Qualitative Hydrogeologic Model of Thermal Springs (maps)
by Michael John Galloway

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
The occurrence of most thermal springs In Montana Is controlled by the transection of crystalline rock by major tectonic structures. Greater than 50% of the thermal springs In the state are known to Issue from crystalline rock at or near the surface. Crystalline rock in Montana includes Precambrian basement complex and Cretaceous-Tertiary intrusives. Preliminary water chemistry studies suggest nearly all thermal springs in Montana with discharge temperatures greater than 38eC circulate initially within crystalline rock.

Fracture-porosity and permeability decrease rapidly with depth. This suggests intense fracturing, such as shearing and brecciation, is required to permit deep circulation of meteoric water. Eight areas in the Boulder, Pioneer, Tobacco Root, and Idaho batholiths were studied. Structures at 7 of these areas can be associated with major, deep-seated linear features.

This interpretation of geothermal systems in Montana implies the following: (1) each thermal spring may actually be a diffuse system, (2) an accurate estimate of reservoir volume may not be possible, and (3) recharge is available from a large area.

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QUALITATIVE HYDROGEOLOGIC MODEL OF THERMAL SPRINGS IN FRACTURED CRYSTALLINE ROCK

by

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of the requirements for the degree

of

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ABSTRACT

The occurrence of most thermal springs in Montana is controlled by the transection of crystalline rock by major tectonic structures. Greater than 50% of the thermal springs in the state are known to issue from crystalline rock at or near the surface. Crystalline rock in Montana includes Precambrian basement complex and Cretaceous-Tertiary intrusives. Preliminary water chemistry studies suggest nearly all thermal springs in Montana with discharge temperatures greater than 38°C circulate initially within crystalline rock.

Fracture-porosity and permeability decrease rapidly with depth. This suggests intense fracturing, such as shearing and brecciation, is required to permit deep circulation of meteoric water. Eight areas in the Boulder, Pioneer, Tobacco Root, and Idaho batholiths were studied. Structures at 7 of these areas can be associated with major, deep-seated linear features.

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INTRODUCTION

Purpose of Investigation

The purpose of this investigation is to develop a hydrogeologic model for hot springs in fractured crystalline rocks. The model is based on detailed geologic and hydrologic studies of selected thermal springs in southwest Montana and a review of hydrologic characteristics of convective geothermal systems and flow through fractured media. The development of a qualitative model would facilitate the evaluation of geothermal potential in any Montana thermal spring area.

Location

The eight study areas in southwest Montana (Fig. 1) are within the Northern Rocky Mountain physiographic province (Thornbury, 1965). The thermal springs, located in mountainous terrain, discharge from the lower elevations at each locality. Elevations, county, and township and range for each spring are listed in Table 1.

Geologic Setting

The most prominent regional geologic features of southwest Montana are the Late Mesozoic batholiths and associated volcanics and the Late Cenozoic block-faulted sediment-filled valleys. Eaton (1976) places all of southwest Montana within the Cordilleran Thermotectonic Anomaly (CTTA).

