THE GEOLOGICAL SURVEY OF WYOMING
Gary B. Glass, State Geologist

EARTHQUAKES AND RELATED GEOLOGIC HAZARDS IN WYOMING

by
James C. Case

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LARAMIE, WYOMING

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Cover photograph - James Goodrich and Mrs. S.W. Nile stand on a new fault scarp across U.S. Highway 191 at sunrise, August 18, 1959, after the Hebgen Lake earthquake. This fault is in Montana, about three miles from the Wyoming-Montana border. (Photograph by William B. Hall.)

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Introduction

An earthquake is generally defined as a sudden motion or trembling in the Earth caused by the abrupt release of slowly accumulated strain. The most common types are caused by tectonic and volcanic forces although they can also result from explosions, cavern collapse and other minor causes not related to slowly accumulated strains. The analysis of the exact causes and mechanisms of earthquakes is extremely complex and will only be addressed in a general manner in this report.

Causes of earthquakes

Faults and elastic rebound theory

In this report the discussion of the causes of earthquakes will be restricted to nonvolcanic deformation of the Earth's crust. Nearly all felt earthquakes occur as a result of abrupt movement along a preexisting or newly created fault. A fault is defined as a fracture or zone of fractures along which there has been displacement of the sides relative to one another parallel to the fracture (**Figure 1**).

Harry F. Reid of Johns Hopkins University proposed a mechanism for failure either along a fault or in unfaulted bedrock. His mechanism is called the elastic-rebound theory of earthquakes (*in* Boore, 1977, p. 68). According to the theory, all rocks are elastic to a degree. A stress applied to a rock will strain (deform) it, but below a certain limit (elastic limit) will not cause permanent deformation. When the stress below that limit is removed, all strains will be instantly and totally recoverable, meaning that the rock will rebound to its original shape. When a rock is strained in such a manner, it stores mechanical energy in much the same way as when a rubber ball is squeezed in your hand. You can feel the ball trying to push back with its stored energy.

When a solid piece of rock is compressed, it deforms in an elastic manner

Figure 1. Vertical displacement due to movement on the active Grand Valley fault. The offset shown is on the east edge of Afton, Wyoming. The Grand Valley fault is present along the eastern edge of Star Valley.

up to its elastic limit, at which point it undergoes permanent plastic deformation until it breaks. This is the basic process through which faults are originally created. When a fault or fracture is already present in a rock body, and a force is applied to the rock masses on both sides of the fault, there is a tendency for stress and strain to build up near the zone of previous failure. When the stress becomes greater than the frictional strength of the fault, failure occurs at the fault's weakest point.

In order to describe faults, an understanding of certain terminology is required (**Figure 2**). The fault strike is a line defined by the intersection of the fault plane and a horizontal surface. A fault is classified on the basis of direction of motion, called slip. In general, there are dip-slip, strike-slip and oblique-slip faults.

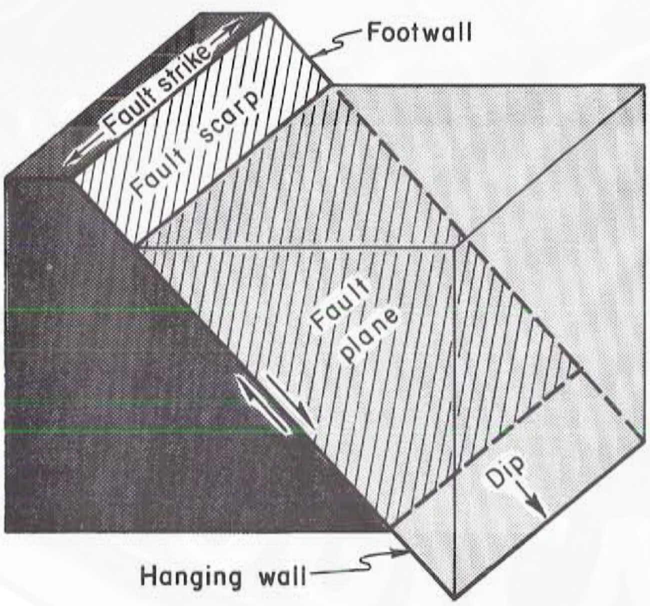


Figure 2. Fault terminology. Single-barbed arrows indicate direction of motion.

There are two types of dip-slip faults in which the component of slip is parallel to the dip of the fault:

Normal fault - A fault in which the hanging wall appears to have moved down relative to the footwall. Normal faults are often caused by either differential vertical or tensional forces (**Figure 3**).

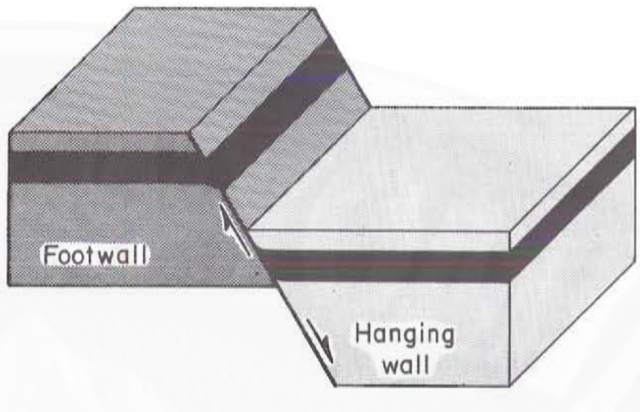


Figure 3. Normal fault. Single-barbed arrows indicate direction of motion.

Reverse or thrust fault - A fault in which the hanging wall appears to have moved up relative to the footwall. These faults are usually caused by compressional forces (**Figure 4**).

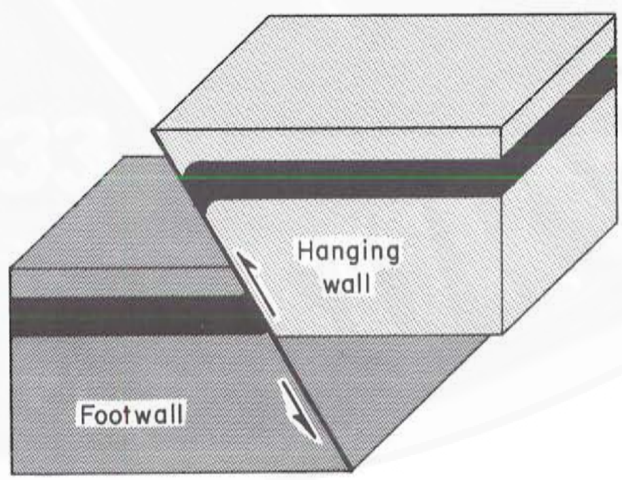


Figure 4. Reverse fault. Single-barbed arrows indicate direction of motion.

There are two types of strike-slip faults in which the component of slip is parallel to the strike of the fault, with no slip in the direction of the fault dip. Strike-slip faults are generally caused by a stress in which there are offset parallel forces acting in opposite directions in a horizontal plane. The two types of strike-slip faults are:

Left-lateral slip fault - A strike-slip fault which, when viewed along the fault strike, shows relative movement of the left block toward the viewer **(Figure 5)**.

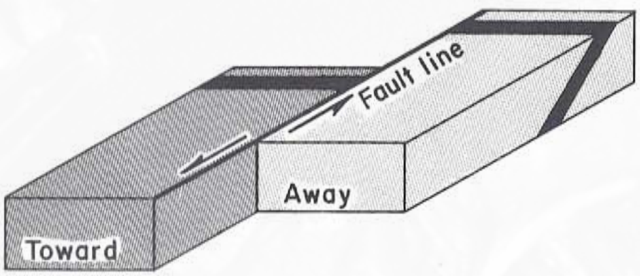


Figure 5. Left-lateral slip fault. Single-barbed arrows indicate direction of motion.

Right-lateral slip fault - A strike-slip fault which, when viewed along the fault strike, shows relative movement of the right block toward the viewer **(Figure 6)**.

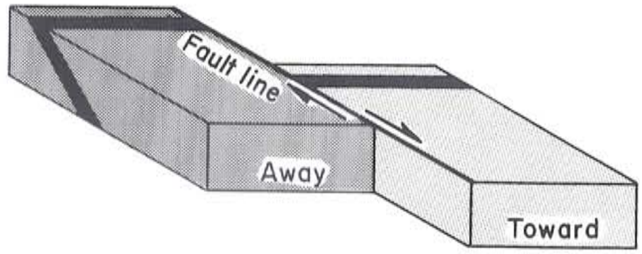


Figure 6. Right-lateral slip fault. Single-barbed arrows indicate direction of motion.

A fault with both a strike-slip and and a dip-slip component is an oblique-slip fault. The net movement with this type of fault is diagonally up or down the fault plane. This type of fault is caused by a combination of forces **(Figure 7)**.

