the bedrock. The existence of high ridges above the new scarps, however, is presumptive evidence that much of the relief is indeed structural. It is instructive to examine the only four localities at which bedrock was found to crop out near the scarp on both flanks.

The new fault scarp east of Grayling Creek is high and impressive where it cuts surficial deposits but abruptly dwindles to small disconnected and highly irregular scarplets as it enters the bedrock hill formed of Pliocene(?) welded tuffs immediately west of Blarneystone Ranch. The rhyolite shows no evidence of older faulting, whereas the surficial deposits were broken in previous late Quaternary time by two generations of major scarps. This suggests that there is no older fault along the course of the new breaks in the bedrock area.

The Red Canyon scarp is continuous and impressive where only surficial deposits are on the downhill and downdropped side; but near its west end, where bedrock is encountered in both blocks, it disintegrates into small, irregular, discontinuous scarps within a single formation.

Near the east end of the new scarp of the Hebgen fault north-northwest of Lakeview, Cambrian carbonates crop out a few yards above a single new scarp that is about 4 feet high, and large outcrops of rhyolite welded tuff lie only 30 yards below the scarp. The linear bedrock contact can reasonably be assumed to be a preexisting fault, and the clean new break between the two rock types reinforces the assumption. A normal fault here has probably dropped the rhyolite down against the limestone, and was reactivated during the 1959 earthquake. It is significant that here, the only place where an old bedrock fault is clearly demonstrable, the new scarp is a continuous feature.

The Hebgen fault southeast of Kirkwood Creek is represented by a continuous single new scarp on each side of a limestone knob. As these scarps approach the large but discontinuous limestone outcrops from both sides, they break up into a number of small scarps. The bedrock of the ridge is shattered by a complex of fissures and small extremely irregular faults, dropped down on either side; where these breaks cut nearly continuous outcrops, there is no clear evidence for older displacements. The new breaks represent no lithologic contacts, and there does not appear to have been a previous topographic scarp. A limestone outcrop west of the new breaks, however, dips eastward moderately, but east of the quake scarplets the dips are overturned. The divergence of dips may indicate an old fault, but the relations suggest instead that the new scarps advanced into previously unfaulted rock at this locality.

Despite this general lack of conclusive evidence for faulting of exposed bedrock, there is much evidence in the surficial material for previous faulting during Quaternary time on some of these same trends. The northernmost fissures of the Corey Spring fault zone, for example, lie along the base of a steep bedrock slope and follow closely an old lineament developed in both bedrock and surficial deposits. Just west of the prominent gully north of Corey Spring, the prequake lineament crossed a colluvial terrace as a narrow straight welt; during the 1959 earthquake this was outlined by new opposed scarplets, largely mapped by Witkind, and thus was made recognizable as a reactivated horst. Half a mile to the west, the old lineament splits: one arm continues westward toward the alluvial fan of Red Canyon Creek, and the other curves northward along the hillside on the east side of the valley; this latter trend is emphasized by a new fissure. The curving features roughly parallel the beds of the Paleozoic sequence higher up the hill, which swing northward into the west flank of the steeply overturned northward-plunging syncline. The old lineament that continues westward apparently marks a line of earlier movement on the Corey Spring fault zone. A thousand feet north of the divergence, new fissures of the two trends cross; this emphasizes the contrast in the structures that control subsidence here. Half a mile north of the divergence, the short irregular fissures of the northward trend give way to the new scarp of the Red Canyon fault, which rises gradually northward as a single scarp to attain an impressive height.

The new scarp east of Grayling Creek is on a fault that has had at least two previous episodes of late Quaternary displacement (figs. 36 and 37). Old and new structures die out southeastward via a monocline of decreasing height; but in the surficial deposits 3 miles east of the end of the structure, another old fault scarp rises gradually in the sand plain to become a prominent structure cutting the east margin of the West Yellowstone basin. Whether or not this old scarp was reactivated in 1959 was not determined.

The bedrock fault between Cambrian limestone and Pliocene(?) rhyolite, reactivated in part as described above, separated varied pre-Tertiary rocks from rhyolite over a considerable distance where it was not reactivated in 1959. East of the newly active part, this fault has probably offset the base of the Paleozoic section by about 400 feet, and all or much of this may have been accomplished since the deposition of the rhyolites. No Quaternary surficial units are recognizably offset along the fault trace, in contrast to the abundant scarps in unconsolidated materials to the south and east. At Kirkwood Creek, however, the surface of the Pinedale fan is furrowed by a prominent old drainage course, now dry, that heads just below the new Hebgen fault scarplets and appears to mark the site of a group of springs that were created by slight prehistoric movement on the fault.

The toe of the Kirkwood fan is cut by a prominent transverse scarp that is interpreted (pl. 5) as an erosionally modified fault scarp of interglacial age which parallels the Hebgen fault, about 350 yards to the east. A Pinedale channel of Kirkwood Creek was deflected along the base of the scarp, at right angles to its upstream course, by a slump graben or moat. The topographic offset across the scarp ranges from 80 to 100 feet and appears to reflect fault displacement of a surface of Bull Lake age. To the southeast, alinement of three springs now within the lake but shown on an unpublished map of the prelake basin, marks the continuation of this apparent fault. The small alluvial cone a mile southeast of Kirkwood Creek is discordantly below the dry gully at its head and seems to have slipped down along an upslope refraction of the fault—much as the fan at Hebgen Dam has done on the refracted Hebgen fault during this earthquake.

The position of the new Hebgen and Red Canyon fault scarps at the base of high bedrock ridges that stand above long talus slopes is convincing presumptive evidence that these ridges are in part due to previous normal faulting. Topographic evidence for such faulting would be ephemeral, as the scarps formed would greatly oversteepen unconsolidated talus which lay already at its angle of repose, and debris shed from cliffs above would hasten the burial or destruction of the scarp. The talus would also bury any outcrops on the downhill side. The bedrock ridges rising above the impressive new fault scarps must be in substantial part products of previous offsets along the same faults.

It is conversely evident that where the new scarps trend into bedrock areas not marked by prequake scarps, as in cases described above, there had not been appreciable late Quaternary movement on faults in that bedrock. As fault scarps are widely preserved in unconsolidated material (other than talus) of varied late Quaternary ages in the region, it is reasonable to expect that bedrock scarps of similar age should be still better preserved; their absence is clear evidence of the lack of late Quaternary faulting of those outcrops.

The new fault scarps north of Hebgen Lake thus seem to be partly on faults which had previous, and perhaps major, offset during late Quaternary time, and partly on extensions of those faults into areas which had little or no such offset. The old faults are increasing in both total offset and lateral extent.

CONTROL OF MODERN FAULTING BY LARAMIDE STRUCTURE

Older Quaternary displacements of the faults north of Hebgen Lake seem to have taken place in the zones of easiest slippage, as determined by the attitudes of the pre-Tertiary rocks, whereas what appear to be new extensions of the faults tend to be in zones where attitudes are less favorable for slippage. This is readily seen upon comparison of plates 2 and 5.

The new Red Canyon fault scarp is parallel to the bedding of the limestones of the high-standing block. The strata dip moderately to steeply lakeward in the strongly overturned limb of a fold and strike through a broad arc which has a curvature of 90°. The Red Canyon scarp closely follows this arc, and for much of its length diverges very little from a stratigraphic horizon on which slipping is expectable—the contact between the massive Mission Canyon limestone and the thin-bedded shaly Amsden formation. Displacement on the fault has been down the dip of the beds and clearly has occurred there because of this favorable attitude for slipping. Above the scarp, the high bedrock ridge, presumably itself a fault scarp, is coextensive with strata with this general attitude.

The scarp abruptly disintegrates into crude zones of small scarps and fissures at both ends where the overturned Paleozoic beds roll to gentle upright attitudes and the ridge loses its commanding height. Where the beds are oriented unfavorably for slippage and must be crosscut by any normal fault developed, there has been little or no previous modern faulting. The Red Canyon fault is now being extended into those areas.

The situation along the Hebgen fault scarp is similar. The trend of the fault is straight so that no concentricity of fault and strata can be demonstrated, but fault and beds are essentially parallel in strike for most of the length of the fault. Southeast of Hebgen Lake Lodge, 1959 displacement in the fault zone is represented by small discontinuous scarps and fissures that extend for 1½ miles in rocks that have a gentle to moderate northeast dip. The beginning of a single continuous scarp is coincident with the abrupt overturning of the beds to a steep southeast dip just east of the lodge. Farther northwestward, the prevailing dip of the beds exposed above the scarp rolls back to an eastward position, and the scarp is parallel to the beds in strike but cuts directly across them in dip.

Throughout a length of more than 4 miles, the Hebgen fault scarp is but a few hundred feet west of a west-dipping thrust fault. Beneath this thrust the beds are overturned almost everywhere and are nearly parallel to the thrust; farther northwest this thrust becomes nearly a bedding-plane feature. It seems that at shallow depth, the Hebgen fault, which at the surface is steeper than the thrust, is either coextensive with the Laramide thrust fault or else lies closely below in the nearly parallel beds of the footwall. North of Hilgard Lodge the scarp cuts obliquely across the overturned sedimentary contact of the Cambrian beds on the Precambrian metasedimentary rocks and remains within the latter as far as Cabin Creek, 1½ miles farther north. At the creek the scarp enters upper Paleozoic limestones—presumably by crossing a normal fault that bounds the north wall of the Madison River canyon, although no such fault is shown on the geologic map (pl. 5)—and remains in them to its oddly curved terminus one-half mile to the northwest. Despite the apparent complexity of these detailed relations, the recently active fault is broadly parallel to the overall attitude of the Paleozoic strata and to thrust faults within the sequence.

The deformation map (pl. 2) shows that faulting and warping went on together along the 1959 scarps. Major scarps of both the Hebgen and the Red Canyon faults lie updip along panels of basinward-dipping planes of easy gliding formed by the overturned limbs of folds. The height of new scarps tends to vary systematically with the dip of the strata. Deformation has occurred by faulting where slippage was easy and by warping where it was more difficult. Presumably this was true in the past also.

Of the three major fault scarps developed during the 1959 earthquake, the two that are in areas of exposed bedrock are rigorously controlled by structures—bedding planes—which can hardly extend to a depth of more than 1 or 2 miles. The surface fault pattern must accordingly differ from the pattern of deeper, and more fundamental, displacements.

STRUCTURES SOUTH OF HEBGEN LAKE

The continuity of the many faults and monoclines, both of 1959 and older, that displace the surfaces of the sand plain, moraines, and other deposits of unconsolidated material, indicates that they are due to displacements of the underlying bedrock, although the bedrock is not exposed within most of the basin. No detailed analysis of the relation between the 1959 faults and monoclines and the older bedrock structures can be made. The West Yellowstone basin is blocked out by faults on both north and south sides, although to a large extent the structural relief between basin floor and adjacent highlands is due to warping rather than faulting. The faults have been repeatedly active during late Quaternary time. As is brought out in a later section, we believe that those faults and monoclines which have easterly trends, most of which are south of the Madison Arm, have a late origin and are not Laramide features.

Three fault zones in the southwestern part of the West Yellowstone basin converge toward Targhee Pass, the low route across the Madison Range west of West Yellowstone. One zone trends northwestward toward the pass along the topographic margin of the basin; the other two curve westward and southward, respectively, toward the pass. All three zones arc together: they do not intersect, they converge. No scarps were recognized to cut the surficial material of the pass, and the bedrock there was not mapped; but this convergence suggests that major structures cross the mountains through the pass. Two of these three fault zones in the basin have displacements that increase the relief of the basin, but the third has the opposite sense of offset: along the zone trending southward along the west side of the basin, the west, or mountain-facing, side is the downdropped one. Viewed from the framework of the Madison Range, however, all three fault zones have the same sense of displacement, as though the Madison Range north of Targhee Pass has been downfaulted along varying combinations of these three fault zones.

EARLIER WARPING OF THE WEST YELLOWSTONE BASIN

The distribution of surficial deposits is evidence that the West Yellowstone basin has been warped during late Quaternary time, before the 1959 earthquake. Major changes in the drainage pattern of the surrounding area are further evidence of long-continued sagging.

SURFICIAL DEPOSITS

Hebgen Dam is at the entrance to the narrow Madison River canyon, where the river profile steepens abruptly (fig. 46). A detailed geologic section made of the dam corewall excavation by Montana Power Company engineers shows that the transverse bedrock surface is nearly horizontal and is lower than the modern gravel-filled channel across the entire length of the dam. At the northeast end of the dam, surficial material is 115 feet thick, and bedrock lies 25 feet below the modern channel. The unconsolidated material, as interpreted by Richmond (chapter T), includes, from the base upwards, obsidian sand; a persistent layer of red clay inferred to be a fossil soil developed on pre-Bull Lake deposits; more obsidian sand; and, intertonguing with and overlying the upper sand, a steep alluvial fan of Bull Lake age. Growth of the fan narrowed the canyon and deflected the river against the southwestern bedrock wall, but there has been no lowering of the bedrock lip of the West Yellowstone basin during late Quaternary time. Sources for the obsidian sand are all to the southeast, and drainage of the West Yellowstone basin through the Madison River canyon is thus established for a considerable part of the late Quaternary.

The base of the Bull Lake till beneath the lake basin has been located uncertainly in well logs. Even the most conservative interpretation indicates a reversed slope of more than 50 feet for the base of these deposits between the dam and the south shore of the Madison Arm and suggests downwarping of at least 80 feet.

If the bedrock floor of the basin once drained to the Hebgen damsite, a greater downwarping is indicated. The obsidian sand has been penetrated only in two oil-test wells, 2 miles northwest of West Yellowstone; there the sand is about 200 feet thick, and its contact with rhyolite bedrock is essentially at the altitude of the bedrock rim of the basin at Hebgen Dam. A water well on the south shore of Madison Arm, 6 miles closer to the dam, did not reach bedrock at a depth of 240 feet and thus reached sand about 130 feet lower in altitude than the bedrock rim at the dam. The bedrock floor of the basin has a minimum closure of 130 feet. Downwarping, with a low between West Yellowstone and Horse Butte and a closure of several hundred feet, is indicated. The axis of 1959 subsidence was about 5 miles north of the apparent low indicated by the well logs.

CHANGES IN DRAINAGE

Long-continued sagging of the West Yellowstone basin is responsible for some major changes in the drainage pattern of the surrounding area. At least part of the area of the present basin probably once drained northeastward to the Gallatin River. The deep canyon of the Gallatin River along the State boundary is continuous with that of Grayling Creek, and so inconspicuous is the low divide separating the oppositely flowing streams as to give the illusion that Grayling Creek even now is simply the upper Gallatin. The upper part of present Grayling Creek flows northwestward, an orientation which suggests that this part formed as a tributary to the north-flowing Gallatin. Numerous other drainage details, among them the barbed tributaries of the now southward flowing part of Grayling Creek, reinforce this suggestion. Drainage of the West Yellowstone basin toward the Gallatin seems to have been interrupted by backtilting. Along the northeast edge of the basin, the present drainage pattern suggests much modification of the earlier north-draining streams by headward erosion of tributaries from the sinking basin. Farther within the basin the drainage is entirely centripetal; a smooth bowl-shaped basin is cut radially by drains which suggests that the streams are consequent on a surface shaped by basinal sagging. This relation is thought to hold for the slopes south and east from Grayling Creek, past the entrenched Madison River, and southwestward to the east margin of the great lobate flow of post-Bull Lake rhyolite south of the town of West Yellowstone (pl. 5).

The extension of this downwarp eastward into the Yellowstone Plateau is shown by a broad syncline, in welded tuff, whose axis strikes about S. 80° E. and lies about 5 miles north of Madison River from the Park boundary to Madison Junction. The youth of this syncline is indicated by the drainage reversals that occur along its northern hinge line. Straight Creek, 17 miles east-northeast of West Yellowstone and just north of the area shown on plate 5, is an underfit stream that flows northward to the Yellowstone River. It heads just north of an imperceptible drainage divide in a conspicuous old channel that is continuous southward to the Gibbon Geyser Basin. At the northwest edge of this basin the old channel, little changed at this point, turns east, and although modified strongly by later erosion and deposition, it appears to continue eastward beneath the Recent rhyolite dome of Gibbon Hill, east of Gibbon Meadows and about 6 miles from its present divide. The southern limit of the essentially unmodified part of the channel is 3 miles from the modern divide at an altitude nearly 150 feet lower than the windgap. Assuming that the now reversed drainage had the same northward gradient as does the present upper part of Straight Creek, a backtilting of about 275 feet is recorded in this 3-mile stretch of channel. Three miles to the east, the upper part of the former channel of Obsidian Creek, which parallels that of Straight Creek, is similarly backtilted. The axis of drainage reversals trends eastward. The syncline defined by these reversals and by the deformation of the welded tuff was deepened as much as 0.7 feet between level surveys, presumably during the earthquake.

Within the West Yellowstone basin, continued sagging has clearly controlled stream processes. South of the present axis of subsidence and west of U.S. Highway 191, the sand plain on the south side of the Madison River is marked by the traces of old meander scars over a belt locally more than a mile wide. These scars show a consistent northward shift of the former channels. The north bank of the river is almost everywhere a single prominent escarpment cut by the modern (prereservoir) channel of the river. Northward tilting during the channel cutting is indicated. East of the highway the headward trend of the river is more to the south, and the stream flows down the slope of the basinal sag; the meander belt narrows upstream, and no shift of the stream toward the right bank is evident. The gradient of the stream in this downslope section is anomalously high, as would be expected (fig. 46).

Tilting is also suggested by the apparent northward migration of Cougar Creek and its abandoned channel to the south. The present southern distributary of the creek probably uses part of the old course.

In the southwest corner of the basin, progressive eastward shifting of meander scars of the South Fork of the Madison River is apparently due to basinal sagging (fig. 35).

The Grayling and Red Canyon fans are near the axis of subsidence of the recent earthquake (pl. 2). The surface of these features is considered by Richmond (pl. 5) to be Pinedale (latest Pleistocene) in age, but their size suggests a considerably greater age for their cores. The lack of dissection of the fans implies that the axis of sagging has been in about the same position—that is, near the point of entry of these streams into the basin—since the inception of the fans.

At some time before the late Quaternary, the basin probably emptied southward into the upper Snake River Plain. The basin is now rimmed to the south by huge flows, each hundreds of feet thick, of late Quaternary rhyolite; the flow immediately south of West Yellowstone is younger than Bull Lake moraines. The Madison Range dwindles southward, and at Reas Pass (altitude 6,932 ft.) the crest is less than 400 feet above Hebgen Lake. Before the eruptions the basin probably drained southward around the end of the Madison Range. The ancestral north fork of the Snake River probably headed where Beaver Creek now flows southward toward Hebgen Dam. Lowering of the site of the present Madison River canyon is thought to have taken place by downdropping of the block south of a fault postulated in the canyon, with consequent reversal of drainage.

RELATION TO REGIONAL STRUCTURE

A comparison of the overall pattern of deformation and seismicity accompanying the Hebgen Lake earthquake with evidence of recent deformation in the surrounding region suggests that the deformation of 1959 is part of a complex of sagging that extends eastward from the Centennial Valley and Henrys Lake depression across the older north-trending structures of the Madison Range and Madison Valley and that may be related to the Snake River Plain to the south.

MADISON VALLEY

The Madison Valley is a north-trending structural trough that lies near the crest of an older Laramide uplift. The valley is bounded on the east side by the impressive but much eroded fault scarp of the Madison Range for most of its length. At the south end of the Madison Range, in the Snake River drainage south of the Madison Valley, faulting is of minor importance. According to J. B. Hadley (oral communication, 1960), the west side of the Madison Val-

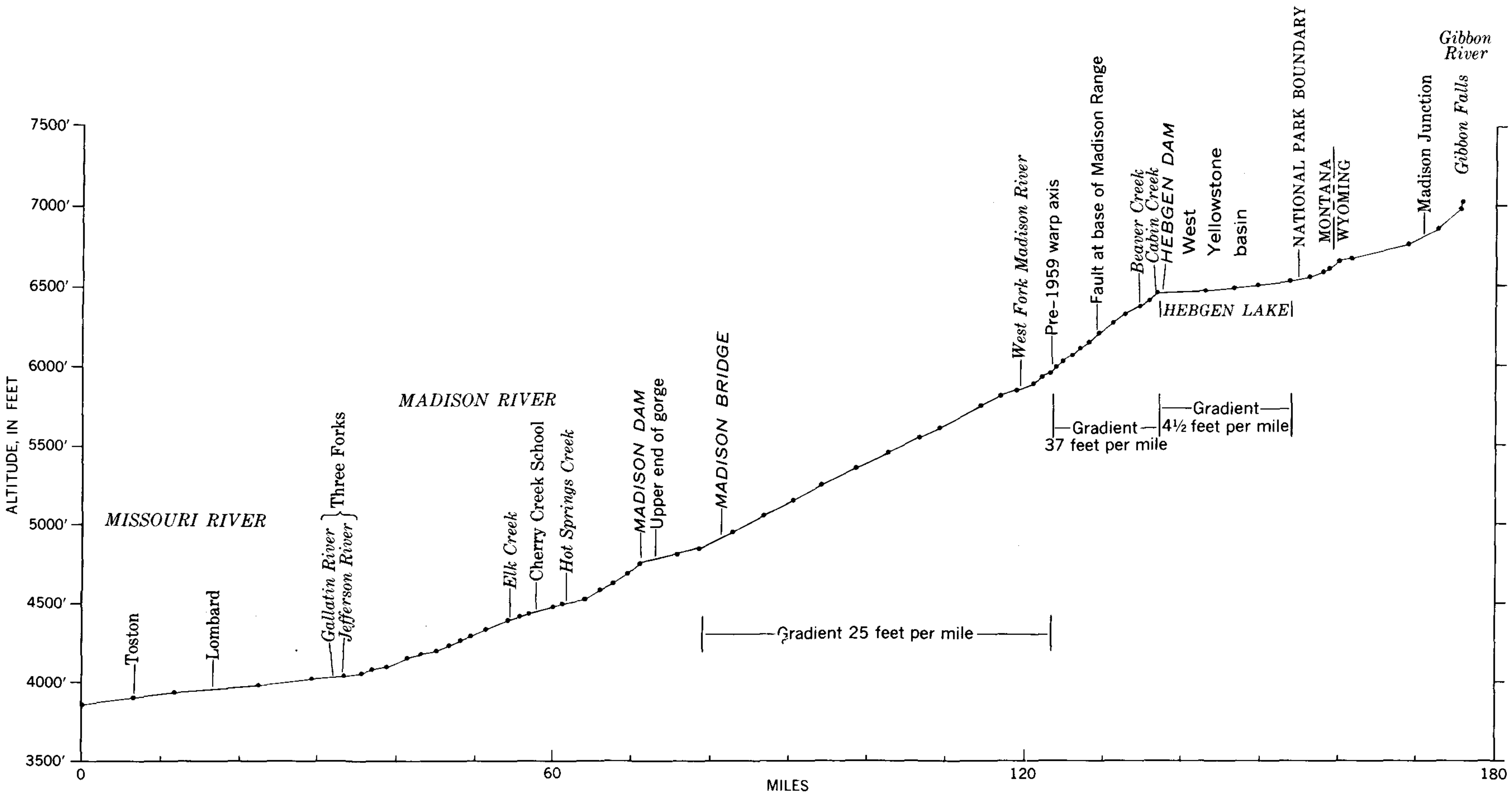

FIGURE 46.—Longitudinal profile of the Missouri, Madison, and Gibbon Rivers, from Toston, Mont., to Gibbon Falls, Wyo. The abrupt flattening of the gradient at Hebgen Dam occurs where there was an abrupt inflection of the north flank of the basin of 1959 subsidence. Within the preexisting West Yellowstone basin, profile changes are probably due largely to deformation: Madison River starts down the south flank of the basin of late Quaternary subsidence just east of the Wyoming State line, crosses the axis within Hebgen Lake, and then flows along the north flank. The known basin of late Quaternary warping whose axis trends along Madison River from its West Fork toward the mouth of the Madison River canyon seems to be reflected by a slight backtilting of the profile near West Fork, but closer to the canyon the gradient is even steeper than it is below the backtilted stretch; either regional uplift south of West Fork, or various hydrologic factors, might have caused this anomaly.

ley is bounded against the Gravelly Range by a series of faults and warps, and the valley is largely younger than the volcanic rocks of Oligocene age and may be of very late Cenozoic age.

A study of aerial photographs of former courses of the Madison River southwest of the mouth of the Madison River canyon indicates an impressive amount of geologically recent tilting in the part of the valley affected by the 1959 deformation (figs. 47, 48). The younger flood plains are cut in unconsolidated fan gravel that is at least in large part of late Pleistocene age; the older flood plains cut rhyolites which are assumed to be correlative with those dated as of Oligocene age in the southern part of the Madison Range. The oldest flood plains are far southwest of the present river and enter the trough of Cliff Lake, part of a north-trending canyon whose present stream now joins the Madison River a few miles northwest of the area in the photographs. Maximum structural relief on these flood plains is about 500 feet. The several younger plains lie successively closer to the modern river. Tilting in the younger plains is difficult to define precisely as they were cut successively deeper in the fill and as they start and end near the present river; but the southernmost and oldest has been tilted northward about 100 feet, and an intermediate one has been tilted about 60 feet.

Until late Pleistocene time, the Madison River thus flowed southwest of its present course to Cliff Lake and thence northward along the course that is now a deeply entrenched canyon. Tilting toward the east-northeast diverted the river closer to its present course, but later the direction of tilting changed to the north-northeast. The modern river flows slightly north of west for nearly 5 miles after leaving the Madison River canyon, and in this distance it flows along a structural axis of late Quaternary subsidence.

HENRYS LAKE BASIN

The Madison Valley is separated from the drainage of the Snake River by low Reynolds Pass. Just southeast of the pass is a broad valley, clearly the southern extension of the Madison Valley structural depression, which trends south-southeastward for about 20 miles and merges with the northeast corner of the Snake River depression. Henrys Lake lies in a relatively depressed part of this valley directly east of the east-trending structural basin of Centennial Valley, which is described next (fig. 49).

The Centennial Mountains jut eastward into the valley south of Henrys Lake and are bounded on the north by faults which lose structural relief eastward. One of the prominent north-facing scarps in surficial deposits dams Henrys Lake. These young faults trend northeastward toward the low transverse valley of Targhee Pass, but do not enter it; at the front of the Madison Range the structures turn abruptly to form low discontinuous scarps that trend northwestward. The latest episode of faulting has thus been limited to the depression of a block at the intersection of the Centennial Valley and Madison Valley subsidence trends. As other young scarps, downfaulted on their north sides during late Quaternary time, trend toward Targhee Pass from the West Yellowstone basin on the east, it seems likely that the underlying faults are continuous across Targhee Pass, and that the low divide is a structural depression much modified by erosion.

CENTENNIAL MOUNTAINS AND CENTENNIAL VALLEY

The Centennial Mountains are a compound tilted fault block that dips smoothly southward beneath the lavas and tuffs of the Snake River Plain, and thus they form the structural border of the plain; the north slope, facing Centennial Valley, is a series of arcuate fault scarps that rise more than 3,000 feet above the valley. Fault scarps young enough to offset glacial deposits, alluvium, and lake sediments of latest Pleistocene and Recent ages lie in many places in the valley and along the mountain front and are as high as 50 feet (fig. 49). Viewed broadly, the valley is a long east-trending sag, bounded on the south by the faulted face of the Centennial Mountains. The eastern faults are en echelon, stepping to the left. The valley formed by collapse of an area once continuous between the Centennial Mountains and the ranges to the north.

Both the Centennial and Madison fronts are about 3,000 feet high, but the Madison scarp has been very much more eroded than has the Centennial scarp. The Madison Range crest lies 1.5 to 5 miles from the fault trace at the foot of the scarp, and permanent streams have cut deeply into the block; Sheep Creek, for example, falls only 400 feet in the mile above the fault trace. The crest of the arcuate Centennial fault block south of Upper Red Rock Lake, by contrast, is only 0.8 mile to 1.5 miles horizontally distant from the fault trace 3,000 feet below, and even the stream most deeply entrenched into the block falls 1,200 feet in its lower mile. Although faulting along the Centennial and Madison fronts has overlapped in time, the Centennial structures are, viewed broadly, much younger than the Madison ones. Perhaps the Centennial structures formed chiefly in Quaternary time and the Madison ones largely in the Pliocene.

Prominent shorelines clearly recognizable on the ground and in aerial photographs show that a large

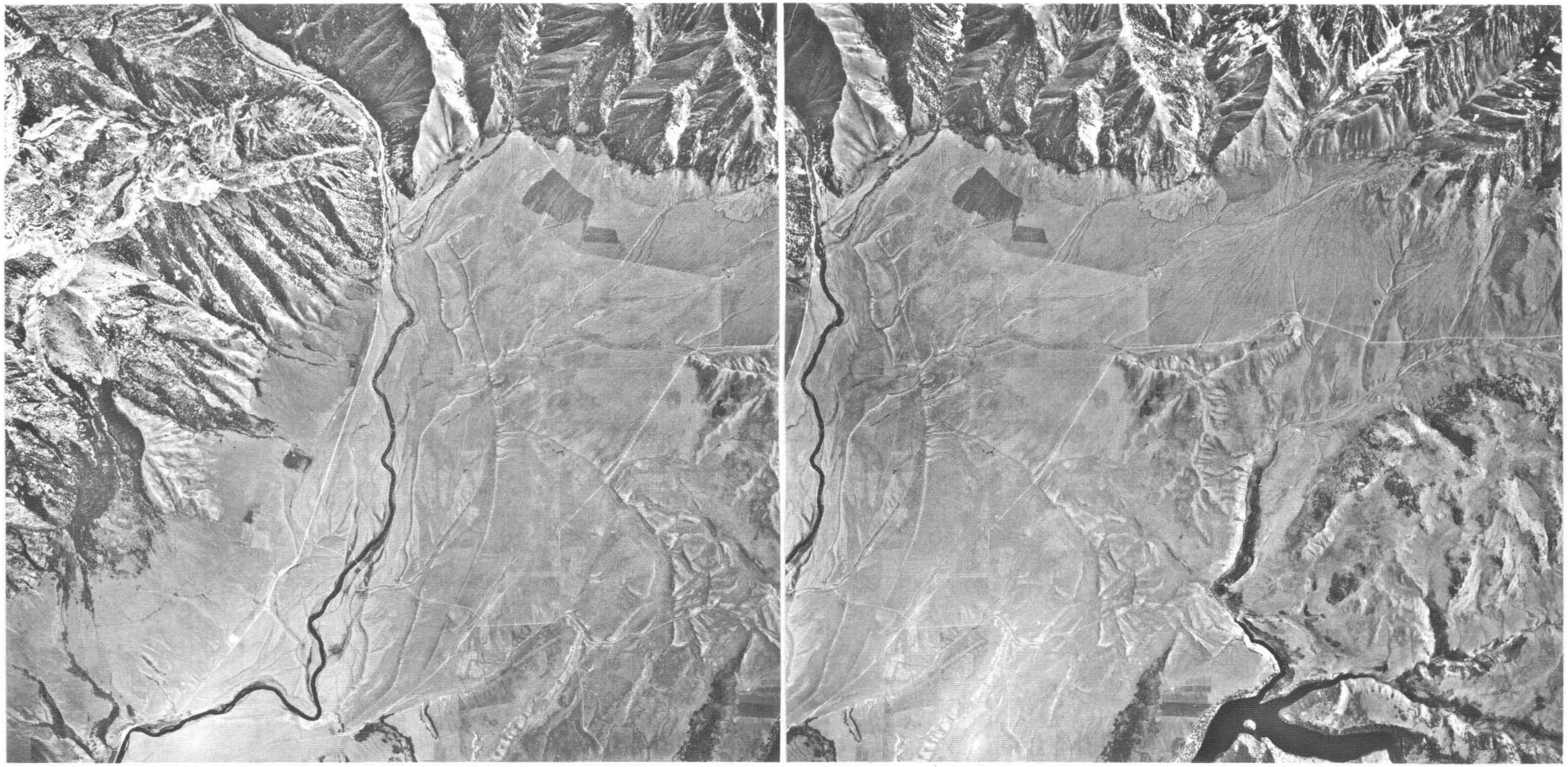

FIGURE 47.—Stereoscopic pair of aerial photographs of Missouri Flats area. See figure 48 for explanation and scale. Photographs by U.S. Army Map Service.

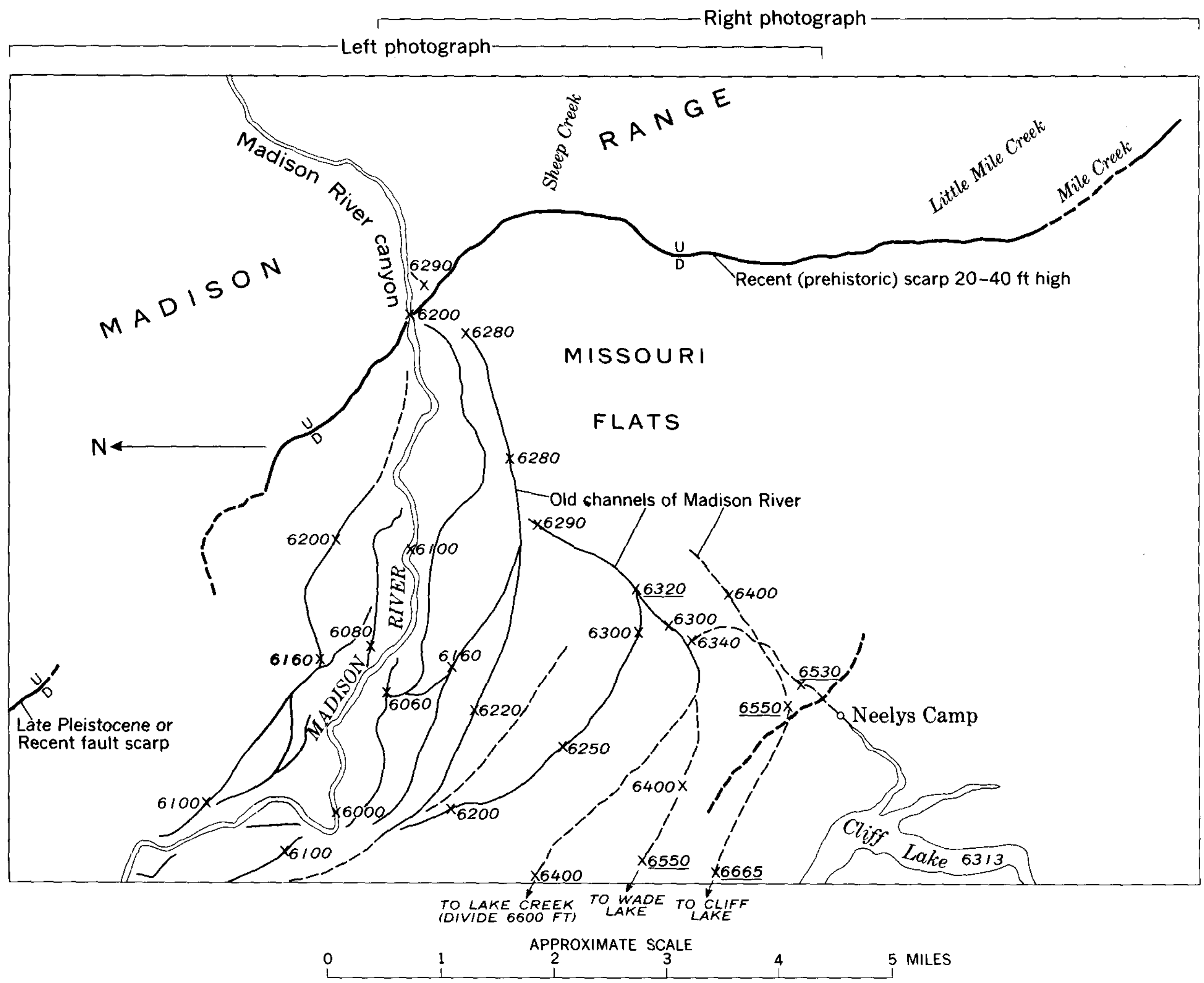

FIGURE 48.—Map of Missouri Flats area, south end of Madison Valley, Montana, showing features in aerial photographs of figure 47. Lines (dashed where uncertain) mark former courses of Madison River. Those within 2 miles of Cliff Lake have been much modified by subsequent warping, faulting, and erosion but show that the river formerly flowed through Lake Creek canyon. These old flood plains have been tilted to the east-northeast and have a structural relief of 400 or 500 feet. Younger courses are confined to the present valley of the Madison and have been tilted north-northeastward with a maximum structural relief of about 100 feet. The altitudes (marked by X's) were determined from 1 : 24,000 photogrammetric manuscript maps with 20-foot contour intervals; underlined altitudes mark drainage divides produced by deformation of initially continuous channels.

lake occupied Centennial Valley during Recent time (fig. 49). Just south of Upper Red Rock Lake the shoreline is cut into alluvial fans and talus cones derived from the youngest Pinedale moraines at the foot of the imposing mountain front. This indicates that the lake was full in post-Pleistocene time, perhaps during the postaltithermal stage of minor glaciation known elsewhere in the region and dated at about 1800 B.C. Despite its extreme youth, this lakeshore has been warped at least 60 feet and broken by faults. The closed basin filled by the lake has been nearly emptied—only the shallow Upper and Lower Red Rock Lakes remain—by this deformation and, to a lesser extent, by erosion of an outlet channel to the west. At the southwest corner of present Upper Red Rock Lake, the shoreline has a minimum altitude of 6,630 feet, but it rises to 6,690 feet in 3 miles to the east; northward from here, the shoreline descends 20 feet and then disappears where it has been cut by the flood plain of modern Red Rock Creek. North of the stream flat and Upper Red Rock Lake, the shoreline reappears and has an even altitude near 6,650 feet. Farther west, close control of shoreline altitude is lacking; but north of Lima Reservoir the lake floor is offset about 20 feet along highly irregular fault scarps. Faulting and much of the warping presumably occurred simultaneously during a major earthquake and probably emptied the lake.

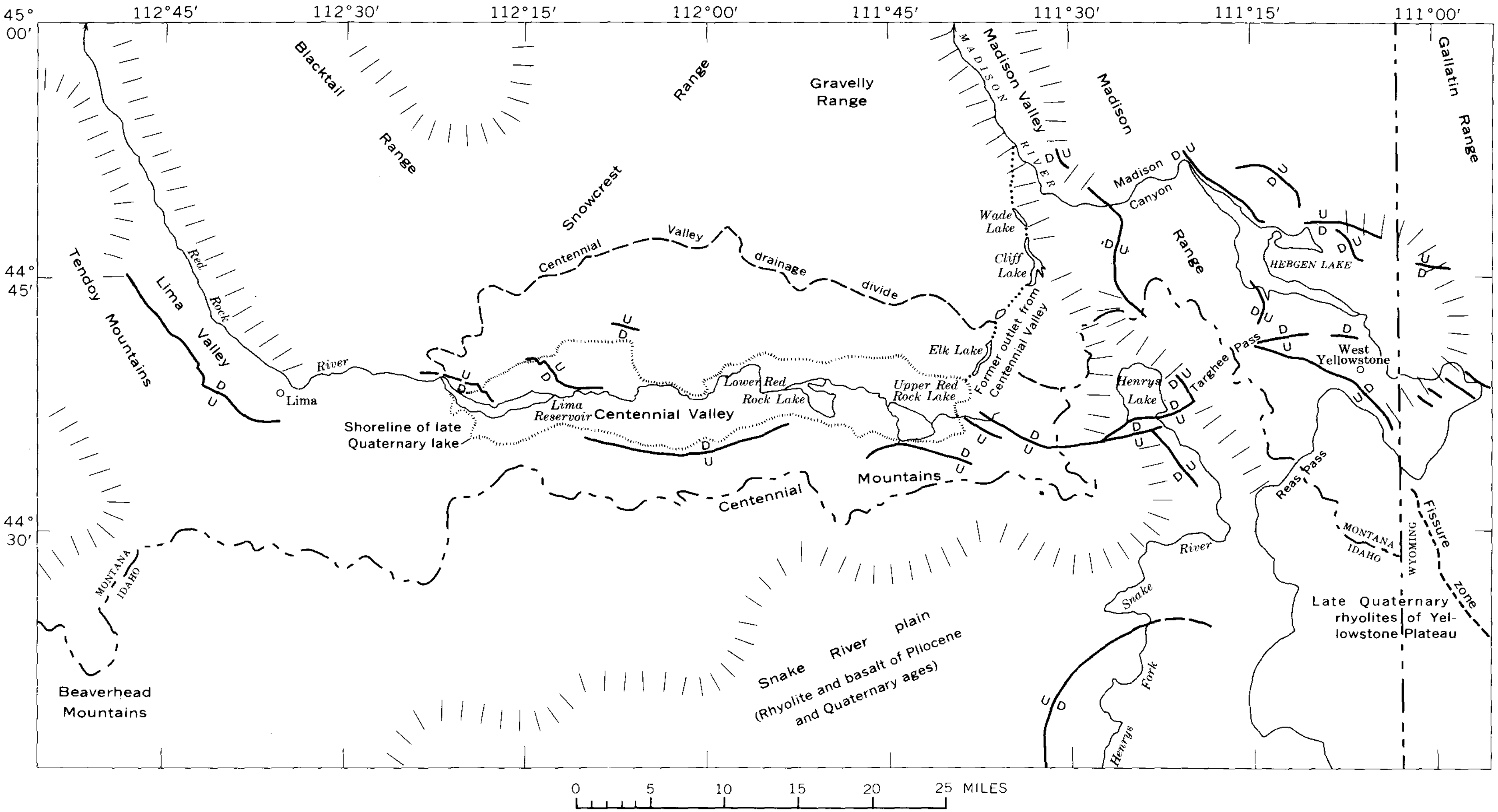

Figure 49.—Late Quaternary structure of the Centennial Valley-West Yellowstone region, Montana, Idaho, and Wyoming. Faults shown have been active during very late Quaternary time. Centennial Valley was occupied by a lake that drained to the Madison River probably until Recent time, when faulting and subsidence caused the lake to drain to the west. Centennial Valley is a structural trough bounded on the south by the composite fault block of the Centennial Mountains. Modern deformation about Henrys Lake and near West Yellowstone apparently indicates eastward extension of the Centennial Valley zone of subsidence across the older structures of Madison Valley and Madison Range. To the west, the Centennial Valley zone abuts the reactivated part of the fault along the front of the Tendoy Mountains.

Bull Lake moraines are well developed in the eastern part of Centennial Valley and indicate glaciation more extensive than that of the subsequent Pinedale stage whose moraines are nested inside them. In the central part of the valley, however, no Bull Lake moraines are exposed: although such moraines must have formed in the vicinity of Upper and Lower Red Rock Lakes about a piedmont glacier more extensive than the small Pinedale glaciers, the old moraines have been buried by lake and other sediments. The exposed alluvial fans along the mountain front are small and extend only a short distance to the edge of the lake sediments; the minimum-altitude valley flats begin as close as 0.1 mile from the trace of the mountain-front fault. These features indicate that subsidence here has been very rapid.

Until late Pleistocene or Recent time, Centennial Valley drained northward from its east end, through the canyon occupied by Cliff Lake, to the Madison River. The once continuous canyon has been segmented by large landslides, perhaps quake triggered, behind which stand the various lakes (Mansfield, 1911); the rock floor beneath the slides probably slopes continuously northward from Elk Lake, which is less than a mile north of Centennial Valley, although Elk Lake now drains into Centennial Valley because of a large landslide dam at its north end. The lake surface is only 25 or 30 feet above the nearest part of the Recent Centennial Valley shoreline; this height may represent either building up of the channel by sedimentary or slide processes or structural tilting. The Centennial Valley lake recorded by the shoreline may have drained northward through the canyon, or this drainage course may have been abandoned somewhat earlier in favor of the westward one.

SUPERPOSITION OF STRUCTURES IN THE MADISON–CENTENNIAL REGION

Although structures of different types are superimposed in parallel array in many regions, successive structures in the Madison-Centennial region have cut markedly across the older structures. This is in large part obvious even on the geologic map of Montana. Madison Valley lies along the core of Precambrian crystalline rocks of a broad Laramide uplift; although oblique to the Laramide structures at only a low angle, the modern valley is where the Laramide range crest once lay. The Centennial Mountains and Valley cut sharply across a northeast-trending Laramide basin in which are preserved high Upper Cretaceous rocks in the west and the Precambrian core of a Laramide uplift—the same one that is dropped down by Madison Valley—in the east. Both Madison and Centennial Valleys truncate a fault basin too old to have much influence on present topography: a synclinal fault trough trends northeastward in the central part of the Centennial Mountains, disappears beneath Centennial Valley, reappears in the Gravelly Range, and is again cut off at an angle of 45° by Madison Valley. This old trough cuts sharply across Laramide structures and drops down volcanic rocks of probable Oligocene age. The deformation of the southern part of Madison Valley by structures trending eastward from the Centennial area was described previously.

In the Madison-Centennial region, Laramide structures that trend broadly northward thus were truncated in Miocene(?) time by a northeast-trending graben. The graben was deformed by north-northwest trending Madison Valley (largely during the Pliocene?), and both the old graben and Madison Valley have been deformed (mostly during the Quaternary?) by east-trending Centennial structures.

SEISMIC EVIDENCE OF MOVEMENT PATTERN

A plotting of recorded earthquake epicenters in the northern Rocky Mountain area (Ross and Nelson, chapter E, fig. 14) emphasizes a major seismic zone that trends northward through Yellowstone National Park. A striking detail of the pattern is the lack of recorded activity in the upper Snake River Plain, a major structural feature that was active in late Tertiary and Quaternary time.

North of the Snake River Plain the plotted epicenter of a major earthquake on November 23, 1947, lies about 40 miles west of the north-trending seismic zone through Yellowstone and 10 miles north of the axis of Centennial Valley. This epicenter (44°47′ N., 112°02′ W.), at the drainage divide between the Centennial basin and the region to the north, may record activity along the east-trending Centennial subsidence zone. No other recorded epicenters plotted by Ross and Nelson seem to be related to this trend.

The epicenters of the Hebgen earthquake and the aftershocks that recurred over a period of 7 weeks (Murphy and Brazee, chapter C, fig. 11) constitute an east-trending array that is twice as long as it is wide. Omitting the uncomputed ones, the computed epicenters form an east-trending group that is four times as long as it is wide and which lies generally north of the West Yellowstone basin, just as the epicenter of the 1947 quake is north of the Centennial basin.

A different pattern of epicenters was determined for a series of aftershocks recorded within the space of a few days by Stewart, Hoffman, and Diment (chapter D, fig. 12). The epicenters fall in a broad arc whose oblique inflection lies a few miles northeast of West Yellowstone. The northern arm of the arc trends about N. 80° W. across the zone of main northwest-trending reactivated fault scarps; the other arm trends southward along the Wyoming State line. The lack of agreement between this pattern and that found by Murphy and Brazee may be due to the shortness of the Geological Survey record.

A first-motion study, based on P waves as recorded at 72 stations, was made for the Hebgen Lake earthquake by Ryall (1962), who used the method of Byerly. Ryall's calculations indicate the initial slipping to have been on a fault plane striking N. 80° ± 10° W., dipping 54° ± 8° S.W., and with dip-slip movement. This calculated strike lies far to the west of the dominant northwestward trend of the major faults reactivated to the surface and is consistent with the interpretations made in this paper.

GEODETIC EVIDENCE OF ACTIVE ZONES

The level line run by the Coast and Geodetic Survey from West Yellowstone to Bozeman in 1934 and rerun after the 1959 earthquake is but a part of a more extensive network of levels. Third-order level lines were run by the Topographic Division of the Geological Survey in 1948 between the older first- and second-order levels. There are systematic large closing errors in these third-order lines across the Centennial basin which strongly suggest that the basin subsided about half a foot between 1934 and 1948, perhaps mostly during the earthquake of November 23, 1947.

By contrast, Coast and Geodetic Survey line 48, in Madison Valley north of the area that subsided during the 1959 earthquake, remained almost stable between 1934 and 1959.

CONCLUSIONS

The 1959 basin of new subsidence has an overall trend of about N. 80° W. and is superimposed at a high angle across the major structures of the Madison Range, which trend north or northwest. A similar crossing of old trends by new is shown by the pattern of the intense late Quaternary deformation of the region. Madison Valley south of the mouth of the Madison River canyon is being deformed by structures that strike east from the Gravelly Range, Centennial Valley, and Centennial Mountains; and other active structures strike east from the West Yellowstone basin far into the Yellowstone Plateau. The Centennial structures, which strike eastward, are largely younger than the north-northwest striking Madison structures.

The broad aspects of deformation during the earthquake, considered with the pattern of Quaternary deformation of the region, suggest that the Centennial structural system is now being extended eastward across the Madison structural system and into the Yellowstone Plateau. Although the large new scarps northeast of Hebgen Lake utilize older Madison structures oriented northwestward, the subsidence breaks across such individual structures to form a composite basin with an eastward trend. It seems that the structural pattern now in early stages of development in the northwestern part of the Yellowstone region will result ultimately in the formation of major east-trending structures, similar to but east of the present Centennial Mountains and Centennial Valley, and in the progressive obscuration of the Madison Valley and Madison Range. Reactivation of preexisting structures with other trends is to be expected in this process.

The subsidence during the earthquake of 1959 increased the relief of structures clearly belonging to the Madison Range system. Both Madison Valley and the West Yellowstone basin were depressed relative to the Madison Range, but the intervening part of the range subsided also, at least near Hebgen Lake and in the Madison River canyon. The long axis of the compound basin of new subsidence is thus defined, independent of the incompletely known behavior of the Madison Range, by the proved subsidence of Madison Valley, the Madison River canyon, and West Yellowstone basin, and the Gibbon River valley to have a general easterly trend across the Madison structures.

The general high regional altitude suggests that the Madison-Centennial region is being broadly elevated. Upon this presumed regional elevation is superimposed the more local subsidence to which are due the downbowed and downfaulted basins. The character of the late Cenozoic structures—normal faults and gentle warps—suggests regional extension that permits the slumping of keystone blocks. Probably the region is being slowly elevated, perhaps without earthquakes, at the same time that it is being stretched; the great earthquakes may accompany the settling of local blocks rather than the uplift of mountains. The warping and faulting of the Recent lake basin in Centennial Valley, the formation of the high Recent scarps along the fronts of the Tendoy and Madison Ranges, and the landslides damming Cliff Lake and the other lakes in its chain may all have occurred simultaneously in an extremely severe earthquake.

The known deformation accompanying the earthquake consisted largely of downward movement. From this we conclude that the immediate cause of the earthquake was collapse in response to gravity, regardless of the ultimate tectonic, magmatic, or other reasons for that collapse. The surface effects can be viewed as due to slumps of various scales and at different depths, variously interacting and interfering, related partly to near-surface directions of easy slipping and partly to the motion of deeper material.

The tight control of the positions of two of the major 1959 faults by the orientation of strata in old structures of shallow extent shows that these faults are in effect giant slump scarps. The utilization of any specific preexisting structure is therefore incidental and dependent on a more fundamental control of the zone of subsidence. As these near-surface faults bound a very large basin of new subsidence, we regard them as passive products of the general collapse rather than as direct products of deep tectonic stresses. The abruptness of the warps and scarps that bound the 1959 basin of subsidence on the north and the termination northward of these northern structures along a vague line trending near N. 80° W., in contrast to the broad gentle south flank of the basin, lead us to infer that the deep structure controlling the surficial subsidence is either a normal fault striking about N. 80° W. and dipping southward or an abrupt monocline with the same orientation.

The earthquake might have been due to fault movement regardless of whether such movement was fundamental to the subsidence or not; but a minority of American geophysicists, and a larger minority in other countries, no longer support the common assumption that simple fault motion causes all major earthquakes, and it does appear that the kinetic energy of the subsiding masses about Hebgen Lake is adequate to explain the earthquake in other terms.

The 1959 collapse has the shape of a broad but markedly asymmetric basin warped gently downward on the south flank and bounded by abrupt structures on the north. We presume that the fundamental deep structure is similarly one sided. The collapse could have been caused by withdrawal of either horizontal or vertical support from beneath the basin; regardless, presumably slow motion over a long period produced stresses which were released suddenly during the earthquake.

The intensity of late Quaternary volcanism in the Yellowstone Plateau and upper Snake River Plain, where individual eruptions of rhyolite have volumes as large as 5 cubic miles, suggests a possible causal relationship between that volcanism and the concurrent deformation of the adjacent Madison-Centennial region. The geometric ties between the volcanic province and the flanking highlands are so numerous that the stress systems controlling them must be closely related. The northeastern part of the Snake River Plain—which includes the Yellowstone Plateau as its high lava-filled end—is rimmed almost continuously by high mountain blocks that dip inward beneath it. The depression trends northeastward at a high angle across both Laramide structures and most of the Tertiary block-fault structures, including the Madison Range, yet is not itself deformed by structures of such trends (Hamilton, 1960). The Centennial Mountains are subparallel to the depression at its edge, and on the opposite side of the depression the Teton Range emerges to complete a pattern of remarkable dynamic symmetry. Jackson Hole, the structural counterpart of Centennial Valley, has also undergone extraordinary tectonic activity during the Quaternary and also cuts sharply across older structures. The intervening Snake River–Yellowstone province has been the site of intense basaltic and rhyolitic volcanism throughout Pliocene, Pleistocene, and Recent time. The regional pattern suggests that southwestern Montana is now drifting relatively northwestward; that the Snake River–Yellowstone depression is a rift zone with a tensionally thinned crust and an abnormally high temperature gradient; and that the Centennial family of structures (and hence the earthquake of 1959) is a product of this horizontal extension.

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A Geological Interpretation of the Epicentral Area—the Dual-basin Concept

By GEORGE D. FRASER, IRVING J. WITKIND, *and* WILLIS H. NELSON

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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A GEOLOGICAL INTERPRETATION OF THE EPICENTRAL AREA—THE DUAL-BASIN CONCEPT

By George D. Fraser, Irving J. Witkind, and Willis H. Nelson

ABSTRACT

During the Hebgen Lake earthquake, the Red Canyon, Hebgen, and Madison Range faults were reactivated, and two basins (Hebgen Lake and Missouri Flats), bounded in part by one or more of these faults, were differentially dropped. These basins trend northwest and flank the core of the Madison Range, which remained relatively stable.

The relief of the Madison Range, as measured from the Hebgen Lake basin across the range to the Missouri Flats basin, has increased. It is suggested that the principal controlling structures for this change in relief are deep-seated earthquake-generating faults which either extend directly to the surface or are represented there by derivative faults.

The basins are modified by warps parallel and transverse to the faults and areally associated with them. The warps die out gradually away from the faults. This episode of basin deformation continues the late Cenozoic deformation which must have included many preceding earthquakes.

Compaction and slump of surficial deposits have masked the true amount of bedrock subsidence. As a result, isobase maps show the deformational pattern of the surface rather than the pattern of deformed bedrock. Inherent uncertainties in the data, the absence of eastward-trending fractures in the Madison Range, and the reactivation of northwest-trending faults all cast doubt on the propagation of an eastward-trending syncline across the Madison Range as suggested by the single-basin concept.

INTRODUCTION

Two different interpretations of the deformation in the epicentral area have arisen from the study of the Hebgen Lake earthquake. One, presented by Myers and Hamilton in chapter I, may be referred to as the single-basin concept. It emphasizes the importance of warping of the lake basin and suggests that the faults north of the lake are secondary features controlled by local structure and do not necessarily reflect a deeper, more fundamental fault displacement. In this view, a single basin of subsidence extends from Hebgen Lake across the Madison Range and the southern part of the Madison Valley, and part of the range has subsided along with the bordering basins (pl. 2). The warping is seen as part of a very young and genetically different structural system that seems to be disrupting the older north- and northwest-trending structural elements. This system includes the eastward trending Centennial Valley and Centennial Mountains as well as the Snake River depression and part of Yellowstone National Park.

Another interpretation, implicit in the preceding chapters by Witkind, Fraser, and Nelson, may be called the fault-controlled dual-basin concept. It emphasizes a close relation between the surface deformation and a fundamental fault, or faults, movement along which gave rise to the 1959 earthquake. A corollary of this view is that the deformation was restricted to two fault-controlled basins on either side of the Madison Range, which itself was not greatly affected (fig. 50). The discernible structural pattern is in no sense unique but merely a duplicate of patterns well known in western Montana where fault blocks have been repeatedly dropped and tilted in fashioning and maintaining the intermontane basins.

STRUCTURAL SETTING

The structural fabric of western Montana was determined to a large extent during the Laramide orogeny, and all subsequent deformation has been guided by these structures. Pardee (1950) notes a regional parallelism between Laramide and late Cenozoic structures but points out a fundamental difference between the two: the Laramide structures are typically tight folds broken by thrust faults; the late Cenozoic structures are characteristically normal faults and gentle warps. This parallelism is apparent on the geologic map of Montana (Ross, Andrews, and Witkind, 1955); the trend of the active Madison Range fault clearly parallels the Laramide thrust and fold axes in the southern Madison Range. Similarly, in the Hebgen Lake earthquake area the Red Canyon and Hebgen faults (probably formed during the late Cenozoic) parallel and appear to be controlled by Laramide structures. This definite relationship

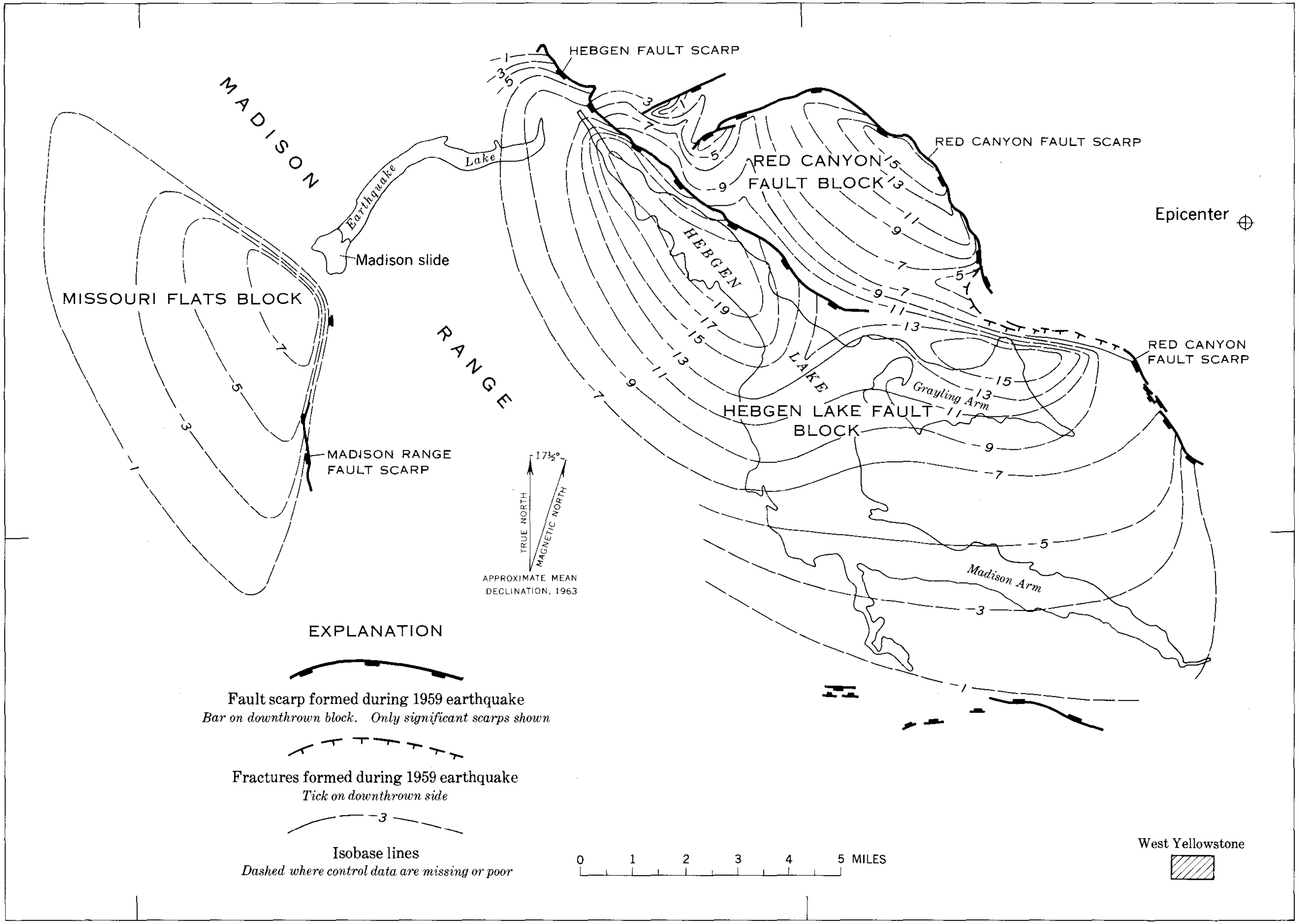


Figure 50.—Isobase map illustrating the dual-basin, fault-controlled pattern of deformation.

with the geologic past is demonstrated by Witkind (chapter G), by Myers and Hamilton (chapter I), and by comparison of plates 2 and 5.

REACTIVATED FAULTS NORTH OF HEBGEN LAKE

The newly formed fault scarps coincide with, or are closely parallel to, the preexistent Red Canyon and Hebgen faults and thus imply that these faults were reactivated during the 1959 earthquake. A basic assumption is that the strike and direction of movement on these faults are related to a deep fault, or faults, required by the elastic-rebound theory or other theories of seismic faulting (Orowan, 1960; Griggs and Handin, 1960). The exposed faults are continuous enough to be considered as surface expressions of deep-seated causative structures. The Red Canyon and Hebgen fault scarps could be surface manifestations of a single fault which split near the surface.

This concept may be supported by recent seismological studies. Ryall (1961) states:

> An attempt to obtain the direction of faulting by the method of P. Byerly yielded incomplete results because of a lack of data from the region north of the epicenter. From 72 observations of the first motion of P, a single plane was defined which had strike N. 80° W. ±10° and dip 54° SW. ±8°. It was not possible to ascertain, however, whether this plane corresponded to the faulting observed in the field (dipole source) or whether it represented the auxiliary plane of a simple force at depth.

The long history of repeated activity on the extensive Red Canyon and Hebgen faults suggests that they are major features of deep-seated origin. To elucidate this history, we will trace the development of a representative fragment of the area—Kirkwood Ridge (pl. 5).

The first stage in the structural growth of Kirkwood Ridge was one of compression in which the rocks were folded, overturned to the northeast, and finally broken by the Divide thrust fault, which now emerges along the northeast edge of the ridge (fig. 51*A*). These events happened in Late Cretaceous to early Tertiary time, during the Laramide orogeny.

The next significant structural stage was regional uplift accompanied by local sagging or uneven subsidence of strips along normal faults. This style of deformation began in the Tertiary and has continued intermittently ever since. Presumably each episode of faulting produced no more than a few feet or tens of feet of structural relief. The trend of the marginal normal faults and the axes of the downwarps were guided by the old Laramide structures. The Red Canyon fault, which outlines the south edge of Kirkwood Ridge, belongs to this younger generation of faults (fig. 51*B*).

In the structural development of Kirkwood Ridge the upright (southwest) limb of an overturned fold was gradually let down in successive small steps, very likely accompanied by earthquakes. The upright limb is now offset from its former position by 1,000 to 2,000 feet (fig. 51*C*), which suggests that it has moved scores, perhaps hundreds of times. In this light, the earthquake of August 17 is merely one episode in a recurrent structural story.

Hebgen Ridge, along the northeast shore of Hebgen Lake, has been formed in the same manner; the northeast edge is bounded by the Wells thrust fault, and the southwest edge is bounded by the Hebgen normal fault (pl. 5). Between these faults the strata are vertical or overturned, as on Kirkwood Ridge.

The impressive heights of both ridges plus the great apparent throw suggest that these normal faults have been recurrently active for a long time.

Folding proxies for faulting in a few areas along and adjacent to the principal scarps. The deformed rocks were offset in many places by movement along old faults, and in such places a distinct scarp broke through the overlying surficial material. In other places the rocks presumably yielded by flexing, the healed fault was bent along with the bowed rocks, and no scarps formed. On the surface these flexed localities are marked by gaping fractures and small breaks which mirror the underlying monoclinal fold.

SIGNIFICANCE OF DATA

Our isobase map (fig. 50) is based chiefly on three sources of data: (1) preearthquake and postearthquake surveys completed by the U.S. Coast and Geodetic Survey, (2) measured changes in the old and new shorelines of Hebgen Lake, and (3) throw along the newly formed fault scarps.

All these sources indicate that a very large area subsided. In places the subsidence was uniform, as from West Yellowstone to the wye formed by the junction of Highways 191 and 499. Elsewhere, as along the northeast shore of Hebgen Lake, subsidence was erratic.

This erratic subsidence may have resulted in some measure from differential compaction of surficial material. In chapter I and in plate 2 the subsidence shown by the bench marks and the road profiles has been interpreted as the uneven warping of downthrown bedrock. We suggest that differential compaction and slumping of the unconsolidated material in which nearly all the bench marks are emplaced and on which all the roads are constructed, may account for some part of the subsidence. What is shown is the configuration of the ground surface, not necessarily the

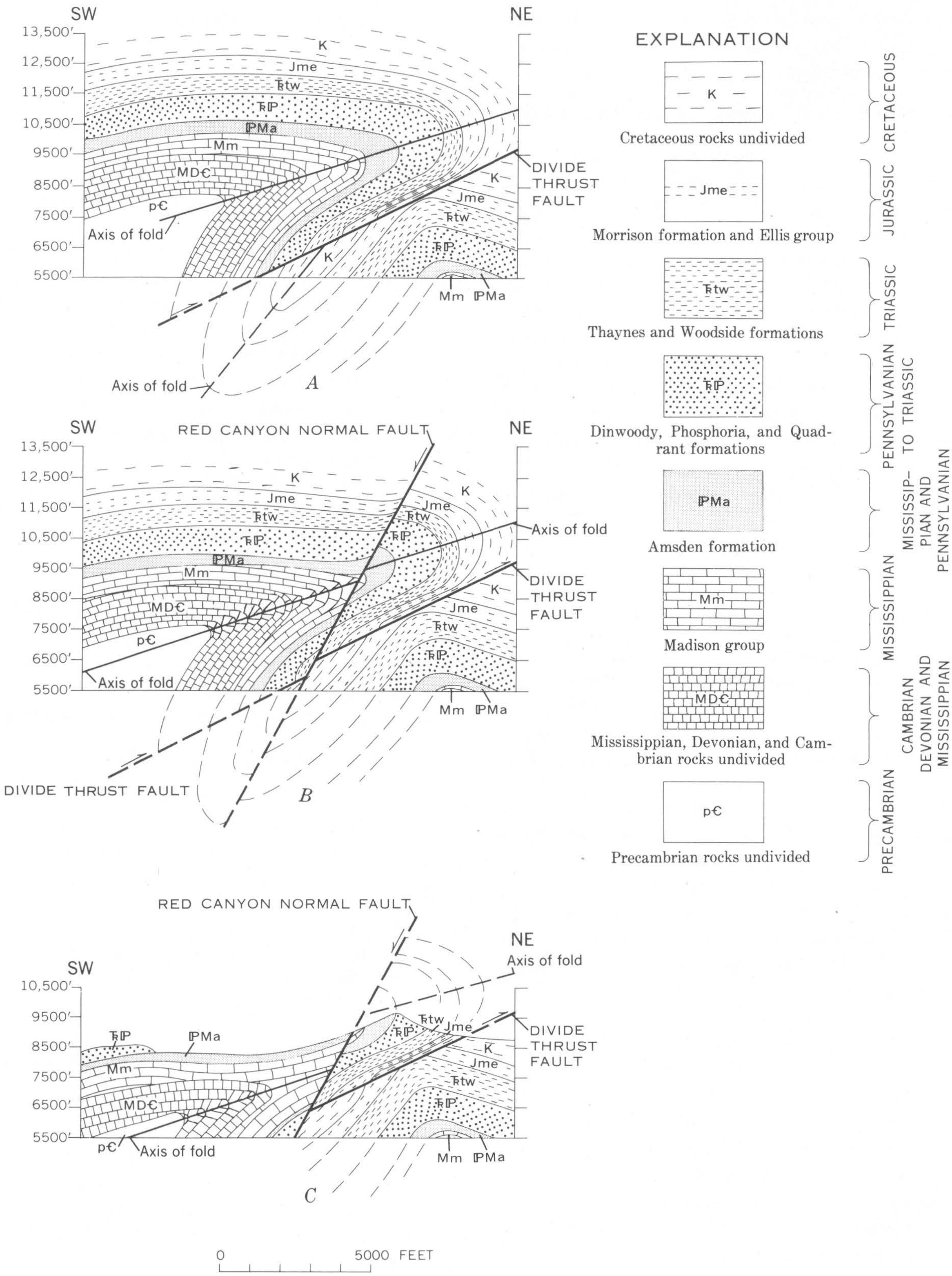


FIGURE 51.—Three sections to illustrate development of Kirkwood Ridge. *A*, As a result of compressive forces, rocks are folded, then turned over, and finally broken along the Divide thrust fault; *B*, during a second episode of deformation the upright limb of the fold is let down along the Red Canyon normal fault, which parallels preexisting Laramide structures; *C*, continued intermittent lowering of the upright limb coupled with unceasing erosion results in the present Kirkwood Ridge—bounded on the northeast flank by a thrust fault and on the southwest flank by a normal fault. The Red Canyon fault scarp coincides with or is closely parallel to the Red Canyon fault.

configuration of the concealed bedrock. Without question, bedrock has subsided, as has been demonstrated in three bedrock areas around the lake. But compaction of surficial deposits has also occurred, as attested by the hundreds of sand spouts along the northeast lake shore. Along highways, compaction of surficial material showed as disrupted pavement, transverse cracks, and uneven sags. Near bridges, the highways locally dropped 1 to 2 feet and gave the erroneous impression that the bridge had been raised. Of the 53 significant bench marks from West Yellowstone to Beaver Creek (via Highway 191 and Route 499), 52 are on alluvium, colluvium, or glacial drift. As each material responds differently to compaction, differential changes can be expected from one material to another. This is suggested by the string of bench marks north of Grayling Arm (pl. 2). Bench marks on both the Grayling and Red Canyon alluvial fans subsided as much as 15 feet. By contrast, the bench marks on the glacial drift at Corey Springs subsided but 7 feet. One possible interpretation of these data, then, is that the alluvium compacted more than the glacial drift; how much either material compacted at any one place is unknown.

Compaction also played a role in the subsidence near Beaver Creek, where Montana State Route 499 crosses a broad basin about 2½ miles wide filled with alluvium and till. Sand spouts are widespread, and the highway was broken in many places by transverse cracks which ranged in width from 3 inches to 1½ feet (L. D. Tingey, written communication, 1961). The highway near the bridge sagged, and the bridge approaches were disrupted and were about 6 inches lower than the road surface across the bridge. The bridge showed evidence of southward movement and, according to a bench mark on one of its abutments, had subsided about 8 feet. These features are indicative of compaction of surficial material; therefore, not all of the 8-foot drop at Beaver Creek bridge is attributable to true bedrock subsidence.

The pattern of subsidence near Beaver Creek is interpreted by Myers and Hamilton (chapter I) as due entirely to bedrock subsidence. Their comparison of road profiles[4] indicates even subsidence of the highway from Beaver Creek bridge to the edge of Earthquake Lake, at which point the highway is 14 feet lower than its preearthquake altitude (pl. 2). From this they conclude that the bedrock in the Madison River canyon has subsided a like amount. But later surveys along this same sector by the Bureau of Public Roads fail to repeat this evenness. Comparisons of these profiles, by contrast, show irregularities and some abrupt changes. Along 300 feet of the highway 2900 feet southwest of Beaver Creek, there was as much as 9 feet of new subsidence between September 1959 and the spring of 1960, which brought the total subsidence to a maximum of 20 feet. Along part of the steep colluvial shore of Earthquake Lake, Bureau of Public Roads surveys record continuing changes in altitude of 1.0 to 1.8 feet. These changes, possibly due to differential compaction or slump of surficial material, indicate the inherent instability of the surficial deposits. They are cited here in support of the view that the data may contain intrinsic uncertainties. What is shown in the Beaver Creek area is the pattern of subsidence of the floor of the Madison River canyon, not necessarily the pattern of deformed bedrock in the canyon or the deformation pattern of the crest of the Madison Range.

[4] One profile was completed in 1932–33 by the U.S. Bureau of Public Roads, and the other was run in late September 1959 by the U.S. Coast and Geodetic Survey.

The measurements completed around the lake comparing preearthquake and postearthquake shorelines may also include surface changes which, to some extent, mask the pattern of the deformed bedrock. Of 60 miles of shoreline, only about 2 miles are bedrock, and this is chiefly in three small areas near the northwest arm of Hebgen Lake. Surficial deposits thus make up about 97 percent of the shoreline. These, at least locally, have been subjected to some compaction and slumping; for in 1960, when the lake was drawn down to permit repairs to Hebgen Dam, the exposed near-shore mud flats were covered locally by a multitude of sand spouts.

From the above it seems that the data are not adequate for the preparation of a precise isobase map showing the definite pattern of the warped bedrock. An isobase is defined as a line of equal deformation, but in an area dominated by surficial deposits, an unknown amount of compaction is included in most control points. Unless some method can be devised to determine how much the bedrock was warped, the true deformational pattern resulting from the 1959 earthquake may never be known.

PATTERN OF DEFORMATION

Our isobase map (fig. 50) is based on data from several sources but disregards other data, chiefly from the Madison River canyon, which we consider to be equivocal or to contain inherent uncertainties. We interpret the data to show a stable core for the Madison Range flanked by two basins, one east of the range, and the other west of it. Both basins subsided unevenly during the earthquake.

