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GRANITE EMPLACEMENT WITH SPECIAL REFERENCE TO NORTH AMERICA

By A. F. Buddington

Abstract

Publications of the last 25 years that discuss the emplacement of granite plutons are reviewed, with special reference to North America. The plutons are classified according to emplacement in the epizone, mesozone, or catazone of the earth's crust. It is found that those emplaced in the epizone are almost wholly discordant; those in the mesozone complex, in part discordant and in part concordant; and those of the catazone predominantly concordant. Granite formed by granitization is considered to be minor or local in plutons of the epizone, common but subordinate in those of the mesozone, and a major factor in plutons of the catazone. The authors of the papers reviewed in general, however, infer that magma was either directly or indirectly the major factor in all the zones. Contrary to some current theories, this review emphasizes the great number and great total volume of granitic plutons emplaced as fluid magma in the epizone and their community of origin with lavas of similar composition directly associated in time and space. Magma is thus inferred to play the major role in Tertiary stocks and batholiths. There appears to be no discontinuity between plutons of the epizone and those of the mesozone, and a major role for magma is indicated for the latter also. The evidence is not clear as to whether plutons of the mesozone are continuous with those of the catazone, have roots in the catazone, or are pinched off from it. Batholiths emplaced in the mesozone are dominant in most basement complexes of Precambrian to Early Cretaceous ages.

Sommaire

Les publications des 25 dernières années qui traitent de l'emplacement de plutons de granit sont passées en revue, en se référant spécialement à l'Amérique du Nord. Les plutons sont classés en fonction de leur emplacement dans l'epizone, la mesozone, ou la catazone de la croûte terrestre. On a découvert que ceux placés dans l'epizone sont presque complètement discordants; ceux dans la mesozone complex, en partie discordants et en partie concordants, et ceux situés dans la catazone concordants de façon prédominante. Le granit forme par le granitisation est considéré comme peu important ou épars dans les plutons de l'epizone, courant mais secondaire dans ceux de la mesozone, et d'importance majeure dans les plutons de la catazone. La plupart des auteurs de ces publications passées en revue déduisent cependant que le magma était, directement ou indirectement, le facteur le plus important dans toutes les zones. Contrairement é certaines théories courantes, cette étude met en évidence le grand nombre et le grand volume total de plutons granitique emplace sous forme de magma fluide dans l'epizone; aussi leur origine communal à celle des laves de composition similaire associées directement à elles dans le temps et dans l'espace. On en déduit donc que le magma joue le rôle de premier plan dans les batholithes et les stocks Tertiairies. Il ne parait pas y avoir de discontinuité entre les plutons de l'epizone et ceux de la mesozone, et, par consequent, on indique que le magma joue également un rôle important dans le mesozone. Il n'est pas clairement évident que les plutons de la mesozone forment une suit ininterrompue avec ceux de la catazone, qu'ils aient des racines dans la catazone, ou bien qu'ils se disjoindre. Les batholithes situés dans la mesozone dominent dans la plupart des complexes profonds de l'epoque Précambrienne jusqu' à l'epoque Crétacee inférieure.

Zusammenfassung

Es werden Veröffentlichungen der letzten 25 Jahre besprochen, welche die Position von Granit-Plutonen, besonders solcher von Nord-Amerika, behandeln. Die Plutone werden entsprechend ihrer Lagerung in der Epizone, Mesozone oder Katazone der Erdkruste Klassifiziert. Es zeigt sich, daß die Lagerung in der Epizone nahezu völlig diskordant ist, diejenige in der Mesozone dagegen komplex, teilweise diskordant, teilweise konkordant, und diejenige in der Katazone vorwiegend konkordant. Granite, welche durch Granitisation geformt wurden, müssen in den Plutonen der Epizone als verhältnismaßig, selten oder nur örtlich angesehen werden, als regelmäßig, jedoch untergeordnet, in denen der Mesozone, und als ein Hauptfaktor in den Plutonen der Katazone. Die Autoren der besprochenen Veröffentlichungen kommen jedoch im allgemeinen zu der Überzeugung, daß das Magma, direkt oder indirekt, in allen zonen der Hauptfaktor war. Im Gegensatz zu einigen anderen umlaufenden Theorien betont diese Zusammenschau die große Anzahl und den großen Gesamtraum von Granitplutonen, welche als flussiges Magma in die Epizone eingedrungen sind und unterstreicht ihren gemeinschaftlichen Ursprung mit Lavamassen ähnlicher Zusammensetzung, mit denen sie zeitlich und räumlich in direkter Verbindung stehen. Das Magma spielt infolgedessen eine Hauptrolle in tertiären Granitstöcken und Batholithen. Es scheint keine Unterbrechung zwischen den Plutonen der Epizone und denen der Mesozone zu bestehen, und für die letztere scheint das Magma ebenfalls eine wesentliche Rolle gespielt zu haben. Die Beweise sind nicht klar, ob die Plutone der Mesozone in diejenigen der Katazone übergehen, ob sie Wurzeln in der Katazone haben oder ob sie von ihr abgeschnitten sind. Die in die Mesozone eingelagerten Batholite sind in den meisten Grundmassiven vom Präkambrium bis zur Unteren Kreide vorherrschend.

ЗАЛЕЖИ ГРАНИТА В СЕВЕРНОЙ АМЕРИКЕ

А. Ф. Буддингтон

Абстракт

Печатные труды последних лет, описывающие положение гранитных плутонов, особенно труды изданные в Северной Америке являются предметом настоящего обзора. Плутоны классифицированы согласно их положению в эпизоне, месозоне или катазоне земной коры. Было найдено что плутоны находящиеся в эпизоне почти полностью несовместимы; теже в месозонном соединении частью несовместимы и частью совместимы тогда как те в катазоне главным образом совместимы. Гранит образованный путем гранитизации разсматривается как второстепенный или местный фактор в плутонах эпизона, обычный но второстепенный в плутонах месозона, и как главный фактор в плутонах катазона. В общем авторы статей настоящего обзора однако допускают что магма непосредственно или посредственно является главным фактором во всех зонах. В противоположность некоторым современным теориям рассмотренным в настоящем обзоре подчеркивается большое количество и большой объем гранитных плутонов внедренных в виде жидкой магмы в эпизоне и их средство по происхождению с лавами подобного же состава непосредственно связанными по времени и пространству. Повидимому не существует непрерывной связи между плутонами эпизона и плутонами месозона, при чем главная роль для магмы указана также для последнего. Еще не ясно являются ли плутоны месозона непрерывно связанными с плутонами катазона, исходят ли они из катазона или отщеплены от него. Батолиты, внедренные в месозоне играют главную роль в большинстве основных масс Прекамбрионовой вплоть до ранней Кретасовой эпох.

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INTRODUCTION AND ACKNOWLEDGMENTS

A wealth of detailed descriptions of the internal structure and external relationships of granitic plutons to country rock has been published since the reports (1931–1935) of Professor Grout's "Committee on Batholith Problems", the review by Daly (1933), and the memoir of Balk (1937). Read has meanwhile (1949; 1951; 1955; 1957) developed a philosophy of the origin and genetic relationships of granitic bodies under the phrase "The Granite Series" which is a major contribution. Read emphasizes that the mechanics of emplacement of granitic masses must be interpreted in the light of their regional setting. The writer proposes to amplify this idea further, largely in the sense of an essay review documented with specific examples, based preponderantly on the pertinent literature of the past 25 years that describes the granitic plutons of North America. The plutons of North America are emphasized because the author is better able to evaluate the implications of the literature on them; and because the phenomena of the plutons of North America and their interpretations have led to an emphasis on certain mechanics and conditions of emplacement that is different from that of much of the current European literature and deserves review and consideration. A few examples of plutons from outside North America will be cited to exemplify or accentuate certain ideas or phenomena.

The writer is indebted to Preston E. Cloud, Jr., H. H. Hess, F. F. Osborne, and Arie Poldervaart for friendly criticism and constructive suggestions. They should not, however, be held responsible for shortcomings of the review.

The problem of the origins of granite is necessarily a factor in the consideration of the mechanics of emplacement of plutons. Many geologists who believe that most granites are formed by "granitization" or "transformation" repeatedly emphasize that the problem must be solved by geology and field evidence. There is also the implicit inference that field geologists, with independent minds and familiar with the current ideas of granitization, will find granitization the best hypothesis to explain most or nearly all granites. Yet the North American literature of the past 25 years emphasizes strongly the role of magma either directly or indirectly in the problem of emplacement of plutons. This despite the fact that the authors quoted, more than 100, are field geologists who accept as valid concepts the potentiality of formation of granite by granitization as well as emplacement by plastic crystalline flow. A review of the literature, in general, makes it obvious that we do not yet have dependable criteria that are acceptable to geologists as a whole to distinguish between the products of the different mechanisms of emplacement. There would also probably be little dissent from the view that such geologic problems will *not* be solved by field geology alone, but by retention of what seems good in old ideas with constant rethinking and co-ordination of new hypotheses, of data from new experiments and new laboratory studies, and of new results of field geology aided by the rare new "flash of insight" idea.

In the discussion to follow the dividing line between a stock and a batholith will be taken as roughly 40 square miles as suggested by Daly. Granite, except where indicated otherwise by the context, will include the family of granitoids such as quartz diorite or tonalite and trondhjemite, granodiorite, quartz monzonite or adamellite, and granite or leucogranite and alaskite. The term pluton will be used here in a very general sense for any body of intrusive igneous (or "pseudo-igneous" by metasomatism or recrystallization) rock of any size or shape. Dikes, sills, and laccoliths will receive but passing mention. The term plastic crystalline flow here means flow of material which remains wholly or predominantly crystalline during thoroughgoing deformation and includes the concept of recrystallization through partial melting or solution and redeposition.

READ'S "GRANITE SERIES"

Read's concept of the "Granite Series' may best be presented by excerpts from his writings.

> (1957, p. 79) "Intrusions have been classed as pre-tectonic, syntectonic or post-tectonic. I have endeavored to codify these relationships in what I call the Granite Series (Read, 1949), a series which relates the nature and form of different types of granitic bodies with their place in the fold-belt and the time of their final solidification. The *Granite Series* can be represented thus:
>
> ——————TIME——————→
>
> ————CRUSTAL LEVEL————→
>
> | | | | |
> |---|---|---|---|
> | Autochthonous granitization granites, migmatites and metamorphites | Parautochthonous granites | Intrusive magmatic granites | Plutons |
>
> Deep in the fold-belt are formed, at an early stage of orogeny, great complexes of granitization granites associated with migmatites and widespread regional metamorphic rocks. As the orogeny continues, a part of these autochthonous granites becomes partly unstuck from its surroundings and moves higher in the fold structure. This process continues with the movement of true intrusive and magmatic portions late and high, and culminates in the emplacement of the granite plutons, highest and latest, pushing their way as almost solid bodies even into the post-orogenic sediments." (1951, p. 21) "The resulting parautochthonous granites show variable marginal relations, in some places migmatitic, in others characterized by an aureole of thermal type. This movement out of the migmatitic-metamorphic setting may continue till the genetic ties are completely severed and *true intrusive granites* emplace themselves in higher levels of the crust maybe as magma but more likely as migma. The final term of the granite series is represented by the high-level *plutons*, intrusive into non-plutonic regions late in the history of the orogen concerned.... The plutons are the domain of the Granite-tektonik of the Cloos school and, as their emplacement produced considerable folding and distortion in the country rock surrounding them, they came in as almost dead bodies."