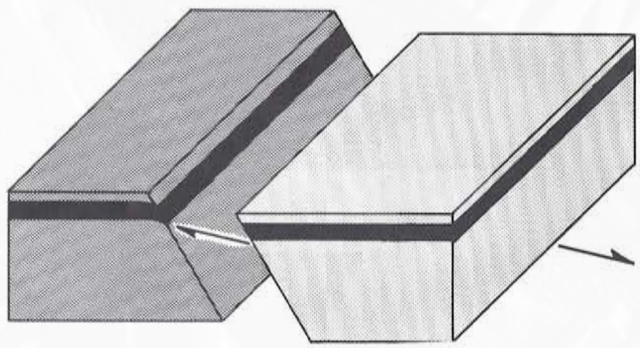


Figure 7. Oblique-slip fault. Single-barbed arrows indicate direction of motion.

All of the faults that have been described are idealized. Often, instead of a single fault there is a zone of faulting in which many of the above types can be found and faults can also show different components of movement with time.

Active faults

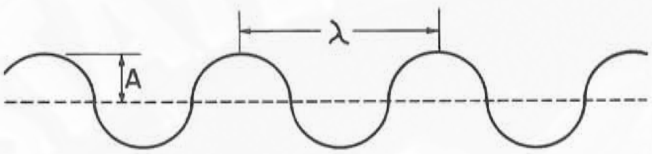
For the purpose of this report, active faults are defined as those that have been active since the beginning of the Quaternary Era, two million years ago. Also considered are faults that have been recurrently active since Early Miocene time (last twenty million years).

Unfortunately, not all active faults are exposed at the surface of the Earth. There are many faults that originate and terminate deep within the Earth. These hidden faults are located and defined by seismic studies or through analysis of information derived from drilling deep oil, gas or water wells. Understandably, not all buried faults have been located.

There is a degree of correlation between active surficial faults and historic earthquake epicenters. **Sheet 1** (back pocket) shows how these two factors correlate in Wyoming. However, there are many areas where suspected active surficial faults have been mapped and there has been no historic activity. There are also many areas where earthquakes have occurred and no active faults have been located at the surface. Only through continuing detailed monitoring of seismically active areas can the buried faults be located.

Seismic waves

When failure or rupture occurs along a fault, the rock bodies on either side of the fault move, releasing the energy that was stored in their strained state. The released energy forms seismic waves (**Figure 8**). An earthquake is felt when the waves reach the surface of the Earth.



λ = wave length

A = amplitude, maximum wave displacement

Frequency = number of wave pulses passing a point per unit time (cycles per second)

Period = 1/ frequency (time for one cycle to pass)

Figure 8. Wave terminology.

There are four basic types of seismic waves that occur during an earthquake. Two of the waves, P (primary) and S (secondary) are body waves that occur within the earth while the other two, Love and Rayleigh, occur on the surface.

P waves - P waves (primary waves) are longitudinal body waves; the direction of particle motion is the same or directly opposite to the direction of wave propagation. These waves are also called compression-dilation (push-pull) waves (**Figure 9**). Their wave propagation is similar to what would be observed if a springlike toy called a "slinky" was held in a straight line and one end of it was alternately pushed and pulled.

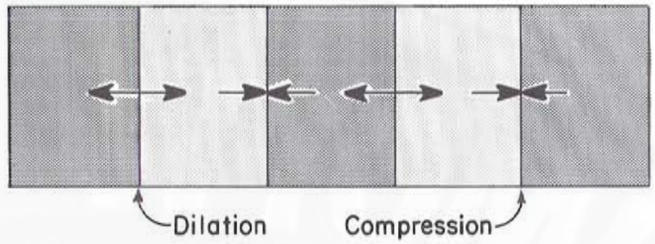


Figure 9. P-wave particle motion.

A pulse made of a compression and an expansion would travel down the "slinky".

When an earthquake occurs, P waves travel outward in all directions from the focus or point of initial rupture of the earthquake. P waves can travel through all portions of the Earth's interior, including the liquid portion of the core. They are the fastest of the waves generated by an earthquake, and can travel through the Earth's crust at approximately 12,000 miles per hour (Van Rose, 1983). The average velocity of P waves through the entire earth is approximately 15,000 miles per hour (Hays, 1981). As a result, they are the first to reach seismic stations. This does not necessarily mean that they are the most destructive. At the surface of the Earth they cause an up-down motion that many structures can tolerate.

S waves - S waves (secondary waves) are transverse waves; the particle motion within the transmitting medium is at right angles to the direction of the wave's propagation (**Figure 10**). S waves are also known as shear waves. To visualize an S wave, tie a "slinky" to a stationary object and give the free end a series of side-to-side jerks. The wave form generated is similar to an S wave.

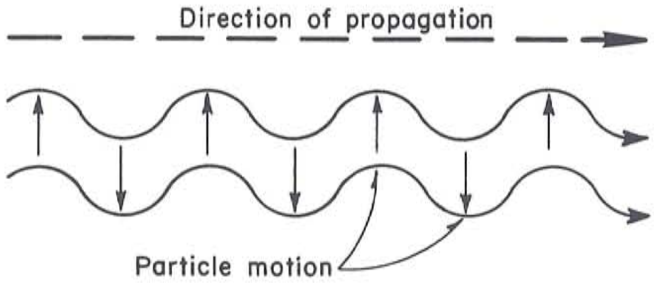


Figure 10. S-wave particle motion.

S waves do not move through liquids, and as a result do not directly penetrate the Earth's core. They are slower than P waves, but can still travel through portions of the Earth's crust at 6,700 miles per hour (Van Rose, 1983). As a result, S waves are the second pulses to be recorded at seismic stations.

S waves are among the most damaging waves because when they reach the surface they cause structures to vibrate from side to side. They are more damaging than P waves for that reason, and also because they have a higher amplitude. Both P waves and S waves are of higher frequency than surface waves, and as a result have a tendency to damage low structures. Harmonic vibrations in low buildings are more likely to be created by the frequencies associated with the P and S waves.

Rayleigh waves - Rayleigh waves are restricted to the Earth's surface layers. The particle motion, always in a vertical plane, is elliptical and retrograde with respect to the direction of propagation. Rayleigh waves are considered to be the principle component of ground roll **(Figure 11)**.

Rayleigh waves and Love waves (see below) transmit the bulk of the energy in shallow earthquakes, and can be very damaging. They are low-frequency waves, in comparison to P and S waves, and are more damaging to high structures than low ones. The frequencies usually observed have tendencies to set up harmonic vibrations in tall buildings. They

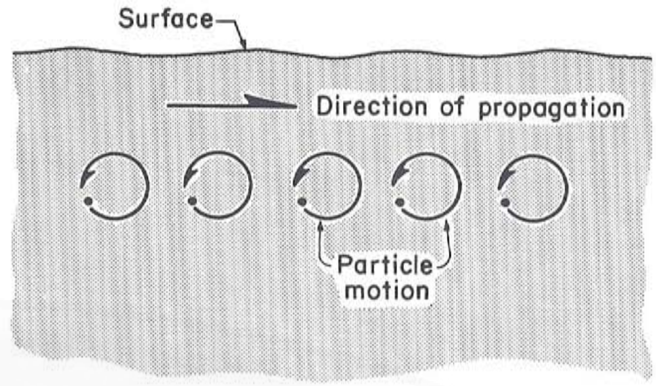


Figure 11. Rayleigh-wave particle motion.

generally have a high amplitude compared to P and S waves. They are slower than body waves, having a velocity that is about nine tenths that of S waves in the same medium.

Love waves - Love waves are also restricted to the Earth's surface layers. They have a particle motion that is horizontal and transverse to the direction of propagation **(Figure 12)**. They are generally observed when there is a low-speed layer overlying a medium in which elastic waves have a higher speed (Dobrin, 1960). As with Rayleigh waves, Love waves have low frequency and high amplitude. These waves also tend to damage tall buildings.

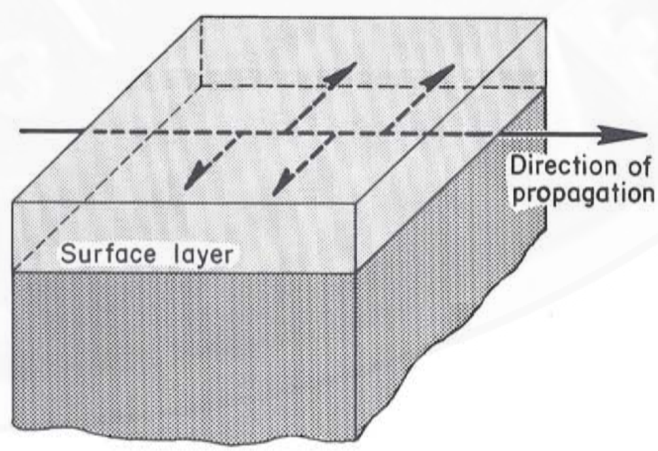


Figure 12. Love-wave particle motion.