The eastern basin consists of the Red Canyon and Hebgen Lake blocks (fig. 50). The Red Canyon block is well defined by marginal faults, but the limits of the Hebgen Lake block are less certain. The southeastern part of the block, northwest of West Yellowstone, may actually be a shallow asymmetrical graben with the largest displacement on the north.

The western basin, a much larger one with northerly trend, coincides roughly with the Madison Valley. Part of this basin is underlain by a third dropped and tilted block which underlies the Missouri Flats of the Madison Valley, directly west of the Madison Range and south of the Madison River canyon.

The northeast limit of significant subsidence seems clearly to have been the Red Canyon and related fault scarps, and the southeast limit was somewhere north of West Yellowstone; the west limits are uncertain (fig. 50). Subsidence of Hebgen Lake, of the Red Canyon fault block, and of Missouri Flats is demonstrated by a combination of geodetic and geologic evidence. Data are lacking or equivocal on how the intervening Madison Range behaved. This gap in information permits differing interpretations of the shape of the warping and of the relation of warping to faulting.

In the single-basin concept the important structure, a syncline, is transverse to other structures in the Madison Range; the core of the Madison Range has been bowed down along an easterly and northeasterly trend that coincides with the Madison River canyon (pl. 2). Extension of the transverse synclinal axis as far west as Cliff and Wade Lakes is part of the single-basin concept.

In the dual-basin concept the important structures are faults which combine to accentuate separate basins of subsidence parallel to the faults; the core of the Madison Range has remained relatively stable. Any deformation in the Cliff and Wade Lakes area is part of the western basin and reflects the Missouri Flats structure.

The pattern of deformation shown near Hebgen Lake and in the Missouri Flats area is to us an uncertain reflection of the subsided basins. In the Madison River canyon, the data—all from the detritus-filled canyon floor—show changes that, we believe, do not accurately mirror the changes in the underlying bedrock or the crest of the Madison Range (p. 103).

No matter what shape is visualized for the warping, the relief of the Madison Range has increased between Hebgen Lake and the reactivated segment of the Madison Range fault. The Hebgen Lake basin sagged along a fault zone, and Missouri Flats was relatively downthrown along the reactivated segment of the Madison Range fault. Thus, a northwest-trending structural and topographic high that existed before the earthquake has been accentuated locally by the earthquake deformation. In a very general sense, the Madison Range south of the Madison River canyon has been titled to the northeast; but more precisely, it has been warped downward along the Hebgen fault zone and (at least relatively) upthrown along the reactivated segment of the Madison Range fault.

RELATION BETWEEN WARPING AND FAULTING

In the dual-basin concept, subsidence is viewed as concentrated along the faults and within the two principal basins. This is consistent with observed earthquake deformation patterns in most seismic areas and with the fault theories of earthquake mechanism. Nearly all detailed studies of deformation associated with earthquake faults which break the surface show that distortion of the surrounding rocks dies out systematically a few miles away from the fault (Richter, 1958).

Both isobase maps clearly show maximum deformation near the faults. The Hebgen–Red Canyon fault zone was activated for about 15 miles, and the most extreme deformation is centered along this zone. A decrease in the amount of sag is apparent toward the ends of this fault zone and perpendicularly away from it. In detail these changes can be defined by synclinal axes which trend in two different ways—parallel and transverse to the fault.

An interpretation of the origin of warping as a consequence of faulting is presented in figure 52. The figure shows an idealized "Hebgen fault," with maximum subsidence midway along its trace and zero displacement at either end. Parallel to the fault, a canoe-shaped syncline develops as the result of differential subsidence combined with upward drag. As the ground is not displaced at either end of the fault, the central part of the subsided block sags as much as the throw on the fault, thus leading to a transverse syncline plunging at a high angle into the sag in the canoe-shaped syncline.

The actual situation is more complex because there is a composite fault zone, the traces curve, there are two sag points instead of one, and one sag may result largely from adjustment of surficial deposits (p. 101–103).

The transverse syncline postulated as a major structural element in the single-basin concept (pl. 2) is interpreted in figure 50 as a minor structure with little genetic significance. The equally long but better defined and more sharply flexed synclinal complex which parallels the fault zone on both isobase maps seems more significant, not only because it is more

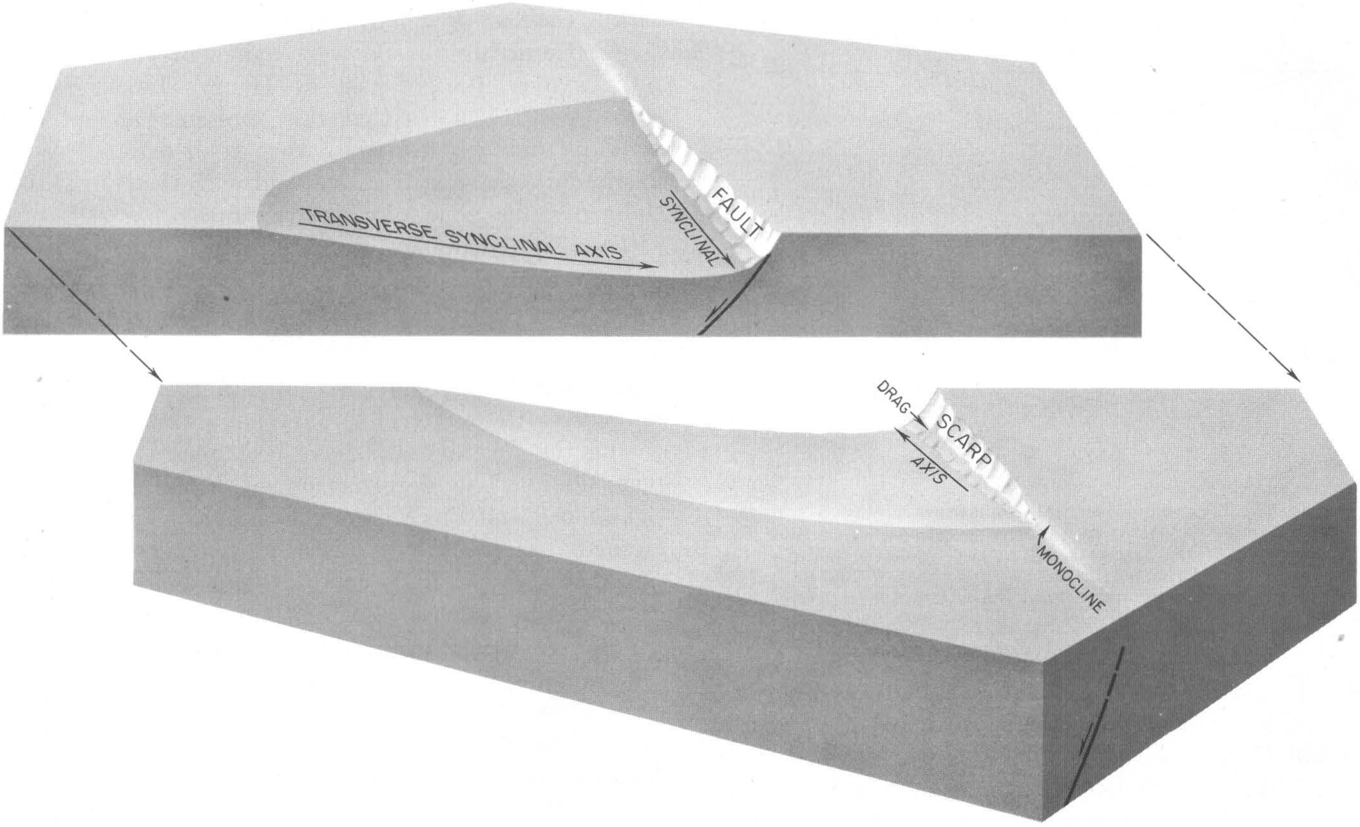


FIGURE 52.—Block diagram of idealized normal fault showing warping as a consequence of faulting. Major displacement is near the midpoint of the fault scarp. Ends of the fault scarp pass into monoclinal folds. The subsided block is marked by two synclinal axes: one parallel to, and the other transverse to the fault scarp.

impressive, but because it is clearly related to the northwest-trending fault zone that accounts for nearly all the subsidence in the area of greatest deformation.

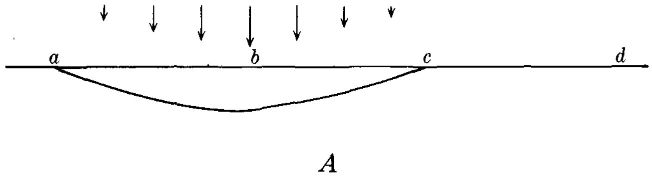
If a syncline formed in the Madison River canyon, its postulated dip was about 2 feet per mile (pl. 2). Such structures, which cannot be mapped by ordinary geologic techniques, probably do not develop into basin-forming structures for reasons stated below. Nor can such structures be construed as disrupting the Madison Range unless it can be shown that similar structures, within and across the range, are not normal parts of any fault-bounded range.

A mountain-front fault like the Madison Range fault probably is not active along its entire length at any one time. Rather it moves first in one place and then another, with a maximum localized throw of a few tens of feet during the major episodes. The end result of the countless separate episodes of faulting is a tilted mountain range separated from an adjoining basin by an impressive fault zone. During this process the transverse warps that form during each episode of faulting interfere with each other. The limbs of transverse folds, side by side along successively reactivated segments of a long fault, tend to cancel out (fig. 53), and the main indication of the composite transverse syncline may be the lip at either end of an elongate basin.

This fault-controlled warp pattern seems well satisfied by the two basins. The eastern basin has tilted northeastward along the Red Canyon–Hebgen fault complex. The Missouri Flats basin has tilted eastward along the Madison Range fault.

DELINEATION OF NORTHWEST-TRENDING BASINS

Western Montana has probably been undergoing intermittent uplift since the Laramide orogeny (Pardee, 1950, p. 403). As a result, segments of the stretched superficial layer of the crust have foundered and formed the intermontane basins that are spread throughout western Montana. All subsequent structures have been guided by the general north and northwesterly trend of these basins, and recurrent structural activity has emphasized the established pattern (p. 99). These basins, probably first well defined in the early Tertiary, thus have been more clearly delineated since then by repeated subsidence. We view



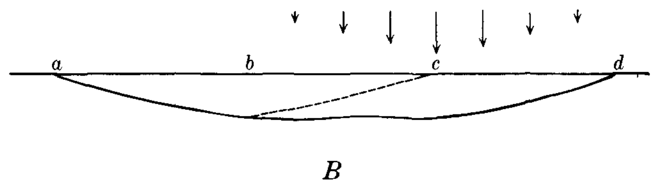


FIGURE 53.—Diagram of vertical displacement below a reference horizon *a b c d* due to two successive episodes of faulting. *A*, The result of the first episode; *B*, the combined results of the two episodes. The lengths of the arrows show the amount of vertical displacement. Synclines shown are the transverse type illustrated in figure 52.

the Hebgen Lake earthquake as merely another episode in a well-established pattern of basin formation. In the Hebgen Lake area, the basins on either side of the Madison Range have been accentuated during the earthquake; as a result, the relief of the Madison Range increased.

The evident late Cenozoic structural unrest of east-trending Centennial Valley (fig. 49) has been cited by Myers and Hamilton (chapter I, p. 96) as regional evidence of a new cross structure being propagated across the Madison Range. Such propagation seems doubtful. The faults that outline Centennial Valley diminish in displacement toward the east and seem to disappear near Henrys Lake; as far as is known, they fail to penetrate the Madison Range. No mention is made of east-trending faults by Freeman, Sweet, and Tillman,[5] who mapped the south end of the Madison Range east of Henrys Lake, nor are any such faults shown on their map. The Madison Range fault is one of the largest Cenozoic structures in the region and probably has been active in a major way within the last few hundred years (Pardee, 1950, p. 373). Referring to the west flanks of the Madison and Targhee Ranges east of Henrys Lake, Pardee (1950, p. 375) notes that they present "abrupt and regular fronts suggestive of worn fault scarps." With this 55-mile scarp still active, and with the subparallel Hebgen fault zone also active, any cross structure without significant active faults would seem to be of secondary importance.

[5] Freeman, L. B. Sweet, J. M., and Tillman, Chauncey, 1950, Geology of the Henrys Lake Mountains, Fremont County, Idaho, and Madison and Gallatin Counties, Montana: University of Michigan unpublished M.S. thesis.

CONCLUSIONS

The fault scarps formed in the 1959 earthquake coincide with or parallel old faults that originated during the Tertiary and have been recurrently active ever since. The deformation of both the Hebgen Lake block and the Missouri Flats block can be satisfactorily explained by renewed movement on the Hebgen–Red Canyon fault complex and on the Madison Range fault, respectively. Subsidence data from several sources show accurately the pattern of deformation of the surface—a pattern that does not necessarily reflect the deformed bedrock. This is especially true in the Madison River canyon where the surficial deposits locally have been unstable since the earthquake. Warping during the earthquake was a minor byproduct of faulting which occurred first at depth and then was propagated to the surface. Reactivation of a deep-seated fault or faults was probably the immediate cause of the Hebgen Lake earthquake of 1959.

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Landslides and Related Phenomena Accompanying the Hebgen Lake Earthquake of August 17, 1959

By JARVIS B. HADLEY

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LANDSLIDES AND RELATED PHENOMENA ACCOMPANYING THE HEBGEN LAKE EARTHQUAKE OF AUGUST 17, 1959

By Jarvis B. Hadley

ABSTRACT

The Hebgen Lake earthquake of August 17, 1959, was accompanied by widespread adjustments of the surficial mantle which can be classed as landslides. These disturbances ranged from ubiquitous small-scale slumps to one of the largest landslides ever recorded in the United States. They were most abundant in an area about 20 miles north-south and 50 miles east-west around the earthquake epicenter near Hebgen Lake.

Many landslides were rock falls and debris avalanches of weathered bedrock on cliffed slopes, set in motion directly by the main earthquake or by aftershocks. Several were debris slides into Hebgen Lake, apparently augmented by the presence of water-filled material along the lake shore. Many others were massive slumps of colluvial mantle jolted downhill by earthquake shocks. The slumps were commonly marked by fissuring along ridge crests and rarely by compressional folds and thrust faults in surficial deposits at the base of the slope. Most of the damage to Hebgen Dam was caused by such massive slumping combined with debris sliding into the lake.

One of the most spectacular features of the earthquake was a very large and rapid rockslide involving nearly 40 million cubic yards of bedrock and colluvium which slid into the canyon of the Madison River below Hebgen Lake. Unlike most other types of landslides that occurred throughout a fairly large area, the Madison Slide owed its size and destructiveness to localized features of topography and to the bedrock structure in the wall of the canyon.

Less spectacular but almost equally impressive is a large earthflow which had developed near Hebgen Lake long before the earthquake, was reactivated by it, and continued to move slowly for nearly a month afterward.

All the fatalities of the earthquake resulted directly from earthquake-induced landslides, as did most of the damage to roads and highways. Damage to roads and highways represents a dollar loss comparable to, if not greater than, that sustained by buildings and other structures.

Evidence of churned ground indicates that surficial material in a few places moved not with the force of gravity but against it. These occurrences were confined to small areas, generally on strong bedrock units with little soil cover.

INTRODUCTION

The effectiveness of earthquakes in causing catastrophic landslides, especially in mountainous regions, is well known. Such landslides have been described from Japan, China, northern India, Italy, Portugal, Peru, Chile, Mexico, and various parts of the United States, particularly California. In fact, the shocks administered to the ground by every major earthquake result in gravitative adjustment of the surficial material, probably over large areas. It is usually, however, only the largest or most destructive of these adjustments that attract attention (Lawson, 1908; Daly, 1926, p. 30–34; Selgado, 1951; Hutchinson, 1957, p. 42, 46).

The Hebgen Lake earthquake was accompanied by its share of such adjustments of the surficial mantle, including a catastrophic rockslide that killed 26 persons. Many less spectacular landslides occurred in an area whose dimensions are still unknown but certainly extended throughout the epicentral area from the southern part of the Madison Valley east to Mammoth Hot Springs and several miles beyond. Most of these features can be classified as rock slides, debris slides, slumps, and rockfalls (Sharpe, 1938, p. 65–80). In addition, a large earthflow (Sharpe, 1938, p. 50–55), which had developed long before the earthquake, was reactivated by it and moved slowly for at least 4 weeks afterward.

ACKNOWLEDGMENTS

Although I was responsible for gathering information on landslides and related phenomena resulting from the earthquake, many helpful observations were contributed by other members of the party and visitors to the project, including D. R. Crandell, Jack Epstein, Warren Hamilton, W. D. Long, W. B. Myers, W. H. Nelson, and I. J. Witkind. D. R. Crandell also gave much helpful advice in preparing the report.

THE MADISON SLIDE

One of the most dramatic and disastrous events connected with the Hebgen Lake earthquake was a landslide in the canyon of the Madison River 6 miles

FIGURE 54.—Madison Slide, upstream view showing the west edge of the slide debris, blocked highway, and dry river bed below the slide. U.S. Forest Service photograph.

downstream from Hebgen Dam (Frontispiece; fig. 54). Thirty-seven million cubic yards of broken rock slid into the canyon, burying a mile of the river and highway to depths of 100 to 200 feet. Water ponded by the slide formed a lake which, 3 weeks afterward, had become nearly 200 feet deep and extended almost to the dam. The slide itself spread upriver to the edge of a campground full of vacationers sleeping in tents, trailers, and on the ground; most of these people were more or less seriously hurt. In addition, some 30 persons had stopped for the night along the riverbank farther downstream, and most were buried by the moving rock. The slide accounted for all but two of the known fatalities resulting from the earthquake, and the other two were also victims of falling rock.

The landslide in the Madison River canyon, by far the most spectacular earth movement resulting from the earthquake, took place 17 miles west of the epicenter of the first and strongest shock and 6 miles west of the area of principal surface deformation at Hebgen Lake. It was clearly not caused by the ground motion alone; other factors were responsible for its location and magnitude. These proved to be purely local structural and geomorphic conditions which had produced a dynamically unstable slope ready for a period of unusual rainfall, a violent flood, or an earthquake to set it in motion.

GEOLOGIC AND TOPOGRAPHIC SETTING

Topographic and geologic conditions in the lower part of the canyon caused slope instability in the slide area. The canyon is narrow and steep walled in the high and rugged southern part of the Madison Range (fig. 55). The walls on the north side are 2,000 to 3,000 feet high in strong, resistant bedrock, that is

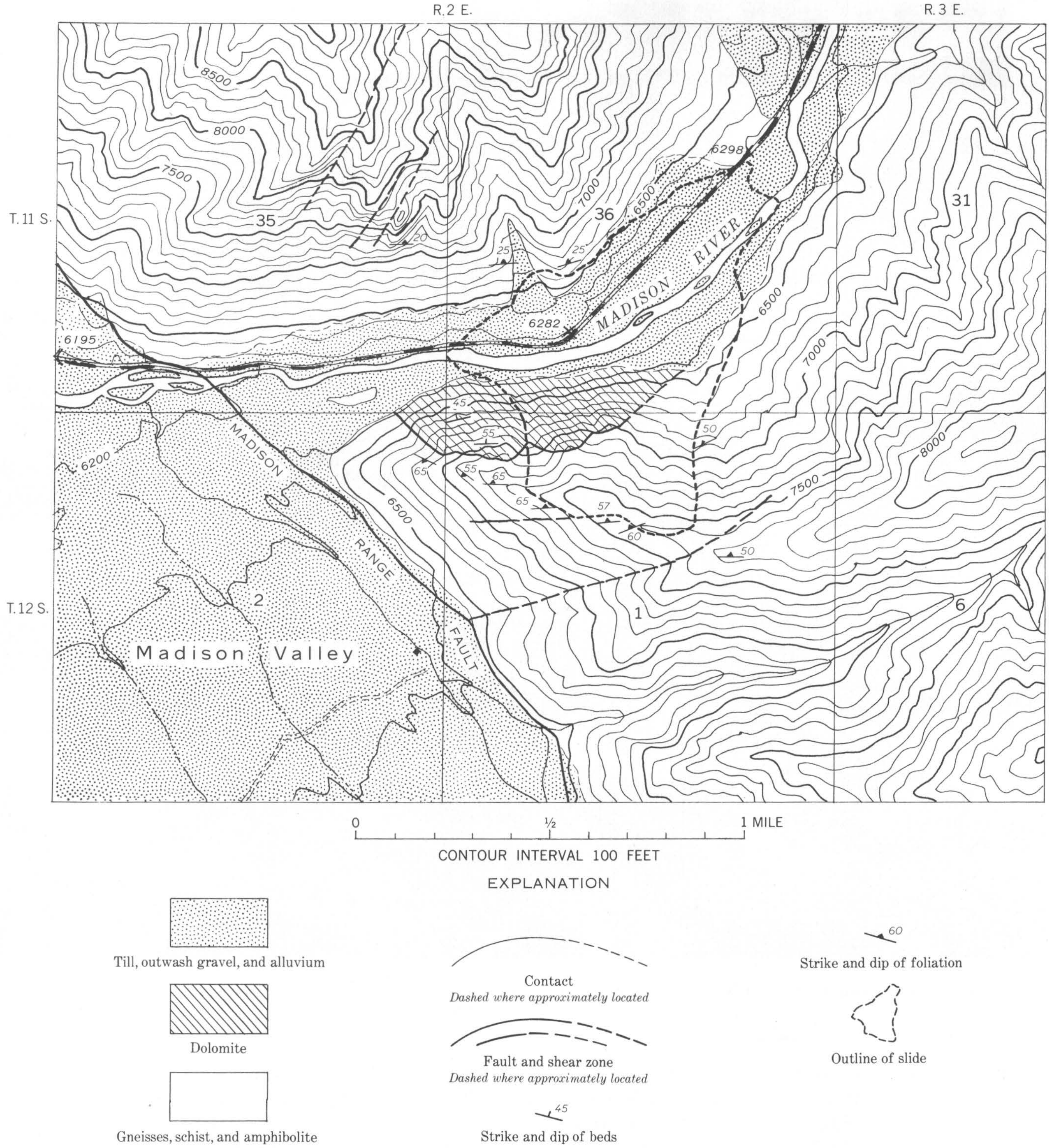

Figure 55.—Geologic map of the Madison Slide area. Datum is mean sea level.

FIGURE 56.—West border of the Madison Slide scar showing dolomite spurs on stable slope. Part of the slide mass in foreground. Photograph by John R. Stacy.

exposed in many ledges and cliffs, and their lower slopes are extensively mantled with talus debris. The wall on the south side, where the slide occurred, is 1,000 to 1,500 feet high and, for the most part, is composed of weaker rocks with less steep slopes mantled by thick colluvial and forest cover. The valley bottom in this part of the canyon is narrow, 500 to 1,000 feet wide, and partly occupied by the Madison River, which is 200 feet wide and 2 to 3 feet deep. The valley is floored by coarse glacial and alluvial gravels 50 or more feet thick.

The south wall of the canyon before the sliding consisted of two parts developed on contrasting types of bedrock. The western part, underlain by dolomite, was characterized by a series of craggy bedrock spurs 300 to 400 feet apart, separated by shallow but steep ravines with slopes approaching 45° (fig. 56). The slopes of the eastern and upper parts, underlain by schist and gneiss, were less steep, averaging 27°. They were marked by minor ridges and gullies 40 to 60 feet deep and 400 to 600 feet apart trending more or less directly down the slope.

The rocks exposed in the canyon walls are entirely Precambrian crystalline rocks of pre-Belt age. Those on the north wall are dominantly amphibolite and hornblende gneiss mixed with quartz-feldspar and other gneisses. The rocks are foliated, moderately layered, and much jointed but not particularly sheared or altered. Rocks on the south wall are micaceous gneiss and schist, which formed most of the slide mass, and dolomite, exposed prominently just west of the slide scar. A small body of coarse diorite or altered diabase makes a small knob surrounded by a gravel terrace at the foot of the south canyon wall just west of the slide.

The gneiss of the south wall is medium to fine grained, mostly moderate- to dark-gray mica-quartz-feldspar gneiss with micaceous partings. Locally it grades into quartzose feldspathic schist or schistose quartzite. The gneiss is generally strongly sheared, and most of the schist probably represents sheared and altered gneiss or other rocks. Some of this highly sheared rock is conspicuously chloritic green and contains many small pods and lenses of vein quartz. Part of the gneiss in the western part of the slide area is dark-gray lime-silicate gneiss characterized by garnet and finely fibrous amphibole. This rock is less schistose than the other rocks and formed large angular blocks conspicuous in the western part of the slide mass.

The schist and gneiss on the south canyon wall are deeply weathered. Exposures in the rim of the slide scar show that these rocks are disintegrated and decomposed to depths of 100 feet or more near the top and more than 50 feet along the east side. Bedrock in the exposure near the top of the scar is soft and friable 50 feet below the ground surface (fig. 57). The rock at the base of this exposure, 150 feet below the crest of the ridge before sliding, is hard but breaks readily along weathered foliation and joint surfaces. Weathered bedrock is probably less thick in the lower parts of the slope, but it is covered in these areas by a thick blanket of colluvium.

Nearly all the schist and gneiss is mantled by variably weathered colluvium whose thickness changes according to the local topography but generally increases toward the bottom of the slope. At one point, where the east edge of the slide scarp cuts through a minor ridge 400 feet above the river, at least 40 feet of coarse blocky and clayey colluvium overlies much weathered and partly disintegrated schist. At the base of the slope in the eastern part of the slide area, the colluvium may be 75 or more feet thick. Along the former crest and uppermost parts of the ridge, weathered but otherwise undisturbed bedrock extends to the grassroots.

The dolomite is a light-gray, locally pink or green tinted medium crystalline marble similar to that in the type Cherry Creek formation of Peale (1896) some 20 miles to the northwest. In the slide area it forms a steep rocky slope with few trees that is easily identified on the ground and in aerial photographs. Here it is marked by relict bedding layers 6 inches to 2 feet thick that are somewhat folded and obscured by joints and foliation. Joints and other breaking surfaces in the dolomite are much less closely spaced than those in the more schistose rocks, so that it was more resistant to weathering and formed much of the coarsest material in the slide mass. Areas underlain by dolomite are either bare rock or mantled by thin debris. Some of the steep ravines had received minor amounts of colluvium from the schist and gneiss higher on the slope.

The most prominent structural element involved in the landslide is the dolomite body, which formed a wedge a mile long and 500 feet wide at river level (fig. 55), that tapers upward and eastward between faults on the south and the erosion surface on the north. This body of relatively strong rock served as a structural buttress bolstering the lower part of the canyon wall. It is cut by joints, conspicuous among which is a set trending northeast and dipping steeply northwest, as well as minor faults with slickensided surfaces. One of these faults, visible in the base of a dolomite spur in the lower western part of the slide scar, dips 50° N. 15° W. and is marked by striae that trend N. 40° W. and indicate reverse movement. Indistinct bedding surfaces in the dolomite dip mostly northward at 45° to 65° but are locally folded.

Northward-dipping foliation and shear zones in the schist and gneiss forming the rest of the south wall of the canyon contributed to the instability and movement of the rock. Most of this foliation and shearing resulted from movements associated with the Laramide faulting along the trend of the canyon and dips northward at angles ranging from 45° to 75°. In the upper part of the canyon wall west of the slide, however, foliation in the gneiss dips steeply southward and thus provides greater stability in that area. A prominent zone of greatly sheared and altered chloritic and sericitic schist 20 to 30 feet wide crosses the ridge obliquely somewhat west of the highest point on the slide scarp. The rock in this zone is par-

FIGURE 57.—View east along the crest of the Madison Slide scar. Highly sheared and weathered schist dipping steeply northward appears in the center; slump fractures in background and in foreground, those in foreground are parallel to schistosity in underlying rock. Photograph by John R. Stacy.

ticularly soft and slippery and partly determined the upper limit of the slide.

The more gently dipping foliation in the rocks of the north wall of the canyon and the south-dipping foliation in the south wall are probably of Precambrian age. The foliation is parallel to compositional layering and is gneissic rather than schistose so that the rocks are not particularly weakened thereby. Intersecting joints provide the principal structural weakness in these rocks, especially a set of strong northeast-trending joint or shear zones that have weathered to produce deep clefts and semidetached pinnacles on the north wall of the canyon.

DIMENSIONS AND SURFACE FEATURES

A section of the south wall of the canyon 2,200

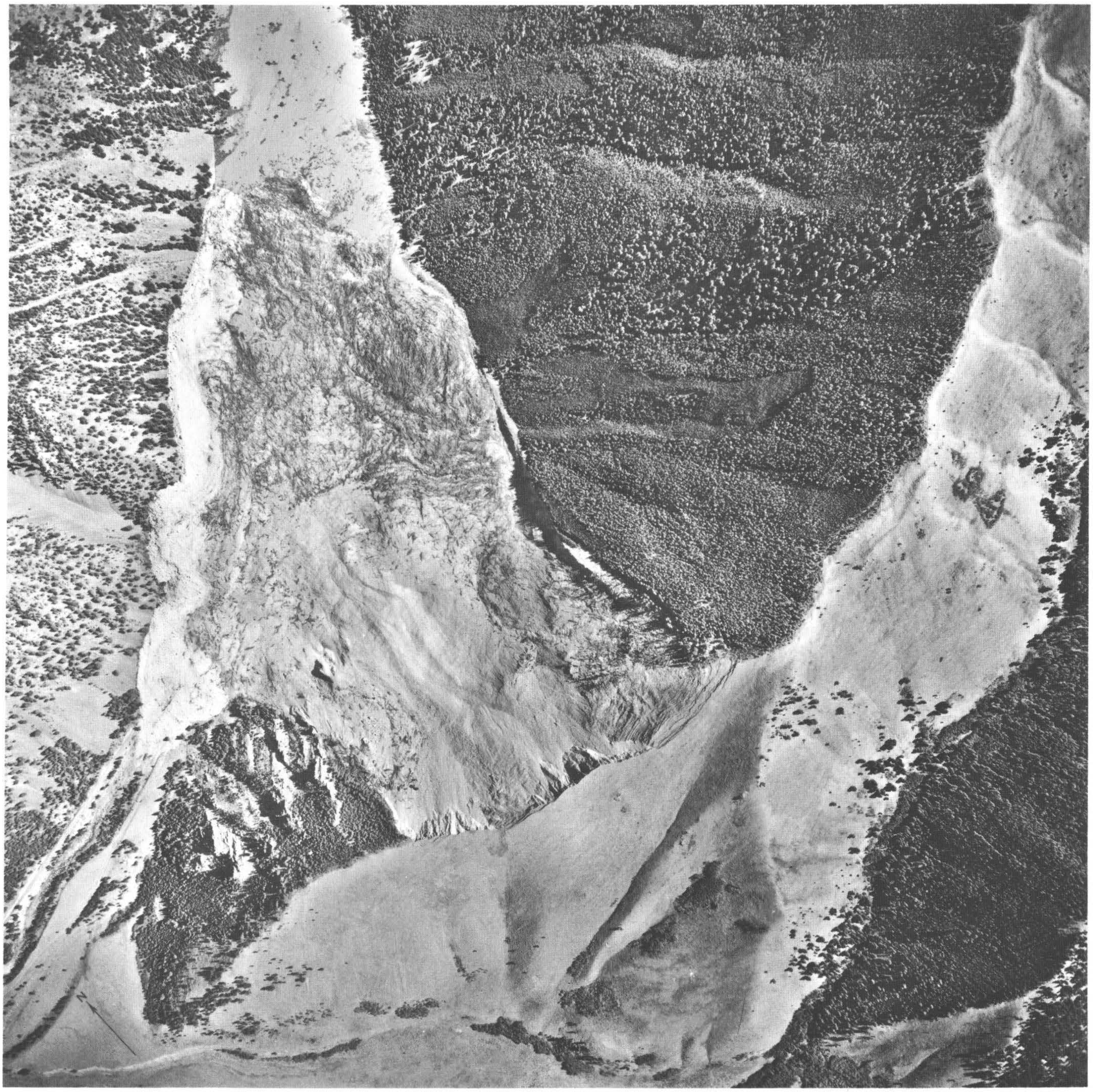

FIGURE 58.—Vertical aerial photograph of the Madison Slide. Recent scarp of Madison Range fault appears southwest of slide. U.S. Forest Service photograph, August 22, 1959.

feet long and 1,300 feet high slid northward into the canyon, spreading over an area of about 130 acres on the canyon floor and adjacent slopes (fig. 58). The farthest point on the toe of the slide mass reached just a mile from the crest of the scar. A dolomite spur 700 feet high and two smaller adjacent spurs were largely removed, as well as a large volume of deeply weathered schist and gneiss uphill and east of the dolomite. The slide scar not only reached the crest of the ridge on the south side of the canyon but in places cut 300 feet beyond or 200 feet vertically down the back slope. The resulting scar is 400 feet

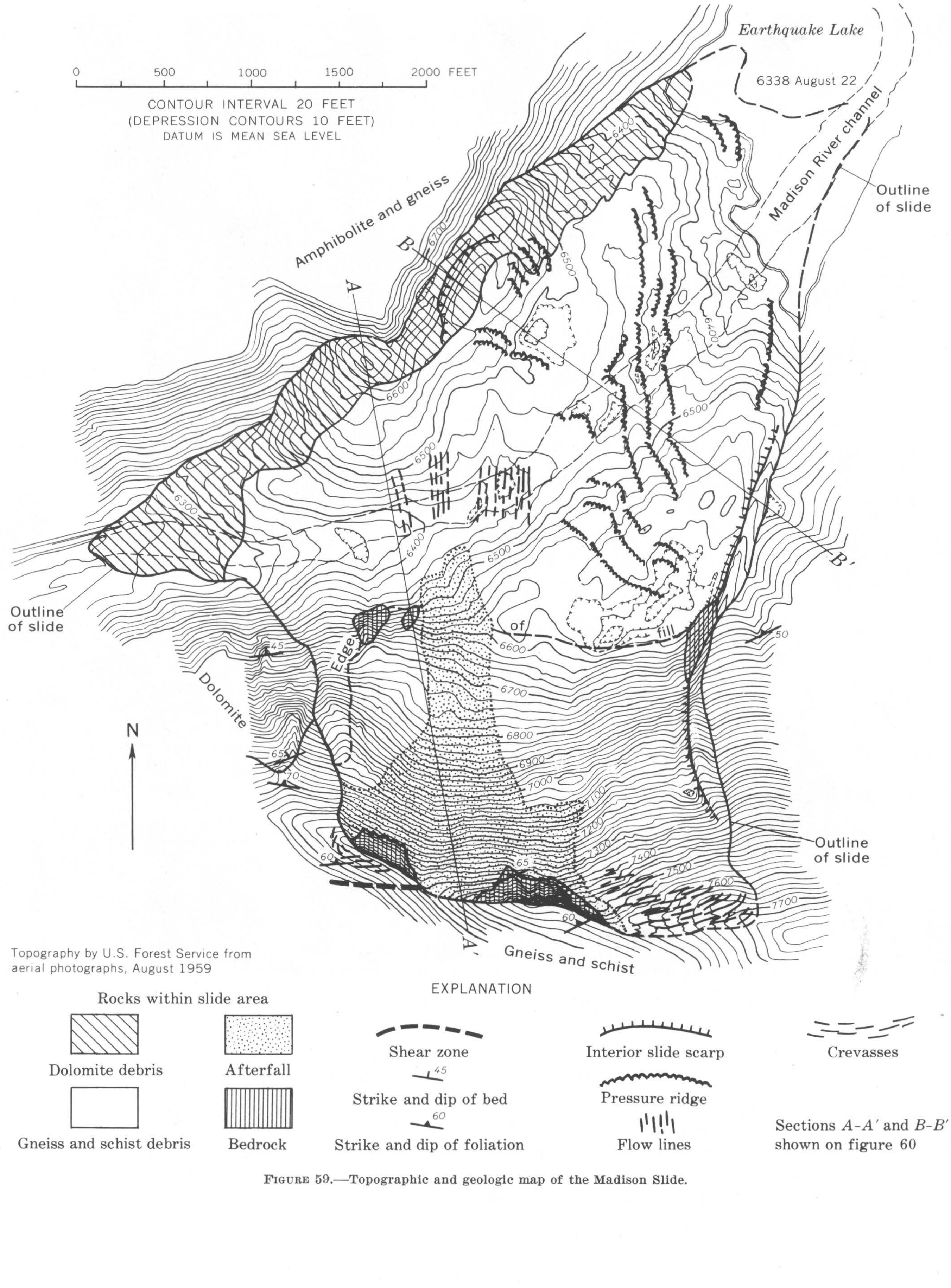

FIGURE 59.—Topographic and geologic map of the Madison Slide.

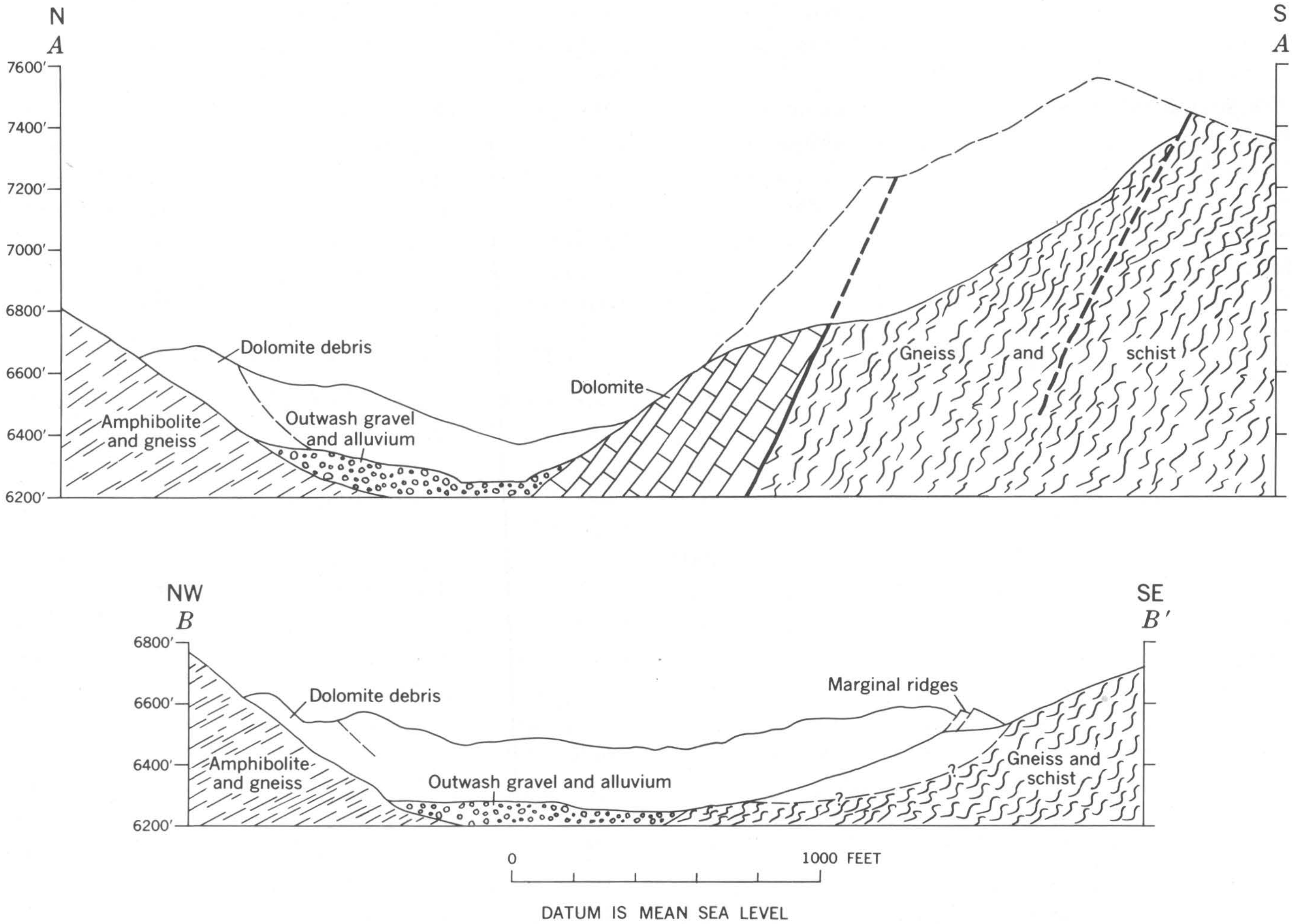

FIGURE 60.—Sections of the Madison Slide. Section lines are on figure 59.

below the former surface in some places and 100 to 300 feet below it in most other places. It thus forms a broadly concave surface more than 2,000 feet wide on the south canyon wall where previously the surface was strongly convex (fig. 59).

The surface of movement of the slide mass cut obliquely across the schist, gneiss, and dolomite, breaking the dolomite at the west end about 350 feet above the valley floor and descending eastward in the weaker rocks probably to near river level (fig. 60). The upper limit of the slide was determined in large part by the weak zone of sheared and altered schist previously mentioned. The top of the scar followed this shear zone in part, but to the east and west it deviated so as to conform more closely to the former ridge crest. Small step faults with 3 to 5 feet of displacement in the ground above the western part of the scar show a continued tendency for sliding to follow bedrock structure (fig. 57). At the east end, the former ridge crest slumped broadly on a series of step faults rather than giving way in a sharp break (fig. 57). The lateral boundaries of the scar were determined by topographic and dynamic rather than structural features. The west boundary followed a steep ravine between two of the larger dolomite spurs. The nearly straight east boundary, which cuts sharply across minor ridges and gullies, appears to have been controlled mainly by the direction of motion of the mass as a whole.

If faults or other through-going structural elements controlled the depth of the surface of rupture, they are not evident. Rather, this seems to have been determined mainly by the depth of weathering and fracturing of the bedrock, the ability of the dolomite buttress to withstand the thrust of the energized slide mass, and probably the energy imparted to the slide mass by the earthquake itself.

The slide debris covers an area of about 130 acres in the bottom of the canyon to an average depth of 150 feet and a maximum depth of 220 feet in and just north of the buried river channel. The channel itself is marked by a pronounced topographic trough and several closed depressions in the slide mass. It was filled with debris that is 220 feet deep at the highest point in the trough but thins gradually toward the upstream and downstream margins of the slide (fig. 59).

Most of the slide debris crossed the river, but perhaps a third remained on the south side. The western part, coming from the steeper part of the original canyon wall, went highest, although not farthest, as its movement was impeded by the opposite wall of the canyon. Nearly all this part of the slide mass crossed the river, leaving only stumps of the dolomite spurs and a few small debris cones on the south side. Its momentum, greater than that of other parts of the slide, carried it to the highest point reached by the mass, 430 feet above the river and directly across from the highest dolomite spur (section *A–A′*, fig. 60; fig. 61). From the section it appears that the center of mass in this part of the slide traveled 1,600 feet on a 30° slope and, assuming an acceleration one-fourth that of gravity, would have acquired a maximum velocity of 100 miles per hour and taken 20 seconds to come down. As the west end of the slide seems to have broken out considerably above the valley floor, it is likely that the violent air blast reported by eyewitnesses (chapter A) resulted from expulsion of air trapped beneath the descending debris.

The eastern part of the slide mass traveled more slowly. Its edge is marked by marginal ridges that resulted from initial swelling and rising of the ground followed by sliding of all but the outer margin of the raised area (fig. 62). These ridges are 40 to 60 feet high, 100 feet or less wide, and 600 to 1,000 feet long. Their inner slopes are steep scarplike cuts in the slide material, whereas their outer slopes are less steep and retain their original forest cover, which was more or less disrupted in the process of raising the ridge. In some places the subsoil was so broken up and the outer slope so steepened that the trees and soil slid away to the bottom of the ridge. North- and northwest-trending pressure ridges developed between the slower and faster moving parts of the slide and produced the closed depressions in the trough of the slide mass (fig. 59).

This part of the slide mass is also covered by felled trees and large blocks of soil and sod thickly scattered over broken and weathered rock. Many trees are alined with their tops pointing backward, opposite to the direction of sliding; others are alined crosswise or in other directions, apparently toppled by heaving ground or swept into alinement with pressure ridges.

COMPOSITION AND VOLUME OF SLIDE DEBRIS

Most of the surface of the slide mass consists of an aggregate of angular blocks conspicuously lacking in fine fragments. Most of the rock is gneiss, but prominent bands of schist, lime-silicate gneiss, and dolomite debris parallel the toe of the slide mass and preserve the order in which these rocks formerly existed on the ridge. The most conspicuous of these bands is that made by the dolomite debris spread out in a prominent white ridge in the highest part of the slide mass along the north wall of the canyon. A tonguelike apron below the bedrock exposures in the upper part of the slide scar consists of afterfall, or debris flows which mantled the scar after the main slide had occurred (fig. 59).

Much of the blocky debris at the surface of the slide mass consists of gneiss fragments ½ to 1 or 2 feet long, although in many places blocks of lime silicate gneiss are 2 to 5 feet long; schist fragments are generally smaller, mostly 1 to 12 inches. The dolomite produced the largest fragments, among which blocks 5 feet or more across are common; some are 20 feet or more in length, and one house-sized block that rode on the front of the slide mass is nearly 30 feet on a side. All the fragments have broken along weathered joint or foliation surfaces, although many are fresh inside. The only surfaces exposing fresh rock are those where the corners or edges of the larger blocks were knocked off in sliding.

Excavations made in preparing a spillway across the slide revealed that only the top few feet consists of coarse fragments without fine material, and that the upper part of the debris includes abundant smaller rock fragments as well as much sand, silt, and clay-sized material (fig. 63). Most of this represents decomposed rock in the original colluvium and weathered bedrock, but presumably some was produced by crushing and grinding in the mass as it descended. Ab-

FIGURE 61.—Dolomite debris at highest point of slide mass against the north wall of the canyon, 430 feet above the riverbed. Photograph by John R. Stacy.

sence of the finer material in the surficial layer of the slide mass and its abundance in the immediately underlying debris are attributed to surface sorting by movements during transport and perhaps also by vibrations from aftershocks of the earthquake.

Topographic maps of the slide area were made with a Kelsh plotter by the U.S. Forest Service from aerial photographs at 1:10,000 and 1:15,000 scale taken before and 5 days after the slide took place. These maps, at a scale of 1 inch equals 200 feet and with contour intervals of 5 and 10 feet, show in great detail the shape of the slide scar, its relations to the former ground surface, and the topographic form of the slide mass. From them, fairly accurate measurements could be made of the volume of rock removed from the scar and of the volume of the slide mass itself.

The volume of rock and earth removed from the

FIGURE 62.—Tree-covered southeastern part of the Madison Slide showing marginal ridge left by movement of the slide mass past it. Earthquake Lake in the background, August 28.

scar, as computed from seven cross sections of the slide 500 feet apart, is 28 million cubic yards, weighing 60 million tons at a density in place of 2.6. In the same manner, the volume of debris deposited in the bottom of the canyon is calculated as 37 million cubic yards. The slide material thus increased in volume by 9 million cubic yards or 32 percent and acquired a porosity of about 25 percent. In terms of comparative volumes of slide debris, the size of the Madison Slide is similar to that of the Turtle Mountain slide of 1903 at Frank, Alberta (35 to 40 million cubic yards), and somewhat smaller than the Gros Ventre slide of 1925 in northwestern Wyoming (50 million cubic yards).

EARTHQUAKE LAKE AND OTHER MARGINAL FEATURES

Evidence at the upstream edge of the slide indicates that a wave of muddy water carrying trees, driftwood, and small rocks was pushed ahead of the slide mass. The wave reached a maximum height of 100 feet above the river on the north wall of the canyon and continued with decreasing height for ¼ to ½ mile upstream beyond the toe of the slide. Its limit was soon inundated by rising water in Earthquake Lake and could not be precisely determined. This wave broke the lower limbs from trees near the edge of the slide and left a swash mark of wood, mud, and matted grass and bushes at its farthest reach. It engulfed the Rock Creek campground and caused much of the injury to persons and damage to property there.

A similar wave at the downstream side of the slide reached a terrace on the south side of the river 15 feet above the channel and flooded it briefly to a depth of 1 or 2 feet for 500 feet along the riverbank, stranding logs and fish among the sagebrush. It carried logs ½ to ¾ mile downriver and swept two badly battered automobiles 100 yards beyond the downstream edge of the slide.

At the upstream side of the slide, the ponded water

FIGURE 63.—Northern part of slide mass from trough in slide mass up dry gully on north side of canyon. Dolomite debris piled high against the side of the gully at upper right. Earth-moving equipment preparing spillway in foreground. Note the fineness of much of the slide debris beneath the surficial cover of blocks. Photograph by John R. Stacy.

of the Madison River formed a new lake that eventually reached the crest of the slide mass and extended upstream 6 miles, or almost to Hebgen Dam (frontispiece, fig. 64, and pl. 2). The level of the new lake rose very rapidly at first because of the small area of the new lake and the arrival of water from the first 3 or 4 surges in Hebgen Lake. By 6:30 on the morning after the earthquake, the water had risen 20 feet or more and had covered all the vehicles in Rock Creek Campground just upstream from the slide. It rose at a gradually decreasing rate thereafter for a little more than 3 weeks (fig. 65). Its rise was decreased to about a foot per day by controlling the flow at the dam during the last 5 days to allow completion of an artificial spillway across the slide mass, and the water began flowing through this spillway on September 10, 24 days after the slide occurred.

Uncontrolled erosion of the slide mass and the sud-

FIGURE 64.—Earthquake Lake seen from the upstream edge of the slide on August 20. Nearer trees in the water are in Rock Creek campground, submerged 20 to 40 feet.

den draining of the lake that would have resulted were prevented by prompt action on the part of the U.S. Army Corps of Engineers, Garrison District, who prepared a spillway 200 feet wide and half a mile long in the natural trough through the slide mass.[6] Time permitted only a minimum operation and the use of only the materials of the slide itself to control an expected flow between 1,000 and 2,000 cubic feet per second across the slide. This proved impossible at the gradient determined by the length of the spillway and the original level of the new lake; so the top of the spillway was lowered by 50 feet. This drained nearly half the water behind the slide and reduced the spillway gradient to one more consistent with the material remaining in the spillway channel.

CAUSES OF THE LANDSLIDE

The fundamental causes of the landslides were slope instability of long standing coupled with the energizing effects of the earthquake. Climatic factors, although instrumental in developing conditions favorable for sliding, were not involved in setting the slide in motion, nor does it seem at all likely that flooding or other activity in the bottom of the canyon served to trigger the slide.

Slope instability in the slide area seems to have been produced by the geomorphic (rather than mechanical) effect of the dolomite wedge in the lower and western parts of the slope on the south wall of the canyon. Here the rock maintained a slope angle of nearly 40°—considerably steeper than that consistent with the strength and resistance to erosion of the schist and deeply weathered rock in the bulk of the ridge behind it. Slopes east of the slide average 25° and seem to represent more nearly an equilibrium profile for this kind of rock. West of the slide, where dolomite forms a larger part of the ridge and the foliation of much of the weaker rock dips southward into the ridge, both structure and rock strength are more consistent with

[6] A comprehensive report on these operations was published in September 1960 by the U.S. Army Corps of Engineers, Omaha District, Omaha, Nebr.

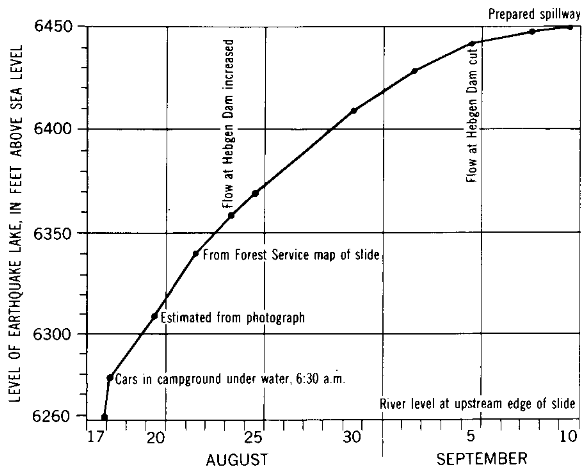

FIGURE 65.—Curve illustrating rise of Earthquake Lake behind the Madison Slide. Points not otherwise indicated are from U.S. Army Corps of Engineers gage record. Marks on the time scale indicate midnight.

the steeper angle, and the slope was stable under the conditions of the earthquake.

The question of the manner in which the slide was set in motion is not so simple. The essential condition seems to be that a mass of weak rock, held in place largely by friction, was suddenly transformed into a mass which literally flowed into the canyon. Also important is the evidence that this fluidity was produced simultaneously over the slide area. Factors tending to reduce the internal friction or resistance to sliding, or to overcome this resistance by imparting momentum to the mass, are presumably responsible for the sudden change from solid to fluid behavior. Increased water content, a common cause of increased mobility in landslides, is not a factor here for there had been little or no rain in the area for 6 weeks prior to the slide and the slide mass was notably dry. Dynamic factors due to the earthquake therefore seem to be primarily responsible.

In addition to the strong ground vibrations in this area, the whole of the Madison River canyon subsided 7 to 8 feet during the earthquake, probably during the first few minutes (Myers and Hamilton, chapter I). A sudden drop of the landslide area would have produced a momentary decrease in the gravitational force acting on the slide mass and would have been most effective in reducing the frictional resistance. It would, however, have equally decreased the force tending to move the mass downhill. On the other hand, the downward momentum imparted to the slide mass could have been considerable and would have both increased the force tending to overcome frictional resistance and decreased that resistance by causing dislocation within the mass itself. Fluidity then would have increased rapidly as the slide gathered speed. Such considerations as well as the abrupt downhill dislocations of the surficial mantle seen throughout the earthquake area suggest that the big landslide was a direct result of the first and strongest earthquake shock.

It is conceivable that the initial shock was insufficient to produce the requisite fluidity in the slide mass and that this was acquired more gradually by repeated shocks and preliminary smaller movements in the slide mass, culminating in a sudden release of the slide as a whole, perhaps during one of the aftershocks. Most contemporary accounts by eyewitnesses, however, did not suggest any great lapse of time between the first earthquake shock and the landslide; and responses to our later inquiries, received from three persons who were in the vicinity of the slide, state definitely that the slide followed almost immediately after the first shock.

Another facet of the question of what caused the landslide is its proximity to the great Madison Range fault, less than half a mile distant, along which the floor of the Madison Valley dropped 25 feet or more only a few hundred years ago. Presumably the slide area was about as unstable then as just before the present slide and the question obviously arises, why did it not occur then? The most likely explanation seems to be that the slide area is on the stable side of the Madison Range fault, rather than on the downdropped side, and therefore did not acquire the energy necessary for sliding.

OTHER LANDSLIDES

Many landslides involving much smaller masses that moved for shorter distances than the slide in the Madison River canyon occurred throughout the earthquake area. Most took place in the surficial mantle rather than in bedrock, although a few occurred in the softer bedrock formations such as the Woodside formation.

An unusual concentration of slides along the northeast shore of Hebgen Lake within a few miles of the dam carried 3,000 feet of State Route 499 and several houses into the lake. The largest of these slides, just north of Hilgard Lodge a mile from the dam, extended nearly 100 feet above the lake level and carried 1,000 feet of the highway and 350,000 cubic yards of earth beneath the surface of the lake (fig. 66). Many smaller but similar slides occurred on both sides of the narrow steep-sided northern part of the lake and carried surficial material and trees into the water (fig. 79).

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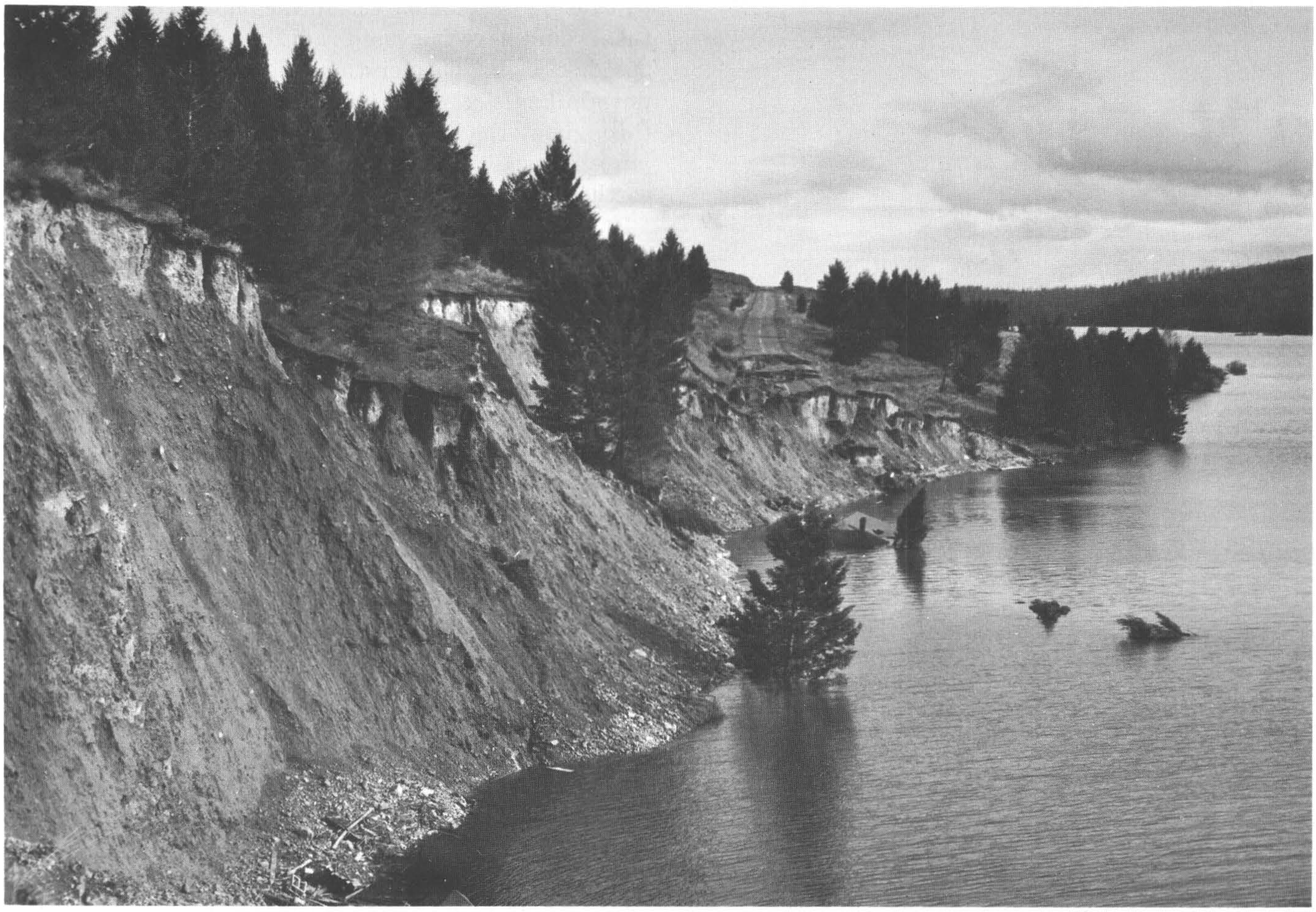

FIGURE 66.—Earthquake-induced landslide on the northeast shore of Hebgen Lake 0.6 mile from Hebgen Dam. Depth of water below the cut is indicated by the floating bungalow in the middle distance. Beyond is the tectonically submerged shore near Hilgard Lodge.

The concentration of landslides in this area is probably due to their proximity to the reactivated Hebgen fault and to the locus of maximum subsidence along the northeast shore of the lake. The slides occurred only in colluvial material on slopes steeper than 10° which extended below water level. Slopes as steep or steeper that were distant from the lake did not slide in spite of the fact that some of them were very close to the Hebgen fault. On the other hand, bottom soundings of the lower end of the lake, described by W. H. Jackson (chapter H), show accumulations of slide material as much as 20 feet thick in the submerged river channel 500 feet from the shore. This suggests that the presence of saturated material in the lower parts of these slopes was a major factor in their instability.

The slides along Hebgen Lake occurred during the first shock or soon afterward, for they had already cut the highway by the time the 200 persons who were caught in the Madison River canyon between the dam and the big slide attempted to leave. They were caused presumably by the momentum produced by the violent subsidence of the lake shore.

Other landslides of various proportions were reported from areas as far away as the valley of the Yellowstone River north of Gardiner and included a large slide on Mount Holmes in the northwestern part of Yellowstone National Park. These could not be investigated in the time available.

ROCKFALLS

Rockfalls were the most common type of landslide resulting from the earthquake. In general these were falls of individual fragments, in small or large numbers, from the upper parts of bedrock cliffs; only in a few places did the volume of falling material approach avalanching. Cliffs of volcanic rocks in the northwestern part of Yellowstone Park and in the vicinity of Cliff and Wade Lakes in the southern part of the Madison Valley were particularly affected, and most

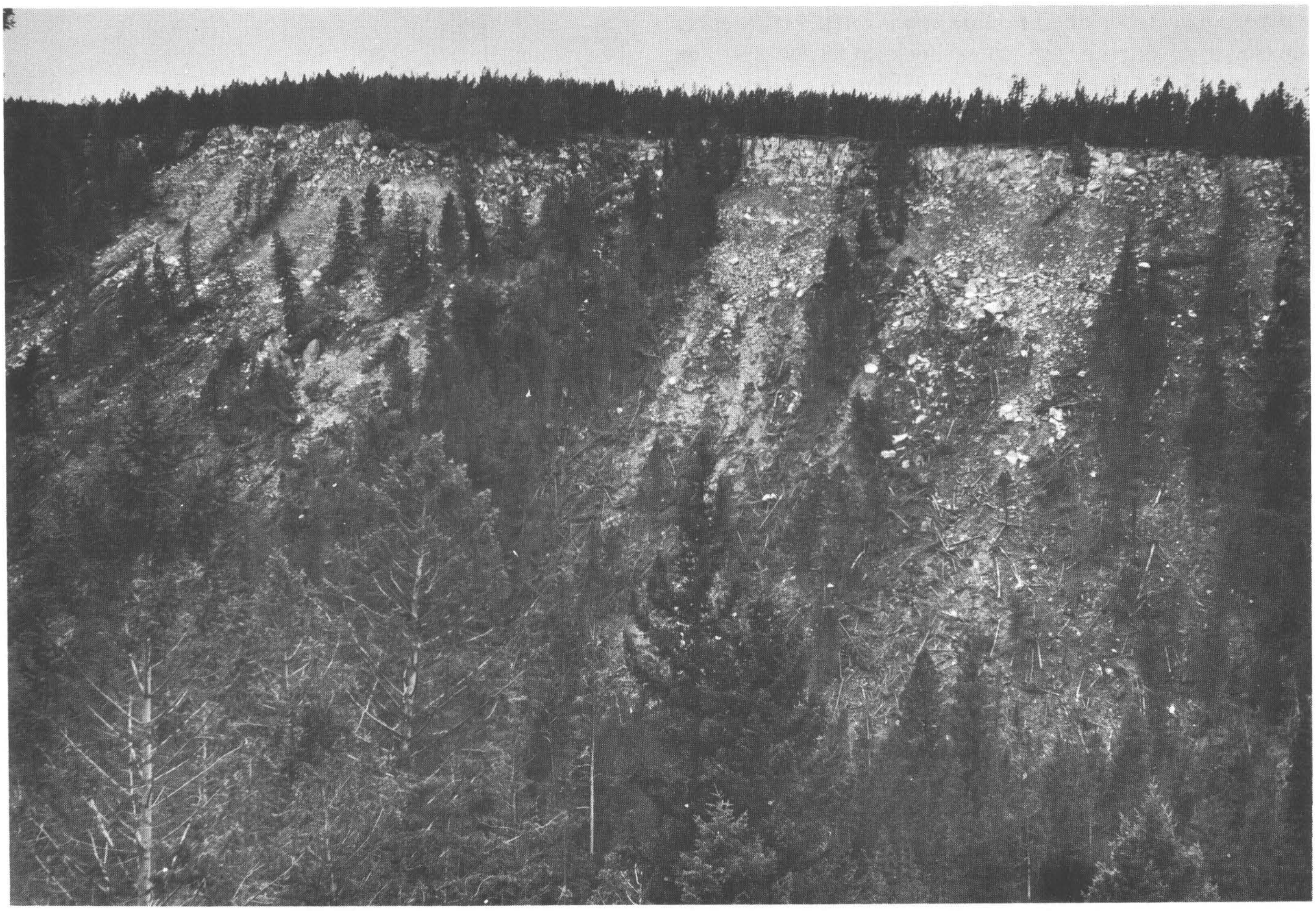

FIGURE 67.—Rhyolite cliffs, half a mile southwest of Wade Lake, from which blocks fell in large numbers, smashing trees on the slope beneath.

of the damage to the park was caused by such falls onto highways.

The upper canyon of the Madison River between West Yellowstone and Madison Junction was the scene of many rockfalls which continued for several days after the first disturbance. Here, large boulders of weathered and variably altered rhyolite rolled down into the canyon from rhyolite cliffs several hundred feet above the canyon floor; some of the largest boulders landed on the highway or bounded across the river. Many of these falls were precipitated by aftershocks, especially that of 8:26 a.m. on August 18, and were witnessed by Park rangers and others. Only minor amounts of material fell from the obsidian walls of the Firehole Canyon just south of Madison Junction.

Similar rockfalls descended from rhyolite cliffs along Cliff and Wade Lakes and along the Madison River for several miles northwest of the canyon below Hebgen Dam. The cliffs surmount old canyon walls cut in nonresistant rhyolite 100 or more feet thick capped by well-consolidated rhyolite tuff 40 to 60 feet thick. This tuff, with a specific gravity considerably less than most rock, is broken by widely spaced joints into blocks several feet to 20 feet on a side and weathered to loosely held residual boulders on the cliff faces. These blocks fell in large numbers during the earthquake, smashing large trees on the slopes below and coming to rest in some places amid great piles of similar blocks which had preceded them in past centuries (fig. 67). At the north end of Cliff Lake, several blocks fell on a small campground; one, landing squarely on a tent with two occupants, accounted for the only fatalities other than those in the Madison slide.

Many similar rockfalls were shaken from the steep rocky slopes in the glaciated Precambrian part of the Madison Range at least as far north as Sentinel and Deadman Creeks and as far south as the head of Mile Creek, and from the Paleozoic rocks at both ends of Boat Mountain just northwest of Hebgen Lake. Several blocks fell on State Route 499 in the canyon just west of the slide area during the early shocks and during the aftershock at 8:26 a.m. on August 18. All these falls contributed to extensive taluses already formed. The talus slopes themselves generally remained stable, for no major sliding in old talus was observed.

KIRKWOOD EARTHFLOW

A remarkable example among the many kinds of landslides associated with the earthquake is a large earthflow in the drainage of Kirkwood Creek about a mile east of Hebgen Dam (pl. 2). This is a tongue-shaped flow 400 to 800 feet wide and half a mile long which began to move at least 5 days after the main earthquake and was still moving slowly on September 18, a full month after the quake. The flow is an old one, developed to its present size many years ago but apparently inactive for years. Renewed movement resulting from the earthquake amounted to at least 100 feet at the time the flow was studied in detail on September 18.

The flow occupies the floor of a southeastward-draining valley tributary to the North Fork of Kirkwood Creek, and its head is just south of the divide between Kirkwood and Cabin Creeks. All the valley floor is occupied by the moving mass, which heads in an area of old and new slide scars and terminates in a steep rounded front marked by abundant crevassing (fig. 68). The overall gradient of the flow from the top of the scarps to the toe is about 10°, but in the middle two-thirds of the flow it is only a little more than 6°. The surface of the flow is in part heavily forested, and the recent movement was evident in hundreds of tilted and fallen trees. Several small sag ponds on the flow before the quake were tilted, broken, and drained by the renewed movement.

The bedrock at the northeast side of the valley consists of the Woodside formation, Thaynes formation, Ellis group, and Morrison formation, in ascending order; all the units dip 10° to 20° NW. These beds are capped by resistant sandstone of the Kootenai formation on the divide at the head of the valley. The southwest side of the valley consists largely of rocks of the Ellis group that strike northwest parallel to the valley and are overturned toward the northeast. The head of the flow is in the soft, clayey Morrison formation, which appears to make up a good deal of the flow. The material seen in crevasses as much as 10 feet deep is soil, subsoil, and colluvium, all very clayey.

The flow boundaries are sharp; those along the sides of the upper half of the flow are well-developed shear

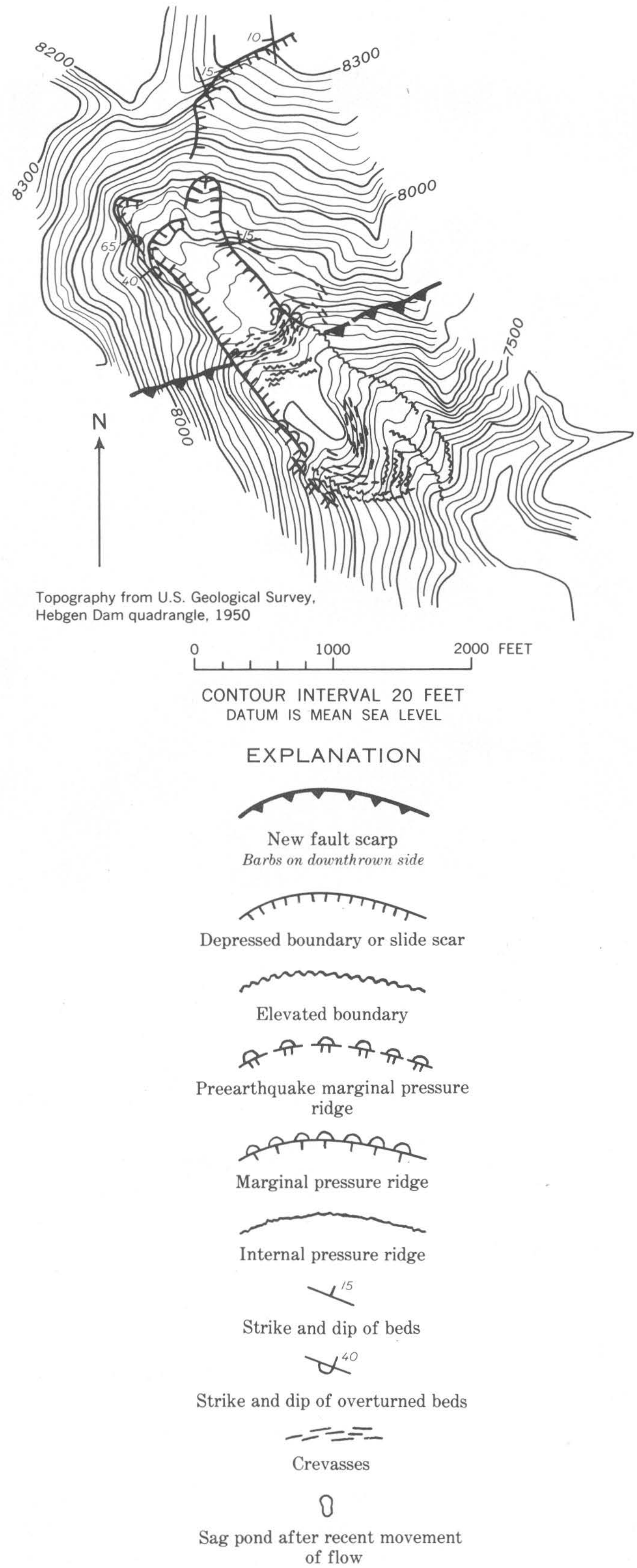

FIGURE 68.—Map of Kirkwood earthflow 1 mile east of Hebgen Dam.

FIGURE 69.—Depressed margin at southwest side of the Kirkwood earthflow 700 feet from its head. The ground surface subsided 10 to 15 feet. Faint grooves on scarp plunge 10° SE (toward left).

surfaces that dip steeply under the flow and are seen as scarps 1 foot to 15 feet high (fig. 69). The scarps are commonly marked by grooves plunging 10° to 15° in the direction of flow and the surface of the flow is locally tilted or stepfaulted downward toward the scarps. The shear-scarp boundary along the southwest side of the flow is remarkably straight and continuous for a quarter of a mile, apparently guided by the steeply dipping beds in the adjacent valley wall.

The margin for 1,000 feet along the northeast side and for a shorter distance along the southwest side of the flow is an elevated boundary along which the flow surface is raised as much as 18 feet above the adjacent ground (fig. 70). Such a margin developed in part where the ground surface at the edge of the flow was locally steeper than the overall gradient so that higher ground on the flow has been brought against lower ground at its margin. In part it also represents thickening and swelling of the flow mass. The shear surface in such places is generally concealed by earth and sod slumped from the edge of the flow.

A most remarkable boundary developed for 350 feet along the southwest edge of the flow where it impinged against the steep side of the valley (fig. 71). Here the flow encountered rising ground instead of level or falling ground and produced a pressure ridge along which the edge of the flow was raised 20 feet, tilting the forested surface from a 30° to a 75° slope and the trees on it to angles of 45°. The boundary shear surface of the flow was turned like a plowed furrow from its normal slope of 65° inward to 70° outward (fig. 72). Grooves on this surface plunge 5° to 10° backward toward the head of the flow and graphically record the raising of the ridge.

About midway in its length, the flow moved over a transverse terrace where a new fault scarp originally crossed the flow, although it was obliterated by subse-

FIGURE 70.—Elevated margin of the Kirkwood earthflow about midway on the northeast side raised 15 feet. Man is standing on a trail whose continuation on the flow is displaced 100 feet to the left.

quent movement of the flow. Many transverse crevasses, similar to those in icefalls on glaciers, developed where the moving earth went over the edge of the terrace and low transverse pressure ridges formed just below them (fig. 73). An acre of large trees were uniformly tilted forward as they went over the edge of the terrace, and trees that formerly grew on the terrace face were tilted backward as they reached the level ground below (fig. 73).