The writer believes that Read's discussion needs some major revision. Plutons with "granite-tektonik" are not the final terms of the granite series. On the contrary they belong almost wholly to the mesozone where, as multiple units, they may form huge batholithic complexes. The final terms of the granite series are the plutons of the epizone. Read minimizes as "few", "puny", and "nearly dead" the bodies emplaced in the upper levels of the crust. This is wholly inconsistent with the number, size, significance, and the evidence for mobility and fluidity of most plutons emplaced in the epizone of the crust in North America. Knopf (1955, p. 697) estimates that plutons of Tertiary age in North America and Greenland (all emplaced in the epizone) have a total area of 52,000 square miles.

Hans Cloos (1931) has presented a succinct pertinent discussion of the possible structural relationships of plutons at different depths. A summary prepared by S. W. Sundeen (1935, p. 48–49) is quoted here

> "In a single mass the inner tectonics differ at different horizons. In the upper horizon there are poor, elusive structures in irregular branching stocks. In the moderate depths there are well-formed arches, schlieren domes, partitions or pendants of wall rock and cleavage with border deformation and mylonites; with or without a stretching of the border spalls. In the deep zone there are arches of gneiss. The whole region is deformed and moves with the magma as an ill-defined

> occurs not only before and during crystallization but in part after. There is thus a lack of any small dynamic border zone and a lack of fracture surfaces, border spalls and most other features of the moderate zone. Cloos sketches a vertical section of a composite of these three horizons, each mass smaller and drained up from the one below, and with less foliation than the one below; but does not deny that a single mass may grade downward without much change in horizontal section into the conditions of the deeper zone."

Zones of Emplacement

The granite emplacement series will be discussed with emphasis on the internal and external structure of the plutons within different temperature–depth or intensity zones of the crust. Under the simplest hypothesis the intensity of regional metamorphism may be expected to increase somewhat uniformly with depth and therefore afford an indicator of the depth. The site and period of granite emplacement, however, is not one of static conditions but of dynamic changes in the environment. This is especially true for the mesozone. The intensity of regional metamorphism in belts away from the intrusives may still be used, however, as a suggestive clue to the depth zone. (*See also* Michot, 1957). Inasmuch as we are particularly interested in the physical conditions of the country rocks at the time and in the region of emplacement, the upper and lower limits of each zone will have a considerable range of depth for different regions and at different times in development. The depths for plutons with similar characters will depend on the temperature, pressure, relative mobilities of the country rocks, and other factors. The term zones as used here thus refers actually in substantial part to *intensity zones* rather than strictly depth zones. At the same level in the mesozone a batholith may be emplaced discordantly in only warm country rock during early stages and conformably in hot rock during late stages, especially in the roof. A predominantly mesozonal pluton may have characters peculiar to the catazone in the roof portion: In some examples the estimation of the physical conditions may be difficult and little better than a guess. Nevertheless examples for which at least fair to good data are available afford the basis for a reasonably consistent picture and afford some additional insight into the problems of emplacement.

Michot (1956, p. 28) suggests that the epizone may be taken as extending from the surface to a depth of 10 km. The mesozone and catazone will be successively below this.

The development of the greenschist facies of metamorphism in rocks could be considered as a characteristic phenomenon of the mesozone. Fyfe, Turner, and Verhoogen (1958, p. 218) estimate that it is unlikely to develop below a temperature of about 300°C. and above a depth of about 10 km. Their curves (p. 182) for rise of temperature with depth suggest that at a depth of about 10 km in a great thickness of sediments, after depression in the earth's crust, the temperature might rise to about 250°C., and where the temperature had been increased by magmatic intrusion it might be as high as 450°C. A rise of temperature in the country rock above an intruding magma is strongly implied by such data as that given by James (1955). The magma may thus be emplaced in rock of temperature higher that that otherwise appropriate for the general region. A depth commonly of 4 miles but with occasional extension, perhaps, to 6 miles seems a reasonable estimate for the base of the epizone.

The depth of the base of the mesozone or top of the catazone where the amphibolite facies starts must likewise have a substantial range, perhaps from as shallow as 5 miles to as deep as 10 miles. Wegmann (1935) estimated the minimum depth of the "migmatite front" at 10 km. The temperature range for the mesozone may be estimated to vary from about 250°–350° at the top to 500°C. at the base.

Tuttle and Bowen (Adams, 1952, p. 38) wrote that "It is improbable that many granites reaching the light of day have crystallized at depth greater than 9 miles". This seems slightly low. The possibility that erosion has exposed levels at a maximum a few miles deeper than this must be considered, but in general it is probably of the right order of magnitude. Gutenberg (1957) cites figures of 35 km for the depth of the "granitic" crust beneath the Alps and 25–30 km beneath the Sierra Nevada. If this is assumed to indicate the depth of the down folded sial, and if reasonable estimates are made for the thickness of eroded material a figure of about 25 miles is arrived at as the normal maximum depth for sialic material. Assuming further that the minimum thickness of sialic basement complexes in the continental shields is about 10–12 miles, the inference may be drawn that present levels of erosion have rarely exposed rocks that were ever at a depth greater than about 12–15 miles.

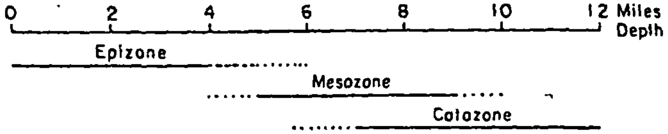

Figure 1.—Schematic Relationships of Emplacement Zones

The rocks of the granulite facies in the Grenville belt appear to represent those formed at some of the deepest depths now exposed in North America. They are estimated to have formed between 600° and 700°C. Fyfe, Turner, and Verhoogen (1958, p. 182) give a curve that indicates that a temperature range of 600°–700°C. could be reached at depths of 9–13 miles where the gradient had been increased by magmatic intrusion. Rosenquist (1952, p. 102) estimates the minimum depth for the development of the granulite facies in this temperature range as 9–10 miles.

The predominance of mesozonal batholiths in basement complexes, however, indicates that only locally has erosion cut very deep.

Possible depth relationships for the zones are shown in Figure 1.

There is in the western Cordillera of North America a rough correlation between the structural relationships and zones of emplacement of the plutons and their ages, especially for those of Late Jurassic and younger age. Plutons of Tertiary age as now exposed were exclusively emplaced in the epizone and may be associated in space and general time with volcanic rocks of equivalent composition and often emplaced in them; those of the Late Cretaceous are also emplaced in the epizone, but there are *more* large plutons; the great composite early Late Cretaceous (?) Southern California batholith has characters transitional between those of the epizone and those of the mesozone, whereas the composite stocks and batholiths of the Late Jurassic and Early Cretaceous were emplaced in the mesozone with earlier members of the largest batholiths emplaced in the catazone.

Similarly, in the Appalachian orogen, the post-Pennsylvanian plutons were all emplaced in the epizone, whereas many earlier plutons of Devonian age were intruded in the mesozone, transitional mesozone-catazone, or locally perhaps the catazone.

For rocks older than the Tertiary, however, there is no necessary systematic relationship between age of rocks and the depth at which they were emplaced. Unmetamorphosed plutons 2.5 billion years old may be of the type emplaced in the mesozone, as in the Keewatin belt of the Canadian Shield where erosion to only moderate depths is indicated.

Plutons of Epizone

"Dr. Hutton's theory of granite . . . at the same time that it conceives this stone to have been in fusion, supposes it to have been, in that state injected among the strata already consolidated." John Playfair, 1802.

Introduction

It is rare that estimates are given in the literature of the depths at which the present exposed parts of a pluton were intruded. Tertiary intrusions as now exposed may be expected to have been emplaced within the epizone; the time for subsequent erosion has been too short to permit deep erosion. A review of the literature shows that Tertiary intrusions have the following characters, and these will be used as criteria in classifying intrusives in the epizone of older ages.

Tertiary stocks and batholiths are largely or wholly discordant to the country rock no matter whether they occur in Precambrian schists and gneisses or in folded Paleozoic and Mesozoic sediments, or, as is common, in gently dipping Tertiary volcanic rocks. Occasionally, as in the Gold Hill stock in Nevada (Nolan, 1935, p. 43–48), part of the walls are controlled by preintrusive faults. They may occur in limestones, a type of rock peculiarly resistant to granitization, without any suggestion of relics or inheritance by replacement. A few granitic plutons may be effectively homogeneous in composition, but most are of composite character caused by a successive series of magma emplacements of diverse composition. The diversity is commonly from syenite or monzonite to granite, or from diorite through quartz diorite, granodiorite, and quartz monzonite to granite. Quartz diorite commonly does not form as large a part of plutons in the epizone as it does in those of the mesozone. Locally or in a few plutons the diversity may be due in part to incorporation, more or less in place, of country rock, especially in border or roof zones, but this is usually relatively unimportant. Roof pendants are common. Many of the plutons are effectively homophanous without lineation or foliation. Some have a primary linear structure, but well-developed planar foliation is uncommon and, where it occurs, is usually restricted to local border facies or is indistinct.

The orientation of lineation in the Jamestown, Colorado, granodiorite stock has been studied by Goddard (1935). He finds that the stock is elongated N.-S., that the lineation along the western part plunges about 70°–80° and in a re-entrant protrusion on the east has a gentler plunge of about 35°–60°. Grout (quoted in Calkins and Butler, 1943, p. 35–36) studied the lineation of the Alta stock in Utah and shows the linear structure plunging in general 80° or steeper in the border zone and gently in the core. The lineation of both stocks thus suggests steep upward flow in the border zones. Grout and Balk (1934, p. 885) find that lineation in much of the Boulder batholith is elusive, but most has a pitch of about 70°, and a steep conformable upward rise is suggested. Both the Alta and Jamestown stocks and also the Mount Princeton batholith (Dings and Robinson, 1957, p. 30) have associated dikes with gently plunging lineation suggesting subhorizontal flow. Other types of orientation of flow lines such as arches of flow lines and disconformable flow lines at an angle to the walls have been referred to by Balk (1937, p. 50–54, 60–63, 69–78). The arches *of flow lines may in some examples, at least,* be suggestive of an arched roof.

Moehlman (1948, p. 118) and others have referred to Tertiary plutons whose walls converge downward.

Volcanic rocks are commonly associated in close genetic relationship with Tertiary plutons, but they need not be with plutons of the deeper part of the epizone. Characteristically, at least, part of the volcanic rocks will have compositions comparable to that of the facies of the plutons themselves although the quantitative ratios may be different. Alper and Poldervaart (1957) have studied the Animas stock in New Mexico and the volcanic rocks it intrudes and have shown that not only is the chemical and mineral composition similar but the zircons of both the tuff and the pluton [illegible] similar habits.

[illegible] of the contact-metamorphic zones, may be relatively unmetamorphosed. If folded and regionally metamorphosed there may be independent evidence that the country rocks were only moderately warm and at shallow depths at the time of emplacement. Zoning of associated mineral veins on a regional scale is common, as are veins of epithermal or xenothermal character in the upper part of the epizone. Zoning of veins by repetitive introduction of solutions of diverse compositions during repeated structural reopenings is common. Peripheral outward deformation of the side walls is a feature of some epizonal plutons. It ranges from locally strongly deformed peripheral folds to gentle parallel peripheral folds; outward thrusting is inferred in one example; rarely there is a local thin zone of foliation or local thin layer of slight plastic crystalline flowage in contact-metamorphic zones or local minor drag folding. Most of the plutons are small, but there are also nevertheless many batholiths. The Paleozoic White Mountain batholith in New Hampshire associated with cauldron-subsidence origin has an outcrop area of 680 square miles, the Upper Cretaceous Boulder batholith an area of 1200 square miles, and the Tertiary Cordillera Blanca batholith of Peru is more than 75 miles long (Egeler and De Booy, 1954). The emplacement of stocks and batholiths associated with ring-dike complexes and cauldron subsidence is uniformly attributed by all authors to subsidence, either subcolumnar block sinking or block or piecemeal stoping.