When an earthquake occurs at some distance from a seismic recording station, the initial arrival of each of the four wave types can be delineated rather easily. However, both body waves can undergo multiple reflections; P waves can travel through the Earth's core where they are refracted upon entering and leaving; P waves can be transformed to S waves after they reflect at the surface; and S waves that reach the core can be converted to P waves and pass through it (**Figures 13 and 14**). Body waves can create surface waves when they contact the crust or surface of the Earth.

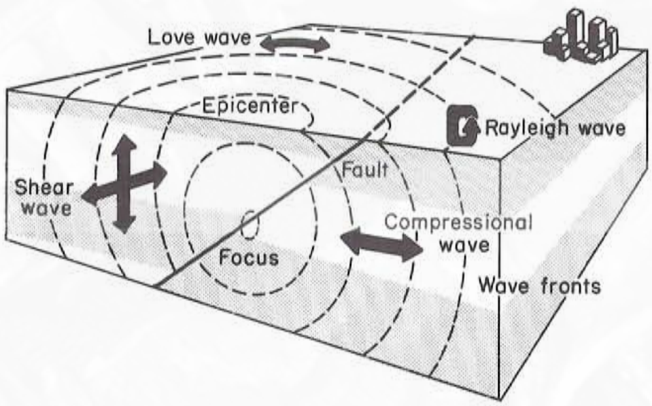


Figure 13. Seismic wave propagation. From Hays (1981).

There are zones between 100° and 140° from the earthquake source where there is no direct penetration of S and P waves. This is due to the reflecting and refracting properties of the Earth's core. These zones are called shadow zones. However, P and S waves can reach this zone through multiple reflections. **Figure 14** shows the simplified paths of seismic waves through and around the Earth.

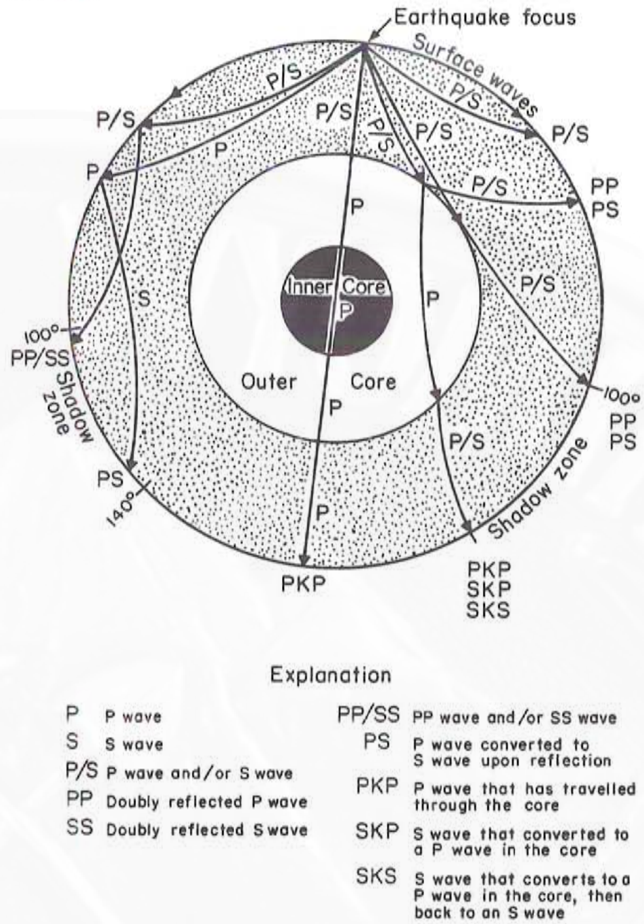


Figure 14. Path of seismic waves through the earth. Modified from Dobrin (1960, page 34) and Bolt (1973).

Earthquake measurements

There are two basic methods of monitoring and measuring various parameters of earthquakes. One method utilizes complex equipment to measure or determine various aspects of seismic waves, including their source area, wave form and the amount of energy they release. The other method utilizes a compilation of personal observations of the effects an earthquake has on the Earth's surface or on specific types of structures at a given locality.

Instrumentally determined earthquake parameters

The instrument that most often comes to mind when discussing an earthquake is the seismograph. It is composed of two major parts, one for detecting and the other for recording. The detecting instruments, or seismometers, are of various types, the most common being a form of pendulum suspended from a weak spring. Measurements of ground motions relative to the pendulum reveal earth motion at a specific site. Another type of detection instrument is a strain seismometer, which measures fluctuations in the length of a rigid rod. Another detection instrument is a long-period vertical seismograph, which measures the fluctuations in gravity caused by very low-frequency waves.

The seismometers relay information to a recording device called a seismogram that plots and records the various types of wave motions, either electronically or on paper attached to a drum. The record produced by the seismogram is called a seismograph. When information has been received from at least three seismic stations, analysis of the seismographs can begin **(Figure 15)**.

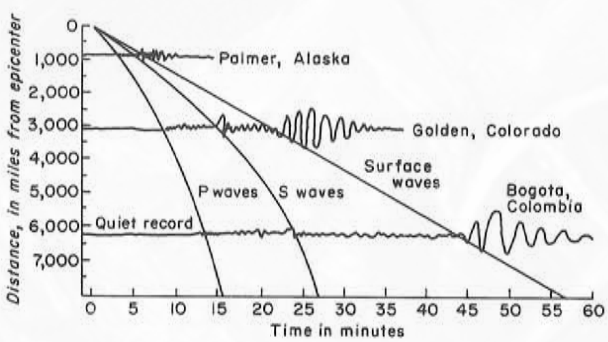


Figure 15. Seismic travel-time curves showing differences in arrival times of P, S and surface waves. From U.S. Geological Survey *Earthquake Information Bulletin, v. 9, no. 4, July-August 1977, p. 15.*

Various parameters of an earthquake are determined from the seismographs. Through a complex form of triangulation both the hypocenter (actual point on a fault where rupture begins) and epicenter (point on the surface of the Earth directly above the hypocenter) can be determined. The dimensions of the ruptured surface, the orientation of the fault, and the amount of slippage that has occurred can also be determined under ideal conditions.

Magnitude

The most commonly referred to earthquake measurement relates to the size of the earthquake and the total energy released. That measurement is the strength, or magnitude. There are one advanced and three basic magnitude scales in use today. They all derive their numerical components in part or entirely from comparing the amplitudes of various seismic waves or seismic wave combinations.

The most famous magnitude scale was developed by Charles Richter in 1935 (*in* Richter, 1958). It was designed to be used where recording stations are within 600 kilometers (370 miles) of an epicenter and hypocenters are fairly shallow. The original Richter magnitude scale represents local magnitude (M_L).

Richter's method utilizes a series of identical Wood-Anderson torsional seismometers to measure the relative or trace amplitude of ground motion caused by seismic waves. Specific wave forms are not delineated with this procedure although most of the waves that are measured have periods (1/frequency) from 0.1 to 2 seconds.

In developing his scale, Richter compared the common logarithms (base 10) of the maximum trace amplitudes of ground motions caused by earthquakes of various sizes to those generated by a standard earthquake. The standard earthquake is defined as one that would produce a trace deflection of 0.001 millimeter on a torsion seismograph that is located at a distance of 100 kilometers (62 miles) from an earthquake source.

In order to have a reference point on the magnitude scale, a standard earthquake was assigned a magnitude of 0. Other points on the scale are defined by the formula $M_L = \text{Log } A - \text{Log } A_o$ **(Figure 16A)**. As M_L represents a number that is a common logarithm (base 10), an earthquake with an M_L of 2 has a maximum amplitude of ground motion that is 10 times larger than an earthquake with an

A. **Local magnitude (Richter scale)** (Stover, 1985)

$M_L = \log A - \log A_o$.

A = Recorded trace amplitude.

A_o = Amplitude of standard earthquake.

Seismometer within 600 kilometers of epicenter.

Focus depth less than 70 kilometers

B. **Surface wave magnitude** (Stover, 1985; Spence, 1977)

$M_S = \log (A/T) + 1.66 \log D + 3.3$.

A = Maximum vertical surface-wave trace amplitude in microns.

T = Period in second ($18 \leq T \leq 22$).

D = Distance from epicenter to recording station in geocentric degrees.

Earthquake must be in a distance range of 20° to 160° geocentric.

Focus depth less than 50 kilometers.

Valid to $M_S = 8.6$.

C. **Body wave magnitude** (Stover, 1985)

$M_B = \log (A/T) + Q (D,h)$.

A = Ground motion amplitude in microns, taken from P-wave group (not necessarily the maximum).

T = period in seconds ($0.1 \leq T \leq 3.0$).

Q = Function of distance between hypocenter and station (D) and depth (h) where $D \geq 5°$ (geocentric).

Valid to $M_B = 6.8$ (Spence, 1977).

D. **Seismic moment magnitude** (Kanamori, 1980; Spence, 1977)

$M_W = (\log M_o/1.5) - 10.7$.

$M_o = \mu S < d >$ = seismic moment.