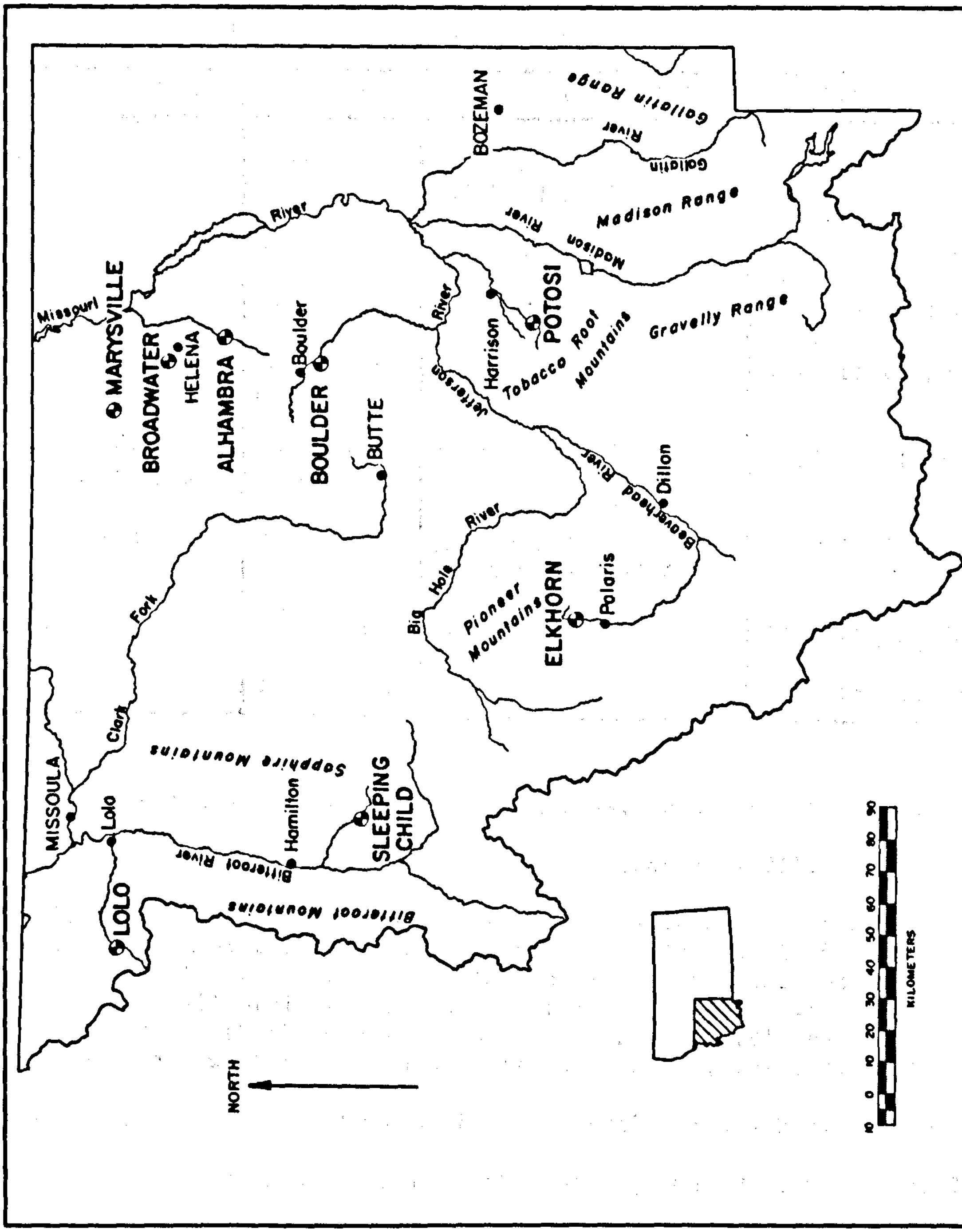


Figure 1. Index Map of Areas Studied. ⊕ shows location of thermal spring or geothermal area studied and ● indicates a town or city.

Table 1. Thermal spring locations and elevations.

THERMAL SPRING	COUNTY	TOWNSHIP & RANGE	ELEVATION
Potosi	Madison	unsurveyed 45° 36'N by 111° 54'S	6100 ft (1860m)
Boulder	Jefferson	T5N R4W	4841 ft (1476m)
Alhambra	Jefferson	T8N R3W	4300 ft (1311m)
Broadwater	Lewis and Clark	T10N R4W	4040 ft (1232m)
Marysville (geo-thermal area)	Lewis and Clark	T12N R6W	5300 ft (1616m)
Sleeping Child	Ravalli	unsurveyed 46° 5'N by 114° 0'W	4800 ft (1463m)
Lolo	Missoula	T11N R23W	4160 ft (1268m)
Elkhorn	Beaverhead	T4S R12W	7240 ft (2207m)

The CTTA is characterized by high heat flow, pronounced Late Cenozoic crustal extension, and elevated topography (Eaton, 1976).

> It extends from the left-lateral strike-slip Garlock fault, southeast California, northward to the Lewis and Clark line of right-lateral strike-slip faults, Idaho-Montana, and from central California to central Utah; its southern half coincides with the Great Basin and its borders (Sierra Nevada and western Colorado plateaus) (Eaton, 1976, p. 850).

Eaton suggests that two-thirds of all thermal springs having base temperatures greater than 90°C in the conterminous U.S. are a result of and controlled by the high heat flow and pervasive Late Cenozoic extensional faulting of the CTTA.

Procedure

Field work was conducted during the summers of 1975 and 1976 plus several days during the winters of the same years. Specific procedures for various portions of this study are presented in the appropriate sections of the text. Except for Marysville, the geology of each area was plotted on 1:21,100 black and white aerial photographs and later transferred to topographic base maps. When suitable scale topographic maps were available, some geologic data were plotted directly on the maps. Plane table base maps were constructed for Boulder and Potosi hot springs because of the need for detailed mapping of the numerous spring vents at each site.

Summary of Previous Work

Waring (1965) presented the most complete catalogue of thermal spring data for southwest Montana. He incorporated many older works such as Peale (1896) and Weed (1905) and included many of his own observations. Other lists such as Balster and Groff (1972) and Renner and others (1976), have been published since 1965 but do not add significantly to the content of Waring's paper.

White and Williams (1975) listed the important thermal spring parameters and provided descriptive data and reservoir assessments for important thermal springs in the U.S., including 9 in Montana.

Several other aspects of geothermal resources such as igneous-related systems, recoverability, and geopressurized systems, are discussed in the paper.

In a paper dealing with the hydrology, activity, and heat flow of a thermal system in Nevada, White (1968) discussed several fundamental geometric and hydrodynamic characteristics of convective geothermal systems. Because these characteristics are fundamental to convective thermal water flow, they are important in the development of models within any geologic setting.

Several papers deal with various aspects of hot water occurrence in southwest Montana. Robertson and others (1975) and Chadwick and Kaczmarek (1975) discussed various structural controls of thermal springs. Chadwick and Kaczmarek (1975) proposed a classification system and conceptual models for typical springs in Montana. Kaczmarek (1974) and Mariner and others (1976) discussed the geochemistry and geothermometry of selected thermal springs. Williams (1975); Hawe (1975) and Qamar (1976); Struhsacker (1976); and Blackwell and others (1973) and McSpadden (1975), described specific thermal areas in Montana: Bearmouth, Camus, LaDuke, and Marysville respectively. Several heat flow studies which cover various portions of Montana have been conducted by Blackwell (1973); AAPG (1973); and Balster (1974).

MAJOR TECTONIC FEATURES OF SOUTHWEST MONTANA

The tectonics of southwest Montana are dominated by four major elements:

1. West-northwest-trending lineaments
2. The Montana Disturbed Belt
3. Extensional faulting
4. Major Igneous Activity

Thermal water systems of southwest Montana are controlled by one or more of these four tectonic elements. The elements will be discussed only to the extent that they relate to the occurrence of geothermal systems.

West-northwest-trending Lineaments

Segments of the west-northwest-trending lineament system of the Northern Rocky Mountains have been discussed by many writers, as early as Calkins and MacDonald (1909) and as recently as Talbot and Hyndman (1973) and Reynolds (1977). J. G. Smith (1965, p. 1400) attempted to synthesize earlier works and his own observations to ". . . encompass completely the regional extent and the significance of these deep-seated zones of lateral movement in the tectonic development of the northern Cordilleran region."