Reference for comparative purposes may be made to the Cenozoic Slaufrudal stock 2 by 7½ kilometers in diameter, that cuts basaltic lavas with intercalated rhyolitic volcanic rocks in Iceland. Cargill, Hawkes, and Ledeboer (1928) describe the stock as consisting of miarolitic granophyre, in part with granitic texture. They suggest that the stock was emplaced by sinking of the replaced mass "en bloc" and that a distinct semihorizontal layering of the intrusion indicates intermittent subsidence, the stock growing by the addition of successive sills or caps. Relations are exceptionally well shown in steep topography.

Many other stocks and batholiths have a domical or a broad arch-shaped roof. This is commonly due in part to angular steplike transection of the roof to yield this kind of shape, in part to doming of the roof either by simple doming, or by doming accompanied by faulting in the roof due to distention.

The earlier members of a complex pluton, in small or moderate volume, will show chill zones against country rock. Dikes, apophyses, or small satellitic stocks related to large volumes of rock will commonly show chill zones or porphyritic characteristics. Many large masses or later members of a composite stock or batholith show no chill zones. There is often a set of late-stage aphanitic or porphyritic dikes. Associated lamprophyre dikes are also common. Distinct pegmatite veins are typically rare or minor although small pegmatitic nests may occur locally. Aplitic veinlets may be present, but aplite dikes or facies are not commonly abundant in the stocks. In some of the batholiths, however, aplite or equivalent alaskite may be well developed, as in the Boulder, Seagull, and Ackley (White, 1940) batholiths. Miarolitic structure is common, especially in leucogranite or alaskite, and it may have pegmatitic facies associated locally. Many aplite dikes are restricted to the border zone of the pluton where they occur both in the roof and in the adjacent igneous rock. Relatively flat-lying sheets of alaskite occur in the Seagull batholith, Yukon Territory, and of micropegmatite in the batholith of the Casto quadrangle, Idaho.

Granophyre may also occur locally as sheets, stocks, domical roof facies of stocks, or as metasomatized country rock. Granophyre, in general, occurs exclusively in the epizone.

Occasionally stocks of the epizone may be accompanied by satellitic laccoliths (Hunt, 1956; Strobell, 1956).

Emplacement predominantly by metasomatism is uncommon in the epizone and will be discussed under the title Pseudo-igneous Emplacement. Several plutons, however, do have an extensive aureole of granite or granite gneiss resulting from granitization of sandstone or metaquartzite. Contacts of pluton and country rock are normally sharp.

Oftedahl (1953, p. 71–74, 92–93) has interpreted the central nordmarkite and monzonite facies of the central part of the Sande stock in southern Norway as the product of assimilation of lavas by an ekerite magma more or less in place.

Such criteria as lack of contact metamorphism and contact metasomatism, lack of chill zones, and the presence of evidence for upward drag of wall rocks have been inferred (Read, 1951, p. 9–10; Tweto, 1951; Hunt, 1953, p. 165· Drewes, 1958, p. 233; Mackenzie, 1958, p. 69) to indicate that many plutons of the epizone have been emplaced as highly viscous magma at relatively low temperatures (600°C. or lower). This interpretation is satisfactory for many porphyritic intrusives with an aphanitic matrix, but we need more data to be sure it is appropriate for plutons of almost exclusively phaneritic rock. Most plutons of the epizone do have accompanying evidence of contact metamorphism and contact metasomatism. Andalusite is developed in shales and wollastonite locally in limestone at many contacts. Pyroxene intermediate between hedenbergite and johannsenite is not uncommon (Allen and Fahey, 1957). Tourmaline is common in many aureoles of epizonal stocks. Many epizonal stocks have at least local miarolitic facies, and they may be phaneritic throughout. These facts are consistent with the probability that the liquid phase of the magma of such plutons was relatively fluid because of its content of volatiles during part or all of its period of crystallization. An excellent comparison of the characteristics of highly viscous and less viscous magma emplacement in sills has been given by Tweto (1951).

In a few examples the smaller plutons are accompanied by a small amount of breccia whose origin is in part interpreted as an explosion breccia and in part as due to upward drag of magma. Examples are a breccia of slightly rounded fragments of sedimentary rocks and of porphyry, in a matrix of similar comminuted rock associated with a diorite stock in the La Plata district (Eckel, 1949, p. 39) and breccia zones on one side of a grano*diorite stock (Goddard, p. 383–384). Tweto* (1951, p. 526–528) has described an intrusion breccia as an advance guard of porphyry sills, formed by explosive intricutsion of fluids or tenuous magma. The breccia may consist of fragments of country rock and of chilled porphyry in a shale matrix or of dirty contaminated igenous material.

The magmatic origin of many salic dikes, stocks, and laccoliths emplaced in the epizone is indicated by the inclusions of deep-seated Precambrian rocks which they contain where emplaced in overlying beds. Examples are the quartz diorite porphyry laccolith described by Rouse (1933, p. 145–146) emplaced in Tertiary volcanic rocks with inclusions of Precambrian rocks which have been brought up for a minimum of 2½ miles, the inclusions of Precambrian rocks in monzonite-diorite porphyry stocks, sills, and sheets emplaced in

Mesozoic beds described by Eckel (1949, p. 34, 41), an example cited by Powers (1915, p. 166–168) in Vermont where bostonite dikes in Middle Ordovician shales contain inclusions of underlying Precambrian rocks, and similarly that by Buddington and Whitcomb (1941, p. 78–79) from New York where small laccolites and sills of quartz bostonite and rhyolite porphyry emplaced in Ordovician shales contain fragments of underlying Cambrian sandstone together with rare fragments of Precambrian basement material.

Some of the Tertiary laccoliths are so closely associated with volcanic rocks that they can confidently be considered to belong to a volcanic association. Hunt (1956, p. 43) writes concerning the laccolithic mountains of the Colorado plateau that "it seems likely that most of the larger stocks in the laccolithic mountains reached the surface and erupted, although probably none of them extruded any great quantity of lava or pyroclastic materials". A paper by Rouse *et al.* (1937) also portrays probable relationships between laccoliths and volcanic rocks. The major bodies originally described by Gilbert as laccoliths are now interpreted by Hunt (1956, p. 42–45) as the upper part of stocks, the latter up to about 2 miles in diameter. The Three Peaks laccolith, Utah, about 5 miles in diameter, has been studied in detail by Mackin (1947). He finds that the Upper Cretaceous (?) laccolith was emplaced under a cover that ranged from 2000 feet to a possible maximum of 8000 feet. The laccolith consists of quartz monzonite porphyry, generally holocrystalline but with some glass in the groundmass near contacts. He infers that the chilled borders and the glass prove that the mass was emplaced as magma. Some of the quartz monzonite porphyry is finely miarolitic.

The numerous diorite and monzonite porphyry sills of the La Plata district (Eckel *et al.*, 1949, p. 34) also belong among the volcanic bodies.

Granitic Stocks and Batholiths Associated with Ring Dikes and Cauldron Subsidence

Introduction.—Granitic stocks and batholiths associated in time and space with ring dikes and cauldron subsidences in direct relation to volcanic rocks occur in many different belts of different and widely spread localities [illegible] There are very pertinent to our

are also independent larger discordant plutons, associated with those directly due to cauldron subsidence, and inferred to be emplaced by block foundering or stoping. The prototype of this kind of complex is the Devonian Glen Coe cauldron subsidence and the associated Starav granite batholith described by Clough *et al.* (1909). Other examples of discordant batholithic intrusion following ring dike and stock emplacement are the Conway granite pluton of the White Mountain batholith complex (680 sq. miles, Fig. 2), the Drammen and other batholiths (Fig. 3), and the Jos-Bukuru pluton complex (285 sq. miles) of northern Nigeria (Jacobson *et al.*, 1958, p. 11; Pl. VII).

Billings (1943) writes that he found descriptions in the literature of 115 ring dikes and 30 ring-dike complexes. He states that 11 of the 30 ring-dike complexes have a central block of volcanic rocks that has subsided. These central volcanic rocks are flows and pyroclastic rocks ranging in composition from basalt through andesite and trachyte to rhyolite; they are, he believes, comagmatic with the rocks in associated ring dikes. He further states that 17 of the 30 ring-dike complexes possess what may be called a central stock, and the central stocks usually consist of quartz-bearing rocks, commonly quartz syenite or granite. Belts in Nigeria and Southwest Africa and examples elsewhere have been described since Billings wrote so that many more ring-dike complexes are now known.

Examples have been chosen from the literature to exemplify the foregoing principles. Before presenting these we might refer to some of the largest masses of acid volcanic rocks which are believed to be of magmatic origin.

The volcano-tectonic depression of the Rotorua-Taupo graben in New Zealand (Marshall, 1935) is about 60 by 15–20 miles in areal dimensions, with several thousand feet of depression and about 2000 cubic miles of welded tuff (ignimbrites) occupying a basin of approximately 10,000 square miles. This would be about equivalent to a batholith of 200 square miles and 10 miles depth. Again, Westervelt (1952, p. 565) has described a Middle Pleistocene rhyolitic tuff blanket in a fault trough in the Lake Toba area of North Sumatra covering 25,000 sq km and with a volume of 2000 cu. km. These are inferred to have formed as a result of the initial break-through of a comparatively shallow acid magma

depression. Ross (1955) has pointed out that pyroclastic rocks of rhyolitic, dacitic, quartz latitic, and some of latitic composition are present in many regions of the world, in volumes which dwarf many batholiths. Larsen and Cross (1956, p. 94) have estimated that the Miocene Potosi volcanic series of the San Juan volcanic rocks in Colorado contains 2300 cubic miles of rhyolitic volcanic rocks and 3000 cubic miles of quartz latite volcanics. The rhyolite volcanic rocks are the equivalent of a granite batholith 230 square miles in area and 10 miles deep. Stocks of monzonite and granodiorite emplaced in the same general period of time as the Potosi lavas are associated with them and range in size from necks up to plutons 2 by 5 miles in diameter.

Late Tertiary cauldron subsidence and intrusion.—The youngest cauldron-subsidence complexes may be expected to be among the least eroded and give the clearest evidence of belonging to a volcanic association.

Two such complexes from the United States will be described.

MEDICINE LAKE HIGHLANDS CALDERA, CALIFORNIA: Where erosion has not cut too deeply, cauldron subsidences occur at the surface as in the Pliocene(?)-Pleistocene(?) volcanic rocks of the Medicine Lake Highlands, California, described by Anderson (1941, p. 358–361). The caldera is an elliptical area about 4 by 6 miles in diameter in a shield volcano of platy olivine andesite about 20 miles in diameter. The rim of the caldera is outlined by nine volcanic vents from which platy andesite (±10 per cent normative quartz and ±61 per cent calcic oligoclase) has issued. The floor of the caldera is inferred to have sunk at least 500 feet through collapse of a central block coincident with ring-dike intrusion and the eruption of andesitic lavas squeezed up the marginal fractures. This hypothesis would necessitate a stock of magma below with the composition of a quartz diorite. Later lavas from the vents include olivine andesites, dacites, and rhyolites and are inferred to represent continued differentiation products discharged from later local vents in part spaced along the margin of the depressed block. This would imply in part the existence of magma below which could yield granitic masses on consolidation. The nature and tectonic relationships of these volcanic rocks may be considered as surface manifestations of ring-dike complexes and associated deeper stocks and batholiths.