S = Area of fault.

$< d >$ = Average displacement.

μ = Shear strength of faulted rock.

Seismic moment information is incorporated into an extension of existing magnitude scale.

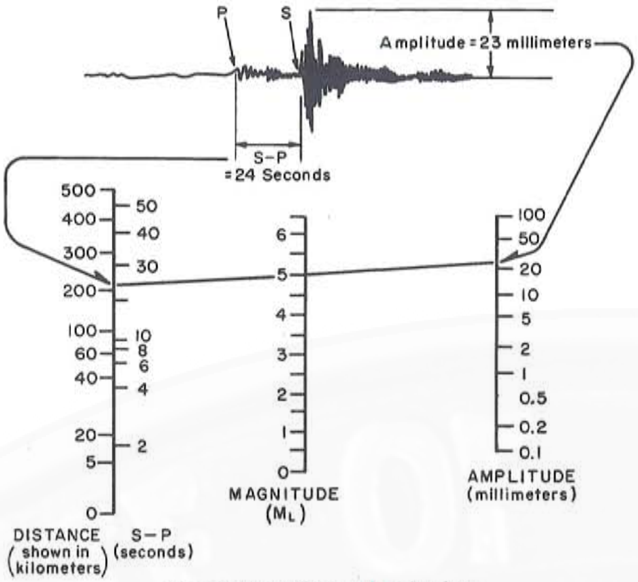
Valid for all earthquakes especially useful for those with magnitudes greater than 7.0.

Figure 16. Methods of calculating earthquake magnitude.

M_L of 1, and 100 times larger than one with an M_L of 0. That is not to say that 10 or 100 times more energy was released. Although the complex relationship between magnitude and earthquake energy is not exact, it has been estimated that with a magnitude increase of one step the associated seismic energy increases about 32 times. **Figure 17** shows how to calculate the Richter (local) magnitude of an earthquake.

It is confusing to many people that an earthquake can have a negative magnitude. Earthquakes with a maximum trace amplitude of ground motion greater than that of a standard earthquake have a magnitude greater than 0 while those with a lesser amplitude would have a negative magnitude. The positive and negative values are for comparison purposes only. As with other magnitude scales, the M_L system is open ended, in that it has no theoretical minimum or maximum limits (no measured earthquake has exceeded a magnitude of 8.9).

Due to the limitations of the M_L system, Richter and Beno Gutenburg developed scales to measure the magnitudes of earthquakes at distant locations and at depths of up to 700 kilometers (435 miles). The scales are based upon the trace amplitude of both surface waves and body waves. Gutenburg realized that for earthquakes at great distances from a recording station, seismic surface waves with a period of 20 seconds are often dominant. His surface wave magnitude scale (M_S) was based upon comparing the trace amplitude of surface waves with periods between 18 and 22 seconds. The formula used is $M_S = \text{Log}(A/T)+1.66 \text{ Log } D+3.3$ (**Figure 16B**). Restrictions on the formula other than those already mentioned are that the earthquake must be in the distance range of 20° to 160° (geocentric), and that depth corrections



Procedure for calculating the local magnitude, M_L:

1. Measure the distance to the focus using the time interval between the secondary (S) and the primary (P) waves (S—P=24 seconds).
2. Measure the height of the maximum wave motion on the seismogram (23 millimeters).
3. Place a straight edge between appropriate points on the distance (left) and amplitude (right) scales to obtain M_L=5.0.

Figure 17. Calculating the Richter magnitude of an earthquake. Modified from Bolt (1978).

have to be made for hypocenter depths greater than 50 kilometers (30 miles) (Stover, 1985).

Gutenburg utilized seismic body waves (P and S) to define another magnitude scale. The formula used by the U.S. Geological Survey is M_B = Log (A/T) + Q (D,h) (**Figure 16C**).

Upon examining the various scales, it becomes obvious that more than one magnitude can be assigned to a single earthquake, depending on the nearness of the hypocenter to recording stations, the depth of the hypocenter and the period of the seismic waves. However, due to refinements and variations of the basic formulas, all types of magnitudes that are computed are fairly comparable with Richter's original M_L concept.

There are certain inconsistencies in the theoretical basis of portions of the magnitude-scale approach to defining large earthquakes. When faults have a very long rupture, seismic waves from the closer end of a fault can reach a seismogram before those from the farther end (Boore, 1977). In addition, when long faults move, seismic waves that are generated can have very low frequencies and extremely large wave lengths. Both the M_S and M_B scales have limitations on the frequencies or periods that can be utilized. As a result, many of the wave forms generated along long faults or large earthquakes cannot be used with the M_B and M_S scales. The measured magnitude, utilizing those scales, may represent only a portion of the energy released by the rupture. The size of the earthquake may be much larger than the magnitude calculation infers. As a result, seismologists are turning to seismic moment in order to more accurately represent the size of large earthquakes.

The method used to calculate the seismic moment ensures that the energy emitted from the entire fault is measured. The formula used to define seismic moment is $M_O = \mu S \langle d \rangle$, where S is the area of the fault, <d> is the average displacement on the fault, and μ is the shear strength of the faulted rock (Spence, 1977). An extension of the standard Richter magnitude scale has been developed by incorporating seismic moment (Kanamori, 1978), M_W=(Log M_O/1.5)-10.7 utilizing the original magnitude scale (**Figure 16D**). The 1964 Alaskan earthquake had an M_S of 8.3. This has been advanced to M_W of 9.2 after incorporating seismic moment calculations.

Parameters determined from personal observations

Although many earthquake parameters are defined instrumentally, personal observations can be valuable. In fact, before the advent of modern instruments, reports on the effects of seismic shaking were the only source of data. Even today, there are portions of the world that are so far removed from seismograph stations that accurate measurements of small earthquakes cannot be made. There are two parameters that are based upon personal observation: one is the intensity or degree of shaking and the other is the sound generated by an earthquake.

Intensity

Intensity is a qualitative measure of the degree of shaking an earthquake imparts on people, structures and the ground. A series of intensity scales have been developed that group earthquake effects into various scale values. The most widely used scale was introduced in Italy by G. Mercalli (1902) and modified by H.O. Wood and Frank Neumann (1931) in California. The scale that resulted is called the modified Mercalli scale of 1931. A shortened version of the scale was printed in the September-October, 1974 issue of the *Earthquake Information Bulletin* (v. 6, No. 5) **(Table 1)**.

For a single earthquake, intensities can vary depending upon the distance from the epicenter. At the edge of the felt zone an intensity of I might be reported, while the intensity at the epicenter would be much higher. Isoseismal maps that are generated from reported intensities can provide valuable information on earthquakes and their effects.

Comparisons of magnitude and intensity

Magnitude and intensity scales have been confused by the general public for years. It is hoped that this report will help to alleviate some of the confusion. Magnitudes are instrumentally determined parameters that address the energy released by an earthquake, and intensities are qualitative estimates of the degree of ground shaking.

There have been some attempts to approximately and indirectly compare magnitudes to maximum expected intensities for earthquakes. These comparisons could only be made with any degree of accuracy if the earth was everywhere of uniform composition with equal layers of bedrock, equal thicknesses and types of surficial deposits, similar fault types and hypocenters at similar depths. In reality, alluvial valleys and soft or unconsolidated sediments will magnify ground shaking and register higher intensities than adjacent areas on solid rock (Anderson, 1971).

Sound of earthquakes

Some earthquakes produce sounds that have been reported by people near the epicenters. In conjunction with the 5.0 magnitude and greater earthquakes that occurred in Wyoming in 1984, many people reported sounds comparable to thunder, passing trains or heavy equipment.

David Hill of the U.S. Geological Survey (Hill, 1976) offered a possible explanation for the sounds. He suggested that some sounds are due to nearly vertical-incident P waves coupling directly into the atmosphere, with the ground in the vicinity acting as a giant loudspeaker driven by the P waves.

There have been reports of hearing an earthquake before feeling it. In these cases the P waves are not strong enough to be felt, but they are strong enough to activate the "loudspeaker". What is felt are the later S waves. As expected, various rock types can produce different sounds. Crystalline basement rock near the surface produces sharp shot-like sounds or booms. If a thick sequence of sedimentary rock is at the surface, the sounds may be soft rumbles.

Table 1. Modified Mercalli intensity scale of 1931 (abridged).