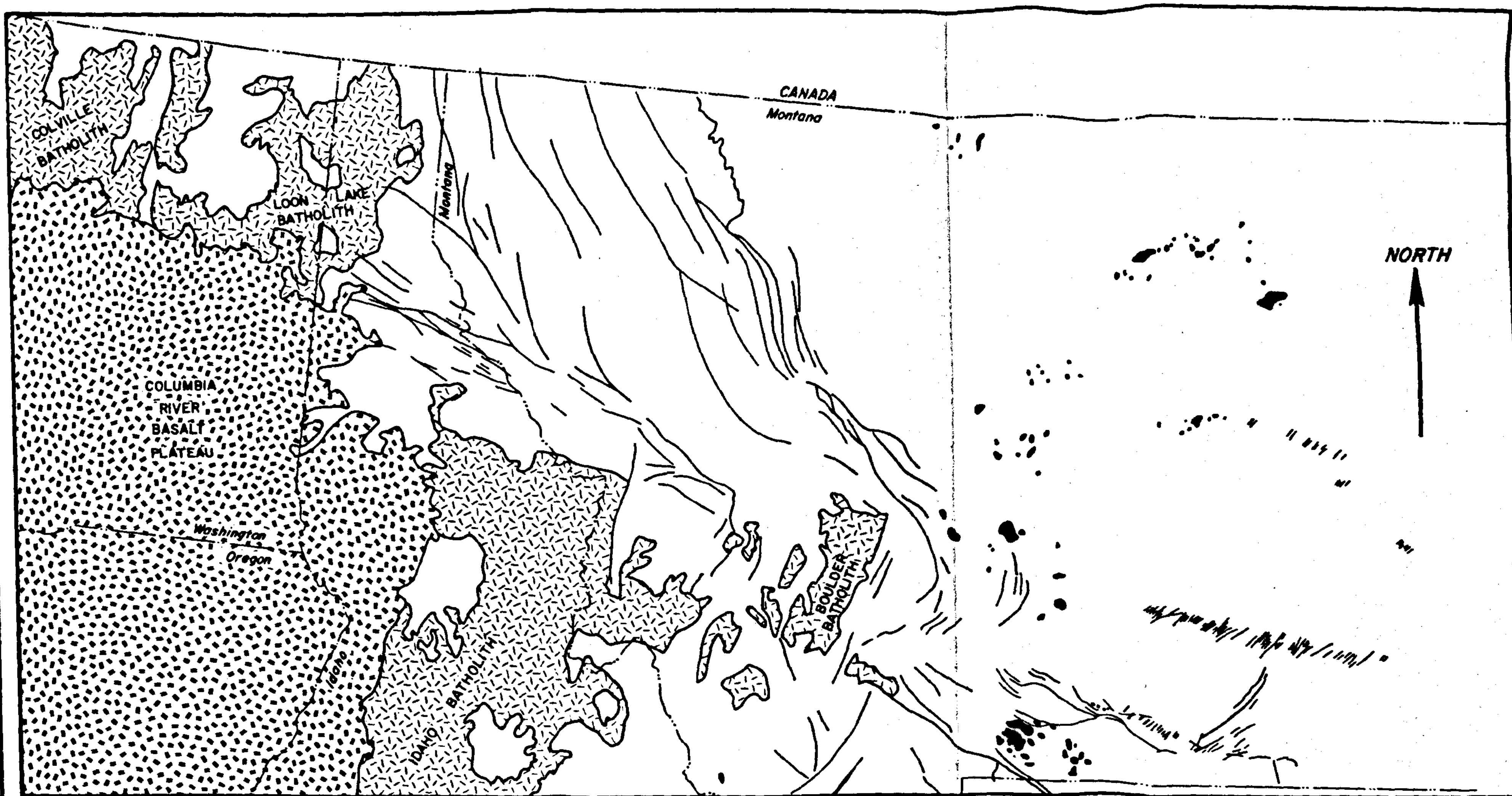
The lineament zone is a structural belt 100-150 km wide and at least 800 km long, extending from central Montana west-northwest across the grain of the Rockies into northern Idaho (Fig. 2) (J. G. Smith, 1965). Smith attributes the various structural elements that compose what he terms the Lewis and Clark lineament to deep-seated transcurrent movements of the earth's crust.

Evidence

Geologic and physiographic evidence strongly favor existence of a major west-northwest structure in the Northern Rocky Mountains. Straight, transverse valleys in the vicinity of Helena and Missoula, Montana which extend to Spokane, Washington coincide with the Osburn fault zone, a major component of the Lewis and Clark lineament (J. G. Smith, 1965). The valleys and fault zone delineate a sharp boundary between the Belt province on the north and the Batholithic and Basement provinces on the south (McMannis, 1965). The exact number, location, and significance of various lineaments within the Lewis and Clark Lineament is controversial.

As a partial solution to the controversy, J. G. Smith (1965) recognized 5 west-northwest structural lineaments (Fig. 3) as superficial expressions of deep-seated lateral movement. The individual lineaments from south to north are described below (see also Fig. 3).