The lower end of the flow is a bulbous ridge with a steep front 250 feet high marked by many large crevasses in its upper part and by pressure ridges and small underthrusts lower down along the toe of the flow. Many trees were uprooted and overturned in an area of deep crevassing on the east side of this ridge, and their roots were still parting with explosive sounds at intervals of half an hour or so while we were studying the flow (fig. 74). Trees along the toe toppled forward as the ground rose, but no large-scale overriding was evident.

That the middle part of the flow moved at least 100 feet was determined in three separate places. A trail crossing the elevated east margin 1,000 feet from the toe (fig. 70) was displaced at least 60 feet and probably nearly 100 feet, although the displaced part was partly obscured by slumping along the margin of the flow. A tree standing on the edge of the western marginal scarp at the northwest end of the pressure ridge lost branches, apparently broken off by a passing tree on the moving margin of the flow. The closest tree that would have been able to do this is 90 feet farther down the flow, indicating a movement of more than 90 feet at this point (fig. 71). Finally, a dry pond that was undisturbed when aerial photographs were taken of the earthflow area on August 22 had moved 100 to 125 feet relative to objects outside the flow by the time we visited the area (fig. 68).

The scarp at the head of the flow shows as much movement as this or more and was the area of greatest

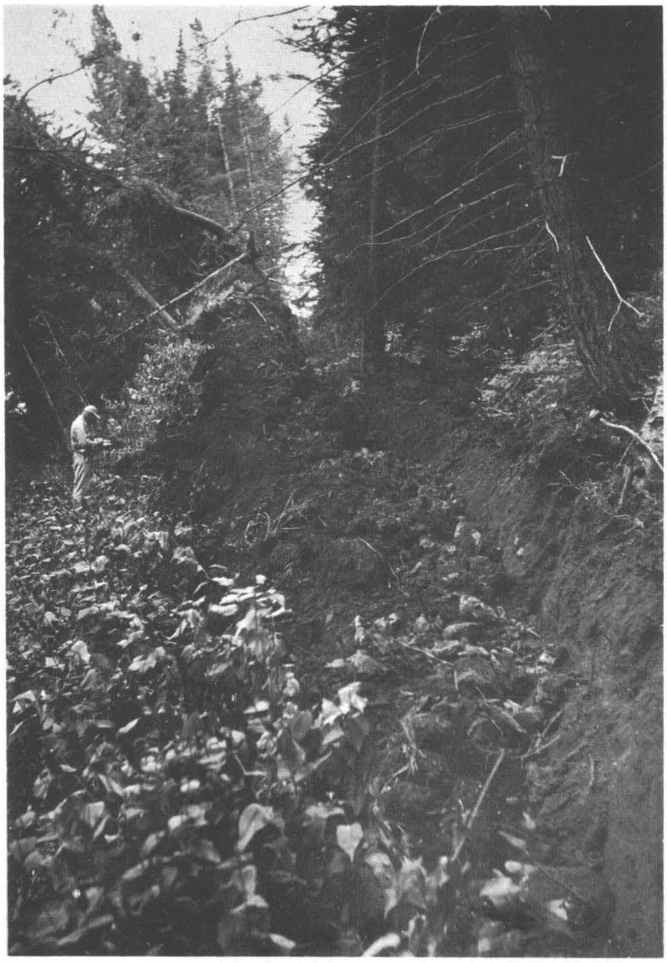

FIGURE 71.—Marginal pressure ridge at southwest side of Kirkwood earthflow 1,500 feet from its head. View is southeast along the boundary shear scarp at the northwest end of the ridge. Toppled tree at crest of ridge moved there from a point behind the camera, breaking branches from trees at the west side of the scarp before it fell.

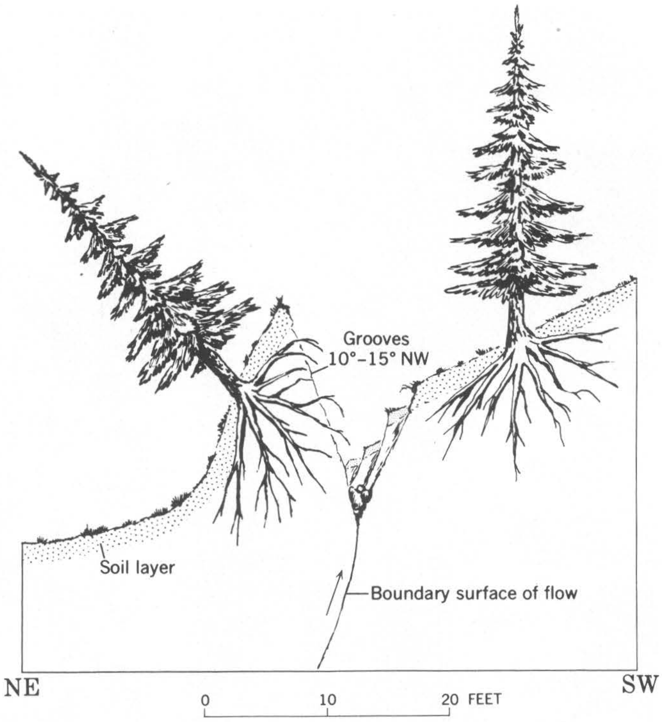

FIGURE 72.—Section of the pressure ridge at the southwest margin of the Kirkwood earthflow.

subsidence and horizontal displacement (fig. 75). Most of the bulk motion was apparently absorbed in raising and thickening the front of the flow and in forming the pressure ridges and underthrusts which are abundant along the toe. The toe itself did not override the ground in front of the flow to any great extent.

Although this movement began at least 5 days and possibly more after the earthquake, it was indirectly caused by the quake. Two effects of the earthquake may have helped to reactivate the flow. One is the increased flow of ground water that was noted soon after the first shock in many parts of the earthquake area. No water was flowing from the earthflow when we studied it, but it seems likely that ground water from adjacent bedrock soaked into its lower part, and increased its capacity to move on the gradient previously established. The other effect is that the previous gradient was increased 35 feet per mile by local warping due to the earthquake (pl. 5), and the lower half of the flow was abruptly dropped about 3 feet along the fault which crosses its midsection. These effects, though small, impelled the flow to take another large step in its gradual progress into the valley of Kirkwood Creek.

MANTLE SLUMPS

On many of the steeper slopes and ridges within several miles of Hebgen Lake, ground fractures were formed as a result of massive downhill slumping of the unconsolidated surficial mantle and, in places, the more weathered upper part of the bedrock. These fractures appear as normal fault scarps a few inches to a few feet high that are distinguishable from tectonic fault scarps only by the fact that they are concentrated along ridge crests and nearly always parallel the crest. Most are within 300 feet of the crests, and on the steeper slopes they show approximately equal vertical and horizontal displacements in the direction of the slope. Those squarely on the crests are gashes which may be several feet wide with little or no vertical displacement. A few scarps are vertical fractures, near the crests, in which the uphill side went down. In such scarps, the fracture beneath

FIGURE 73.—Aerial view of the midsection of the Kirkwood earthflow where it went over a bedrock terrace. The elevated east margin is at lower center; crevassed ground and tilted trees are in center; and the depressed west margin is in the upper part of the picture. The new fault scarp appears at the top, and a nearly drained sag pond can be seen at the upper right. Photographed September 3. More trees were tilted and the sag pond had moved considerably by September 18. Photograph by John R. Stacy.

FIGURE 74.—Aerial view taken September 3 of the lower part of the flow as seen from the east. The high front of the flow is the partly wooded ridge in left center marked by crevasses and twisted trees. Many low pressure ridges are visible in the treeless area at left. Area at extreme right also shows in figure 73. Photograph by John R. Stacy.

FIGURE 75.—Scarp area at the head of the Kirkwood earthflow. Shows maximum distension of the surface, toppled trees, soil destroyed, and much subsoil. View eastward from the west edge of the flow.

the slumped mantle seems to have been deeper than usual and to have crossed under the crest to the slope opposite that which moved.

Scarps and fissures due to mantle slumps were seen on nearly every major ridge visited within several miles of Hebgen Lake and as far as Cliff and Wade Lakes, Coffin Mountain, and the Lion Head Ski area, 1½ miles northwest of U.S. Highway 20 west of West Yellowstone. Unusually large scarps and fissures extend for nearly a mile along the ridge east of the Madison Slide. In some places they indicate downslope movement of several feet.

Most of this movement appears to have been absorbed by compaction of the mantle lower on the slopes, and rarely are complementary compressional features seen. In two places, however, notable downslope effects of mantle slumping occurred. One is on State Route 499 just west of Corey Spring; the other is at Hebgen Dam, which was considerably damaged by such movements.

UNDERTHRUSTING BY DOWNSLOPE MASS MOVEMENT

By WARREN HAMILTON

An excellent opportunity to study in detail the effects of mass downslope movement of surficial materials at the base of a steep slope was provided by a fortunate combination of natural and manmade features a quarter of a mile west of Corey Spring in the Teepee Creek quadrangle. Here, State Route 499 and flanking board fences run along the break in slope between a moraine and the alluvial plain at the north side of Hebgen Lake. A new fault scarp, across which the throw was relatively down on the lake side, trends along the slope above the moraine 300 feet higher than the road.

The moraine is composed of relatively fine material, its clasts being mostly less than 6 inches in diameter, and has a thin soil cover. The slope at the base of the hill consists largely of transported soil at least

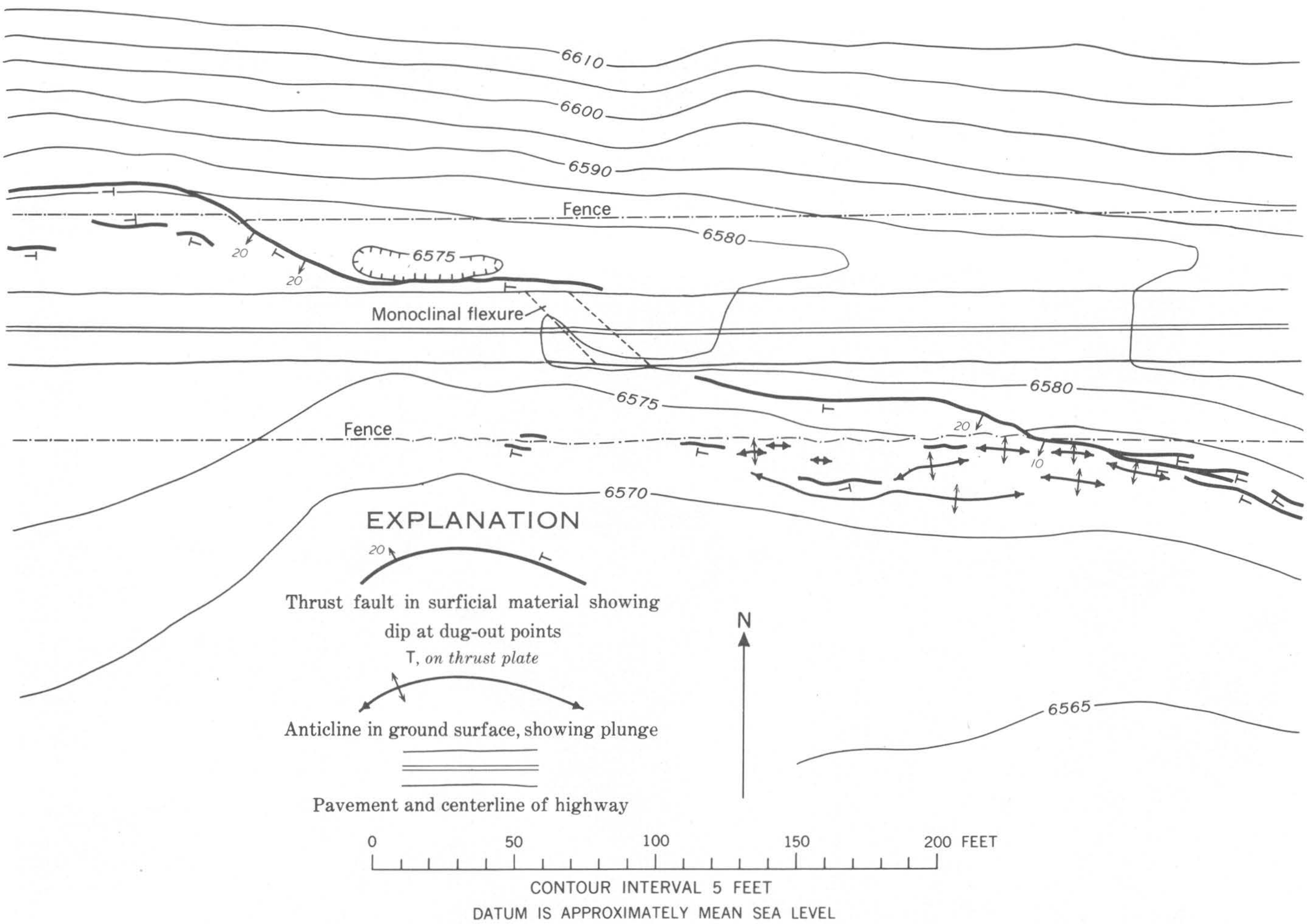

FIGURE 76.—Map of part of State Route 499 near Corey Spring on the north side of Hebgen Lake, showing folds and thrust faults in alluvium resulting from mantle slumping. Planetable map by Warren Hamilton, W. H. Hays, and William Quinlivan, September 3, 1959.

2½ feet thick, which is much thicker than that on the hill above. At one point on this slope, a 4-foot bulldozer cut exposed grayish-brown topsoil 8 inches thick, beneath which there is a smooth gradation to grayish-yellow-brown soil at the bottom of the cut; pebbles are sparse and small. A 2½-foot hole dug in the broad alluvial slope 100 feet south of the south fence did not penetrate the soil. The grayish-brown topsoil here is 8 inches thick and is underlain by reddish-brown crumbly clayey silt; there are no pebbles or visible sand.

Presumably bedrock beneath both hillside and flat dropped against the fault exposed higher on the hill but, in addition to this displacement, there was also much downslope shifting of surficial material. This was largely accomplished by pervasively distributed movement that caused few breaks in the surface. At the base of the morainal slope, the down-moving hillside material underthrust the much flatter alluvium. The compression of the alluvium was taken up partly on a single major thrust fault, partly on imbricate thrusts, and partly by folding of the alluvium (fig. 76).

The major thrust zone lies near the break in slope and crosses the road and the fences obliquely. Along the thrust, the morainal material plowed beneath the alluvium, carrying grass under for 2 feet normal to the strike and arching and variably crumpling the alluvium. The main fault was dug out in five places. At each of these the overlying alluvium lay on grass for 2 feet back from the trace, and beyond that the break continued downward at, or slightly steeper than, the angle of slope at that point. In the eastern part of the mapped segment, numerous imbricate thrusts developed; and two overlapping faults were seen also in one of the holes dug in the western part, with grass dragged under each, although the lower had been overridden and did not crop out.

Subsidiary short thrust faults developed in back of the main fault. The single backthrust, shown near

FIGURE 77.—Displaced pavement and fences along State Route 499 near Corey Spring. View west. Man is standing on crest of fold in pavement. Photograph by John R. Stacy.

the west side of figure 76, crops out at the upper edge of the roadside ditch beneath an unusually high position of the main thrust and was obviously a result of outward pushing of the upper plate there.

The highway was bent and offset by the fault as it went beneath the asphalt, but the road surface was not broken. The center line was displaced 3½ feet, and a double bend was formed in 30 feet of road length (fig. 77).

Many small anticlines were formed in the surface of the alluvium and are indicated on the map. These have structural heights ranging from a few inches to a little more than a foot. Possible broader folds were observed to the west, but the undulations there might have antedated the earthquake and are not shown on the map.

The fences on both sides of the highway were broken by the thrust fault, and the posts on opposite sides were offset directly across the fault strike—1.7 feet at the north fence, 2.1 feet at the south. There was almost no strike-slip component.

The character of the deformation suggests that the soil on the alluvium thrust and rumpled as a thin coherent cover above unconsolidated material, which in turn was thickened by the compression it received rather than deformed in the manner of the more coherent soil. The south fence was compressed throughout the length of the area mapped, and the compression indicated the component of the shortening due to rumpling of the alluvium. Posts are 9 feet apart, and the resultant postquake curvatures of the boards (that is, the measurements normal to the chords of the arcs now formed by the boards) ranged from a fraction of an inch to 14 inches. Despite the striking appearance of the fence (fig. 78), the shortening was not great; it ranged from 0.1 percent to 1 percent when

FIGURE 78.—Fence along south side of State Route 499 near Corey Spring, offset and compressed by hillside slump. View west. Shovel at lower right indicates location of underthrust fault in alluvium. Photograph by John R. Stacy.

FIGURE 79.—Hebgen Dam, view upstream from a point 0.3 mile northwest of the dam. Repair is underway, but the exposed east end of the core wall, the damaged spillway, and the eroded downstream face can still be seen. The lake shore, except near the dam, is submerged, and trees have been carried into the water by landslides. Photograph by John R. Stacy.

measured for 50-foot lengths, and averaged nearly 0.4 percent.

HEBGEN DAM

Hebgen Dam trends northeastward across the narrow upper part of the canyon of the Madison River below Hebgen Lake. One abutment rests against dolomite bedrock on the southwest side of the canyon and the other against an alluvial fan on the northeast side. The dam is constructed of earthfill on both sides of a concrete core 87 feet high that is footed in bedrock for most of its 700-foot length. Its east end rests on the fan and is interrupted by a spillway gate 80 feet wide. A concrete flume 1,000 feet long conducts the overflow from the gate across the fan and into the river channel considerably below the toe of the dam (fig. 79).

The new fault scarp of the Hebgen fault along the northeast side of the canyon passes within 1,000 feet of

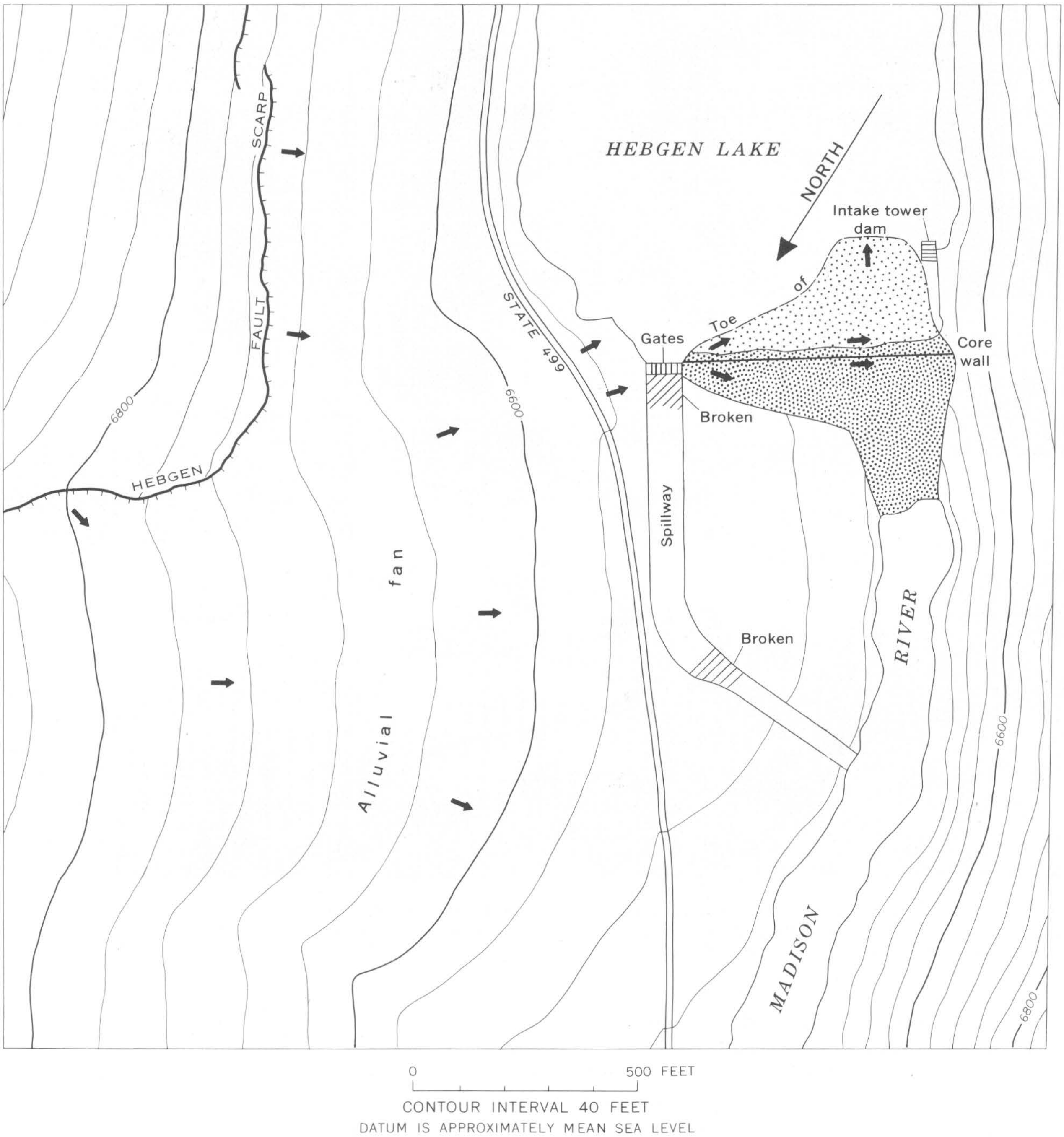

FIGURE 80.—Sketch map of Hebgen Dam and vicinity showing relations to the alluvial fan and the Hebgen fault scarp. Arrows indicate direction of surficial movements caused by the earthquake.

the east end of the dam but deviates upslope around the upper end of the fan. This indicates that most of the fan participated in the subsidence of the downdropped block (fig. 80; see also fig. 20).

During the earthquake, the spillway and both embankments of the dam subsided and were shifted southwestward relative to the core wall, which alone remained fixed to the bedrock at the southwest end of the dam. The southwest end of the spillway gate moved 1.8 feet past the end of the core wall, which in turn was pushed upstream 2.8 feet and dropped nearly a foot along several vertical tension fractures 1 to 6 inches wide (fig. 81). Both embankments sank and exposed the core wall, which was originally cov-

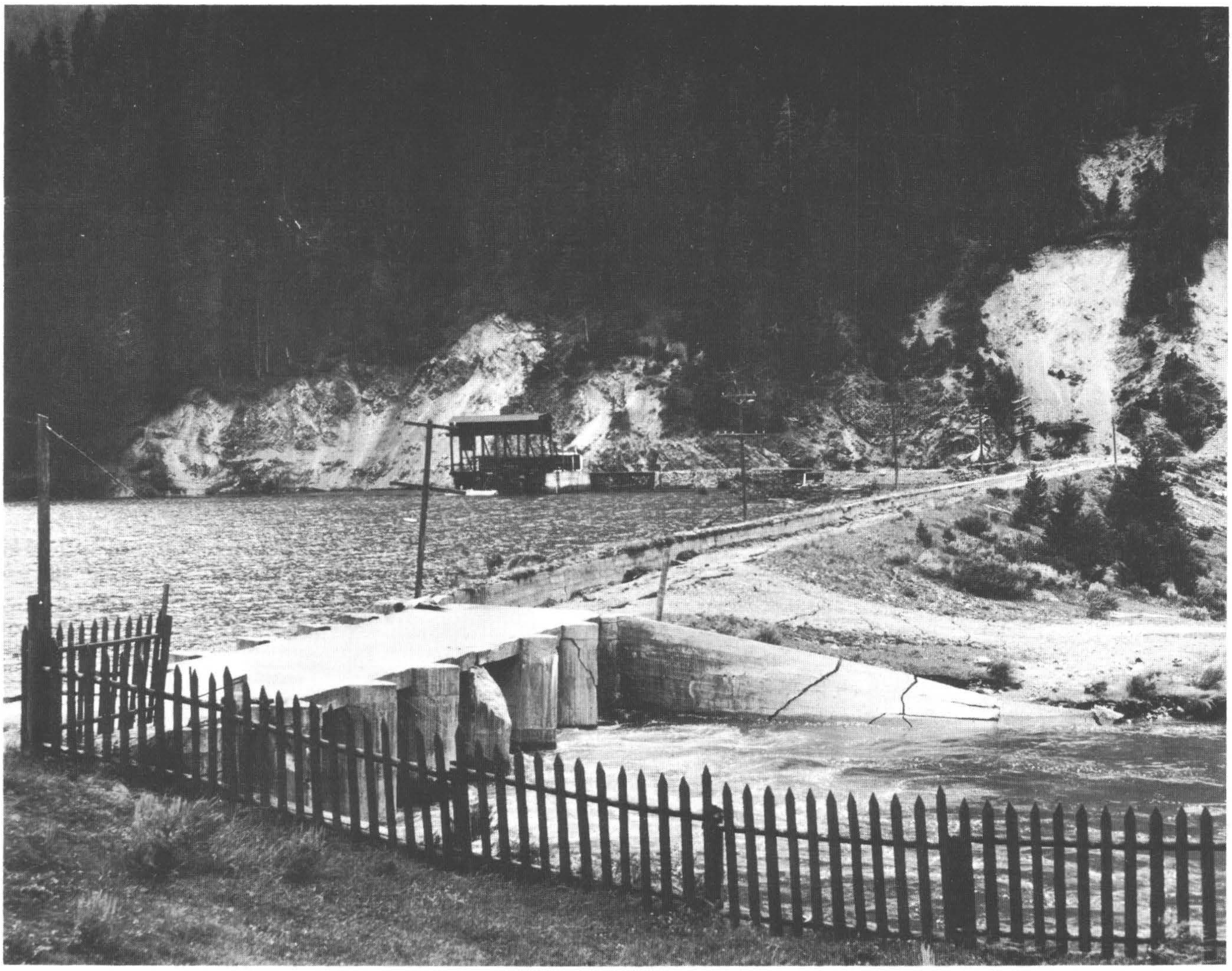

FIGURE 81.—The crest of Hebgen Dam seen from the northeast abutment the day after the earthquake. Visible are the sunken downstream embankment, the bared core wall and its bent east end, and the broken spillway gates and cover. The structure at the opposite shore is the intake tower. Photograph by I. J. Witkind.

ered by 8 or 10 inches of dirt and sod flush with the top of the embankments. The subsidence was greatest in the eastern parts of both embankments and decreased westward to a few inches at the west end of the core wall. The downstream embankment sank a maximum of 2.4 feet except immediately adjacent to the spillway gates. The upstream embankment sank in places 5 feet below the top of the core wall, and its easternmost part moved 3½ feet southwestward and caused gaps of this magnitude in the concrete wing wall attached to the core at the southwest side of the spillway gates. Scratches made by the larger rock fragments in the embankment against the newly exposed face of the core wall show that the embankment moved at first more horizontally than vertically and that its west end actually moved upward against the canyon wall. These scratches are 15 inches long at the east end of the dam and decrease to an inch or two at the west end, thus indicating that the horizontal movement decreased regularly from east to west. Subsequently, the embankment sank, exposing the core wall, and spread laterally beneath the lake. A detailed resurvey of the dam by the Montana Power Company shows that parts of the upstream toe of the dam rose several feet and that the surface of the fan between the spillway and the river channel had bulged as much as 5 feet above its former position.

It seems clear from these observations that much of the damage to the dam was caused by massive downslope movement of unconsolidated material above the northeast abutment, set in motion by abrupt subsidence on the Hebgen fault. Only in this way can the compression of the spillway gates and the strongly horizontal initial movement of the embankments be accounted for. Water-filled material flowed out from beneath the upstream part of the dam, probably in a manner similar to the landslides along the highway, and caused additional settling of the upstream embankment.

CHURNED GROUND

Surficial disturbances quite different from the gravitational effects previously described were seen at several points in the epicentral area of the Hebgen Lake earthquake. These are the phenomena sometimes referred to as churned ground, in which parts of the ground and objects resting on it are thrown into the air or overturned with little horizontal displacement. Such dislocations seem to result from ground vibrations whose vertical accelerations approach or exceed that of gravity. They are not commonly observed or recorded, but they apparently occurred in the great Indian earthquake of 1897 (Richter, 1958, p. 50–51) and in the Cedar Mountain earthquake of 1932 (Gianella and Callaghan, 1934, p. 2).

In the Hebgen Lake area, churned ground or other evidence of unusually strong vertical acceleration was found at six localities within 3 miles of Hebgen Dam. They were all found in the course of other earthquake investigations and no special search for churned ground was made. At each locality, outcrops of fractured bedrock were broken up and the fragments thrown about so as to merit the term "exploded outcrop". The outcrops are all of strong bedrock units, including the Madison group, Kootenai formation, Shedhorn sandstone, and Precambrian quartzite, and all were found on ridge crests or other topographic eminences.

One area of churned ground, reported by Warren Hamilton, is a small bedrock knob 20 to 30 feet high of Precambrian quartzite at an altitude of 6,820 feet near the southwest shore of Hebgen Lake, 1.45 southeast from the dam in the SE¼ sec. 26, T. 11 S., R. 3 E. The quartzite is much jointed, and blocks were jostled about; many were overturned, and many previously in place were thrown out of their sockets. Several fallen trees on the knob were bounced severely and moved several feet; their courses are marked by bark and shredded wood left on the rock. Neither nearby talus nor other outcrops on the hillside showed such effects.

A larger area was seen at the east end of Boat Mountain near the center of sec. 3, T. 11 S., R. 3 E., 2.7 miles north-northwest from Hebgen Dam. As described by W. H. Nelson,

> The ridge crest here is nearly flat and covered with frost-heaved limestone blocks a few inches to 1½ feet in diameter of the underlying Madison group, which dips 60° SW and makes up the greater part of the ridge. The high point of the ridge and an area extending for 100 feet along the crest in either direction has been severely churned. Ninety-five percent of the limestone fragments have been overturned so that their gray weathered upper sides are underneath and their yellowish undersides, formerly buried, are now on top. Prior to the earthquake, this part of the ridge had a cover of small Alpine plants growing in soil a few inches thick. Most of this soil is now inverted so that the plant roots extend upward into the air and the leaves are underneath the soil.

A similar locality that is not so severely churned is on the crest of Hebgen Ridge 0.8 mile northeast of Hebgen Dam.

Two exposures of the basal sandstone of the Kootenai formation were also churned. One of these is on the ridge between Cabin Creek and its lowest major tributary in the SE¼ sec. 11, T. 11 S., R. 3 E. The sandstone, which dips 35° SW, here makes a small rocky knob where it crosses the ridge. Several joint-bounded blocks of the sandstone as much as 2 feet in diameter were thrown or rolled from deep sockets on the side of the knob. The other locality is in the SW¼ sec. 13, T. 11 S., R. 3 E., on the divide between Cabin Creek and Kirkwood Creek, where the sandstone on a knob at an altitude of 8,580 feet has been jostled and thrown about in much the same way.

Shedhorn sandstone dipping 35° NW. on a ridge crest in the NE¼ sec. 24, T. 11 S., R. 3 E., 1.8 miles due east of Hebgen Dam, was also jostled, and a few blocks were overturned.

The bounced trees near Hebgen Lake and the overturned soil on Boat Mountain seem to be evidence that the ground motion in these places acquired a vertical acceleration exceeding that of gravity. In the other localities cited, horizontal and vertical accelerations less than gravity may have combined to dislodge the blocks of sandstone from the sides of the knobs. No areas underlain by thick soil, other surficial material, or even the weaker bedrock formations are known to have been churned, and it is unlikely that ground accelerations in most of those areas were more than a small fraction of gravity. This is in accordance with Richter's surmise (1958, p. 60) that even the strongest seismic waves lose much of their energy where they emerge in thick unconsolidated rocks. The restriction of the churned areas to small isolated parts of the many bedrock exposures in the epicentral area suggests not only that seismic waves traveled along strong bedrock units and emerged at the surface with little loss of energy, but also that they may have been reinforced locally by interference of two or more waves.

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Part 2

Hydrologic Effects

Effects of the Hebgen Lake Earthquake on Surface Water

By FRANK STERMITZ

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

GEOLOGICAL SURVEY PROFESSIONAL PAPER 435-L

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CONTENTS

EFFECTS OF THE HEBGEN LAKE EARTHQUAKE ON SURFACE WATER

By Frank Stermitz

ABSTRACT

The tremor of August 17, 1959, and some subsequent tremors were noted on the water-level recording charts at nearby and distant gaging stations. Waves of water were forced out of the Madison River by ground motion and by the landslide which created Earthquake Lake. Rhythmic waves of water overtopped Hebgen Dam; the highest had a peak discharge of about 14,000 cfs (cubic feet per second). The capacity of Hebgen Lake increased about 9,000 acre-feet with respect to the datum of the dam. Prompt increases in discharge of 0.5 to 0.2 cfs per square mile occurred on many streams in the earthquake area, and subsequent seasonal recession was less than usual.

INTRODUCTION

The increase in the base flow of streams after the earthquake of August 17 was noteworthy, and the sudden damming of the Madison River presented a problem in water management that required the collection of surface-water data. These data were obtained from existing stream-gaging stations and from stations established at new sites. The U.S. Army Corps of Engineers and the Montana Power Co. aided in the collection of hydrologic data.

SURFACE DRAINAGE

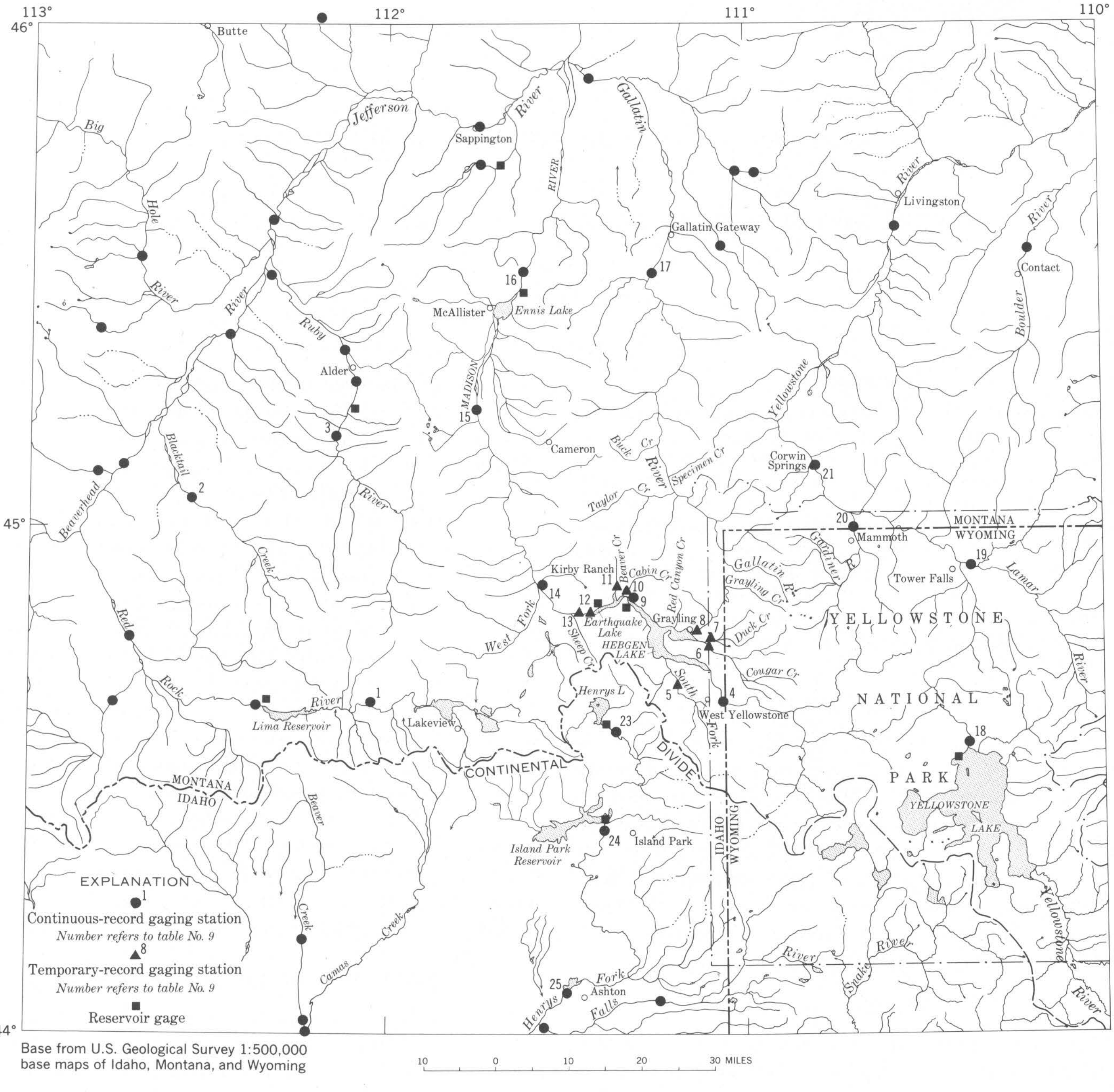
The Continental Divide, which separates the Missouri and Snake River basins, crosses high plateaus in the region of the severe earthquake disturbance. The Madison River drains the central part of the earthquake area (fig. 82); the Jefferson and Gallatin Rivers, which combine with the Madison River at Three Forks, Mont., to form the Missouri River, drain most of the western part; the Yellowstone River drains the area to the north and east; and the Snake River drains the southern part of the area.

Precipitation is relatively well distributed throughout the year, but July to September is the driest period. The annual precipitation, which averages more than 20 inches, falls mainly in the form of snow. At West Yellowstone, the average depth of the snow on April 1 is 37 inches, and the content of water is 12 inches. The Twenty-One Mile snow course, which lies at an altitude of 7,150 feet near the Madison-Gallatin divide, has an average depth of snow of 53 inches and an average water content of 18 inches on April 1. High base flows are the result of permeable glacial deposits, broad mountain meadows, and lakes and springs. Hydrographs of the May to October period of streamflow are shown in figure 83. The Madison River near West Yellowstone drains an area of 419 square miles, and the average annual runoff is about 15 inches. The Gallatin River near Gallatin Gateway drains an area of 828 square miles, and the average annual runoff is 12 inches. The more subdued peak and more gradual decrease in streamflow of the Madison River are indicative of the relatively greater ground-water storage in that basin. By late October the Madison River generally reaches a stable flow that continues until the spring rise. The decrease of streamflow in the Gallatin River normally continues through the winter.

COLLECTION OF DISCHARGE DATA

The collection of streamflow and reservoir data was aimed at obtaining information that would be immediately useful in solving the problem of water management that resulted from the damming of the Madison River about 6 miles downstream from Hebgen Lake. Some of the data collected were useful in determining the effects of the earthquake upon streamflow; but it was not possible to obtain data for all streams that may have been affected, and antecedent information was somewhat inadequate for comparisons.

The gaging-station records listed in table 9 are those considered pertinent to this report. Temporary gages installed on streams directly tributary to Hebgen Lake and to the lake formed by the landslide were



Base from U.S. Geological Survey 1:500,000 base maps of Idaho, Montana, and Wyoming

Figure 82.—Map showing surface drainage, and locations of gaging stations.

TABLE 9.—*Gaging-station records pertinent to observation of discharge effects of Hebgen Lake earthquake*

Number in fig. 82	Gaging station	Location[1]			Drainage area (sq mi)	Type and period of record	Remarks
		Lat N.	Long W.	Description			
1	Red Rock River at Kennedy Ranch near Lakeview, Mont.	44°39′	112°39′	14 miles NW of Lakeview	323	Recorder, seasonal 1936	Diversion for 6,000 acres. Some regulation possible.
2	Blacktail Creek near Dillon, Mont.	45°03′	112°03′	12½ miles SE of Dillon	312	Recorder, 1946	Diversion for 4,000 acres. Many springs.
3	Ruby River above reservoir, near Alder, Mont.	45°11′	112°09′	10 miles S of Alder	538	Recorder, 1938	Diversion for 3,000 acres. Some springs.
4	Madison River near West Yellowstone, Mont.	44°39′	111°04′	1½ miles E of West Yellowstone.	419	Recorder, 1913	No diversions. Many springs.
5	South Fork Madison River near West Yellowstone, Mont.	44°41′	111°12′	5 miles W of West Yellowstone.	110	Intermittent staff readings, Aug. 28–Oct. 22, 1959.	Many swamps and springs. Poorly defined surface drainage.
6	Cougar Creek near West Yellowstone, Mont.	44°46′	111°07′	7 miles N of West Yellowstone.	90	Intermittent staff readings, Aug. 26–Oct. 22, 1959.	Do.
7	Duck Creek near West Yellowstone, Mont.	46°47′	111°07′	8 miles N of West Yellowstone.	25	Intermittent staff readings, Aug. 26–Oct. 23, 1959.	Do.
8	Grayling Creek near West Yellowstone, Mont.	44°47′30″	111°08′	10 miles N of West Yellowstone.	80	do	Some swamps and springs.
9	Madison River below Hebgen Lake, near Grayling, Mont.	44°52′00″	111°20′15″	1,500 ft downstream from Hebgen Dam.	904	Recorder, 1938	Regulated by Hebgen Dam. Daily lake level for this study
10	Cabin Creek near West Yellowstone, Mont.	44°52′	111°21′	At State Highway 499	30	Intermittent staff readings, Aug. 25–Oct. 23, 1959.	Direct tributary to landslide lake.
11	Beaver Creek near West Yellowstone, Mont.	44°52′	111°21′30″	do	40	Intermittent staff readings, Aug. 26–Oct. 23, 1959.	Do.
12	Madison River below landslide lake.	44°49′40″	111°26′	At downstream toe of slide		Staff and recorder readings, Aug. 26–Sept. 12, 1959.	Installed for observation of landslide dam seepage.
13	Madison River above Sheep Creek, near West Yellowstone, Mont.	44°49′40″	111°26′40″	½ mile below slide		Staff and recorder readings, Sept. 8–Oct. 3, 1959.	Installed for discharge control during spillway changes.
14	Madison River at Kirby Ranch, near Cameron, Mont.	44°53′	111°34′	22 miles S of Cameron and above West Fork.		Staff readings Aug. 31, Oct. 2, 1959; Recorder Oct. 3, 1959.	Installed for sediment data and better discharge data.
15	Madison River near Cameron, Mont.	45°14′00″	111°45′00″	4 miles NW of Cameron	1,669	Recorder, 1951–58; Aug. 26, 1959	Diversion for 5,300 acres.
16	Madison River below Ennis Lake, near McAllister, Mont.	45°29′25″	111°38′00″	1½ miles downstream from Ennis Dam.	2,186	Recorder, 1938	Regulated for power. Daily lake level for this study. Diversion for about 23,000 acres.
17	Gallatin River near Gallatin Gateway, Mont.	45°30′	111°16′	8 miles S of Gallatin Gateway.	828	Recorder, 1889–94; 1930	Diversion for 1,400 acres.
18	Yellowstone River at Yellowstone Lake Outlet, Yellowstone National Park.	44°34′	110°23′	¼ mile downstream from outlet.	1,010	Recorder, 1926	No regulation or diversion.
19	Lamar River near Tower Falls ranger station, Yellowstone National Park.	44°56′	110°22′	¾ mile upstream from mouth.	640	Recorder, 1922	Do.
20	Gardiner River near Mammoth, Yellowstone National Park.	45°00′	110°42′	1½ miles N of Mammoth	200	Recorder, 1938	Do.
21	Yellowstone River at Corwin Springs, Mont.	45°07′	110°48′	8 miles N of Gardiner	2,630	Recorder, 1910	Upstream diversion for about 960 acres.
22	Snake River at Moran, Wyo.	43°51′	110°35′	1,000 ft downstream from Jackson Lake Dam.	824	Recorder, 1903	Flow regulated for irrigation.
23	Henrys Fork near Lake, Idaho.	44°36′	111°21′	4 miles S of Lake	98	Recorder, 1920	Excessive regulation. Hides earthquake effects.
24	Henrys Fork near Island Park, Idaho.	44°24′59″	111°23′41″	1 mile W of Island Park	481	Recorder, 1933	Flow regulated for irrigation at Henrys Lake and Island Park Reservoir.
25	Henrys Fork near Ashton, Idaho.	44°05′	111°30′	3 miles W of Ashton	1,040	Recorder, 1902–08; 1926–55	Flow regulated as above. Diversion for about 18,000 acres.

[1] Locations given with reference to local landmarks.

read intermittently, and the discharge measurements and gage readings provided fairly accurate records of trends of the flow of these streams. Temporary recording stations were also established on the Madison River immediately below the landslide dam, above Sheep Creek (near the slide), and at the bridge at the Kirby Ranch (formerly Hutchin's Ranch) to observe flow through and over the landslide dam and incremental tributary flows.

IMMEDIATE EFFECTS OF EARTHQUAKE

RANGES OF STAGE CAUSED BY GROUND MOTION

The major earthquake at about 11:38 p.m. on August 17 and some of the aftershocks that continued through August 18 were noted by water-level recorders at stream-gaging stations in the general area and, to some extent, by those hundreds of miles from the epicenter. Sensing elements of water-level recorders are copper floats about 10 inches in diameter which rest on the water surfaces of stilling wells. The wells are generally rectangular with water-surface areas ranging from 9 to 25 square feet. Motion of the floats is transmitted by lines or tapes to float pulleys on horizontal shafts, and rotation of a shaft directs the stylus of the recorder. The rise and fall indicated by the recorder can be a crude measure of the intensity of the earth shock. Several variables may be encountered in measuring the earth shock by this means; these are: the relative position of the float in the well, the size of the well, the depth of water in the well,

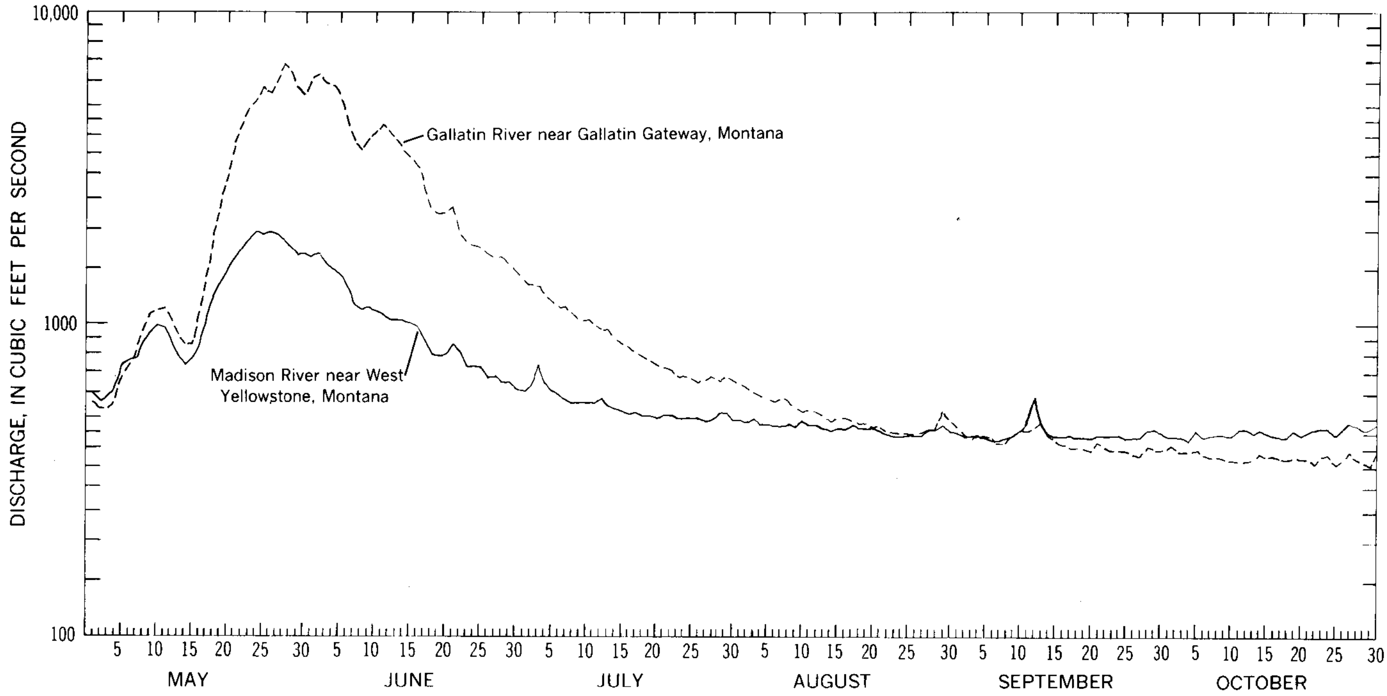


FIGURE 83.—Hydrographs of Gallatin and Madison Rivers during part of 1956.

the plane of the float pulley with respect to the direction of the ground motion, the frictional resistance of individual recorders, the duration of the tremor, and the foundation material. The time scale of a recorder chart is usually 2.4 inches per day, which, with the uncertainties of time correction to the chart, limits the accuracy of timing lesser shocks.

The water-level chart for the Madison River near West Yellowstone (table 9, No. 4) recorded the major shocks of about 11:38 p.m. on August 17 and 10 aftershocks on August 18. The vertical fluctuations of the traces were: August 17, 11:38 p.m., 1.07 feet; August 18, 12:05 a.m., 0.15 foot; 12:25 a.m., 0.07 foot; 1:15 a.m., 0.06 foot; 1:45 a.m., 0.12 foot; 3:45 a.m., 0.03 foot; 4:05 a.m., 0.06 foot; 8:25 a.m., 0.18 foot; 4:45 p.m., 0.04 foot; and 9:00 p.m., 0.01 foot. This gaging station is 1½ miles east of West Yellowstone and 17 miles southeast of Hebgen Dam. The chart for the Yellowstone River at Corwin Springs, Mont. (table 9, No. 21), about 30 miles northeast of West Yellowstone, showed a vertical range of 0.34 foot for the major shock of August 17. The chart also showed that nine aftershocks which occurred on August 18 had approximately the same relative magnitude as those recorded at the Madison River station near West Yellowstone. The chart for the Madison River below Hebgen Lake, near Grayling (table 9, No. 9; and fig. 84), showed that the major shocks were so rapid and violent that the pen failed to ink a continuous line and that some time slippage of the chart occurred. The total time slippage by 8:25 a.m. on August 18 was about 3 hours. Surges of river discharge caused by waves in Hebgen Lake were so pronounced and numerous that they obscured any recordings of separate shocks until the severe tremor of 8:25 a.m. on August 18. Mr. George Hungerford, the attendant at Hebgen Dam, states that the discontinuous trace on the recorder chart was noted during a brief inspection of the stream gage shortly after the major shock on August 17. At that time, he and Mr. Lester Caraway, his assistant, saw a wave of water coming down the river channel toward the gage and left hurriedly, assuming that Hebgen Dam had failed.

At the gaging station on Henrys Fork near Lake, Idaho, 13 miles southwest of West Yellowstone, the tremors were so severe that the float wire was disengaged from the float wheel, and record was terminated. Another gaging station, on the Snake River at Moran, Wyo. (table 9, No. 22; and fig. 82), showed pronounced movement of the recorder stylus. This station, which is just downstream from Jackson Lake Dam and about 90 miles south of West Yellowstone, showed a stage range of 1.23 feet at 11:40 p.m. on August 17, 0.51 foot at about 1:00 a.m. on August 18, and 0.51 foot at about 8:20 a.m. on August 18. A range of stage of 0.70 foot was recorded at the gaging station on the Red Rock River at Kennedy Ranch near Lakeview, Mont. (table 9, No. 1), about 47 miles

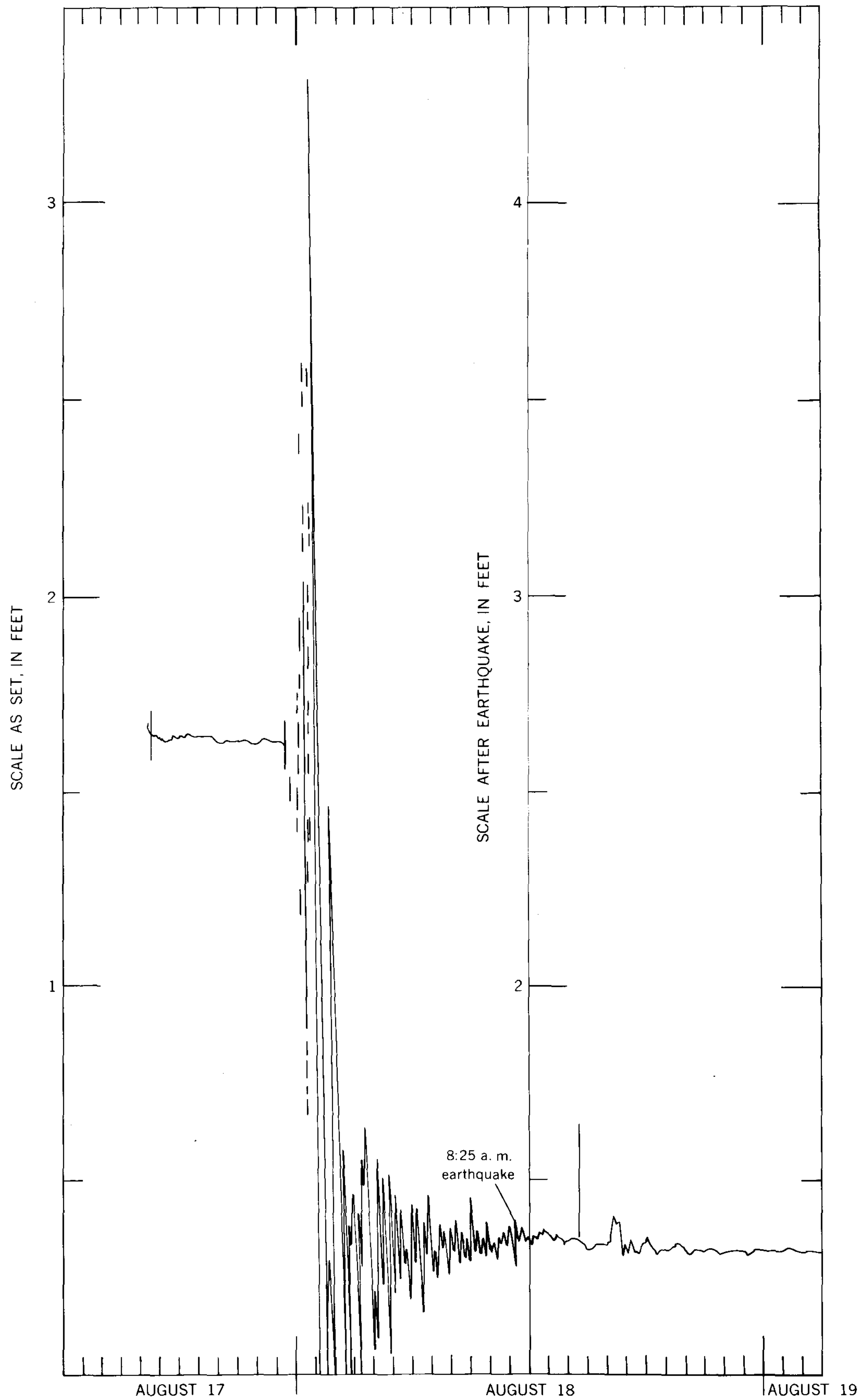


FIGURE 84.—Water-stage recorder chart from gaging station below Hebgen Dam showing violent movement of pen in response to the earthquake and aftershocks.

TABLE 10.—*Range of stage recorded in stilling wells at selected gaging stations as the result of earth tremors of about 11:38 p.m., Aug. 17, 1959*

Gaging station	Distance and direction from West Yellowstone: Miles	Distance and direction from West Yellowstone: Direction	Range (ft)
Red Rock River at Kennedy Ranch near Lakeview, Mont.	45	W	0.70
Blacktail Creek near Dillon, Mont.	75	W	.20
Ruby River above reservoir near Alder, Mont.	65	NW	.28
Ruby River near Twin Bridges, Mont.	85	NW	.15
Jefferson River near Twin Bridges, Mont.	90	NW	.05
Jefferson River near Sappington, Mont.	85	N	.14
Madison River near West Yellowstone, Mont.	1½	E	1.07
Madison River below Hebgen Lake, near Grayling, Mont.	19	NW	(1)
Madison River below Ennis Lake, near McAllister, Mont.	65	NW	.03
Gallatin River near Gallatin Gateway, Mont.	60	N	.04
East Gallatin at Bozeman, Mont.	70	N	.13
Missouri River at Toston, Mont.	100	N	.11
Missouri River near Ulm, Mont.	190	N	.21
Marias River near Shelby, Mont.	260	N	.02
Missouri River below Fort Peck Dam, Mont.	330	NE	.10
Milk River at Milk River, Alberta	320	N	.03
Willow Creek near Glasgow, Mont.	320	NE	.04
Yellowstone River at lake outlet, Yellowstone National Park.	35	E	.06
Lamar River near Tower Falls ranger station, Yellowstone National Park.	40	NE	.47
Gardiner River near Mammoth, Yellowstone National Park.	30	NE	.22
Yellowstone River at Corwin Springs, Mont.	35	NE	.34
Yellowstone River near Livingston, Mont.	70	N	.04
Boulder River near Contact, Mont.	75	NE	.02
Sunlight Creek near Painter, Wyo.	80	E	.02
Clarks Fork Yellowstone River at Chance, Mont.	100	E	.17
Yellowstone River at Billings, Mont.	150	NE	.08
Wind River near Dubois, Wyo.	95	SE	.02
Shoshone River at Byron, Wyo.	130	E	.05
Bighorn River at Kane, Wyo.	200	E	.02
Yellowstone River at Miles City, Mont.	280	NE	.05
Powder River at Arvada, Wyo.	250	E	.01
Snake River at Moran, Wyo.	90	S	1.23
Pacific Creek near Moran, Wyo.	65	S	.06
Snake River near Irwin, Idaho	90	S	.22
Henrys Fork near Lake, Idaho	13	SW	(1)
Henrys Fork near Island Park, Idaho	23	SW	.20
Henrys Fork near Ashton, Idaho	45	SW	.17
Teton River near St. Anthony, Idaho	55	SW	.18
Great Western Canal near Idaho Falls, Idaho	95	SW	.78
Medicine Lodge Creek at Ellis Ranch, near Argora, Idaho.	75	W	.16
Big Lost River near Arco, Idaho	130	SW	.04
Malad River near Gooding, Idaho	220	SW	.01
Snake River at King Hill, Idaho	240	SW	.17
Snake River at Weiser, Idaho	300	W	.12
Salmon River at Salmon, Idaho	140	W	.04
Clearwater River at Spalding, Idaho	300	SW	.04
Kootenai River at Libby, Mont.	340	NW	.04
Clark Fork below Missoula, Mont.	210	NW	.10
Clark Fork at St. Regis, Mont.	270	NW	.05
Flathead River near Columbia Falls, Mont.	290	NW	.50
Flathead River at Columbia Falls, Mont.	300	NW	.02
Clark Fork near Plains, Mont.	270	NW	.04
Green River at Warren Bridge, near Daniel, Wyo.	120	S	.03
Blacks Fork near Green River, Wyo.	220	S	.01

[1] Not determined; chart indicates violent action.

west of West Yellowstone. Other distinctive ranges of stage were recorded at the Great Western Canal near Idaho Falls, Idaho (0.78 foot), and at the Flathead River near Columbia Falls, Mont. (0.50 foot), about 90 and 300 miles, respectively, from West Yellowstone.

A tabulation of the recorded range of stage at selected gaging stations is given in table 10. The wide ranges of stages with respect to distance from the epicenter of the earthquake seem to be more indicative of the physical arrangements in the gage stilling wells and the relative plasticity of the materials underlying the wells than to the local severity of the tremor.

PROPULSION OF WATER FROM STREAMS

The severity of the tremors at 11:38 and 11:46 p.m. on August 17 may have generated violent waves in streams or lakes. A reasonably distinct water mark observed on August 21 on the river bank at the gaging station on the Madison River below Hebgen Dam was a good indication of the magnitude of wave action. This mark, about 6.0 feet above gage datum, was 4.4 feet higher than the stage immediately preceding the tremors. At a gage height of 5.4 feet, bent grass and bits of debris were considered to be evidence of downstream flow. Waves of propulsion from the Madison River may have exceeded the waves caused by the rockslide for a few miles downstream from Hebgen Dam. (See chapter K.) An examination of the riverbank on August 28 in the vicinity of the stream-measuring cableway, about a quarter of a mile downstream from the gaging station, disclosed inconclusive evidence of ejection waves. Many drift marks of subsequent discharge waves that resulted from flow over Hebgen Dam were plain, however. Indirect reports that campers on the low riverbank near the cableway were struck by a wave of water almost at the time of the tremors indicate the occurrence of ejected waves before the arrival of the first discharge wave over Hebgen Dam. Supporting these observations is the report that approximately half the volume of water in the weighing bucket of the tipping-bucket rain gage near Hebgen Dam was splashed out by the tremors.

WAVES OVER HEBGEN DAM

A rhythmic wave pattern was induced in Hebgen Lake by the severe earthquake of August 17. The early waves overtopped Hebgen Dam, and subsequent waves of lesser intensity broke over the spillway for hours. The peak discharge of the greatest wave recorded at the gaging station has been computed by indirect methods to be 10,200 cfs (cubic feet per second).

Messrs. George Hungerford and Lester Caraway, attendants at Hebgen Dam, and their wives, were the only persons who saw the waves go over the dam, and their recollections are naturally hazy because of the tremendous mental stress at that time. About 11:50 p.m. the two men reached the nearby recording gage and noted that the level of the stream was unexpectedly low. Mr. Hungerford recalls that the water-stage recorder pen, observed by flashlight, had traced only discontinuous lines since the tremors began. They could not have been at the gage more than a minute or two when they saw a wave of water approaching rapidly. Mr. Caraway judged that it may have been 5 feet high. They ran to the road about 100 feet away because they assumed the dam was failing; and, with their wives, they hurried along the road toward the dam, which is about 1,500 feet upstream from the river gage. It may have been as early as 11:55 p.m. when they reached the dam. The dam had been wetted by overflow. They also noticed that some of the concrete core wall of the dam was newly exposed on the downstream side. Shortly after their arrival they saw the lake level gradually swell, or rise, and overtop the dam by an estimated 2 to 4 feet. Mr. Caraway said that the swell appeared to increase in height just as it met the upstream side of the core wall, and then plunged over. The first observed overtopping may have continued for a few minutes, and was followed by a gradual lowering of the lake level to the point where much of the upstream face of the dam was exposed. The two men are in reasonable agreement that they saw 3 or 4 swells of diminishing intensity before the water was contained in the spillway. They estimated the time interval between swells to have been about 10 minutes and acknowledged that their sense of passage of time may have been greatly disturbed.

The peak discharge at the dam was computed at 14,000 cfs and is the sum of the calculated flow over the dam, through the spillway, and over the stop logs of the outlet structure. The flood marks used for this determination were clearly defined along the right, or north, shore above the dam and less definitely on the left, or south, shore. Assumptions concerning the height of the earth cover on the core wall of the dam and the exact position of the stop-log crests in the spillway and outlet structures may further detract from the reliability of results. The determination is classified as poor.

A slope-area measurement by F. C. Boner indicates a peak discharge of 10,200 cfs near the gaging station. A two-section reach of 188 feet, with its lower section about 200 feet upstream from the recording gage, was used. The flood marks on the right bank were reasonably definite but indicated a rather irregular water slope. Flood marks on the rocky left bank were few and irregular; therefore, the resulting discharge computation is classified as poor. A logarithmic extension of the stage-discharge relation at the gaging station also indicated a peak discharge of about 10,000 cfs.

Much of the reduction in peak discharge between 14,000 cfs at the dam and 10,200 cfs at the gaging station can be accounted for by channel storage in the reach. Farther downstream, reduction in peak discharge would have been even more pronounced.

The water-stage recorder chart for the Madison River below Hebgen Dam indicates that the interval between the lesser wave crests of the early morning of August 18 averaged about 15 minutes. (See fig. 84.) Reversals on the recorder graph at the baseline, and the probable tracing of aftershocks with unknown amounts of slippage caused by the shocks, made it impossible to determine the time interval between the early waves. In calculating the discharge for the period of unreliable record, it has been assumed that the first wave crest was the highest, that the subsequent waves were spaced at intervals of 15 minutes, and that the minimum river discharge between waves consisted of leakage of 50 cfs through the timber stop logs of the outlet structure.

EFFECTS OF ROCKSLIDE

The violent displacement of the water of the Madison River by the rockslide about 6 miles downstream from Hebgen Dam overturned camping trailers, bowled over tents, and caused human injuries and loss of life. Barked trees and debris were found scattered high along the riverbanks for half a mile below the slide. The upstream wave resulting from the landslide was probably dissipated more rapidly.

The result of the gouging and pulverizing action of the slide was the formation of a fairly impervious dam which stopped the flow of the Madison River for weeks. The abundance of spring inflow to the Madison River a few hundred yards downstream from the slide, however, kept the river alive.

APPARENT CHANGE OF CAPACITY OF HEBGEN LAKE

During the interval between 5:00 p.m. on August 17 and about 8:30 a.m. on August 18, the level of Hebgen Lake declined 0.74 foot, according to gage readings by Montana Power Co. personnel. Subsequent spirit levels indicated no apparent change in the altitude of the gage with respect to the outlet structure on the core wall of the dam near the left or south shore. This lowering of the lake represented a change in storage of 9,400 acre-feet, according to the capacity table of the company. Calculations indicate, however, that outflow during the time interval exceeded inflow by about 400 acre-feet, thus the lowering of the lake level reflected an increased storage capacity of about 9,000 acre-feet.

TABLE 11.—*Daily discharge, in cubic feet per second, at selected stations*

Date	(1) Red Rock River at Kennedy Ranch, near Lakeview, Mont.	(2) Blacktail Creek near Dillon, Mont.	(3) Ruby River above reservoir, near Alder, Mont.	(4) Madison River near West Yellowstone, Mont.	Computed inflow to Hebgen Lake (3-day average)	(9) Madison River below Hebgen Lake, near Grayling, Mont.	(13) Madison River above Sheep Creek, near West Yellowstone, Mont.	(14) Madison River at Kirby Ranch, near Cameron, Mont.	(15) Madison River near Cameron, Mont.	Computed inflow to Ennis Lake	(16) Madison River below Ennis Lake, near McAllister, Mont.	(17) Gallatin River near Gallatin Gateway, Mont.	(20) Gardiner River near Mammoth, Yellowstone National Park	Computed natural inflow to Island Park Reservoir, near Island Park, Idaho	Computed natural flow, Henrys Fork near Ashton, Idaho
August 1959															
1	48	39	106	378	469	1,000	----------	----------		1,480	1,340	794	147	----------	----------
2	46	45	114	378	412	1,000	----------	----------		1,450	1,350	830	155	----------	----------
3	46	49	118	384	438	989	----------	----------	[1] 1,220	1,440	1,330	838	159	----------	----------
4	43	44	114	378	594	989	----------	----------		1,330	1,330	766	143	----------	----------
5	43	36	110	378	851	989	----------	----------		1,460	1,150	738	136	----------	----------
6	43	31	114	371	810	1,020	----------	----------		1,470	1,330	719	133	540	----------
7	42	28	116	364	814	1,020	----------	----------		1,410	1,580	706	130	540	1,230
8	41	27	114	364	660	1,000	----------	----------	[1] 1,200	1,380	1,400	686	130	511	1,100
9	38	28	114	364	688	1,020	----------	----------		1,330	1,310	680	127	526	1,091
10	37	25	112	364	646	1,000	----------	----------		1,370	1,310	667	124	500	1,156
11	37	25	112	364	640	1,030	----------	----------		1,280	1,320	648	124	529	1,120
12	36	24	110	364	677	1,000	----------	----------		1,360	1,320	648	130	530	1,129
13	35	27	110	364	566	989	----------	----------		1,330	1,310	660	133	510	1,140
14	32	26	99	364	497	1,040	----------	----------	[1] 1,180	1,360	1,300	634	133	511	1,090
15	33	25	92	364	411	1,020	----------	----------		1,320	1,300	615	136	538	1,061
16	32	23	88	357	552	1,000	----------	----------		1,310	1,290	602	133	540	1,038
17	31	20	85	357	[2] 667	1,030	----------	----------		1,370	1,290	589	133	540	1,200
18	34	20	90	442	[2] 4,048	859	----------	----------	[1] 700	905	4,300	660	163	655	1,270
19	33	24	99	505	[2] 385	847	----------	----------	[1] 500	970	1,880	787	195	703	1,355
20	38	36	106	514	[2] 765	928	----------	----------	[1] 400	678	678	822	200	700	1,443
21	38	38	112	505	935	1,060	----------	----------	[1] 350	580	671	822	195	646	1,443
22	37	33	110	497	1,390	1,240	----------	----------	[1] 300	447	593	801	184	613	1,354
23	36	34	108	488	1,360	1,310	----------	----------	[1] 300	573	482	801	179	633	1,321
24	35	33	108	488	1,220	1,920	----------	----------	[1] 290	451	324	794	179	618	1,341
25	36	34	106	488	1,230	2,370	----------	----------	[1] 280	538	210	787	179	605	1,326
26	35	36	106	488	960	2,640	1.1	----------	280	483	210	787	179	604	1,339
27	36	36	110	488	1,110	2,970	2.5	----------	280	542	360	787	179	592	1,344
28	35	35	116	480	1,100	2,820	3.7	----------	280	586	586	766	175	592	1,342
29	36	36	116	473	1,050	2,740	9.4	----------	280	487	560	766	175	588	1,422
30	34	38	116	473	912	2,660	10	----------	280	679	515	766	175	595	1,398
31	35	38	116	480	960	2,580	13	148	275	472	508	766	175	631	1,355
September 1959															
1	34	38	116	480	1,100	2,500	12	139	280	577	522	773	175	588	1,411
2	34	38	108	473	1,170	2,410	15	139	285	559	541	752	175	585	1,378
3	36	38	110	473	951	2,350	17	139	290	534	534	745	175	571	1,355
4	38	36	110	473	902	1,690	18	166	290	655	528	732	171	567	1,381
5	41	36	112	480	867	1,300	19	155	297	528	528	745	171	585	1,527
6	40	39	120	480	999	969	29	148	318	528	528	759	171	673	1,625
7	40	38	131	473	1,020	770	[3] 57	175	325	528	528	732	171	614	1,583
8	41	35	129	473	939	770	108	184	346	522	522	719	171	624	1,624
9	34	35	118	465	956	770	110	184	360	588	515	712	167	595	1,624
10	36	35	118	465	942	781	271	416	384	663	554	712	167	578	1,595
11	40	33	116	465	980	792	690	868	725	785	658	700	167	----------	1,608
12	41	33	112	465	1,030	378	792	804	1,020	1,200	534	693	167	----------	----------
13	45	33	106	473	981	825	738	780	874	1,050	541	693	163	----------	----------
14	48	34	108	473	1,040	825	820	954	1,040	1,060	710	693	163	----------	----------
15	52	36	112	488	963	880	834	1,000	1,080	1,130	962	700	167	----------	----------
16	53	36	116	488	975	1,000	1,110	1,180	1,180	1,150	1,150	706	167	----------	----------
17	50	39	114	488	969	1,080	1,220	1,240	1,350	1,360	983	706	167	----------	----------
18	54	41	110	488	958	1,240	1,360	1,440	1,460	1,610	1,380	706	167	----------	----------
19	57	42	110	514	930	1,260	1,390	1,480	1,530	1,410	1,450	712	167	----------	----------
20	55	45	129	574	1,150	1,260	1,440	1,560	1,650	1,630	1,440	766	190	----------	----------
21	65	56	160	593	1,260	1,270	1,600	1,620	1,820	2,040	1,730	822	195	----------	----------
22	64	54	148	540	1,270	1,380	1,680	1,680	1,760	1,990	2,260	780	175	----------	----------
23	66	50	139	522	1,080	1,410	1,680	1,750	1,890	1,980	2,090	745	171	----------	----------
24	66	47	137	505	1,050	1,350	1,600	1,700	1,830	1,890	2,160	732	163	----------	----------
25	68	46	137	505	1,120	1,230	1,450	1,580	1,730	1,920	2,260	738	163	----------	----------
26	62	46	137	505	1,070	1,770	1,680	1,790	1,760	1,740	2,040	745	163	----------	----------
27	64	46	137	505	1,000	1,700	2,210	2,120	2,100	2,270	1,820	752	167	----------	----------
28	64	49	144	505	1,100	1,440	1,810	1,970	2,040	2,280	1,730	738	163	----------	----------
29	74	49	142	497	1,160	752	1,540	1,480	1,640	1,930	1,970	712	155	----------	----------
30	79	49	142	497	1,120	1,100	2,050	1,720	1,890	1,790	2,170	719	155	----------	----------

[1] Estimated on basis of records for Madison River below Hebgen Lake and relations in prior years.
[2] Not averaged with adjacent daily discharge; data for August 17–20 reflect changes in capacity of Hebgen Lake.
[3] Flow prior to September 8 near toe of slide with about 10 cfs inflow between sites.