I.	Not felt except by a very few under especially favorable circumstances.
II.	Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
III.	Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing vehicles may rock slightly. Vibration like passing of truck. Duration estimated.
IV.	Felt indoors by many during the day, outdoors by few. At night, some persons are awakened. Dishes, windows and doors are disturbed; walls make cracking sound. Sensation like heavy truck striking building; standing vehicles are rocked noticeably.
V.	Felt by nearly everyone; many awakened. Some dishes, windows, etc. broken; a few instances of cracked plaster; unstable objects overturned. Disturbance of trees, poles and other tall objects sometimes noticed. Pendulum clocks may stop.
VI.	Felt by all; many persons are frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster or damaged chimneys. Damage slight.
VII.	Everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate damage in well-built ordinary structures; considerable damage in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving vehicles.
VIII.	Damage slight in specially designed structures; considerable damage in ordinary substantial buildings, with partial collapse; extensive damage in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments and walls. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well water levels(?). Disturbs persons driving vehicles.
IX.	Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; extensive damage in substantial buildings, with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken.
X.	Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable along river banks and steep slopes. Shifted sand and mud. Water splashed (slopped) over banks.
XI.	Few, if any masonry structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipe lines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.
XII.	Damage total. Waves seen on ground surfaces. Lines of sight and level distorted. Objects are thrown upward into the air.

Earthquake potential of Wyoming

This report presents selected information that is useful in visualizing the past, present and suspected future effects and source areas of earthquakes in Wyoming. The sources of information are historic epicenter maps, seismic risk maps and maps showing probabilistic estimates of maximum expected acceleration in rock.

Historic epicenter maps

Knowledge of where earthquakes have occurred in the past is necessary before any type of prediction can be made about the future. With historic knowledge, a state or country can more accurately be divided into zones reflecting levels of past activity. An earthquake epicenter map for Wyoming has been generated based upon information supplied by the National Earthquake Information Service (U.S. Geological Survey), the Solar Terrestrial Data Center (National Oceanic and

Atmospheric Administration) and from unpublished listings at the Geological Survey of Wyoming (**Sheet 1**, back pocket).

The historic earthquake record in Wyoming begins in 1871 and continues to the present day. Most of the measurements of earthquake size were derived from intensity scales until the mid-1960s. Since then, magnitude expressions dominant. A few earthquakes with epicenters outside of Wyoming were large enough to affect extensive portions of the State.

Seismic risk maps

Seismic risk maps (**Figure 18**) break study areas into zones of maximum expected damage in an attempt to show the relative earthquake hazards to people and structures. The maximum expected damage is based upon tectonics of the area, the distribution and amount of energy released by historic earthquakes, and an estimate of the amount and type of damage caused by previous earthquakes.

Seismic risk maps have certain limitations:

1. In many cases, the historic record is not well documented, requiring an estimate of past activity. In many sections of the country there are no records of earthquakes over 200 years old.
2. Seismic risk maps show zones of maximum expected damage, not earthquake frequency. Several intensity VI earthquakes may be as damaging as one of intensity VII.
3. The lines drawn on the maps to separate zones are misleading. Actually, most of the boundaries are areas of gradual transition.
4. Future earthquakes may not occur where they have in the past, especially when compared to the limited historic record. Similarly, future earthquakes may be larger than the historic maximum size in a specific area.

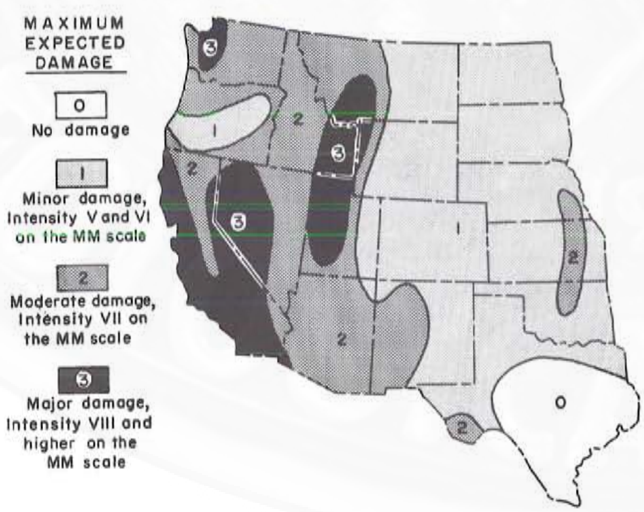


Figure 18. Seismic risk map of the western United States (map is based on damages associated with historic earthquakes). MM = Modified Mercalli intensity scale. Modified from Algermissen (1969).

Maps showing probabilistic estimates of maximum acceleration in rock

As stated in the previous section, seismic risk maps have many limitations. One of the most important is that earthquake occurrence and size frequencies are not incorporated. In order to address this problem, S.T. Algermissen, D.M. Perkins, and others at the U.S. Geological Survey incorporated probabilities of ground motion occurrences into their maps and descriptions.

Their process utilized not only historic seismicity data but also information on seismic source areas (faults) in order to predict future motions. Through a complex process described in Algermissen and others (1982), they produced preliminary maps on the maximum expected horizontal accelerations in rock for 10-, 50- and 250-year periods at a 90 percent probability level of nonexceedance. In other words, there is a 90 percent chance that the maximum expected horizontal accelerations presented on their maps will not be exceeded during the 10-, 50- and 250-year periods under consideration, or a 10 percent chance they will be exceeded in those time periods. This equates to exceedance return periods of 95-, 475- and 2,373-years for the 10-, 50- and 250-year maps **(Figures 19, 20, 21).** The maps do not present information on the last exceedance occurrence.

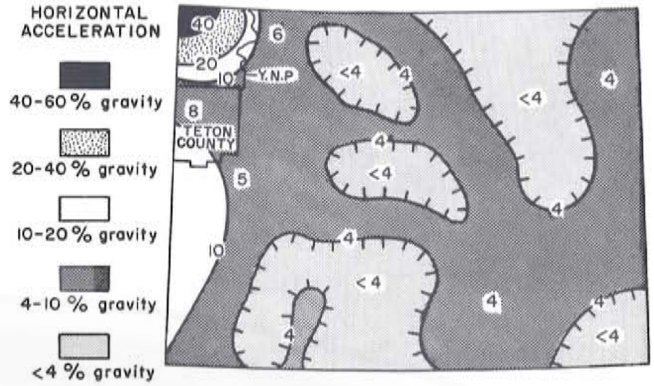


Figure 20. Preliminary map of horizontal acceleration (expressed as a percent of gravity) in rock, with 90 percent probability of not being exceeded in 50 years. Note: statistically the horizontal accelerations will be exceeded every 475 years. Modified from Algermissen and others (1982). Y.N.P. = Yellowstone National Park.

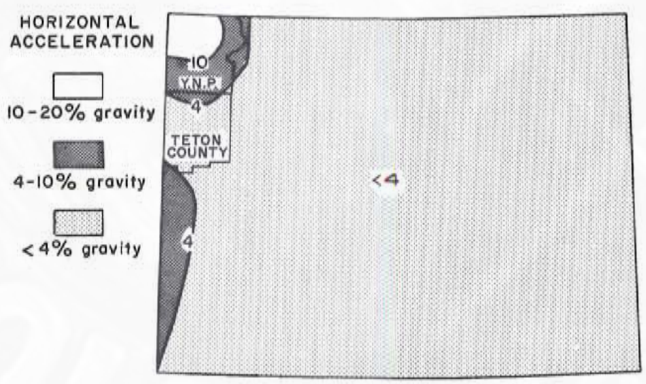


Figure 19. Preliminary map of horizontal acceleration (expressed as a percent of gravity) in rock, with 90 percent probability of not being exceeded in 10 years. Note: statistically the horizontal accelerations will be exceeded every 95 years. Modified from Algermissen and others (1982). Y.N.P. = Yellowstone National Park.

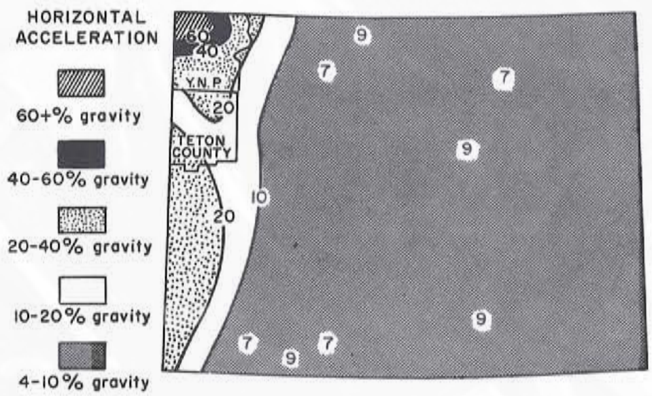


Figure 21. Preliminary map of horizontal acceleration (expressed as a percent of gravity) in rock, with 90 percent probability of not being exceeded in 250 years. Note: statistically the horizontal accelerations will be exceeded every 2,373 years. Modified from Algermissen and others (1982). Y.N.P. = Yellowstone National Park.

There are fairly accurate comparisons that can be made between some intensities and ground accelerations expressed as fractions of g, the acceleration due to gravity at the Earth's surface (32 feet/second/second) **(Table 2).** Buildings in earthquake hazard zones are often designed to withstand at least 0.1g of acceleration.

Table 2 (opposite). Comparisons of intensities and ground motion accelerations. Ground motion accelerations from F. Neumann (1954).