1. Nye-Bowler lineament is defined by the Nye-Bowler fault zone described by Wilson (1936).



LEGEND

EOCENE INTRUSIVES

TERTIARY BASALT

CRETACEOUS-TERTIARY INTRUSIVES

UNDIFFERENTIATED

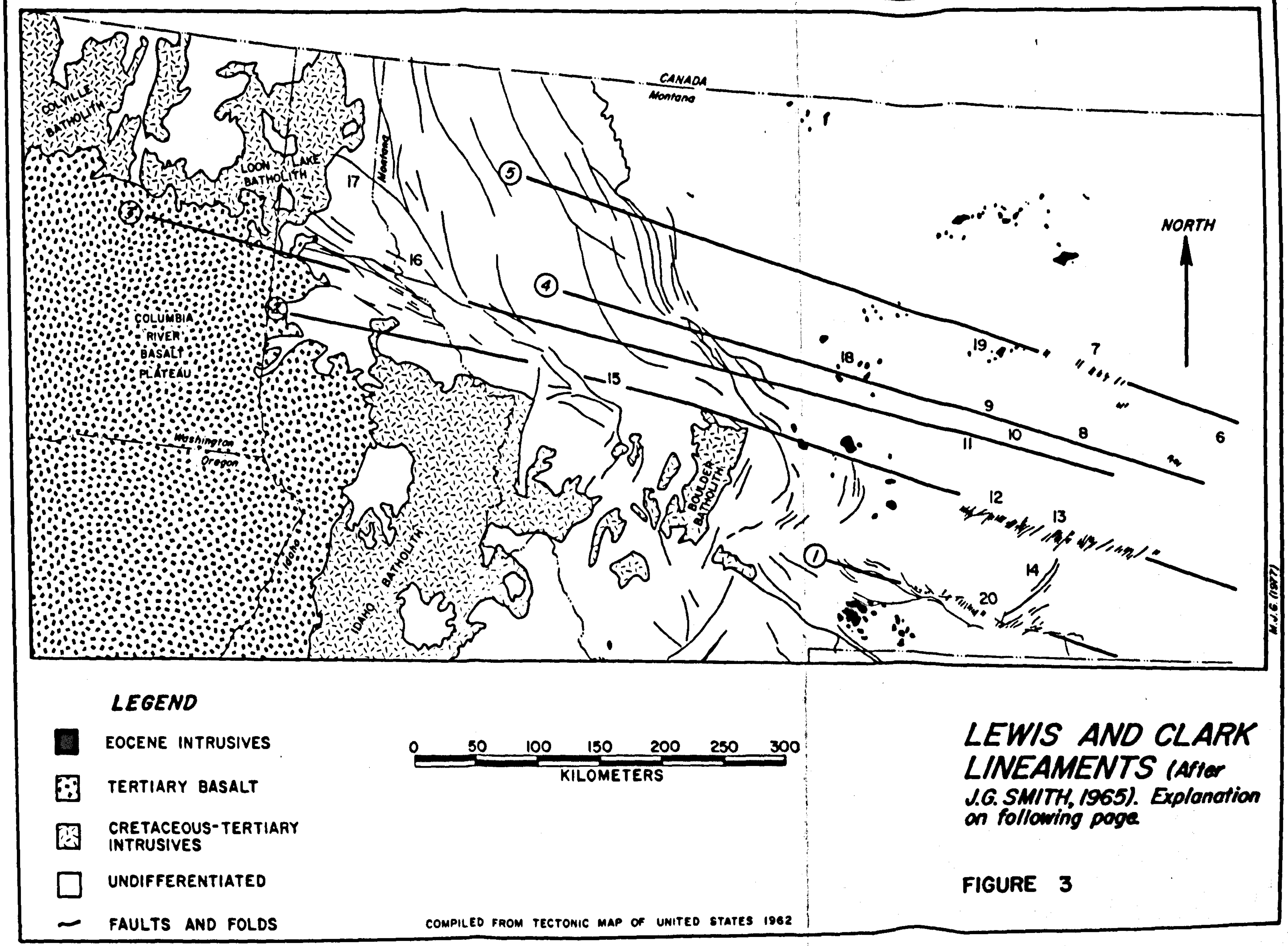
FAULTS AND FOLDS

0 50 100 150 200 250 300
KILOMETERS

COMPILED FROM TECTONIC MAP OF UNITED STATES 1962

GENERALIZED TECTONIC MAP OF A PORTION OF NORTHWESTERN U.S.

FIGURE 2



LEWIS AND CLARK LINEAMENTS (After J.G. SMITH, 1965). Explanation on following page.

FIGURE 3

EXPLANATION FOR FIGURE 3

LEWIS AND CLARK LINEAMENTS

(surface traces of deep-seated megashears) (After J. G. Smith, 1965)

(1) Nye-Bowler lineament

(2) Lake Basin lineament

(3) Osburn lineament

(4) Big Snowy lineament

(5) Cat Creek lineament

CONSTITUENT STRUCTURES

6 Porcupine dome

7 Cat Creek anticline and fault zone

8 Big Wall anticline

9 Big Snowy uplift

10 Woman's Pocket anticline

11 Shawmut anticline

12 Big Coulee-Hailstone dome

13 Lake Basin fault zone

14 Pryor terrace

15 Clark Fork en echelon folds

16 Osburn fault zone

17 Hope fault

18 Little Belt uplift

19 Judith Mountains intrusives

20 Nye-Bowler fault zone

2. Lake-Basin lineament is defined by the linear trend of the Lake Basin fault zone, en echelon folds in the Drummond-Helena area (Clark Fork en echelon folds), southern flexure in the Laramide folded belt and the associated tear faults in the Philipsburg area, and truncated northern ends of the Boulder and Idaho batholiths.
3. Osburn lineament "defined by a series of anticlinal flexures and monoclines developed on the central Montana platform (Sonnenburg, 1956) which includes the Woman's Pocket anticline" (J. G. Smith, 1965, p. 1400). Traced westward, the lineament is aligned with the Osburn fault zone and the southern terminous of the Loon Lake batholith.
4. Big Snowy lineament is marked by structures within the central Montana platform. These include the Little Belt Mountains, southern margins of Big Snowy uplift, and Porcupine dome.
5. Cat Creek lineament includes the northern Montana oroclines and forms the northern margin of the central Montana platform: Judith Mountains uplift, Porcupine dome, and en echelon Cat Creek fault zone.