SILVERTON CALDERA, COLORADO: Another

dera, described by Burbank (1941). The Silverton caldera is a minor unit areally and structurally of the volcanic field of the San Juan Mountains. The caldera was formed in the late Tertiary by gradual downwarping and faulting of a large shield-shaped block of crust about 8 miles in diameter. As downwarping of the basin became accentuated with thickening of the volcanic accumulations, ring faults and associated radial fractures developed. Intrusive bodies forced their way upward along certain more strongly accentuated regional rifts, and great numbers of smaller intrusive bodies and volcanic pipes penetrated the broken rocks of the fault ring. Burbank suggests that both the rock alternation and the concentration of intrusive bodies indicate that at moderate depths below the surface the margin of the caldera is underlain by a more or less continuous ring of intrusive rock. The intrusive rocks consist of gabbro-diorite, andesite, latite, quartz latite porphyry, and rhyolite. The volcanic rocks consist of andesite, "latites", quartz latite, and rhyolite. Larsen and Cross (1956, p. 227) describe one of the quartz monzonite stocks as 2 by 5 miles in diameter.

Early Tertiary cauldron subsidence and plutons.—Early Tertiary plutons associated with cauldron subsidence may be expected to include some which have been eroded to a deeper level than those of the Late Tertiary, and this is the fact.

QUITMAN COMPLEX: The following summary of the Early Tertiary Quitman complex is based on that by Huffington (1943). A series of lava flows ranges in composition from basalt to trachyte and rhyolite; rhyolites are most abundant. They are associated with pyroclastic rocks, and the whole has an approximate thickness of 3500 feet. Late basining, probably due to magmatic subsidence below the volcanic rocks, has dropped the central portion of the volcanic rocks approximately 4500 feet. A discontinuous elliptical ring of intrusives around the volcanic rocks is interpreted as a ring dike about 4 miles in diameter. The earliest intrusion in the area was a diorite; the ring-dike intrusion averages quartz monzonite. There is a related stock of quartz monzonite adjacent to the ring dike. The quartz monzonitic stock is subcircular with a diameter of about 3.5 miles and is separated from the parts of the ring dike by a septum about half a mile wide of Lower Cretaceous sedimentary rock. Belts of aplite and granite porphyry monzonite. Granites form less than 10 per cent of the plutons, and monzonite and syenite each about 10 per cent. The ring dike is inferred to have been emplaced in large part by stoping along the ring fracture.

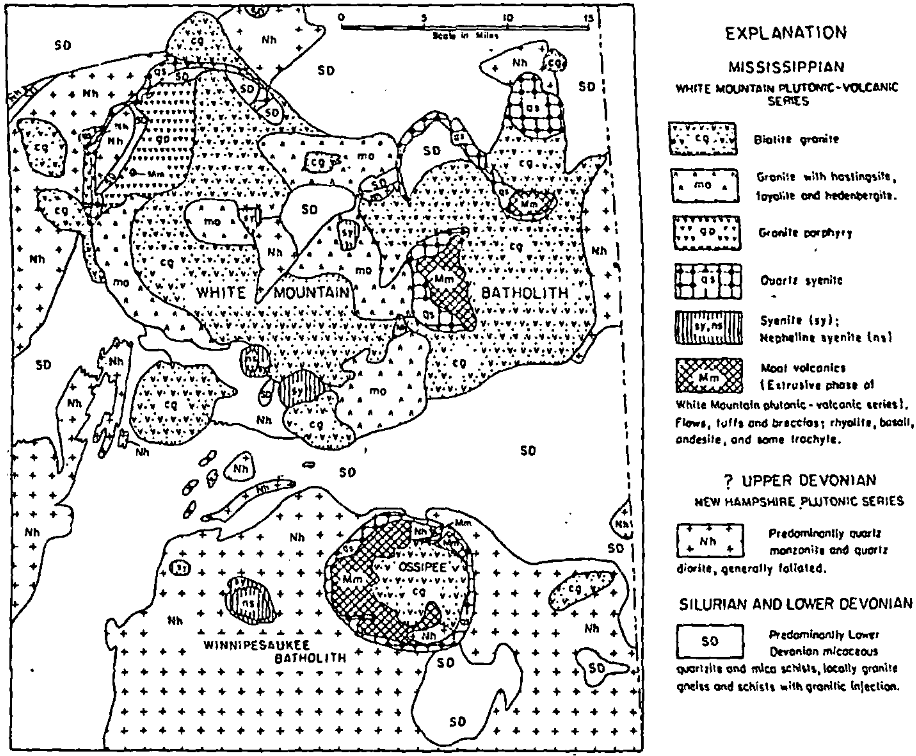

FIGURE 2.—NEW HAMPSHIRE BELT OF PLUTONS

Ossipee complex ring dike and biotite granite stock of cauldron subsidence; White Mountain batholith of cauldron subsidence and coalescing rim fracturing; both Mississippian (?) of epizone; and older Winnipesaukee batholith of transitional (?) mesozone-catazone; Modified after M. P. Billings, 1956; N. H. Planning and Development Commission, Division of Geol. Sci., Harvard University and U. S. Geological Survey

Permian, Mississippian(?), and Devonian plutons associated with cauldron subsidence.—Four great belts, three of them more than 200 miles long, each with numerous granitic stocks, batholiths, and ring-dike complexes in direct tectonic, geographic, and time relationships with volcanic rocks of related compositions, have been described in recent years. The belt of Tertiary ring dikes and caldera subsidences with associated granite stocks of Scotland should also be noted. The four belts include that of New Hampshire (Billings, 1945; 1956) of Mississippian(?) age, the Oslo district in Norway (Holtedahl, 1943; Oftedahl, 1953) of Permian age, a belt in Nigeria (Greenwood, 1951; Jacobson *et al.*, 1958), and one of late Karroo age in Southwest Africa (Korn and Martin, 1954). These plutons occur in the "epizone" of the crust and can approximately be considered to belong to a volcanic association. In addition to the annular dikes they range in size from small plugs and dikes to batholiths underlying as much as 680 square miles. The rocks of these belts may be thought of as representing in part levels of deeper erosion than the Tertiary plutons and in part the rise of large masses of granitic magma to relatively high levels.

Chapman and Williams (1935, p. 507) find that granite forms more than 78 per cent of the plutonic complexes of the New Hampshire belt, syenite and quartz syenite 20 per cent, and gabbro, diorite, and monzonite less than 2 per cent. Jacobson *et al.* (1958, p. 7) estimate that granite forms 94 per cent of the plutons of the belt in Nigeria and mafic to intermediate rocks only 6 per cent. By way of comparison granite and granophyre form 63 per cent of the area of intrusive rock in the belt of Tertiary plutonic complexes of the British Isles, gabbro and dolerite 33 per cent, and ultrabasic rocks 3 per cent according to Richey (1948, p. 55).

NEW HAMPSHIRE BELT OF MISSISSIPPIAN(?) PLUTONS: The ring-dike complex of the Ossipee Mountains, New Hampshire (Kingsley, 1931), has one of the most nearly perfect ring dikes of the New Hampshire belt and exemplifies the general character of the complexes (Fig. 2). The ring dike is described as subcircular with a diameter of a little more than 8 miles and is composed of porphyritic quartz nordmarkite (13 per cent normative quartz). Within the central complex is an arc-shaped mass of the Moat volcanic rocks. These have an approximate thickness of 7000 feet and consist of basalt (4 per cent normative quartz), andesite (13.4 per cent normative quartz), and quartz porphyry flows, and equivalent tuffs and breccias. The remainder of the central complex, except for a small block of country rocks, is composed of a coarse-gained biotite granite (25.8 per cent normative quartz). Kingsley estimates that the minimum subsidence of the volcanic rocks on the borders is 4500 feet and in the center 12,500 feet. The biotite granite is inferred to have been emplaced as a result of either piecemeal stoping or a columnar-block subsidence.

The White Mountain batholith (fig. 2), another member of the New Hampshire central complexes, is significant because of its size. The following description is condensed from that of Billings (1928). The batholith lies about 4 miles north of the Ossipee Mountains complex. It underlies about 680 square miles and consists predominantly of granite with subordinate nordmarkite and great blocks of volcanic rocks, from a few to 8 miles in diameter, that have settled into the batholith. The volcanic rocks consist of siliceous flow rocks and interbedded tuffs and breccias to a thickness of about 11,800 feet. The siliceous flows are largely comendites (24–33 per cent normative quartz) or quartz porphyries. Trachyte (18.3 per cent normative quartz) also occurs. The plutonic phases include a small mass of diorite, nordmarkite (4–13 per cent normative quartz), and granite (23.6–28.5 per cent normative quartz). Some nordmarkite porphyry occurs as small satellitic stocks or as chilled border facies whose groundmass is dense. It is noteworthy that the extrusive comendites are comparable to the plutonic granite and the extrusive trachyte to the plutonic more siliceous facies of the nordmarkite. The volcanic rocks seem to have settled at least 5000 feet and probably at least 17,000 feet in one area since they do not occur in the adjoining region. The emplacement of the batholith is inferred to result from roof subsidence. Magma moved upward along great fractures and issued at the surface as great flows and pyroclastic deposits.

The belt of epizonal plutons is slightly discordant to the trends of the older country rocks.

OSLO BELT OF PERMIAN PLUTONS: The Oslo, Norway, belt of cauldron subsidences or ring-dike and central complexes has been described by Holtedahl (1943) and Oftedahl (1952; 1953). It is of such general significance that the writer cannot refrain from including a summary here. (*See also* Figure 3.) The plutonic complexes were emplaced in a lava plateau consisting of 2000–3000 m of basaltic and rhomb porphyry flows. The plutonic phase began with the consolidation of larvikite plutons, the magma of which corresponds to that of the rhomb porphyries. During later periods, with the formation of syenitic and granitic stocks there occurred the cauldron subsidences. Quartz porphyry annular dikes occur in three of the four described cauldrons as an early phase of intrusion.