Intensity	Ground acceleration as percent of g	
	Average	Range
II	0.23%	0.1 - .5%
III	0.31%	0.1 - 0.8%
IV	0.93%	0.2 - 4.6%
V	1.33%	0.2 - 7.5%
VI	4.0 %	0.5 - 17.5%
VII	6.7 %	1.8 - 14 %
VIII	17.2 %	5.1 - 35 %
IX	25 %	----------

Historic earthquakes

Wyoming earthquakes generated more public interest in 1984 and 1985 than in the last 25 years. In large part, this was due to their locations, magnitudes and intensities. The intensity V earthquakes that occurred west of Gillette, in May and September of 1984, were among the largest earthquakes in that part of the State in the last 100 years. The October, 1984, magnitude 5.5 earthquake in the northern portion of Albany County was one of the largest quakes ever recorded in the eastern half of the State. The November, 1984, quake (intensity VI) near Atlantic City was also one of the largest earthquakes in that part of the State in recent times **(Sheet 1,** back pocket). Persons who are not aware of Wyoming's historic seismic record may wonder if inactive areas are suddenly reactivating. The historic record, however, illustrates that all of the regions of the State that had an earthquake in 1984 have had a few and in some cases many earthquakes in the past.

Selected earthquakes in seven areas of the State are discussed below, emphasizing areas that are less active than Yellowstone National Park, Jackson Hole and the Overthrust Belt.

Powder River Basin

Very little damage has been attributed to earthquakes in the Powder River Basin. In October, 1922, an intensity IV tremor that originated north of Casper was felt all the way to Sheridan (*Laramie Boomerang*, October 26, 1922; *Casper Daily Tribune*, October 26, 1922; and *Sheridan Post*, October 26, 1922). In Casper, it was thought that there had been an underground gas explosion. In Sheridan, houses shook and dishes rattled. Although the event was described as an intensity IV earthquake, it could have been larger.

In October of 1954, an intensity IV earthquake was reported between Douglas and Wheatland. Although there was no damage, houses were shaken and people were awakened. Train traffic in the area was halted until it was determined that the tracks were not damaged.

In 1964, there was a series of quakes in the Lusk area. In August, people attending a concert in the new Lusk High School thought the furnace blew up when a tremor occurred (*Casper Star-Tribune*, August 23, 1964). That earthquake had a magnitude of 4.5 and a maximum intensity of V. It was felt within a 50-mile radius.

In 1976, a 4.8-magnitude quake occurred approximately 15 miles south of the 1984 tremors near Gillette. In 1967, there was a 4.2-magnitude earthquake located about 42 miles south of Gillette. There have been five earthquakes with magnitudes ranging from 4.0 to 5.1 within 50 miles of Gillette since 1967.

Casper area

Two of the earliest recorded earthquakes in Wyoming occurred near Casper (Mokler, 1923). The first was on June 25, 1894, and had an estimated maximum intensity of V. In residences on Casper Mountain, dishes rattled to the floor and people were thrown from their beds. Water in the Platte River changed from fairly clear to reddish and became thick with mud. On November 14, 1897, there was an intensity VII quake. A person sitting in a chair was thrown to the floor. Guests in the Grand Central Hotel ran into the streets without bothering to dress. The hotel was considerably damaged, with cracks in the structure up to four inches wide. There have been eight intensity IV or greater and three intensity V or greater earthquakes within 40 miles of Casper since 1894.

Northern Albany County

In 1984, the 5.5 magnitude earthquake in northern Albany County created much concern. At least seven tremors have been recorded within 30 miles of the 1984 site since 1947. A February, 1983, quake of 4.0 magnitude and maximum intensity IV was centered only a few miles south of the 1984 tremors. This earlier earthquake was felt as far away as Laramie although there was no damage (*Laramie Daily Boomerang*, February 15, 1983). In 1947, an intensity V earthquake was centered approximately 25 miles to the northeast.

Laramie area

The Laramie area has hosted a number of earthquakes. In 1931 and 1935, slightly felt tremors rattled dishes (*Laramie Republican - Boomerang*, September 21, 1931; and November 11, 1935). In January, 1954, an intensity V earthquake in the area was thought to be an explosion in the downtown area (*Laramie Republican and Boomerang*, January 21, 1954). The newspaper reported that "stories were getting more colorful by the hour", which is exactly what happened with the next earthquake account. In 1955, an intensity IV earthquake occurred near Woods Landing (*Laramie Republican and Boomerang*, May 23, 1955). Reflecting the fears of the time, one person thought that an atomic bomb had dropped on Denver, and she was preparing to accept the evacuees. Also during that quake, a group of fishermen camping near Woods Landing reported that they were rolled around in their tent. The earthquake was not felt in Laramie. Six quakes of intensity IV or greater have occurred within 30 miles of Laramie since 1931.

Bighorn Mountains-Bighorn Basin

There have been several earthquakes reported on the fringes of the Bighorn Basin. A number were located near Ten Sleep. Most notable was one that occurred in 1925. It was of intensity V and was accompanied by a great roar. Some of the cabins in the area reportedly shook like aspen leaves in the wind. The quake was felt all the way to Sheridan (*Sheridan Post-Enterprise*, November 18, 1925).

The Thermopolis area has also experienced a number of earthquakes. There have been approximately eight earthquakes of intensity IV or greater within a 25-mile radius of Thermopolis since 1928. Most of the quakes caused no more damage than the rattling of dishes. The largest quake occurred in 1972. It was of intensity V and magnitude 4.1. The only reported damage was a cracked ceiling at a rest home in Thermopolis. Many residents mistook the quake for a sonic boom (*Casper Star-Tribune*, December 9, 1972).

Lander area

As most residents of the State know, the Lander-Atlantic City area was subjected to an intensity VI earthquake in 1984. However, a few quakes in the past have caused as much concern. In 1934, an intensity V quake was centered northwest of Lander. For a radius of 10 miles around Lander, residents reported that dishes were thrown from cupboards and pictures fell from walls. Residents ran into the streets and expressed considerable concern. Cracks were found in buildings in two business blocks, and the brick chimney of the Fremont County Courthouse was moved two inches away from the building. The shock was felt at Rock Springs and Green River (*Casper Tribune-Herald*, November 25, 1934). There have been 11 quakes within 30 miles of Lander since 1923, seven of which had intensities greater than IV.

Jackson area

Although many earthquakes have occurred in the Jackson area, very few have caused damage in recent times. In 1932, an intensity VI quake did cause some damage; plaster broke off walls, foundations cracked, and because it happened at night people were thrown from their beds. Several people ran from their homes without waiting to grab any clothing. As it was January, none stayed outside for long (*Jackson Hole Courier*, January 28, 1932). There is great potential for strong earthquakes to occur in the Jackson area. The fact that there has been little damage in the last 100 years does not mean that there won't be any in the future.

Yellowstone National Park

Yellowstone National Park is among the most seismically active areas in the State. A complete discussion of the seismicity of the park is beyond the scope of this report. The most notable recent activity was the Hegben Lake earthquake on August 17-18, 1959 **(Cover photograph)**. Although the epicenter was located just outside Wyoming, its disasterous effects were felt well into the State. The quake had a magnitude of 7.1 with a maximum intensity of X. A major landslide was initiated, which damned the Madison River. So much material was displaced so quickly that a great blast of air was generated that blew people about like leaves (Witkind, 1964). Twenty-eight people were killed by that earthquake and its effects. New fault scarps were noted near Hegben Lake, and many geysers were adversely affected in the park. There were 156 aftershocks recorded during the first 21 hours after the main shock. Minor earthquake activity continues in the Hebgen Lake area. Very few years pass without publicized earthquake activity in Yellowstone. The area is one of the greatest seismic hazards in the State.

Geologic hazards related to earthquakes

A number of geologic hazards have components that can be affected by earthquakes. In some cases the relationship is obscure, with special conditions being required for an effect.

Liquefaction

During liquefaction, materials composed primarily of water-saturated sands and silts lose their strength and behave as viscous fluids. (Quicksand is a result of liquefaction of water-saturated fine-grained sands.) Seismic waves passing through saturated materials may compact them, raising the pore water pressure. When the pore water pressure rises to about the pressure of the weight of the overlying materials, liquefaction occurs **(Figure 22)**.

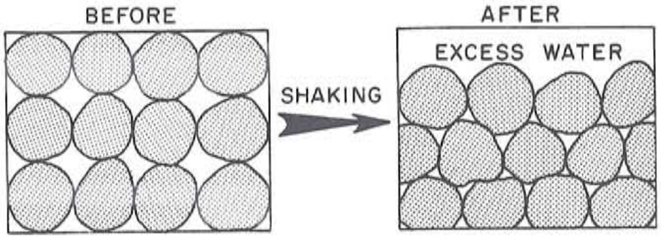


Figure 22. Simplified liquefaction process.