In addition to geological evidence, Douglas (1972) summarizes geophysical studies, including his gravity and magnetic surveys which support the existence of a major west-northwest-trending structure.

Displacements

Several writers have attempted to explain the paradox of strike-slip movements with opposite sense along the Lewis and Clark lineament. The Lake basin and Cat Creek fault zones and the Clark Fork en echelon folds are classic examples of simple shear deformation resulting from sinistral displacement (J. G. Smith, 1965); the overall appearance of deformation along the lineament is also sinistral. However, indisputable evidence summarized by Wallace and others (1960) has been reported for dextral displacement, up to 25 km, along the Osburn fault zone.

Smedes (1958) proposed that both dextral and sinistral Laramide movement could have occurred simultaneously along a composite, east-pointing wedge with its apex near Missoula. The wedge is bounded on the south by the Lake Basin lineament and along the north by the Osburn lineament. During Laramide compression, westward movement of the wedge produced the two opposing displacements.

Poulter (1959) suggested dextral displacement representing earlier Laramide, Nevadan, or even Precambrian deformation preceded the Laramide sinistral movements. He also suggested dextral movement could have been restricted to the west end of the lineament during Laramide time.

J. G. Smith (1965), however, proposed that all observed displacements can be explained by an overall deep-seated sinistral movement. Deep-seated sinistral movement initially produced folds with north-trending axes. Continued sinistral movement rotated the fold axes westward, at the same time producing complimentary dextral shears which displaced the earlier folds. As counter-clockwise rotation continued, dextral displacement increased and folds and fractures were rotated to their present configuration, subparallel to the Lewis and Clark lineament trend.

Another model (Talbot and Hyndman, 1973) explains the dextral displacement in the vicinity of the Idaho and Shuswap batholiths as the net displacement due to spreading along glide vectors associated with the rising infrastructure and sinistral movement along the Lewis and Clark lineament. Their model is analogous to a ridge-ridge-transform fault system.

The magnitude of the lateral displacement is difficult to estimate because individual fractures are not clearly defined but rather are broad zones of deformation. The apparent offset of the Loon Lake and Idaho batholiths suggests up to 175 km of left-lateral offset since middle Cretaceous time (J. G. Smith, 1965) (Fig. 3). In addition to 175 km of Post-Cretaceous offset, Smith suggested vertical movement has played an important role in the development of the Northern Rockies. He proposed at least 5 phases of vertical uplifts since Precambrian time can be identified on the basis of sedimentary rock distribution, specifically the areal extent of the Protozoic Belt, Ordovician Big Horn, and Big Snowy Group.

Age

Uranium mineralization along fractures within the Osburn fault zone has been dated at 1250 million years (Eckleman and Kulp, 1957) which suggests a lineament existed in the general position of the Lewis and Clark lineament in Precambrian time. In the Tobacco Root Mountains, Precambrian strike-slip movement along northwest-trending faults (Reid, 1957) supports the existence of pre-Belt activity along megashears in the northern Rockies (J. G. Smith, 1965).

Paleozoic strike-slip movement has not been documented, although vertical movement may have influenced distribution of Paleozoic sediments (J. G. Smith, 1965). Stocks, dated at 100 m.y., have been cut by a fault in the Osburn system (Wallace and others, 1960) and post-Eocene movement has been reported in the Lake Basin fault zone (Alpha and Fanshawe, 1954). Reynolds (1977) reports Precambrian through Holocene movement along the Lewis and Clark lineament. Evidence of Holocene movement has been observed along the North

Meadow Creek portion of the northwest-trending Bismark-Spanish Peaks system (Andretta and Alsup, 1960).

Although the exact nature of the Lewis and Clark lineament is still controversial, the existence of a west-northwest-trending fundamental structure is well documented as part of the tectonic framework of the Northern Rocky Mountains.

Montana Disturbed Belt

The rocks involved in Disturbed Belt folding and thrusting range in age from Precambrian (Belt Supergroup) to Tertiary. Thrusting and folding resulted in a minimum crustal shortening of 47 km (Mudge, 1970). Uplift to the west may have exceeded 14,000 meters during the period from very Late Cretaceous to Late Eocene (Mudge, 1970).

The Lewis and Clark lineament divides the Disturbed Belt into two distinct units (Scholten, 1970). Mudge (1970, p. 377) describes the northern part of the Disturbed Belt in Montana as "a northwesterly trending zone of closely spaced westerly dipping thrust faults, many folds, and some longitudinal normal faults and transverse faults" (Fig. 2). In contrast, the southern part is generally more complex. Whether this greater complexity is due to "essentially different tectonic mechanisms" (Scholten, 1970) or one mechanism, complicated in the south by the Mesozoic-Cenozoic batholiths, is unresolved.

Extensional Faulting

The formation of many mountain ranges and intermontane basins of western Montana has been attributed to block faulting similar to that of the Basin and Range Province (Pardee, 1950). The appearance of Oligocene basin sediments and the change to bimodal volcanism in early Oligocene, suggests that block faulting was initiated at this time (Chadwick, 1977). Normal faults which bound various ranges trend north to northeast in western Montana. Seismic activity has continued into Holocene time, much of which has been attributed to extensional faulting (Smith and Sbar, 1974).

Although not necessarily related, the Intermountain Seismic Belt will be discussed with Extensional faulting.

Intermountain Seismic Belt

The Intermountain Seismic Belt (ISB) is a north-trending seismically active zone 100km wide and 1300km long in the western U.S., extending north from Arizona to the Northern Rockies of Montana (Smith and Sbar, 1974). These authors (p. 1205) describe the seismicity as "characterized by shallow focal depths, most less than 15km, and by earthquake swarms that are coincident in some cases with geothermal features and areas of high heat flow."