The Oslo belt is very instructive because of the association of large batholiths of larvikite, nordmarkite, and biotite granite which form a belt about 200 km long; the four major cauldron complexes lie within the central part of this belt (Oftendahl, 1953, Fig. 1). One of these batholiths, the Drammen biotite granite (Oftedahl, 1953, p. 103), encloses the Drammen cauldron in the form of a huge cylindrical block or roof pendant. The biotite granite magma is younger than the rocks of the cauldron, and the stoping followed the ring fault nearly exactly. It has a quartz porphyry border against the effusives of the Drammen cauldron. The Drammen pluton is about 55 km long. Against the rocks of the Giltrevann cauldron it has a border facies of quartz porphyry with an aplitic groundmass. Oftendahl (1953, p. 58) further suggests that the granite magma was at a relatively high temperature, perhaps superheated. An ekerite batholith is younger than the rocks of the Sande cauldron. The ekerite is full of pegmatite nests and has a porphyritic border zone. Holtedahl (1951, p. 90) concludes with respect to the mechanism of emplacement of the batholiths that "huge subterranean crustal blocks sank to an

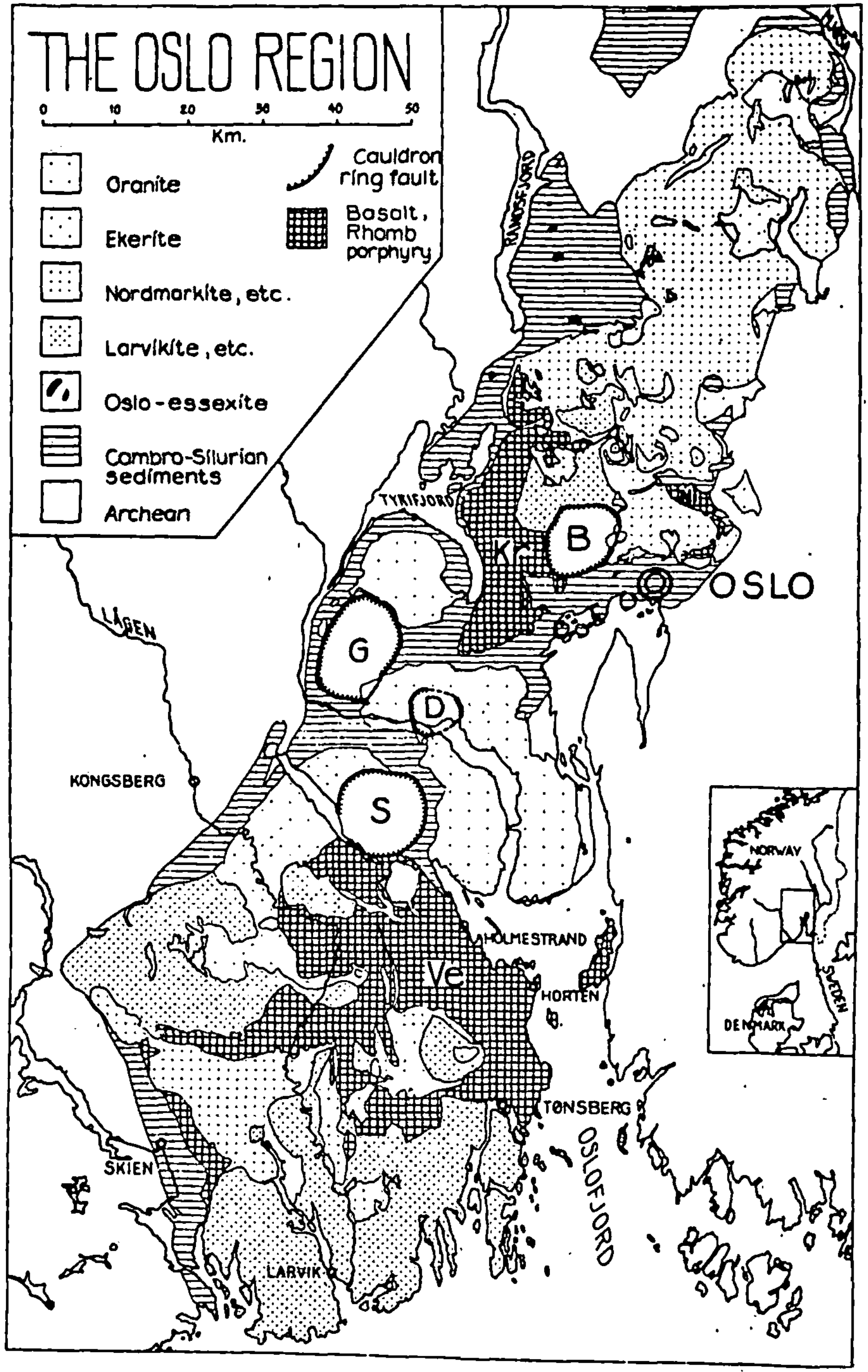

FIGURE 3.—OSLO REGION, NORWAY

Four Permian complexes (B, D, G, S) of lavas, ring dikes and stocks of cauldron subsidence, and associated younger discordant batholiths; all of epizone. After Oftedahl (1953)

unknown depth along curved fracture lines, with magma occupying the vacated space". Oftedahl (1952, p. 58) infers a monzonitic magma batholith at least 100 km long and 20-40 km wide, slightly superheated, to explain the rhomb porphyry lava flows. He further suggests (1952, p. 60) that stoping normally was arrested close to the surface (about 100-500 m below the surface), but occasionally the magma stoped its way clear to the surface to form areal eruptions.

Stocks Primarily by Cauldron Subsidence, but Accompanied by Outward or Upward Pressure

General discussion.—The preceding discussion has dealt with stocks that are assumed to have been emplaced effectively by subsidence of columns or blocks. There are a few examples in which subsidence of blocks or columns is inferred to have been the major mechanics of emplacement, but they are also accompanied by evidence of deformation of country rock due to outward side pressure or uplift of roof.

The basaltic lavas around the Mull complex (Bailey *et al.*, 1924, Chapters XII and XIII) show several concentric folds with dips of 10°-30° in discontinuous arcs or circles. Tyrrell's description (1928) of the northern granite mass of the island of Arran, Scotland shows that along one part of the border the country-rock schists have been dragged around by an uplift produced by the granite magma so that their strike swings into approximate parallelism. In another part of the border there has been upward movement by faulting and mylonitization.

Mt. Monadnock, Vermont.—The Mount Monadnock pluton in Vermont is inferred by Chapman (1954) to result from the settling of a domical reservoir which developed cylindrical and radial fractures with consequent cauldron subsidence and stoping of large arcuate-shaped slabs as the major method of emplacement. North and south of the stock, however, quartzites and schists which in general strike about north and dip east have been strongly twisted and their schistosity thrown out of regional strike. Chapman suggests that during the very early stages of intrusion positive magma pressure was sufficiently vigorous for a brief period of time to deform the immediately surrounding metamorphic rocks.

stock about 17 by 27 km in diameter, in Finland. The following summary is based on the work of Savolahti (1956). The earliest intrusions form an arc and consist of several successive intrusions of gabbro with anorthositic differentiates. The younger gabbro intrusions have chilled contacts against the older. The main core of the complex is a stock of biotitic rapakivi granite. The youngest intrusions are granite porphyry dikes with fine-grained facies in contact with all older rocks. Savolahti (1956, p. 83) writes of the rapakivi granite in general that "At an early period the hypabyssal, partly effusive characters of rapakivis were described. Likewise, the occurrence of miarolitic cavities was recognized to be typical, and the scarcity of aplites and pegmatites was known". The stock is in general discordant to the country rock which consists of migmatites and older microcline granite, but locally the trend of the foliation of the migmatites has been deformed into conformity with the boundary of the intrusive complex. Some of the rapakivi granites in this region have been determined by Kuovo and Gast (1957, p. 30) to be about 1650 million years old, on the basis of Rb-Sr, K-A, and U, Th-Pb.

Other Discordant Plutons of Epizone

Introduction.—The characteristic setting for discordant batholiths associated in space, time, and tectonics with ring dikes, cauldron subsidence, and stocks is a peneplaned surface on a "basement complex" or series of folded and faulted beds overlain unconformably by a veneer of lavas. Such plutons occur in the New Hampshire, Oslo, Nigeria, and Southwest Africa belts and the Glen Coe district that have been referred to. They may be very appropriately called subvolcanic. Although many stocks and batholiths are thus either directly or indirectly associated with cauldron subsidence and ring-dike complexes, many plutons of the epizone are not, although they are in general discordant. Many of these latter plutons are associated with volcanic rocks in space and time, but the evidence is commonly not sufficiently direct to tie both assuredly to the same tectonic history. Some of these plutons may not have had connections that broke through to the surface to yield lava flows. Others probably did have such connections, but the roof either remained intact or broke down and sank in the magma in such a way that direct connection be-associated with present evidence for caldera subsidence.

The Bayview and Packsaddle Mountain stocks, Idaho, described by Sampson (1928) possibly afford a link between the directly subvolcanic type of pluton and plutons not now associated with volcanic rocks. The stocks are predominantly granodiorite and are significant because probably subsidence of roof rocks has accompanied their emplacement. The stocks are discordant with the relatively flat lying structure of the surrounding rocks, which are predominantly the Precambrian "Belt" series of sedimentary rocks. Block faulting has resulted in downdropping of blocks of Cambrian sedimentary beds. The block faulting is found only about the stocks and is related to the intrusion. The Cambrian rocks are found only where they are more or less engulfed in igneous rock. The granodiorite is coarse-grained up to contacts. The collapse structure recalls that associated with cauldron subsidence. There is no pegmatite in the granodiorite, but there are a few aplite dikes.

An excellent example of control of emplacement of a Precambrian quartz monzonite stock of fractures of the country rock has been described and mapped by Steven (1957, p. 365–375). One border has many *angular* step-like irregularities, and another has a zone 1.5–2 miles wide with a complex network of dikes.

Several stocks or lines of stocks and several batholiths of the epizone will be summarily described. These may be either parallel or discordant to regional structure.

Stocks and volcanic rocks of western Cascade Mountains, Oregon.—The Miocene (?) lavas and the line of Miocene (?) intrusive stocks of the western Cascade Mountain range in Oregon (Callaghan, 1933; Buddington and Callaghan, 1936) afford an excellent example of an association of lavas and plutons of similar composition, and of association in geography and time (fig. 4). The belt about parallels the trends of the substructure. The lavas range from basalt to rhyolite but are characterized by andesite. The intrusive stocks in Oregon occur at intervals along a line about 200 miles in length. They range in size from small plugs to a stock 1½ by 2½ miles in diameter. The rock of the smaller bodies is generally porphyritic aphanitic, that of the largest body is even-grained granular. The rocks range in composition from augite diorite through augite dacite porphyry and augite granodiorite porphyry to granite. The larger masses are in general more siliceous. In the Bohemia district (Buddington and Callaghan, 1936, p. 426) the intrusions occur in an arc and as radial dikes. The line of intrusions has a general northerly trend, but the individual intrusions and veins trend mainly to the west or northwest. Epithermal and xenothermal metalliferous veins are associated with the stocks. An extension of this belt into Washington is shown on the map by Waters (1955, Pl. I).

A belt of Quaternary and late Tertiary volcanoes lies in general to the east and roughly parallel to the line of older Tertiary intrusive stocks.

Snoqualmie batholith, Washington.—The Snoqualmie batholith of Washington (Fig. 4) of early Miocene or late Oligocene age (Grant, 1941, p. 590–593) has been described by Smith and Calkins (1906). It is composed of granodiorite and biotite granite, is miarolitic (Waters, 1955, p. 711), and has porphyritic modifications on the margins of large masses and in dikes. It is emplaced in folded sedimentary beds and in the Keechelus volcanic rocks. The latter are in part gently folded and in part have only initial dips. They consist of pyroxene andesite, dacite, rhyolite, and basalt with the first two greatly preponderating. An analysis of a sample of the andesite shows 18 per cent normative quartz and is very similar in composition to the granodiorite which intrudes it. The batholith is about 10 miles in diameter and is inferred by Smith and Calkins to have consolidated about 4000 feet below the surface. Knopf (1955, p. 695) states that the batholith is roughly 250 square miles in exposed area and may be the top of a mass 4000 square miles in extent not yet uncovered by erosion.

Batholith, Casto quadrangle, Idaho.—The Miocene batholith described by Ross (1934) from the Casto quadrangle, Idaho, is also noteworthy because of its size. It is at least 30 miles long and averages 7 miles wide. Much the most abundant rock according to Ross is granite, but there are small masses of quartz monzonite, granite porphyry, dacite porphyry, quartz diorite, and granophyre. The batholith cuts Oligocene(?) volcanic rocks composed preponderantly of rhyolite and quartz latite, and Ross states that some of the late rhyolitic flows may possibly be related to members of the pluton. The depth of intrusion he infers to have been not much more than 2 miles. The granite is locally finer-grained close to the contacts and granite porphyry may be a marginal facies. Micropegmatite occurs in nearly horizontal ribs. The overlying strata were domed into broad low arch by the granite rather than crosscut. The granite locally has followed faults.