There have been very few documented cases of liquefaction in Wyoming, in part due to the abundance of coarse materials in the saturated alluvium of the State's streams. The most notable occurrence was in the West Yellowstone area during the 1959 Hegben Lake earthquake. Fissures opened in many fields, through which water and sand were ejected (Witkind, 1962). The fissures reached the saturated sands, and because the sands were compacted with increased pore-water pressure, a slurry was injected up the cracks. There appears to be evidence for liquefaction in the Teton Mountain area. The U.S. Bureau of Reclamation has identified features in that area that may have been caused by liquefaction (Dean Ostenaa, personal communication, 1986).

There are some saturated alluvial deposits in the State with components that may be prone to liquefaction. The deposits contain saturated sand and silt that are relatively recent (last 10,000 years), in addition to a water table that is within 30 feet of the surface (Hays, 1981). The liquefaction hazard may be increased during spring runoff; during years of greater than normal precipitation; in areas of extensive irrigation; and in or around seismically active areas.

Assuming a severe earthquake shock, the areas of the State that have deposits with liquefaction-prone components are portions of the Bear River valley, Star Valley, Snake River valley, Yellowstone National Park, Yellowstone River valley and the New Fork River valley **(Figure 23)**.

Slope movements

There are many slope movements and slope movement types in Wyoming. Slumps and earth flows are quite common. Most movements are complex, or combinations of two or more simple types. Slump-earthflows are probably the most common in the State although there have been spectacular slides classified as rock block slides.

Primary causitive factors for slope movement are moisture, lithology, slope angle and seismicity. Generally, moisture is a factor in all landslides with the other factors having varying influence. Determining the influence of earthquakes on landslides is difficult for those landslides older than the historical earthquake record. However, there are a number of landslides that have been either directly or indirectly correlated with ground shaking. The most famous is the lower Gros Ventre landslide that occurred on June 23, 1925 (*Jackson Hole News*, June 26, 1975). Although the rock materials were saturated and dipping, it is felt that the earthquakes that occurred shortly before destabilization of the mass were a contributing factor. The lower Gros Ventre Slide is one of the largest historic landslides in the United States.

The Yellowstone National Park area has had many landslides that can be correlated with seismic activity. A magnitude 6.4 earthquake on June 30, 1975, caused a number of landslides in the park (*Jackson Hole News*, July 3, 1975). Twelve miles of a road between Norris Junction and Madison Junction were closed due to landslides as was the secondary route around the Virginia Cascades, which was determined to be the epicenter location. A series of three- to four-foot deep cracks that were fifteen to twenty feet long were found near the epicenter.

Slope movements have also been correlated with earthquakes at Thermopolis. In October, 1944, an intensity IV earth-

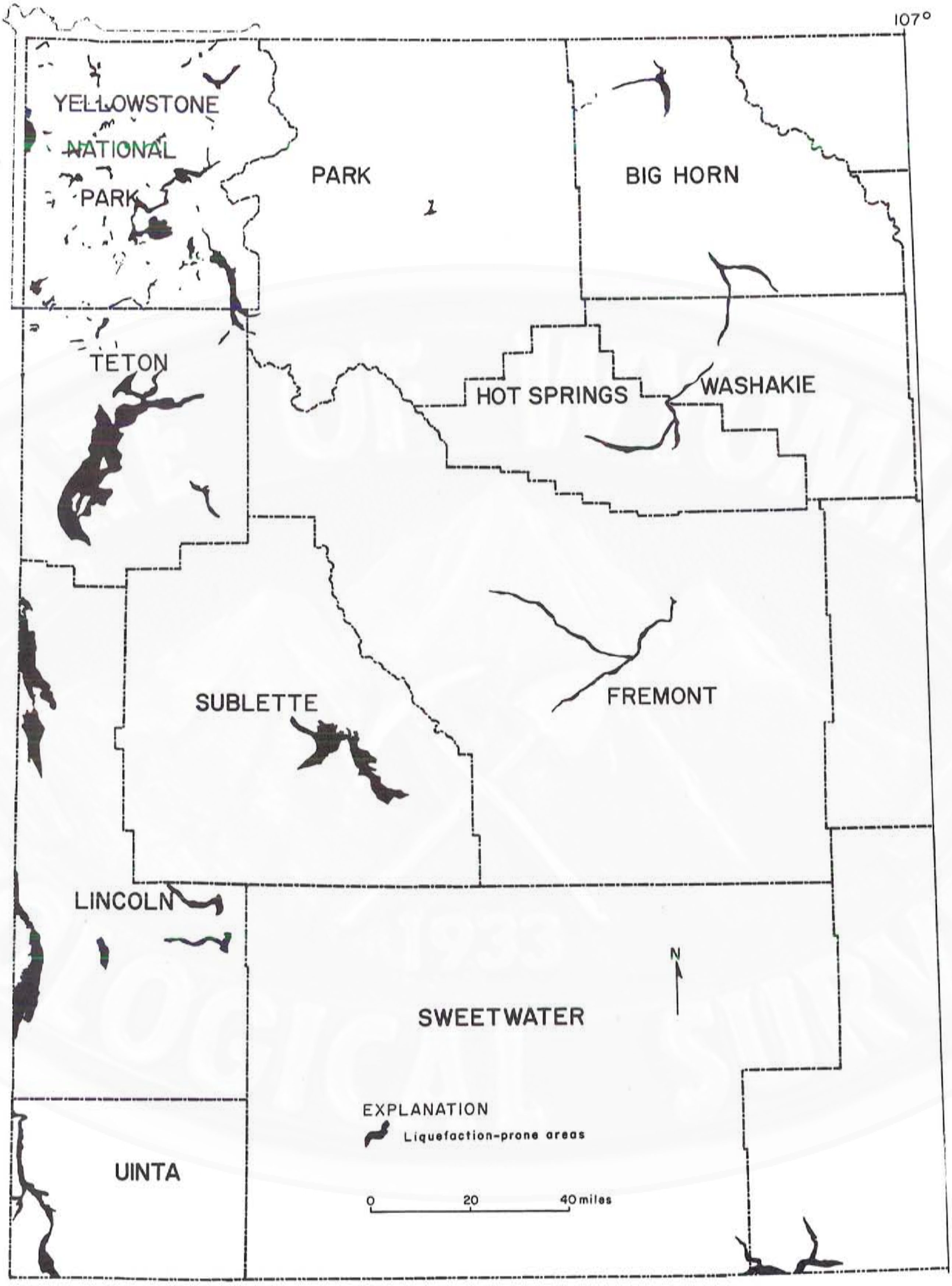


Figure 23. Preliminary map of liquefaction-prone areas, western Wyoming.

quake caused rockfalls in the Wind River Canyon. At Hot Springs State Park, it was reported that there was a "caving of earth on the south rim of the large hot spring in the park" (*Casper Tribune-Herald*, October 13, 1944).

Many older slope movements around the State (Case, 1986) were probably earthquake induced. Unfortunately, there is no accurate method to determine which movements were activated by earthquakes.

Mine subsidence

There are abandoned underground coal mines in many areas of Wyoming. Depending on local bedrock types, type of mine, and size of the mined-out area, subsidence has followed mining in many cases **(Figure 24)**. Based upon old newspaper reports, it appears that there have been a few instances of earthquake-induced mine subsidence in the Rock Springs area.