In the Northern Rockies, earthquake activity has centered around Flathead Lake, Butte, and Helena, Montana, and the Yellowstone-Hebgen Lake area. The largest documented event, magnitude 7.1, along the ISB was the 1959 Hebgen Lake earthquake which produced a vertical displacement of 6.7m (Myers and Hamilton, 1964) along normal faults. Several magnitude 6 earthquakes and numerous smaller events have affected the Helena area in the past 50 years.

Seismicity is extremely important to geothermal activity, particularly in systems which are actively depositing sinter or travertine (Marler and White, 1975). Earthquakes provide the necessary mechanism for reopening clogged discharge channels. Since mineral deposition seems to play only a minor role in Montana geothermal systems, seismic activity is of limited importance in this respect. However, seismic activity may indicate ongoing tectonic processes which propagate permeable conduits in otherwise impermeable rocks.

Igneous Activity

Extrusives

Volcanism in Montana can be divided into 5 major episodes (Chadwick, 1972).

1. Precambrian: basaltic lavas in Belt rocks (Ross, 1959).
2. Cretaceous: The Elkhorn Mountains and related volcanics represent deposits from major volcanic centers during Late Cretaceous time. Predominantly andesitic flows, tuffs, breccias, and reworked volcanoclastics accumulated to 4600 m. Dacitic to rhyodacitic lava flows and volcanoclastics are exposed near Bozeman, Montana. The Madison and Beartooth Ranges were also the site of volcanism; andesitic and dacitic respectively.
3. Eocene: Absaroka-Gallatin region was the site of andesitic to dacitic flows and breccia outpourings. Quartz latite was extruded in the Boulder batholith region (Lowland Creek volcanics) and alkalic igneous activity in the central Montana province.
4. Mid-Tertiary: A major rhyolitic field in the northern Boulder batholith and adjacent areas has recently been dated at 36-37 m.y. (Galloway, 1977a; Chadwick, 1977). Basalts were extruded 23-34 m.y. ago in west-central and southwest Montana.

Chadwick (1977) suggests this bimodal activity reflects initiation of extensional faulting in southwest Montana.

5. Pliocene-Pleistocene: This period represents voluminous, sometimes violent, eruptions of rhyolitic lava and ash with interspersed basaltic lava flows in the Island Park-Yellowstone plateau region.

Because of the lack of recent volcanism in Montana outside the Yellowstone Park region, most Montana thermal springs appear unrelated to cooling igneous bodies. Intrusive igneous bodies, 5-10 km thick and older than 2-10 m.y. would no longer supply heat to thermal systems (Smith and Shaw, 1975). Excluding the Pliocene-Pleistocene activity in the Yellowstone Park area and eastern Snake River Plains, activity in Montana is too old to be supplying heat to any thermal system.

Intrusives

Boulder batholith: The Boulder batholith is a Late Cretaceous Early Tertiary epizonal composite mass of at least 15 plutons. Composition ranges from syenogabbro to alaskite, but the dominant exposed rock type is quartz monzonite. The batholith proper, exposed over 6000 km^2, is flanked by numerous smaller satellitic plutons (Fig. 2), including the Tobacco Root and Pioneer batholiths. The Boulder batholith intrudes over 6 km of folded Belt, Paleozoic, and Mesozoic marine sedimentary strata of shelf or miogeosynclinal facies (Klepper and others, 1971). Roof rocks consist almost exclusively of Elkhorn Mountains volcanics, which are related chemically and isotopically to rocks of the batholith (Tilling, 1973). The volcanics are slightly older (78 m.y. compared to 68-78 m.y.) but volcanism probably overlapped with the initial phases of intrusion. Contact metamorphism of the intruded rock is slight, ranging from nearly unmetamorphosed host rock to epizonal skarn-hornfels (Knopf, 1957). The batholith is the host rock for several rich ore deposits, such as the deposits around Butte, Montana.

The geometry and method of emplacement of the Boulder batholith is hotly debated. The same geological and geophysical observations have been interpreted differently by various workers. The two basic interpretations are: 1) The Boulder batholith is a classical steep-sided, deep-rooted (greater than 15 km), intrusive (Klepper and others, 1971), 2) The Boulder batholith formed as a thin (less than 5 km) and shallow mass that spread over a floor of pre-magmatic rocks and was roofed largely by its own volcanic ejecta (Hamilton and Myers, 1967). For a complete review of evidence supporting each interpretation, see Klepper and others (1971, 1974) and Hamilton and Myers, 1974a and b).

From an extensive gravity survey in southwest Montana, Burfeind (1967) suggests Cretaceous-Tertiary granitic intrusives underlie much of the area. All intrusives exposed at the surface are connected at depth and mountain ranges such as the Madison, Bridger, and Gallatin are underlain by calc-alkaline intrusives at 4300-4900 m below the surface. Plutons exposed at the surface have either been faulted in place and/or magma intruded along pre-existing faults.

Burfeind (1967) suggested that calc-alkaline intrusives of southwest Montana were derived from a parent magma which was responsible for all Cretaceous and Tertiary calc-alkaline intrusives of the Cordilleran.

Hyndman and others (1975) have shown that emplacement of the Boulder batholith may be closely related to the detachment of the Sapphire block from the rising Idaho batholith infrastructure. Emplacement of the 10-15 km thick Sapphire block onto the crust would have depressed the region in front of the block and possibly controlled the flow of magma toward the east into the depression (Hyndman and others, 1975). The Boulder batholith and associated plutons may, therefore, be related temporally and spatially to the detachment and emplacement of the Sapphire tectonic block.