TABLE 11.—*Daily discharge, in cubic feet per second, at selected stations*—Continued

Date	(1) Red Rock River at Kennedy Ranch, near Lakeview, Mont.	(2) Blacktail Creek near Dillon, Mont.	(3) Ruby River above reservoir, near Alder, Mont.	(4) Madison River near West Yellowstone, Mont.	Computed inflow to Hebgen Lake (3-day average)	(9) Madison River below Hebgen Lake, near Grayling, Mont.	(13) Madison River above Sheep Creek, near West Yellowstone, Mont.	(14) Madison River at Kirby Ranch, near Cameron, Mont.	(15) Madison River near Cameron, Mont.	Computed inflow to Ennis Lake	(16) Madison River below Ennis Lake, near McAllister, Mont.	(17) Gallatin River near Gallatin Gateway, Mont.	(20) Gardiner River near Mammoth, Yellowstone National Park	Computed natural inflow to Island Park Reservoir, near Island Park, Idaho	Computed natural flow, Henrys Fork near Ashton, Idaho
October 1959															
1	86	49	140	488	1,000	1,400	2,030	1,720	1,830	2,150	2,340	706	155	----------	----------
2	86	49	132	488	967	2,220	2,900	2,540	2,320	2,390	2,500	693	155	----------	----------
3	85	49	138	488	891	892	1,850	1,940	2,220	2,670	2,670	706	151	----------	----------
4	86	47	135	480	877	892	----------	1,510	1,670	1,960	2,470	712	155	----------	----------
5	88	47	135	473	900	904	----------	1,490	1,730	1,870	2,280	726	155	----------	----------
6	92	47	138	473	904	904	----------	1,510	1,620	1,910	2,340	732	159	----------	----------
7	----------	50	149	522	964	904	----------	1,950	2,080	2,250	2,800	759	163	----------	----------
8	----------	----------	149	514	986	916	----------	1,740	1,960	2,690	3,050	719	159	----------	----------
9	----------	----------	----------	522	994	916	----------	2,360	2,380	2,240	2,730	732	163	----------	----------
10	----------	----------	----------	531	1,050	928	----------	2,260	2,340	2,560	2,690	700	159	----------	----------
11	----------	----------	----------	514	1,080	928	----------	2,040	2,260	2,770	2,730	719	159	----------	----------
12	----------	----------	----------	548	1,140	940	----------	2,340	2,400	2,690	2,650	759	167	----------	----------
13	----------	----------	----------	531	1,060	952	----------	2,120	2,320	2,660	2,700	726	163	----------	----------
14	----------	----------	----------	522	1,010	952	----------	2,560	2,560	2,770	2,680	712	159	----------	----------
15	----------	----------	----------	522	975	952	----------	2,140	2,430	2,850	2,760	732	159	----------	----------
16	----------	----------	----------	531	1,020	964	----------	2,010	2,160	2,300	2,720	752	163	----------	----------
17	----------	----------	----------	522	1,020	964	----------	2,220	2,470	2,850	2,700	712	159	----------	----------
18	----------	----------	----------	505	1,210	964	----------	1,870	2,080	2,380	2,650	706	151	----------	----------
19	----------	----------	----------	514	1,140	2,040	----------	1,950	2,060	2,320	2,520	712	147	----------	----------
20	----------	----------	----------	514	1,150	2,680	----------	3,480	3,330	3,370	2,660	706	147	----------	----------
21	----------	----------	----------	514	----------	2,660	----------	3,890	4,090	----------	----------	706	147	----------	----------
22	----------	----------	----------	514	----------	2,860	----------	----------	----------	----------	----------	706	151	----------	----------

INCREASE IN STREAMFLOW

An increase in streamflow was recorded at a number of gaging stations in the general area of the earthquake as early as August 18. The Madison and Gallatin Rivers were particularly affected, and lesser effects were noted on the upper Red Rock and Ruby Rivers. Blacktail Creek near Dillon, Mont., a tributary of the Red Rock River, and Gardiner River at Mammoth, Wyo., a tributary of the Yellowstone River, showed marked and sustained increases. There was no apparent effect on the Yellowstone River except that on the Gardiner River tributary. An increase in the flow of the North Fork Musselshell River about 130 miles north of Hebgen Lake seems to have been larger than might result from measured precipitation. The appreciable increase in flow of streams in the Snake River basin seems to have been confined to Henrys Fork south of Hebgen Lake.

The natural regimen of many streams is disturbed by reservoir regulation and irrigation to the point that changes in flow caused by the earthquake might have been obscured. Records of discharge are collected on relatively few small streams, and substantial changes in flow of such streams may be unnoticed. Where current precipitation may be a factor, the effects of the earthquake could not be completely isolated. The effects of precipitation were varied, however, and did not, in many instances, mask the effects of the earthquake on streamflow. It is fortunate that most streams in the immediate area were at base-flow levels at the time of the earthquake, and that there was little precipitation for nearly a month after. The daily records of discharge from many streams in the area are given in table 11, and the hydrographs in figure 85 may aid in visualization.

As might be expected, the increases in the flow of streams attributable to the earthquakes (fig. 85) were most pronounced in the Madison River basin. Before the earthquake the flow of the Madison River near West Yellowstone was decreasing gradually, and its discharge was 357 cfs on August 16 and 17. An increase in discharge was noted about 3:00 a.m. on August 18, and by midnight it was about 500 cfs. (See fig. 85*A*.) The high rate of flow of 514 cfs on August 20 may have resulted partly from rain. Light, unrecorded rains may have caused slight deviations in the gradual decrease in streamflow that continued to mid-September, when heavier rains produced a general increase and slightly higher level of base flow. The difference in flow on August 17 and 19 represents an increase of 41 percent in discharge, or 0.35 cfs per square mile of drainage area. It is surprising that the increased flow diminished so slightly in the weeks following.

Further evidence of increased base flow is obtained from records of the reach between the Madison River

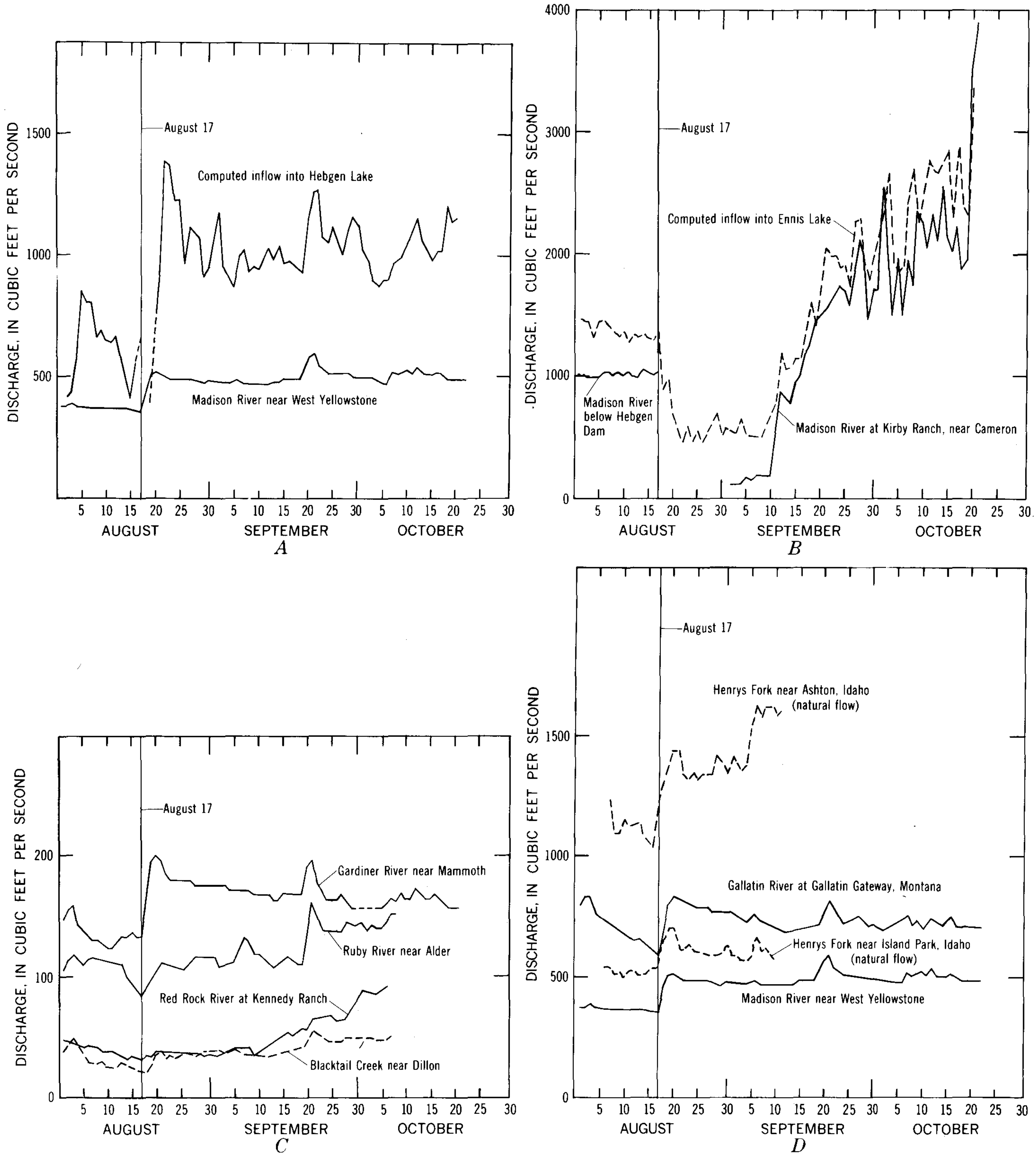


FIGURE 85.—Hydrographs of larger streams in the vicinity of the earthquake area. *A*, Computed inflow into Hebgen Lake (3-day average except August 16–20) and flow of Madison River near West Yellowstone. *B*, Flow of the Madison River below Hebgen Dam. Water began flowing over the landslide dam on September 10. *C*, Flow of Gardiner, Ruby, and Red Rock Rivers, and Blacktail Creek. *D*, Flow of Gallatin and Madison Rivers and computed natural flow of Henrys Fork.

near West Yellowstone and the Madison River below Hebgen Dam. The computed net inflow to Hebgen Lake averaged 610 cfs in the period August 1 to 16 and 1,100 cfs during the period August 21 to 31 (fig. 85*A*). The difference of 490 cfs represents an increase of 80 percent, or 0.54 cfs per square mile of drainage area. The inflow averaged 990 cfs during the first half of September and 1,080 cfs during the last half of September; the increase probably was a reflection of precipitation during that period. The apparent changes in the capacity of Hebgen Lake obscured the pattern of inflow for a few days following the severe earth tremors.

The landslide dam 6 miles downstream from Hebgen Lake interrupted and modified the flow of the Madison River and thus made comparisons of prior and subsequent flow more involved and less definite. Although temporary gages were established after the earthquake on the principal tributaries, Cabin Creek and Beaver Creek, no records of discharge before the earthquake are available, and no conclusions can be drawn regarding increased base flow.

The net inflow to the Madison River between Hebgen Dam and Ennis Lake, near McAllister, was 370 cfs during the period August 1 to 17 (fig. 85*B*). The depletion of the river by diversions for irrigation in this reach is not known. During the period August 21 to 31, the net inflow averaged 530 cfs in the reach from the landslide dam to Ennis Lake. As the discharge of Cabin and Beaver Creeks was not included in the net inflow for the latter period, an approximate flow of 120 cfs should be added to 530 cfs for proper comparison. Some changes in diversion may have occurred to partly account for the inflow gain of 76 percent over the period August 1 to 17. The increase in inflow of 280 cfs is equivalent to 0.22 cfs per square mile of intervening drainage area. A like comparison for the period September 1 to 15 shows the average inflow decreased to about 545 cfs. Flow over the landslide dam began on September 10, and close comparisons after that time should take into consideration changes in channel storage that have not been determined. The hydrographs of figure 86 may be helpful in understanding the relations of the discharge records collected at various points in the long reach between Hebgen Lake and Ennis Lake.

Significant increases in discharge after the earthquake were measured in the Gardiner, Ruby, and Red Rock Rivers, in Blacktail Creek (table 9; fig. 85*C*), and in all tributaries of the Jefferson and Yellowstone Rivers. The increase in discharge of the Gallatin River between August 17 and 21 was about 40 percent, or 0.28 cfs per square mile of drainage area. Buck, Taylor, and Specimen Creeks were turbid after the quake, which indicates that their flow had also increased.

The earthquake caused no appreciable change in the flow of the Yellowstone River at the outlet of Yellowstone Lake; the slow decrease in streamflow continued undisturbed. The Lamar River tributary showed no change attributable to the earthquake, but the Gardiner River near Mammoth became more turbid, and by August 18 its flow increased. The increase from August 17 to 20 was 67 cfs, or about 50 percent, equivalent to 0.33 cfs per square mile of drainage area. Records for the Yellowstone River at Corwin Springs, Mont., indicate that the Gardiner River was the only major tributary that increased in flow.

The complex pattern of reservoir storage and irrigation use makes it difficult to analyze the records of streamflow in the Snake River basin for earthquake effects; it seems that Henrys Fork was the only stream that was clearly affected (fig. 85*D*). Diversions for irrigation and other depletions above Henrys Lake prevent close comparison, although it is reported that the average daily inflow to Henrys Lake was 63 cfs during the period September 11 to October 8. The inflow averaged only 27 cfs during equivalent periods in earlier years, which indicates that a substantial increase occurred as a result of the quake. Miscellaneous discharge data in sec. 18, T. 15 N., R. 43 E., on Duck Creek, a tributary to Henrys Lake are as follows:

Date	*Discharge (cfs)*
Aug. 10	4
Aug. 14	4
Aug. 18	6
Aug. 21	15
Aug. 30	14
Sept. 8	15
Sept. 11	15
Oct. 8	16

The measurements indicate a substantial increase in flow that may be attributed to the earthquake, although the increase may be due to variation in irrigation diversion and to precipitation.

Calculations of daily natural flow can be somewhat misleading if variations in the velocity of the flowing streams and channel and streambank storage are not considered. H. C. Eagle, under whose direction the data in the upper Snake River basin were collected, states that after allowance for these factors, his calculations indicate an increase of about 150 cfs in natural inflow to Island Park Reservoir for a few days following the earthquake. (See fig. 85*D*.) This is an increase of 28 percent, or 0.4 per cfs square mile of drainage area. He states that the increase was prob-

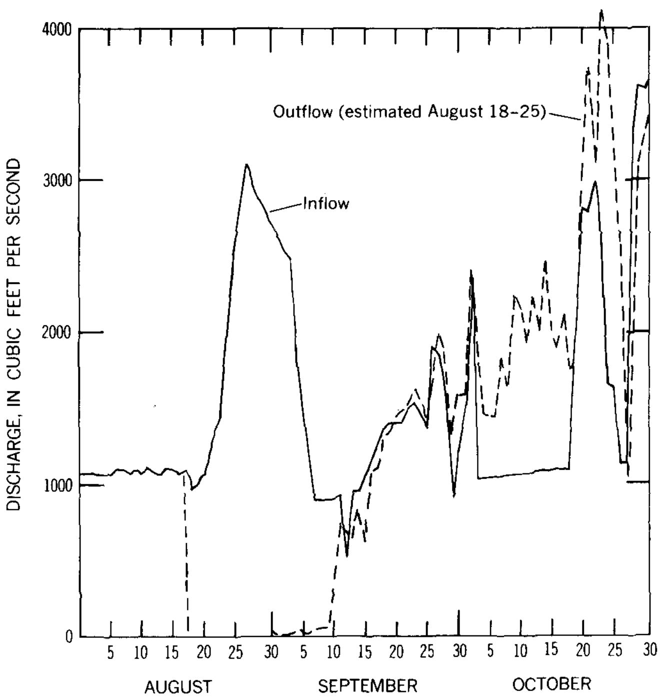


FIGURE 86.—Hydrographs of inflow and outflow of Earthquake Lake.

ably about 50 cfs by September 10. The calculations of natural flow for the gaging station on Henrys Fork near Ashton, Idaho, show an apparent increase of 300 cfs, or about 28 percent, a gain of about 0.3 cfs per square mile of drainage area (table 11, and fig. 85*D*).

MANAGEMENT OF EARTHQUAKE LAKE

The landslide dam across the Madison River posed some serious problems. If the accumulating water had been allowed to rise to the top of the dam, it would have endangered the downstream toe of the earthquake-damaged Hebgen Dam, thus possibly causing failure of this dam. As the stability and permeability of the slide dam under various levels of impoundment could only be estimated, the safety of thousands of people and extensive valuable property were threatened. The U.S. Army Corps of Engineers studied the problem with the aid of consultants, and a plan of action was initiated under their emergency flood-control powers. The saddle, or lowest point, of the slide was lowered to an altitude of approximately 6,450 feet by means of hastily assembled earth-moving equipment that worked at full speed 24 hours a day. The resulting lake, named "Earthquake Lake," then contained about 80,000 acre-feet of water, and its upstream limit was about half a mile downstream from Hebgen Dam. The first overflow occurred on September 10. The prepared spillway did not appear sufficiently stable to provide reasonable safety to downstream residents and property; hence, degradation of the spillway was begun by utilizing the transporting properties of the overtopping water. This involved narrowing the spillway to produce rapid erosion. Earth-moving equipment was used to dislodge the larger boulders in the channel and much of the material on the spillway banks. When, on October 27, the slide dam had been lowered further to an altitude of about 6,400 feet, the resultant lake contained about 45,000 acre-feet; and the passage of the water of the Madison River through the channel with reasonable safety then appeared to have been accomplished.

Figure 86 is a hydrograph obtained partly from estimated flows that show the water management of Earthquake Lake. Inflow to the lake was determined from the records of the Madison River below Hebgen Lake, near Grayling, and also from estimates of tributary inflow. The tributary inflow was estimated to be 80 cfs during the period August 1 to 17 and 130 to 140 cfs after August 17, based on miscellaneous measurements of the discharge of Cabin and Beaver Creeks. The outflow of Earthquake Lake was computed from incomplete records of the flow of the Madison River immediately below the slide, of the Madison River below Sheep Creek near West Yellowstone, and of the Madison River at Kirby Ranch near Cameron. Essentially, the outflow is 130 cfs less than that shown in table 11 for the Madison River at Kirby Ranch.

Sediment Studies on the Madison River After the Hebgen Lake Earthquake

By THOMAS F. HANLY

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

SEDIMENT STUDIES ON THE MADISON RIVER AFTER THE HEBGEN LAKE EARTHQUAKE

By Thomas F. Hanly

ABSTRACT

After the earthquake on August 17, 1959, all the streams tributary to Hebgen Lake appeared to be carrying heavy loads of sediment. Analyses showed that the material in suspension consisted primarily of fine clay with a particle size bordering on the colloidal, and the sediment loads were relatively unimportant. With respect to sediment, the principal result of the earthquake was the increase in sediment transported from the Madison Slide by the drainage from Earthquake Lake. Water with a high velocity due to the steep slope eroded the spillway, and this resulted in large quantities of sediment being transported downstream beyond Cameron. Aggradation in the river and consequent erosion of the banks was very pronounced for about 2 miles downstream from the toe of the slide.

INTRODUCTION

The earthquake not only had a profound effect upon the flow of streams in the Hebgen Lake area, but it also affected the sediment load carried by the streams. Samples of water from five streams entering Hebgen Lake were analyzed to determine the concentration of suspended sediment attributable to the effects of the earthquake. Samples were also collected from the Madison River at gaging stations 1 and 2, at Cliff Lake bridge, at Kirby Ranch, and from the West Fork of the Madison River near the mouth at the Kirby Ranch. (See fig. 87.)

INFLOW OF SEDIMENT TO HEBGEN LAKE

Although the concentrations of suspended sediment were small during the period immediately following the earthquake, they were unusually high for that time of the year. The streams are normally clear in September, and appreciable quantities of sediment are usually carried only during spring runoff. Most of the streams became clear rather quickly after the earthquake, but the South Fork of the Madison River appeared very muddy as late as September 17. Samples obtained on that date, however, indicate that the appearance was misleading; the material in suspension consisted primarily of fine clay with a particle size bordering on the colloidal. As indicated in table 12, the sediment inflow into Hebgen Lake was negligible and had very little, if any, effect upon the capacity of the lake.

Table 12.—*Daily discharge and concentration of sediment in streams flowing into Hebgen Lake*

Source of sample	Date (Sept. 1959)	Discharge (cfs)	Concentration (ppm)
Madison River near West Yellowstone	7	457	4
South Fork Madison River near West Yellowstone	6	120	306
	10	120	257
	17	120	27
Cougar Creek near Grayling	6	30	256
Duck Creek near Grayling	6	37	178
Grayling Creek near Grayling	6	75	134

Only two streams of appreciable size, Cabin and Beaver Creeks, enter the Madison River between Hebgen Dam and the Madison Slide. Both streams were clear on September 7 and remained clear during the period of this study. Sand fountains were active in Beaver Creek valley but probably had little or no effect on the sediment concentrations of either the Madison River or Earthquake Lake.

SEDIMENT IN MADISON RIVER BELOW THE SLIDE

Sampling of suspended sediment from the Madison River below the slide began on September 3 at gage 1, which was 300 feet downstream from the toe of the slide. The flow of the river there at that time was 17 cfs (cubic feet per second). Between September 3 and the time the water began flowing over the slide dam on September 10, concentrations of suspended sediment averaged 45 ppm (parts per million) with an average daily suspended-sediment load of 3.5 tons. The concentrations increased abruptly to 2,800 ppm as the first flow to top the spillway reached the gage. As the flow increased, the concentrations decreased somewhat; but the daily load of suspended sediment averaged about 1,500 tons until September 13, when it became necessary to abandon the station. The daily flow and daily suspended-sediment concentration and load are shown in table 13.

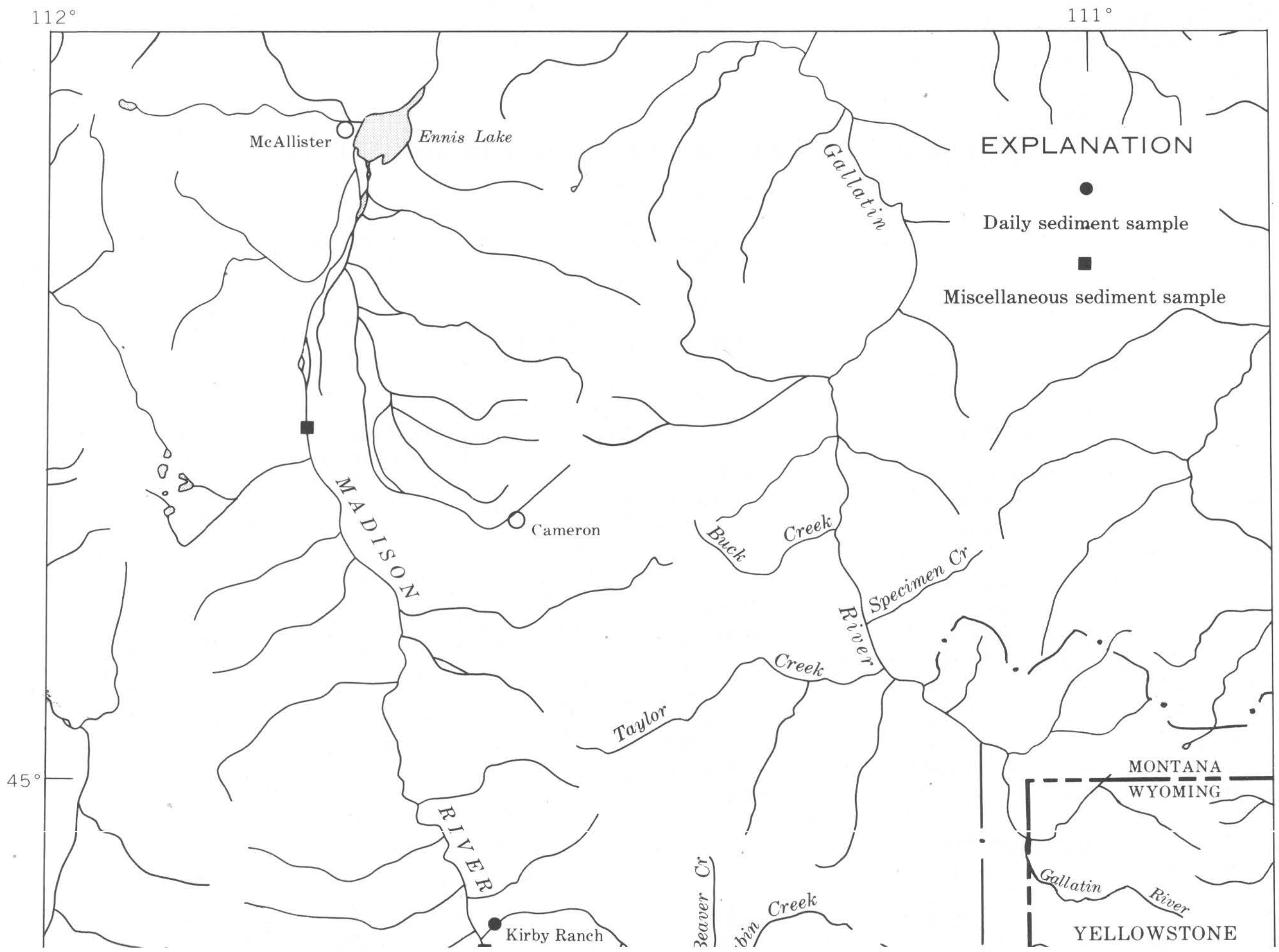
112°
111°
45°
EXPLANATION
Daily sediment sample
Miscellaneous sediment sample
McAllister
Ennis Lake
Gallatin
MADISON
Cameron
Buck Creek
Specimen Cr
River
Taylor Creek
RIVER
Kirby Ranch
Beaver Cr
Creek
MONTANA
WYOMING
Gallatin River
YELLOWSTONE

TABLE 13.—*Daily discharge of Madison River at gage 1 and daily suspended-sediment concentration and load*

Date (Sept. 1959)	Discharge (cfs)	Concentration (ppm)	Load (tons per day)
4	18	45	2
5	19	35	2
6	29	46	4
7	56	45	7
8	67	46	8
9	74	46	9
10	265	1, 130	809
11	655	960	1, 700
12	700	810	1, 530
13	670	750	1, 360

Samples were collected at gage 1 three times on September 10 for analysis of particle size of the suspended sediment. The median particle size was 0.012 mm at 11:35 a.m., 0.017 mm at 2:30 p.m., and 0.021 mm at 11:15 p.m. (fig. 88*A*).

TABLE 14.—*Daily discharge of Madison River at gage 2 and daily suspended-sediment concentration and load*

Date (Sept. 1959)	Discharge (cfs)	Concentration (ppm)	Load (tons per day)
10	271	1, 100	805
11	690	1, 020	1, 900
12	792	740	1, 580
13	738	700	1, 390
14	820	380	841
15	834	440	991
16	1, 110	470	1, 410

When it became evident that gage 1 would have to be abandoned, gage 2 was established on September 10 at a point about 2,640 feet downstream from gage 1. A 4-day overlap of record showed similar patterns of sediment concentration at the two stations. The daily flows of the stream and the concentrations of suspended sediment, as well as the daily sediment loads, are shown in table 14.

Samples of suspended sediment were collected at gage 2 on September 14 and 16 for analysis of particle size. The median size on September 14 was 0.029 mm, about 40 percent larger than the median size at gage 1 on September 10, the last day of its operation. The median size on September 16 at gage 2 was 0.042 mm. (See fig. 88*B*.)

Sampling of the Madison River at the Cliff Lake bridge, about 4 miles downstream from gage 1, was begun on September 5. In the interval between September 5 and the time when the water flowing over the slide reached the bridge, the concentration of suspended sediment averaged 9 ppm, and the daily load averaged 2.6 tons. When the flow over the slide reached the bridge, the concentration of suspended sediment had increased to 400 ppm and the average daily load to 805 tons. The daily flows, concentrations, and loads are shown in table 15.

TABLE 15.—*Daily discharge of Madison River at Cliff Lake bridge and daily suspended-sediment concentration and load*

Date (Sept. 1959)	Discharge (cfs)	Concentration (ppm)	Load (tons per day)
5	87	6	1
6	92	[1] 8	2
7	112	[1] 10	3
8	129	[1] 10	3
9	138	12	4
10	346	470	439
11	835	710	1, 600
12	940	350	888
13	772	500	1, 040
14	920	270	671
15	718	100	194

[1] Estimated.

Daily sampling was discontinued at this site on September 16, but samples collected on October 8 and 13 contained 402 and 640 ppm of sediment, respectively, indicating daily suspended-sediment loads of 1,780 and 3,630 tons. Three samples were collected at the Cliff Lake bridge for particle-size determination; the median sizes were 0.012 mm on September 10, 0.030 mm on October 8, and 0.045 mm on October 13. (See fig. 88*C*.)

On September 9, a daily sampling station was installed on the Madison River at the Kirby Ranch bridge, 10.9 miles downstream from gage 1 and 0.5 mile upstream from the West Fork of the Madison River. This station was established to learn the effects of high runoff over the slide and in the river downstream from the slide. Daily flows, concentrations, and loads are shown in table 16.

TABLE 16.—*Daily discharge of Madison River at Kirby Ranch bridge and daily suspended-sediment concentration and load*

Date (Sept. 1959)	Discharge (cfs)	Concentration (ppm)	Load (tons per day)	Date (Oct. 1959)	Discharge (cfs)	Concentration (ppm)	Load (tons per day)
9	184	5	2	1	1, 720	200	929
10	416	179	201	2	2, 540	1, 470	10, 100
11	868	348	816	3	1, 940	1, 520	7, 960
12	804	[1] 300	651	4	1, 480	[1] 750	3, 000
13	780	[1] 280	590	5	1, 460	[1] 500	2, 000
14	954	260	670	6	1, 480	[1] 400	1, 600
15	708	120	229	7	1, 960	[1] 340	1, 800
16	1, 180	190	605	8	1, 740	330	1, 550
17	1, 240	110	368	9	2, 380	[1] 340	2, 200
18	1, 440	240	933	10	2, 280	[1] 360	2, 200
19	1, 480	190	759	11	2, 060	[1] 400	2, 200
20	1, 560	210	885	12	2, 360	520	3, 310
21	1, 620	270	1, 180	13	2, 140	620	3, 580
22	1, 680	170	771	14	2, 580	1, 000	6, 970
23	1, 750	130	614	15	2, 160	680	3, 970
24	1, 700	100	459	16	2, 020	550	3, 000
25	1, 580	120	512	17	2, 240	740	4, 480
26	1, 790	408	1, 970	18	1, 880	470	2, 390
27	2, 120	325	1, 860	19	1, 960	680	3, 600
28	1, 970	90	479	20	3, 440	2, 370	22, 000
29	1, 480	70	280	21	3, 840	2, 460	25, 500
30	1, 720	90	418				

[1] Interpolated.

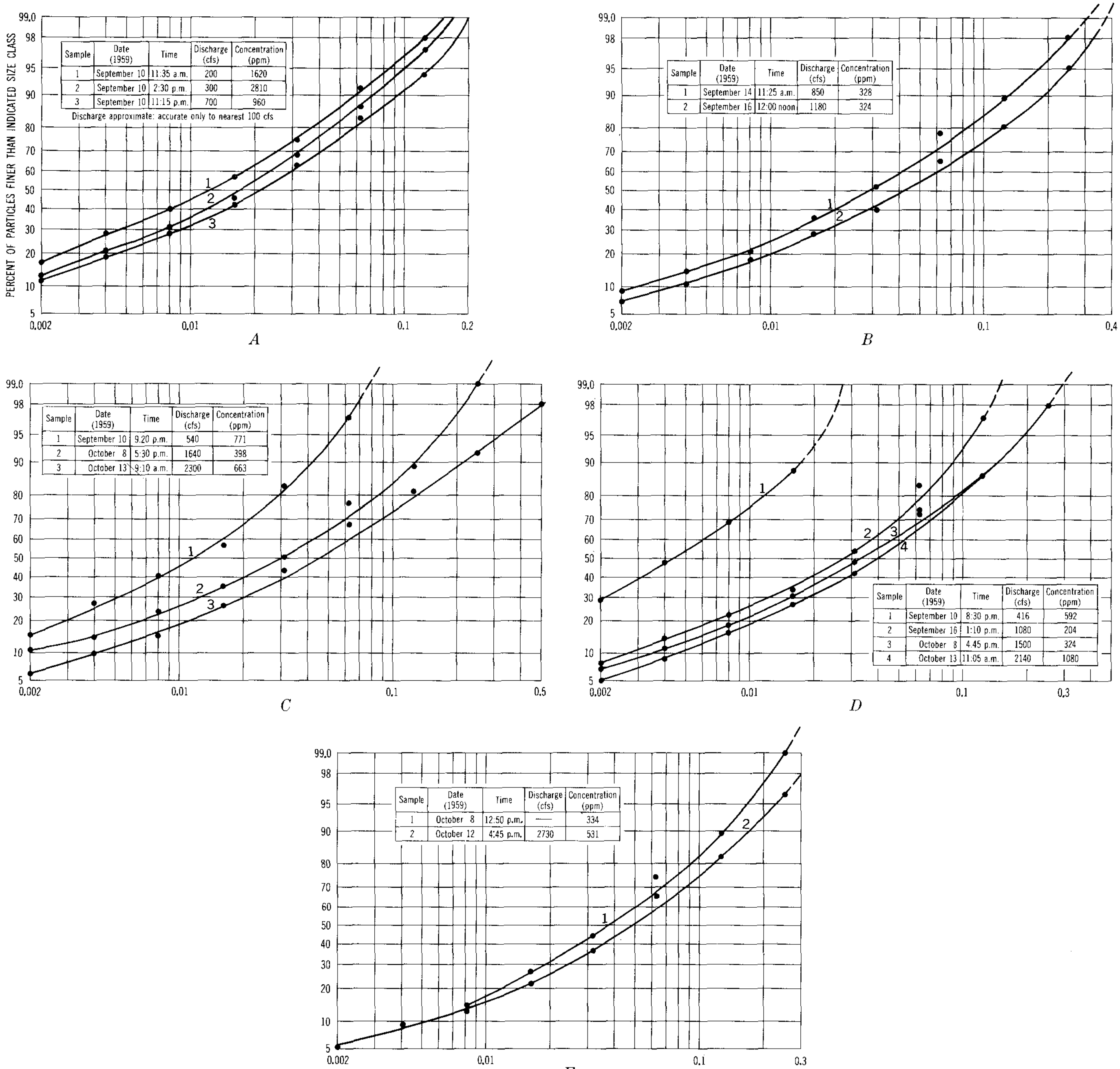

FIGURE 88.—Particle size of suspended sediment in Madison River. *A*, At gage 1; *B*, at gage 2; *C*, at Cliff Lake bridge; *D*, at Kirby Ranch bridge; *E*, at gaging station near Cameron.

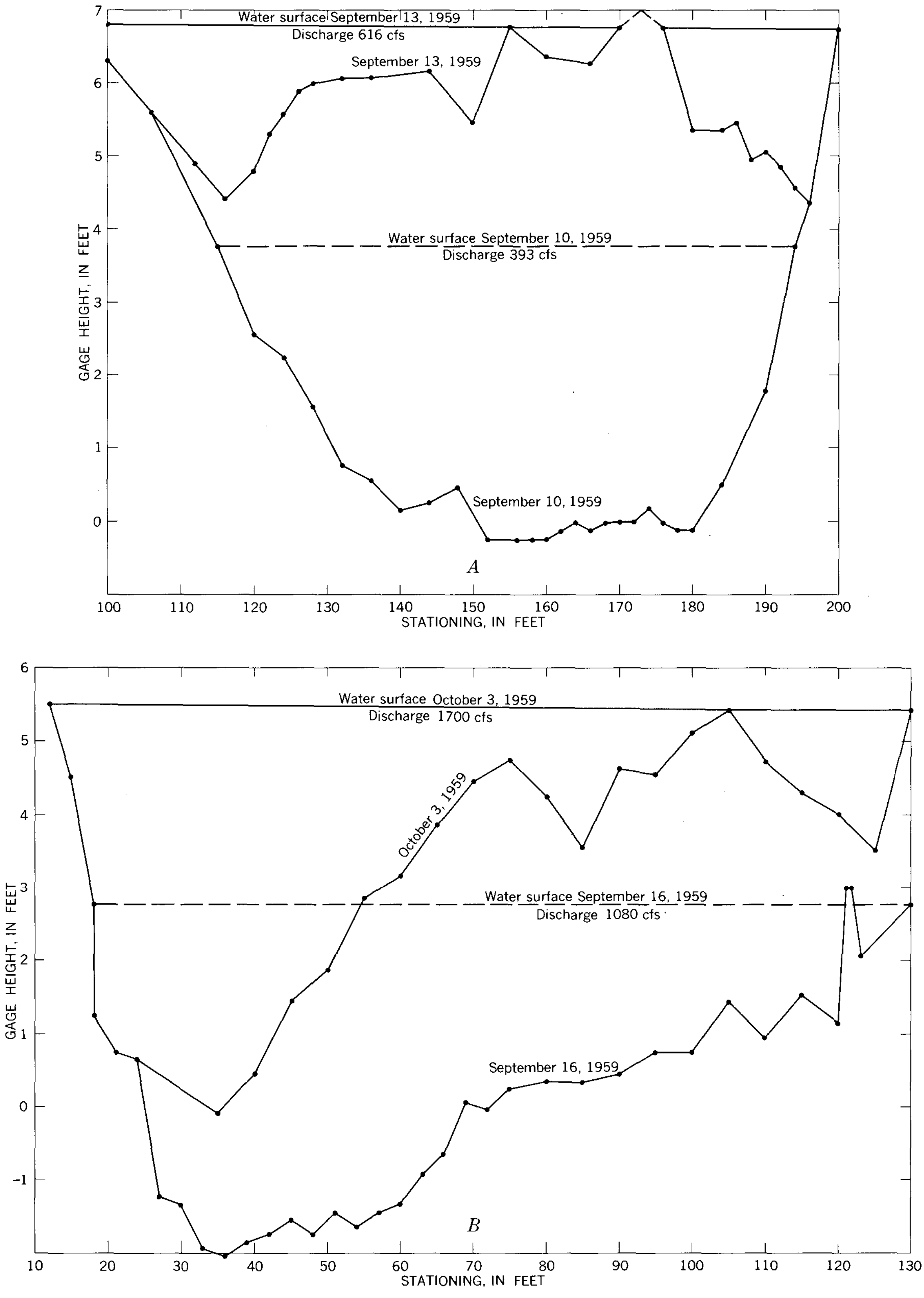

FIGURE 89.—Sections of the channel of the Madison River below the landslide. *A*, at gage 1; *B*, at gage 2.

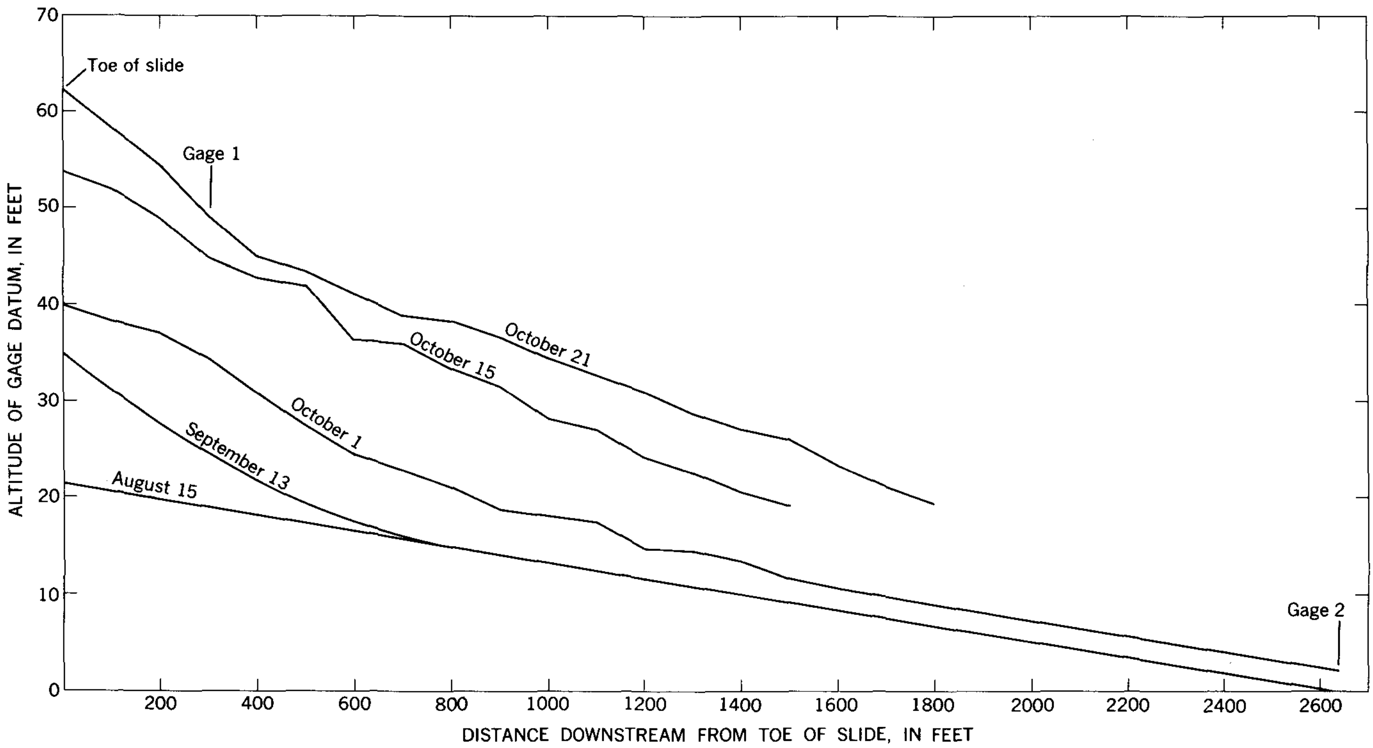

Figure 90.—Profiles showing aggradation of the channel of the Madison River between the toe of the slide and gage 2.

The suspended-sediment concentration in the river at this station was 5 ppm on September 9, and the daily load was 2 tons. During September 10 to 30, the concentration of sediment fluctuated considerably as a result of the work being done at the spillway across the slide. In this period the concentration of sediment averaged 200 ppm, and the daily loads averaged 727 tons. The concentration and load increased appreciably on October 1 when the work of deepening the spillway began. Four samples were collected during the period September 10 to October 13 for analysis of particle size. The median particle sizes were 0.004 mm on September 10, 0.027 mm on September 16, 0.033 mm on October 8, and 0.040 mm on October 13. The median particle size increased tenfold during this period. (See fig. 88*D*.)

Several samples of water were collected from the Madison River at the gaging station near Cameron, where the suspended-sediment concentration increased from 7 ppm on September 9 to 531 ppm on October 12. Sediment load increased from 6 tons to 2,730 tons per day during this period. Two samples were collected at this site for particle-size determinations. The median particle sizes were 0.038 mm on October 8, and 0.05 mm on October 12. (See fig. 88*E*.)

In addition to the periodic and daily sampling mentioned above, samples were collected on September 16 and October 26 from the Madison River at the gaging station downstream from Ennis Lake, and on September 5 from the West Fork of the Madison River near its mouth at Kirby Ranch. The concentrations of suspended sediment at the gaging station below Ennis Lake were 20 and 118 ppm on September 16 and October 26, respectively. The concentration of the sample from the West Fork of the Madison River was only 6 ppm on September 5, and as late as September 16 the stream was still clear.

MOVEMENT OF BED MATERIAL BELOW THE SLIDE

The movement of bed material from the slide into the channel became evident soon after the water started flowing through the new spillway (figs. 89

and 90). It was not possible to measure the movement of bed material as the material was much too large for the available sampling equipment.

Aggradation in the channel at gage 1 averaged about 5½ feet in depth by September 13, and, as the stage-discharge relation was changing so rapidly, it became necessary to abandon the station. By October 21 the channel at gage 1 had aggraded to an average depth of about 30 feet.

The movement of bed material continued (figs. 91, 92, 93), and by October 3 it was necessary to abandon gage 2. Sections of the channel at gage 1 on September 10 and 13 and at gage 2 on September 16 and October 3 are shown in figure 89. Profiles of the channel between the toe of the slide and gage 2 are shown in figure 90.

FIGURE 91.—Madison River, view upstream from the site of gage 1 to the landslide. *A*, September 10, 1959; *B*, September 14, 1959.

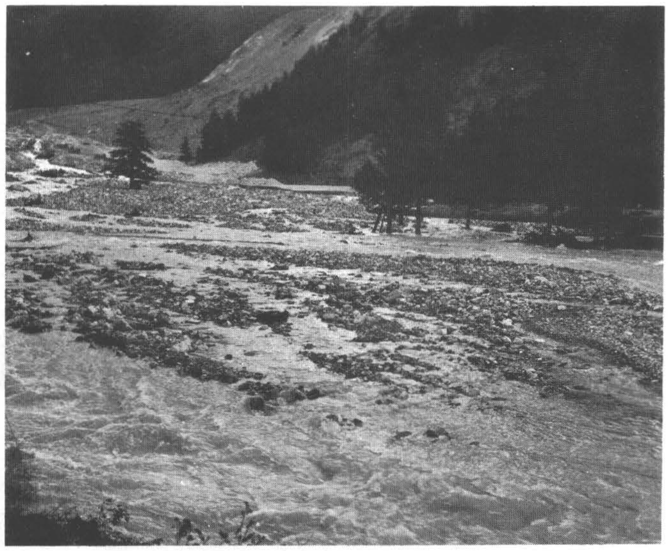

FIGURE 92.—The Madison River at the site of gage 1, September 14, 1959.

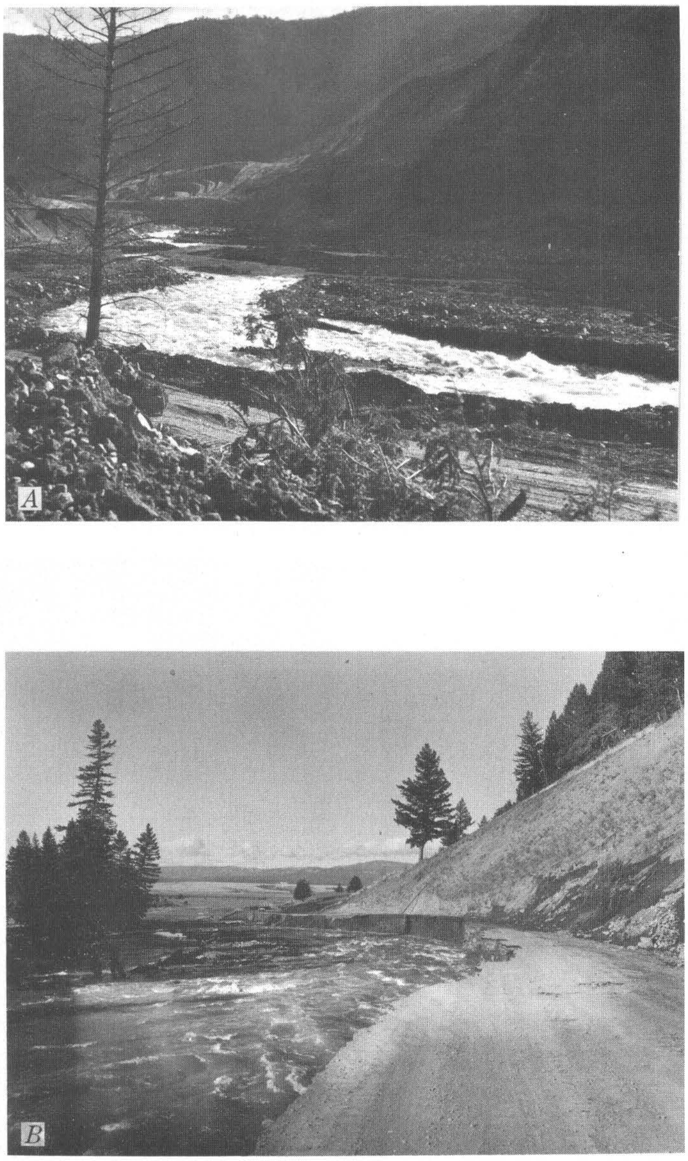

FIGURE 93.—Madison River below the landslide, October 28, 1959. The outlet channel across the slide dam at this time had been lowered about 50 feet from its original position, thus draining part of Earthquake Lake. *A*, View upstream from site of gage 1; *B*, view downstream from gage 2 showing erosion of bank and highway.

Ground-Water Phenomena Associated With the Hebgen Lake Earthquake

By FRANK A. SWENSON

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

GROUND-WATER PHENOMENA ASSOCIATED WITH THE HEBGEN LAKE EARTHQUAKE

By Frank A. Swenson

ABSTRACT

The Hebgen Lake earthquake had great effects on the ground- and surface-water regimen of the water-rich West Yellowstone area. Most streams are spring fed, and after the earthquake the water discharged was very turbid. Most large springs are associated with volcanic rocks, but others issue from older sedimentary beds or from the alluvium. Some springs increased greatly in flow; others ceased to flow. The temperature of the water from some springs reportedly was higher after the earthquake, and new thermal springs issued from an active fault zone.

Wells were also influenced by the earthquake. Whereas water issued under pressure from at least two wells in the area which had never been known to flow, water levels in several other wells declined noticeably. Effects of the earthquake were noted in wells far from the epicenter. Flow began or flow pressures increased in wells within a radius of 200 miles.

Numerous sand spouts were formed where ground water issued under pressure. In some places these spouts were alined along cracks; in other places discrete craters were formed.

Two large sinkholes were formed on a steep alluvial fan underlain by cavernous bedrock. Some water was erupted in the early stages of formation, but most of the unconsolidated alluvial material was carried downward into the underlying bedrock.

GENERAL GROUND-WATER CONDITIONS

The rich ground-water resources of the West Yellowstone–Hebgen Lake area are utilized only slightly by man, but they are replenished abundantly by nature. About 21 inches of precipitation falls at West Yellowstone yearly, and considerably more falls in the surrounding highlands. Precipitation falls mainly during the winter, accumulates as snowpack, and discharges as the snow thaws in spring and summer. Records for the 15-year period 1938–52 show that an average of 11.8 inches of precipitation accumulates as snowpack by April 1 of each year at a snow course near West Yellowstone.

Springs, some fairly large, are numerous in the West Yellowstone area and are the most important source of ground-water discharge. Most of the springs are associated with volcanic rocks that border the area on the south, east, and northeast, but many issue from older sedimentary rocks or rise from the alluvial plain. The location of the springs and wells discussed in this report is shown on plate 2.

The city of West Yellowstone does not have a municipal water system; homes and places of business are supplied from individual private wells. Many of the older wells were dug by hand and later deepened by driving sand points or drilling. Most of the newer wells, and those wells supplying water for public consumption, were drilled and cased from the surface. Outlying ranches, developed campgrounds, and summer cabins use water from springs or from drilled or driven wells. The water level beneath the plain bordering Hebgen Lake on the south and east is, in general, shallow and has considerable seasonal fluctuation. Changes in the level of the lake are reportedly reflected in the water level in wells adjacent to the lake.

EFFECTS OF THE EARTHQUAKE ON SPRINGS

The majority of streams in the area are fed by springs that normally discharge crystal-clear water. After the earthquake, water from many of the springs became brown and turbid. The rate of discharge from many springs also was affected; some springs reportedly discharged as much as three times their normal flow, whereas the flow from others decreased, or ceased entirely. In addition, many small ephemeral springs were created.

Three principal springs and numerous smaller ones are the main source of water for the South Fork of the Madison River. The flow of the river in early September was less than 0.3 cfs (cubic feet per second) upstream from a series of springs in the northern part of sec. 19, T. 14 S., R. 5 E. The water issues under pressure into the stream from the base of a 40-foot vertical cliff composed of columnar-jointed rhyolite. The joints in the massive volcanic rock provide channelways for the water. Just downstream from the largest group of visible springs, the flow of the stream was gaged at 8.60 cfs on September 3. The water had a temperature of 40°F and was extremely turbid.

Small springs are numerous downstream for the next 3 miles; their combined flow was estimated to be about 10 cfs. Another large spring issues at the base of a steep slope between the road and the railroad in the NE¼NW¼ sec. 5, T. 14 S., R. 5 E. Its discharge could not be gaged because of the backwater from beaver dams, but it was estimated to be about 8 cfs. The water issues from jointed ryholite, and, when sampled on August 28, it had a temperature of 47°F and was extremely turbid.

The largest observed spring tributary to the South Fork of the Madison River, known as Black Sand Spring, is in the SW¼NE¼ sec. 31, T. 13 S., R. 5 E.; its flow was 16.34 cfs when gaged on August 31. The flow at that time was reportedly somewhat less than normal. The water issues from around the edges and at the bottom of a pool about 100 feet in diameter at the base of a gentle slope. Some of the water issues from a deposit of stream-rounded pebbles and cobbles stratigraphically underlying slabby volcanic rock, which in part borders the pool. The water had a temperature of 48°F when sampled and was extremely turbid when first observed after the earthquake. By early September the water issuing from the western part of the spring was almost clear, but the water in the remainder of the spring was still turbid. The water was still somewhat milky on October 23, and the flow seemed to be greater at this time than it was immediately after the earthquake.

It was reported that the flow of Big Springs, about 13 miles southwest of West Yellowstone in Idaho, was highly turbid immediately after the earthquake. These springs, with a combined flow of more than 180 cfs, still had a milky green color on September 4. The springs issue from the base of a rhyolite cliff, and the geologic setting is similar to that of the springs that issue into the South Fork of the Madison River.

Another large spring that is associated with volcanic terrane is along lower Campanula Creek in Yellowstone Park at long 111°04′13″ W. and lat 44°47′30″ N. The spring, here designated Duchess Spring, had a flow of more than 3.5 cfs when visited on August 27, had a water temperature of 47°F, and was extremely turbid. It issues from the base of a cutbank near the north edge of a distinctive structural valley that extends westward from the spring 2½ miles to Grayling Creek. No stream occupies this valley, which is marked by recent faulting about half a mile west of the spring. The position of Duchess Spring is governed by the bedrock structure.

The largest spring observed issues from the base of a steep slope about 30 feet above the level of the Grayling Arm of Hebgen Lake (SW¼NW¼ sec. 7, T. 12 S., R. 5 E.). This spring was formerly known as Corey Spring, but it is on land now owned by Mr. E. S. Armstrong. According to the owner, the flow of the spring was gaged at 15 cfs in 1935; its temperature was 42°F in July 1959; and the water was cloudy for about 48 hours after the earthquake. On September 2, the spring had a gaged flow of 16.73 cfs and a temperature of 46°F. Paleozoic rocks form the high mountain mass immediately north of the spring, and it is believed that they are the source of the water, although the spring issues from colluvial material.

The other springs that were visibly affected by the earthquake issue from Precambrian rocks west of Hebgen Lake. One is in the NE¼ sec. 4, T. 13 S., R. 4 E., and furnishes water for the Lonesomehurst summer-home area. Its flow ceased at the time of the earthquake, and it remained dry for about 2 weeks. The flow was said to be somewhat greater several weeks after the earthquake than it had been before the earthquake. The other spring, in the SW¼ sec. 20, T. 12 S., R. 4 E., furnishes water for the Rumbaugh Ridge summer-home area. Its flow was considerably greater than normal when observed in early September, and there were flakes of weathered mica in its water at that time.

A series of new springs emerged from an active fault zone that cuts Precambrian rocks in the southwestern part of Horse Butte. The fault zone extends along the hillside between the shore road and the road to the lookout tower on Horse Butte. Numerous springs issued from the fault zone and made the shore road impassable in the northern part of sec. 35, T. 12 S., R. 4 E. When these springs were visited on September 4, it was estimated that the total flow entering the reservoir along the quarter mile of spring flow totaled 170 gpm (gallons per minute). The flow of the springs had decreased to only about 25 gpm by October 22. The temperature of the water ranged from 50° to 53°F, or high enough above the mean annual air temperature for the water to be considered thermal.

A large spring issues from alluvium on the T. S. Povah ranch in the SE¼SW¼ sec. 14, T. 13 S., R. 4 E.; it was flowing 8.01 cfs when gaged on August 31 and had a temperature of 47°F. The water became extremely turbid after the earthquake, but in about

2 weeks it began to clear, and by mid-September it had only a slight milky appearance. The water issues from numerous openings in a shallow depression in the alluvial plain.

Several springs rise in alluvium near the Basin Ranger station near the SE corner sec. 9, T. 13 S., R. 4 E. They were reported to flow about 6 cfs but ceased flowing after the earthquake. They were still not flowing when observed in late August, although water was standing in several depressions. The flow was estimated to be between 2 and 3 cfs on September 24 and between 4 and 5 cfs on October 23.

A brief change in Radium Hot Springs in the Kootenay National Park of British Columbia illustrates the random effects on springs outside the area described in this report. Radium Hot Springs are west of the Continental Divide and about 500 miles northwest of Hebgen Lake. About an hour after the first shocks, the water in Radium Hot Springs turned dark-chocolate brown, gradually clearing in 36 to 40 hours. The volume of flow of the springs was approximated by timing the rate of filling of an adjacent basin, and an increase of flow of about 50 gpm was determined in this manner. A temperature rise of about 2°F also was observed. Normal conditions were restored in about 48 hours.

EFFECTS OF THE EARTHQUAKE ON WELLS

Widespread effects of the earthquake on wells were noted in the area. (See da Costa, chapter O.) The effect on wells close to the epicenter of the earthquake was in places spectacular. At The Narrows Motel in the NE¼SE¼ sec. 10, T. 12 S., R. 4 E., which is very near Hebgen Lake, the water level in a well 58 feet deep cased with 3⅝-inch steel pipes reportedly fluctuated with the level of the reservoir. The cylinder pump was securely bolted to the flanged top of the casing. Immediately after the earthquake, water under high pressure containing fine-grained mica sand was forced out of the narrow opening between the base of the pump and the flanged top of the casing. Mr. Frank Jans, the owner, reported that the water and sand were forced out horizontally a distance of 30 feet from the well. As pressure ebbed, this distance slowly decreased, and after 3 hours the water was dripping off the top of the casing and continued to do so for some time. The casing, with attached pump, was lifted 6 to 8 inches above the concrete slab on which it had rested.

Similar conditions may have prevailed at a stock well in the NE¼SE¼ sec. 25, T. 12 S., R. 4 E., on Horse Butte peninsula. Water and coarse sand presumably were forced up an 81-foot drilled well—probably outside the casing. A cone of coarse sand and finer grained material about 6 inches high and extending 10 feet from the well was built up around the casing. It seems that after emitting the granular material, the well ejected clearer water which somewhat eroded the deposited cone.

On the Deep Well Ranch, owned by T. S. Povah, an oil-test well drilled some 20 years before the 1959 earthquake tapped artesian water, and the well now flows 2.02 cfs. The well is in the NE¼NW¼ sec. 23, T. 13 S., R. 4 E., and is reported to be 280 feet deep. Normally the water from the well is clear and colorless, but for several weeks after the earthquake it was milky. The water, which is at a temperature of 44°F, is forced under considerable pressure from an 8-inch casing, and now forms a fountain above the top of the casing, which is about 5 feet above the ground.

There was an appreciable lowering of the water level in several wells in the area. The water level in a dug well in the NE¼NW¼ sec. 1, T. 13 S., R. 4 E., which was formerly used as a water supply by the Madison Arm resort, was about 9.5 feet lower after the earthquake. As the surface of Hebgen Lake at this location was about 8 feet lower after the earthquake, it is not surprising that the water level in this well declined. The lowering of the water in the well and lake resulted partly from the earthquake and partly from a later intentional lowering of the lake to determine the extent of damage to the dam.

A well in the basement of a house at the R. B. Whitman ranch in the SW¼NE¼ sec. 16, T. 12 S., R. 5 E., had a static water level about 11 feet below the basement floor before the earthquake. During the earthquake, and perhaps for some period afterward, water flowed from the well into the basement. By the morning after the earthquake, however, a jet pump, which was set at 15 feet, failed to pump water. The pump intake was reset 10 feet lower, and by November 1959 the well was furnishing an adequate amount of water. The well is about 150 feet from a prominent fault scarp, on the relatively upthrown block.

Most drilled wells in the town of West Yellowstone are 100 to 125 feet deep and obtain water from relatively permeable obsidian sand and gravel. Water from most of the drilled wells was slightly turbid for a few hours after the earthquake. Dug wells generally extend a few feet below the water table, which, in West Yellowstone, is about 40 feet below ground surface. They generally are cribbed with pine boards and timbers and may be open to the bottom or backfilled with earth around casings of tile or culvert material. Both types of dug wells failed after the earthquake, owing either to collapse of cribbing or to local, and possibly temporary, lowering of the water table.

The effect of the earthquake on wells was not limited to the West Yellowstone area. A well on the Roy Keller ranch about 150 miles from West Yellowstone near Billings, Mont. (SE¼SE¼ sec. 16, T. 2 S., R. 25 E.), began to flow the day after the earthquake and was still flowing in November 1959. The well had a small flow from a depth of 81 feet while it was being drilled in 1949, but the flow later ceased. The water level before the earthquake was about 1 foot below ground surface; it is now 0.2 foot above the surface.

It was reported that the shut-in pressure of water in an 800-foot artesian well at a creamery in Lewistown, Mont., about 190 miles northeast of West Yellowstone, increased by 5 to 10 pounds per square inch after the earthquake. This is equivalent to a rise in pressure head of between 11.5 and 23 feet. Another artesian well, 2 miles east of Lewistown, ceased flowing in about 1949 but again flowed about 1 gpm after the earthquake. Near Garneill, about 160 miles northeast of West Yellowstone, the flow of a well doubled as a result of the earthquake.

In addition to the effect on wells, it is reported that the water level in a mine in Custer County, Idaho, rose to the extent that considerable additional pumping was required.

SAND SPOUTS

The earthquake caused cracks to open in the earth in many places, and ground water and sand were erupted onto the surface. These phenomena are termed "sand spouts" or "earthquake fountains" (Byerly, 1942) and have been observed in connection with such major earthquakes as New Madrid, 1811 (Fuller, 1912), Charleston, 1886 (Dutton, 1889), Assam, 1897 (Oldham, 1899), and San Francisco, 1906 (Lawson, 1908). The ejected ground water was under considerable hydrostatic pressure in some places, as indicated by the swampy land west of the lower end of Beaver Creek (sec. 21, T. 11 S., R. 3 E.), where the confining bed was tough, peaty sod. Fragments of this sod 6 to 8 inches thick and several feet square were torn loose or folded back by the surge of confined water. This mechanism may also explain the masses of peaty material as large as 20 × 6 × 4 feet that were found floating in Hebgen Lake or thrown onto the shore. The peaty material is made up of intergrown roots and stems and is difficult to pull apart.

The sand spouts near the mouth of Beaver Creek were formed in or near a glacial moraine and in alluvial and paludal deposits behind the moraine. The material ejected from some sand spouts on the edge of the moraine consists of fresh angular rock fragments about the size of fine-grained sand. The moraine contains large glacial boulders as well as clay-size material, but the erupted sand is of very uniform size. So much sand was ejected in several places that the confining bed of peaty sod caved in, leaving steep-walled holes several feet in diameter that extend below the water table.

FIGURE 94.—Open cracks with ridges, about 6 inches high, of erupted fine-grained sand. The case is about 4 inches long.

Sand and water were ejected along the entire length of cracks in some areas (fig. 94), whereas discrete craters were formed along the cracks in other areas. There seems to be no pattern of distribution or alinement of the earth cracks in the Beaver Creek area.

Another conspicuous area of numerous sand spouts on the northeast side of Hebgen Lake extends across the northern part of sec. 17, T. 12 S., R. 5 E. The

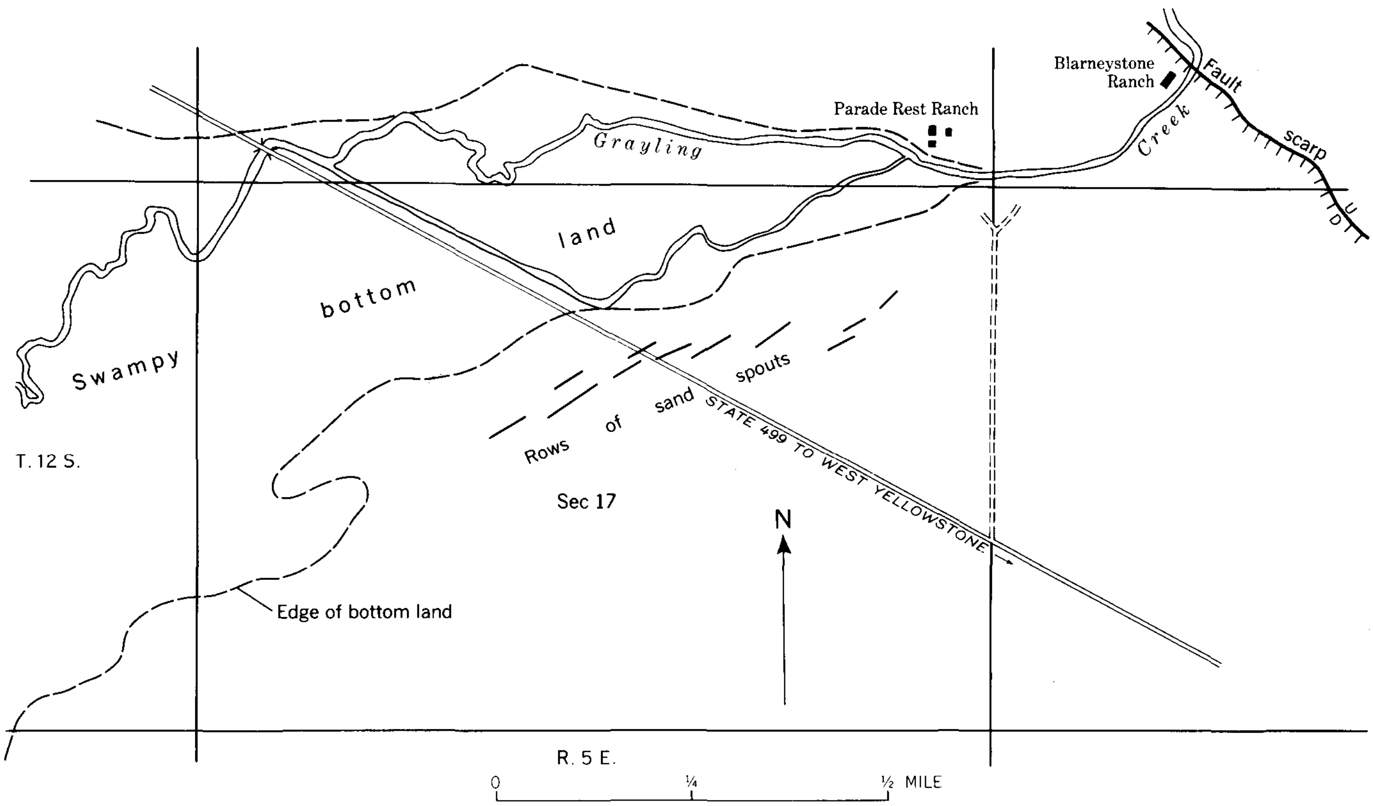

FIGURE 95.—Sketch of alinement of cracks and sand spouts in sec. 17, T. 12 S., R. 5 E.

earth cracks and associated sand spouts are alined fairly well on a general trend of N. 60° E., which is about normal to the fault scarp half a mile to the northeast. The cracks are arranged in a crude en echelon pattern (fig. 95). So much material was erupted in places that the ground between parallel adjacent cracks has settled more than 1 foot. These sand spouts are on the alluvial flats adjacent to the lower swampy course of Grayling Creek. The material erupted is largely fine-grained sand with minor amounts of coarser and finer grained material.

A third area of sand spouts is in the SW¼ sec. 24 and NW¼ sec. 25, T. 12 S., R. 4 E., on Horse Butte peninsula. There is no definite trend to the cracks at this location. One of the larger groups of sand spouts is along a crack that developed across the top and side of a hill; other cracks are in a broad shallow depression and are alined approximately normal to the strike of the depression. The sand spouts in this area are largely in glacial deposits plastered on the east flank of the bedrock of Horse Butte. The material erupted consists of coarse- to fine-grained sand which is very similar to that in the deposits found around the well about half a mile to the southeast.

Two symmetrical sand spouts (fig. 96) were observed on the emergent south shore of the Madison Arm of Hebgen Lake near the middle of sec. 6, T. 13 S., R. 5 E. They are the only ones that were found on a fault trace; all the others observed were on earth cracks having no discernible displacement. The ejected material consists of lake silt and fine-grained sand. The fact that these cones show no signs of wave erosion indicates that they were built after the Hebgen Lake shoreline emerged.

SINKHOLES

On the lower part of the rather steep alluvial fan of Dave Johnson Creek in the NE¼SE¼ sec. 5, T. 12 S., R. 4 E., two large sinkholes appeared after the earthquake. The larger of these holes (fig. 97) is 45 feet long, 15 feet wide, and visibly 12 feet deep. The ends of the hole are undercut 4 or 5 feet. The sinkhole is oriented in a northwest-southeast direction, normal to the ground slope but approximately parallel to the strike of the Paleozoic rocks in the mountains to the east. A small amount of sand and gravel containing cobbles up to 6 inches in diameter was formed along the lowest edge of the sinkhole and indicates that there was some eruption of material, similar to a sand spout, during the early stages of

FIGURE 96.—Row of sand spouts with two principal craters on emergent floor of Hebgen Lake. The row of sand spouts is on trace of fault with downdropped block to the right. The fault displacement of about 2 feet is masked by the movement of saturated silt with the recession of the lake water. The craters were formed after the shore emerged and thus were not destroyed by lake waves. Tobacco can gives scale.

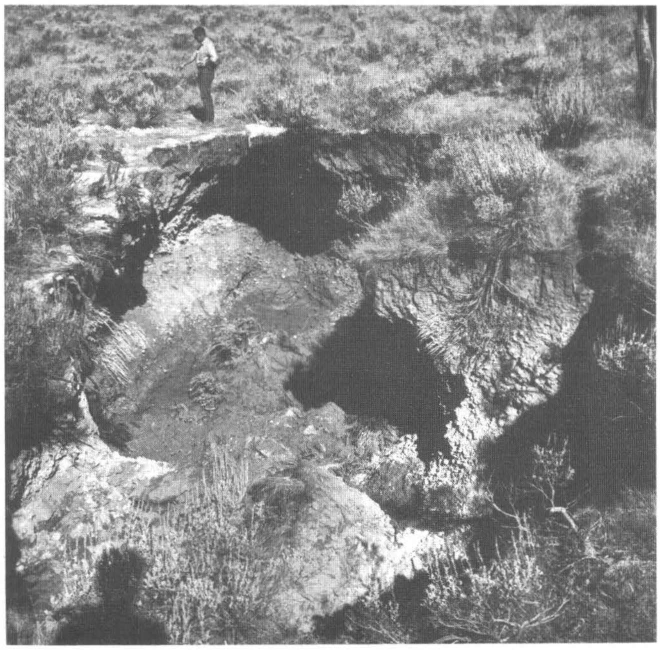

FIGURE 97.—Large sinkhole developed on steep alluvial fan of Dave Johnson Creek near Hebgen Lake Lodge. Hole is 45 feet long, 15 feet wide, and at least 12 feet deep. Photograph by John R. Stacy.

formation of the sinkhole. The amount of material erupted is very small compared to the size of the hole. Exposed in the sides of the hole is coarse cobble alluvium that could not have been washed from the hole without the eruption of considerably more water than the evidence indicates.

The smaller sinkhole, about 200 feet west of the larger one, is about 15 feet long, 5 feet wide, and 6 feet deep. At the northwest end is a round depression 15 feet in diameter where the sagebrush and sod have sunk. A deposit of mud on the sod indicates that a few inches of muddy water filled the depressed basin some time after the quake. After the muddy water drained away or evaporated, the area continued to sink, and when observed in late October 1959 the depression was 2 feet deep.

In the spring of 1960, when the level of Hebgen Lake had been lowered to permit repair of the dam and power installations, some additional observations were made. Much of the upper part of the reservoir bottom was exposed and sublake sand spouts and other features were found.

On the strike of the large sinkhole, described above, and about 1,000 feet southeast of it, a large sand-spout sinkhole was visible (fig. 98). This feature was 10 feet in diameter and 3 feet deep with a broad shallow cone of erupted sediments. The flow of a small intermittent stream had discharged into the crater and eroded the cone of fine sediments to the gravelly lake bottom, but there was no visible evidence of surface flow from the crater. This sandspout sinkhole was formed under the lake and was exposed only when Hebgen Lake was lowered.

Many small to large sublake sand spouts were observed by F. W. Woodward[7] in the southwestern part of sec. 14, T. 12 S., R. 4 E., and in adjacent sections. These features are on the strike of the sinkholes and the sand-spout sinkhole above described. This strike is parallel to that of the northeast shore of Hebgen Lake and to that of the Hebgen fault. Many of the sublake sand-spouts observed by Woodward are small and resemble prairie dog towns, but one com-

[7] Consultant, Billings, Mont., oral communication.

posite cone had a diameter of 20 feet. Without question, such sublake sand spouts contributed to the turbidity of the lake.

A row of sand spouts extends along the lower edge of the smaller sinkhole and makes an angle of at least 10° with the strike of the hole. Much more material was erupted from this row of sand spouts than from the sinkholes. All the material erupted by the sand spouts is fine-grained sand and silt, which is certainly not representative of the coarse cobble alluvium exposed in the walls of the sinkholes.

It is believed that the rocks underlying the alluvial fan at depth are cavernous Paleozoic limestones that were dropped down by faulting from the rocks exposed in the mountains to the east. At the time of the earthquake, compressive forces were exerted on the ground water contained in caverns and open fissures in rocks beneath the fan. The compression was great enough to force water through zones of weakness to the land surface, and the initial surge of this water carried some sand and gravel. With release of pressure, the water returning into the ground carried material down into the caverns and joints below. It is quite possible that there was a series of surges of water, each of which carried more rock material downward until support for the sod roof was weakened to the point of collapse.

FIGURE 98.—Sublake sand-spout sinkhole on floor of Hebgen Lake in SW¼ sec. 4, T. 12 S., R. 4 E. A shallow cone of erupted sediments can be seen around the crater. This feature is about 1,000 feet southeast of the sinkhole shown in figure 97. Crater is 10 feet in diameter and 3 feet deep.