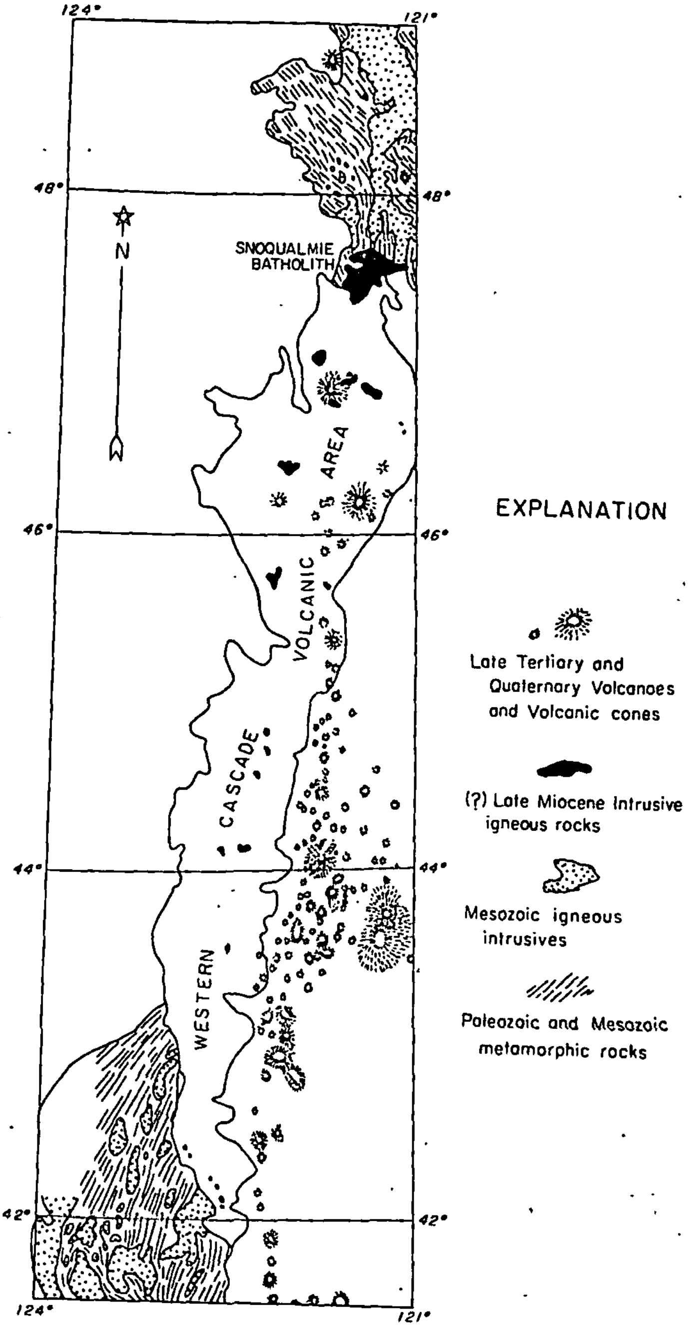

FIGURE 4.—BELT OF TERTIARY INTRUSIVE STOCKS EMPLACED IN VOLCANIC ROCKS OF WESTERN CASCADE MOUNTAINS IN OREGON AND WASHINGTON
Modified after Tectonic Map of United States, Am. Assoc. Petroleum Geologists, 1944

Boulder–San Juan discordant belt of plutons, Colorado.—A series of plutons of Tertiary age occurs in a belt extending from a little northwest of Boulder to the San Juan district in Colorado (Fig. 5), a distance of about 200 miles. The belt is strongly discordant to the regional

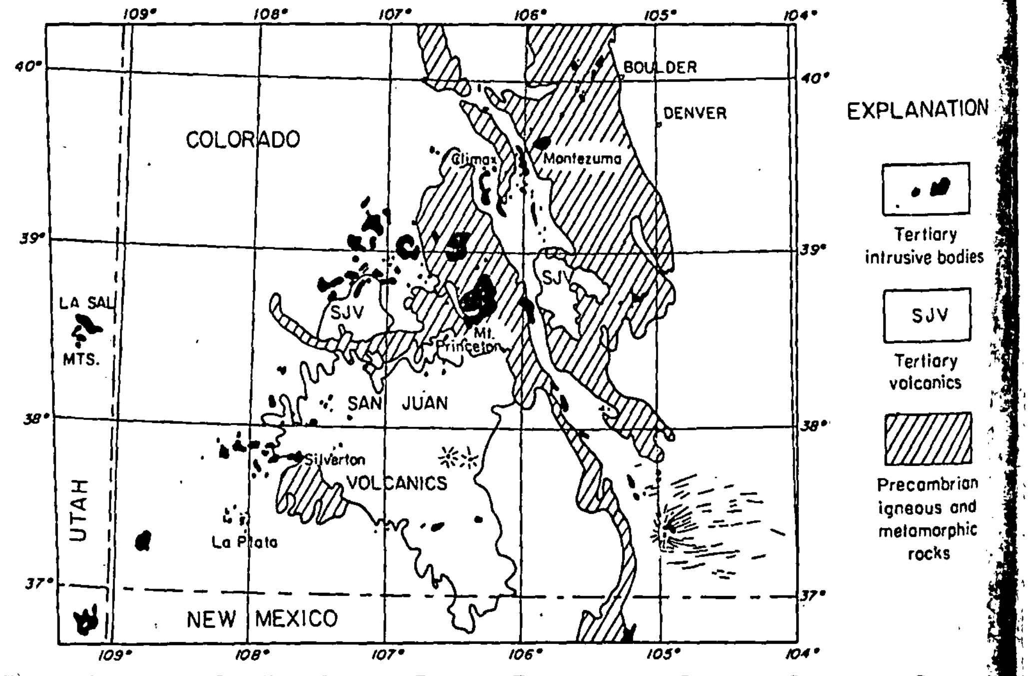

Figure 5.—Boulder-San Juan Belt of Plutons Discordant to Regional Structure, Colorado
Modified after Geologic Map of United States, U. S. Geol. Survey, 1932

structure. Among others it includes the Jamestown, Montezuma, Silverton, and La Plata stocks and the Mount Princeton batholith.

Lovering and Goddard (1950) have described the belt. They state that some stocks probably occupy old volcanic throats, but many others were roofed with pre-Denver (Upper Cretaceous and Paleocene) rocks and probably forced their way in by stoping and intrusive faulting. The earliest intrusives show structure concordant with the country rock much more commonly than do the later, and among the latest of the intrusives explosion breccia is common. Most of the intrusives lie between monzonite and quartz monzonite in composition. The stocks are associated with dikes and sills. The texture in general ranges from coarse porphyritic and medium-grained to porphyritic aphanitic. The intrusion continued intermittently throughout the considerable span of time during which the Laramide revolution was in progress. Local transverse fracture zones marked by intrusive activity followed a period of regional northwesterly folding and faulting. The transverse structure is explained as follows; (Lovering and Goddard, 1950, p. 63).

> lies perpendicular to the direction of Laramide compression, it is possible that tensional forces of some magnitude were present here during the folding of the region. The belt of porphyry stocks occupies a position on the northwestern side of a tectonic transition zone between two types of regional deformation. Fault movements in the porphyry belt suggest that the northern part of the transition zone was one of shearing with nearly horizontal movement as well as one of tension. It is believed that these two stresses—shearing and tension—were in part responsible for the rise of the magma that formed the porphyry stocks."

Dings and Robinson (1957) have described the Mount Princeton batholith which is one of the largest bodies in the belt. It is about 14 by 19 miles. They describe it as quartz monzonite with small areas of younger granite, intrusive quartz monzonite porphyry, and quartz latite porphyry. Only locally does the rock of the batholith change texture to a porphyry at the border, but apophyses are finer-grained. Small aplite dikes occur throughout. Pegmatite dikes are rare. The younger granite, a leucogranitic type, however, is locally miarolitic and has numerous pegmatite veins. Beryl occurs in the granite, the miarolitic cavities, and the pegmatite. There are two granite stocks, one

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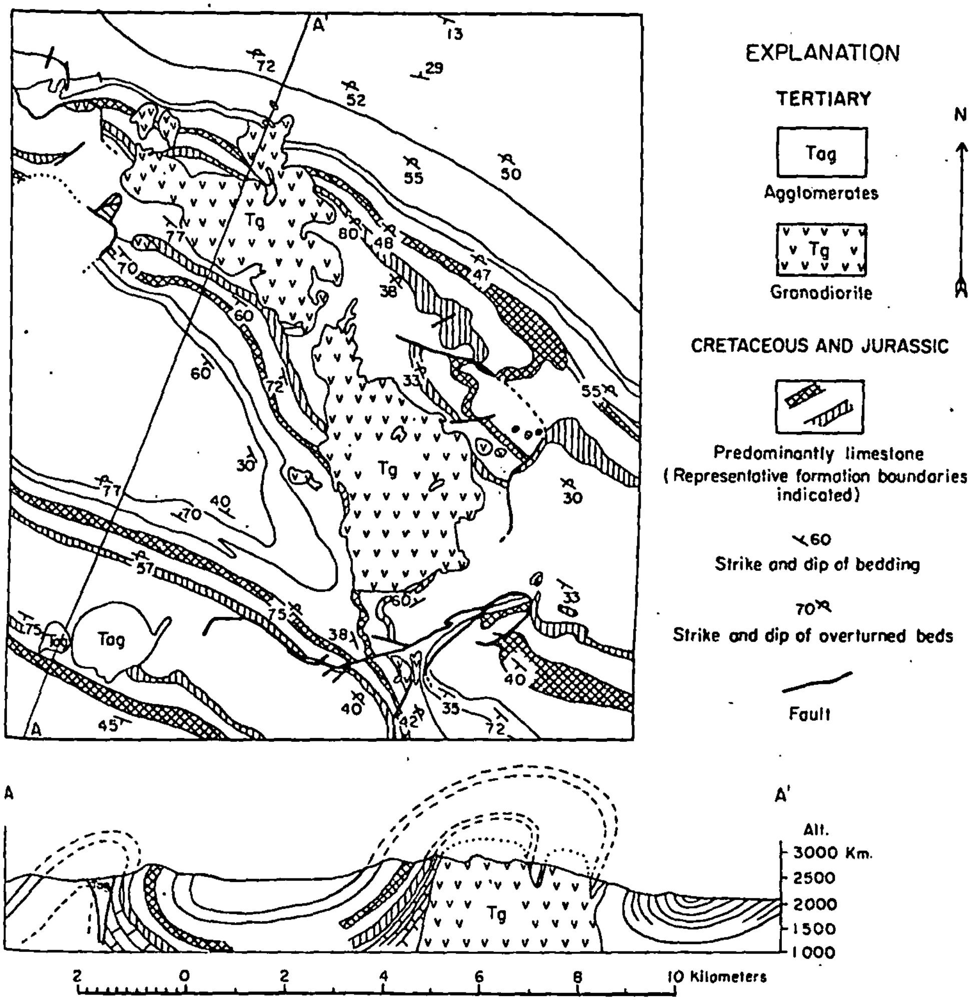

Figure 6.—Tertiary Granodiorite Stocks Emplaced Discordantly in Epizone
In asymmetrical anticline of limestone, near Concepcion del Oro, Mexico. After Rogers, Tavera, and Ulloa (1956)

The Tertiary Montezuma quartz monzonite stock is an example intruded discordantly in the epizone almost exclusively in Precambrian migmatites, schists, and gneisses (Lovering, 1935).

The mineralization and by inference some of the associated plutons of the Boulder–San Juan belt has been stated (Eckelmann and Kulp, 1957) to be 59 ± 5 million years of age.