Idaho batholith: The Idaho batholith is a Late Mesozoic catazonal intrusive, exposed over approximately 40,000 km^2 in central Idaho and extreme western Montana (Fig. 2). A full range of compositions exist, but quartz monzonite is the most abundant rock type. The batholith syntectonically intruded high-grade regionally metamorphosed rocks (Hyndman and others, 1975), many of which form roof pendants or inliers (Larson and others, 1958). In the south, batholithic rocks are in contact with and may underlie Tertiary volcanics (Larson and others, 1958). Although local portions of the batholith may have been formed by replacement, the Idaho batholith is considered to be principally the result of intrusion by a granitic melt at approximately 20 km depth (Larson and others, 1958; Chase, 1973). Numerous plutons and northeast-trending dikes intruded the batholith during Eocene time. These plutons, including the Lolo pluton, are generally more felsic and are thought to have been emplaced at a much higher level than the Mesozoic plutons of the batholith (Williams, 1977).

HYDROLOGIC CHARACTERISTICS OF FRACTURED MEDIA AND GEOTHERMAL SYSTEMS

Fracture Phenomena[1]

According to Stearns and Friedman (1972), there are two types of fractures, shear and extensional. Movement is perpendicular to the fracture plane in extensional-fractures and parallel in shear-fractures. Two potential shear-fracture orientations and one extensional-fracture orientation are possible in any triaxial stress state, where $\sigma_1 > \sigma_2 > \sigma_3$. The axis of principle stress, σ_1, bisects the acute angle formed by the two shear-fracture planes. The axis of least principle stress, σ_3, bisects the obtuse angle.

Nearly 200 years ago, Coulomb (1776) stated that shear-fracturing is controlled by the material's initial shear strength and internal friction. Experimental work has shown this to be true; shear fractures do not occur along planes of maximum stress (45° to σ_1), but rather at some angle less than 45°. The departure from 45°, Φ, is the angle of internal friction, which is a property of the material.

Ductility is probably the most important factor in fracture development. Laboratory deformation of rocks of varying ductilities is summarized in Figure 4. Ductility increases from left to right.

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[1] Portions of this section are summaries of "Reservoirs in Fractured Rock," Stearns and Friedman, 1972.

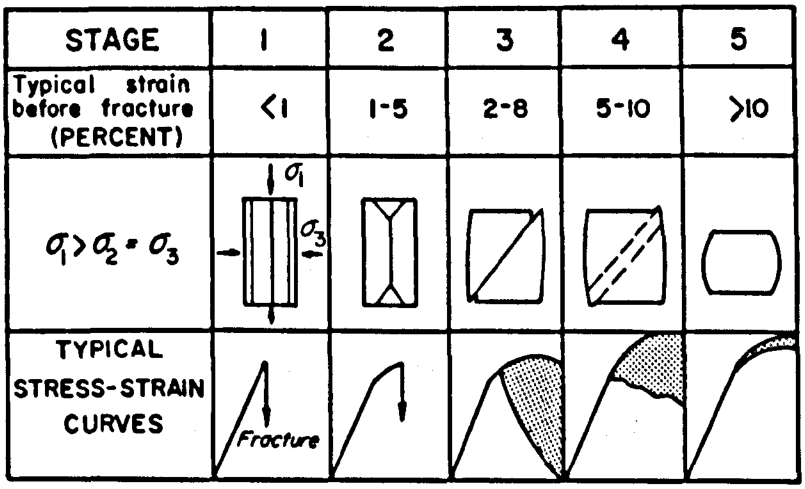

Figure 4. Summary of Rock Deformation. Ductility increases from left to right. (After Griggs and Hardin, 1960)

The linear relationship in stage 1 between stress and strain and the dominant occurrence of extension-fractures, is typical of brittle rock.

A single fracture with permanent distortion (barreling) occurs in stage 3. With increasing ductility, major movement is along a shear zone made up of many subparallel shear-fractures and minor extensional-fracturing. Stage 5 represents very ductile rocks which undergo permanent strain with no macroscopic fracturing.

Factors which affect ductility and therefore control rupture are rock type, temperature, effective confining pressure, and shear rate. The study of natural fracture systems is complicated with respect to quantification of the various factors because of numerous possible parameter combinations.

Natural Fracture Systems

Stearns and Friedman (1972) subdivide pervasive and consistently oriented fracture systems into two major classes: structure-related fractures and regional orthogonal fractures. Structure-related fractures are associated with specific faults and folds, whereas regional orthogonal fractures are involved in the structural development of an entire region.

The origin of regional orthogonal fractures is still unresolved. Several explanations appear in the literature, such as Blanchet (1957); Price (1959); and Hodgson (1961), but no explanation accounts for all of the observed features. Fractures of this type are commonly continuous for several kilometers as single breaks or as narrow zones. They have been observed to be vertically continuous up to 150 m. Their orientations are extremely consistent over large areas.

Structure-related fractures are assignable to the same stress state as the associated structure. Shear-fractures mimic orientation and sense of shear of the major associated fault. Therefore, both shear and extensional-fractures can help delineate faults, particularly in homogeneous rocks. The relative development of the three potential fracture orientations associated with a fault cannot be predicted, although displacement along the major fault can be estimated by measuring the cumulative slip along parallel fractures.

If all other factors which control fracturing are constant, the number of fractures is most affected by variations in lithology (Stearns and Friedman, 1972). If lithology is constant, as in a granitic pluton, the variations in fractures must be a result of variations in dip and strike and spacing can be attributed to changes in structural trend of individual large and small scale structures. This suggests that a thorough examination of fracture patterns may reveal structures which would otherwise go unobserved in homogeneous rock.

Fractures in Crystalline Rocks

Flat-lying Fractures

Stearns and Friedman (1972) discussed fractures which occur in all lithologies. However, in considering only intrusive rocks, the occurrence of flat-lying fractures is an important additional source of fracture-permeability. This fracture type is particularly important to the discussion of thermal water systems because they represent the only fractures which can maintain artesian pressure. Vertical fractures in the near-surface environment should lose pressure along their entire permeable length. Artesian pressures within horizontal fracture systems have been reported (for example, LeGrand, 1949; Galloway, 1977a).