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Effect of Hebgen Lake Earthquake on Water Levels in Wells in the United States

By JOSE A. DA COSTA

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EFFECT OF HEBGEN LAKE EARTHQUAKE ON WATER LEVELS IN WELLS IN THE UNITED STATES

By Jose A. da Costa

ABSTRACT

The Hebgen Lake earthquake caused fluctuations of water levels in wells throughout the conterminous United States and as far away as Hawaii and Puerto Rico. The main shock was recorded by fluctuations that ranged from several feet near the earthquake epicenter in southwestern Montana to a few hundredths of a foot in more distant localities. Some of the weaker aftershocks caused smaller, but recordable, fluctuations.

INTRODUCTION

Records of fluctuations resulting from the Hebgen Lake earthquake were obtained from wells under observation by the Geological Survey in 30 States. This is the widest coverage received by any earthquake occurring in the United States since the beginning of extensive continuous recording in 1943. This does not mean that the Hebgen Lake earthquake was the largest in that period, but that more observation wells are now available than previously, and that the water-level records have been studied more closely for the effects of earthquakes. Nevertheless, the effect of the Hebgen Lake earthquake on water levels in wells is the most pronounced ever recorded in the United States.

SUMMARY OF WATER-LEVEL AND PRESSURE FLUCTUATIONS

The water-level and pressure fluctuations and other well data given in table 17 were obtained from records of water-stage recorders and pressure gages installed by the U.S. Geological Survey and its cooperating agencies. The fact that no fluctuations are shown for some States does not necessarily mean that the earthquake did not cause any; it may mean that no recorders were installed in wells.

The time scale of the recorders is ordinarily not more than 2 inches per 24 hours, and thus the earthquake-induced fluctuations are usually represented on the recorder charts by a thick vertical line normal to the direction of movement of the pen along the time scale. The time of fluctuation can normally be read only to the nearest 10- or 15-minute interval. Consequently, it is difficult to correlate the water-level fluctuations in wells with different types of seismic waves, and the only useful data that recorder charts normally provide are the approximate time and maximum amplitude of the fluctuations.

The best records of earthquake-induced water-level fluctuations are those obtained by Rexin (1952) with a specially constructed recording gage. This instrument is, in essence, a water-level recorder that has an expanded time scale (0.915 cm per min) and a vertical scale (1:5) (E. E. Rexin, oral and written communications, 1959). Water-level records such as those obtained by Rexin show that the largest water-level and pressure fluctuations are produced by Rayleigh waves, although small fluctuations may be correlated with the arrival of both P and S waves (Vorhis, 1955, p. 50).

The main earthquake was followed by more than 200 aftershocks (U.S. Coast and Geodetic Survey, 1959, p. 8). There were four major aftershocks on the day of the earthquake and four on the next day, as reported by Murphy and Brazee, in chapter C, table 1 and repeated for convenience in table 18.

Of the approximately 600 water-level recorders operated by the Geological Survey in the conterminous United States, 185 showed water-level fluctuations caused by the Hebgen Lake earthquake. The maximum fluctuation observed exceeded 10 feet. The fluctuations were more than 1 foot in wells in 9 States, and between 0.5 and 1 foot in wells in 5 States. In some States there were water-level adjustments in wells in which earthquake fluctuations had not previously been recorded.

The seismic waves cause compression and dilation of the aquifer which in turn cause rises and declines of the water level or artesian pressure in wells. Further detailed study of this relation may lead to a better understanding of the elastic properties of aqui-

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TABLE 17.—*Fluctuations of water levels and pressure heads*

[Age: p€. Precambrian; €, Cambrian; O, Ordovician; S, Silurian; D, Devonian; M, Mississippian; ℙ, Pennsylvanian; P. Permian; TR, Triassic; K, Cretaceous; Cz, Cenozoic: T, Tertiary; E, Eocene; Mi, Miocene; Pl, Pliocene; Q, Quaternary; Pt, Pleistocene; R, Recent. Time: The Greenwich civil times refer to August 18, the day of the main earthquake, except those marked * (asterisk), which refer to August 19, and those marked † (dagger), which refer to August 20. Where no figures are given, time was not available. Tr, trace]

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
Alabama									
Jef-1	Sec. 29, T. 15 S., R. 3 W	Jefferson	Birmingham	140	Bangor limestone	M	6:40	43.4	1.3
COL-1	Sec. 30, T. 3 S., R. 10 W	Colbert	Muscle Shoals	265	Fort Payne chert and Tuscumbia limestone.	M	6:40	28.50	.73
ELM-1	Sec. 14, T. 19 N., R. 20 W	Elmore	Eclectic	402	Augen gneiss	p€	6:40	9.06	.60
MAR-1	Sec. 32, T. 12 S., R. 13 W	Marion	Guin	520	Pottsville sandstone	ℙ	6:40	7.82	.20
TW-7	Sec. 23, T. 14 S., R. 8 E	Calhoun	Jacksonville	95	Conasauga limestone	€	6:40	13.51	>1.0
						€	*16:00	13.55	.06
TW-14	Sec. 6, T. 15 S., R. 8 E	do	Anniston	213	do	€	6:40	50.65	.85
						€	*16:00	50.69	.05
TW-2	Sec. 32, T. 16 S., R. 7 E	do	do	355	Shale, streaks of limestone	€	6:40	20.44	.64
TW-9	Sec. 7, T. 13 S., R. 10 E	do	Piedmont	145	Rome formation: limestone	€	6:40	11.79	.09
C-33-T	Sec. 21, T. 5 S., R. 4 W	Morgan	Decatur	295	Fort Payne chert	M	6:40	13.41	.1
J-OBS-1	Sec. 5, T. 3 S., R. 4 W	Limestone	Athens	142	do	M	6:40	17.55	.4
CT-2	Sec. 5, T. 3 S., R. 4 W	do	do	133	do	M	6:40	13.55	.5
TW-11	Sec. 33, T. 2 S., R. 11 W	Lauderdale		227	do	M	6:40	14.14	>1.0
TW-42	Sec. 22, T. 4 S., R. 11 W	Colbert	Tuscumbia	325	Tuscumbia limestone	M	6:40	28.80	.65
TW-16	Sec. 4, T. 4 S., R. 11 W	do		327	Fort Payne chert	M	6:40	74.29	.23
TW-24	Sec. 15, T. 4 S., R. 11 W	do	Tuscumbia	352	Tuscumbia limestone	M	6:40	60.40	.42
TW-15	Sec. 9, T. 4 S., R. 11 W	do	do	286	do	M	6:40	21.42	.04
W-6	Sec. 12, T. 5 S., R. 1 W	Madison					6:40	11.90	.93
CT-79	Sec. 32, T. 3 S., R. 1 W	do	Huntsville	146	Fort Payne chert	M	6:40	11.89	.35
MT-3	Sec. 29, T. 5 S., R. 1 E	do	do	356	Tuscumbia limestone	M	6:40	16.87	.29
D-3	Sec. 25, T. 3 S., R. 1 W	do	do		do	M	6:40	35.01	.15
Arizona									
(D-4-2) 13bcc	Sec. 13, T. 4 S., R. 2 E	Pinal		534	Alluvium	Q	7:00	73.73	0.05
(D-5-8) 16dda	Sec. 16, T. 5 S., R. 8 E	do		200	do	Q	7:00	129.08	.06
(D-6-5) 25ccc	Sec. 25, T. 6 S., R. 5 E	do		102	do	Q	7:00	54.25	.30
Arkansas									
6S-5W-21cab	Sec. 21, T. 6 S., R. 5 W	Jefferson	0.35 mi. E of Reydel	75	Alluvium: coarse sand	Q	6:30	11.86	0.01
6S-7W-32ccc	Sec. 32, T. 6 S., R. 7 W	do	10 mi. SE of Pine Bluff	115	Terrace dep: medium to coarse sand.	Q	7:15	19.48	.02
7S-3W-6ddd	Sec. 6, T. 7 S., R. 3 W	Arkansas	1 mi S of Gillette	113	Very coarse sand	Q	6:50	52.14	.01
8S-3W-tract 2286	T. 8 S., R. 3 W	do	6 mi S of Gillette	125	Terrace deposit: medium sand.	Q	6:30	19.03	.01
8S-3W-33abd	T. 8 S., R. 3 W	Desha	8 mi NE of Dumas	60	Terrace deposit: medium sand, gravel.	Q	6:50	11.83	.01
11S-2W-3cca	T. 11 S., R. 2 W	do	14 mi SE of Dumas	754	Claiborne group: sand	T	7:00	14.84	.06
California									
2N/7E-2C1	Sec. 2, T. 2 N., R. 7 E	San Bernardino	Twenty-nine Palms, Surprise Springs.	400	Gravel, sand, and clay lenses.	T and Q	7:00	40.96	0.56 .22
2N/7E-4H1	Sec. 4, T. 2 N., R. 7 E	do	Surprise Spring Basin, Twenty-nine Palms.	500	Alluvium: gravel, sand, and clay lenses.		7:00	191.36	.12
3N/8E-29C1	Sec. 29, T. 3 N., R. 8 E	do	Deadman Basin, Twenty-nine Palms.	800	Alluvium: sand, gravel, and clay lenses.		7:30	88.97	.14
8S/2W-20L1	Sec. 20, T. 8 S., R. 2 W	Riverside	Wolf Valley, near Temecula.	524	Alluvium	Q	7:00	40.35	.06
8S/2W-20B2	Sec. 20, T. 8 S., R. 2 W	do	Pauba Valley near Temecula.	213	Older alluvium	Pt	7:00	25.98	.22
26/40-17N1	Sec. 17, T. 26 S., R. 40 E	Inyokern Nots	China Lake, Indian Wells Valley.	178	Alluvium: gravel, sand, and clay.	Pt	7:00	100.90	[2] .06
26/40-22N1	Sec. 22, T. 26 S., R. 40 E	Kern	China Lake, Indian Wells Valley.	203	Sand, silt, and clay	Q	7:30	72.21	.08
26/40-22P1	Sec. 22, T. 26 S., R. 40 E	do	do	850	Alluvium: sand and clay	Pt	7:00	67.96	.19
17/17-21N1	Sec. 21, T. 17 S., R. 17 E	Fresno	Mendota-Huron	1,005	Alluvium	Pt(?) and R	8:00	295.03	.10
2 3/25-16N3	Sec. 16, T. 23 S., R. 25 E	Tulare	Terra Bella-Lost Hills	430	do	Pt and R	8:00	237.64	.50
23/25-16N4	Sec. 16, T. 23 S., R. 25 E	do	do	250	do	Pt(?) and R.	6:00	118.62	0.10
23/25-17Q3	Sec. 17, T. 23 S., R. 25 E	do	do	355	do	Pt(?) and R.	6:00	121.71	1.07
12/12-16H5	Sec. 16, T. 12 S., R. 12 E	Fresno	Mendota-Huron	720	do	Pt(?) and R.	6:00	130.00	.27
14/14-5H1	Sec. 5, T. 14 S., R. 14 E	do	do	401	do	Pt(?) and R.	7:00	94.88	.04

See footnotes at end of table.

TABLE 17.—*Fluctuations of water levels and pressure heads*—Continued

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
				California—Continued					
15/16-20R1	Sec. 20, T. 15 S., R. 16 E.	Fresno	Mendota-Huron	350	Alluvium	Pt(?) and R.	8:00	79.93	1.11
						Pt(?) and R.	*18:00	79.91	.04
15/16-34E1	Sec. 34, T. 15 S., R. 16E.	do	do	500+	do	Pt(?) and R.	7:00	179.99	.16
16/16-2M1	Sec. 2, T. 16 S., R. 16 E.	do	do	1,047	do	Pt(?) and R.	6:00	184.94	.43
26/26-10R1	Sec. 10, T. 26 S., R. 26 E.	Kern	Terra Bella-Lost Hills	1,000	Older alluvium	Pl and Pt(?).	6:00	373.88	.39
						Pl and Pt(?).	13:00	373.96	.04
6/30-29D3	34°34′25″, 120°04′15″	Santa Barbara	Santa Ynez Valley	43	Alluvium: coarse sand and gravel.	R	6:30	14.18	.04
7/34-12E1	Sec. 12, T. 7 N., R. 34 W.	do	Near Lompoc	385	Careaga sand: massive, fine to coarse sand.	Pl	7:00	305.50	.02
7/35-22N2	Sec. 22, T. 7 N., R. 35 W.	do	Santa Ynez Valley, 6 mi W of Lompoc.	194	Alluvium: sandy gravel	R	6:30	6.34	.70
7/35-33R1	34°38′47″, 120°33′54″	do	Lompoc Canyon, 6 mi W of Lompoc.	420	Careaga sand: coarse gravel.	Pl	6:30	109.70	.06
11/9-13L1	Sec. 13, T. 11 N., R. 9 E.	Kern	Boron area, Antelope Valley.	462	Sand, clay, and gravel; basalt.		7:00	152.37	.67
11/9-36C2	Sec. 36, T. 11 N., R. 9 E.	do	do	372	Sand, silt, clay, and gravel		7:00	107.86	.34
				Colorado					
C-23-42-13	Sec. 13, T. 23 S., R. 42 W.	Prowers	1.5 mi E of Holly	47	Alluvium: sand and gravel.	Pt and R.	3:18	7.09	0.03
				Florida					
Col-9	Sec. 5, T. 4 S., R. 17 E.	Columbia	Lake City	836	Limestone	T	6:45	87.28	0.01
H-13	Sec. 21, T. 27 S., R. 18 E.	Hillsborough	11 mi NW of Tampa post office.	300	do	T	6:30	6.43	.04
H-30	Sec. 32, T. 31 S., R. 19 E.	do	1.7 mi N of Ruskin	500	do	T	6:15	[3] +9.40	.06
H-500	Sec. 9, T. 32 S., R. 20 E.	do	Wimauma	330	do	T	6:30	49.19	.12
Pasco 16	Sec. 34, T. 25 S., R. 21 E.	Pasco	6.5 mi S of Dade City	1,008	do	T	6:45	59.09	.19
Volusia 31	Sec. 11, T. 18 S., R. 32 E.	Volusia	Alamana	113	do	T	6:15	4.61	.08
V-909-106-4	Sec. 33, T. 15 S., R. 32 E.	do	1.3 mi SW of Daytona Beach.	234	do	T	6:30	5.28	.14
V-910-105-1	Sec. 27, T. 15 S., R. 32 E.	do	0.8 mi SW of Daytona Beach.	498	do	T	6:15	14.26	.18
P-45	Sec. 36, T. 28 S., R. 23 E.	Polk	4 mi S of Lakeland post office.	768	do	T	6:55	62.56	.13
F-210	Sec. 13, T. 53 S., R. 41 E.	Dade		112	do	T	6:50	[4] 2.37	.19
G-553	Sec. 16, T. 55 S., R. 40 E.	do	12 mi SW of Miami	91	do	T	6:50	[4] 6.75	.14
G-580	Sec. 11, T. 55 S., R. 40 E.	do	Miami Springs	100	do	T	6:50	[4] 3.95	.23
S-19	Sec. 25 T. 53 S., R. 40 E.	do	do	95	do	T	6:50	[4] 1.45	.48
S-68	Sec. 19, T. 53 S., R 41 E.	do	do	61	do	T	6:50	[4] .52	.23
G-820	Sec. 9, T. 49 S., R. 42 E.	Broward	2 mi NW of Oakland Park.	224	do	T	6:50	[4] 5.35	.17
				Georgia					
7	32°08′17″, 81°22′54″	Effingham	Meldrim	431	Ocala limestone	E	6:00	14.74	0.18
17	32°03′30″, 81°05′45″	Chatham	Savannah	505	do	E	6:00	108.50	.23
63	32°05′07″, 81°05′49″	do	do	525	do	E	6:00	103.25	.22
82	32°06′10″, 81°07′55″	do	do	332	do	E	6:00	104.42	.10
99	32°04′30″, 81°03′30″	do	do	500	do	E	6:00	74.20	.27
317	32°02′00″, 80°54′15″	do	Cockspur Island	354	do	E	5:30	22.50	.37
382	32°06′10″, 81°08′55″	do	Savannah		do	E	6:00	92.66	.35
				Hawaii					
83	21°18′20″, 157°51′05″	Honolulu	Island of Oahu, Beretania and Kapiolani Sts., Honolulu.	509	Koolau volcanic series: basalt.	Pl(?)	5:00	1.31	0.0
193	21°23′37″, 157°56′52″	do	Island of Oahu, Waimalu Valley.	363	do	Pl(?)	6:00	5.36	.10
T-24	21°21′27″, 157°53′10″	do	Island of Oahu, Manaiki Gulch.	115	do	Pl(?)	7:30		Tr
T-52	21°24′20″, 157°55′45″	do	Island of Oahu, Waimalu Valley.	321	do	Pl(?)	7:30	148.41	.06

See footnotes at end of table.

TABLE 17.—*Fluctuations of water levels and pressure heads*—Continued

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
				Idaho					
7N-36E-22ac2	Sec. 22, T. 7 N., R. 36 E.	Jefferson	4 mi S of Hamer	35	Snake River basalt	Pl to R	7:00	8.02	[5] 3.35
						Pl to R	8:30	8.00	.22
						Pl to R	9:00	7.98	.04
						Pl to R	15:30	7.95	[5] .42
						Pl to R	*4:30	7.98	.05
7N-35E-20cb1	Sec. 20, T. 7 N., R. 35 E	do	North Lake game refuge	65	do	Pl to R	6:00	50.75	.24
						Pl to R	7:00	50.76	.05
						Pl to R	14:00	50.79	.03
7N-34E-4cd1	Sec. 4, T. 7 N., R. 34 E.	do	3 mi E, 1 mi N of Monteview.	57	do	Pl to R	6:30	24.94	4.91
						Pl to R	8:00	24.94	.68
						Pl to R	9:00	24.94	[6] .22
						Pl to R	11:00	24.95	.16
						Pl to R	15:30	24.95	[7] 1.76
						Pl to R	*4:00	24.96	.30
5N-34E-9bd1	Sec. 9, T. 5 N., R. 34 E.	do	4.5 mi S of Mud Lake	553	Basalt	Pt	6:30	258.95	4.67
						Pt	8:00	258.95	.17
						Pt	15:30	258.96	.55
53N-2W-9aa1	Sec. 9, T. 53 N., R. 2 W.	Kootenai	6 mi E, 1 mi N of Athol	351	Fluvioglacial gravel and sand.	Pt	6:30	230.99	.26
						Pt	15:30	231.01	.05
						Pt	*4:00	231.02	.02
5S-17E-26ac1	Sec. 26, T. 5 S., R. 17 E.	Lincoln	2 mi N of Shoshone	255	Snake River basalt	Pl to R	6:00	186.62	[8] >10.0
						Pl to R	7:00	186.62	.10
						Pl to R	15:00	186.71	.24
5S-33E-35cc1	Sec. 35, T. 5 S., R. 33 E.	Power	6 mi NW of Pocatello	60	Gravel	Pl to R	6:00	24.25	.10
4N-45E-13ad1	Sec. 13, T. 4 N., R. 45 E.	Teton	3.5 mi SE of Driggs	304	Alluvium	Pl to R	7:00	172.93	[8] >5.00
						Pl to R	7:30	173.02	.11
						Pl to R	8:00	173.07	.38
						Pl to R	9:00	173.08	.20
						Pl to R	11:30	173.15	.10
						Pl to R	15:30	173.25	.63
						Pl to R	*4:00	173.52	.21
4N-1W-35aa1	Sec. 35, T. 4 N., R. 1 W.	Ada	2 mi N, 1 mi W of Meridian.	44	Sand and gravel	Pl to R	18:00	5.18	.16
3N-2E-25bb2	Sec. 25, T. 3 N., R. 2 E.	do	2 mi SE of Boise	52	do	Pl to R	17:00	9.17	.09
1S-30E-15bc1	Sec. 15, T. 1 S., R. 30 E.	Bingham	7 mi W, 8 mi S of Atomic City.	752	Basalt	Pt	7:30	712.90	.69
5S-31E-27ab1	Sec. 27, T. 5 S., R. 31 E.	do	1 mi N, 1 mi E of Aberdeen.	46	do	Pt	8:00	12.23	.03
1S-19E-3cc2	Sec. 3, T. 1 S., R. 19 E.	Blaine	Near Gannett	51	Clay, silt, sand, and gravel.	Q	7:00	13.24	1.68
						Q	8:00	13.24	.10
						Q	15:00	13.26	.30
						Q	18:00	13.29	.04
2S-20E-1ac2	Sec. 1, T. 2 S., R. 20 E.	do	Near Picabo	209	Snake River basalt and alluvium.	Pl to R	6:00	143.32	.95
						Pl to R	14:00	143.33	.12
						Pl to R	*3:00	143.36	(?) .04
						Pl to R	†24:00	143.36	.09
						Pl to R	†17:00	143.44	(?) .06
6N-31E-13db1	Sec. 13, T. 6 N., R. 31 E.	Butte	12 mi W, 1 mi N of Mud Lake.	326	Snake River basalt	Pl to R	5:00	215.14	[8] >2.00
						Pl to R	6:00	215.14	.62
						Pl to R	9:00	215.13	.12
						Pl to R	13:00	215.20	[8] >2.00
						Pl to R	*3:00	215.20	.27
5N-29E-23cd1	Sec. 23, T. 5 N., R. 29 E.	do	3 mi S, 2 mi E of Howe	401	do	Pl to R	7:30	272.31	3.82
						Pl to R	8:30	272.30	.09
						Pl to R	9:00	272.30	.05
						Pl to R	16:00	272.28	.36
						Pl to R	*4:30	272.27	.05
3N-29E-14ad1	Sec. 14, T. 3 N., R. 29 E.	do	17 mi E, 3 mi S of Arco	588	basalt	Pl to R	4:00	452.06	[8] >2.00
						Pl to R	6:00	452.06	[8] >2.00
						Pl to R	8:00	452.04	.11
						Pl to R	13:00	452.04	[8] >2.00
						Pl to R	*2:00	452.03	.17
2N-27E-33ac2	Sec. 33, T. 2 N., R. 27 E.	do	12 mi S of Arco	1,200	Snake River basalt	Pl to R	*7:00	985.56	.04
				Illinois					
ANL-9	Sec. 9, T. 37 N., R. 11 E.	DuPage		140	Niagara dolomite	S	6:30	94.61	0.14
ANL-10	Sec. 9, T. 37 N., R. 11 E.	do		198	do	S	6:30	79.85	.96
ANL-20	Sec. 10, T. 37 N., R. 11 E.	do		168	do	S	6:30	42.74	.11
ANL-38	Sec. 5, T. 37 N., R. 11 E.	do		173	do	S	6:30	101.76	.02
				Indiana					
Cs. (25/3-10A1)	Sec. 10, T. 25 N., R. 3 E.	Cass	Bunker Hill A. F. Base, Peru.	130	Limestone	S	6:45	20.96	0.11
Md 8	Sec. 15, T. 21 N., R. 6 E.	Madison	Elwood	415	do	S	7:00	30.77	.07
Ma 31	Sec. 6, T. 16 N., R. 5 E.	Marion	Indianapolis, Fort Benjamin Harrison.	347	Niagara dolomite	S	7:00	102.1	.56
Ma 32	Sec. 36, T. 17 N., R. 3 E.	do	Indianapolis	322	do	S	4:00	13.05	.05
Mi(25/3-1H1)	Sec. 1, T. 25 N., R. 3 E.	Miami	Bunker Hill A. F. Base, Peru.	182	Limestone	S	6:30	21.06	.49
						S	15:30	21.04	.03
Mi(25/3-1H2)	Sec. 1, T. 25 N., R. 3 E.	do	do	183	do	S	6:30	18.09	.17
						S	15:30	18.08	.01
Mi(25/3-1H4)	Sec. 1, T. 25 N., R. 3 E.	do	do	182	do	S	6:45	23.20	.35
Mi.(26/3-35C1)	Sec. 35, T. 26 N., R. 3 E.	do	do	143	do	S	6:30	35.55	.30
						S	15:30	29.70	.03
Pu 6.(29/4W-4L1)	Sec. 4, T. 29 N., R. 4 W.	Pulaski	Francisville	663	Niagara dolomite	S	6:45	16.15	.20

See footnotes at end of table.

TABLE 17.—*Fluctuations of water levels and pressure heads*—Continued

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
				Kentucky					
8245-3745-404	37°46′, 82°45′	Johnson	West Van Lear	115	Breathitt formation: sandstone.	P	9:00	28.25	0.28
8400-3705-40	37°07′, 84°04′	Laurel	London	400	Lee formation: sandstone	P	*3:00	48.14	.06
				Maryland					
Cal-GD6	38°19′58″, 76°27′08″	Calvert	Solomons	493	Aquia greensand	E	9:00	34.44	(?) 0.04
						E	12:40	34.75	(?) .05
				Michigan					
7N7E17-1	Sec. 17, T. 7 N., R. 7 E.	Genesee	Flint; Franklin Street and Sunnyside Ave.	222	Saginaw formation: sandstone, shale, coal, and limestone.	P	6:45	35.68	0.60
						P	15:45	35.83	.06
						P	*4:15	36.13	.07
2N4E9-1	Sec. 9, T. 2 N., R. 4 E.	Livingston	Near Howell	280	Saginaw formation: Parma sandstone, shale, coal and limestone.	P	6:45	112.04	.21
3N10E32-1	Sec. 32, T. 3 N., R. 10 E.	Oakland	Pontiac; Walnut and Wessen Streets.	160	Alluvium: sand and gravel.	Pt	6:45	116.40	.14
39N23W28-3	Sec. 28, T. 39 N., R. 23 W.	Delta	Near Escanaba	530	Munising sandstone	€	7:00	3.11	.26
41N5W23-1	Sec. 23, T. 41 N., R. 5 W.	Mackinac	Near Moran	47	Salina formation: limestone	S	7:00	14.45	.13
				Minnesota					
28.22.22bdd1	Sec. 22, T. 28 N., R. 22 W.	Dakota	South St. Paul	264	Jordan (?) sandstone	€	7:00		0.16
117.22.5abd2	Sec. 5, T. 117 N., R. 22 W.	Hennepin	Wayzata	483	Jordan sandstone	€	6:30		.93
117.22.8dbd2	Sec. 8, T. 117 N., R. 22 W.	do	Woodland	503	do	€	6:30		.27
117.23.11bbd1	Sec. 11, T. 117 N., R. 23 W.	do	Orono	437	do	€	6:30		.20
117.23.34daa2	Sec. 34, T. 117 N., R. 23 W.	do	Excelsior	467	do	€	6:30		.67
						€	15:30		(?).05
58.17.7daa1	Sec. 7, T. 58 N., R. 17 W.	St. Louis	Virginia	357	Biwabik iron formation	p€	(?)6:00		.33
58.17.18ccc1	Sec. 18, T. 58 N., R. 17 W.	do	do	118	Virginia shale: fractured argillite.	p€	6:30		.16
				Montana					
6-20-8aa	Sec. 8, T. 6 N., R. 20 W.	Ravalli	3.5 mi NE of Hamilton	20.5	Fluvial sand and gravel	Q	6:45	4.96	0.12
7-20-21cd	Sec. 21, T. 7 N., R. 20 W.	do	7 mi NE of Hamilton	13.7	do	Q	6:45	4.45	.14
						Q	15:30	4.46	.03
						Q	19:00	4.50	.01
						Q	20:30	4.51	.01
						Q	*4:00	4.54	.01
8-20-19ba3	Sec. 19, T. 8 N., R. 20 W.	do	2 mi N of Victor	52.0	Sandy clay	T	6:45	13.19	>.01
9-20-12bb	Sec. 12, T. 9 N., R. 20 W.	do	3.5 mi N of Stevensville	19.7	Alluvium	Q	6:45	5.46	.18
9-20-26ba1	Sec. 26, T. 9 N., R. 20 W.	do	Stevensville	22.7	do	Q	6:45	4.22	.40
						Q	15:30	4.25	.08
						Q	*4:00	4.29	.04
10-20-14ba1	Sec. 14, T. 10 N., R. 20 W.	do	Florence	43.8	Glaciofluvial deposit	Q	6:45	24.99	.16
13-19-8cc	Sec. 8, T. 13 N., R. 19 W.	Missoula	1 mi W of Missoula	112.3	Sand, gravel, and silt	Q	6:45	38.13	.65
						Q	15:30	38.21	.12
						Q	*4:00	38.28	.05
13-19-29da	Sec. 29, T. 13 N., R. 19 W.	do	Missoula		do	Q	6:45	39.39	1.76
						Q	8:00	39.40	.05
						Q	8:45	39.40	.01
						Q	15:30	39.42	.15
						Q	*4:00	39.48	.14
14-21-13bb	Sec. 13, T. 14 N., R. 21 W.	do	12 mi NW of Missoula	25	Sand and gravel	Q	6:45	10.91	.55
						Q	8:00	10.91	.01
						Q	8:45	10.90	.01
						Q	15:30	10.89	.03
						Q	*4:00	10.87	.04
35-24-9bd	Sec. 9, T. 35 N., R. 24 E.	Blaine	6 mi SE of Hogeland	35	Flaxville formation	Mi or Pl	7:00	33.63	.14
						Mi or Pl	10:50	33.62	.12
						Mi or Pl	*2:00	33.62	.04
				Nebraska					
A10-6-36cdd		Lancaster	Lincoln, Irvingdale Park	170	Dakota sandstone	K		67.31	0.23

See footnotes at end of table.

TABLE 17.—*Fluctuations of water levels and pressure heads*—Continued

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
			Nevada						
S19/60-9bcc1	Sec. 9, T. 19 S., R. 60 E.	Clark	Las Vegas Valley	830	Alluvium	Pl and Pt(?)	[9] 7:00 [9] 8:30	99.56 99.70	>1.00 .03
S19/60-33baa1	Sec. 33, T. 19 S., R. 60 E.	do	do	1,008	do	Pl and Pt(?)	[9] 7:00	21.86	[10] .45
S22/61-4bcc1	Sec. 4, T. 22 S., R. 61 E.	do	do	355	do	Pl and Pt(?)	[9] 7:00	99.0	(?) [11] .29
			New Jersey						
25.14.8.2.2		Morris	Madison	181	Wisconsin drift: coarse sand and gravel.	Pt	7:00	[4] 189.64	0.08
25.14.9.4.1		do	do	100	Wisconsin drift: sand, gravel, and clay.	Pt	7:00	174.18	.09
26.22.4.4.4		Union	Hillside well 4	400	Brunswick formation: red shale.	R̄	7:00 16:00	24.42 24.57	.72 .06
31.1.6.4.8		Camden	Camden, Cooper St. and Delaware Ave.	300	Raritan formation: sand and gravel.	K	7:00	[3] +8.80	.09
25.14.3.5.5		Morris	About 3 mi E of Whippany.	170	Wisconsin drift: sand, gravel, and clay.	Pt	7:00	175.89	.02
26.21.5.4.6		Union	Near Kenilworth	290	Brunswick formation	R̄	7:00	60.15	.04
			New Mexico						
10.24.9.333	Sec. 9, T. 10 S., R. 24 E.	Chaves	Berrendo well	258	San Andres limestone	P	6:40	48.67	0.22
10.24.21.212	Sec. 21, T. 10 S., R. 24 E.	do	Berrendo-Smith well	324	do	P	6:40	47.37	.75
13.25.27.211	Sec. 27, T. 13 S., R. 25 E.	do	Greenfield	880	do	P	6:40	145.43	1.34
11.23.3.342	Sec. 3, T. 11 S., R. 23 E.	do	Mask	595	do	P	6:40	181.71	.04
20.26.17.334	Sec. 17, T. 20 S., R. 26 E.	Eddy	Seven Rivers well	70	Seven Rivers formation	P	6:40	55.18	.05
22.26.2.242	Sec. 2, T. 22 S., R. 26 E.	do	Hillcrest School		Capitan limestone	P	6:40	57.47	.09
24.29.17.444	Sec. 17, T. 24 S., R. 29 E.	do	USGS no. 11, Carlsbad	246	Rustler formation	P	6:40	2.88	.02
24.29.30.414	Sec. 30, T. 24 S., R. 29 E.	do	USGS no. 13, Carlsbad	66	do	P	6:40	18.34	.14
12.11.9.221	Sec. 9, T. 12 N., R. 11 W.	Valencia	Bluewater	500	San Andres limestone	P	6:40	168.19	.59
6	Sec. 4, T. 14 S., R. 4 W.	Sierra	Dakos, Hot Springs	105	Magdalena group	P̄ and P	6:40	[3] +.66	(?)1.34
			New York						
SA 529	43°03′27″, 73°47′53″	Saratoga	Saratoga Springs	189	Little Falls dolomite	Є	[9] 7:15	51.44	0.28
Cu 10	42°08′07″, 79°12′14″	Chautauqua	Near Jamestown	150	Gravel	Pt	[9] 7:00	48.14	.22
			Ohio						
Cl-1	39°55′50″, 83°51′12″	Clark	Springfield	57	Glacial outwash gravel	Pt		4.57	0.24
Dl-3	40°21′42″, 83°04′00″	Delaware	Delaware Dam	135	Columbus limestone	D		30.00	.11
Ge-3a	41°25′, 81°22′	Geauga	Chagrin Falls	120	Cuyahoga formation	M		20.41	.08
Gr-3	39°49′, 83°53′	Greene	Yellow Springs	116	Brassfield limestone	S		57.54	.10
Ho-1	40°36′, 81°55′	Holmes	Millersburg	43	Logan formation	M		4.86	.04
Mu-2	39°57′, 82°02′	Muskingum	Zanesville	53	Glacial outwash	Pt		8.40	.02
R-2	40°41′, 82°35′	Richland	Lexington	129	Black Hand sandstone	M		26.89	.02
			Oklahoma						
66	Sec. 33, T. 4 N., R. 8 W.	Grady		254	Rush Springs sandstone	P	6:00 14:00	85.67	0.28 (?).05
			Pennsylvania						
7			Mine shaft in northern anthracite field (p. 176).				6:30	256.70	0.10
14			do				6:30	215.00	.17
Henry Mine			do				6:30	217.92	.05
			Puerto Rico						
24-66.3-2			Barrio Hato Rey Sur, Municipio Rio Piedras.	105	Alluvial deposit		11:20	1.86	(?)0.01
			South Carolina						
BFT-119(M-2)	32°26′15″, 80°46′30″	Beaufort	3 mi W of Burton	93	Limestone	E	7:00	24.10	0.10
46	32°18′05″, 81°58′10″	Jasper	3 mi S of Okatie	334	do	E	5:30	20.67	.29
BFT-121(M-4)	32°28′, 80°44′	Beaufort	Marine Corps Air Station.	105	do	E	7:00	14.17	.15
BFT-113(P1-25)	32°27′30″, 80°45′	do	Burton	107	do	E	7:00	25.20	.15

See footnotes at end of table.

TABLE 17.—*Fluctuations of water levels and pressure heads*—Continued

Well	Location [1]	County	Area	Depth of well (ft)	Water-bearing formation and (or) type of material	Age	Time	Depth to water before disturbance (ft)	Double amplitude of fluctuation (ft)
Tennessee									
7:1-6	36°35′, 84°04′	Campbell	Jellico	620	Limestone and sandstone	P	7:00	76.09	0.34
						P	16:00	76.14	.03
Texas									
B-166		Lamb	About 4 mi S of Earth	202	Ogallala formation	Pl	6:50	37.56	0.15
436		Bexar	San Antonio	756	Edwards limestone	K	6:45	61.87	.60
S1-18		Dimmit	4 mi SW of Carrizo Springs.	320	Carrizo sand	E	6:55	131.35	.20
Utah									
(B-2-1) 24bad-3		Davis	Woods Cross	386	Valley fill	Cz	6:50	[3] +24.6	3.8
(C-16-7) 12dcd-5		Millard	5 mi N of Delta	480	do	Cz	7:10	[3] +22.2	5.1
(C-7-8) 10cbd-1		Tooele	Dugway	175	Gravel	Cz	6:40	88.30	.55
(C-2-4) 33add-1		do	Erda	165	Sand and gravel	Cz	6:40	11.76	[8] >1.00
(C-35-11) 33dbc-1		Iron	Cedar City	140	Gravel	Cz	6:55		.78
Washington									
20/3-18-C1	Sec. 18, T. 20 N., R. 3 E.		Tacoma	185	Sand and gravel	Pt	6:45	96.23	1.15
						Pt	8:00	96.22	.09
						Pt	8:30	96.22	.03
						Pt	15:30	96.25	.27
						Pt	*16:10	96.23	.00
20/25-21A2	Sec. 21, T. 20 N., R. 25 E.	Grant	Roads 9 NW and 1 NW.	652	Columbia River basalt	M	6:00	19.54	.29
22/27-19N2	Sec. 19, T. 22 N., R. 27 E.	do	Soap Lake (Div. St. and 4th Ave.).	275	do	T	1:00	35.69	[9] .05
10/17-5E2	Sec. 5, T. 10 N., R. 17 E.	Yakima	White Swan	15	Alluvium: sand and gravel.	R	6:15	7.05	.04
						R	14:30	7.06	.03
						R	*14:15	7.08	.14
10/19-1N1	Sec. 1, T. 10 N., R. 19 E.	do	McKinley and Ft. Simcoe Roads.	21	do	R	6:45	6.52	.03
26/45-32J2	Sec. 32, T. 26 N., R. 45 E.	Spokane	Campbell and Trent Roads.	155	Sand and gravel	Pt	7:15	123.29	.35
						Pt	14:00	123.33	.03
Wisconsin									
Lf-57	Sec. 33, T. 1 N., R. 2 E.	Lafayette	Shullsburg	265	Galena dolomite	O	6:30	134.57	>1.00
						O	8:45	134.53	.03
						O	15:30	134.54	.18
						O	*4:00	134.47	.04
Lf-121	Sec. 35, T. 1 N., R. 2 E.	do	do	300	do	O	7:00	74.64	.55
						O	16:30	74.80	.03
						O	*5:00	74.80	.02
Lf-183	Sec. 34, T. 1 N., R. 1 E.	do	New Diggings	268	do	O	6:00	91.47	.46
						O	7:00	91.48	.01
						O	14:30	91.49	.02
Ml-121	Sec. 13, T. 5 N., R. 22 E.	Milwaukee	South Milwaukee	268	Niagara dolomite	S	6:00	67.82	.07
Fl-12	Sec. 11, T. 15 N., R. 17 E.	Fond du Lac	Fond du Lac	817	Galena, Platteville, and St. Peter formations.	Є and O	6:00	69.25	.10

[1] Where section, township, and range are not given, figures refer to latitude north and longitude west.
[2] Single-amplitude fluctuation.
[3] Water level is plus when piezometric surface is above land surface.
[4] Water levels are referred to mean sea level.
[5] Small fluctuations about 10 minutes later.
[6] Small fluctuations about 30 minutes later.
[7] Two small fluctuations about 15 and 30 minutes later.
[8] Complete revolution of chart; pen retraced mark; unable to determine highest and lowest points of fluctuation.
[9] Within half an hour, more or less.
[10] Float sticks occasionally. Residual change in water level may be due to malfunction of instrument.
[11] Float rubs on casing; double amplitude probably greater than 0.29 foot.

TABLE 18.—*Magnitudes of the main earthquake and the aftershocks*

Main quake or aftershock	Date	Time (GCT) [1]	Location		Magnitude
			Lat. (N.)	Long. (W.)	
Main quake	Aug. 18	06 37 15	44°50′	111°05′	7.1 (Pasadena).
Aftershock 1	do	07 56 18	45°	110.5°	6.5 (Berkeley).
2	do	08 41 50	44.8°	110.7°	6.0 (Berkeley).
3	do	11 03 52	44.8°	111.1°	5.5–5.75 (Berkeley).
4	do	15 26 06.5	44°53′	110°44′	6.5 (Pasadena).
5	Aug. 19	04 04 03	44°54′	111°38′	6.0 (Berkeley).
6	do	19 06 29	45°	111.4°	
7	do	19 43 47.5	45°	110.5°	
8	do	21 45 57	45°	111.5°	

[1] Greenwich civil time.

fer material, an important factor governing the storage characteristics of artesian reservoirs. Data on the wells and their water-level fluctuations are given in table 17. Résumés of the fluctuations, by States, follow.

Alabama.—Twenty wells, whose depths range from 95 to 402 feet, showed fluctuations. The double amplitude exceeded 1 foot in three of the wells (Jef–1, TW–7, and TW–11); it was between 0.5 and 1 foot in six others. Two fluctuations caused by aftershocks were recorded in well Jef–1, in addition to that caused by the main quake. The first of these occurred 2 hours after the main fluctuation and had a double amplitude of 0.1 foot; the other, about 8 hours after the main one, had a double amplitude of 0.2 foot. Wells TW–7 and TW–14 also recorded an additional fluctuation about 11 hours after the main one; they may have been affected by aftershock 4 (table 18).

Alaska.—Of the four recorders installed on wells at Anchorage, two showed no fluctuations; one, no record; and 1, a questionable fluctuation of 0.3 foot.

Arizona.—Fluctuations were recorded in three wells; the double amplitudes were 0.04, 0.06, and 0.30 foot.

Arkansas.—The double amplitudes of the fluctuations in six wells ranged from 0.01 to 0.06 foot. Five of the wells are drilled in alluvium and terrace deposits of Quaternary age.

California.—Fluctuations were recorded in 24 wells. The maximum double amplitude was 1.11 feet in a well 350 feet deep, drilled in Recent and Pleistocene(?) alluvium. Five other wells showed fluctuations of 0.5 foot or more. The wells range in depth from 43 to more than 1,000 feet. Water-level fluctuations that might be attributed to aftershocks were observed in wells 15/16–20R1 and 26/26–10R1 (table 17). The time correspondence is very poor, however, and it is possible that these fluctuations were due to other causes.

Colorado.—Only one well showed fluctuations that might be attributed to the earthquake. This well (C–23–42–13), in Prowers County, is 47 feet deep and is drilled in sand and gravel of Recent to Pleistocene age; the thickness of the saturated material is about 40 feet. The double amplitude of the fluctuation was 0.03 foot. The time of occurrence, however, according to the recorder chart, was about 3:18 GCT (Greenwich civil time) on August 18, or more than 3 hours before the earthquake. The time correspondence is therefore very poor, probably due to malfunction of the recorder timing mechanism.

Florida.—Fluctuations were recorded in 15 wells. The maximum double amplitude ranged from 0.01 to 0.48 foot. All the wells are drilled in limestone of Tertiary age, and their depths range from 61 to more than 1,000 feet. Although wells in Florida ordinarily show significant fluctuations in response to earthquakes, none corresponding to the aftershocks were observed.

Georgia.—Water-level fluctuations ranging in maximum double amplitude from 0.1 to 0.37 were observed in seven wells. The wells are 332 to more than 500 feet deep and are drilled into limestone of late Eocene age.

Hawaii.—The fluctuations in four wells had a maximum double amplitude range of from a trace to 0.10 foot. In one well, operated by the Scripps Institute of Oceanography, a large pen displacement stopped the recorder; it was not possible to fix the time or cause. The wells range in depth from 115 to 509 feet and are drilled into basalt of the Koolau volcanic series of Pliocene(?) age.

Idaho.—Water-level fluctuations were observed in 18 wells. The maximum double amplitude ranged from 0.10 foot in well 5S–33E–35ccl to more than 10 feet in well 5S–17E–26acl near Shoshone, Lincoln County. Wells in Jefferson County near Monteview and Mud Lake fluctuated 4.91 and 4.67 feet, respectively.

A well at Mud Lake collapsed as a result of the earthquake. It was 387 feet deep and was drilled to 104 feet in sedimentary deposits; at this depth the casing was set at the top of the underlying basalt. The casing and the pump in the well dropped 10 to 15 feet, and an area about 10 feet in diameter caved around the well. It is uncertain if the earthquake did any more than trigger the collapse, for another well in the same vicinity and in similar rocks collapsed before the earthquake.

Of particular interest are the fluctuations caused by aftershocks (figs. 99, 100) that were observed in 12 of the 18 wells in Idaho. Wells 7N–34E–4cdl and 4N–45E–13adl showed fluctuations that can be identified with five of the aftershocks. Water levels in another 10 wells fluctuated in response to 4 or fewer after-

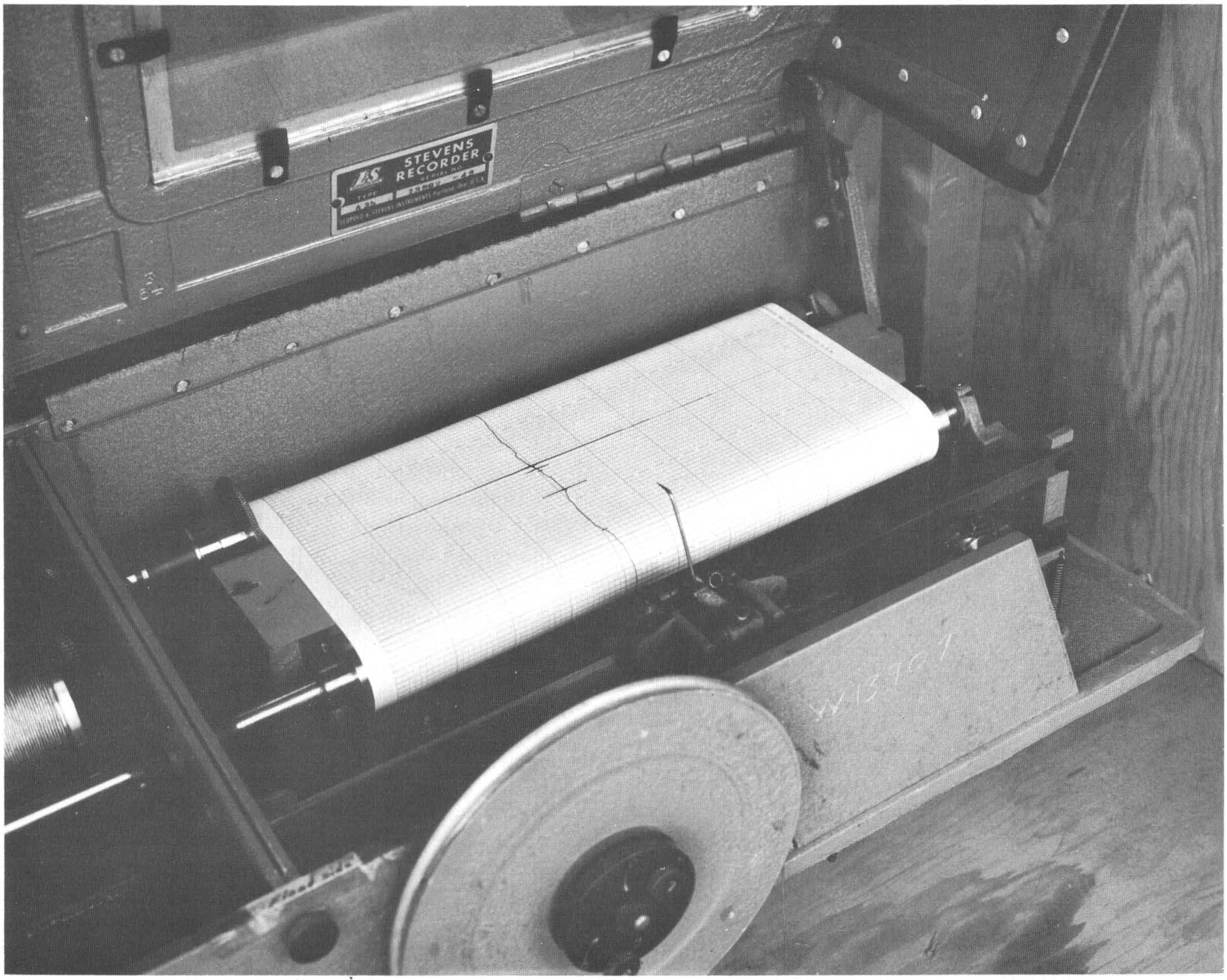

FIGURE 99.—Water-level recorder showing fluctuations caused by the Hebgen Lake earthquake. Well in sec. 22, T. 7 N., R. 36 E., Jefferson County, Idaho. Photograph by Atomic Energy Commission.

shocks. Aftershock 4 was detected in 11 wells, which was the greatest number of wells affected, and aftershock 1 was the next most effective. Both aftershocks were of magnitude 6.5 (table 18). The maximum double amplitude of the fluctuations recorded for aftershocks 4 and 1 in wells in Idaho was 0.68 foot and more than 2 feet, respectively (table 17).

The effects of aftershock 5, which occurred the next day (August 19), were observed in seven wells. The maximum observed double amplitude was 0.30 foot.

Illinois.—Water-level fluctuations were observed in four observation wells at the Argonne National Laboratory; the maximum double amplitude was 0.96 foot in well ANL–10. All the wells are drilled into the Niagara dolomite of Silurian age.

Indiana.—Nine wells showed fluctuations. In two, the fluctuations were about 0.5 foot. In the other seven the range was from 0.01 to 0.35 foot. Four wells fluctuated at times corresponding to aftershock 4. The maximum fluctuation caused by this aftershock was 0.05 foot in well Ma 32.

Kentucky.—Water-level fluctuations were observed in two wells in the Eastern Coal Field region; the maximum double amplitudes were 0.06 and 0.28 foot.

Maryland.—One well, Cal–GD6, showed fluctuations that might be attributed to the earthquake and to one of the aftershocks. Both the time of occurrence of the fluctuations and their unusual pattern, however, make it doubtful that they were induced by earthquakes.

Michigan.—Five of the observation wells showed maximum fluctuations of from 0.06 to 0.60 foot. The

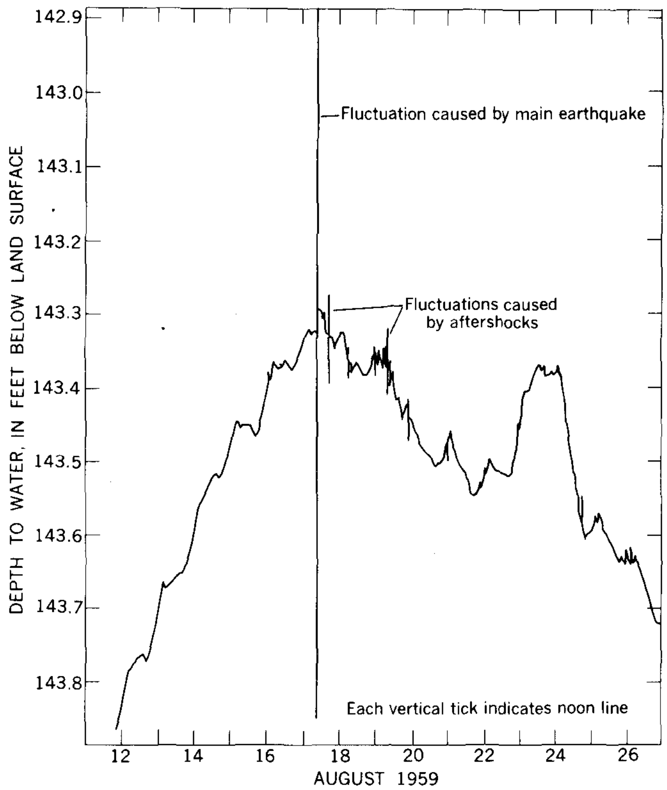

FIGURE 100.—Hydrograph of well 2S–20E–1ac2, Blaine County, Idaho, showing water-level fluctuations caused by the main earthquake and aftershocks.

wells range in depth from 47 to 530 feet. The water level in well 7N7E17–1 fluctuated 0.7 foot in response to aftershock 5.

Minnesota.—Maximum fluctuations of 0.05 to 0.93 foot were observed in seven wells whose depths range from 118 to 503 feet.

Montana.—Ten wells showed that maximum fluctuations resulting from the main quake ranged from 0.01 to 1.76 feet. In addition, six of the wells registered fluctuations of 0.01 to 0.15 foot that may have resulted from aftershocks.

Nebraska.—The only record available, that of a well in Lincoln that ends in the Dakota sandstone (Cretaceous), showed a maximum fluctuation of 0.23 foot.

Nevada.—The water level in one well (S19/60–9bccl) fluctuated more than 1 foot in response to the main quake and slightly in response to an aftershock. The fluctuation shown in table 17 for wells S19/60–33baal and S22/61–4bccl may be considerably smaller than the real fluctuation, as the floats in both these wells occasionally stick in the casing.

New Jersey.—Maximum fluctuations of 0.02 to 0.72 foot were observed in six wells. The water level in one of the wells (26.22.4.4.4) fluctuated 0.06 foot in response to aftershock 4. The wells range in depth from 100 to 400 feet.

New Mexico.—The maximum double amplitudes in the 10 observation wells that registered fluctuations ranged from 0.02 foot to 1.34 feet. The Dakos well (No. 6) in the Hot Springs area, Sierra County, had such an abrupt fluctuation that the float tape was disengaged from the spines of the float wheel. The largest fluctuation was in the Greenfield well, which is the deepest (880 feet) of the 10 observation wells.

New York.—No fluctuations were observed on Long Island. In upstate New York, however, water levels in two wells fluctuated; the maximum double amplitude was 0.28 foot in well SA 529 in the eastern part of the State. The water level in this well fluctuated also in response to two of the aftershocks, but records of the magnitude are not available.

Ohio.—Fluctuations were observed in seven wells; the maximum double amplitudes ranged from 0.02 to 0.24 foot.

Oklahoma.—One well, No. 66, in Grady County, showed fluctuations corresponding to the main shock and to an aftershock, probably No. 3 or 4 (table 18).

Pennsylvania.—No fluctuations in wells were reported, but some unusual fluctuations were observed in the water levels of three mine-water pools in the northern anthracite field in northeastern Pennsylvania, 10 miles north of Wilkes-Barre. The double amplitudes ranged from 0.05 to 0.17 foot.

Puerto Rico.—Six observation wells on the island are equipped with recorders. Only one of these (24–66.3–2), in San Juan, showed a maximum fluctuation of as much as 0.01 foot. This well is the only one that, in the past, has shown water-level fluctuations that might be associated with earthquakes whose epicenters were near Puerto Rico.

South Carolina.—The maximum fluctuations observed in four wells ranged from 0.10 to 0.29 foot. The time correspondence of the fluctuations in three of the wells with the earthquake is very poor, and it is possible that they were not induced by the earthquake.

Tennessee.—The water level in a 620-foot well (7:1–6) at Jellico fluctuated in response to both the main quake and to aftershock 4. The maximum double amplitude in response to the main quake was 0.34 foot.

Texas.—The records from three wells show that the maximum double amplitudes of water-level fluctuations caused by the quake ranged from 0.15 to 0.60 foot.

Utah.—Records of five observation wells show maximum water-level fluctuations ranging from 0.55 foot

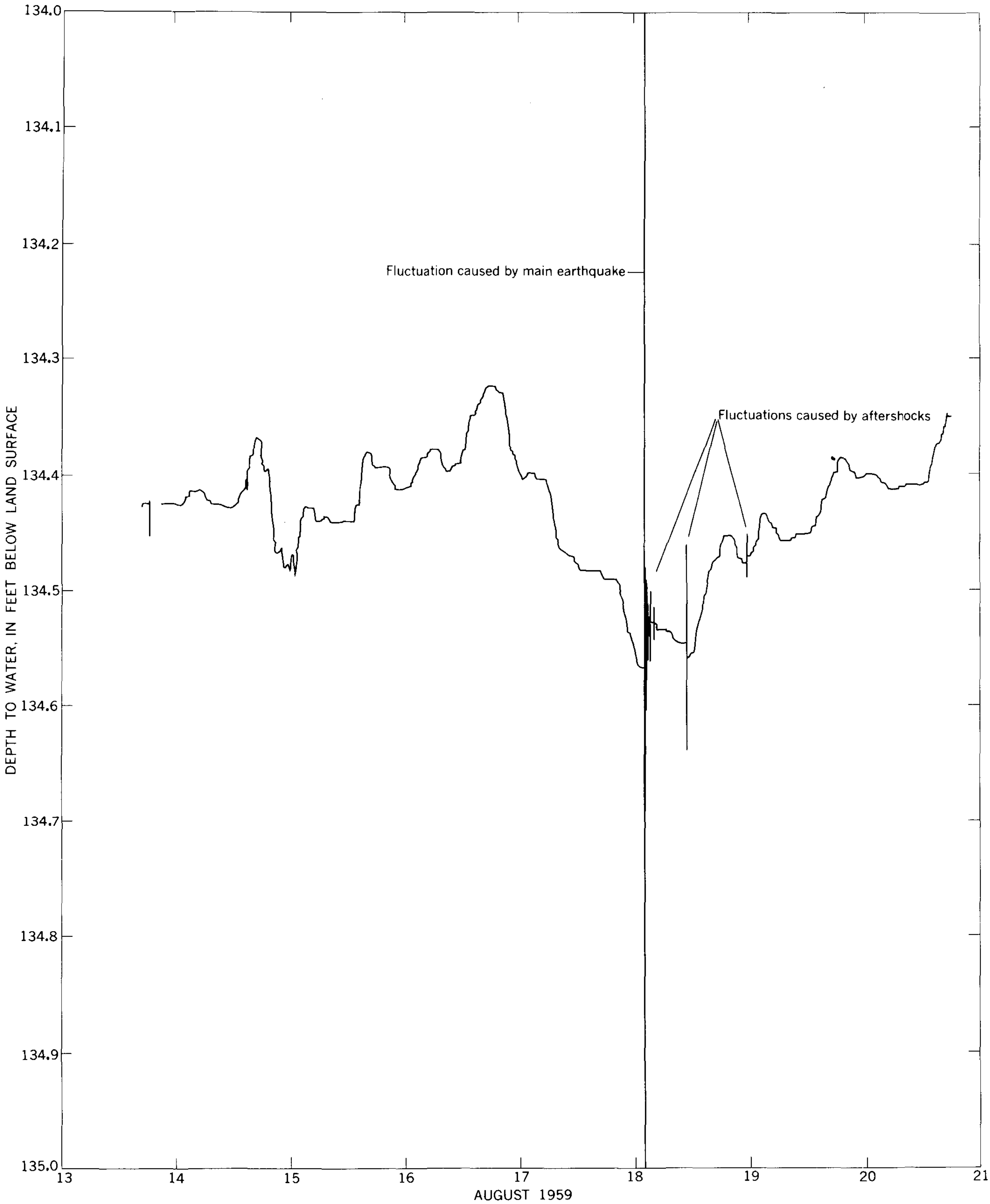

FIGURE 101.—Hydrograph of well Lf–57, Lafayette County, Wis., showing water-level fluctuations caused by the earthquake and aftershocks.

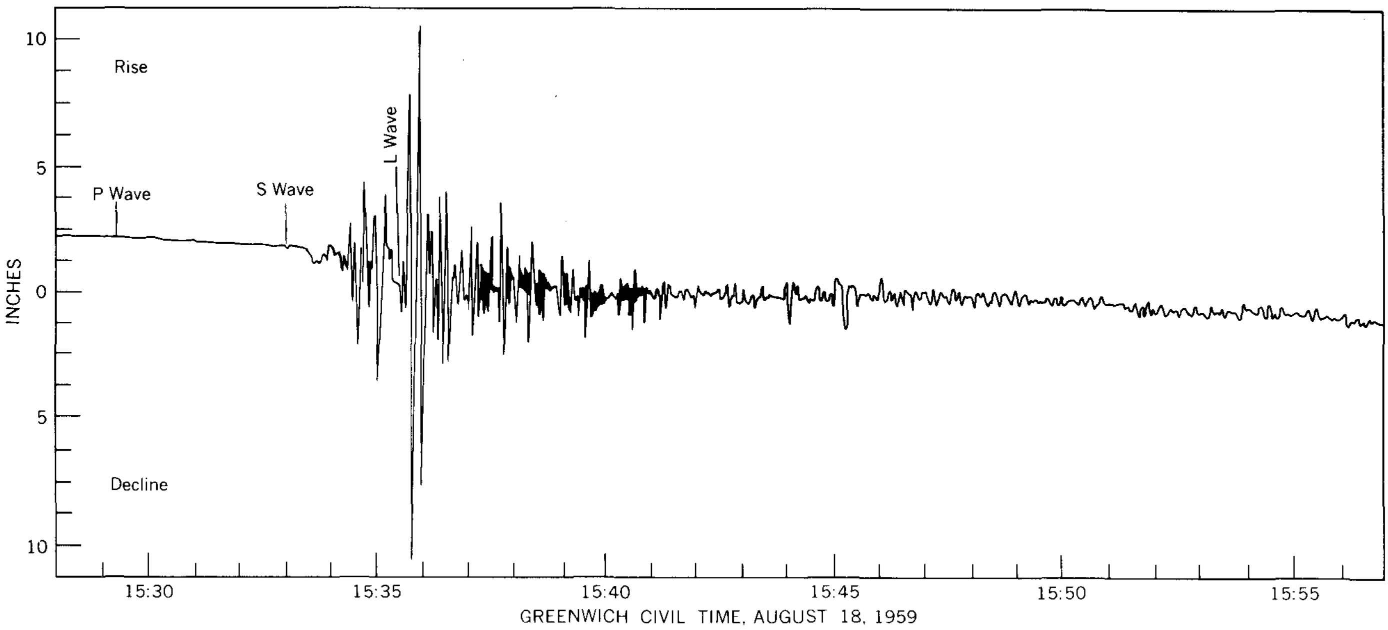

FIGURE 102.—Hydrograph of Nunn-Bush Co. well in Milwaukee County, Wis., showing water-level fluctuations caused by one of the aftershocks of the earthquake. (Hydrograph furnished by E. E. Rexin, and published with his permission.)

to more than 5 feet. The pressure head of well (B–2–1) 24bad-3, which is equipped with a pressure gage, fluctuated 3.8 feet, and that of well (C–16–7) 12dcd-5 fluctuated 5.1 feet.

Washington.—Maximum fluctuations of from 0.03 foot to 1.15 feet were observed in six wells. Well 20/3–18–C1 showed significant fluctuations not only in response to the main earthquake but also to aftershocks 1, 2, 4, and 5.

Wisconsin.—The maximum water-level fluctuations in five observation wells ranged from 0.07 to more than 1.0 foot; the water levels in three of these wells responded to aftershocks. The water level in well Lf–57 fluctuated more than 1 foot in response to the main quake, and slightly in response to at least three of the aftershocks (fig. 101). Water levels in wells Lf–121 and Lf–183 fluctuated in response to the main quake and to two of the aftershocks. All these wells are in sandstone and limestone of Paleozoic age.

The water level in the well of the Nunn-Bush Shoe Co., Milwaukee County, Wis., fluctuated with both the main quake and the aftershocks. The data on the fluctuations in this well were not available for inclusion in table 17; but, through the courtesy of E. E. Rexin, a hydrograph that illustrates the fluctuations in response to one of the aftershocks is shown in figure 102. The main quake caused a fluctuation of more than 10 inches on the chart, and the limits of the chart prevented the recording of the full amplitude. The major fluctuation continued with diminishing amplitude until after 3:00 GCT on August 18. Rexin (written communication, 1959) reported that the arrival times of the P, S, and L waves for the main shocks were $06^h38^m20^s$, $06^h41^m9^s$, and $06^h43^m00^s$ GCT, respectively.

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Records of Postearthquake Chemical Quality of Ground Water

By ROBERT C. SCOTT

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CONTENTS

RECORDS OF POSTEARTHQUAKE CHEMICAL QUALITY OF GROUND WATER

By Robert C. Scott

ABSTRACT

Ground-water data of radioelement and other chemical components were obtained for immediate and future identification of geochemical changes that may be attributed to the Montana earthquake.

Turbidity caused by clay minerals and other weathering byproducts was the most pronounced change in ground water that could be related with certainty to the earthquake. Some springs and geysers contained seemingly anomalous amounts of radioelements as well as some of the more common chemical components. These were tentatively interpreted as effects from the earthquake. Because little preearthquake information is available for the area affected most by the earthquake, it may be several years before much of the data presented in this paper can be adequately interpreted.

INTRODUCTION

Many physical changes in the ground-water regimen were reported after the earthquake by rangers of Yellowstone National Park and others living in the area. Samples of water were therefore collected from the more readily accessible thermal basins in the park and from a few nonthermal sources in the earthquake area (fig. 103) for the purpose of determining any significant chemical changes. Some of the chemical characteristics are believed to be directly related to the earthquake; other chemical characteristics are more difficult to interpret because preearthquake data are not available or are so old that they cannot be compared reliably with modern analyses. The information in this report may be useful in evaluating some of the more obscure effects of the earthquake when new analyses are obtained after the ground-water regimen has returned to normal. It may be several years, however, before earthquake produced suspended material and other chemical constituents are flushed from the system and there is a return to normal dynamic equilibrium.

Samples 3439 through 3442 were collected shortly after the earthquake by Survey personnel. These samples were collected from three springs that became turbid immediately after the earthquake. The samples were separated into two fractions—water and suspended material. Each fraction was analyzed for as many chemical and physical characteristics as the amount of sample would permit.

SAMPLING METHODS

Nineteen samples of ground water were collected from thermal and nonthermal springs, geysers, and wells in the earthquake area. Their location, geologic environment, and other data are given in table 19. Samples for radioelement analysis were collected in separate containers and treated immediately after collection with about 8 ml (milliliters) of acetic acid and 2 ml of chloroform to prevent loss of the radioelements from the solution. Samples 1466 and 2050, collected in 1956 and 1957, respectively, are the only preearthquake samples available that were collected and analyzed by the same methods as those used for the postearthquake samples.

CHEMICAL CHARACTERISTICS OF THERMAL WATER

Physical changes were reported by park rangers to be more noticeable in the thermal water of the Midway, Lower, and Upper Geyser Basins than in other areas; therefore, most samples were collected in these basins. Samples were mainly from springs or geysers that showed radical changes in color or turbidity of the water, temperature, or volume of flow.

LOWER GEYSER BASIN

Preearthquake analyses of Black Warrior Spring and Leather Pool are not available, but postearthquake analyses of samples from these two sites can be compared with analyses of water from other sites in Lower Geyser Basin (Allen and Day, 1935, p. 277, 307; Clarke, 1914, p. 20). Black Warrior Spring (table 20, analysis 3411) contained less sodium and lithium, and Leather Pool (table 20, analysis 3419) contained more sulfate and less fluoride than any of

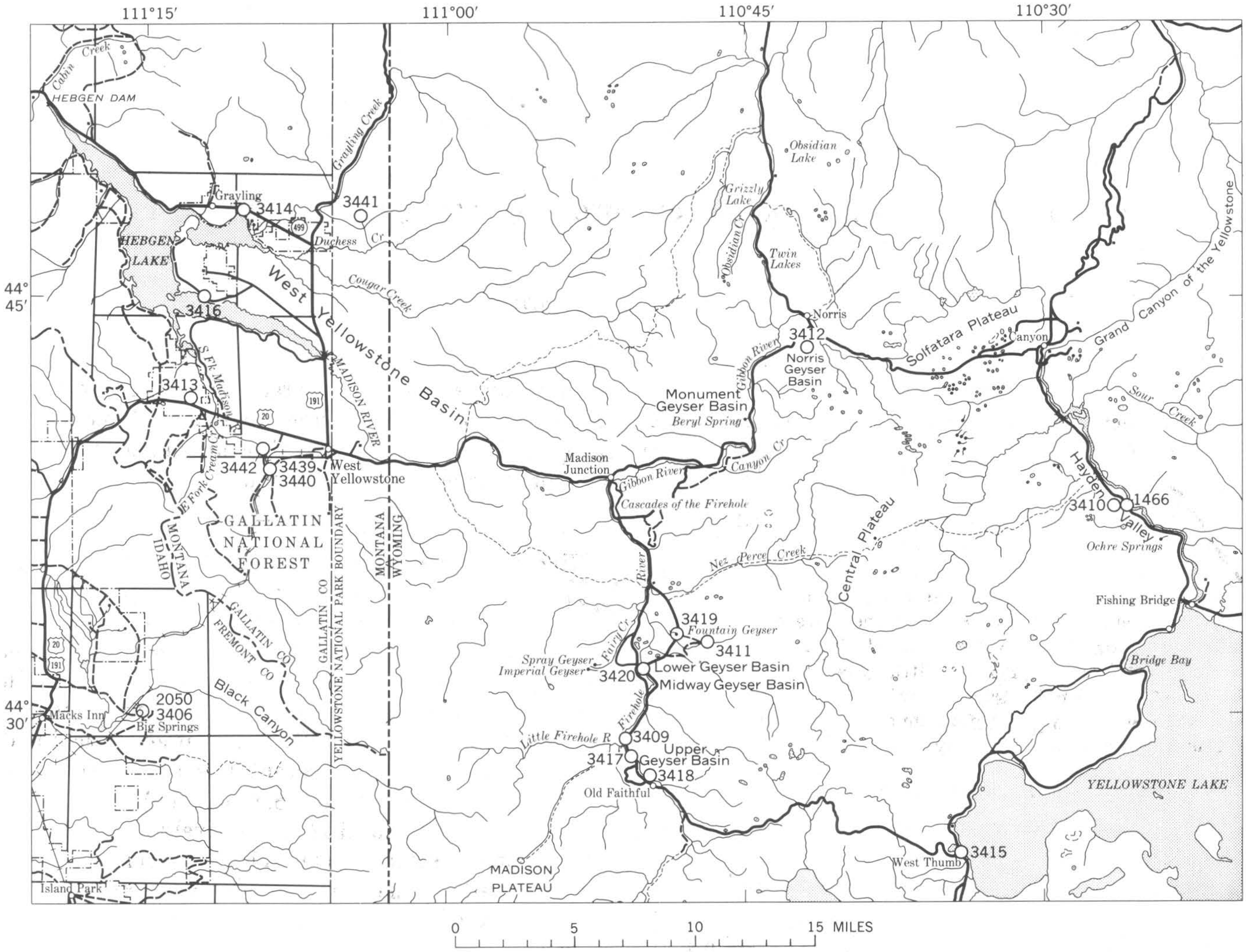

FIGURE 103.—Location of water-sampling sites.

TABLE 19.—*Field data for water samples*

Sample	Location	Geologic terrane	Temperature (°F)	Yield (gpm)	Appearance	Date collected
1466	Spring below Dragons Mouth Spring, Mud Volcano area, Hayden Valley, Yellowstone National Park, Wyo.	Silicic volcanic rocks	119	[1] 15	Turbid	July 10, 1956.
2050	Big Springs, NE¼ sec. 27, T. 14 N., R. 44 E., Fremont County, Idaho (Stream comprising flow from all springs in group).	do	52	83, 500	Clear	July 23, 1957.
3406	Big Springs, NE¼ sec. 27, T. 14 N., R. 44 E., Fremont County, Idaho (Largest spring in northeast group of springs).	do	56	[1] 500	Slightly turbid	Sept. 4, 1959.
3409	Sapphire Pool, Biscuit Basin, Upper Geyser Basin, Yellowstone National Park, Wyo.	do			Turbid	Sept. 5, 1959.
3410	Churning Cauldron, Mud Volcano area, Hayden Valley, Yellowstone National Park, Wyo.	do	91		Colored	Sept. 5, 1959.
3411	Black Warrior Spring, Lower Geyser Basin, Yellowstone National Park, Wyo.	do	149		Clear	Sept. 1, 1959.
3412	Echinus Geyser, Norris Geyser Basin, Yellowstone National Park, Wyo.	do	180		Slightly turbid	Sept. 3, 1959.
3413	Flowing artesian well, 280 feet deep, Deep Well Ranch, NW¼ sec. 23, T. 13 S., R. 4 E., Gallatin County, Mont.	Fluviatile sand and gravel	44	895	Turbid	Sept. 4, 1959.
3414	Corey Spring, NW¼ sec. 7, T. 12 S., R. 5 E., Gallatin County, Mont.	Silicic volcanic rocks (?)	49	7, 320	Clear	Sept. 3, 1959.
3415	Abyss Pool, West Thumb area, Yellowstone National Park, Wyo.	Silicic volcanic rocks	178		do	Sept. 5, 1959.
3416	"New" Spring, south side of Horse Butte, NE¼ sec. 35, T. 12 S., R. 4 E., Gallatin County, Mont.	Precambrian metamorphic complex.	54		do	Sept. 4, 1959.
3417	Sprite Pool, Upper Geyser Basin, Yellowstone National Park, Wyo.	Silicic volcanic rocks	162		Turbid	Sept. 1, 1959.
3418	Tortoise Shell Spring, Upper Geyser Basin, Yellowstone National Park, Wyo.	do	203	[1] 15	Clear	Sept. 1, 1959.
3419	Leather Pool, Lower Geyser Basin, Yellowstone National Park, Wyo.	do	168		Turbid	Sept. 5, 1959.
3420	Excelsior Geyser, Midway Geyser Basin, Yellowstone National Park, Wyo.	do	146		Clear	Sept. 5, 1959.
3439	Railroad Spring, NW¼ sec. 5, T. 14 S., R. 5 E., Gallatin County, Mont.	do	47	[1] 1, 500	Turbid	Aug. 23, 1959.
3440	Do	do	47	[1] 1, 500	do	Aug. 28, 1959.
3441	Duchess Spring on Campanula Creek, Yellowstone National Park, Mont.	do		[1] 1, 300	do	Aug. 27, 1959.
3442	Black Spring, NE¼ sec. 31, T. 13 S., R. 5 E., Gallatin County, Mont.	do	47		do	Aug. 28, 1959.

[1] Estimated.