Stocks of Concepcion del Oro, Mexico.—Two closely adjacent granodiorite stocks of Tertiary age are well exposed northwest of Concepcion del Oro in Mexico (Fig. 6). They have been significant because the great relief and mine workings permit an accurate picture of their shape and geologic relationships. They are emplaced discordantly in the core of an asymmetrical anticline of Jurassic and Cretaceous sedimentary beds. Limestones are predominant with minor shaly limestone. The roof is faulted in a manner which may be explained as due to a slight doming by upward magma pressure. Roof pendants lead to an inference of a relatively flat archlike roof with re-entrants. The emplacement seems clearly to be due to sub-considered a "resister" to granitization. The mineral deposits are zoned, in part with a vertical distribution, from hypothermal deposits at lower levels to epithermal chimney deposits in the higher parts of roof.

Boulder batholith, Montana.—A pluton of Late Cretaceous age, of great size for the epizone, and presumably somewhat older than the plutons of the Boulder-Breckenridge belts, is represented by the Boulder batholith in Montana. Knopf (1948, p. 666) states that a great volume of andesite and latite was erupted, probably just before emplacement of the Boulder batholith, which rose to such a high level that it invaded the pile of lavas. Balk (1937, p. 91) states that the Boulder batholith approached the surface somewhere between 2000 feet and 10,000 feet. The batholith is 70 miles long with an area of 1200 square miles. Grout and Balk (1934, p. 880) state that flow structures are poorly developed and (p. 885) that linear structures are much more widespread and uniform, and probably more significant than the foliation as indicating the intrusive movement, although they are also elusive.

Knopf (1957, p. 81) states that the batholith is composite and—that the order of intrusion is (1) basic hypersthene-bearing granodiorite, (2) granodiorite, (3) porphyritic granodiorite, (4) biotite adamellite, and (5) muscovitic biotite granite. Alaskite and aplite are abundant. Contact metamorphism has locally developed sillimanite - cordierite - microperthite hornfels. Knopf describes the emplacement of the batholith as the problem of how five different magmas in turn made room for themselves in the higher level of the crust and built up a composite batholith. He writes that near the batholith the invaded country rock has been more closely folded than at a distance from the contact. In places the strata adjacent to the batholith stand vertically and have even been overturned. Locally a series of reverse faults has developed along the eastern border of the batholith. The intrusive magma according to Knopf has manifestly made room for itself by crowding aside the enveloping rocks, by close appression of the beds, by overturning them, and by imbricate high-angle thrusting.

Knopf reports (p. 90) the age of the batholith as determined by the Larsen zircon method to range from 62–72 m.y. and by the potassium-argon method, 87 m.y.

Seagull batholith, Yukon Territory.—The Seagull batholith in Yukon Territory has been based on his work. The batholith is about 6 by 28 miles and is emplaced largely in the trough of a syncline of Paleozoic rocks (Fig. 7). The batholith could be of mid-Cretaceous or Late Cretaceous age. It is in a deeply dissected mountain country with great relief and can be shown to have steep walls and relatively flat undulatory roof with several of the mountain peaks capped with hornfels and the valleys in quartz monzonite. The country rocks are regionally metamorphosed and belong to the muscovite-chlorite subfacies. There is no evidence for forceful intrusion or side thrusting. The carbonate rocks have diopside, tremolite, and garnet in the contact zone, at one locality with wollastonite. The rock of the batholith is a coarse-grained leuco-quartz monzonite with flat sheets of fine-grained and porphyritic alaskite. Alaskite in near-horizontal layers up to 20 feet thick forms 5–25 per cent of the mass. There are miarolitic cavities in the quartz monzonite with quartz and tourmaline. Dense spherical aggregates of quartz and tourmaline also occur as replacements in the quartz monzonite and alaskite. There is no pegmatite.

Precambrian Plutons of Epizone

General statement.—The major types of epizonal plutons may be of Precambrian as well as of Tertiary age. Discordant plutons of Precambrian age have been described by Anderson, Scholz, and Strobell (1955) from the Bagdad area, Arizona, and by Kalliokoski from the Weldon Bay area, Manitoba. Other types of Precambrian plutons are referred to below.

Granophyre.—Extensive granophyre sheets associated with gabbroic or diabasic stratiform complexes have been described or redescribed in recent years from the Precambrian of Minnesota (Schwartz and Sandberg, 1940), Wisconsin (Leighton, 1954), the Wichita Mountains of Oklahoma (Hamilton, 1956; Merritt 1958), and the Sudbury complex in Ontario (Thomson, 1956). All these were emplaced in the epizone. Hamilton writes that in the Wichita Mountains, granophyre, granite, and rhyolite form a sheet complex of dozens of separate plutons, of which many are sills and funnel-shaped masses. He states that granite and rhyolite may be either lateral equivalents of granophyre or intercalations in granophyre, and that the granites are probably in general younger than the granophyres. Rhyolite inclusions are abundant in some

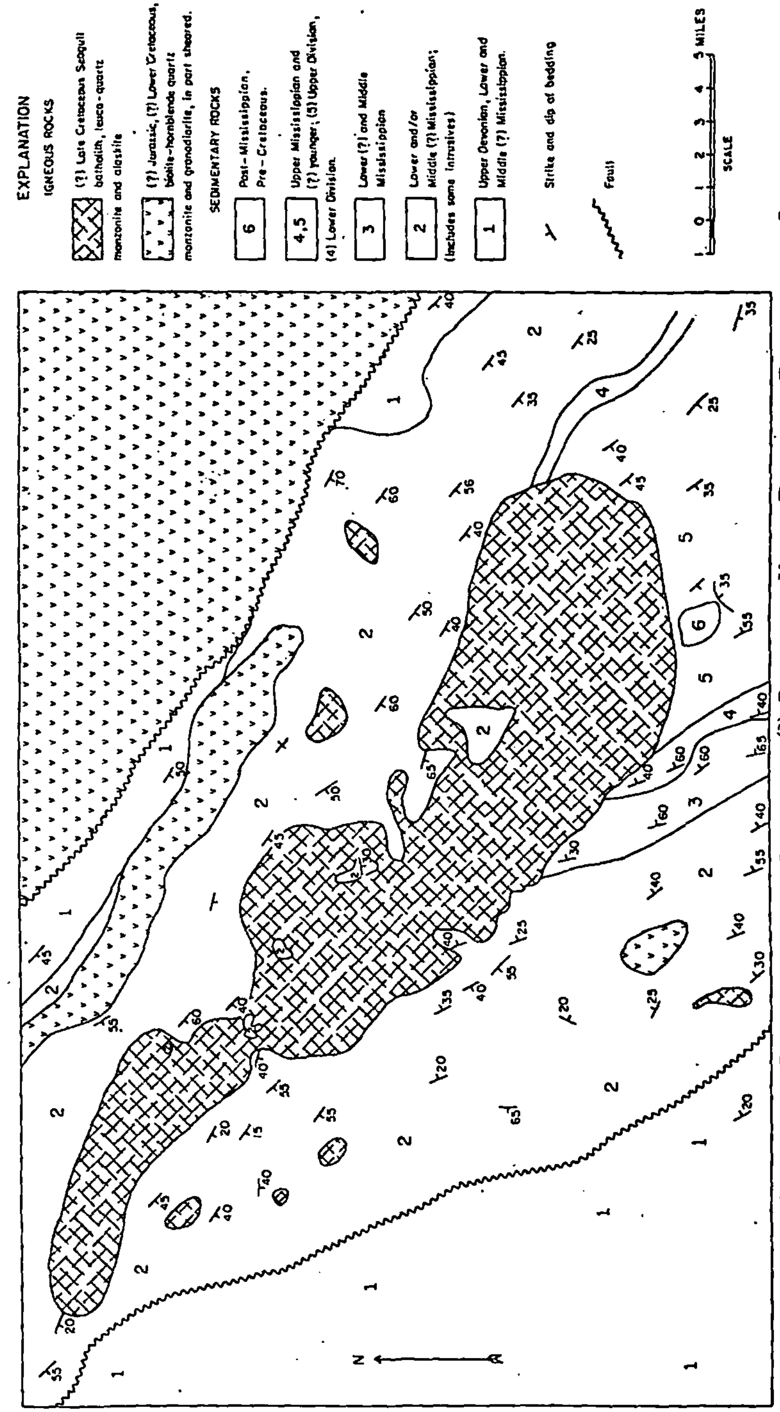

FIGURE 7.—SEAGULL DISCORDANT UPPER CRETACEOUS (?) BATHOLITH, YUKON TERRITORY, EMPLACED IN A SYNCLINE
The quartz monzonite is locally miarolitic. The roof pendants cap the tops of mountains and are underlain by granite. Modified after Poole (1955).

cavities are common in one of the granite masses and that it may have been emplaced as a batholith. A younger granite mass has chilled facies against older rocks. The younger granite is also miarolitic. K/A and Rb/Sr ages of a biotite from the younger granite are reported by Merritt (p. 62) to be 480 m.y. and 500 m.y. respectively.

Thomson (1956, p. 43–45) has proposed that the Sudbury Basin is a volcano-tectonic depression surrounded by a ring complex of dikelike and sill-like character, and that the granophyre (micropegmatite) was emplaced as a ring structure inside the norite.

Ring-dike complexes.—The Ahvenisto pluton of Finland has already been referred to as an example of a Precambrian ring-dike complex. Two other ring-dike complexes, the Chatham-Grenville and Rigaud stocks of probable Precambrian age in Quebec, have been described by Osborne (1934). Chill facies such as quartz porphyry and syenite porphyry are found in both complexes, and miarolitic structure is found in the syenite of the Rigaud stock.

Aureoles of Pseudo-igneous Emplacement

Description.—The contacts of the plutons of cauldron subsidence are almost universally sharp as are the contacts of most other stocks of the epizone. Extensive metasomatism of earlier gabbro members by younger granitic magma, however, has been described by Korn and Martin (1954) from the Messum complex in southwest Africa. The Bingham, Cassia and La Plata plutons in Utah, Idaho and Colorado respectively have been interpreted as cores of magmatic origin with aureoles of pseudo-igneous granite, the product of replacement of quartzite or sandstone. Loughlin and Koschman (1942, p. 41–42) have also described a small body of granophyre interpreted to result from metasomatism of sandstone by emanations from an adjacent Tertiary granite body.

Bingham, Utah, stock.—Stringham (1953) has described a small stock at Bingham, Utah, where granite forms about two-thirds of the area of the pluton and is inferred to be the product of granitization of quartzite. Some of this granite is exceptionally high in K_2O. The core of the stock is granite porphyry which is interpreted to be of magmatic origin. In the area to the south (Gilluly, 1932, p. 65) a series of volcanic flows in which latite is overwhelmingly predominant are cut by intrusives inferred to be roughly correlated with the Bingham Oligocene age.

Cassia batholith, Idaho.—Anderson (1934) has described the present 60 square miles exposure of the Cassia batholith to consist of about two-thirds porphyritic, usually gneissic, granite and one-third granodiorite. He believes the porphyritic granite gneiss is a replacement of metaquartzite and the granodiorite a crystallized magma. The batholith has a dome-shaped roof, and the metasomatic porphyritic granite extends at least 1800 feet down. The batholith is of late Cretaceous or early Tertiary age. The data are not too definitive, but the batholith is here classified as emplaced in the epizone.

La Plata stocks, Colorado.—Eckel (1949) has inferred the emplacement of small Late Cretaceous or Tertiary diorite and monzonite stocks in the La Plata District, Colorado (Fig. 5) to be in part by replacement or assimilation of country rock. He finds (p. 39) that where a monzonite stock transects beds of sandstone there are inclusions from an inch to several hundred feet in length which retain the attitude and position of the beds from which they were derived. However, he also finds that in many places contacts between the diorite and monzonite and the host rock are sharp and that the diorite and monzonite locally contain fragments of Precambrian rocks even where they are in Paleozoic or younger sedimentary beds.