Flat-lying fractures or sheet structure tend to be better developed in relatively unfractured granite (Jahns, 1943; Hills, 1963), suggesting stresses which cause sheeting are taken up by existing fractures in highly fractured granites (Hills, 1963). Fracture spacing tends to increase with depth, but has been reported to be prominent below 100 m in New England quarrys (Hills, 1963). Flat-lying fractures have also been reported in the Butte mines at 1500 m (R. Miller, oral comm.) and in the deep drill-hole at Marysville, Montana at a depth of 2000 m (McSpadden, 1975). There is some question as to the origin of flat-lying fractures in the Marysville well (see p. 142).

LeGrand (1949) discussed possible origins of flat-lying fractures, such as: 1) expansion due to hydration of feldspar to form kaolin,

2) seasonal temperature variation, and 3) expansion due to removal of overlying material. He concluded that removal of overburden is the most likely cause.

Smedes (1973) interprets flat-lying fractures of the Homestake pluton, within the Boulder batholith, as controlled by flat-schlieren. The schlieren are mafic-rich quartz monzonite and granodiorite, locally injected and replaced by pegmatite. This suggests flat-lying fractures in this pluton are of a primary origin related to flow structure near a flat roof of a partially solidified magma (Smedes, 1975).

Fracture Propagation

Knapp (1976) has shown heating host rocks by an intruding body increases pressure within fluid-filled pores due to differential expansion of rock and fluid. Fluid pressure increases until it is approximately equal to the least compressive stress, the rock then yields by fracturing.

In shallow environments, a 20°C increase will increase the pore fluid pressure by 150 bars; this is sufficient to initiate vertical fracturing under the appropriate stress conditions (Knapp, 1976). "The ensuing increase in permeability increases the convective heat flux which accelerates the upward movement of the fracture front to a few cm/year" (Knapp, 1976, p. 957).

Fracture-Porosity and Permeability

Fracture-porosity and permeability are controlled by fracture width or aperture, area, spacing, surface roughness, and filling (Stearns and Friedman, 1972; Parsons, 1972). Fracture-porosity values reported in the literature range from less than .05% (Snow, 1968) to 6% (Regan and Hughes, 1949). Estimated fracture-permeabilities range from a few millidarcys to many darcys (Stearns and Friedman, 1972).

Estimation of Fracture-porosity and Permeability

Porosity: Several methods for estimating fracture-porosity have appeared in the literature. Elkins (1953) calculated fracture porosity by direct measurement of fracture aperture and spacing from cores. Murray (1968), working with folded sedimentary rocks, showed fracture porosity (ϕf) is directly proportional to the product of bed thickness and curvature.

$$\phi f = \frac{T}{2R} = \frac{1}{2} T \frac{d^2 z}{dx^2} . \qquad (1)$$

where T = the bed thickness, and

R = radius of curvature. From equation (1),

$\frac{d^2z}{dx^2}$ = the curvature.

Fracture-porosity is not dependent upon aperture or spacing in equation (1).

Using aquifer test data from selected dam sites, Snow (1968) described a method for calculating fracture-porosity, openings, and spacings:

$$\Theta = 5.45 \left(\frac{k}{\Delta^2}\right)^{\frac{1}{3}}, \qquad (2)$$

where Θ = porosity

k = intrinsic permeability

Δ = average spacing.

Assuming all openings are equal,

$$2B = \Theta\frac{\Delta}{3}, \qquad (3)$$

where B is the half-aperture. Digital studies reveal an error of ±20% in computing average porosity (Snow, 1968).

Fracture-porosity can also be calculated from conventional electric logs if the porosity of the unfractured rock is known (Pirson, 1967).

<u>Permeability</u>: Parallel plate flow, which is applicable to both single-fracture and multi-fracture studies, is the basis of fracture flow analysis. The successful use of the concept in single fracture studies depends on the variation between ideal and actual fractures. In multi-fracture-rock-mass studies,

> individual fissures [fractures] may be assumed to function as ideal plate openings, but their collective response must be evaluated in terms of the elastic properties of the mass and the geometric parameters which account for the dispersion of orientations and spacings of fissures (Parsons, 1972, p. 90).

Media with Parallel Conduits: Using a nonconducting solid cut by smooth parallel openings (Fig. 5), permeability K can easily be calculated if discharge and potential gradient are known.

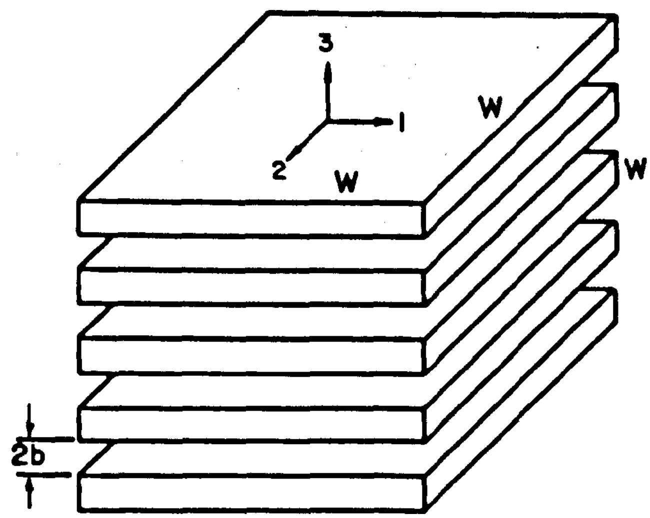

Figure 5. A Solid Volume Cut by Parallel Plane Conduits. Solid volume has dimensions of W and parallel plane conduits have an opening of 2b (After Snow, 1969).