TABLE 20.—*Chemical analyses of water samples*

[Results in parts per million unless otherwise specified]

Chemical characteristics	Sample																		
	1466	2050	3406	3409	[1] 3410	3411	[2] 3412	3413	3414	3415	3416	3417	3418	3419	3420	3439	3440	3441	3442
Silica (SiO_2)	230	46	50	244	225	175	250	21	7.6	197	20	299	259	284	255	38	41	27	39
Aluminum (Al)	.5	.0	[3].1	[3].3	38	.00	[3]1.6	[3].1	.0	.2	.1	.1	[3].1	[3].1	.2	[3].0	[3].0	[3].0	[3].0
Iron (Fe)	[3].14	.03	[3].10	[3].33	55	.00	[3]2.3	[3].00	.64	.00	.14	.00	[3].00	[3].00	.00	[3].06	[3].13	[3].10	[3].04
Manganese (Mn)	----	.00	[3].00	[3].00	3.6	.11	[3].36	[3].00	.00	.00	.00	.50	[3].00	[3].00	.00	[3].00	[3].00	[3].00	[3].00
Calcium (Ca)	79	5.6	5.6	1.6	171	18	5.6	38	50	1.6	21	3.2	1.6	3.2	1.6	6.4	6.4	8.0	6.4
Magnesium (Mg)	28	.5	1.0	.0	69	.0	4.9	6.8	24	.0	14	.0	.0	.0	.5	.5	.9	1.5	.7
Sodium (Na)	65	13	21	420	150	96	164	4.9	5.0	444	65	280	408	268	408	11	11	5.0	13
Potassium (K)	46	3.0	3.6	18	46	12	54	1.0	.6	20	6.0	12	19	16	12	2.0	2.0	1.4	2.0
Zinc (Zn)	----	----	.0	.0	.05	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	----	----	----	----
Arsenic (As)	----	----	.00	2.5	.00	.00	.08	.00	.00	2.0	.02	.60	2.5	.6	2.0	.00	.00	.00	.00
Molybdenum (Mo)	----	----	.003	.055	<.001	.002	.002	.002	.001	.085	.005	.015	.045	.038	.032	----	----	----	----
Lithium (Li)	----	----	.4	2.2	.6	.9	1.2	.6	.3	3.4	.00	2.4	4.8	1.8	2.6	----	----	----	----
Strontium (Sr)	----	----	----	----	1.17	----	.07	----	----	.08	----	.06	.02	.05	.04	----	----	----	----
Copper (Cu)	----	----	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	----	----	----	----
Ammonium (NH_4)	----	----	----	.0	3.2	.5	1.4	----	----	.0	----	.0	.0	.2	.0	----	----	----	----
Bicarbonate (HCO_3)	5	41	56	332	0	155	0	134	188	598	288	269	174	24	527	34	36	40	38
Carbonate (CO_3)	0	0	0	99	0	0	0	0	0	0	0	0	102	0	20	0	0	0	0
Sulfate (SO_4)	494	2.9	6.2	22	1,440	35	310	25	79	55	11	77	27	225	21	2.5	2.5	3.7	3.7
Chloride (Cl)	2.2	2.5	4.0	309	20	53	105	2.0	2.0	298	10	208	382	248	278	4.0	4.0	1.0	4.0
Fluoride (F)	.0	3.2	4.0	18	.6	11	6.0	1.1	.2	18	1.0	18	18	6.0	18	3.6	3.6	.3	3.6
Nitrate (NO_3)	.0	.6	.0	.0	.0	.0	.0	.0	.4	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0
Phosphate (PO_4)	.10	.00	.00	1.1	.00	.00	.35	.00	.00	1.9	.16	1.2	1.3	3.1	2.0	.00	.00	.00	.00
Boron (B)	----	----	.05	2.6	.34	.42	1.0	.02	.02	3.0	.12	1.6	2.2	2.4	2.5	.05	.07	.02	.06
Bromide (Br)	----	----	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	----	----	----	----
Iodide (I)	----	----	.5	.5	.5	.0	.5	.0	.5	.0	.0	.0	.0	.0	.0	----	----	----	----
Suspended solids	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----	524	359	3,454	480
Particle size (percent <0.001 mm)	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----	99	100	100	100
Dissolved solids (residue on evaporation at 180° C)	990	91	120	1,240	2,240	464	935	158	261	1,260	278	1,040	1,220	1,110	1,240	89	94	76	93
Specific conductance (μmhos at 25° C)	1,020	92	127	1,940	2,520	527	1,110	245	418	2,070	475	1,340	1,970	1,440	1,870	84	84	70	89
pH	4.6	6.9	6.8	9.2	3.0	7.4	3.5	7.5	7.7	8.2	7.7	7.3	9.2	5.8	8.4	7.3	7.3	7.3	7.3
Uranium (U) (μg/l)	<.1	.6	.9	.5	3.6	.5	2.3	1.6	1.0	.2	.6	1.1	<.1	41	.6	1.1	.9	.3	.9
Radium (Ra) (μμc/l)	.1	<.1	.1	.2	.1	.3	3.3	.3	.1	<.1	2.2	.4	<.1	1.3	.1	----	----	----	----

[1] Hydrogen=55 ppm; calculated acidity (as H_2SO_4) from pH readings adjusted to pH 4.5=0.
[2] Hydrogen=16 ppm; calculated acidity (as H_2SO_4) from pH readings adjusted to pH 4.5=0.
[3] In solution at time of analysis.

the published analyses. Other chemical components common to the earlier analyses are within the ranges of the maximum and minimum values reported earlier. The differences noted above, however, cannot be attributed with certainty to changes effected by the earthquake. At any given time, great differences are found in a few chemical components in many springs and geysers in any one basin, and also any individual thermal spring or geyser having a history of erratic flow can be expected to show chemical changes with time.

UPPER GEYSER BASIN

Samples were collected from Sprite Pool (table 20, analysis 3417), Tortoise Shell Spring (table 20, analysis 3418), and Sapphire Pool (table 20, analysis 3409). Analyses of water from the last two sites were published by Allen and Day (1935, p. 249), and many of the chemical constituents are comparable to those of current analyses. The agreement between the concentrations of most components is remarkably close in both springs. A slight increase in iron, aluminum, and sulfate ions and a small decrease in alkalinity (HCO_3^{-1} and CO_3^{-2}) were noted and may have been produced by the earthquake. The water in a slow-circulating system such as a restricted channel may show the effects of longer contact time with metallic sulfides. Subterranean fracturing of rocks enclosing a system not only could increase the circulation rate but also could bring water into contact with unaltered sulfide minerals. Changes in the circulation pattern might also introduce acid water or acid-producing gases which would be capable of dissolving large amounts of metals and bringing about the observed changes.

A comparison of the analysis of Sprite Pool (table 20, analysis 3417) with the analyses of water from other sites in the area published by Allen and Day (1935, p. 249) and Clarke (1914, p. 21 and 22) shows that each component falls within the same range in each analysis.

MIDWAY GEYSER BASIN

The analysis of a sample from Excelsior Geyser (table 20, analysis 3420) is similar to that reported by Allen and Day (1935, p. 268), but both differ in many respects from an analysis of water from the same geyser, made by Gooch and reported by Clarke (1914, p. 20). The results of the present analysis and the one reported by Allen and Day (1935) agree well enough to suggest no change since the earthquake.

OTHER THERMAL LOCALITIES

Samples were collected from a single locality each in the West Thumb area (table 20, analysis 3415), the Mud Volcano area (table 20, analysis 3410), and the Norris Geyser Basin area (table 20, analysis 3412). Park rangers reported little or no physical change in these areas after the earthquake. Analyses of water from the same sites were not reported by Allen and Day (1935), and the chemical composition of the sample from Norris Geyser Basin (Echinus Geyser) was reported only by Clarke (1914, p. 19).

The analyses in table 20 of samples from these three sites are in reasonably close agreement with analyses reported by Allen and Day (1935, p. 328, 419, and 469) for nearby sites in each of the areas. Sample 1466 (table 20), collected before the earthquake from a small spring in the Mud Volcano area, is not directly comparable to the postearthquake sample 3415 (table 20) from Abyss Pool, and neither is typical of the area with respect to chemical character.

Although the analysis of water from Echinus Geyser, in the Norris Geyser Basin, seemingly is different from the analysis reported by Clarke (1914, p. 19), it is not unusual for analyses of samples collected approximately 50 years apart to differ to some extent. The relatively high concentrations of iron and aluminum found in the recent analysis of Eichinus Geyser is commensurate with the amounts of oxides of these metals found in sinter in the pool in the vent of the geyser (Allen and Day, 1935, p. 485).

RADIOELEMENT CHARACTERISTICS OF THERMAL WATER

Uranium and radium concentrations in thermal water are commonly small. The radioelement concentrations in the samples from Leather Pool (table 20, analysis 3419) and Echinus Geyser (table 20, analysis 3412) therefore may be considered anomalous. At first glance, the uranium content of the water from Churning Cauldron (table 20, analysis 3410) also seems anomalous, but the uranium is not as abundant in proportion to the dissolved mineral content as it is in some of the more dilute water.

The radioelement content of Echinus Geyser is anomalous not only because the radium content is large in proportion to the dissolved solids in the water and is large compared to the radium content of the other samples, but also because the radium is present in amounts greater by a factor of about 4 than would be found if it were in radioactive equilibrium with the uranium present. Radium generally occurs in water of the bicarbonate or sulfate type in a concentration below the equilibrium value determined by the uranium present. This is because uranium forms stable, soluble anionic complexes with these radicals. The radium, which is present only as a cation, commonly is removed from solution by ion exchange with clays or other minerals having ion-exchange capacity. The water in Echinus Geyser was turbid when collected, and the radium may have been brought to the surface in the suspended material. The earthquake could easily have dislodged this suspended material without increasing the uranium concentration in the water. The radium would be leached from the suspended material during the time the sample was in transit and in storage. It is also possible that the acetic acid and chloroform added at the time the sample was collected aided the leaching process.

An additional factor that was considered, but not tested, is that perhaps not all the radium is a daughter product of uranium. Radium–224, a member of the thorium series, also is included in the analytical method used. The amounts of each radium isotope can be determined by more laborious analytical methods, and such work may be necessary before the radium anomaly can be more fully explained.

The uranium content of the water from Leather Pool is anomalous not only when compared with that of the other samples in Yellowstone Park but also when compared with that shown in analyses of thermal water from other areas. The amount of radium present in Leather Pool is about that expected for this type of water and geologic terrane, but the amount of uranium is greater by approximately a factor of 20 than that commonly found in most thermal water. The uranium also exceeds the anomaly threshold for nonthermal fresh water in this geotectonic region (Scott and Barker, 1958).

The large amount of uranium present in the water of Leather Pool lends support to the conclusion that the ground-water environment beneath Leather Pool probably was altered by the earthquake. It is very possible that the earthquake fractured some of the rocks beneath this spring and exposed fresh uraniferous minerals to circulating ground water. Careful systematic monitoring of Leather Pool for several years should show whether this interpretation is tenable. It may be that the geologic terrane in this area is inherently rich in uraniferous minerals. If so, ground water circulating in the area should contain relatively constant amounts of uranium over a long period of time. Recent exposure of fresh uraniferous minerals in a few new fractures should not supply large amounts of uranium to water for very long.

A comprehensive study of the radioactivity of thermal water in Yellowstone National Park was made by Schlundt and Moore (1909). They measured only the radioactivity of radon (radium emanation) present in the water, however, and the values they reported for uranium and radium are those that would be expected if the radon were in radioactive equilibrium with the uranium and radium. Radium and uranium seldom, if ever, occur in water in amounts that are in equilibrium with respect to radon; thus the values reported by Schlundt and Moore (1909) bear no relation to the amounts of radium and uranium actually present. For this reason, the analysis of water samples for radioactivity reported by Schlundt and Moore (1909, p. 23, 24, 27, and 28) cannot be compared with the radioelement determinations in this report.

CHEMICAL CHARACTERISTICS OF NONTHERMAL WATER

Samples were collected from six nonthermal springs and one well east and southeast of Hebgen Lake. Except for Big Springs, Idaho, no analyses of preearthquake samples are known. People living in the area reported that all the springs and the well discharged clear water before the earthquake. At the time the samples were collected, the flow from each was still turbid. Big Springs, however, which is most remote from the earthquake area, and the well at Deep Well Ranch, drilled in sand and gravel, had become almost clear by the time the samples were collected.

Effects of the earthquake are evident in the dissolved constituents of sample 3406 (table 20), collected at Big Springs. Compared with the analysis of the sample collected in 1957 (table 20, analysis 2050), the dissolved-solids content and all chemical constituents except nitrate showed an increase. Except for the increase in nitrate, which is not significant, the chemical changes as well as the turbidity can be attributed to the earthquake.

Interpretation of the chemical changes in other water must await analyses of samples collected later, when the earthquake effects have decreased.

CHARACTERISTICS OF THE SUSPENDED SOLIDS IN GROUND WATER

The three springs that became most turbid after the earthquake were sampled (table 19, samples 3439, 3440, 3441, and 3442), and the suspended material was separated from the water with an ultracentrifuge. The water fractions were analyzed for all the chemical constituents (table 20) for which methods were available. The suspended-material fraction of each sample was used for particle-size determination, clay-mineral analysis, and semiquantitative spectrographic analysis (table 21).

TABLE 21.—*Semiquantitative spectrographic analyses, in weight percent*

[Analysts for: suspended material for water samples, R. G. Havens; rhyolite samples, J. C. Hamilton. Figures are reported to the nearest number in the series 7, 3, 1.5, 0.7, 0.3, 0.15, and so on. M, major constituent greater than 10 percent; d, detected but not measurable. Also looked for but not found in amounts greater than standard detectabilities, P, As, Au, B, Bi, Cd, Eu, Ge, Hf, Hg, Ho, In, Ir, Li, Lu, Os, Pd, Pr, Pt, Re, Rh, Ru, Sb, Sm, Ta, Tb, Te, Th, Tl, Tm, U, and W]

Elements	Suspended material in indicated water samples				Rhyolite samples	
	3439	3440	3441	3442	227	228
Silicon (Si)	M	M	M	M	M	M
Aluminum (Al)	M	M	M	M	7	3
Iron (Fe)	3	3	3	3	.7	.3
Magnesium (Mg)	1.5	1.5	1.5	1.5	.03	.015
Calcium (Ca)	.7	.7	.7	.7	.15	.07
Sodium (Na)	.7	.7	.3	.7	3	1.5
Potassium (K)	3	3	1.5	1.5	7	3
Titanium (Ti)	.15	.15	.3	.15	.07	.03
Manganese (Mn)	.15	.15	.03	.07	.015	.003
Silver (Ag)	d	d	d	d	0	0
Barium (Ba)	.03	.03	.07	.03	.015	.015
Beryllium (Be)	.0007	.0007	.0007	.0007	.0003	.0003
Cerium (Ce)	.07	.07	.07	.07	d	0
Cobalt (Co)	.0007	.0007	.0007	.0007	0	0
Chromium (Cr)	.007	.007	.007	.007	0	0
Copper (Cu)	.03	.03	.015	.015	.0003	.00015
Dysprosium (Dy)	d	d	0	d	0	0
Erbium (Er)	d	d	0	d	0	0
Gallium (Ga)	.0007	.0007	.0007	.0007	.0015	.0007
Gadolinium (Gd)	d	d	0	d	0	0
Lathanum (La)	.03	.07	.03	.03	.007	.003
Molybdenum (Mo)	[1] 0	0	0	0	.0007	.0007
Niobium (Nb)	.0015	.0015	.0015	.0015	.003	.003
Neodymium (Nd)	.07	.07	.03	.03	d	0
Nickel (Ni)	.003	.003	.003	.003	0	0
Lead (Pb)	.007	.007	.003	.007	.003	.0015
Scandium (Sc)	.0015	.0015	.0015	.0015	0	0
Tin (Sn)	.003	.003	.0015	.0015	0	0
Strontium (Sr)	.003	.007	.007	.003	.0007	d
Vanadium (V)	.007	.007	.007	.007	0	0
Yttrium (Y)	.07	.07	.015	.07	.003	.003
Ytterbium (Yb)	.003	.003	.0007	.003	.0003	.0003
Zinc (Zn)	.015	.015	0	.03	0	0
Zirconium (Zr)	.015	.015	.015	.015	.015	.015

[1] Looked for but not found.

One spring, Railroad Spring, was sampled twice (samples 3439 and 3440), 5 days apart. No significant changes were detected in either the dissolved mineral content or the chemical character of the suspended material during this 5-day period.

The mineral composition of each sample of suspended material was determined by X-ray diffraction. No detectable qualitative or quantitative differences in mineral composition were found among the samples; the following list applies to all samples:

Minerals present	*Estimated amount (parts in 10)*
Montmorillonite	6
Mica	1
Kaolinite(?), chlorite(?)	1(?)
Feldspar	1
Quartz	Trace
Cristobalite	Trace(?)

Because the estimated amounts are derived from relative intensities of the diffraction lines, which are affected by many factors in addition to the quantity of a mineral, these estimates are not intended to give more than a very general indication of the relative amounts of the various minerals present.

The semiquantitative spectrographic analyses show that the trace-element content of the suspended material is more like that of ordinary granitic pegmatite than it is of rhyolite. Comparison of the spectrographic analyses of the suspended material (table 21, analyses 3439, 3440, 3441, and 3442), with spectrographic analyses of rhyolite country rock collected near Duchess Spring (table 21, analysis 227) and near Railroad Spring (table 21, analysis 228) indicates that the suspended material has a concentration of rare-earth and other elements in the proportional amounts commonly found in pegmatite. (Note especially: silver, cerium, copper, lanthanum, neodymium, scandium, tin, and zinc.) This enrichment of trace elements in the suspended material is the result of geochemical processes in the ground-water regimen.

The rhyolite terrane from which the water samples were collected has conspicuous columnar jointing. Ground water passes through these joints and discharges through springs and seeps. During the time ground water is in contact with the rhyolite, chemical erosion of the joint faces occurs; feldspar and other silicate minerals are altered, and soluble compounds of the minor elements are removed. The relatively insoluble clay minerals that are formed remain on the joint faces and surfaces of cavities. Many trace elements are easily adsorbed by clay minerals and, once adsorbed, are strongly held in the exchange positions. Thus the clay is enriched with rare earths and other minor elements previously leached from the rhyolite.

Apparently the "rocking" action of the earthquake rubbed the joint faces together and loosened some of the clay. The particles became suspended in the water, producing turbidity in the springs, and provided the anomalous amounts of minor elements found in the residue. Because of the large areal extent of the rhyolite terrane and the ground-water reservoir, it is probable that many of the springs will yield turbid water for months or even years.

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Effects of the Hebgen Lake Earthquake of August 17, 1959 on the Hot Springs of the Firehole Geyser Basins Yellowstone National Park

By GEORGE D. MARLER

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

EFFECTS OF THE HEBGEN LAKE EARTHQUAKE OF AUGUST 17, 1959, ON THE HOT SPRINGS OF THE FIREHOLE GEYSER BASINS, YELLOWSTONE NATIONAL PARK

By George D. Marler [8]

ABSTRACT

The 1959 Hebgen Lake earthquake resulted in great changes in the functional behavior of most of the thermal springs along the Firehole and Gibbon Rivers in the Yellowstone National Park. The effects were particularly marked along the Firehole drainage. This is the region of Yellowstone's most famous geysers and hot springs. The initial tremors served as an impetus to bring about eruptive activity in scores of geysers and quiescent springs. Many long-dormant geysers erupted; a few new geysers were born. Some hot springs seemed to be affected adversely, but on the whole there was a noticeable increase in thermal energy that, with the passing months, gave evidence of persisting. This paper gives a résumé of the immediate effects of the earthquake upon the springs in the Firehole geyser basins; and then, by means of tables, presents data which show preearthquake and postearthquake conditions of the better known hot springs. These data deal with functional behavior, temperature, discharge, and general characteristics.

INTRODUCTION

Many earthquakes have been recorded in Yellowstone National Park since its establishment in 1872. Until August 17, 1959, none was of sufficient intensity to produce observable changes in any hot spring, and this despite the fact that many important changes in the hot springs have taken place since their discovery.

The Hebgen Lake earthquake, with a magnitude of 7.1, resulted in severe jarrings all over the park. It was only in western sections, however, that severe damage resulted to man-made installations. In this area there were great rock slides from numerous escarpments with resultant timber destruction and road blockage; also, there were great changes in the behavior of hot springs. These changes were particularly pronounced in the geyser basins along the Firehole River 25 miles southeast of the epicentral area at Hebgen Lake. There was also extensive changes in a number of hot springs on the Gibbon River drainage in the northwestern part of the park. Hot springs were affected to a lesser degree in eastern sections.

[8] U.S. National Park Service.

Because the author is particularly familiar with the thermal areas in the Firehole geyser basins and because preearthquake records of thermal activity are far more abundant for this area than for others in the park, this paper deals mainly with the Firehole basins, the region of Yellowstone's most famous geysers and hot springs.

FIRST IMPRESSIONS

As soon as daylight of August 18 permitted, a hurried reconnaissance was made of the geyser basins. Most of the springs visited on that and succeeding days showed evidence of functional alteration. One of the noticeable changes was the marked increase in discharge of hot water and steam. As the reconnaissance proceeded, the evidence of a general increase in temperature and activity became even more apparent. Scores of hitherto quiescent springs with no previous record of geyser activity were either boiling or showed clear evidence of having erupted. Large fragments of sinter scattered around the craters of some springs indicated a major increase of activity and forceful eruption.

The earthquake not only stimulated geyser activity in many formerly quiescent springs, but most of the geysers with established patterns of play erupted at shorter intervals. Only a few of the well-known geysers seemed to be affected adversely in that their activity decreased. During the first few days a number of springs seemed to have escaped any change in activity or temperature, but time and additional observation disclosed that the earthquake had affected in varying degree practically every spring in the main basins.

Cracks and breaks in surface deposits showed in many places in all three basins, more particularly in the Lower Geyser Basin. Some of these cracks had fumarolic action. In the Firehole Lake area of the Lower Geyser Basin the cracks were crisscrossed with numerous minor displacements but with no apparent system of alinement. Careful measurements indicated that the total length of the cracks near Firehole Lake was 9,072 feet, or 1.72 miles. Cracks extended through the centers of craters of a few springs.

One of the unusual conditions produced by the earthquake was murkiness of the formerly limpid water of hundreds of springs. The Lower Basin had a larger proportion of murky springs than the Upper Basin, and the water in many of the geysers was tinted by fine sediment.

Not only did the Lower Basin, which is nearer the Hebgen Lake area than the Upper Basin, show an increased number of turbid springs, but also the water levels in a large proportion of the springs had fallen. The great majority of springs whose levels were a few inches to a few feet below previous normal states showed distinct evidence that the ebbing had been preceded by surging and discharge. Throughout the basins there was evidence that the earthquake had acted like a giant hand which suddenly applied enormous pressure to the rocks beneath the hot springs, forcing water from their conduits in a manner comparable to the squeezing of a sponge.

Conditions around the hot springs indicated that a great increase in discharge took place during or immediately after the initial big tremor. Its jarring served as a trigger to start discharge from hundreds of springs. Had this happened in the daytime, the spectator would have witnessed geyser activity on a scale never even closely approximated since Yellowstone's discovery. Even so, during the days following August 17 a spectacle without precedent was observed. For example, in the Fountain Group in the Lower Basin, major geysers with no previous record of concerted action erupted simultaneously and for sustained periods never previously known.

OBJECT OF PAPER

The following discussion shows statistically the nature and degree of change in the behavior of most of the important and better known hot springs in the Firehole area. The 173 springs considered constitute but a minor fraction of the thermal units in this area, but they are the only ones for which comparable pre-earthquake data are available. The general state of some other springs, as they were observed immediately following the earthquake, is also tabulated.

During the past 20 seasons as a naturalist in Yellowstone National Park, the author has had the opportunity to become acquainted in considerable detail with the thermal springs in the Firehole geyser basins. The data gained during this period are here compared with the information secured during a 3-month period of study following the earthquake. To facilitate the postearthquake study, the National Park Service set aside funds and created a project to help catalog changes induced by the August 17 earthquake. This project was given the somewhat formidable title "Emergency Interpretive Study of Earthquake Phenomena, Yellowstone National Park." The first phase of the study was completed late in December 1959.

RESULTS OF INITIAL SURVEY

As indicated above, a cursory inspection of all groups of springs was made during the first three days following the earthquake. The data sought in this preview were changes in the state of water and in eruptive activity. A summary of the state of springs and geysers observed at that time and their location as to group and basin is given in table 22. The general location of the basins and thermal groups as well as some of the better known thermal units is shown in plate 4.

RESULTS OF SURVEY OF AUGUST TO DECEMBER 1959

Between the time of the preliminary survey and the end of December, practically every spring listed as being normal (table 22) underwent changes that seem related directly or indirectly to the earthquake. If the postearthquake observations have revealed any one fact, it is that many earthquake-induced changes are progressive. Few days passed without new fumaroles being developed and previously quiescent springs becoming eruptive.

The effects of the earthquake on thermal water in the park are shown in tables 23 and 24. Table 25 summarizes data for some of the springs, pools, and geysers not affected by the earthquake. The classification of geysers, springs, and pools is to a considerable extent arbitrary and is used for convenience in presentation. It is not based on any natural or special distinction, rather upon long-accepted local designations. During almost any year some named spring or pool may erupt and be given the status of a geyser, and many so-called springs and pools are true geysers. All these thermal units are technically hot springs; a geyser is a special type of hot spring that erupts periodically.

Changes in many of the geysers and hot springs in the Firehole River drainage are summarized in tables 22–24. Temperature and discharge increased in most hot springs, but declined in a few. The marked rise in the temperature of springs, pools, and geysers, as determined 3 to 4 weeks after the earthquake, indicates a great and sustained increase in discharge of thermal energy. The average temperature of 73 geysers after

TABLE 22.—*State of activity of hot springs and geysers of the Firehole drainage area immediately after the Hebgen Lake earthquake of Aug. 17, 1959*

Group	Number that erupted	Number with no previous record of eruption	Murky	Clear	Ebbed	Flowing	Normal	More active
Lower Geyser Basin								
Fairy Meadows	11	9	31	16	45	7	0	16
Hotel	5	3	11	0	3	5	1	9
Quagmire	3	2	26	3	28	0	3	5
Kaleidoscope	29	22	72	16	18	12	0	31
Pithole Springs	5	1	17	0	5	10	1	7
Fountain	17	2	14	13	2	12	0	19
Firehole Lake	12	10	28	2	17	1	0	3
Pink Cone	8	2	5	12	4	0	1	11
Great Fountain	21	10	26	12	32	6	0	22
Totals	111	61	230	74	154	53	6	123
Midway Geyser Basin								
Excelsior	0	0	5	0	3	2	0	0
Flood	13	1	25	3	11	16	1	16
Rabbit Creek	2	2	107	4	35	4	14	7
Totals	15	3	137	7	49	22	15	23
Upper Geyser Basin								
Sapphire	24	17	18	7	14	12	0	22
Cascade	35	23	34	5	5	19	3	28
Morning Glory	3	3	12	7	3	17	3	7
Grotto	8	0	9	8	5	5	4	4
Giant	2	0	3	10	1	3	4	6
Round Springs	3	1	7	2	2	1	3	2
Daisy	6	1	1	17	6	3	1	10
Black Sand Basin	10	2	4	16	2	11	1	7
Grand	13	3	7	41	12	7	11	9
Castle	13	2	5	13	2	4	2	13
Geyser Hill	28	20	3	47	15	18	0	26
Myriad	27	24	125	5	93	13	4	54
Totals	172	96	228	178	160	113	36	188
Grand totals	298	160	595	259	363	188	57	334

the earthquake was 195.1°F, or 2.3°F higher than in the summers of 1958 and 1959. The average temperature of 89 hot springs and pools after the earthquake was 185.6°F, or 8.7°F higher than in 1958–59. Activity strongly increased among the nearly 300 springs that erupted during or shortly after the earthquake; of these, 160 had no previous record of eruption.

Owing to the difficulty of determining the discharge of many springs and geysers, the effects of the earthquake on discharge are less easily determined than the effects on temperature. It is evident, however, that the overall discharge after the earthquake was much greater.

The tables show that the earthquake caused a marked increase in discharge of thermal energy in the named springs; no comparable data are available for the hundreds of unnamed springs. Table 22 lists the impressive number of quiescent springs that erupted immediately after the earthquake. Owing to a lack of earlier data, few of these are included in the other tables. In many groups of springs, particularly the Sapphire, Fountain, Pithole Springs, Kaleidoscope, and Sprinkler groups, fractures in the sinter have permitted the development of a steadily increasing number of fumaroles. If it were possible, therefore, to tabulate the total increase in discharge of thermal energy, it would be more than the totals shown on tables 23 and 24 for the named springs. Some geysers, such as Trail, Cascade, and Economic, erupted only periodically for a score of days after August 17; but

TABLE 23.—*Summary of characteristics of geysers affected by Hebgen Lake earthquake*

[U, undetermined; NM, not readily measurable; PF, periodic flow]

Name	Observations prior to Aug. 17, 1959						Observations after Aug. 17, 1959					
	Eruption frequency	Discharge 1959 (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1951	Temperature (° F) Summers 1958–59	Characteristics	Date	Eruption frequency	Discharge (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1959	Characteristics
UPPER GEYSER BASIN												
Daisy Group:												
Daisy	128 min	[1] 1,250	75–80	196	196	Plays at angle; has preliminary overflow.	Aug.–Dec.	58 min	[1] 1,250	75–80	200	Earthquake doubled frequency of activity; no preliminary overflow.
Comet	4 min	0	4–5	200	200	Water of eruption falls into crater.	do	Constant	0	4–5	202	Water level ebbed 12 in.
Splendid	Infrequent	[1] 4,160	125	200	200	Series of eruptions when in active phase.	do	One eruption	[1] 4,160	125	202	Played following quake, increased boiling since.
Daisy's Thief	do	[1] 80	20	199	199	When active, eruption precedes and stops Daisy.	do	None	0	0	200	12-in. ebb in water level.
Morning Glory Group:												
Fan	do	8	100–125	199	199	Several vents in cracked sinter.	do	do	10	0	202	Increased boiling and temperature.
Mortar	do	U	30–40	199	199	Plays in concert with Fan.	do	do	0	0	200	Water in crater level higher; hotter.
Sentinel	Dormant	25	0	200	200	Constant boiling	do	do	25	0	202	Erupted after quake.
Grotto Group:												
Riverside	7 hrs 32 min	U	75–80	200	200	Plays at angle, preliminary overflow.	do	6 hrs 28 min	U	75–80	200	5- to 6-hr intervals first few days following earthquake.
Link	Dormant	0	0	202	157	Had ebbed below overflow.	do	None	35	0	163	Murky; increased temperature and discharge.
Culvert	Steady	0	2	198	200	Steady boiling from road shoulder.	do	Steady	0	3	202	Increased temperature and vigor of boiling.
Grotto and Rocket (function as single unit).	8 hrs	[1] 470	10–50	200	200	Active about half of the time; longest eruption 12 hrs.	do	4–10 hrs	[1] 470	10–50	201	Active at times for more than 30 hrs.
Grotto Fountain	8–24 hrs	[1] 430	50	200	200	Precedes Grotto's eruptions.	do	Occasional	[1] 430	50	198	Rejuvenated, 8 to 24 hrs, in October.
Spa	8 hrs	[1] 45	1-ft surge	182	183	Action follows Grotto	do	4–10 hrs	[1] 45	1	183	Murky following earthquake.
Giant Group:												
Giant	Dormant cycle.	0	0	203	202	Periodic cycles	do	None	0	0	203	More sloshing in crater since earthquake.
Catfish	Dormant	0	0	200	198	Active and dormant cycles.	do	Dormant	0	0	202	More vigorous boiling.
Mastiff	do	0	0	204	203	Active when Giant is hot.	do	do	0	0	205	Increased boiling.
Turtle	do	0	0	199	200	Active during active cycle of Giant.	do	do	U	0	200	Periodic overflow following earthquake.
Oblong	6–8 hrs	U	20	197	198	Voluminous discharge	do	3–5 hrs	NM	20	200	Murky; more frequent activity.
Black Sand Basin:												
Whistle	Dormant	.04	0	199	143	When active has long and violent steam phase.	do	Dormant	.05	0	144	Murky first week after earthquake.
Cliff	Infrequent	NM	25–30	198	197	Three eruptions in 1959.	do	Active daily	NM	25–30	200	Erupts 3- to 10-hr periods daily.
Spouter	Near constant	80	3–4	197	198	Erupts most of the time.	do	Nearly constant.	80	3–4	200	Increase in temperature; murky Aug. 18.
Castle Group:												
Castle	14–17 hrs	U	75–100	200	200	Steam phase eruptions	do	4–9 hrs	U	75–100	201	Eruptions about twice as frequent.
Sprinkler	80–100 hrs	do	3	194	194	Splashing eruptions	do	56–60 hrs	U	3	198	Increased frequency.
Deleted Teakettle	Dormant	0	0	200	200	Rarely erupts	do	Dormant	6.5	25	199	Erupted following earthquake; flowing.
Churn	do	0	0	151	155	do	do	Occasional	[1] 3	Boiling eruption	161	Did not rejuvenate until November.
Sawmill	2–3 hrs	[1] 190	20–40	167	189	Plays at about time of overflow.	do	1–3 hrs	[1] 190	20–40	194	Erupted almost continuously first 4 days following earthquake.
Tardy	do	[1] 25	6–10	194	194	Connected with Sawmill.	do	do	[1] 25	6–10	196	Increased activity following earthquake.
Spasmodic	U	[1] 45	1–2	199	199	Boiling-type activity	do	2–4 hrs	[1] 45	6–10	203	Boiling more vigorous.

Grand Group:												
Old Tardy	Infrequent	[1] 35	6	201	200	Jet-type eruption	do	1 or more times daily.	[1] 35	6	202	Increased activity since earthquake.
Bulger	Daily	[1] 30	6–8	198	199	do	do	do	[1] 30	6–8	202	More frequent activity.
Triplet	8–12 times daily.	[1] 25	2–3	184	185	Boiling-type eruption	do	Dormant	6 PF	0	182	Dormant since earthquake.
Grand	3 times daily	60 (quiet phase)	150–180	169	176	Rocket-type eruption	do	do	70	0	165	One eruption following quake; increased flow.
Turban	15–25 min	do	3–6	196	197	Boiling-type eruption	do	do	25	1	187	Partial rejuvenation since September.
Economic	Dormant	0	0	144	154	Jet-type eruption	Aug.–Sept. 15	5–20 min	7.5	6–10	200	Dormant since Sept. 15.
Geyser Hill Group:												
Infant	Only with Giantess.	.4	1	199	198	Boiling-type eruption	Aug. 17	Played for 100 hrs.	3	2	201	Has played occasionally since Aug. 17; usual level −3 in.
Giantess	Infrequent	U	20–50	200	198	Rocket-type eruption	Aug.–Dec.	1 eruption	U	20–150	200	Triggered by earthquake, eruption lasting 3 times normal duration; increased boiling; inactive in January 1960.
Vault	5–7 days	1.3 (quiet phase)	8–10	198	184	Dome-type eruption	Oct.–Dec.	Most days several eruptions.	1.3 in quiet phase	8–10	197	Inactive Aug. and Sept.; marked rejuvenation since September.
Cone	Dormant	0	0	193	191	----------	Nov.–Dec.	Several daily	U	25–30	194	No previous record of activity.
Aurum	Cyclic	0	0	201	201	Was in dormant period	Aug.–Dec.	U	35	12–15	201	Rejuvenated to active geyser by earthquake.
Sponge	45 sec–1 min	[1] 3	.5	202	202	Boiling-type eruption	Aug.–Sept.	Dormant	0	0	200	Water ebbed 6 ft; no activity.
							Oct.–Dec.	45 sec–1 min	3	2/3	203	Increased boiling.
Model	Dormant	U	3–5	192	192	Jet-type eruption	Aug.–Dec.	U	U	3–5	194	Occasionally active since quake.
Lion	4 times weekly.	do	50–60	201	200	do	Aug.–Nov.	Very infrequent.	U	50–60	200	Fewer eruptions following earthquake, partial rejuvenation in December.
Big Cub	Dormant	do	40	200	200	do	Aug.–Dec.	Dormant	0	0	203	Water level, +4 in.; increased boiling.
Lioness	do	do	20–30	200	200	Dome-type eruption	do	do	0	0	201	Water level, +6 in.; increased boiling.
Little Cub	1–4 hrs	1	2 to 3; occasionally 10	200	200	Jet-type eruption	do	10 min to 4 hrs.	1.5	3–10	200	More 10-ft eruptions than preceding earthquake.
Depression	2–3 hrs	[1] 62	3	182	186	Splash-type eruption	do	1–2 hrs	[1] 62	3	188	Increased frequency of activity.
Beehive	1 to 2 times weekly.	NM	150–180	203	203	Powerful cone-type eruption.	Aug.–Sept.	Dormant	0	0	203	One eruption following earthquake.
							Oct.–Dec.	3 to 4 eruptions weekly.	NM	150–180	----------	Rejuvenation and marked increase in activity from October to December.
Cascade	Dormant	0	0	186	186	No known activity in more than 40 years.	Aug.–Sept.	20–60 min	400 (approximate)	20–30	200	Marked rejuvenation by earthquake; dormant since Sept. 11.
Plume	60–70 min	[1] 50	20–25	195	197	Cyclic, active since 1942	Aug.–Dec.	42–46 min	[1] 50	20–25	201	New spring flowing into crater slows activity; when checked, interval shows increase noted.
North Anemone	Every 9 to 11 min.	[1] 20	3	197	198	Erupting water drained into north vent.	do	Infrequent	[1] 25	3	198	Most of energy shifted to South Anemone vent.
South Anemone	do	[1] 1.5	1.5	197	197	Chain action, eruption followed North Anemone.	do	Almost constant.	[1] 2	2	197	Chain action in Anemone group has not been characteristic since quake.
Old Faithful	61.8 min	[1] 2,750	130	200	200	Average 61.8 min (based on check of 1,158 eruption intervals between May 1 and Aug. 18, 1959).	Aug	62.1 for 175 intervals.	[1] 2,750	130	200	There has been a somewhat steady increase in length of interval since quake. No observations for length of interval were made in November.
							Sept	65.0 for 239 intervals.	----------	----------	----------	
							Oct	66.8 for 47 intervals.	----------	----------	----------	
							Dec	67.4 for 255 intervals.	----------	----------	----------	
Cascade Group:												
Hillside	1–2 hrs	[1] 200	2–3	200	200	Was in minor eruption cycle.	Aug.–Sept.	20 min–1 hr	[1] 1,300	20–30	202	Earthquake initiated major eruption cycle.
Cauliflower	50 min	[1] 63	1	192	190	Quite regular	Aug.–Dec.	50–60 min	[1] 63	1	192	Eruption a rolling boil; murky after earthquake.
Sapphire Group:												
Black Pearl	Dormant	0	0	199	199	Cyclic	do	Steady	0	3	200	Ebbed 12 in.; steady boiling.
Shell	Irregular	[1] 4.5	1	198	198	Boiling-type eruption	do	2–50 min	[1] 4.5	1	202	Most of the water flows back into crater.
MIDWAY GEYSER BASIN												
Excelsior Group:												
Excelsior	Dormant	3,600	0	199	199	Large steady discharge	do	None	3,600	0	190	Muddy first few days after earthquake.

See footnote at end of table.

TABLE 23.—*Summary of characteristics of geysers affected by Hebgen Lake earthquake*—Continued

[U, undetermined; NM, not readily measurable; PF, periodic flow]

Name	Observations prior to Aug. 17, 1959						Observations after Aug. 17, 1959					
	Eruption frequency	Discharge 1959 (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1951	Temperature (° F) Summers 1958–59	Characteristics	Date	Eruption frequency	Discharge (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1959	Characteristics
LOWER GEYSER BASIN												
Great Fountain Group:												
Great Fountain	12 hrs	U	100	202	202	Eruption readily predictable.	Aug.–Dec.	3–9 hrs	U	100	203	Change in preeruption symptoms; more frequent activity.
White Dome	15–90 min	U	20–25	199	199	Shift in pattern of length of interval.	Aug. 17–21 Aug. 21–Dec.	Dormant 15–60 min	0 U	0 20–25	199 ----	Dormant first 3 days after quake; then resumed normal function.
Pink Cone Group:												
Pink Cone	2–3 times weekly.	[1] 35	12–15	200	200	Jet-type eruption	Aug.–Dec.	1–4 hrs	[1] 35	12–15	201	Tremendous increase in activity.
Bead	32–33 min	[1] 45	15	182	183	Very regular	Aug.–Sept. Oct.–Dec.	55–60 min 15–16 min	[1] 45 [1] 45	15 15	177 ----	Longer interval; shortened to 15 or 16 min after September.
Narcissus	4–5 hrs	[1] 80	12	171	171	Easy to predict	Aug.–Dec.	5–6 hrs	[1] 80	12	168	Slightly longer interval.
Firehole Lake Group: Artesia.	Dormant	0	0	200	200	Cyclic	Aug.–Oct.	Steady	50	3	201	Rejuvenated by earthquake; ceased in October, but steady flow continued.
Fountain Group:												
Jelly	Irregular	[1] 81	2–10	193	192	Different types of eruptions.	Aug.–Dec.	Nearly steady first few days.	[1] 81	10–15	193	Increased frequency and vigor of activity.
Spasm	30 min to 2 hrs.	U	20–30	196	196	Two types of activity	Aug.–Sept. Sept. 10–Dec.	30 min–1 hr Dormant	U 0	20–30 0	200 0	Eruptions accompanied by pronounced steam phase never observed before. Crater emptied and stayed empty.
Bellefontaine	Dormant	0	0	198	198	Splash-type eruption	Aug.–Dec.	30 min–1 hr	25	5–6	200	Rejuvenated by earthquake.
Fountain	2 eruptions in 1959.	U	----	157	159	Fountain type eruption; duration 45 min.	Aug. 18 Aug. 19–Dec.	All day Dormant	U 0	20–50 0	159 ----	Played in concert with Morning and Clepsydra; no such previous record.
Morning	1–2 eruptions weekly.	170	50–150	189	188	Powerful fountain-type eruption.	Aug. 18–31	Almost constant.	170	50–150	193	Activity started by quake; dormant since Sept. 1, with steady overflow.
Clepsydra	1–2 wild-phase eruptions weekly.	[1] 150	----	198	196	Activity initiated by Morning.	Aug.–Dec.	Constant	200	20–30	201	Wild phase initiated by earthquake; has not ceased playing (January 1960).
Sub	Dormant	0	0	184	188	Nearly empty crater	do	Steady	12	0	196	Has not ceased playing (January 1960).
Jet	4–5 min when in active period.	[1] 25	10–15	198	199	Jet-type eruption	do	5–12 min	[1] 25	10–15	199	Action less frequent since Morning and Fountain ceased playing.
Hotel Group: Thud	Dormant	0	0	180	180	Cyclic, infrequent	Aug. 18	None	0	12–15	183	Erupted following quake; dormant since; murky.
Total of maximum rates observed.	----	14,400	----	----	----	----	----	----	16,000	----	----	
Average temperatures.	----	----	----	193.8	192.8	----	----	----	----	----	195.1	

[1] In eruption.

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TABLE 24.—*Summary of characteristics of springs and pools affected by Hebgen Lake earthquake*

[U, undetermined; NF, no flow since quake; I, infrequent; PU, periodic, undetermined; NM, not readily measurable]

Name	Observations prior to August 17, 1959						Observations after August 17, 1959					
	Eruption frequency	Discharge, 1959 (gpm)	Normal height of eruption (ft)	Temperature (° F)		Characteristics	Date	Eruption frequency	Discharge (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1959	Characteristics
				September 1951	Summers 1958–59							
UPPER GEYSER BASIN												
Morning Glory Group:												
Morning Glory Pool.		25		169	163	Steady overflow	Aug.-Dec		0		169	Murky first few days, water 2 to 6 in. below rim of crater.
Spiteful Spring		1.3		198	199		do		1.5		200	Murky first few days.
Grotto Group:												
Bottomless Pit		U		140	195	Discharges into Chain Lake; sudden temperature increase in July.	do		U		198	Discharges into Chain Lake; murky; increase in temperature.
Square Spring		0		201	199	Active geyser in 1950	do		0		200	Murky first few days.
No. 7 Grotto Group.		5		177	175	Steady overflow	do		16		184	Increase in temperature and overflow.
Giant Group:												
South Purple Pool.		15		140	145	Algal-lined pool	do		0		182	Murky first few days, sharp increase in temperature.
East Purple Pool.		5		196	196	Ebbs 3 ft following Giant's eruptions.	do		NF		196	Murky first few days.
North Purple Pool.		0		196	197	do	do		0		197	Do.
Chromatic Spring		0		171	164	Exchange of flow with Beauty Pool.	do		0		164	One-inch ebb following quake.
Beauty Pool		30		161	164	Exchange of flow with Chromatic.	do		58		169	Increase in flow and temperature.
Inkwell Spring		20		200	200	Steady boiling	do		22		201	Increase in boiling and temperature.
Daisy Group:												
Bonita Pool		I		140	104	Ebbs with eruptions of Daisy and Splendid.	do		0		105	No overflow since quake; usual level −3 in.
Brilliant Pool	Erupts following Splendid.	PU	3–20	194	194	Ebbs 1 ft following Daisy's activity.	do		0		203	Ebbed 1 to 3 ft; boiling most of the time.
Punch Bowl Spring		3.3		201	201	Steady boiling	do		3.8		202	Increased temperature and boiling.
Black Sand Basin:												
Black Sand Pool		95		197	197	Clear, deep blue	do	4–6 min	117	1, surge	200	Geyser activity, plus increased temperature.
Green Spring	Cyclic	125	3	159	177	Periodic variation in temperature.	do		125		179	Murky first few days.
Emerald Pool		10		155	147	Bowl lined with yellow algae.	do		10		150	Do.
Handkerchief Pool.		0	3–4	176	177	Occasional eruptive activity.	do		0		165	3-in. ebb following quake.
Rainbow Pool	Cyclic	U	20–100	166	166	Occasional seasonal eruptions.	do		0		163	6-in. ebb following quake; murky.
Sunset Lake		NM		185	183	Heavy overflow color rings.	do		Heavy		190	Eruption following quake, not observed, murky several days.
Myriad Group:												
North Three Sisters.	Cyclic	15	0	181	182	Shift of activity in main bowl.	Aug.–Nov	18–20 min	0	15–20	198	Murky, 3-ft ebb.
							Dec	18–20 min	0	15–20		Jet-type activity.
Middle Three Sisters.		U		171	171	Constant level; flowed into North Bowl.	Aug. 17–31		0		185	3-ft ebb, murky.
							Sept	Occasional	U	15	199	Sudden heavy surge into eruption.
							Oct.–Dec		0			3-ft ebb.
South Three Sisters.		U		183	183	do	Aug. 17–31		0	0	185	Active through September.
							Sept	Occasional	U	15		
							Oct.–Dec		0	0		
Orange Spring Group:												
Orange Spring	Cyclic	0	0	138	139	Eruptions frequent when in active cycle.	Aug.–Dec	Near constant spouting.	U	3	180	Active cycle following quake.
No. 22		Seep		137	137	Slight steady flow	Aug.–Nov		0		141	6-in. ebb; murky first few days.
							Dec		0			2-in. ebb.

TABLE 24.—*Summary of characteristics of springs and pools affected by Hebgen Lake earthquake*—Continued

Name	Observations prior to August 17, 1959						Observations after August 17, 1959					
	Eruption frequency	Discharge, 1959 (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1951	Temperature (° F) Summers 1958–59	Characteristics	Date	Eruption frequency	Discharge (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1959	Characteristics
UPPER GEYSER BASIN—Con.												
Round Spring Group:												
Round Spring		0		178	137	No flow in 1959 preceding quake.	Aug.–Dec.		0		142	Increase in temperature plus 4-in. rise in water level from mid-1959 level.
North Round Spring.		0		151	150	Infrequent overflow	do		0		165	Same as Round Spring.
Pear Spring		0		166	165	Algal-lined spring	Aug.–Dec.		0		194	Erupted following quake.
West Round Spring.		0		142	142	Lined with brown algae	do		0		160	Erupted violently following quake.
Castle Group:												
Castle Group No. 26.		0		199	199	Steady boiling, no overflow.	do		0		200	Murky first few days; steady boiling.
Crested Pool	Rare	15	3	200	200	Normally steadily flowing spring.	do		0		155	Water level −12 in.; large drop in temperature.
Chimney Fumarole.		0		200	200	Steady steam vent	Aug.–Dec.		0 10		200 200	Steam vent steady with development of new spring on side of chimney.
South Scalloped Spring.		0		198	198	3-in. ebb below rim	do		5		202	Increase in temperature and discharge.
Scalloped Spring		0		200	198	34-in. ebb in crater	do		0		201	24 in. ebb, and increase in temperature.
Frog Spring		0		64	64	Frogs in spring	do		2.3	U	196	Erupted following quake; has remained hot with overflow.
Liberty Pool		Seep		132	133	Pool lined with brown algae.	do		.9		163	One known eruption following quake; big increase in temperature.
Oval Spring		0		199	199	1-ft ebb in water level	do		0		201	Erupted following quake.
Belgian Pool		0		164	163	do	do		0		161	Murky, ebb 18 in.
Terra Cotta Spring.	Cyclic	2	0	196	196	Dormant before quake	do	U	15	6–8	197	Active cycle initiated by quake
Grand Group:												
Wave Spring		5		131	134	Steady overflow	do		12.6		163	Erupted following earthquake; hotter with increased flow.
Calida Spring		0		183	183	1-in. ebb below rim	do		1		192	Murky plus overflow, increase in temperature.
Witches Cauldron		80		200	200	Steady boiling overflow	do		80		200	Murky water, only observed effect.
Milk Cauldron		0		190	194	28-in. ebb in crater	do		0		200	Increase in temperature and boiling plus 20 in. ebb.
Geyser Hill Group:												
Teakettle Spring		0		199	199	1-ft ebb; crater empties when Giantess plays.	do		0		203	Increase in temperature and ebb 30 in.
Rock Pool		Seep		198	198	Constant level, clear water	do		0		198	4 in. ebb plus murky water.
Dragon Spring		5		198	198	Steadily flowing spring	do		0		200	Erupted following quake; water then stayed at 35 in. ebb.
West Doublet Pool.		20–40		178	180	Increased flow with East Doublet surge.	do		20–40		193	Murky following quake, increase in temperature.
East Doublet Pool.		U		193	193	Surge about every 5 min; flowed into West Doublet.	do		U		196	Murky following quake; increase in temperature; flowed into West Doublet.
Beach Spring	U	1.5	1–3	178	185	Cyclic in activity	do	Several daily	1.5	1–3	201	Increase in temperature and frequency.
Ear Spring		30		200	200	Superheated	do	Rare	[1] 125	1	202	Erupted following quake; increase in temperature.
West Goggle Spring.	Intermittent	4	Quiet overflow.	196	196	Periodic overflow	do	Overflow every 20 to 30 min.	4–5	Surge	200	Increased temperature and overflow.
Heart Spring		5		200	199	Steady flow	do		0		200	Murky, ebb 1 in. below overflow.
Arrowhead Spring.		0		177	175	Water 4 in. below rim	Aug.–Sept.	30 min.–1 hr	[1] 54	3	200	Became eruptive following quake.
							Oct.–Dec.		1.5			Overflow but not eruptive.
Old Faithful Group:												
Chinaman Spring		.5		200	200	Constant boiling	Aug.–Dec.		.5	20–30	201	Erupted following quake.
East Chinaman Spring.		.3		196	196	do	do		.3		198	Murky; increased temperature.
Blue Star Spring		7		184	185	Clear, shelved spring	do		7		186	Murky following quake.

Cascade Group:												
Gem Pool		85		188	189	Blue, constant level	do		88		194	Murky; increase in temperature.
Sprite Pool		0		167	164	Water 2 in. below rim	do		.8		181	Murky; discharge; increased temperature.
Calthos Spring		0		184	184	Water 6 in. below rim	Aug.-Nov. Dec.	 2 known eruptions.	0 100	 U	195 	Erupted following quake; became hotter in Dec.
Pulcher Spring		0		199	199	Water level 3 ft below rim.	Aug.-Oct. Nov.-Dec.		150 0		201 	Murky, overflowing. Water 6 in. below rim.
Mirror Pool		22		183	183	Constant overflow	Nov.-Dec.		24		184	Murky following quake.
Sapphire Group:												
Black Opal Spring		175		172	172	Erupts on rare occasions	Aug.-Dec.		175 (approximate).		173	Has stayed murky since quake.
Sapphire Pool	15 min	75 (approximate).	3-6	202	202	Deep blue color: true geyser, superheated.	Aug. 17-Sept. 4. Sept. 5-13 Sept. 13-28 Sept. 29-Dec.	 2 hrs 30 min-2 hrs	125 [1] 2,250 125 [1] 1,000-3,000	6-8 100-150 6-8 60-150	204 204 204 204	Constant surging. Major geyser activity. Constant surging. Major geyser activity.
West Mustard Spring.		0		164	163	Pool lined with yellow algae.	Aug.-Dec.		U	3	200	Constant spouter following quake; loss by splashing.
East Mustard Spring.		0		136	157	do	do		U	3	200	Do.
Avoca Spring		3		198	198	Constant boiling	do		0		202	Water −4 ft.; steam vent activity.
MIDWAY GEYSER BASIN												
Midway Geyser Basin Group:												
Indigo Spring		90		199	199	Steady flowing	do		0		201	Murky; ebb 18 in. below overflow.
Grand Prismatic Spring.		U				do	do		U			Ebbed 8 ft. first few hrs. following quake; crater slightly tilted to NE.
Opal Spring		0		162	162	Infrequent eruptions	do		0		180	Murky, hotter.
Turquoise Pool		0		54	54	Filled by overflow from Prismatic.	do		0		132	8 ft. ebb; increase in temperature; murky.
LOWER GEYSER BASIN												
Great Fountain Group:												
Lemon Spring		2.2		147	143	Algal-lined pool	do		2.2		137	Murky following quake.
Broken Egg Spring.		1		151	152	Constant level	do		0		187	Murky, 15 in. ebb, refilled in December.
Firehole Pool		40		196	196	Constant boiling	do		40		196	Murky following quake.
Surprise Pool		25		200	200	Superheated spring	Aug.-Sept. Oct.-Dec.	 U	25 40	 1	202 	Murky; hotter. Periodically active.
Pink Cone Group:												
Shelf Spring		85		199	199	Superheated	Aug. Sept.-Dec.	4 hrs 	350 85	10-12 	202 201	Activated by quake. No further eruptions.
Firehole Lake Group:												
Zomar Spring		800		166	166	Steady flowing spring	Aug.-Dec.		80		167	Very turbid following quake.
Black Warrior Spring.		U		195	195	do	do		0		177	Erupted following quake; ebbed 12 in.
East end Firehole Lake.		0		165	165	Flame-like ascent of gas bubbles.	do		0		174	Water murky and hotter.
East side Firehole Lake.		0		150	151	do	do		0		148	Water murky and cooler.
Fountain Group:												
Silex Spring		13		190	190	Occasional overflow	Aug.-Sept. 15 Sept. 15-Dec.	Constant 	0 13	3-4 0	200 196	Water muddy, steady spouting; ebbed 3 ft. Crater refilled, steady overflow.
Celestine Spring		66		196	196	Steady overflow; clear blue spring.	Aug. 17-28 Aug. 28-Sept. 14 Sept. 15-Dec.	None Constant 	0 0 65	0 6-8 	200 202 199	Water ebbed 50 inches; muddy. Steady jetting of muddy water. Crater refilled with steady overflow.
Fountain Paint Pot.		0		188	192	Active mud pots at south end of crater.	Aug. 17-21 Aug. 21-28 Aug. 29-Dec.		0 0 0		198 200 196	Slow rise in level, mud volcano action. New mud pots at north end; new fumaroles outside main crater. Mud volcano action subsided; new mud pots and fumaroles persisted.

See footnote at end of table.

TABLE 24.—*Summary of characteristics of springs and pools affected by Hebgen Lake earthquake*—Continued

Name	Observations prior to August 17, 1959						Observations after August 17, 1959					
	Eruption frequency	Discharge, 1959 (gpm)	Normal height of eruption (ft)	Temperature (° F)		Characteristics	Date	Eruption frequency	Discharge (gpm)	Normal height of eruption (ft)	Temperature (° F) September 1959	Characteristics
				September 1951	Summers 1958–59							
UPPER GEYSER BASIN—Con.												
Hotel Group:												
Thud Spring (Fungoid).		8.5		179	179	Steady flowing spring	Aug.–Dec.		8.5		189	Murky, erupted night of quake.
Stirrup Spring		6		191	190	do	do		0		191	Do.
Gourd Spring		0		158	157	Algae-lined pool	do		12		186	Murky, erupted night of quake; steady discharge thereafter.
Jug Spring		Slight		163	163	Seep overflow	do		Seep		161	Murky after quake.
Cliff Spring		1.3		197	197	Steady flow	do		1.3		196	Murky, erupted night of quake.
Oak Leaf Spring		Seep		196	196	Quiescent spring	do		0		196	Murky, erupted night of quake; ebbed 6 in.
Kidney Spring	25 min	[1] 12	3–4	195	198	Periodically eruptive	do	25 min	[1] 12	3–4	199	Murky following quake.
Total maximum discharge.		1,250							1,970			
Average temperature.				176.7	176.9						185.57	

[1] In eruption.

TABLE 25.—*Summary of characteristics of springs, pools, and geysers not affected by Hebgen Lake earthquake*

[Observations before and after Aug. 17, 1959. All springs, pools, and geysers not affected by the earthquake are in the Upper Geyser Basin]

Name	Eruption frequency	Discharge (gpm)	Normal height (ft)	Temperature (° F)	Characteristics
Castle Group:					
Tortoise Shell Spring		37	0	202	Steady violent boiling.
Grand Group:					
Lime Kiln Spring		8.8	0	200	Steady boiling.
Geyser Hill Group:					
Pump Geyser	Steady	9	2	199	Constant spouting.
Pendant Spring		3.5	0	201	Steady boiling; overflow.
East Scissors Spring		1	0	200	Do.
Old Faithful Group:					
Cone Fumarole		0	0	200	Gas evolution.
Cascade Group:					
Artemisia Geyser	24 hrs	[1] 1,850	25–30	180	Clear blue spring.
Atomizer Geyser	1–3 hrs	[1] 10	25–35	198	Two eruption types.
Iron Spring		0	0	193	Muddy; water level −5 ft.
Total maximum rates of discharge (rounded)		1,919			
Average temperature				197	

[1] In eruption.

when their eruptions ceased, nearby quiescent springs became active geysers. This suggests that there was no local diminution in total amount of thermal energy.

RÉSUMÉ OF EARTHQUAKE CHANGES

The following is a general résumé, by areas, of the effects of the earthquake upon the hot springs. It gives a general picture as well as some detailed information regarding marked changes that cannot be described adequately in summary tables.

LOWER GEYSER BASIN

During the first few days following the earthquake it seemed that changes in the Lower Geyser Basin greatly exceeded those in basins farther up the Firehole River valley; a greater number of springs were murky, and the major geysers in the Fountain Group were in almost continuous spectacular eruption. There were numerous crisscrossing cracks and evidence of much slumping in the Firehole Lake area (pl. 4). Although the changes in the Upper Geyser Basin were not as immediately spectacular as those in the Lower Basin, time has proved that they were just as great.

The simultaneous and sustained activity of Morning, Fountain, and Clepsydra geysers following the earthquake was entirely without precedent. One of the park naturalists noted that this activity began at the time of the first big tremor. Previous functioning of these geysers was either independent or in chain action. Morning Geyser was rejuveňated in 1946 from long dormancy and has been periodically active since. An eruption of Morning was ordinarily followed by an eruption of Clepsydra and then Fountain. Following the earthquake all three started playing simultaneously, and this activity lasted all August 18. Fountain was dormant following the 18th, but steady play of Clepsydra and periodic eruptions of Morning lasted until September 1, when Morning ceased erupting. Clepsydra was still erupting continuously early in January and the daily discharge of water from these three geysers was in excess of that before August 18.

No immediate earthquake effects were noted in the Fountain Paint Pot area, but by August 21 it was evident that activity in the group was increasing. New mud pots with more violent explosions began to form in the northern end of the crater, and new mud pots and steam vents developed outside the main bowl. In January, most of the mud pots that developed to the north of the crater were roaring steam vents.

A new geyser named "Earthquake Geyser" developed on an old fissure northwest of Fountain Geyser (pl. 4). The new geyser played more than 100 feet high and discharged a large volume of water. After several days of activity a steam explosion along the same fissure opened a new orifice that diverted the discharge and caused less frequent activity of Earthquake Geyser. Its activity increased again, however, just before the end of the year.

All major quiescent springs in the Fountain area were greatly changed by the earthquake. Celestine, Silex, Leather, and Gentian, which were large and steadily overflowing springs, were in a state of ebb the morning of the 18th. The ebbing continued for several days, after which Celestine, Silex, and Leather became violently active and ejected muddy water. This condition continued until September 15, when Celestine and Silex refilled and overflowed in their normal manner. Leather ceased erupting, but its water stayed murky and at a low level. After ebbing 41 inches, the water in beautiful Gentian Pool began to rise slowly, but at the end of the year it was still 6 inches below overflow.

In the Great Fountain Group, both Pink Cone and Great Fountain geysers and several small unnamed units nearby began playing at greatly shortened intervals. The pattern of eruption of Great Fountain Geyser has been almostly completely modified. Two springs near Great Fountain erupted so violently during the night of the 17th that many large pieces of sinter were strewn around their craters.

MIDWAY GEYSER BASIN

Nearly all the springs in Midway Geyser Basin were turbid after the earthquake, and a few ebbed markedly. Although the earthquake failed to trigger an eruption of Excelsior Geyser, the spring with the most voluminous discharge in the park, it did change the color of the water from rich blue to muddy gray. Turquoise Pool not only lost its appealing color, but the water ebbed 8 feet and left the crater nearly empty. The Grand Prismatic and Indigo Springs both ebbed following the earthquake. Grand Prismatic began overflowing again on August 18, 1959, but Indigo was still at a low level in January 1960. The crater of Grand Prismatic was slightly tilted, and most of the flow now goes to the east. This is the largest single hot spring in the park, with a pool approximately 300 feet in diameter. Precise changes in altitude are not known, but the east side probably dropped from ½ to 1 inch relative to the west side of the pool. Flood Geyser became dormant following the quake but was rejuvenated in about mid-November.

UPPER GEYSER BASIN

As in the Lower Basin, the earthquake triggered eruptive activity in most of the geysers in the Upper Basin. Two Upper Basin geysers, Cascade and Economic, began periodic activity after nearly 40 years of dormancy. Giantess, which had not erupted since May 17, played for more than 100 hours; its longest previously observed activity was 36 hours. Vault Geyser, which is connected underground with Giantess, was dormant following the quake until mid-October. From this date through December it played with much greater frequency than ever previously observed. Immediately following the quake, Daisy, Castle, Riverside, Grotto, Oblong, and other quite regular geysers began playing at shorter intervals, and this behavior persisted throughout the remainder of the year. The durations of Grotto's eruptions were much longer than before August 17. An active cycle began in Cliff, Hillside, and Baby Daisy geysers; and eruptions continued into January. Morning Glory and Rainbow Pools ebbed about 6 inches and remained at these levels past the end of the year.

As of January 1960, only one of the big geysers seems to have been affected adversely. It is certain that the Grand Geyser erupted immediately after the earthquake, but it has been dormant since. The Grand, one of the most spectacular in the park, had previously been playing at a 10-hour interval. After August 17, geysers near the Grand were playing at greatly shortened intervals, and one unnamed, formerly quiescent, spring was erupting on a major scale. There is much evidence that the Grand is connected subterraneously with several of these nearby activated springs and particularly with Castle Geyser. It seems, therefore, that an exchange of function, a common occurrence in the geyser basins where springs are connected subterraneously (Marler, 1951, p. 329–342), was caused by the earthquake.

In the Myriad Group, a greater percentage of springs became murky than in any other group in the Upper Geyser Basin; also, a large proportion of its quiescent springs erupted during the night of the earthquake. Activity in the Myriad Group was continuing, and one of the new geysers, the Trail, was playing about 50 feet high in January 1960. The largest springs in the Myriad Group, known as the Three Sisters, began a slow ebb after August 17, and this condition continued to the end of the year. Since the end of November, two of Three Sisters' vents have erupted frequently and regularly. Occasionally other vents in the main craters have erupted.

There was more evidence of surface fracture in the Sapphire Group than in any other group in the Upper Geyser Basin. A large fissure formed parallel to the craters of Black Opal Spring and Wall Pool, and water discharged from Sapphire Pool flowed into it. A number of fractures in this group emitted steam, and many inactive springs became active geysers. Mustard Springs changed overnight from algal-coated pools to steady geysers, and this activity persisted into January 1960.

Perhaps the most spectacular change in the hot springs occurred in Sapphire Pool, the main spring in the Sapphire Group. Before August 17 it played about every 17 to 20 minutes, erupting to a height of 3 to 6 feet. Following the quake, and until September 5, it surged 6 to 8 feet high constantly. On September 5, its steady boiling and surging became periodic, and the spring changed into a major geyser. Its eruptions were quite regular, occurring about every 2 hours, and were massive and spectacular; some of the bursts were fully 150 feet high and 200 feet broad. From September 14 to 29 it reverted to the previously steadily surging cauldron. On the 29th it again became a major geyser, and this activity has persisted. The extreme discharge has eroded and disrupted the surface forms over a wide radius around Sapphire's crater, and tons of old sinter have been washed into the Firehole River.

Whether by coincidence or prime cause, it was after a strong tremor on the night of September 4 that Sapphire Pool began major eruptive activity. Between the time of a similar tremor on the night of the 13th and another one on the night of the 28th, it reverted to constant surging, but on the morning of the 29th it was again in a major eruptive cycle. Other tremors of similar magnitude have occurred since the 28th but without stopping eruptive activity or producing any discernible effects.

For the first few days after the earthquake Old Faithful appeared to have survived the big shock without change. It was observed to be more erratic than usual, with successive longer and shorter intervals between eruptions, but Old Faithful occasionally functioned that way in normal times (Marler, 1957, p. 7). The average interval between eruptions during the summer before the earthquake was 61.8 minutes, the shortest seasonal average on record. By September 1 it had increased to 62.1 minutes, and it continued to increase at least to January 1960. Two hundred and fifty-five eruption intervals timed during the last 10 days of December showed an average interval of 67.4 minutes. This increase is not due to the season, for past observation has indicated that the cold of winter does not affect Old Faithful's frequency of eruption (Marler, 1954, p. 53).

CONCLUSIONS

Most of the hot springs in the Firehole Geyser Basins have their craters and lateral ramifications in glacial gravel, and much of their immediate water supply comes from this gravel. Gravel is especially susceptible to change by strong earthquakes. The Hebgen Lake earthquakes not only produced considerable slumping of the gravels, as in the Firehole Lake area, but they also altered the underground "plumbing" and thereby changed the location and degree of surface manifestations of thermal energy.

Earthquakes have played, and no doubt will continue to play, an important role in the evolution of Yellowstone's thermal areas. For example, the new fractures noted in hot spring areas after the earthquake suggest that similar fractures noted earlier in old sinter sheets and mounds resulted from past earthquakes rather than from dehydration or other causes.

When describing the characteristics of Old Faithful to park visitors, the author has frequently pointed out that Old Faithful might have come into being as the result of an earthquake, as indicated by the large crack that crosses the geyser mound and from which Old Faithful issues. In referring to an intermediate spring, a direct precursor to Old Faithful, the author wrote (Marler, 1956, p. 620):

> In my opinion the intermediate spring which built the terraces and interred the stumps began to flow when some "mechanical adjustment" broke the geyserite shield, thereby tapping arteries that had been sealed off in their upper reaches.

Effects of the Hebgen Lake earthquake do much to confirm this speculation on the genesis of Old Faithful.

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Part 3

General Geology

Pre-Tertiary Stratigraphy and Structure of the Hebgen Lake Area

By IRVING J. WITKIND, JARVIS B. HADLEY, *and* WILLIS H. NELSON

THE HEBGEN LAKE, MONTANA, EARTHQUAKE OF AUGUST 17, 1959

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CONTENTS

PRE-TERTIARY STRATIGRAPHY AND STRUCTURE OF THE HEBGEN LAKE AREA

By Irving J. Witkind, Jarvis B. Hadley, and Willis H. Nelson

ABSTRACT

Dolomite, schist, gneiss, amphibolite, and quartzite of pre-Belt age are widespread in the Madison Range and are the oldest rocks exposed in the Hebgen Lake area. These metamorphic rocks form a basement complex that is overlain by about 8,000 feet of bedded sedimentary rocks representing all the Paleozoic systems except the Ordovician and the Silurian and all three systems of the Mesozoic.

The area contains three northwest-trending structural units: (1) the Madison fault block, which forms the high mountains along the west edge of the area; (2) the Cabin Creek zone of much folded and faulted Paleozoic and Mesozoic rocks north of Hebgen Lake; and (3) the Pika Point area of broad folds and minor high-angle faults, which occupies most of the northeast corner of the area.

The crystalline rocks of the Madison fault block, capped locally by Paleozoic strata, have been thrust northeastward along the Beaver Creek thrust fault onto the rocks of the Cabin Creek zone. The west edge of the Madison fault block is delineated by the Madison Range normal fault, which has been active in Recent time.

The Cabin Creek zone, immediately east of, and parallel to, the Madison fault block, consists of folded and faulted Paleozoic and Mesozoic sedimentary rocks, many of which have been overturned. These blocks of overturned strata are bounded locally on their northeast flanks by thrust faults and on their southwest flanks by normal faults parallel to the overturned beds. Two of these normal faults, the Red Canyon and Hebgen faults, were reactivated during the earthquake. The geologic history of the area suggests that the blocks bounded by these reactivated faults have subsided repeatedly in the past. This pattern of subsidence was continued during the earthquake of August 17, 1959.

The Pika Point area of broad folds is cut by northwest-trending high-angle faults that may be thrust faults associated with the Laramide deformation.

ROCKS OF PRECAMBRIAN AGE

Precambrian crystalline rocks are widespread in the Madison Range and near Hebgen Lake (pl. 5) where they crop out locally beneath the Paleozoic rocks or are overlapped by Tertiary volcanic rocks. Good exposures are southeast of Targhee Pass and east of Hebgen Lake. These crystalline rocks are of pre-Belt age and include various kinds of gneiss, schist, dolomite, amphibolite, and quartzite. Granitic rocks also are in the Madison Range north of the canyon below Hebgen Dam and in the southernmost part of the area west of West Yellowstone.

Gneiss of various kinds, prevalent in much of the area, includes quartzose and feldspathic gneiss of sedimentary or volcanic origin as well as hornblende and amphibolite gneiss which mainly represent mafic igneous rocks. In the southern part of the Madison Range these rocks are fine to medium grained, but northward and southeastward they are coarser and are associated with migmatitic gneisses and pegmatite. Most are micaceous rocks containing varied amounts of biotite and muscovite as well as considerable feldspar. Sillimanite-bearing rocks are locally present, indicating that the metamorphic grade is fairly high, and sheared and retrograded rocks are common.

Dolomite, similar to that in the type Cherry Creek formation of Peale (1896), occupies considerable parts of the Madison Range south of the Madison River canyon but is not abundant elsewhere in the area. It is conspicuous in the south wall of the canyon, where it forms one of the abutments of Hebgen Dam and played a critical part in the landslide near the mouth of the canyon (chapter K). Here, it is a thoroughly recrystallized fine- to medium-grained rock which forms steep craggy slopes with little soil and few trees; characteristic slopes are well displayed just west of the slide. Manifest in most of the dolomite are relict bedding layers 6 inches to 2 feet thick and commonly folded. Also evident are one or more sets of well-developed joints or shear zones.

Minor amounts of micaceous quartzite and biotite-quartz-mica schist are associated with quartz-feldspar gneiss, especially near Hebgen Lake. Most of the quartzite is light colored, fine to medium grained, and muscovitic. Locally it is strongly colored by partings of bright-green mica. Along the southwest shore of the lake, opposite the mouth of the Madison Arm, much quartzite is associated with fine-grained amphibolite intrusives that superficially resemble basalt.

East of Hebgen Lake the Precambrian rocks are coarsely crystalline pink quartzite associated with thin units of dolomite and abundantly intercalated fine-grained hornblende gneiss and schist. Farther north, along the lower part of Cabin Creek and adjacent slopes, much of the Precambrian rocks are dark-gray or greenish-gray phyllite.

The Precambrian rocks at the south end of the Madison Range, where they are overlapped by volcanic rocks of the Yellowstone region, are dominantly light-gray gneissic and migmatitic quartz monzonite that is medium grained, locally porphyritic, and contains small amounts of biotite and hornblende. Intercalated with these rocks are abundant concordant layers of white pegmatite and minor mafic rocks, among which biotite amphibolite is most conspicuous.

ROCKS OF PALEOZOIC AND MESOZOIC AGE

The Precambrian crystalline rocks are overlain in the Hebgen Lake area by about 8.000 feet of bedded sedimentary rocks that represent five of the seven Paleozoic systems and all three systems of the Mesozoic. Only rocks of the Ordovician and Silurian systems are missing (table 26 and pl. 5).

Flathead sandstone.—The Flathead sandstone of Middle Cambrian age rests unconformably on weathered dolomite, gneiss, schist, and other crystalline rocks of Precambrian age. The sandstone is poorly exposed and normally crops out as rounded ledges and isolated knobs. Typically, the Flathead is a light-brown thick-bedded and crossbedded medium- to coarse-grained quartzose sandstone and conglomeratic sandstone. The matrix is moderately well cemented by silica, and in a few places the unit is a quartzite. Scattered irregularly through the sandstone are angular to well-rounded pebbles of quartzite, quartz, and chert, most of which are about half an inch in diameter. These form thin lenses that are common in the basal beds. Near Hebgen Lake the Flathead is about 125 feet thick, although it probably ranges from 75 to 125 feet in thickness in southwestern Montana (Sloss and Moritz, 1951, p. 2142).