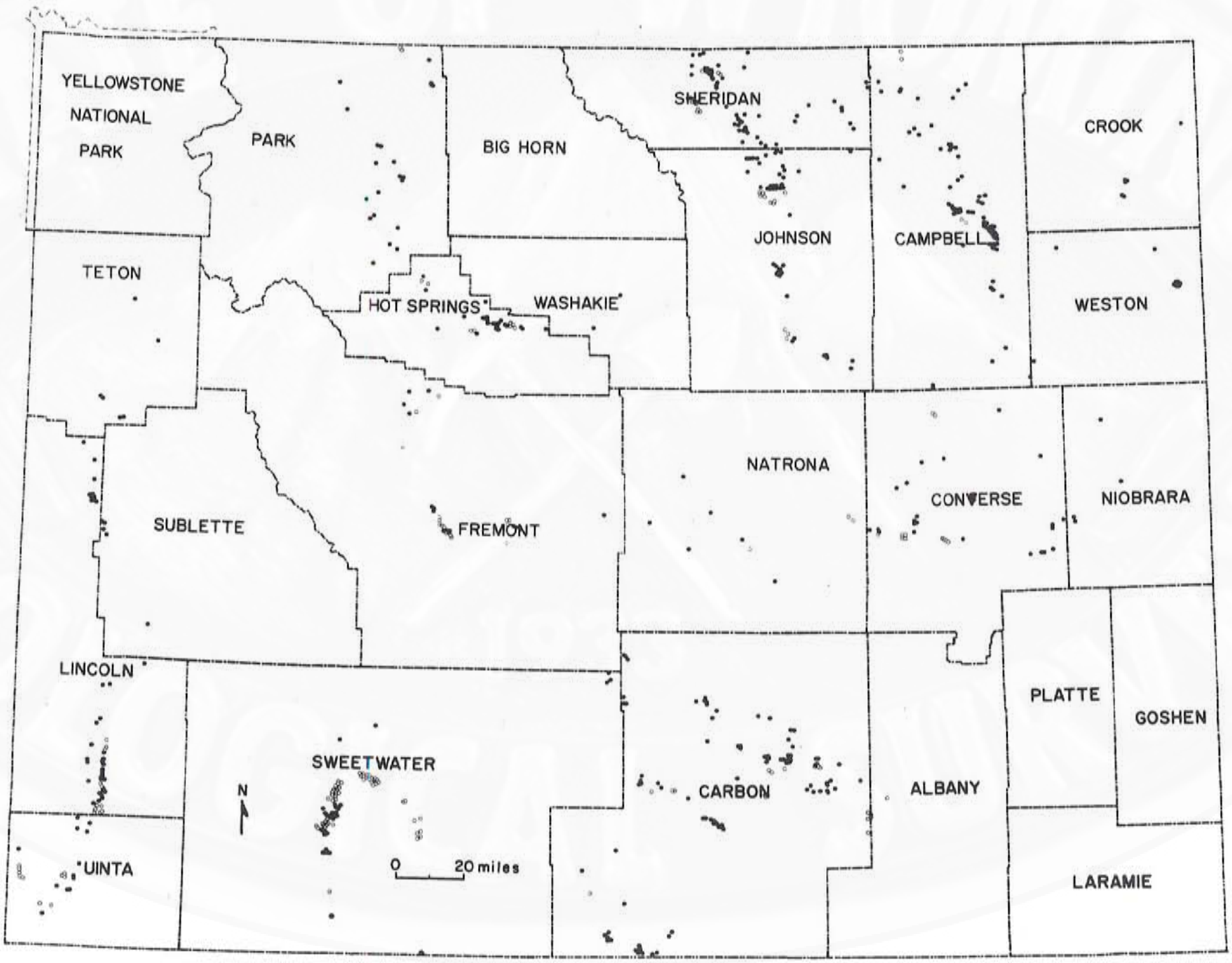


Figure 24. Preliminary map of mined-out areas and mine subsidence in Wyoming. Open circles represent mined-out areas with subsidence; solid circles represent mined-out areas with no known subsidence.

In July, 1910, shocks from a nearby intensity V earthquake partially collapsed the Union Pacific No. 1 mine in Rock Springs (*Cheyenne State Leader*, July 26, 1910). The mine had to be abandoned. Scores of houses were also shaken during the tremor. At the time, some people thought that rapid mine subsidence caused an earthquake-like effect. In fact, the *Cheyenne State Leader* called it an "imitation earthquake". However, it was later determined that an earthquake had actually occurred.

On July 28, 1930, an intensity IV earthquake occurred near Rock Springs (*Casper Daily Tribune*, July 28, 1930). A portion of a mine at Reliance caved in during the disturbance. The tremor was felt at Rock Springs, but there was no appreciable damage. Many residents, however, believed the shaking was again due to settling of the old No. 1 mine.

In September of 1948, there was another intensity IV earthquake centered near Rock Springs that caused no damage. The *Casper Tribune-Herald* (September 27, 1948) attributed the tremor to mine subsidence.

Shrinking-swelling clays

There are many rock formations in the State that have beds of bentonite, or rocks that have a bentonite component (bentonitic shale, claystone). When wetted, the bentonite clay structure expands, resulting in a loss of strength and a slippery exposed surface. Structures located on wet bentonite could slip during a severe earthquake if they are not firmly anchored to a more stable substrate. The clays are fairly stable when dry.

Conclusions

Although all of Wyoming is not as seismically active as Yellowstone National Park and the Overthrust Belt, earthquake hazards in the less active portions of the State should not be ignored. By the same token, earthquake hazards in the State should not be overstated. The earthquakes of 1984 reinforced the historical records for their respective areas.

There are a number of geological hazards that are either caused by earthquakes or are influenced by earthquakes. There should be cautious planning in areas where those hazards are present. Through wise planning, future damage can be minimized in most portions of Wyoming.

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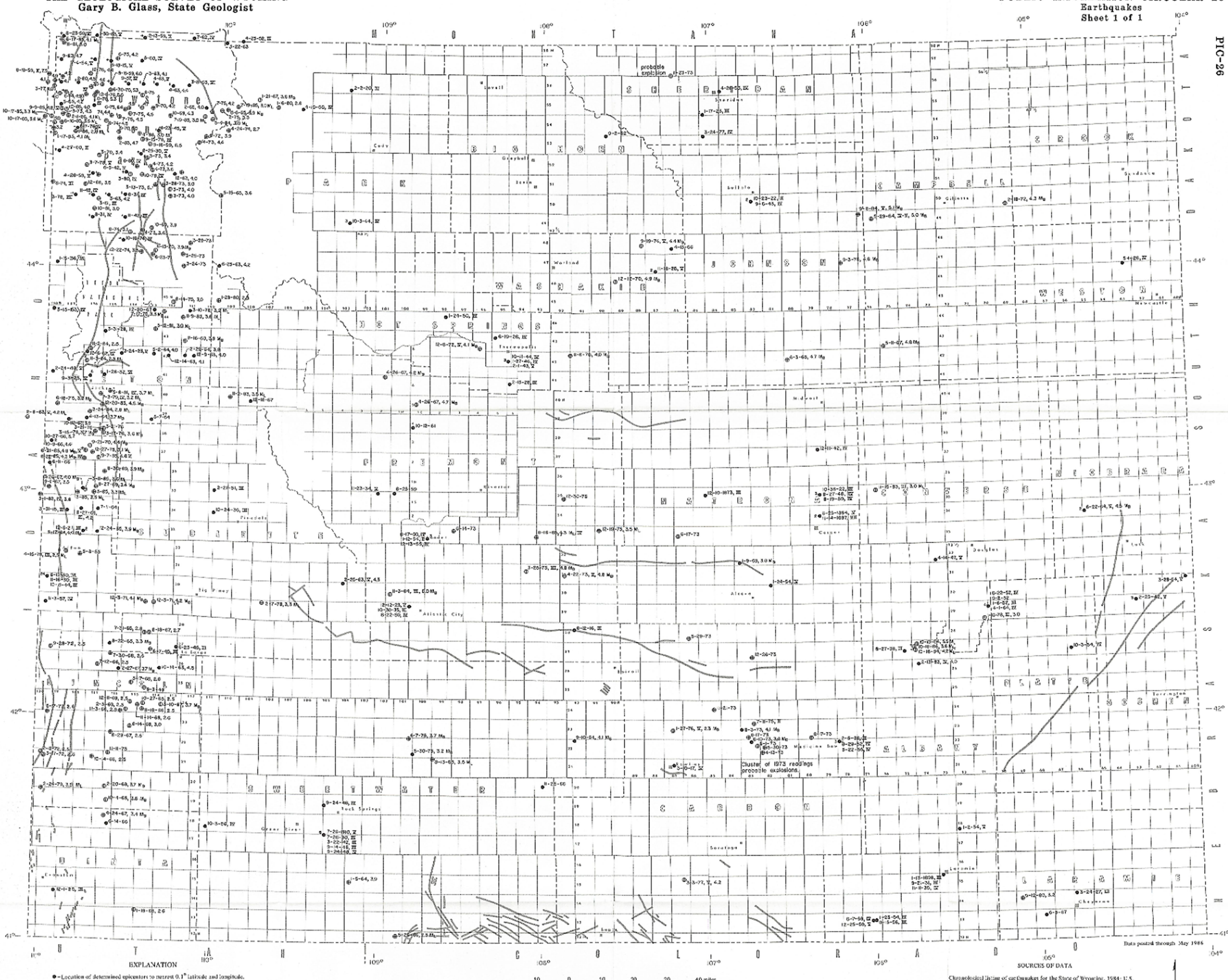
Earthquakes and related geologic hazards in Wyoming

PIC 26

STATE OF WYOMING GEOLOGICAL SURVEY 1933

Geology--Interpreting the past to provide for the future



EXPLANATION

● – Location of determined epicenters to nearest 0.1° latitude and longitude.

⊙ – Location of determined epicenters to nearest 0.01° latitude and longitude.

³● – Number of events at location.

*● – Multiple events in vicinity.

I-XII – Intensities (degree of shaking) derived from Modified Mercalli Intensity Scale of 1931.

2.0-7.5 – Magnitudes (instrumental readings of relative earthquake size).

M_L – Local magnitude (Richter).

M_B – Body wave magnitude.

All magnitudes and intensities shown are highest recorded at location.

Sheet 1 - EARTHQUAKE EPICENTERS AND SUSPECTED ACTIVE FAULTS WITH SURFICIAL EXPRESSION IN WYOMING

Compiled by James C. Case, Cynthia S. Boyd and J.C. Cannia.
Faults compiled by James C. Case.

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