Wolsey shale.—The Wolsey shale of Middle Cambrian age overlies the Flathead sandstone conformably. It is a weak and soft fissile unit that weathers to form broad gentle tree- and grass-covered slopes and is poorly exposed. It consists of greenish-gray to gray fissile shales interlaminated with light-brown calcareous fine- to medium-grained sandstones rich in glauconite. A few very thin bedded crystalline limestones are also interspersed through the formation. The Wolsey is about 150 feet thick in the Hebgen Lake area; it ranges in thickness from 125 to 200 feet in southwestern Montana (Sloss and Moritz, 1951, p. 2142).

Meagher limestone.—The Meagher limestone of Middle Cambrian age rests conformably on the Wolsey shale and ordinarily appears as a prominent cliff above the broad slopes formed on the Wolsey. The Meagher is a gray to light-gray thin-bedded limestone marked by many small yellow to orange mottles and nodules. This mottling is so widespread that the formation is known throughout southwestern Montana as the blue and gold limestone—the "blue" referring to the matrix and the "gold" to the yellow mottles. Locally, gray thin fissile shale laminae separate the limestone beds. The Meagher resembles the Upper Cambrian Pilgrim limestone but differs from that unit in that it lacks both intraformational conglomerates and large amounts of glauconite, both of which are common in the Pilgrim. Minor amounts of glauconite are, however, in the uppermost beds of the Meagher. In the Hebgen Lake area the Meagher is about 450 feet thick; it averages 500 feet in thickness in southwestern Montana (Sloss and Moritz, 1951, p. 2143).

Park shale.—The Park shale of Middle Cambrian age conformably overlies the Meagher limestone. The Park shale is poorly exposed; it weathers easily and forms a gentle slope between the cliffs formed by both the overlying Pilgrim and the underlying Meagher limestones. The Park is a grayish-green to maroon uniformly thin-bedded fissile shale which breaks into minute flakes that commonly show as float. The formation is about 135 feet thick in the Hebgen Lake area; it is 100 to 150 feet thick in southwestern Montana (Sloss and Moritz, 1951, p. 2144).

Pilgrim limestone.[9]—The Pilgrim limestone is the lower formation of the Upper Cambrian series and, in the Hebgen Lake area, rests conformably and transitionally on the Park shale. Commonly the Pilgrim crops out in cliffs and steep slopes. The formation is a gray to light-gray massive to thick-bedded crystalline limestone marked by many yellow and orange mottles. It resembles the older Meagher limestone but differs from that formation in several respects. Mottles in the Pilgrim are larger and coarser, the formation contains many intraformational conglomerates, and glauconite is widespread and characteristic of the formation. Intraformational conglomerates predominate in the lower part and are less prevalent in the upper part, where oolitic limestones predominate. Beds of intraformational conglomerate range in thick-

[9] Since this report was prepared in 1960, fossil determinations indicate that the Snowy Range formation of Late Cambrian age is included in beds here mapped as Pilgrim limestone.

TABLE 26.—*Generalized section of pre-Tertiary sedimentary rocks exposed in Hebgen Lake area*

System	Series	Group, formation, and member	Thickness (ft)	Lithology
Cretaceous	Upper and Lower Cretaceous	Colorado group	2,500±	Basal unit is a yellowish-brown massive crossbedded sandstone about 40 ft thick that is tentatively correlated with the Fall River formation. This is overlain by a dark-gray fissile shale, about 250 ft thick, correlated with the Thermopolis shale. Uppermost unit mapped is a yellowish-brown sandstone about 100 ft thick, provisionally correlated with the Muddy sandstone member of the Thermopolis shale. These units are overlain by an unknown thickness of dark-gray fissile shale, mudstone, and sandstone.
	Lower Cretaceous	Kootenai formation	400±	Conglomerate, conglomeratic sandstone, sandstone, and claystone; clastic rocks composed of well-rounded pebbles and fine to coarse grains of black chert and quartz. Claystone generally grayish red to dark orange pink. Uppermost beds are light-gray fossiliferous limestone.
Jurassic	Upper Jurassic	Morrison formation	225	Mudstone, variegated; contains light-brown massive to thick-bedded crossbedded fine-grained sandstone lenses.
		Ellis group: Swift formation	50	Sandstone, olive-gray to brown, thin- to medium-bedded, crossbedded, medium-grained. Contains many oolites, shell fragments, and light-brown well-rounded chert grains.
		Ellis group: Rierdon formation	250	Claystone, calcareous, silty, light-gray to yellowish-gray; locally altering to thin bedded limestone; commonly weathers to gentle slope.
Unconformity				
Triassic	Lower Triassic	Thaynes formation	15±	Limestone and interbedded claystone; limestone is yellow orange, and a ledge former. Claystone is medium yellowish brown on weathered surface, forms moderate to steep slopes.
		Woodside formation	725	Sandstone, dark-red, thin-bedded, locally fissile, slightly calcareous, very fine grained. Forms moderate to steep slopes.
		Dinwoody formation	265±	Sandstone, light-brown, thin-bedded, tightly cemented by calcite, very fine grained. Contains thin greenish-gray shale partings along bedding planes.
Permian		Shedhorn sandstone	150±	Sandstone containing chert nodules; sandstone is medium olive gray to medium yellowish brown, medium bedded, faintly crossbedded, fine to medium grained.
Carboniferous: Pennsylvanian		Quadrant formation	260+	Sandstone, light-yellow to yellowish-gray, thin- to medium-bedded, locally massive, moderately crossbedded, fine-grained.
		Amsden formation	150±	Siltstone, shaly, bright red; intercolated limestone, light gray; limestone marked by pale-red mottles in light-gray matrix.
Carboniferous: Mississippian	Lower Mississippian	Madison group	1,450±	Includes Lodgepole limestone at base and Mission Canyon limestone; Lodgepole limestone is light-gray thin-bedded dense cherty fossiliferous limestone; Mission Canyon limestone is light-gray to medium-gray massive dense saccharoidal relatively nonfossiliferous limestone.
	Lower Mississippian and Upper Devonian	Three Forks shale: Sappington sandstone member	150±	Siltstone, calcareous, yellow-brown and light-brown; overlain by a thin light-gray to green shale. Shale fragments interspersed with small yellow to orange siltstone flags derived from the Sappington sandstone member.
Devonian	Upper Devonian	Jefferson limestone	300±	Limestone, dolomitic, gray to grayish brown, thin- to thick-bedded, dense, medium to coarsely saccharoidal; locally has petroliferous odor.
Unconformity				
Cambrian	Upper Cambrian	Pilgrim limestone	300±	Limestone, gray to light-gray, locally thick-bedded, finely saccharoidal; marked by yellowish-gray silty limestone mottles in brownish-gray limestone matrix; contains thin seams of glauconitic sandstone and mud-pebble conglomerates.
	Middle Cambrian	Park shale	135	Shale, greenish-gray to maroon, fissile; weathers to gentle slope.
		Meagher limestone	450	Limestone, gray to light-gray, thin-bedded, finely saccharoidal; marked by distinctive mottles of yellow silty limestone in dense brownish-gray limestone matrix.
		Wolsey shale	150	Shale, greenish-gray to gray; locally contains thin beds of light-brown quartzitic fine- to coarse-grained sandstone. Glauconite grains common.
		Flathead sandstone	125	Conglomeratic sandstone and sandstone, light-brown, massive- to thick-bedded, crossbedded; matrix consists of medium- to very coarse-grained sandstone. Pebbles are angular to well-rounded quartzite, quartz, and chert.
Unconformity				
		Total	8,050	

Precambrian metamorphic rocks

ness from 6 inches to as much as 2 feet and are composed of flat angular pebbles alined more or less parallel to the bedding planes. Individual pebbles are 2 to 4 inches long, about 1 inch wide, and about a quarter of an inch thick.

The Pilgrim is about 300 feet thick in the Hebgen Lake area; it ranges in thickness from 200 to 300 feet in southwestern Montana (Sloss and Moritz, 1951, p. 2145).

Jefferson limestone.—In the Hebgen Lake area the basal unit of the Upper Devonian series is the Jefferson limestone, which unconformably overlies the Pilgrim limestone. The Jefferson normally forms steep slopes that are more or less covered by dense vegetation and scree. On the basis of scattered outcrops, the Jefferson is a light-brown to brown thin- to thick-bedded saccharoidal dolomitic limestone, mottled locally by dark-gray irregular-shaped petroliferous patches. Locally at least two evaporite-solution breccias, each about 10 feet thick, are in the upper part of the Jefferson. The breccias are composed of light-brown angular fragments of limestone in a matrix

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of light-gray limestone. The limestone fragments range in size from 1/8 inch to about 3 inches on a side; most are about 1 inch on a side. The Jefferson is about 300 feet thick in the Hebgen Lake area.

Three Forks shale.—The Three Forks shale of Late Devonian and Early Mississippian age consists chiefly of yellow-brown and light-brown calcareous shaly siltstone beds. These are overlain by a thin light-gray to green shale, which in turn is capped by a yellow thin-bedded calcareous siltstone tentatively correlated with the Sappington sandstone member. None of these units are well exposed in the Hebgen Lake area. As far as is known, the Three Forks shale is conformable and transitional with both the underlying and overlying units. Commonly the Three Forks appears as a broad grass-covered slope or bench between the steep slopes formed on both the underlying Jefferson limestone and the overlying Madison group. The basal siltstone and the Sappington sandstone member are represented by thin chips, flakes, and fragments of yellow and orange calcareous siltstone and very fine grained sandstone. The Three Forks shale, including the Sappington sandstone member, is about 150 feet thick in the mapped area.

Madison group.—The Madison group of Early Mississippian age consists, in ascending order, of the Lodgepole limestone and the Mission Canyon limestone. Both these formations form ridges and crop out as steep slopes and cliffs that are more or less concealed beneath dense scree and thick foliage. Well-preserved fossils abound, and in most places they appear as fragments in a fossil hash. The Lodgepole is a light-gray thin-bedded dense finely crystalline fossiliferous limestone, through which are interspersed nodules and thin beds of chert. Thin shale seams locally separate the limestone beds. The Mission Canyon is a light- to medium-gray massive to thick-bedded coarsely crystalline limestone that in many places is comparatively unfossiliferous. The uppermost beds of the Mission Canyon locally are light-yellow to yellowish-gray dolomitic limestone. In the Hebgen Lake area a coarse solution breccia that contains angular fragments and blocks of light-gray fossiliferous limestone in a light-brown sandy limestone matrix is about 100 feet below the top of the Madison group. Locally the breccia is deep red.

The Madison group is about 1450 feet thick in the Hebgen Lake area; this thickness may include, however, some repetition of Madison strata as a result of bedding-plane faults.

Amsden formation.—The Amsden formation of Mississippian and Pennsylvanian age rests uncomformably on the underlying Madison group. The Amsden is one of the most striking units in the Hebgen Lake area, and it typically appears as an irregular sequence of light-gray thin limestone beds which alternate with bright-red shaly siltstone and shale beds. The limestone is resistant and stands as steep ridges and small cliffs; conversely, the shale and siltstone are easily eroded and form strike valleys between the flanking limestones. In general the limestone is light gray, dense, very finely crystalline, and characterized by small chert nodules that are bright red, white, gray, or black. The limestone beds in the Amsden contain few fossils or are unfossiliferous. Many of the light-gray limestones are marked by irregular-shaped pale-red mottles and splotches which give the limestones a distinctive appearance not duplicated by any of the other limestones. The bright-red shaly siltstones weather as thin flags and chips which are most arresting in the float. The Amsden ranges in thickness from about 150 feet to about 225 feet.

Quadrant formation.—The Quadrant formation of Pennsylvanian age crops out in the Hebgen Lake area as an unbroken sequence of sandstone beds. Most of the ridges in the area are underlain by the sequence. The Quadrant rests conformably and transitionally on the Amsden formation. In general the Quadrant is a light-yellow to yellowish-gray thin- to medium-bedded moderately crossbedded fine-grained calcareous sandstone. In many places the sandstone is weakly cemented by carbonate and silica and is friable. Elsewhere it is well cemented chiefly by silica and is a quartzite. Each of the sandstone beds ranges in thickness from 2 to 12 feet. The Quadrant ranges in thickness from 260 feet to as much as 620 feet.

Shedhorn sandstone.—The Shedhorn sandstone of Permian age intertongues with the Phosphoria formation of Idaho. In the Hebgen Lake area, all beds between the underlying Quadrant formation and the overlying Dinwoody formation are here grouped with the Shedhorn sandstone. The Shedhorn is conformable with these units and crops out in much the same manner. The Shedhorn is a brown to gray massive to medium-bedded poorly crossbedded fine- to medium-grained sandstone through which are disseminated nodules and tubercles of white to dull-gray chert. The matrix is composed of white quartz sand, through which are scattered minor amounts of amber and black chert as well as white and brown shell fragments. The Shedhorn is about 150 feet thick.

Dinwoody formation.—The Dinwoody formation of Early Triassic age conformably overlies the Shedhorn sandstone. In most places the Dinwoody forms steep slopes mantled by colluvial deposits which support a thick foliage. Where exposed the Dinwoody consists of light-brown thin-bedded dense calcareous very fine grained to fine-grained sandstone and siltstone that are interbedded at irregular intervals with thin greenish-gray shales. Many outcrops of the Dinwoody are characteristically tawny brown. Locally small distinctive manganese dendrites coat joints and bedding planes. The Dinwoody is about 265 feet thick in the Hebgen Lake area.

Woodside formation.—The Woodside formation, the middle unit of the Lower Triassic series, is conformable and intertongues with the underlying Dinwoody formation. Commonly the Woodside weathers as a broad strike valley on which a red soil has developed. In a few places it weathers to form steep slopes. The headwaters of Red Creek rise in the Woodside formation, and during periods of heavy runoff the normally clear water is colored deep red by sediment derived from the Woodside. In general the Woodside is a thin- to medium-bedded sequence of dark-red very fine grained sandstones and shaly siltstone, all of which are characterized by ripple marks, mud cracks, and other evidence of shallow water deposition. The formation seems to be barren of fossils. In the Hebgen Lake area the Woodside is as much as 725 feet thick.

Thaynes formation.—The Thaynes formation, uppermost unit of the Lower Triassic series, is conformable with the underlying Woodside formation. The bulk of the Thaynes consists of thin yellow orange limestone and claystone beds, and the unit consequently is weak and easily eroded. Both it and the Woodside form broad strike valleys. Where the Thaynes is covered, its presence is attested by scattered light-gray fragments of limestone derived from the underlying parent rock. On steep slopes the claystones disintegrate into angular fragments about half an inch on a side which move downslope and mantle the underlying strata.

In the Hebgen Lake area the Thaynes is about 15 feet thick.

Ellis group.—The Ellis group of Middle and Late Jurassic age consists of three formations in southwestern Montana; in ascending order these are the Sawtooth, the Rierdon, and the Swift. Of these, the Sawtooth formation seems to be missing in the Hebgen Lake area, so that the Rierdon rests unconformably on the Thaynes. The Swift formation commonly appears as benches from which younger, less resistant strata have been stripped.

The Rierdon overlies the Thaynes formation with regional uncomformity which is, however, imperceptible on outcrop. As this unconformity is traced northward, the underlying strata are beveled, and the Rierdon rests on progressively older beds. The Rierdon is a light-gray limy fossiliferous claystone that commonly forms broad grass-covered gentle slopes. It is well exposed near the head of Red Creek and in Cabin Creek where it is about 250 feet thick.

The Swift is an olive-gray to brown thin- to medium-bedded crossbedded medium-grained quartzose sandstone. Well-rounded light-brown chert grains are conspicuous, and locally shell fragments abound, thus giving the sandstone the appearance of a shell breccia. Glauconite is common. The Swift is about 50 feet thick in the Hebgen Lake area.

Morrison formation.—The Morrison formation is the uppermost unit of the Upper Jurassic series and seems to overlie the Swift formation conformably. The Morrison forms extensive gentle to moderate slopes that are underlain by thin-bedded sandstone and layers of intercalated claystone. In general the Morrison is composed of light gray to light brown thin-bedded crossbedded moderately friable very fine grained sandstone and light-gray, green, and pale-red mudstone, calcareous claystone, and siltstone. Several sandstone beds as much as 30 feet thick are interleaved in the variegated mudstone-claystone sequence. These sandstones are light brown, massive to thick bedded, crossbedded, and fine grained. Commonly they are moderately well cemented by calcite and silica. The Morrison is about 225 feet thick in the Hebgen Lake area.

Kootenai formation.—The Kootenai formation of Early Cretaceous age rests conformably on the Morrison formation. Three major units are distinguishable. At the base is a coarse bench-forming conglomerate and conglomeratic sandstone composed of well-rounded pebbles of black chert and white quartz. The matrix of the conglomerate is a medium to coarse sand composed of black chert and white quartz grains. Locally the larger clasts are missing and the resultant sandstone is known as salt and pepper sandstone. This basal conglomerate is overlain by a thick sequence of variegated mudstone, claystone, and siltstone beds in which pink, light gray, green, and brown predominate. The uppermost beds of the unit are thin light-gray fossiliferous limestones in which gastropod molds abound. The Kootenai is about 400 feet thick in the Hebgen Lake area.

Colorado group.—Above the Kootenai formation are at least 2,000 feet of sandstone, shale, and mudstone belonging to the Colorado group of Early and Late Cretaceous age. Most of these rocks are light or brownish-gray medium-grained thick-bedded sandstones with subordinate thin-bedded sandy mudstone and shale. They underlie much of the valleys of Cabin and Sentinel Creeks, where they are largely

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covered by glacial deposits and are exposed mainly in stream banks and old landslide scars. The lower part of the Colorado group, immediately above the Kootenai formation, consists of a distinctive threefold sequence that includes a basal sandstone overlain by shale and a second sandstone. These three units lithologically resemble, and may be equivalent to, the Fall River formation, Thermopolis shale, and Muddy sandstone member of the Thermopolis shale, respectively, of eastern Montana and Wyoming. The upper and lower sandstones normally form benches from which younger strata have been removed. By contrast, the shale is nonresistant and forms wide elongate strike valleys.

STRUCTURAL GEOLOGY

The structural events that accompanied the Hebgen Lake earthquake took place in a complex pattern of older structures recorded in the rocks of the Hebgen Lake–West Yellowstone area (pl. 5). In essence, an epoch of late Tertiary faulting and basin formation is imposed on regional folds and thrust faults of Laramide age. Major Cenozoic block faults parallel the Laramide trends in many places but are otherwise quite different in behavior. The Laramide structures are marked by compression; overturned folds and thrust faults are common. By contrast, the Cenozoic structures are marked by tension; normal faults, partly bounding fault blocks, are widespread. The last deformation, partly recorded by the 1959 earthquake, may be a continuation of an episode of block faulting that has been going on since the late Tertiary.

The structural framework of the epicentral region includes a number of different tectonic elements. On the west is the Madison Valley, an intermontane basin which trends northwest and is about 55 miles long and 5 to 12 miles wide. The valley is bounded on the east by the towering mass of the Madison Range, whose west flank is outlined by the Madison Range fault, one of the larger Cenozoic block faults.

Near Hebgen Lake, the Madison Range, consisting largely of Precambrian crystalline rocks, is thrust northeastward over a somewhat narrower belt of folded and faulted rocks of Paleozoic and Mesozoic age called the Cabin Creek fold and thrust zone. This in turn is bounded on the northeast by an area of broadly arched structures called the Pika Point area.

All these elements trend northwest and conform to the regional grain of southwestern Montana. Southwest of the epicentral area, however, the Centennial Mountains trend eastward and end against the south end of the Madison Valley (fig. 49). The north face of the Centennial Mountains is interpreted as "the worn scarp of a fault that has elevated the range and tilted it gently southward" (Pardee, 1950, p. 374). This fault, known as the Centennial fault, has been active in Pleistocene and Recent time. Sluggish drainage and ponds near the fault line indicate that the valley block in front of the range has been depressed (Pardee, 1950, p. 376).

Toward the east, southeast, and south, all these tectonic elements disappear beneath the younger rhyolites of the Yellowstone Plateau, and the Snake River Plain.

MADISON VALLEY

The broad valley west of the Madison Range is bordered on both sides by Precambrian crystalline rocks and is overlapped from the south by Oligocene(?) rhyolite tuff which is downfaulted into the valley. Thin tuffaceous sediments and freshwater limestone of later Cenozoic age lie on the Precambrian rocks in the northern part of the valley, but their relation to the rhyolite in the southern part of the valley is not known. The extent of both groups of rocks is concealed in most parts of the valley by cobble gravel several hundred feet thick that is largely of Pleistocene age.

MADISON RANGE

Near Hebgen Lake the Madison Range is composed dominantly of crystalline rocks of Precambrian age which are locally covered by rocks of Paleozoic age. The internal structure of the Precambrian part of the block is complex and was not studied in detail. Most of the strata are folded and strongly metamorphosed stratified rocks now characterized by variably developed foliation and compositional layering which formed in Precambrian time. Considerable areas of more massive migmatitic gneisses and granitic intrusive bodies are also present. In the area involved in the earthquake study, most of the foliate structures that were observed strike between east and northeast and dip variably north or northwest. Dips are generally steep, although in some places, as in the north wall of the Madison River canyon below Hebgen Dam, the dip is as low as 25°.

The patches of Paleozoic and younger sedimentary rocks in the southern part of the Madison block are gently warped and, in a few places, moderately to tightly folded. They are dragged up steeply along the west side of the fault south of the junction of Watkins and Coffin Creeks; and a narrow syncline trends north-northwest through the ridge southeast of Coffin Lake and on into the southeastern spur of Coffin Mountain.

The moderate folding of the Paleozoic rocks suggests that the Madison block locally yielded to the compressional forces of the Laramide orogeny. The block as a whole, however, was thrust eastward over the rocks of the Cabin Creek zone on the major Beaver Creek thrust fault. This fault, with moderate to low southwest dip, brings Precambrian rocks against rocks ranging from the Amsden formation to the Colorado group and must have a displacement of at least 2 miles. It can be traced along the east side of the Madison block from the north edge of the map area to somewhat north of Hebgen Lake. Near the lake it is covered for a mile or two by glacial deposits in the valley of Beaver Creek. Where it reappears just north of the lake, the fault, or a branch of it, swings southwestward and probably continues in the Precambrian rocks of the canyon below Hebgen Dam. Either the Beaver Creek fault or a related one may extend beneath the lake along the southeast edge of the Madison Range, for west of West Yellowstone, Cretaceous rocks are exposed immediately northeast of the Precambrian rocks of the Madison block. Cenozoic volcanic and glacial deposits have largely concealed the older rocks south of the lake where this fault might emerge.

MADISON RANGE FAULT

The Madison Range fault at the west edge of the Madison Range is a very young normal fault that can be traced northwestward for 70 miles from near Big Springs, Idaho, to a point near Ennis, Mont. The fault borders the east edge of the Madison Valley and is east of Henrys Lake and Reynolds Pass. The fault's length and the continuously high mountain front along it make it one of the largest Cenozoic faults in the region. It is actually a fault system with many local deviations and branches which extend into the blocks on either side. Movement on the fault system began at least by Pliocene time, because Oligocene volcanic rocks in the southern part of the valley and basalt flows in the northern part were downfaulted. Movement has occurred along most of the fault system within the last few hundred years and has resulted in well-preserved scarps locally 10 to 30 feet high and traceable from near Ennis to a point several miles south of the west end of the Madison River canyon (Pardee, 1950, p. 369–373). These scarps are exceptionally well displayed in displaced terraces at the mouth of the canyon. The southernmost scarps extend into the Madison block along Mile Creek where the fault displaces alluvium and talus on the canyon floor.

A small segment of the Madison Range fault was reactivated during the Hebgen Lake earthquake. Two fresh west-facing scarps in surficial material were formed parallel to or coincident with the Madison Range fault. The first scarp, about 400 feet long, is in the N½ sec. 12, T. 12 S., R. 2 E.; the second, about 1½ miles long, extends along the east edges of secs. 14, 22, and 26, of T. 12 S., R. 2 E. (See chapter I.)

CABIN CREEK ZONE

The Cabin Creek zone of folded and faulted Precambrian, Paleozoic, and Mesozoic rocks is immediately east of and parallel to the Madison fault block. This zone, 2 to 4 miles wide, trends northwest near Hebgen Lake and bends northward in the northern part of the area. It is probably continuous with a belt of similar structures that form the west edge of the Madison Range 20 miles farther northwest near Ennis, Mont. Along its east margin the Cabin Creek zone merges imperceptibly with the broad structures of the Pika Point area.

The northern part of the Cabin Creek zone, near Hilgard Creek (SW¼ T. 10 S., R. 3 E., unsurveyed), consists of a large overturned syncline of Mesozoic rocks bounded on the west by the Beaver Creek thrust fault. The southern part of the zone, northeast of Hebgen Lake, is more complex. It consists of two units of vertical or overturned strata exposed in Hebgen and Kirkwood Ridges, with more gently dipping strata between and on either side. These structures are interpreted as major folds whose steep overturned limbs form the ridges, and whose upright limbs lie in the structural and topographic troughs between and beyond the ridges. The folds were overturned and thrust toward the northeast; thrust faults that dip to the southwest appear discontinuously along the squeezed synclinal axes or in the overturned limbs of synclines (fig. 51).

The upright limbs of these overturned folds have been dropped repeatedly and tilted along normal faults probably formed in the late Tertiary. Both Hebgen and Kirkwood Ridges, therefore, are bordered on their north sides by thrust faults and on their south sides by normal faults. It is these normal faults which were reactivated during the earthquake (Witkind, chapter G, p. 37).

HEBGEN RIDGE

A high, narrow ridge, here called Hebgen Ridge, extends along the northeast shore of Hebgen Lake from Dave Johnson Creek to Cabin Creek. The ridge trends about N. 45° W., is about a mile wide, 5¾ miles long, and is 1,300 to 2,000 feet high. It is separated into two almost equal parts by the valley of Kirkwood Creek.

Most of the Ridge is composed of Mississippian, Pennsylvanian, and Permian strata that are nearly vertical, vertical, or overturned (pl. 5, structure section *A–A′*). The northeast edge of the ridge is outlined by the Wells thrust fault, which extends 8¾ miles from its southeast end in the SW¼ sec. 33, T. 11 S., R. 4 E., to a point northwest of Boat Mountain, where it is overlapped by the Beaver Creek fault. The fault surface is not exposed, and consequently its precise location and attitude are unknown. Beds just above the fault dip southwestward from 20° to 50°, which suggests that the fault itself is undulatory and has a correspondingly varied dip. For much of its extent, Carboniferous and Permian strata have been thrust over rocks of like age. Near the northwest end of Hebgen Ridge, Triassic strata are thrust onto Jurassic rocks; and on Boat Mountain, rocks ranging from Cambrian to Triassic are thrust onto Cretaceous rocks.

A second thrust fault, the Johnson fault, is along the southwest flank of Hebgen Ridge and brings steeply dipping Cambrian strata against Mississippian strata along much of the ridge. This fault is relatively inconspicuous and has not influenced the topography of the ridge as markedly as the Wells thrust. It extends for 7½ miles from the W½ sec. 4, T. 12 S., R. 4 E. to the NW¼ sec. 14, T. 11 S., R. 3 E., where it apparently ends as a bedding-plane fault.

The straight southwest flank of Hebgen Ridge is attributed to a normal fault, here called the Hebgen fault. It closely parallels the Johnson thrust fault and for most of its extent is separated from the latter by about 1,000 feet. The Hebgen fault dips southwestward at about 75°, and Cambrian strata are on both sides of it for much of its length. Near its southeast end in sec. 31, T. 11 S., R. 4 E., the fault is marked by juxtaposition of different Cambrian units and structural discordance between beds. Southwest of the fault, Cambrian strata dip about 40° NE; on the opposite side, Cambrian strata dip about 75° SW.

Farther northwest the rocks southwest of the fault are concealed beneath colluvium, but the displacement apparently decreases, for the Cambrian-Precambrian contact cut by the fault near Hilgard Lodge is little offset.

The trace of the Hebgen fault across the southern tip of Boat Mountain is also mostly concealed and is based in great measure upon the location of the fresh fault scarp. In this area the fault probably dips southwest at a high angle, and its displacement is small, involving only Mississippian rocks.

KIRKWOOD RIDGE

A long narrow discontinuous and arcuate ridge north of Hebgen Ridge is another example of a thrust block bordered along its north flank by a thrust fault and along its south flank by a normal fault. The local name "Kirkwood Ridge" is used in this report to refer to this topographic complex, which is cut by Red Creek and extends about 9½ miles from the NE¼ sec. 7, T. 12 S., R. 5 E. to the E½ sec. 24, T. 11 S., R. 3 E. The ridge is convex to the northeast, from ½ to ¾ mile wide, and has about 1,500 feet of relief.

For much of its length the ridge is composed of near-vertical, vertical, and overturned late Paleozoic and Mesozoic strata (pl. 5, structure section *A–A′*). The curving north flank of the ridge is defined by the Divide thrust fault that trends about N. 45° W., is about 7½ miles long, and locally dips southwestward at about 20°. The fault plane is nowhere exposed. Near its southeast end the thrust fault is concealed beneath rhyolites at the west margin of the Yellowstone Plateau. Farther northwest the fault plane is hidden beneath colluvial debris. Despite the cover, it is clear that the rocks of Kirkwood Ridge have been thrust onto Mesozoic rocks which form part of the southwest flank of a broad shallow syncline that plunges northwest (pl. 5, structure section *A–A′*).

The Red Canyon normal fault outlines the south edge of Kirkwood Ridge. The Madison group, Amsden formation, and Quadrant formation are well exposed on and near the crest of the ridge, but the slopes below are everywhere covered by debris concealing the trace of the fault. The Amsden and Quadrant formations are downfaulted against rocks of the Madison group near the center and at the west end of the fault, which suggests that the ground south of the fault has been dropped.

The fault extends for about 9½ miles from its west end in the SW¼ sec. 19, T. 11 S., R. 4 E. to the NW¼ sec. 8, T. 12 S., R. 5 E., where it is concealed beneath a thick earthflow. It may extend southeastward for an additional 3½ miles and end in the center of sec. 22, T. 12 S., R. 5 E. If so, it has a total length of about 12¾ miles.

Dense foliage and thick debris conceal the trace of the Red Canyon fault along the southeastern part of Kirkwood Ridge. In the S½ sec. 36, T. 11 S., R. 4 E. and the N½ sec. 1, T. 12 S., R. 4 E., the plane of the Red Canyon normal fault may coincide with the plane of a preexisting reverse fault. If so, it seems likely that the Cambrian and Devonian strata west of the fault, which were once upthrown, have been dropped back toward their former position. Along the south end of Kirkwood Ridge the Red Canyon fault is not exposed, but it is inferred, chiefly as a result of repeated beds.

PIKA POINT AREA

Northeast of the intensely folded and faulted rocks of the Cabin Creek zone is an extensive area characterized by broad anticlines and domes and wide synclinal troughs. Only the southwestern part, referred to as the Pika Point area, was studied.

Sedimentary rocks ranging in age from late Paleozoic to late Mesozoic are well exposed in the flanks of upwarps centered in the vicinity of Pika Point and Monument Mountain. Locally these broad structures are cut by high-angle faults in which stratigraphic displacements are about a few hundred feet. Most of these faults trend north or northwest parallel to the folds and faults of the Cabin Creek zone; the ground west or southwest of these faults has been relatively upthrown. These faults may be high-angle faults associated with the Laramide deformation. Other faults, in which the ground west and southwest is relatively downthrown, may be normal faults associated with late Cenozoic faulting.

CENTENNIAL VALLEY

The Centennial Valley (fig. 49) can be traced eastward from Lima Reservoir (T. 14 S., R. 6 W., Mont.) for about 40 miles to its junction with the Madison Valley at Henrys Lake (T. 14 N., R. 43 E., Idaho). The valley is about 5 miles wide, and its floor is nearly level. All streams drain westward, and their sluggish courses are interrupted by many ponds, marshes, and sloughs. The south margin of the valley is determined by the Centennial fault which has elevated the Centennial Mountains and tilted them southward (Pardee, 1950, p. 374). The fault has been active in Pleistocene and Recent times, with the valley block downthrown (Pardee, 1950, p. 376).

VOLCANIC ROCKS

All tectonic elements except the eastward trending Centennial Valley disappear to the east, southeast, or south beneath volcanic rocks of the Yellowstone Plateau and the Snake River Plain.

In the eastern, southeastern, and southern parts of the Hebgen Lake–West Yellowstone area, the older rocks and structures are covered by the northwest margin of the great pile of later Tertiary and Quaternary volcanic rocks that forms the Yellowstone Plateau. Here the Tertiary rhyolites dip north and northwestward at low angles, presumably with initial dip from eruptive centers within the plateau. The Quaternary rhyolite and obsidian flows followed this slope in part, but also appear to have filled a caldera-like depression originally suggested by Boyd (1961, p. 412). Minor high-angle faults, recognized mainly by their topographic expression, cut both the Tertiary and Quaternary volcanic rocks; one such group of northwest-trending faults near the mouth of the Madison Canyon east of West Yellowstone may be related to the subsidence of the West Yellowstone basin.

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Volcanic Rocks of the West Yellowstone and Madison Junction Quadrangles Montana, Wyoming, and Idaho

By WARREN HAMILTON

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VOLCANIC ROCKS OF THE WEST YELLOWSTONE AND MADISON JUNCTION QUADRANGLES, MONTANA, WYOMING, AND IDAHO

By Warren Hamilton

ABSTRACT

The southern part of the Madison Range is composed chiefly of a sequence of volcanic rocks about 1,500 feet thick. Welded rhyolite tuff, olivine basalt, olivine trachyandesite, and leucite basalt are present. A fossil pollen flora found low in the sequence dates it as of Oligocene age. The volcanic rocks have been uplifted several thousand feet and much eroded.

Welded tuffs and flows of rhyolite of probable Pliocene age form large sloping plateaus north of Madison River, smaller areas elsewhere, and probably underlie much of West Yellowstone basin. Associated with these rhyolites are small flows of olivine basalt.

The Madison and Central Plateaus are formed mostly of enormous flows of rhyolite of late Quaternary age. The flows have lithoidal interiors and obsidian breccia upper portions; individual flows have volumes of as much as 5 or 10 cubic miles. The youngest flow in the area overlies moraines of both advances of the Bull Lake glaciation but was eroded by a Pinedale valley glacier. The rhyolite flows of the Madison Plateau were erupted from a crestal fissure zone, whereas those of the Central Plateau came from randomly oriented vents.

INTRODUCTION

The volcanic rocks of the Madison Junction, West Yellowstone, and part of the Targhee quadrangles—the southern half of plate 5—were mapped during September of 1959. Fieldwork was thus of reconnaissance nature. As the volcanic rocks are mostly very young, their topography is largely constructional; and as most individual units are large, very rapid fieldwork was possible. With the use of aerial photographs and the excellent topographic maps, it was further possible to extend the outline of many volcanic units beyond the areas of field observations. The regional relationships of the rocks were described by Hamilton (1960).

About 40 thin sections, and several times as many hand specimens, all from the West Yellowstone and Madison Junction quadrangles, were studied in the laboratory.

OLDER VOLCANIC ROCKS OF MADISON RANGE

The rocks of the Madison Range south of U.S. Highway 191 and west of the South Fork of the Madison River are dominantly volcanic, and similar rocks occupy smaller areas to the north (pl. 5). Most of these rocks belong to a volcanic sequence about 1,500 feet thick that has shared in the development of the present Madison Range and has been uplifted several thousand feet and much eroded. Welded rhyolite tuff, olivine basalt, olivine trachyandesite, and leucite basalt are among the rock types present.

Outcrops are poor. The bedrock is almost completely obscured by colluvial rubble; attitudes can seldom be measured; and few contacts between rock types, no matter how distinctive, can be located with confidence.

WELDED RHYOLITE TUFF

The dominant volcanic rock type of the Madison Range is welded rhyolite tuff, which is in all the areas of volcanic rocks. Some of the tuff is much deformed or intercalated in the known older sequences, and it is clearly part of an assemblage believed to be largely of Oligocene age (see below), but some of the rhyolite designated on plate 5 as belonging to the Oligocene assemblage was found on the basis of subsequent work (Hamilton and Leopold, 1962) to be part of the much younger Snake River–Yellowstone assemblage.

The rhyolites of the Oligocene sequence are welded tuffs that carry abundant small phenocrysts of clear sanidine and fewer of quartz and oligoclase; biotite and hornblende are lacking. The rhyolites are thus compositionally of the same type as those of Pliocene and Quaternary ages of the Yellowstone Plateau and upper Snake River Plain and unlike biotitic and hornblendic rhyolites such as are common in other regions.

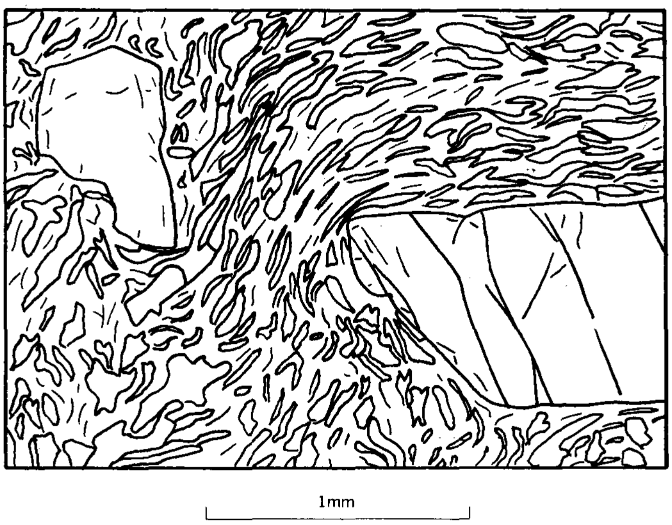

FIGURE 104.—Photomicrograph of welded rhyolite tuff. Two crystals of sanidine lie in a matrix of squashed shards of clear brown glass that displays secondary flow structure. West of peak 7307, 4.5 miles west-southwest of West Yellowstone.

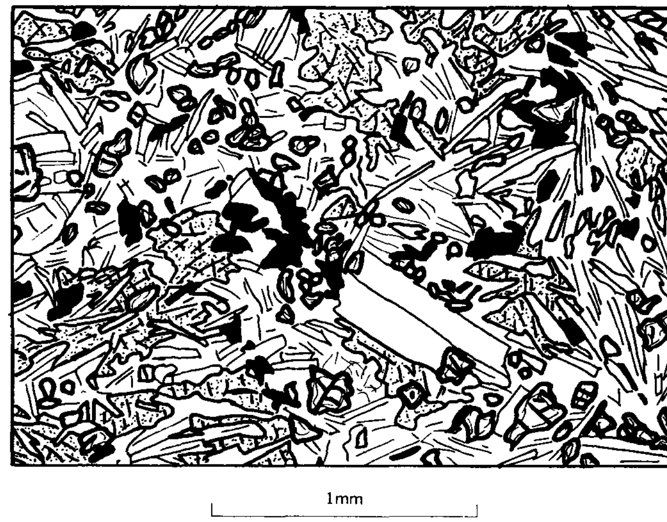

FIGURE 105.—Photomicrograph of olivine basalt. Laths and phenocrysts of labradorite are partly enclosed in subophitic plates of augite (stippled) and interstitial magnetite and ilmenite; granules of magnesian olivine, altered marginally to iddingsite, are strewn about the section. Northwest side of hill 6777, 3.5 miles west-northwest of West Yellowstone.

Undevitrified welded tuff is dark and dense, with the luster of pitchstone, and consists of big shards of brown glass that are thoroughly mashed together and show secondary flowage. Such rock was found, for example, at an altitude of 7,000 feet in the floor of the main north-trending valley 4.5 miles west of West Yellowstone (fig. 104). A chemical analysis of this rock was given by Hamilton and Leopold (1962).

OLIVINE BASALT

A common mafic rock type of the older volcanic sequence is olivine-augite basalt, which was found on hill 6777, 3.5 miles west of West Yellowstone. It occurs as intercalations, both thick and thin, between welded tuffs west and southwest from hill 6777 to the west side of the Madison Range.

The only specimen of basalt from the Madison Range that was studied petrographically came from hill 6777. This is a light-gray rock; like many of the olivine basalts of the Yellowstone region, it is deceptively light colored because of its abundant near-white plagioclase. It has a diabasic texture, with big (0.5 mm) laths of labradorite, plates of augite, and granules of olivine ($-2V \approx 85°$) that have been altered to iddingsite along cracks (fig. 105).

RED OLIVINE-AUGITE TRACHYANDESITE

Distinctive porphyritic red trachyandesite at least 250 feet thick caps hill 7202 (7204?), 3 miles west-southwest of West Yellowstone. It extends at least 2 miles westward from there, and its base rises to an altitude of at least 7,100 feet in that distance; it may be more than 500 feet thick at hill 7663. This is a grayish-red rock studded by very conspicuous stubby euhedral prisms 2 to 7 mm long of black augite, in which are enclosed granules of red iddingsite (fig. 106). There are fewer phenocrysts of normally zoned labradorite and numerous pseudomorphs of red iddingsite after small phenocrysts and granules of olivine. Much of the olivine is enclosed in, or is at the rim of, augite phenocrysts. The dark-red groundmass consists of andesine laths, augite granules, iddingsite granules, magnetite, and interstitial clay(?). Hamil-

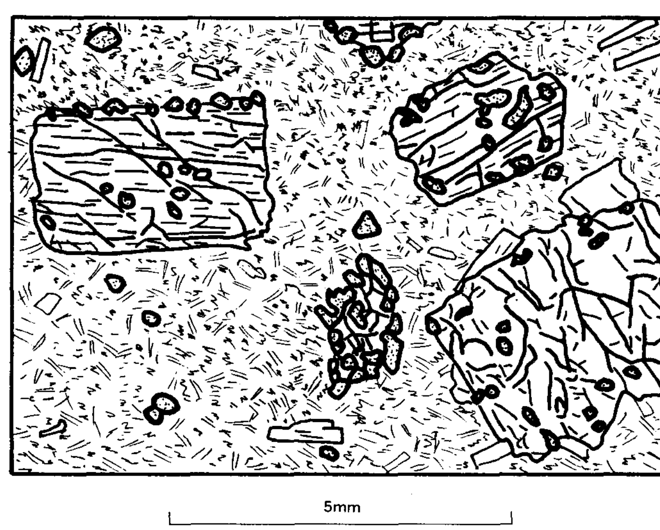

FIGURE 106.—Photomicrograph of olivine-augite andesite. Large phenocrysts of augite, enclosing, and partly rimmed by, granules of olivine (stippled; completely altered to iddingsite), lie in a fine-grained matrix of andesine laths, augite granules, clay, and magnetite. Northeast of peak 7307, 3 miles west-southwest of West Yellowstone.

ton and Leopold (1962) gave a chemical analysis of the rock.

OLIVINE-AUGITE-LEUCITE BASALT

A rock of rare type, olivine-augite-leucite basalt, occurs as a unit about 150 feet thick. The rock has a distinctive appearance; it is dark gray in a freshly broken hand specimen and thickly studded by large (to 5 mm) euhedral phenocrysts of augite (greenish black), olivine, and smaller (1 mm) trapezohedra of leucite, in a matrix of dark glass. The glass weathers to a brownish-yellow surface against which the phenocrysts are conspicuous.

This rock was found on the west slope of hill 7883 (7890?), 6 miles west-southwest of West Yellowstone in the SW¼ sec. 3, T. 13 S., R. 4 E., Montana; apparently the same rock underlies the ridge 1 mile to the southwest, in the SW¼ sec. 10, T. 15 N. R. 44 E., Idaho. On the first-noted hill, it occurs between the altitudes of 7,550 and 7,700 feet, where it lies between sheets of welded rhyolite tuff.

In thin section the phenocrysts are seen to make up about half the rock; the amounts of leucite and augite are subequal and greater than that of olivine (fig. 107). The groundmass is of pale-green glass, completely undevitrified, and contains small brown radial-structured amygdules of low-birefringent material with or without nearly isotropic cores. Opaque material is limited to a very small amount of magnetite. The olivine is colorless, sharply euhedral, unaltered, and has a 2V of 90°. The augite is light green, also unaltered and sharply euhedral, and has a +2V near 65°; the relative richness of the color indicates an unusual composition. The leucite is similarly sharply euhedral and shows neither alteration nor zoning.

There are no crystals of plagioclase, and the name "basalt" is based on the assumption that plagioclase would have crystallized from the glass. The presence of leucite and olivine, however, suggests that the alumina content may be so extremely low that no plagioclase would have formed even had crystallization been complete. Hamilton and Leopold (1962) gave a chemical analysis of this rock.

A remarkable feature of this rock is the complete lack of alteration, embayment, or reaction-rimming of any of its components; liquid (glass), olivine, augite, and leucite were clearly in mutual equilibrium. Despite its remarkable composition, the paragenetic relations seem to indicate simple magmatic crystallization.

The red olivine-augite trachyandesite of the same volcanic sequence was found only 1 mile east of the specimen described above. Despite its red color, this

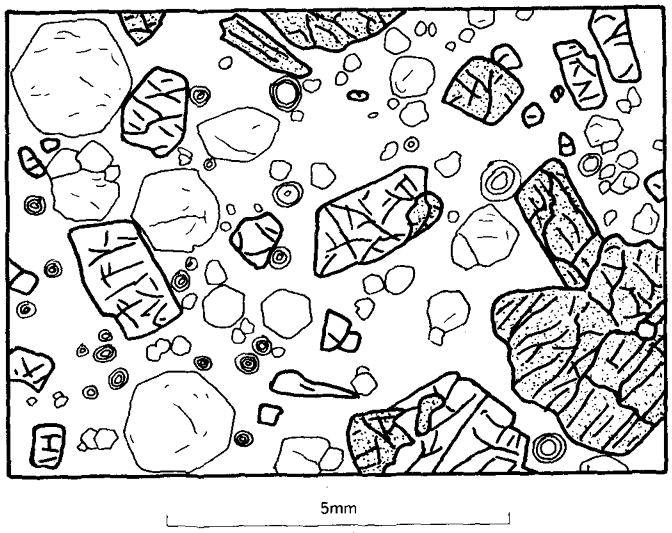

FIGURE 107.—Photomicrograph of olivine-augite-leucite basalt. Euhedral phenocrysts of olivine, green pyroxene (stippled), and leucite, in a matrix of greenish-brown glass which contains zoned spheres of low-birefringent material. SW¼sec. 3, T. 14 S., R. 4 E.

andesite is quite similar in hand-specimen appearance to the olivine-augite-leucite basalt, as it alone of the rocks seen in the region is also studded with big phenocrysts of augite. The base of the red andesite is only 300 feet lower than that of the leucitic rock a mile away, and it is possible that the contrasting types grade together; the intermediate area was not visited. The red andesite differs by its lack of leucite, by the presence of plagioclase phenocrysts, and by the occurrence of olivine as small crystals completely altered to iddingsite rather than as large fresh phenocrysts.

SEDIMENTARY DEPOSITS

Soft sedimentary rocks were found at and near the base of the volcanic section in a number of cuts along logging roads about 6 miles west of West Yellowstone, in the SW¼ sec. 34, T. 13 S., R. 4 E. and in the W½ of the adjacent sec. 3, T. 14 S., R. 4 E. Their aggregate thickness is several hundred feet. The sediments are dominantly water-laid punky tuffaceous siltstones with subordinate sandstone, shale, and mudstone. The rocks are gray to buff where fresh, and these colors in addition to darker greenish browns where weathered. They are laminated to massive and weather to chips. The siltstones contain abundant pollen, studied by Estella B. Leopold (see below), and diatoms.

A few feet of sediments are exposed in the road cut (at the 20°-dip symbol on the map) 3.1 miles north of Big Springs Lookout, at the extreme south end of the Madison Range. Here, basalt agglomerate at least several hundred feet thick is overlain by 20 feet of flaggy rhyolite tuff, that by 6 inches of soft punky white rhyolite ash, that by 2 to 3 feet of stream-laid sand, and that by massive pink welded rhyolite tuff. The soft ash was studied by Howard A. Powers (oral communication, 1960) who found it to be composed of highly silicic glass and relatively large fragments of magnetite and hornblende. No volcanic rocks of comparable composition are known in the West Yellowstone region, and Powers suggested that the ash was derived from the hornblende rhyolites of the Challis volcanics of south-central Idaho.

AGE

Pollen found in the sediments intercalated in the volcanic rocks was dated by Estella B. Leopold as of Oligocene age. The age of the whole sequence is accordingly considered to be Oligocene, although the collection was made from only one locality, and other ages may be represented in other parts of the section.

Tuffaceous siltstones intercalated in the volcanic sequence along a logging road 6 miles west of West Yellowstone were found by Leopold to contain a rich pollen flora with as many as 500 specimens of pollen per gram. She recognized the following plant groups in the pollen:

Conifers: *Sequoia*, *Picea* [spruce], *Pinus* [pine], *Abies* [fir], and *Tsuga* [hemlock].

Broadleaf trees: *Carya* [hickory], *Pterocarya* [lingnut], *Juglans* [walnut], *Betula* [birch], *Alnus* [alder], *Ulmus* [elm] or *Zelkova* [Asiatic elm], and *Ilex* [holly].

Pollen forms, unclassified as to plant group: *Stephanoporopollenites*. Further data are given by Hamilton and Leopold (1962).

According to Leopold (written communication, 1961),

All of these plants were common during Oligocene time. A number of them are not present in the living flora of the West Yellowstone region, were rare or lacking during the Pliocene, and were common during the Oligocene and early Miocene; some are characteristic of the *Oligocene*. *Zelkova* now grows only in Asia, but was common regionally in the middle Tertiary. *Carya* is now limited to the eastern and central United States, but was common in Colorado and Wyoming during the middle Tertiary, and so was *Juglans*, which now lives no closer than the Ozark plateau and northern Mexico. *Sequoia* is now limited to local areas in the western states, and to Asia, but were common regionally in the middle and late Tertiary. *Tsuga* now lives in local stands 200 miles to the northwest, and was common locally in middle and late Tertiary and Pleistocene sediments. *Ilex* now grows in eastern United States and northern Mexico; in the western United States, it has been found only in Montana in beds of Oligocene age. The form species present of *Stephanoporopollenites* is known only from the Oligocene of Wyoming and Colorado.

The absence of several chapparral genera that are characteristic, and often numerically dominant, in Miocene and Pliocene floras, is negative evidence for an Oligocene age. Except for *Ilex*, the assemblage is typical of the White River group of Wyoming. An Oligocene age is highly probable.

A moist-temperate climate is clearly indicated.

The collection has been assigned USGS Paleobotany No. D1430.

RHYOLITE OF PLIOCENE AGE

Welded tuffs and flows of rhyolite of probable Pliocene age form large sloping plateaus north of the Madison River, smaller areas elsewhere, and probably underlie much of West Yellowstone basin.

The plateau north of the canyon is composed of thoroughly devitrified streaky light-gray rhyolite welded tuff that consists of crystals and fragments of sanidine, and a few of oligoclase, in a haze of minute quartz and feldspar and considerable opaque dust. Faint outlines of squashed shards are visible, and there are rare fragments of clinopyroxene crystals. Outcrops are largely limited to valley walls, as the plateau surface is mantled by till. Bedding is parallel to the surface of the plateau, which is probably a constructional plane. Despite the greater dissection of the plateau south of Cougar Creek than of that to the north, rock on opposite sides of the stream appears identical, and there is no appreciable change in surface altitude across it. Cougar Creek follows the approximate axis of a broad basin structure that plunges west-northwest, and the difference in dissection seems related to the positions of consequent streams upon this structure.

The broad dissected sheet of welded tuff 3 miles southeast of West Yellowstone is of similar streaky light-gray to light-red-purple devitrified rhyolite with phenocrysts of glassy sanidine. The tuff weathers to sand and granules (fig. 108) which cover the surface; outcrops are rare. Glacial erratics are almost completely absent from the surface southeast of West Yellowstone, and soil on it is thin.

The northern part of the hill 2 miles east of West Yellowstone is formed on devitrified rhyolite that contains clear sandine and is rich in spherulites.

The mountain including Mount Haynes, which forms the south side of Madison Canyon near its mouth, consists of welded tuff and probably also rhyolite flow-rock. The crest of the northern part of the mass consists of little-devitrified welded tuff that is in part thinly laminated. Beneath and south of this there is massive devitrified spherulite that might have been produced from either a glassy flow or a welded tuff. Both types are very light gray to purplish gray and carry glassy sanidine.

Welded tuff along the South Fork of the Madison River south of the latitude of West Yellowstone prob-

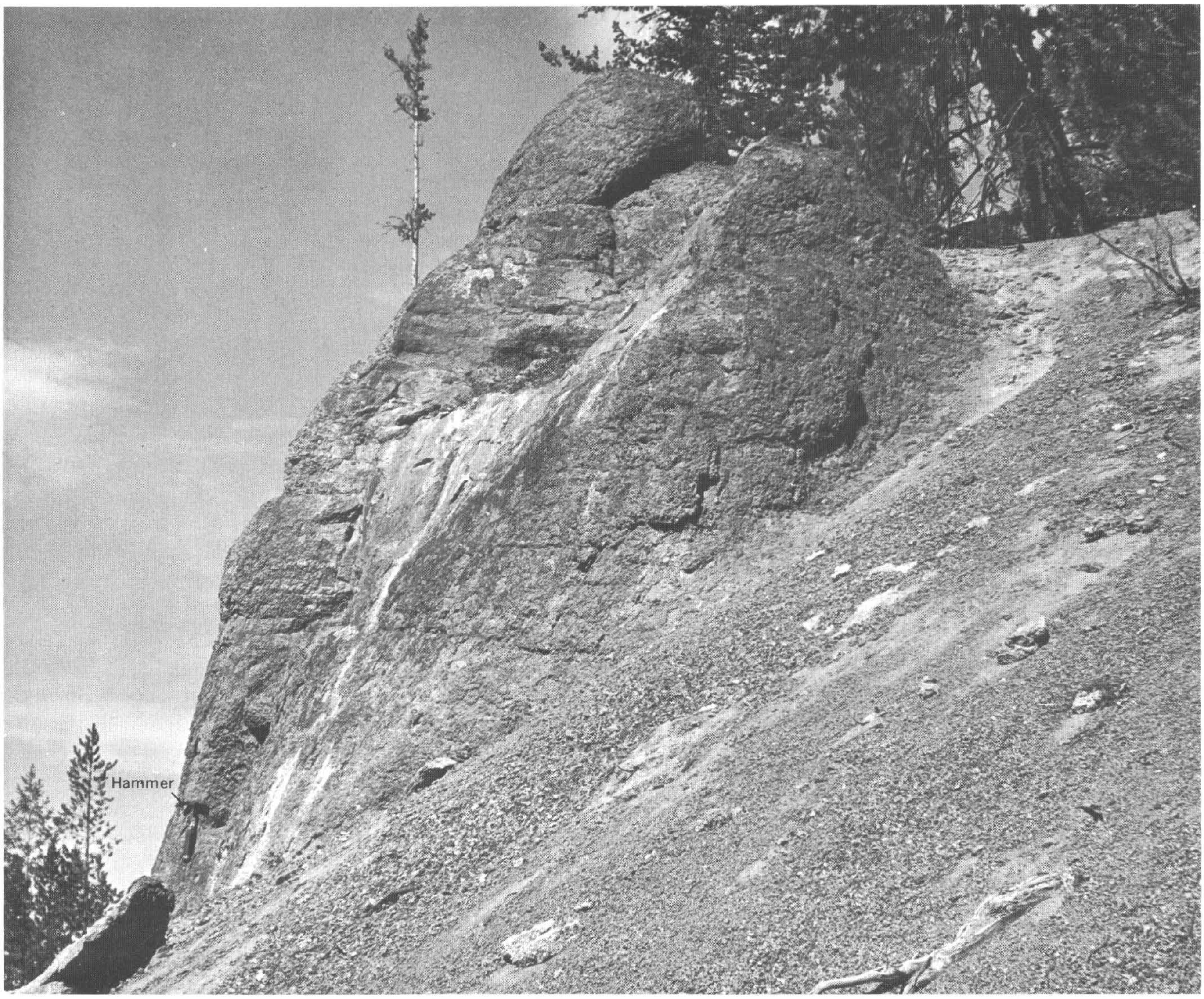

FIGURE 108.—Welded tuff. Squashed-shard layering in this devitrified tuff dips 3° to the right. Rock disintegrates to granular debris and forms rounded outcrops. Hammer at lower left gives scale. Southeast of West Yellowstone 3.5 miles.

ably represents only a single eruption. It is exposed almost continuously in river, road, and railway cuts for 6 miles along the river and dips mostly less than 2°, everywhere less than 5° (fig. 109). Most of the tuff is streaky slabby-weathering rock; less is massive. The rock is light gray with a purplish cast, and the streaks are nearly white where fresh, yellow where weathered. The matrix is of squashed and welded shards that have been thoroughly devitrified to hazes of fine quartz and feldspar and opaque dust. Phenocrysts are clear sanidine and subordinate oligoclase (An_{20}). Tiny fragments of clinopyroxene and red-altered olivine(?) are rare. Biotite and hornblende are lacking.

Some of the welded tuff in the Madison Range is of very similar appearance; although designated on the map (pl. 5) as belonging to the Oligocene sequence, much of it probably formed instead during Pliocene or early Pleistocene time.

Age.—The welded tuff capping the plateau north of the Madison River is of Pliocene or early Pleistocene age, provided the correlation of that tuff with the one in the Morgan-Martzell well at West Yellowstone (see below) is correct. From water-laid rhyolite tuff (unit 6, table 1) immediately beneath the welded tuff in the well, Estella B. Leopold identified pollen dominantly of *Pinus* (pine) and *Artemisia* (sagebrush), with subordinate *Picea* (spruce) and unidentified Com-

FIGURE 109.—Welded rhyolite tuff. The squashed-shard bedding in this thick unit is nearly horizontal. Railroad cut, 3 miles southwest of West Yellowstone.

positae and Chenopodiaceae. According to Leopold, the Compositae first appeared in the region in the lower Miocene, and *Artemisia* in the middle Miocene. All forms still grow in the area, and the material is probably young Pliocene or Pleistocene; the assemblage suggests a pine-and-sagebrush community like the present one.

Similar rhyolites elsewhere in the region have been firmly dated as Pliocene, so these rocks are reasonably designated as of probable Pliocene age.

BASALT OF PLIOCENE OR QUATERNARY AGE

Small flows of olivine basalt occur along the Madison River east of West Yellowstone, and others of basaltic andesite occur between the Madison River and Cougar Creek.

The Madison River occurrences are in three hills, now separated by sedimentary deposits. The hill south of the river, 2.5 miles east of West Yellowstone, is composed of equigranular medium-light-gray dense basalt. The rock consists of flow-alined laths of labradorite (nearly white in hand specimen), semiophitic plates of light-green augite (black in hand specimen), granules and tiny phenocrysts of olivine ($-2V \approx 80°$), and euhedral magnetite and ilmenite. The hill northwest across the river from this locality is of similar rock.

The easternmost exposure of this basalt is just south of the river 4 miles east of West Yellowstone, where the relations between rhyolite and basalt are well displayed in a highway cut (fig. 110). Olivine basalt disconformably overlies welded tuff and was in turn intruded by a dike of basalt. The rhyolite tuff is a light red-purple rock, obscurely streaky, that contains numerous small phenocrysts of glassy sanidine and fewer of quartz and oligoclase; the groundmass fabric is one of big squashed, welded, and partly devitrified shards which contain much hematite dust. The extrusive basalt rests on the eroded surface of the tuff and consists of interlayered lenses of scoriaceous lava and agglomerate composed of dark brownish-gray mafic intergranular olivine basalt; plagioclase laths have a rather feathery appearance; small subhedra of olivine are partly iddingsitized; augite occurs as granules and subophitic plates; and magnetite, ilmenite, and hematite are the opaque constituents. Both the rhyolite and basalt have nearly horizontal attitudes. Cutting the extrusive basalt is a dike of similar composition whose flow structures, as indicated by elongate vesicles, are parallel to its walls; the dike may become a sill at its west side. The dike is composed of dark bluish-gray basalt that is very fine grained and of intergranular fabric. There are sparse small phenocrysts of plagioclase, but most of the plagioclase (normally zoned sodic labradorite) is in tiny laths; augite is light green and occurs in tiny granules; olivine ($-2V \approx 85°$) is in small subhedra; and magnetite and ilmenite are the opaque minerals.

Olivine-augite basaltic andesite overlies welded tuff and is in turn overlain by till 7 miles east-northeast of West Yellowstone. This rock is medium light gray, finely vesicular, and contains sparse small phenocrysts

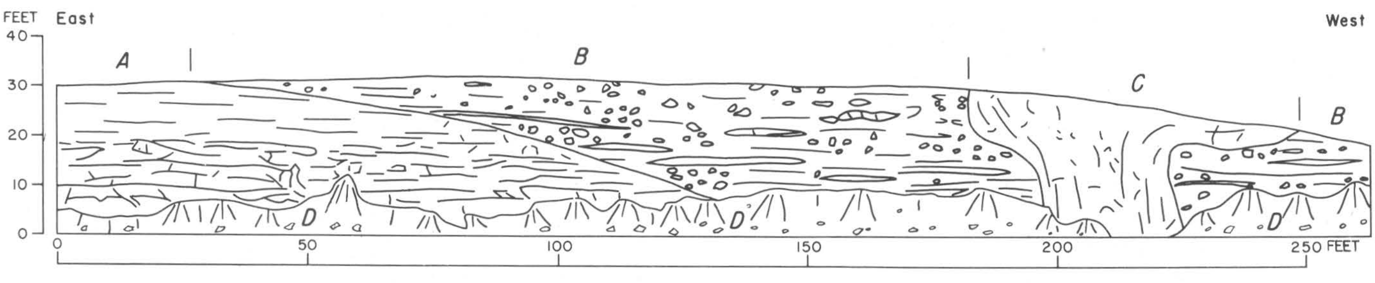

FIGURE 110.—Sequence of rhyolite and basalt of Pliocene or early Pleistocene age. Welded rhyolite tuff (*A*) was eroded, then overlaid disconformably by interlayered agglomerate and scoriaceous lava of olivine basalt (*B*). This in turn was intruded by a steep dike of olivine basalt (*C*) with flow structures subparallel to its walls, (*D*), talus. Highway cut 4.1 miles east of West Yellowstone.

of labradorite in a groundmass composed of laths of calcic andesine, granules of augite, a little very fresh olivine, and magnetite.

The Yellowstone folio (Hague, Weed, and Iddings, 1896) shows several other small flows of olivine basalt within the area of plate 5, both north and south of the Madison River. These were not visited in the present work.

As the basaltic rocks overlie rhyolite of Pliocene or early Pleistocene age, and as they are faulted and considerably eroded, they are considered to be also of Pliocene or early Pleistocene age.

VOLCANIC ROCKS BENEATH WEST YELLOWSTONE BASIN

A section 1,000 feet thick within the volcanic rocks beneath the West Yellowstone basin was cored discontinuously in the Morgan-Martzell well, which was drilled to a depth of 1,390 feet, 1.2 miles north of West Yellowstone in sec. 22, R. 5 E., T. 13 S. The well was drilled in 1953 with the hope of finding oil, because shallow water wells in the vicinity were contaminated by oil seeps at depths of 40 to 100 feet immediately after the Helena earthquakes of 1935. The information and core samples from this oil well were obtained by W. B. Myers from William Martzell of West Yellowstone, whose full cooperation is most appreciated. The well is on the sand plain at an altitude of about 6,645 feet. There is no record of the material in the top 375 feet of the Morgan-Martzell well, but driller's notes (supplied by Martzell) on this interval are available from the Kornowsky well, 0.6 mile west of the Morgan-Martzell. The composite section is as follows:

Section of volcanic rocks and alluvium in Morgan-Martzell well, 1.2 miles north of West Yellowstone

[The top 375 feet are taken from the driller's log of the Kornowsky well, 0.6 mile west of the Morgan-Martzell]

	Thickness (ft)	*Depth, to bottom of unit (ft)*
Collar of well, altitude about 6,645 ft.		
1. Alluvium	200	200
2. Red lava	100	300
3. Lava; probably part of 4, but no samples available	75	375
4. Basalt or andesite, dark-gray to dark-greenish-gray, nonporphyritic; in part vesicular	100	475
5. Rhyolite welded tuff. Light-gray, light-purplish-gray, and light-pinkish-gray rocks composed of squashed and fused shards with many cavities. Scattered crystals of sanidine and fewer of oligoclase and quartz. Glass contains abundant microlites, many of which are pyroxene, and shows practically no devitrification. Flow layers diversely oriented	550±	1,020±20
6. Perlite. Light-gray perlitic glass studded with abundant 2-mm crystals of white sanidine and sparse oligoclase. Microlites of plagioclase and pyroxene in nondevitrified glass. In part spherulitic, with abundant pink spherulites, concentrated in streaks; flow structures (shown by microlites) of the glass pass through the spherulites without interruption, showing that the spherulites are devitrification features. Flow layers diversely oriented. Unit might be part of a single large welded tuff composed of units 5, 6, and 7, as designated here. No cores were available from contact zones	70±	1,090
7. Rhyolite welded tuff. Crystals of sanidine, and fewer of quartz, in matrix of interlensed grayish-orange-pink and grayish-brown squashed and fused shards of perlitic glass. Many of the microlites are of iddingsite pseudomorphous after olivine. Flow layers diversely oriented	30	1,120
8. Water-laid rhyolite tuff. Silty and pumiceous tuffs, in part punky, of angular and unsquashed clasts of glass, and fewer of quartz and sanidine, in dirty-gray to brown matrix composed in large part of slightly altered vitric ash. Maximum size of clasts varies from 0.1 to 0.5 mm. Abundant diatoms, and pollen of *Pinus, Picea, Artemisia*, Compositae, and Chenopodiacaea (as identified by Estella B. Leopold). Bedding planes are essentially horizontal	35	1,155
9. Olivine basalt. Dense medium-dark-gray rock with brownish cast; vitric sheen when viewed with a lens. Composed of large (0.5 mm) laths of plagioclase (An_{50}; dark in hand specimen) partly enclosed in ophitic plates of augite, partly in greenish-brown concentrically zoned chlorophaeite, partly in radial-structured carbonate (not calcite), and partly in brown glass. Minor olivine is in subhedral granules, commonly aggregated. Ilmenite is common	15	1,170
10. Rhyolite tuff. Medium-gray stony tuff, not welded, composed of 1–5 mm clasts of vesicular glassy rhyolite tuff, pumice, and obsidian, cemented but unsquashed. Appearance is similar to cinder-block concrete	15	1,185
11. Augite andesite. Medium-gray dense rock composed of labradorite phenocrysts in a groundmass of andesine laths (very light gray in hand specimen), with subordinate augite, ilmenite, magnetite, carbonate, clay(?), yellow isotropic material, and brownish-green glass. Local zones of amygdules	10	1,195

Section of volcanic rocks and alluvium in Morgan-Martzell well, 1.2 miles north of West Yellowstone—Continued

	Thickness (ft)	*Depth, to bottom of unit (ft)*
12. Olivine basalt. Scoriaceous medium-gray rock which has a granular-vitric appearance viewed with a lens. Composed of laths of plagioclase (An_{55}) in a "dirty" matrix of augite granules, iddingsite pseudomorphs after olivine, ilmenite, magnetite, carbonate, chlorophaeite, and glass. Vesicles partly filled by hemispheres of spherically zoned dark noncalcitic carbonate	20±	1, 215±10
13. Augite rhyodacite(?) of two textural types; either one flow of variable texture or four of alternate textures may be represented. Two of 4 core samples are dark medium-gray rocks with abundant light-gray mottles about 3 mm in diameter. This rock type is composed of phenocrysts of andesine, and fewer of clinopyroxene, in a matrix of flow-alined tiny laths and microlites of calcic oligoclase, tiny prisms of clinopyroxene, granules of magnetite, and interstitial sanidine(?), tridymite(?) and brown glass. There are locally a little poikilitic biotite and veinlets of radial-structured carbonate. The other type is similar except that the material interstitial to the oliogoclase and clinopyroxene of the groundmass is wholly glass; also, the rock lacks mottles. A sample of the nonmottled type was analyzed only for alkalis by Paula Montalto, who found it to contain 3.15 weight-percent Na_2O and 2.72 percent K_2O. This suggests that the rock is a rhyodacite with a very high silica content. Flow structure is horizontal	105	1, 300
14. Andesite agglomerate. Very irregular angular fragments of medium-light-gray vesicular andesite in a yellowish-gray matrix of altered rock of the same composition, interpreted as a flow breccia, although a crushed-rock origin is possible. Unaltered rock is composed of phenocrysts of labradorite, laths of calcic andesine, subordinate subophitic clinopyroxene, much ilmenite and magnetite, minor iddingsite (after olivine), and interstitial high-iron(?) carbonate	20	1, 340
15. Olivine basalt. Light-gray rock composed of near-white plagioclase (phenocrysts, An_{70}; laths, An_{50-60}), ophitic plates of black augite, pseudomorphs of iddingsite and carbonate after small crystals of olivine, ilmenite, and a little green chlorophaeite and carbonate. Vesicles partly filled with radial-structured carbonate	30	1, 365
16. Altered rhyolite tuff. Cherty grayish-yellow rock, partly porcellaneous, with the appearance of fine-grained glass sand	5	1, 370
Drilling abandoned.		

Units 5 to 8, composed of rhyolite tuff about 700 feet thick, probably are equivalent to the section exposed north and 8 miles east of Madison Canyon, and the microfossils date unit 8 as of Pliocene or Pleistocene age. The overlying basalt or andesite (4) corresponds to the small flows above the rhyolite in the outcrop region.

Units 9 to 16 have an aggregate thickness of only a little more than 200 feet and comprise many thin members of diverse types. The basalt, calcic andesite, and rhyolite are similar to those of both the Yellowstone and Madison Range asssemblages—that is, to those of both the Pliocene-Pleistocene and Oligocene-Miocene suites. The rhyodacite(?) has no known outcrop equivalents, and the distinctive porphyritic andesite and leucite basalt of the Madison Range are not in the well section. There is thus ambiguity as to the correlation of the lower part of the well section. It may be equivalent to part of the younger sequence to the east, or to the older one to the west, or to neither.

RHYOLITE OF QUATERNARY AGE

RHYOLITE OF MADISON PLATEAU

The Madison Plateau is chiefly formed of huge flows of viscous rhyolite. A single enormous flow caps most of the Madison Plateau within the area of plate 5. This flow has the plan shape of an irregular V, of which the smaller arm, 10 miles long and 1 to 5 miles wide, extends northwest to within 4 miles of West Yellowstone; the larger arm, slightly longer and 3 to 6 miles wide, extends northeast to Madison Junction. The base of the V is 3 miles south of the map boundary. The northwest arm passed through an older valley near its source and then spread out below with a form rather like that of a piedmont glacier. Its surface is marked by hundreds of flow ridges that are concentric to the lobate form downslope but become tangent to the margin upslope. The northeast arm has a lobate margin more complex than the single mass of the northwest arm, and its steep front forms the north edge of the Lower Geyser Basin, the west wall of Firehole Canyon, and the south side of Madison Canyon from Madison Junction through National Park Mountain almost to Mount Haynes. The front of the flow has been almost unmodified by erosion except in the lowest mile of the Firehole Canyon and along the Madison River, where it was cut back by a valley glacier. The flow surface has been ice-scoured only near its east margin; over most of its extent, the original surface and front of the flow can be seen. Total area of the one flow is 80 square miles, and its volume is 5 to 10 cubic miles.

Presumably, the source of the flow was at its crest. Within the West Yellowstone quadrangle, above and south of the topographic constriction through which the rhyolite flowed to form its northwest lobe, a discontinuous narrow ridge trends south along the western part of the flow. This ridge, with crestal altitudes of nearly 8,600 feet, stands 100 to 150 feet above the adjacent surface of the flow and is 3 miles long. It proably represents a fissure through which much of the lava was extruded. The actual apex of the flow, however, is 2½ miles south of the map area, almost at the point of the V defined by the flow; here, the surface reaches an altitude of 8,800 feet in a broad domelike rise that is steeper to the east than to the west. It is likely that this dome was subordinate to the fissure ridge as a source of lava.

Parts of at least two other flows are present within the area of plate 5. Southwest of West Yellowstone, the youngest flow, described above, overlaps from the north a lobe of a very similar flow, and that in turn overlaps another. The degree of erosional dissection, though slight on even the oldest of these three flows, increases with age. On the east side of the plateau, lobes of two flows are similarly older than the youngest flow, although the correlations with the western masses that are indicated on the map are only presumptive.

The rhyolite flows were exceedingly viscous. Their profiles are markedly convex, and their fronts commonly slope 20°, locally steeper. Above the first few hundred feet of rise, the slopes become more gentle; but even so, the flow surfaces are 400 to 1,000 feet higher a mile in than they are at their edges.

The thick flows consist of an upper zone 100 feet or so thick of obsidian agglomerate. This merges downward with flow-contorted rhyolite several hundred feet thick which in turn merges with massive rhyolite several to many hundred feet thick that rests on an obsidian-agglomerate base.

The upper agglomerates consist dominantly of blocks of black obsidian in a matrix of sandlike unconsolidated glass shards. Exposures are particularly good along the new logging-access road 10 miles south-southwest of West Yellowstone (fig. 111). Obsidian clasts range from granules to blocks more than 50 feet across. Almost all the obsidian is minutely fractured and crumbles readily to sand and granules; massive conchoidal glass is very rare. Many blocks of unweathered obsidian are so granulated that they can be dug open with a shovel. Small phenocrysts of glassy sanidine are ubiquitous, and much of the obsidian is spherulitic. Spherules are variously gray, brown, and, most commonly, pink, and 1 mm to 3 cm in diameter. They are generally clustered in layers parallel to the flow structures, and many blocks are composed of alternate spherulitite and obsidian. Flow structures are widespread, but they are restricted to individual blocks and are chaotically oriented (fig. 112). Blocks of pumice, scoriaceous glass, brown obsidian, and light-gray lithoidal rhyolite are greatly subordinate to black obsidian.

The matrix of the agglomerate is composed of shards, short fibers, and sandlike grains of clear glass that are rather evenly sized 0.02 to 0.2 mm and unsquashed and unconsolidated but generally cohesive where freshly excavated. Larger grains are ropy and pumiceous. Most is colorless, but some is very light gray or light yellowish gray. Small granules of black obsidian are mixed with the glass. Harder matrix, superficially resembling colored concrete, is uncommon (fig. 113). No welded tuff was seen.

The forested surfaces of the flows are thickly strewn with obsidian blocks which project through the soil cover. The unconsolidated matrix is exposed naturally only where there has been particularly rapid erosion. The proportion of the surface represented by exposed blocks decreases markedly with increasing age of flow in the three flows southwest of West Yellowstone.

Sporadic pipelike masses of steeply dipping gray pumice probably represent surges of lava upward after differential movement within the agglomerate was essentially complete.

The only thin section studied from an obsidian block revealed abundant phenocrysts of sanidine, and fewer of quartz, in glass which had a strong flow structure shown by microlites of pyroxene (euhedral prisms less than 0.005 mm long), opaque oxides, and feldspar(?) (needles 0.01 mm long with no measurable thickness). The rock contains trains of pink spherulites, which in section are seen to be feathery to granular and uncolored except for brown rims. The flow structures pass right through the spherulites, which indicates that the spherulites are devitrification products (fig. 114).

The agglomerates have been altered locally by fumarolic activity that was presumably contemporaneous with the cooling of the interior of the flows. Siliceous sinter is widespread as thin coatings on obsidian blocks, and locally it has so impregnated the agglomerate as to thoroughly cement it. Uncommon masses of the unconsolidated-glass matrix have been colored red by discontinuous coatings of tiny granules of hematite, and other masses, representing a further stage of iron addition, have been densely impregnated by limonite and hematite. Sulfur is abundantly pres-

FIGURE 111.—Obsidian agglomerate at top of rhyolite flow. Blocks of black obsidian lie in an unconsolidated matrix of shards, strands, and granules of clear glass. The matrix is dark at top and at lower right because the shards there are coated by iron-oxide dust. Roadcut, 20 feet high, 2.4 miles north-northeast of Big Springs Lookout.

ent with sinter in rare places and may constitute as much as 5 percent of masses of many cubic yards. These features suggest that a chemical study would demonstrate that the less conspicuous magmatic components of the Yellowstone hot springs—halides, arsenates, and others—could also be found in the agglomerates.

The rhyolite flows have been so little eroded that only the upper obsidian-agglomerate zone is exposed in most of them. Few gullies penetrate this zone. The interior of the big flow at the north end of the Madison Plateau, however, is exposed continuously along the lower Firehole River and along the Madison River, where the front of the flow was cut back by

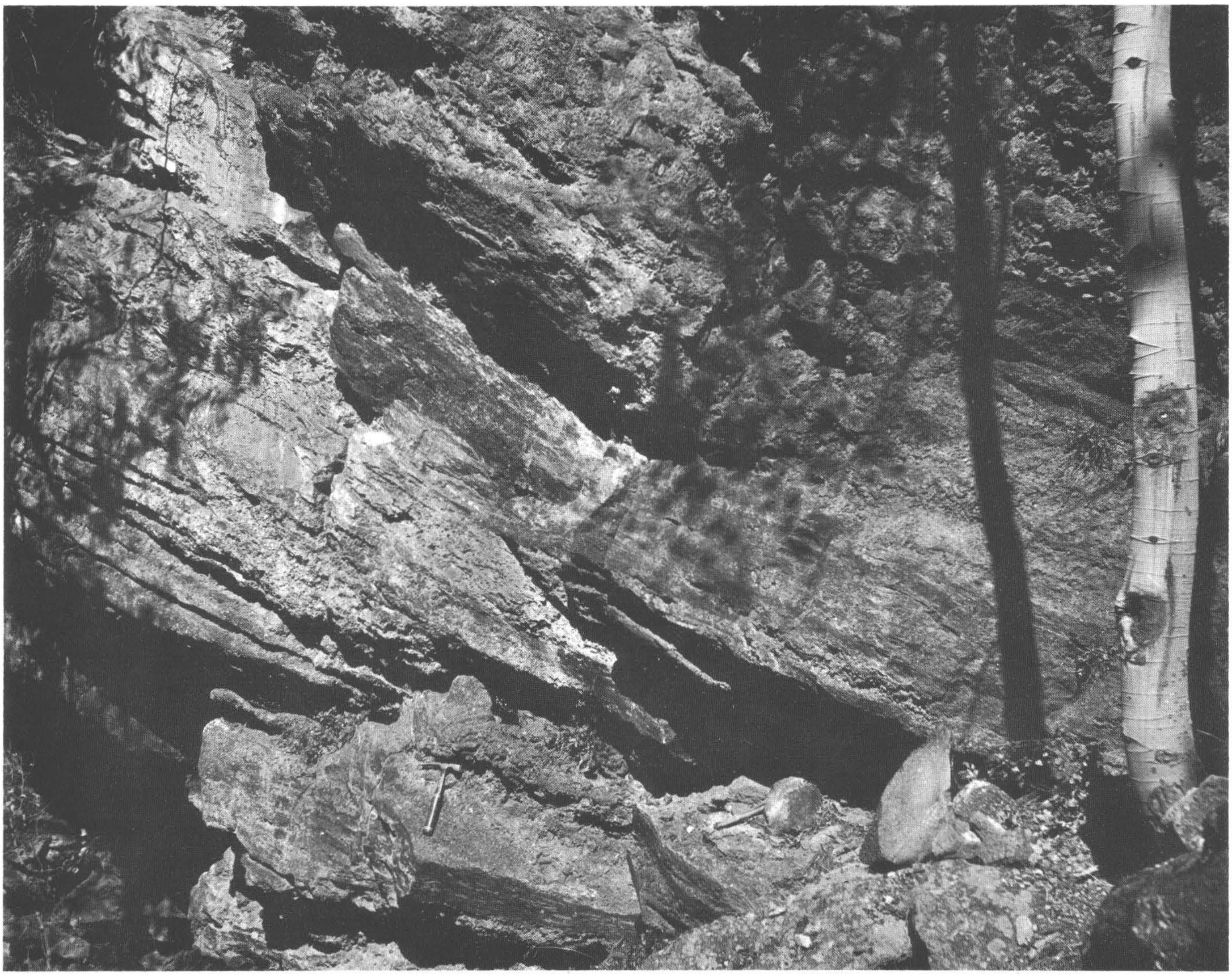

FIGURE 112.—Flow-layered spherulitic obsidian, part of a single huge block within the upper, agglomeratic part of a rhyolite flow. Layers vary from pure glass to nearly pure spherulites and dip back into the flow at this locality near the flow front. Five miles southwest of West Yellowstone.

a valley glacier. The flow-contorted rhyolite near the top of the mass is well exposed along the Firehole River and also on the cliffs 1.5 miles east of Mount Haynes. The main thickness of massive rhyolite also is well exposed east of Mount Haynes, where the one flow is 1,000 feet thick. Although it appears very massive from a distance, F. R. Boyd (written communication, 1961) describes the flow as "finely flow-banded, lithoidal rhyolite."

Boyd also reports seeing the base of one of the flows exposed in Bechler Canyon in the southwestern part of Yellowstone Park. Above the basal obsidian agglomerate is perlitic and spherulitic obsidian which becomes increasingly spherulitic upward and "passes into gray, lithoidal, flow-banded rhyolite" 30 feet above the base.

Age.—The big flow on the northern part of the Madison Plateau overrides moraines of both advances of the Bull Lake glaciation 3 miles southwest of West Yellowstone.[10] The flow was in turn eroded in the Madison Canyon from Madison Junction to Mount Haynes, and its northeastern part was overridden by ice in the vicinity of the Lower Geyser Basin during the earliest advance of the Pinedale glaciation (Richmond and Hamilton, 1960). The flow

[10] The dating of these glacial deposits, information on the soils, and some of the specific relationships with the flows, were determined by G. M. Richmond. (See chapter T.)

FIGURE 113.—Obsidian agglomerate at top of rhyolite flow. Blocks of porphyritic black obsidian are enclosed in cementlike welded shards. Near flow front, 5 miles southwest of West Yellowstone.

is thus precisely dated as within this late Pleistocene interglacial interval.

The next-youngest lobate flow in the West Yellowstone quadrangle is 9 miles southwest of the town and was formed early in the Bull Lake–Pinedale interval. This flow was not overridden by ice during the glaciation. It is covered by a sheet of loess a few inches to several feet thick upon which is developed a soil like that on loess overlying Bull Lake moraines; this suggests that the flow formed while the broad alluvial valley to the southwest was bare and was the source of fine sediment carried away by winds. This second-youngest flow is overlapped by, and is more dissected than, the youngest flow, and has lost some of its smaller scale concentric flow-structure topography.

The third-youngest—the oldest—lobate flow in the West Yellowstone quadrangle is 12 miles southwest of the town and is overlapped by the second youngest. It is considerably more dissected, has lost much of its primary topography, and its front is overlapped from the west by obsidian sand. It may have formed before Bull Lake glaciation.

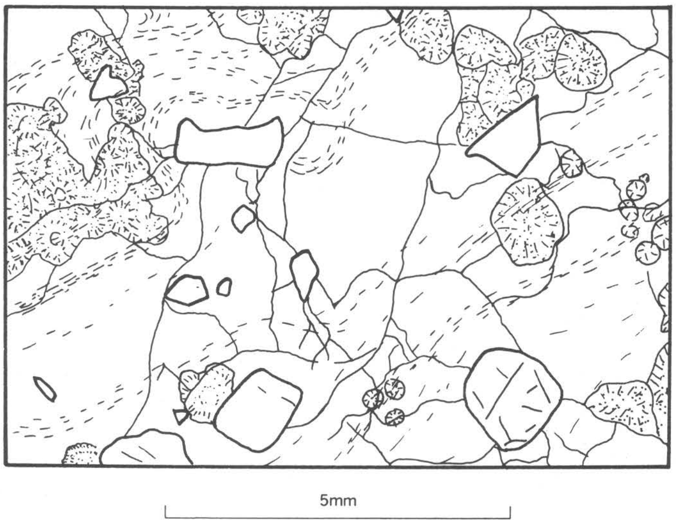

FIGURE 114.—Photomicrograph of spherulitic obsidian. Phenocrysts of quartz and sanidine in clear glass containing spherulites. Flow structures shown by chains of microlites, the size of which is much exaggerated here. Glass is intricately fractured. Obsidian block in capping agglomerate of rhyolite flow, 10 miles southwest of West Yellowstone.

Small post-Pinedale flows lie along the crest of the Madison Plateau south of the area of plate 5 (Hamilton, 1960).

RHYOLITE OF CENTRAL PLATEAU

Large elliptical to irregular flows of rhyolite and obsidian form the Central Plateau shown in the eastern part of plate 5. Like the flows of the Madison Plateau, these flows have steep fronts and surfaces formed of obsidian agglomerate and flow-contorted rhyolite. They differ, however, in having been extruded from diversely oriented vents throughout the plateau rather than from a simple fissure system, and their mutual relations are accordingly much more complex. Most of these flows are smaller than those of the Madison Plateau although still huge by comparison to rhyolites of other regions: on the order of one to several cubic miles each, rather than 5 to 10. There are also some rhyolite domes.

The flows of the Central Plateau have been little eroded, but in places, as in Gibbon Canyon, their interior facies of lithoidal rhyolite are exposed. The upper parts of the flows are well exposed in Firehole Canyon.

The flows of this group studied by G. M. Richmond (oral communication, 1959) are all older than the Pinedale glaciation. Post-Pinedale (and thus Recent) rhyolites are probably present in the Central Plateau, however, and among them is the dome of Gibbon Hill, just east of the area shown on plate 5.

OBSIDIAN SOUTH OF COUGAR CREEK

At least at its southeast end, hill 7085, 6 miles east-northeast of West Yellowstone, is composed of black obsidian. The rock has abundant gray and pink spherulites and lithophysae that are concentrated in layers which dip 3° westward and thus bear no relation to the shape of the hill. Spherulites are as much as 3 inches in diameter and give the rock the superficial appearance of an agglomerate.

In a thin section from this locality, the swirled flow structures (marked by alined microlites) of the glass continue without interruption through the spherulites and indicate that the spherulites are devitrification features. The spherulites are mostly pink in hand specimen and brown in thin section. The glass also contains partial (wedge-shaped) spherulites and blotchy patches of partially devitrified glass. There are scattered phenocrysts in both glass and spherulites; most are sanidine and micropegmatite, but some are oligoclase and fayalite ($-2V \approx 50°$). The obsidian is closely fractured and crumbles readily to granules.

The smoothly elliptical shape of the hill suggests it to be an extrusive rhyolite dome.

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Glacial Geology of the West Yellowstone Basin and Adjacent Parts of Yellowstone National Park

By GERALD M. RICHMOND

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GLACIAL GEOLOGY OF THE WEST YELLOWSTONE BASIN AND ADJACENT PARTS OF YELLOWSTONE NATIONAL PARK

By Gerald M. Richmond

ABSTRACT

Large lobes of glacial ice have extended at least three times into the West Yellowstone Basin from an ice cap that spread across the Yellowstone Plateau from sources in the Absaroka Mountains. The extent of both the lobe and the cap was markedly different each time. Correlative glaciers in the adjacent mountains failed to reach the basin floor except during the oldest glaciation.

The oldest drift crops out only locally but underlies the basin at depth and is overlain by lake beds of glacial silt and younger marls and sands that appear to be of interglacial origin. A thick red clayey soil is developed on these deposits.

The intermediate till comprises two sets of moraines that appear to represent two stades of a single major glaciation, the Bull Lake. These deposits are overlain by glacial-lake silts, by one of a number of obsidian flows that poured out on the Madison Plateau, and by obsidian sand. The post-Bull Lake soil is younger than this sequence.

Moraines made up of a younger till record three stades of Pinedale glaciation. A distinctive soil profile containing thin lenses of volcanic ash is developed on deposits of the Pinedale and locally on older deposits. Both soil and ash are believed to have accumulated during the altithermal interval.

Glacial deposits in the canyons of the Madison Range represent the same succession of Pleistocene glaciations as do those in West Yellowstone basin. In addition, small rock glaciers and talus flows in the cirques record two minor episodes of glacial climate in postaltithermal time.

GENERAL STATEMENT

The West Yellowstone basin has been glaciated at least three times. During each glaciation, a large lobe of ice extended into the basin from an ice cap which spread across the Yellowstone Plateau from sources in the Absaroka Mountains. The extent of both the lobe and the cap was markedly different during each glaciation. Correlative glaciers existed in the mountains adjacent to the West Yellowstone basin but failed to reach the basin floor except during the oldest glaciation.

The oldest drift on the basin floor is correlated with the Buffalo glacial stage of Blackwelder (1915). It crops out only locally but underlies the basin at depth and is overlain by lake beds consisting of glacial silt and younger marls and sands that appear to be of interglacial origin. A thick red clayey soil is developed on these deposits.

An intermediate till is correlated with the Bull Lake glacial stage of Blackwelder (1915). It forms two sets of moraines whose relations suggest that they represent two stades[11] of a single major glaciation. These deposits are locally overlain by glacial lake silts and by a younger obsidian sand that was deposited during and shortly after an outpouring of obsidian flows on the Madison Plateau in the interglaciation between Bull Lake and Pinedale. One of these flows overlies till of the Bull Lake glaciation in the southwestern part of the West Yellowstone basin. A distinctive soil is developed on this sequence of deposits.

A younger till, correlated with the Pinedale glacial stage of Blackwelder (1915), forms a large moraine at the mouth of the canyon of the Madison River at the west edge of Yellowstone National Park. The till also forms two sets of younger end moraines upstream. Three stades of Pinedale glaciation are thus recorded.

A distinctive soil profile containing thin discontinuous lenses of volcanic ash is developed on deposits of the Pinedale glaciation and locally on older deposits. Both the soil and the ash are believed to have accumulated during the altithermal interval.

Glacial deposits in the canyons of the Madison Range represent the same succession of Pleistocene glaciations as those in West Yellowstone basin. In addition, small rock glaciers and talus flows in the cirques record two minor episodes of glacial climate in postaltithermal time.

[11] The terms "glaciation" and "stade" are applied informally and without capitalization throughout this report in the sense recommended by the American Stratigraphic Commission (1959), in lieu of the terms "stage" and "substage".

PRE-BULL LAKE GLACIATION

Blackwelder (1915) first applied the term "Buffalo stage" to pre-Bull Lake glacial drift on the Buffalo Fork of the Snake River, Wyoming, and intimated that, in places, more than one glaciation might be encompassed by such drift. The Kennedy drift in Glacier National Park, which Alden (1926) correlated with the Buffalo stage, is also known to contain deposits of more than one glaciation (Alden and Stebinger, 1913; Alden, 1924, 1932, and Richmond, 1957). Thus the term "Buffalo stage" implies merely pre-Bull Lake Quaternary deposits. In terms of continental glaciation, pre-Bull Lake is considered equivalent to pre-Wisconsin; and where the deposits comprise three superposed tills (Richmond, 1957), they are correlated with the Nebraskan, Kansan, and Illinoian tills of the midcontinent region.

GLACIATION OF POSSIBLE PRECANYON ORIGIN

The presence of high-level till deposits and erratics (up to an altitude of 9,525 feet) in the headwaters of the Gallatin River led Hall (1959) to suggest that, at the maximum of at least one early glacial advance, the ice was essentially continuous from the crest of the Madison Range eastward across the Gallatin Range to the Absaroka Mountains. Reconnaissance observations by the writer and his colleagues failed to disclose erratics on the high divides of the Madison Range in the area affected by the Hebgen quake. The possibility, however, that some pre-Bull Lake glaciation may predate canyon cutting in the mountains is suggested by the size and shape of certain broad cirque basins whose floors are in general accordance with high rock spurs along canyon walls. In these cirques, younger canyon glaciers seem to have been underfit, for their erosive effects have subdivided once-single cirques into tributary parts below the level of the older cirque floor.

POSTCANYON GLACIATION

Ample evidence demonstrates that a glacier from the southeast invaded the floor of West Yellowstone basin at least as far as Horse Butte peninsula (pl. 5) in pre-Bull Lake time. Alden (1953, p. 157) reported erratic boulders of gneiss, rhyolite, and basalt along the crest of Horse Butte, which is above the upper limit of moraines of Bull Lake glaciation on the lower south slopes of the butte.

Till of pre-Bull Lake age is also exposed in quake-caused slump scarps along the shore of Hebgen Lake northeast of Horse Butte. Here, beneath till and associated lake deposits of the Bull Lake glaciation, a thick red clayey soil is developed on eolian silt that rests on a few feet of lacustrine marls and silty clays which in turn overlie an older till. The till is a light-gray to light-yellowish-brown massive clayey silt that contains numerous striated stones and, although not unlike the overlying till of Bull Lake age, clearly is separated from the Bull Lake till by lake deposits and a soil of interglacial origin. The lower part of the lake deposits consists of interlayered light-gray fine-grained sand, clayey silt, and fine gravel that seem to be of late glacial origin. The upper part, however, consists of gray to light-yellowish-brown thin-bedded silty calcareous marl and gray, locally rusty, fine-grained calcareous sand interpreted to be of interglacial origin.

The soil on the lake beds is 24 to 30 inches thick. It consists of a sticky, plastic silty clay that has a strong coarse columnar structure. Its pH is 5.5 to 6.0, and its color ranges from light reddish brown (5YR 6/4) when dry to reddish brown (5YR 5/4) when wet. The unconformity between the soil and the overlying till of Bull Lake age is sharp and locally cuts out the soil completely. At some of these places, smears and foliated lenses of the soil have been incorporated in the overlying till.

At Horse Butte, a similar red soil is developed on bedrock consisting of quartz-rich schist and micaceous quartzite. Here the soil is a sticky, very plastic clay that is about 6 feet thick and has a coarse columnar structure. Its pH is 5.5 to 6.0, and its color ranges from reddish yellow (5YR 6/6) when dry to red (5YR 5/6) when wet. Glacial erratics occur both in and on this soil.

The extent of these deposits of pre-Bull Lake age beneath the West Yellowstone basin is uncertain. At Hebgen Dam (pl. 5), logs of core borings and a geologic section, made in 1911 by B. R. Wallace during construction of the dam, show 12 to 20 inches of red clay at an average depth of 96 feet below the prequake high-lake-level altitude of 6,544 feet. The red clay overlies about 10 feet of obsidian sand that rests on bedrock but includes, in its basal part, a few feet of gravel and boulders. These deposits are here interpreted to be outwash of pre-Bull Lake age beneath interglacial obsidian-sand beds. The deposits overlying the red clay consist of local fan gravel of Bull Lake and younger age and lenses of obsidian sand deposited along the course of the Madison River.

In the southern part of the West Yellowstone basin, logs of water wells, made available by Mr. Carl Hollensteiner, well driller, record 30 to 50 feet of volcanic sand and gravel containing local lenses of blue clay beneath till of the Bull Lake glaciation at depths ranging from 90 to 130 feet below the ground surface.

These deposits are here interpreted to be of pre-Bull Lake glacial and interglacial origin.

The pre-Bull Lake glacier responsible for these deposits in the West Yellowstone basin was a lobe from an icecap that extended over the Yellowstone Plateau into the drainage of both the Madison and the Gallatin Rivers. Alden (1953) reports erratic basalt boulders on the hills southeast of Grayling Creek along the border of Yellowstone National Park, and I. J. Witkind (oral communication, 1959) found patches of old till and erratics as high as 8,400 feet in altitude on the broad ridge that separates Grayling Creek from Tepee Creek, above the upper limit of fresher and thicker till of the Bull Lake glaciation along Tepee Creek (pl. 5).

Pre-Bull Lake local mountain glaciation is indicated by erratic boulders of Precambrian rock that were derived from the drainage of Beaver Creek at the southeast corner of Boat Mountain (pl. 5). Their position 300 to 400 feet above moraines of the Bull Lake glaciation and 600 to 700 feet above the Madison River indicates that the glacier which deposited them must have blocked the canyon of the river to an altitude of about 7,000 feet and formed a lake in the northern part of the West Yellowstone basin. The ice probably extended down the canyon to a terminus at or beyond its mouth, although all traces of terminal deposits have been removed.

BULL LAKE GLACIATION

The Bull Lake stage of glaciation was named from moraines at Bull Lake, Wyo., (Blackwelder, 1915). The till is younger than the very strongly developed reddish soil on deposits of pre-Bull Lake glaciations and is separated from the deposits of the next younger glaciation, the Pinedale, by a distinctive strongly developed soil whose distribution suggests complete deglaciation at the time it formed. Correlation of the Bull Lake glaciation with events in the midcontinent region is uncertain, as no major withdrawal of the ice is known to separate the several subdivisions of the classical Wisconsin glaciation. It has been suggested (Flint, 1956), that the Bull Lake glaciation may represent an episode after Illinoian time and before classical Wisconsin time that is only beginning to be suspected in the midcontinent region (Flint and Rubin, 1955; Forsythe, 1957).

GLACIATION IN THE WEST YELLOWSTONE BASIN

In the West Yellowstone basin, deposits of Bull Lake glaciation are much more readily recognized than are those of preceding glaciations. A large morainal complex composed mostly of till, but including some kame sand and gravel, forms a northwestward projecting loop that extends across the narrows of Hebgen Lake, wraps around the southeast end of Horse Butte below an altitude of 6,800 feet, forms a second discontinuous loop across the Madison Arm of the lake, and overlaps a rocky promontory on the west shore (sec. 33, T. 12 S., R. 4 E.). Alden (1953) included both this till and the erratics on Horse Butte in what he called Illinoian or Iowan glaciation.

In detail, the deposits form two distinct broad moraines, an inner and an outer, which, along the east side of Horse Butte, are separated by a distinct outwash channel. A similar channel lies between the outer moraine and bedrock underlying the butte. The crest of the outer moraine is 50 to 100 feet above that of the inner along the slope of the butte, but at the Narrows north of the butte the inner breaches and extends slightly beyond the terminal arc of the outer.

The two moraines are much alike. Both are large and bulky; their slopes are smooth and rolling but modified by erosion and slumping to an extent that enclosed depressions, although present, are uncommon. Boulders are scattered over the surface of both moraines but are not abundant, and a layer of eolian silt 6 inches to 2 feet thick mantles the till. Lithologically the till in both moraines is similar, but there is one significant difference. On the west and south sides of Horse Butte both tills are composed predominantly of volcanic rock, mostly rhyolite and obsidian, although some is bluish-gray vesicular basalt. To the northeast, the lithology of the outer moraine changes rapidly to include Precambrian granite, gneiss, schist, and quartzite, and various Paleozoic limestones and sandstones derived from the drainage of Tepee Creek and Grayling Creek. The inner moraine, however, maintains its predominantly volcanic lithology to a point nearly half a mile northeast of the change in the outer moraine. This indicates that the two moraines are related to different glacial regimens, here interpreted as representing two distinct stades of the ice front. The fact that the moraine of the late stade breaches that of the early stade at the Narrows further supports this conclusion. No evidence indicative of the extent of interstadial recession was found.

On the northeast side of the West Yellowstone basin, east of the Narrows of Hebgen Lake, the two moraines are covered by an alluvial fan at the mouth of Red Canyon and are indiscernible from the slump and talus debris along the base of the ridge of Paleozoic rocks to the east. The moraines are distinguishable at Corey Spring, however, where the outer contains considerable kame sand, and both can be traced along the mountain slope to the mouth of the canyon of Grayling Creek. Here both moraines are offset topographically about 120 feet upward to the west on opposite sides of the creek, a situation which the writer believes may be due to faulting. East of Grayling Creek the inner moraine progressively overlaps the outer moraine to a point beyond which only a single moraine, marking the outer limit of Bull Lake glaciation, can be traced across Campanula Creek and up the crest of the ridge to the southeast in Yellowstone National Park.

On the southwest side of West Yellowstone basin the outer moraine is displaced by faults on the point opposite Horse Butte, and the inner is preserved only as low knobs projecting above the alluvial floor of the basin. To the southeast, however, both moraines are clearly discernible along the slopes of the basin to the South Fork of the Madison River, although they are cut out by a younger alluvial fan complex along Denny Creek. The crest of the outer moraine rises from an altitude of 6,550 feet at the shore opposite Horse Butte to 6,960 feet on the west side of the South Fork of the Madison River; the inner moraine rises more gradually through the same distance to an altitude of 6,650 feet. Southeast of the South Fork the moraines form broad ridges more than a hundred feet high that pass beneath the front of a large obsidian flow which forms the southern margin of the West Yellowstone basin (pl. 5). As thus outlined, a broad glacial lobe, bounded on the southwest by the South Fork of the Madison River and on the northeast by the ridge between Gneiss Creek and Campanula Creek, flowed from the Yellowstone Plateau into the West Yellowstone basin as far north as the Narrows and Horse Butte during the Bull Lake glaciation. It appears to have had at least two significant pulsations, now marked by the position of an outer and an inner moraine.

The soil on both moraines of the Bull Lake glaciation is similar and is commonly developed through the full thickness of the eolian mantle and into the underlying till. Fresh exposures along slump scarps caused by the earthquake reveal that the thickness of the soil averages 18 to 24 inches and is as much as 36 inches in places. The profile consists of a thin gray A horizon 2 to 8 inches thick over a compact sandy silt B horizon 20 to 30 inches thick in which the material is light brown (7.5YR 6/4) to light yellowish brown (10YR 6/4), nonsticky, nonplastic, and has a moderate, medium angular blocky structure. Its pH ranges from 6.0 to 6.5. The fresh till is light gray (10YR 7/2) to very pale brown (10YR 7/3) and is compact but friable.

Between the B horizon and the fresh till is a zone ranging from a few inches to over a foot in thickness in which fractures in the till and the pebbles and cobbles are coated with a white noncalcareous material. The material was examined in the laboratory by Ray E. Wilcox, who reported as follows:

> The coatings are composed of isotropic to weakly anisotropic material through which are scattered particles (up to 0.01 mm diameter) of definitely anisotropic material.
>
> The isotropic matrix has a refractive index that ranges from about 1.460 to about 1.456. It is probably opaline silica in varying degrees of dehydration and transition to a crystalline lattice, such as cristobalite.
>
> The definitely anisotropic material has much higher refractive indices and appears to consist entirely of original components of the till.

A report of an X-ray analysis of the coatings by Fred A. Hildebrand indicates that the material is

> principally cristobalite with moderate amounts of quartz and feldspar, possibly sanidine. Amorphous silica, which may be present, cannot be detected by X-ray powder diffraction methods.

LOCAL VALLEY GLACIERS

Moraines of the Bull Lake glaciation occur in many of the canyons of the Madison Range surrounding the West Yellowstone basin. A rapid reconnaissance was made in only three of these canyons. In general, though the glaciers were much less extensive than that in the West Yellowstone basin, the surface characteristics of the moraines, their degree of dissection, and the soil developed on them are much the same, except that no zone of opaline silica accumulation is present in the soil.

For example, at the mouth of Beaver Creek, north of Hebgen Dam, at an altitude of 6,550 feet, is a large moraine that is believed to represent the early stade of Bull Lake glaciation which at one time blocked the course of the Madison River. Alden (1953, p. 178) considers it to be of Wisconsin age. The deposit has relatively smooth boulder-littered slopes and is dissected by both Beaver Creek and the Madison River. It is more hummocky than other moraines of Bull Lake glaciation and contains two undissected depressions. The till is composed largely of material derived from Precambrian crystalline rocks; and the soil on it, although not well preserved along the highway, is typical of that on other moraines of Bull Lake glaciation.

A second moraine, representing the late stade of Bull Lake glaciation, lies at the junction of Beaver Creek and its West Fork, where glaciers from both canyons merged. A post-Bull Lake soil is exposed on the moraine, which is trenched as much as 40 feet by outwash of a younger glaciation. A few cobbles of dense percussion-marked quartzite, typical of Tertiary conglomerates not present in the drainage of Beaver Creek, were noted in the till and were believed to have been reworked from deposits of pre-Bull Lake glaciation.

Other pairs of moraines with characteristics common to those of Bull Lake glaciation were noted in Red Canyon northeast of Hebgen Lake and along Watkins Creek northwest of the lake (pl. 5). I. J. Witkind (oral communication, 1959) has observed till of Bull Lake age in the basin of Tepee Creek, and most of the mountain canyons probably were glaciated at this time. Glaciers along streams in the headwaters of the Gallatin River, however, such as the upper course of Grayling Creek, were derived from the ice cap on the Yellowstone Plateau.

OTHER DEPOSITS ASSOCIATED WITH BULL LAKE GLACIATION

Several kinds of deposits other than till can be shown to have formed, at least in part, during Bull Lake glaciation through their stratigraphic relations to the till or to the post-Bull Lake soil.

LAKE BEDS

Along the southeast side of the loop formed by the inner moraine of the Bull Lake glaciation at the Narrows of Hebgen Lake are local exposures of light-gray (2.5YR 7/2) bedded silty lake clay containing scattered coarse to fine sand grains and thin lenses of fine gravel. The beds are dense, 2 to 6 inches thick, and weather buff. They rest on a pre-Bull Lake reddish soil and are overlain by stony colluvium derived from the moraine upslope. From their stratigraphic relations they seem to represent deposits formed in a lake retained back of the inner moraine after recession of the ice. Drillers' logs of water wells throughout the basin southeast of the Narrows record, beneath about 80 feet of obsidian sand, a bed of "yellow clay" and local interlayered sand that ranges in thickness from 10 to 30 feet and rests on bouldery till of the Bull Lake glaciation.

The lake beds are reported by the driller, Mr. Carl Hollensteiner, to have a green cast locally and to become more sandy along the edges of the basin. An exposure of green hued crossbedded sand exposed disconformably beneath the surface alluvium in a faulted terrace along the South Fork of the Madison River (center sec. 14, T. 13 S., R. 4 E.) may represent such a deposit. The crossbedding is of the long deltaic foreset variety.

OUTWASH GRAVEL

Other than kame deposits along moraines, very little outwash gravel of Bull Lake age was observed. Outwash that may have been deposited in front of the moraines is covered by Hebgen Lake. A small deposit was noted on the north side of the Madison River at the mouth of its canyon in Yellowstone National Park, north of, and beyond, the outer limits of a moraine and associated outwash of the Pinedale glaciation. Other small deposits were noted below moraines of Bull Lake age in the mountains adjacent to the basin.

ALLUVIAL-FAN GRAVEL

North of the Narrows, Hebgen Lake is bordered by numerous alluvial fans, most of which seem to be of post-Bull Lake age. Some, especially along the east side of the lake, bear a soil which, although similar to that on moraines of Bull Lake glaciation, is redder where developed on material derived from red shale and has a Cca horizon where developed on gravel derived from Paleozoic or Mesozoic limestone. These deposits are either trenched by younger fan gravels or occur as remnants projecting above younger fan gravels.

INTERGLACIAL OBSIDIAN SAND

The surface of the West Yellowstone basin southeast of the terminal moraines of the Bull Lake glaciation is a broad plain which rises to the southeast at a gradient of 15 to 20 feet per mile and is underlain by sand composed predominantly of obsidian. Drillers' logs of water wells show that this deposit is from 40 to 100 feet thick (average 80 to 90 feet) and rests on the lake beds retained behind the terminal moraines of Bull Lake age. Downstream from the moraines the sand occurs in lenses along former courses of the Madison River and its South Fork. Such a lens about 5 feet thick in fan gravel 25 feet above the pre-Bull Lake red clay was found in borings made at Hebgen Dam during construction.

Exposures in the plain show that the sand occurs in beds 4 to 12 inches thick and displays crossbedding and channel-and-fill structure typical of shallow aggrading streams.

The material ranges in texture from fine-grained sand to fine gravel and contains thin beds of clean very fine grained sand and silt and a few thin pebble bands. The sand is very angular to subangular, glossy, and sharp edged. According to Warren Hamilton (written communication, 1959) it consists predominantly of glassy rhyolite—obsidian, perlite, and pitchstone—sugary welded tuff particles, and lesser amounts of sanidine, pumice, quartz, oligoclase feldspar, and lithoidal rhyolite. The fine-grained sand or silt layers are ¼ to 4 inches thick and are composed of very clean, sharp angular grains, about two-thirds of which are of mineral matter—dominantly sanidine with subordinate amounts of oligoclase feldspar and quartz—and one-third are of clear silicic glass sand. A few colonial diatoms were noted. Hamilton suggests that much of this material could have been derived directly from the upper parts of obsidian flows in Yellowstone National Park, which contain abundant sand-sized matter. It could also have been derived through breakdown of the obsidian, which is very brittle, in the course of transport by streams.

The upper few feet of the sand deposit contains more gravel than at depth, and gravel is especially conspicuous at the surface of the plain along the course of the Madison River. It is composed of stones as much as 3 inches long that are rounded and have duller surfaces than those in the material beneath. The smaller pebbles are mostly of obsidian, and the sand component is like that in the underlying beds. The larger pebbles are mostly of lithoidal rhyolite and glassy tuff—some welded—like that on the north side of the canyon of the Madison River in Yellowstone National Park. These rocks disaggregate as a result of weathering in place into subrounded rubble much like the pebbles.

The obsidian-sand plain is mantled with eolian silt like that on the moraines of the Bull Lake glaciation, and a soil like that on the moraines is developed through the silt and into the sand. Younger dunes, both stabilized and active, lie on bluffs cut in the sand along the east side of the Madison River. The obsidian sand thus was deposited at some time after recession of the ice but before the soil-forming optimum of the Bull Lake–Pinedale interglaciation. The structure, texture, and especially the lithology of the obsidian sand, however, suggest that it is unrelated to glaciation. Tracing the borders of the plain along the margin of the lake basin showed that in many places along the northeast and southwest sides of the basin the sand tends to backfill into tributary valleys, thus indicating, as does also the lithology, that it received no significant component from these tributaries. Along the south side of the basin, however, between the Madison River and its South Fork, the plain steepens headward into alluvial fans composed of the same material. These fans head not only in the major tributaries but also along the front of the large obsidian flow which overlaps moraines of the Bull Lake glaciation as discussed above. The obsidian sand occurs in channels cut in the moraines and grades headward into an obsidian boulder apron along the front of the flow (pl. 5). The sand plain thus seems to have been constructed shortly after emplacement of the flow in post-Bull Lake time but before the development of the post-Bull Lake soil. At the mouth of the canyon of the Madison River in Yellowstone National Park the sand plain is overlapped by the terminal moraine of the next younger glaciation, the Pinedale, and is entrenched by its associated outwash gravel.

LATE PLEISTOCENE AGE OF THE OBSIDIAN FLOWS

The obsidian flow that overlaps moraines of the Bull Lake glaciation along the south margin of the West Yellowstone basin has a relatively fresh aspect, and its surface is unglaciated. It is capped by eolian silt on which the post-Bull Lake soil is developed. Where one lobe of the flow extended into the canyon of the Madison River in Yellowstone National Park, however, its margin has been deeply eroded by Pinedale ice (fig. 115). Glacial erosion of the margin of the flow also has occurred along the cliffs that border the Firehole River as far south as the Upper Geyser Basin. West of the Lower Geyser Basin a lateral moraine of the Pinedale glaciation rests on the margin of the flow, and a series of melt water channels extends from the moraine across the flow (pl. 5). The obsidian flow and the obsidian-sand plain thus are both younger than lake deposits of the late stade of Bull Lake glaciation and older than the post-Bull Lake soil and the subsequent Pinedale glaciation.

A second obsidian flow with an unglaciated and relatively fresh aspect terminates southwest of Hebgen Lake basin between Reas Pass and Black Canyon. It is more dissected than the flow in the Hebgen basin and is therefore probably somewhat older. It rests unconformably on rhyolites and tuffs of Oligocene age and is overlain by eolian silt on which the post-Bull Lake soil is formed. An obsidian-sand plain like that in the West Yellowstone basin extends westward from a boulder apron at its base, but no direct stratigraphic relations to glacial deposits can be demonstrated. The flow is therefore judged to have formed at some time during or shortly after Bull Lake glaciation.

Examination of aerial photographs reveals that most of the flows at the surface of the Madison Plateau, and the Pitchstone Plateau to the south, are unglaciated except along their east margins where they have been scoured by ice of the Pinedale glaciation. At one point between the headwaters of the Firehole

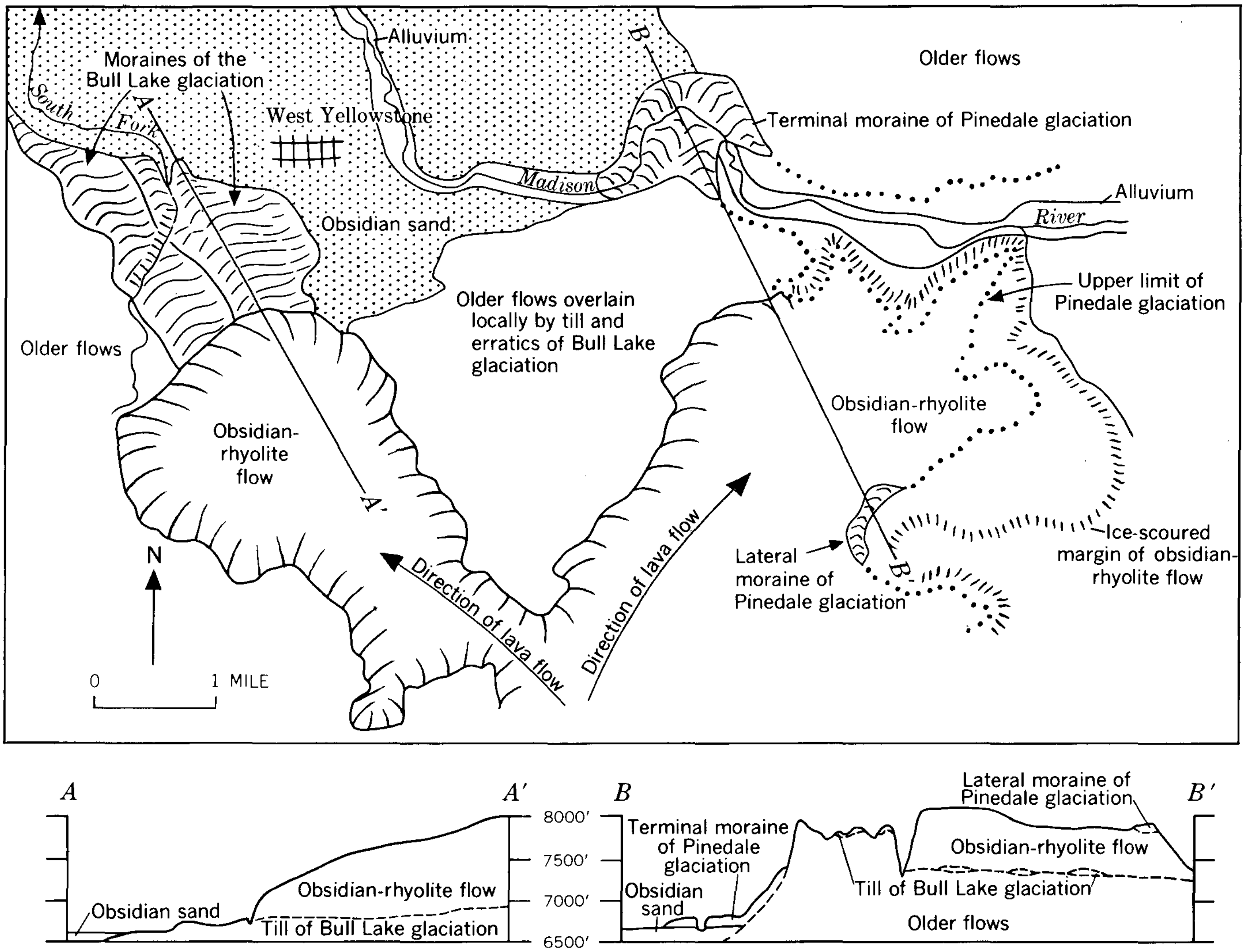

FIGURE 115.—Sketch map and section of West Yellowstone area showing relations of rhyolite-obsidian flow to moraines of the Bull Lake and Pinedale glaciations.

River and Ferris Fork of the Bechler River in the southwestern part of Yellowstone National Park, Pinedale ice crossed the Pitchstone Plateau and flowed down the canyon of the Bechler River, at the mouth of which it deposited a large terminal moraine that extends in a broad arc from Robinson Creek to Falls River (fig. 116). Much more extensive moraines of Bull Lake glaciation have been observed by the writer downstream along Falls River, and extensive deposits of till of a pre-Bull Lake glaciation derived from Yellowstone National Park have been described in the area south of Ashton (Stearns, Bryan, and Crandall, 1939). Aerial photographs show that the southern margin of the flow traversed by Pinedale ice abuts abruptly against, and appears to have overridden, markedly ice-scoured hills in the vicinity of Mountain Ash Creek that lie upstream from moraines of Bull Lake glaciation and beyond the outer limits of Pinedale glaciation. It seems logical to conclude, therefore, that most of the surface flows of the Madison Plateau and Pitchstone Plateau to the south are younger than Bull Lake and older than Pinedale.[12]

PINEDALE GLACIATION

The Pinedale stage of glaciation was named from moraines that enclose Fremont Lake and other large lakes in the vicinity of Pinedale, Wyo., (Blackwelder, 1915). The deposits are younger than the interglacial soil developed on till of Bull Lake age and represent the last major advance of Pleistocene ice in the region.

[12] Examination of aerial photographs of the area southwest of Shoshone Lake shows that a number of small unglaciated flows of fresh appearance overlap part of the area crossed by Pinedale ice along the divide between Shoshone Lake and the canyon of the Bechler River (fig. 116). These flows include Trischman and Douglas Knobs and the relatively flat-topped crest of the divide. Surfaces scoured by Pinedale ice cross the divide to the south of them, and large glacial grooves extend from the base of their steep lobate west margin toward the canyon of the Bechler River. These observations suggest that some volcanic activity may have persisted on the plateaus until after at least the early stade of the Pinedale glaciation.

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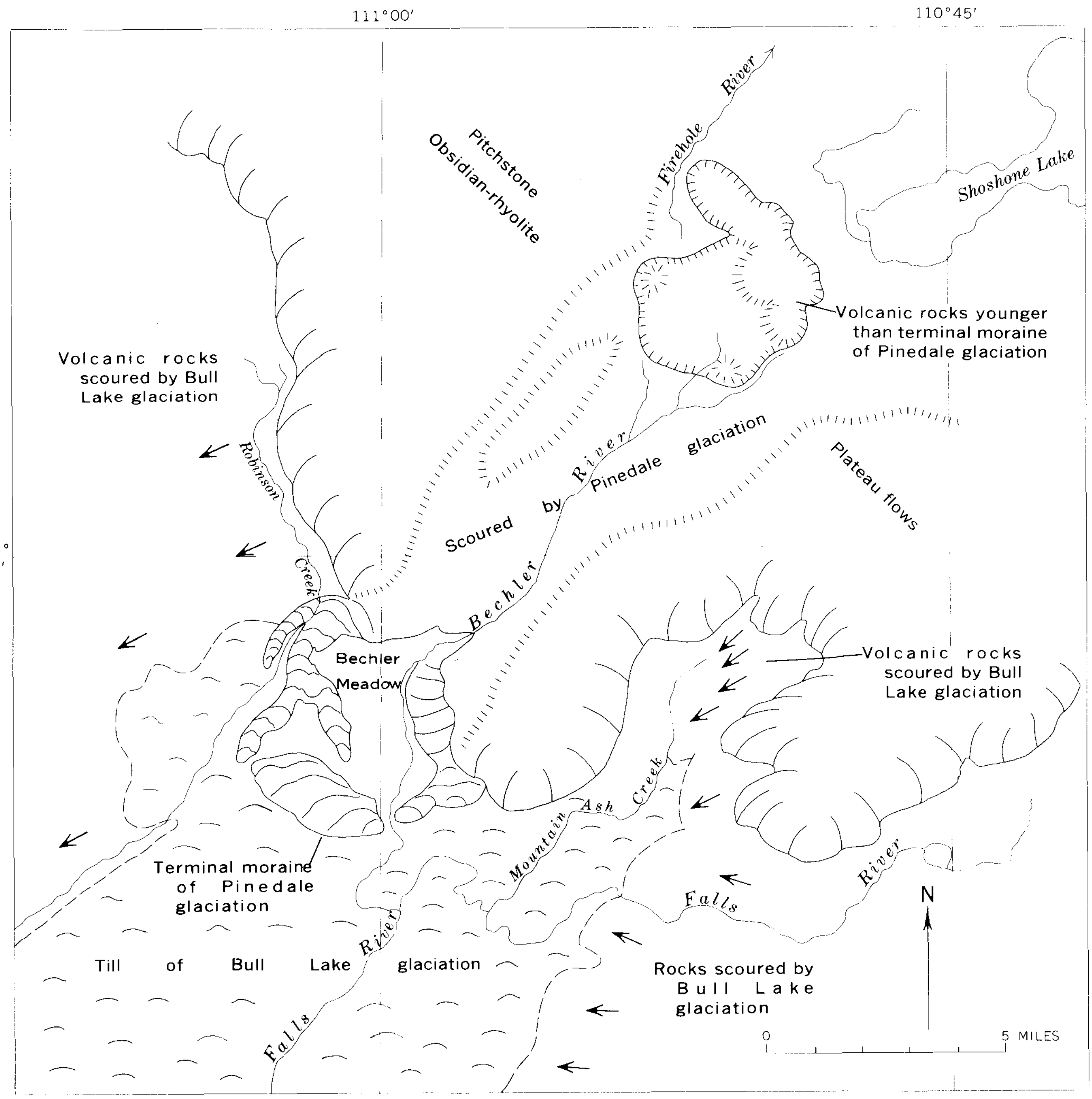

FIGURE 116.—Sketch map drawn from aerial photographs showing relations of obsidian-rhyolite flows of the Pitchstone Plateau to features of the Bull Lake and Pinedale glaciations and the location of volcanics younger than the terminal moraine of Pinedale glaciation.

Pinedale glaciation is thus probably equivalent to the classical Wisconsin glaciation of the midcontinent region. The moraines are commonly more bouldery than those of the Bull Lake age and are but little dissected. Boulders are fresh, and striations commonly are well preserved. The soil on the deposits is less well developed and distinctively different from that on deposits of the Bull Lake or pre-Bull Lake glaciations.

GLACIATION IN THE WEST YELLOWSTONE BASIN AND WESTERN PART OF YELLOWSTONE NATIONAL PARK

Only the southeast corner of the West Yellowstone basin was invaded by Pinedale ice. The maximum advance of the ice is represented by a large forested terminal moraine consisting of several discontinuous arcuate ridges that lie at the mouth of the canyon of the Madison River near the west edge of Yellowstone National Park. The moraine was assigned a late Wisconsin age by Alden (1953). Its surface is hummocky, littered with boulders, lacks a cover of eolian silt, and, in general, has a much more youthful aspect than moraines of Bull Lake glaciation in the central part of the West Yellowstone basin. Enclosed depressions contain water seasonally, and the slopes are but little dissected except for 40 to 50 feet of downcutting along the Madison River.

Lateral moraines rise steeply from the terminal moraine terminus to the canyon wall: that on the south wall pinches out against bedrock at an altitude of 7,400 feet; that on the north wall pinches out at 7,600 feet, slightly farther upstream.

The till consists of a gray massive fairly loose stony sand and silt. Stones range from pebbles to boulders and are subrounded to subangular. Many display glacial soles, facets, or striations. Most stones are obsidian, although there is a considerable proportion of red and pink lithoidal rhyolite and dacite. The matrix of the till is largely glassy and lithoidal rhyolite, welded tuff, pumice, sanidine, oligoclase feldspar, and quartz.

The soil on the till is of two kinds; in most places it consists of 2 to 4 inches of gray humus over 10 to 18 inches of light-gray (7.5YR 7/1–7/2) silty sand that is soft, friable, nonplastic, and has a weak fine-crumb structure. The pH of the soil ranges from 6.0 to 6.5. The light-gray color is imparted by volcanic ash mixed with the mineral matter. In places the ash is so abundant that it makes up most of the soil. A second kind of soil consists of 2 to 4 inches of humus over 6 to 12 inches of light brownish-gray (10 YR 6/2) nonashy silty sand whose other characteristics are essentially the same as those of the more common ashy phase.

The moraine overlies the obsidian-sand plain described above, on which the post-Bull Lake soil is developed; and outwash from the moraine forms a series of low terraces that now are trenched below the level of the plain along the Madison River. The moraine is thus clearly separated from those of Bull Lake glaciation by interglacial deposits and a soil.

The course of the Pinedale glacier was confined to the canyon of the Madison River by the preceding interglacial obsidian flows. Ice essentially filled the canyon to an altitude of about 7,600 feet from the terminal moraine upstream to Madison Junction. From this point, the surface of the ice rose gradually to join an icecap whose borders lay at about 8,000 feet in the area between the Gibbon and Firehole Rivers and may have attained a maximum altitude of about 8,600 feet on the plateau to the east.

As is common in the Rocky Mountains, the Pinedale glaciation embraces three sets of moraines that mark successive halts or minor advances of the ice, here called stades. In the West Yellowstone region, the early stade is represented by the moraine at the mouth of the Madison Canyon in Yellowstone National Park. A younger moraine, which marks the middle stade, lies about 2 miles upstream from the mouth of the canyon, just east of the foot of Mount Jackson. It forms an arcuate ridge, 40 to 80 feet high, that extends along the south side of the canyon and grades into a broad terminal moraine which encloses a bedrock knob about 450 feet high. An area of hummocky ground moraine extends from the terminal moraine upstream for nearly a mile and is trenched about 15 feet by the modern flood plain. Kame deposits related to the recession of the ice from this moraine are abundant on the floor of the valley near Madison Junction, and ground moraine overlies the low hills along the south side of the Gibbon River as far as the east edge of the map area (pl. 5). In this distance the surface of the glacier rose gradually from 6,000 feet at its terminus to 7,250 feet at the mouth of Gibbon Canyon, which was blocked by the north lateral moraine of the glacier. Deposits of the lake thus formed occur along the upper part of Gibbon Canyon and underlie Gibbon Meadows. They consist of sand, fine gravel, and layers of clay derived from hydrothermally altered rhyolite bedrock in the vicinity of hot springs in Gibbon and Norris Geyser Basins.

A second moraine of the middle stade of Pinedale glaciation was formed by a glacial tongue which descended Nez Perce Creek from cap ice on the plateau to the east and extended into the lower Geyser Basin above Firehole Canyon.

The moraine is rather complex. A low ridge of till extends from the valley of Nez Perce Creek into the Lower Geyser Basin southeast of Porcupine Hills. East of Paint Pot Geyser the ridge is underlain largely by kame sand that merges into a distinct ridge of till which crosses the northern sector of the loop road to Firehole Lake and is cut out by the hot-spring terrace below that lake. The ridge reappears southeast of the southern sector of the loop road, passes east of Great Fountain Geyser, and at one time blocked the valley of White Creek. Just west of White Creek the moraine rests on an old geyser cone. Here, the lower 10 to 15 feet of the till—total thickness about 40 feet—is thoroughly altered to red clay which grades upward into fresh till in the upper third of the deposit. To the west the moraine merges with a kame terrace underlain by sand and fine gravel and extends northwest across Firehole River to Fountain Freight Road. Goose Lake, Feather Lake, and Lower Basin Lake are on this terrace. West of the crossing of Fountain Freight Road and the Firehole River, a distinct morainal ridge bounds the large westward loop of the river, and ground moraine extends northward along the base of the cliffs to Firehole Canyon, down which the ice extended an unknown distance.

The Upper Geyser Basin was not invaded by ice during the middle stade of Pinedale glaciation.

The third, or late, stade of Pinedale glaciation was restricted to the Yellowstone Lake basin, where its outer margins lay at an altitude of about 8,000 feet. Although melt water from the ice drained through several outlets, the lowest point reached by the glacier itself was in Hayden Valley at an altitude of 7,800 feet. Here, a long narrow tongue of ice along the Yellowstone River blocked tributaries from the west including Elk Antler Creek, Trout Creek, and Alum Creek, along whose low-lying valleys lake clays accumulated.[13]

South of Yellowstone Lake, where the Continental Divide is crossed by the South Entrance Road, a distinct end moraine of the late stade of Pinedale glaciation extends to the east at an altitude of about 8,000 feet. It is bounded successively by a conspicuous outwash channel followed by Riddle Lake trail, by Riddle Lake, and by Solution Creek.

LOCAL VALLEY GLACIERS

In Pinedale time, as during Bull Lake glaciation, many small glaciers formed in the canyons of the Madison Range adjacent to the West Yellowstone basin. North of the basin, along the West Fork of Beaver Creek, a low moraine overlaps rock ledges along the north wall of the valley at an altitude of 7,640 feet and marks the early stade of Pinedale glaciation. The moraine is composed of fresh, very bouldery till derived wholly from Precambrian crystalline rocks. The soil on the till consists of a thin humus horizon overlying a pale brownish-gray B horizon 6 to 10 inches thick. Locally light-gray volcanic ash forms a thin layer between the A and B horizons or is mixed with the material of the B horizon.

Downstream from this moraine the valley is bordered by steep convex slopes underlain by alluvial gravel, slopewash, and solifluction debris derived in large measure from older till plastered on the valley wall. These deposits bear a soil like that on till of Bull Lake age lower in the valley and are dissected from 10 to 30 feet by gullies containing younger alluvial fans and talus rubble derived from cliffs along the upper slopes.

A second and more prominent moraine, believed to mark the middle stade of Pinedale glaciation, terminates at an altitude of 8,850 feet. Its characteristics are the same as those of the lower moraine, and its lateral moraines can be traced upslope into the cirque to the south which contains Dome Lake and into the large compound cirque to the west that contains Avalanche Lake. The moraine was thus formed by the confluence of glaciers from both major tributaries of the West Fork of Beaver Creek.

A third moraine, believed to mark the late stade of Pinedale glaciation, lies at an altitude of 8,880 feet below the compound cirque containing Avalanche Lake. It has a very abrupt front and in places overlies bedrock ledges. A broad outwash plain extends along its lower margin and downstream across till of the middle stade. The B horizon of the soil on both tills is brownish yellow (10YR 6/6), which is somewhat darker than at lower altitudes. Locally, light-gray post-Pinedale volcanic ash forms a thin layer between the A and B horizons of the soil or is mixed with material of the B horizon.

A similar succession of moraines of the Pinedale glaciation occurs along Watkins Creek and its tributaries west of Hebgen Lake (pl. 5).

East of Hebgen Lake, in the basin at the head of Red Canyon, is a large hummocky lobe of debris about a mile in length which might be considered a glacial deposit. The writer, however, believes it to be either a landslide or solifluction deposit because the cliffs at its head do not have the oversteepened appearance common to cirques of Pinedale glaciation, and because the rim of the basin is at an altitude of only 8,900 feet, which is considerably lower than the rims of cirques known to have contained Pinedale glaciers.

OTHER DEPOSITS ASSOCIATED WITH PINEDALE GLACIATION

Deposits other than till that are associated with Pinedale glaciation include outwash and alluvial gravelly sand, kame sand and gravel, alluvial fan gravel, and alluvial obsidian sand.

OUTWASH AND ALLUVIAL GRAVELLY SAND

A broad outwash apron composed mostly of lithoidal rhyolite and obsidian gravel in an obsidian–sand matrix borders the moraine of the early stade Pinedale

[13] These lake claps lie between higher hummocky remnants of outwash and overlying lake silts of an earlier stade of Pinedale glaciation, when, according to Howard (1937, p. 92–100), a glacier from the drainage of the Lamar River backfilled the Grand Canyon of the Yellowstone River and extended south into the north edge of Hayden Meadows where it dammed a lake in which silt was deposited to "elevations higher than 8,000 feet."

glaciation at the mouth of the canyon of the Madison River. It overlaps the older obsidian-sand plain and thins westward to a featheredge. Along the river, however, the outwash lies in a broad channel that trenches the obsidian-sand plain and extends northwest along the river to a point where it is submerged by Hebgen Lake. North of West Yellowstone the channel is marked by numerous cut-and-fill meander scars bordered by low terraces whose pattern suggests gradual degradation to a level only 5 to 10 feet above the flood plain. The gravelly outwash is only a few feet thick and rests disconformably on obsidian sand. It bears a soil like that on the moraines of Pinedale glaciation, which includes in many places from 4 to 6 inches of post-Pinedale volcanic ash.

Deposits of alluvial gravelly sand, similarly trenched beneath the level of the obsidian-sand plain and terraced a few feet above the modern flood plain, occur locally along South Fork of the Madison River and Cougar Creek. Along the canyon of the Madison River north of Hebgen Lake are deposits of coarse gravel and cobbles composed predominantly of Precambrian crystalline and Paleozoic sedimentary rock types with lesser amounts of rhyolite and obsidian. The sand in these deposits is composed mostly of quartz, feldspar, and mafic minerals derived from the Precambrian rocks of the Madison Range.

KAME SAND AND GRAVEL

Ice-contact deposits of sand and gravel are abundant along the canyon of the Madison River between the moraines of the early and middle stades of Pinedale glaciation and near Madison Junction in Yellowstone National Park. An extensive kame field associated with the early stade is crossed by the Mesa Road east of Firehole Canyon, and kame deposits outline part of the terminus of the ice of the middle stade in the Lower Geyser Basin. All these deposits consist predominantly of obsidian and rhyolite sand with minor amounts of gravel. Their local hydrothermal alteration by hot spring activity is widespread in association with both active and extinct spring deposits.

Along the Gibbon River near Norris Junction, for example, rhyolite bedrock that has been completely altered to a greasy clay by still-active hot springs underlies as much as 40 feet of kame sand and gravel. Cobbles and boulders of rhyolite in the basal part of the gravel rest disconformably on the clayey bedrock but are themselves only altered to a crumbly kaolinitic grus, a situation which suggests that hot-spring activity may have preceded Pinedale glaciation by a sufficient time to permit these differences in alteration. The next higher 10 to 15 feet of the kame deposits is altered and leached to a whitish kaolinitic quartz sand in which lenses and irregular permeable zones are stained bright orange, and cavities and fractures are locally lined with sulfur crystals. This alteration dies out upward along streaks in the upper 15 to 20 feet of the deposits which otherwise appear essentially fresh. That large quantities of melt water flowed from the icecap on the Yellowstone Plateau throughout Pinedale glaciation is suggested by the fact that kames and kame terraces are more widespread than morainal forms. The highest observed ice-contact deposits of Pinedale age are at an altitude of 8,400 feet on both the east and west sides of the ridge along the road between Old Faithful and West Thumb. The ridge attains an altitude of 8,610 feet northeast of Shoshone Lake.

ALLUVIAL-FAN GRAVEL

Numerous alluvial fans were constructed around the margin of the West Yellowstone basin during Pinedale glaciation. Some fans truncate, or head in terrace gravels that truncate, moraines of Bull Lake age. Many are graded to alluvial-gravel terraces that lie below the level of the post-Bull Lake obsidian-sand plain. Still others partly fill gullies trenched beneath the level of alluvial fans formed during Bull Lake glaciation. All bear the post-Pinedale soil profile.

The fan deposits are typically of angular to subrounded gravel derived from local bedrock sources upstream. On the west side of the basin, alluvium in the large fan at the mouth of Denny Creek is mostly derived from Precambrian crystalline rocks, Paleozoic limestone and sandstone, rhyolite, and basalt that crop out in the region of Targhee Pass. The fan at the foot of Watkins Creek is predominantly of Precambrian crystalline rocks and Cambrian sandstone from Coffin Mountain. On the east side of the basin, the fan at the foot of Grayling Creek contains a wide variety of Precambrian crystalline, Paleozoic and Mesozoic sedimentary, and younger rhyolitic rock types; that at Red Canyon is mainly derived from Paleozoic red shaly limestone. Fan deposits along the north margin of the lake are mostly of Paleozoic limestone.

ALLUVIAL OBSIDIAN SAND

Alluvium just above the modern flood plain in the Lower Geyser Basin and in many other basins in the southwestern part of Yellowstone National Park is predominantly of obsidian sand, in places containing a little gravel, which overlies or is trenched into till and outwash deposits of the early and middle stades of Pinedale glaciation. The alluvium forms gently sloping fans that head in streams draining obsidian flows. Many of the fans are swampy and have active hot springs and geyser cones on them. They support vegetation and do not appear to be aggrading at present.

POST-PINEDALE DEPOSITS

Post-Pinedale deposits include volcanic ash of altithermal age and younger rock glaciers, talus flows, talus, and alluvium whose development coincided with two minor stades of Neoglaciation in the Wind River Mountains. Landslides in the form of large blockfalls, slumps, mud flows, debris flows, earth flows, and rock avalanches seem to have occurred throughout post-Pinedale time without significant relation to climatic change.

VOLCANIC ASH

A pale-gray layer of silt-sized volcanic ash rests on, or is intermingled with, the B-horizon of the soil on Pinedale and older deposits. It is most common on alluvial deposits associated with Pinedale glaciation and is overlain in places by younger Recent alluvium. It was not found on any deposits of post-Pinedale origin and therefore is believed to have accumulated during the post-Pinedale interglacial or altithermal interval. In the West Yellowstone basin it attains a thickness of as much as 8 inches of essentially pure ash or as much as 18 inches of ash intermingled with soil material. In Yellowstone National Park, along the road between Gibbon and Norris Geyser Basins, the ash, which overlies till of the early stade of Pinedale glaciation, is 14 inches thick and is essentially pure. It also occurs on till of the middle stade in Madison Canyon and on till of the upper stade at an altitude of 8,800 feet along the West Fork of Beaver Creek north of West Yellowstone basin.

Laboratory examination of the ash was made by Howard A. Powers, who reported as follows:

> The ash consists of particles, all less than 100 microns and most less than 37 microns in diameter, of volcanic glass clouded with a cryptocrystalline mineral, probably feldspar. Equally small euhedral crystals of hypersthene and brown-green hornblende are present, perhaps up to nearly a percent of the total volume. The ash is similar in appearance and in its phenocryst content to some of the very young ash deposits that have been erupted both from Mount Saint Helens and from Glacier Peak, volcanoes in the northern Cascade Range.

ROCK GLACIERS AND TALUS FLOWS

Rock glaciers and talus flows were observed in the high cirques of the Madison Range at the head of the drainage of Beaver Creek and in the vicinity of Coffin Mountain. Three rock glaciers and a series of talus flows in the broad cirque containing Avalanche Lake at the head of the West Fork of Beaver Creek seem to be typical.

One of the rock glaciers terminates at the south shore of Avalanche Lake; a second lies adjacent to the west; and the third lies north of the second in the shelter of a cliffed promontory that separates them (pl. 5). They range from 400 to 800 yards in length. All are characterized by steep fronts 40 to 50 feet high, by arcuate ridges and furrows parallel to the front, and by local enclosed depressions. Blocks at their surfaces range from 1 foot to 15 feet in diameter, which is larger than the blocks in the talus above them. All are of Precambrian crystalline rock, predominantly granite. Fine-grained debris is present in the deposits at depth and is sufficiently abundant at the surface to support a tundra cover in many places and scrub spruce locally. The rock glaciers are not at timberline, however, for trees are present on the crest of the cirque headwall at an altitude of 10,000 feet, some 600 feet above them. A weakly developed soil profile consisting of a few inches of humus and 3 to 6 inches of rust-colored B-horizon is present beneath the tundra-covered areas. In contrast, till of the late stade of Pinedale glaciation that is in front of the rock glaciers bears a soil profile in which the B-horizon is 10 to 16 inches thick. The rock glaciers are therefore considered to postdate that soil-forming optimum and to correlate with the Temple Lake stade of Neoglaciation in the Wind River Mountains (Hack, 1943; Moss, 1951a, 1951b).

The rock glaciers grade headward into tundra-covered talus, but they are clearly separated from modern talus accumulations by irregular depressions that are commonly the site of seasonal snowbanks.

Downslope from the depressions, blocks along the inner edge of the rock glacier are stained yellowish brown and bear no tundra or lichen. In places channels underlain by such stained blocks extend outward across the rock glacier and cut across tundra-covered areas. In at least one place, a secondary bulge of stained and lichen-free blocks lies in front of a snowbank depression into which the fresh modern talus extends. These several features are interpreted to represent an attempt at rejuvenation of the rock glaciers during the post-Temple Lake readvance of the glaciers in the Wind River Mountains in historic time.

Talus flows occur at the foot of cliffs on the south-facing side of the cirque that contains Avalanche Lake and form a series of irregular lobate arcs whose width along the cliffs is much greater than their length. They represent local slow flowage or solifluction at the toe of talus slopes and have fronts from 15 to 20 feet high. The deposits, like the rock glaciers, are mostly tundra covered and grade headward into grassed-over talus slopes on which almost no accumulation is taking place at present. They have a soil profile comparable to that on the rock glaciers and, like them, locally overlap till of the Pinedale glaciation. Therefore, they too are inferred to have formed during the Temple Lake stade of Neoglaciation in the Wind River Mountains. No evidence of subsequent talus flowage was noted on the south-facing side of this cirque.

ALLUVIUM

Recent alluvium under the flood plains of the major streams in and tributary to the West Yellowstone basin is in most instances more fine grained than that of Pinedale age. It consists commonly of humic sandy silt or silty sand with local gravel bands, especially near the base. Although buried humus layers were noted locally, no attempt was made to determine whether the alluvium might be stratigraphically subdivisible. The alluvium is believed to include stream deposits of both the Temple Lake and the historic glaciations. In places the alluvium is trenched by arroyos, especially on fans along the east side of the basin where uplift has been effected by postglacial faulting, including the present quake displacements.

LANDSLIDES

Numerous post-Pinedale landslides and small slumps have occurred in the West Yellowstone basin. As the most significant of these are discussed in other parts of this report, mention will be made here of only one.

Between Grayling Creek and Red Canyon, a large hummocky debris flow has overflowed alluvial fan gravel of Pinedale age and spread out onto the modern flood plain. Its surface is hummocky, displays distinct flow line, and has depressions which contain water. I. J. Witkind (oral communication, 1959) who has studied this deposit, believes that it perhaps first moved some time during Pinedale glaciation and that it has continued to flow intermittently to the present.

SUMMARY OF QUATERNARY HISTORY

The main events of Quaternary history in the region of the West Yellowstone basin may be summarized as follows. It is possible that the area was glaciated during a broad valley stage of erosion before the canyons in the Madison Range were cut, but the evidence is inconclusive. After the canyons were cut, ice of a more clearly defined pre-Bull Lake glaciation invaded the basin from an icecap in Yellowstone National Park. It covered Horse Butte and terminated at some point between the Narrows of Hebgen Lake and Hebgen Dam. The character of the topography over which this glacier advanced across the western part of Yellowstone National Park is unknown, but the major drainage from the ice was probably southwestward to the Snake River. A local valley glacier extended from the headwaters of Beaver Creek into the canyon of the Madison River north of Hebgen Dam.

During the subsequent interglacial interval, lake marl and 100 feet of obsidian sand derived from flows in Yellowstone National Park were deposited on the basin floor, and a thick reddish clayey soil was formed.

An icecap again formed on the Yellowstone Plateau during Bull Lake glaciation, and from it a glacial lobe nearly 13 miles wide flowed northwest across the rim of the plateau into the West Yellowstone basin. This lobe deposited the large moraines at the Narrows of Hebgen Lake but did not cover Horse Butte.

This glaciation consisted of two stades, as marked by two distinct moraines that circumscribe its outer limits. Local glaciers in the Madison Range also display evidence of two, but smaller, stades.

The character of the topography over which the cap ice extended in the western part of Yellowstone National Park cannot be determined owing to a cover of subsequent lava flows, but an ancestral course of the Firehole River is believed to have reached northwest from the Lower Geyser Basin to a confluence with the Gibbon River at about the present mouth of Madison Canyon. At the southwest corner of the basin, the South Fork of the Madison River was blocked by the lateral moraine of the glacier and flowed temporarily to the southwest.

Following recession of the glacier, a lake, in which 20 to 30 feet of silty clay was deposited, developed behind the terminal moraine. Later, during interglacial time, a large obsidian flow poured from the Madison Plateau into the southwest corner of the basin where it overrode the moraines of Bull Lake glaciation. Other arms of the flow extended northeast over the rim of the canyon of the Madison River and east to form the present west walls of the Lower and Upper Geyser Basins. It is probable that most of the surface flows of the Madison Plateau and the Pitchstone Plateau to the south formed during the post-Bull Lake–pre-Pinedale interglacial interval, but volcanic activity may have persisted into Pinedale time on the divide southwest of Shoshone Lake.

During and following the outpouring of lava, 80 to 100 feet of obsidian sand, derived in part through disaggregation of the cooling obsidian flows, was deposited by shallow streams on the basin floor upstream from the Narrows of Hebgen Lake. A loess mantle, which has been accumulating on deposits of Bull Lake glaciation since withdrawal of the ice, is also found on the obsidian flows and on the sand plain. Its source was probably the Snake River plain to the southwest.

Later in interglacial time, a distinctive soil profile developed on this eolian mantle, on exposed deposits of Bull Lake glaciation, and on the obsidian-sand plain.

The Yellowstone Plateau was again the site of an icecap during Pinedale glaciation. Accumulation of the post-Bull Lake obsidian flows, however, had raised the plateau surface sufficiently to confine the ice to the canyon of the Madison River in Yellowstone National Park.

Pinedale glaciation consisted of three stades. During the early stade, ice covered the central part of the plateau west of Yellowstone Lake and abutted the rim of the Madison Plateau west of the geyser basins along the Firehole River. This ice was approximately 600 feet thick over the site of Old Faithful Geyser. A long tongue followed a new course of the Firehole River and carved deeply into the post-Bull Lake obsidian flow along Firehole Canyon and along the Madison River. Melt-water channels also cut across the flow from the surface of ice which lay against the west rim of the Lower Geyser Basin and from a moraine which overlaps the top of the flow at one point. The terminus of the ice was at the mouth of the Madison Canyon at the southeast corner of the West Yellowstone basin.

During the middle stade of Pinedale glaciation the icecap was restricted to an altitude below 8,400 feet in the area west of Yellowstone Lake. A tongue from the icecap extended along the canyon of the Gibbon and Madison Rivers to a point about 3 miles upstream from the terminal moraine of the early stade. Another tongue along Nez Perce Creek terminated in the Lower Geyser Basin. The Upper and Midway Geyser Basins were free of ice, as were also the Gibbon and Norris Geyser Basins; and the Midway and Gibbon Basins were the sites of shallow lakes.

During the late stade the icecap was restricted to the basin of Yellowstone Lake, where its outer margin lay at an altitude of about 8,000 feet. It extended to an altitude of 7,800 feet into Hayden Meadows as a narrow tongue along the Yellowstone River, and to the Continental Divide along the South Entrance Road.

Pinedale glaciation in the local canyons of the Madison Range was parallel to that on the Yellowstone Plateau, but the glaciers were much smaller.

A distinctive soil profile, less strongly developed than those of preceding interglacials has developed in post-Pinedale time. It is locally overlain by, or intermixed with, a layer of volcanic ash that is probably of altithermal age.

Cooling of the climate in postaltithermal time has led to the development of rock glaciers in some of the higher cirques of the Madison Range. These features display two cycles of development which are correlated with two minor stades of Neoglaciation in the cirques in the Wind River Mountains. The older deposits bear a thin, weakly developed soil profile; the younger are fresh and unweathered. Except as influenced by the earthquake, present-day talus is accumulating, for the most part, only in north-facing cirques.

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