Supplementary Descriptions of Plutons Emplaced in the Epizone

New Cornelia quartz monzonite stock, Arizona.—The New Cornelia quartz monzonite stock in the Ajo district of Arizona exhibits many features characteristic of the Tertiary stocks of the southwestern United States. The following description of it is condensed from that by Gilluly (1946).

The great copper mine at Ajo is opened in a low-grade epithermal deposit of chalcopyrite and bornite disseminated in the New Cornelia quartz monzonite that is tentatively referred to Early Tertiary age. The New Cornelia stock is exposed over an area of about 6 square miles. There is a discontinuous border facies consisting of fine-grained quartz diorite. The preponderant rock is an equigranular quartz monzonite with a poorly developed linear structure. The country rock is predominantly the Concentrator volcanic rocks consisting of andesite, keratophyre, and quartz keratophyre flows, breccias, and tuffs with highly altered and complex structure. The Cornelia intrusion Gilluly infers probably took place under a moderate over of rocks after folding and perhaps faulting of the Concentrator volcanic rocks (Cretaceous ?). He states that it is possible that the Cornelia quartz monzonite and Concentrator olcanic rocks represent the same magmatic ycle. The disregard of older structures by the ntrusive, the absence of wall-rock structures oncordant with its contacts, and the sporadic reservation of fine-grained (chilled) border acies indicate that the intrusion took place at relatively shallow depth. The weakness of neation in the rock is inferred by Gilluly to ndicate that there was little motion in the magma at late stages of its consolidation. After ts consolidation, however, it was fractured long westward-trending fissures, and aplite ikes were injected in large quantities. Northard-trending fractures followed and were ccupied by pegmatites in the apical part of e stock. At later stages solutions chloritized nd sericitized the rocks of the apical part of e stock and deposited cupriferous and assoated metallic minerals of the New Cornelia e body.

Hanover stock, New Mexico.—The Hanover anodiorite stock, New Mexico, has been deribed by Paige (1916), Schmitt (1933), and err *et al.* (1950). Where emplacement has been ccompanied by uplift and outward deformaon. It is about 2.5 miles long, north to south, nd less than a mile wide. The Paleozoic bedded cks dip away from the stock on the east and est flanks so that it has the relationship of an nticlinal structure. At the south the bedded ocks have been deformed into an overturned ncline and an asymmetrical anticline beyond. he fold axes form arcs parallel to the edge of e stock as though the folds had been formed y lateral pressure from the overriding stock. ocally folding was so intense that thrusting ccurred on a low-angle fault plane. Kerr *et al.* 950, p. 301–302) state that the near-by nta Rita stock arched and cut through overing sedimentary rocks and quartz diorite lls.

Marysville stock, Montana.—The development by Barrell (1907) of the hypothesis of ock subsidence in magma as the mechanism emplacement of the Marysville granodiorite ock has made the latter one of the classic amples of this phenomenon. Knopf (1950) rites that the stock is 6 miles north of the orthern end of the Boulder batholith and obably an outlying cupola. It is only 3 square iles in area. Barrell states that the roof sedients were domed over the granodiorite to the tent of 1000 feet, possibly 3000 feet; that faulting immediately preceded the invasion of the stock caused by upward pressure of the igneous mass; and that at numerous places the roof of the stock passes beneath the sedimentary cover at a flat angle. Barrell further writes that the granodiorite maintains granularity up to the contact and is medium- to coarse-grained but becomes markedly more porphyritic in outlying tongues and wedges. Aplite occurs within the margin of the stock and in rocks of the border zone but is rare in the interior. There is also minor pegmatite in the same zones as the aplite. Knopf comments (5950, p. 840–842) on the remarkable contact-metamorphic aureole. The argillites are converted to cordierite hornfels, and the limestone to diopsidic and tremolitic hornfels.

Organ Mountain batholith, New Mexico.—A small discordant Tertiary batholith of about 55 square miles in area has been described by Dunham (1935) from the Organ Mountains of New Mexico. He infers that Tertiary andesite flows form the roof for the intrusive comparable with the roof of a laccolith and that in depth the body is crosscutting with steep outward dips and was emplaced by piecemeal stoping. There is evidence that some xenoliths have sunk not less than 1400 feet and probably much more. The dip of the andesite roof is away from the batholith, exceeds 50°, and swings around with the contact. Locally the wall rocks have been powerfully distorted by magmatic pressure. The batholith is composite and consists of three distinct bodies, a monzonite, quartz monzonite and quartz-bearing monzonite. The quartz monzonite has tiny miarolitic cavities as a widespread feature of a porphyritic fine-grained facies. The intrusive process according to Dunham was accompanied by a progressive concentration of volatile fluxes. There are no pegmatites or aplites in the monzonite; aplites but no pegmatites in the quartz monzonite (17.5 per cent normative quartz) and aplites, pegmatites, and mineral veins in the quartz-bearing monzonite (10 per cent normative quartz). The aplites are very small dikes and veinlets. Quartz porphyry sills and dikes occur in the country rock (Dunham, 1935, p. 84). Wollastonite (p. 100) has been formed locally in the limestone contact with the late intrusions.

Paleozoic leucogranite and granite porphyry batholiths, Newfoundland.—Two examples of Paleozoic miarolitic leucogranite or alaskite batholiths emplaced in the epizone have been described by White (1940) and by Van Alstine (1948), both from the south coast of Newfoundland. The roofs of the batholiths are relatively flat, in part covered with roof pendants, and have gently outward dipping walls. The granite is miarolitic and homophanous. Chilled contacts are rare but were observed. The St. Lawrence batholith (Van Alstine, 1948) of Devonian (?) age has related rhyolite porphyry dikes in the country rock and epithermal fluorite deposits. It is elongated in a direction approximately normal to the trend of the major folds and thrust faults of the Cambrian and Ordovician (?) rocks and is inferred by Van Alstine to have been emplaced by stoping. The Ackeley batholith (White, 1940, p. 969) is believed to occupy more than 300 square miles. A miarolitic alaskite facies is associated with molybdenite and muscovite metasomatism. Locally where the batholith is bordered by volcanic rocks there is an agmatite zone 3 miles wide.

Another discordant Devonian (?) batholith of the epizone has been described from the north coast of Newfoundland by Snelgrove (1931, p. 24–25), Baird (1951, p. 49–52), and Neale (1957). The batholith occupies at least 75 square miles in area southwest of Confusion Bay, west of Notre Dame Bay. The rock is described as a granite porphyry or quartz porphyry. Foliation and lineation are marked in marginal facies. Many inclusions of Devonian (?) rhyolitic volcanic rocks occur locally in the border facies. Neale (1957, p. 59) suggests that the volcanic rocks and the prophyry are intimately related, and all the authors infer shallow intrusion. So large a batholith of granite porphyry is very unusual, and it deserves additional detailed study.

PLUTONS OF TRANSITIONAL(?) EPIZONE–MESOZONE

Introduction

There are several batholiths whose description suggests characters appropriate in part for the epizone and in part for the mesozone. They are therefore here included in a transitional (?) group until more definitive classification can be made. The Texas Creek granodiorite batholith (Buddington, 1929a), a unit of the Coast Range intrusives of Southeastern Alaska, is representative of the kind of problem involved. The batholith has a sharply discordant broad relatively flat roof, common aplite and pegmatite dikes in the contact zones only, and associated porphyritic aphanitic dikes. These [illegible] with the epizone. On the other hand it has a foliation throughout and is emplaced in close-folded country rocks—characters appropriate for the mesozone.

Examples of plutons with transitional characteristics appear to be not uncommon in Europe. Several late Hercynian granite complexes in Portugal described by Westerveld (1955) are post-tectonic and almost wholly discordant, but also have steeply dipping planar foliation in substantial part parallel to the borders of the pluton, but locally at an angle.

Southern California Batholith

The huge batholith of Southern California displays predominant discordant contacts and certain other features which ally it with plutons of the epizone, whereas the near absence of chill zones, and the occurrence of internal border foliation and local concordance of structure of country rock to contacts tie it to batholiths of the mesozone.

The northern part of the batholith of Southern California has been described by Larsen (1948), from whose work the following summary is made. The batholith is exposed for a length of 350 miles and a width of about 60 miles. It has a length of probably over 1000 miles if discontinuous bodies at the southern end are included. The batholith was intruded in early Late Cretaceous time. In the area studied by Larsen the batholith was emplaced by more than 20 separate injections. The country rocks were regionally closely folded, metamorphosed, and intruded by earlier granitic rocks, perhaps in late Paleozoic and Triassic sediments, the orientation of the inclusions and other structures of the batholith, the elongation of the batholith, and the strike of the major faults are in about the same direction. Larsen infers that the batholith was emplaced by stoping and not by forceful injection. Forcing apart of the walls may have been important in furnishing room for some of the elongate members of the batholith, but it could not have furnished a large part of the space for the batholith as a whole. He concludes that there is no relation between proximity to granitic bodies and degree of metamorphism, except for local contact metamorphism and for thin screens between intrusive masses and small inclusions. For the most part there is little contact metamorphism where the granitic rocks intrude large bodies of older schists, slates, and quartzites. Some of the screens and inclusions in the granitic rocks, however, have been greatly metamorphosed. The slates were changed to mica schists and to injection schists, the quartzites to mica-garnet-quartz rocks or to quartz-sillimanite rocks. Local bodies of marble around tonalite have been replaced by garnet, diopside, wollastonite, idocrase, feldspar, etc. Many of the intrusive bodies have zones next to their contacts that are banded or gneissoid. Gabbro forms about 7 per cent of the batholith, tonalite 63 per cent, granodiorite 28 per cent, and granite 2 per cent. Larsen is of the opinion that in the area described the rocks had only a moderate temperature when the first member, the San Marcos gabbro, was intruded, as small bodies of that rock are rather fine-grained. Locally aplite dikes are miarolitic. Larsen does not discuss the possible depth at which the present level of the Southern California batholith was emplaced but states (1945, p. 404) that it was probably intruded to within a few kilometers of the surface. Chayes (1956) is of the opinion that the tonalitic rocks are the product of a mechanical mixture and interaction between granodioritic magma with previously solidified gabbro. In most places contacts of the intrusive bodies are sharp. Chilled borders were not found except in one granodiorite dike. Merriam (1946) describes concordance of structure of country rock with batholith contacts in the Ramona quadrangle.

Batholith of Southwestern Nova Scotia

The geology of the great batholith of southwestern Nova Scotia (Fig. 8) has been summarized by Wright (1931). It is more than 110 miles long, 20–30 miles wide, and together with satellitic bodies has an area of 4000 square miles. The eastern part consists of biotite and muscovite granite that rarely shows gneissoid or banded structure. Aplitic and pegmatitic facies are common, and the texture of the granite is maintained up to the contacts. The contacts are sharp. Andalusite hornfels has locally been formed in adjoining slate. The country rocks are predominantly folded late Precambrian chloritic and carbonaceous slate and sandstone. The beds are in broad folds which are transected by the batholith, without appreciable distortion or deflection. The invading magma is inferred by Wright to have displaced its host without appreciable lateral thrusting or doming of the roof. Mesothermal gold-quartz veins occur in the country rocks. Wright quotes Fairbault as estimating that 9 miles of the Precambrian Goldenville formation had been eroded, mostly before Mississippian time. Locally the batholith is reported to cut fossiliferous beds of Silurian and Devonian age. Most of the data of Fairbairn (1957), however, based on rubidium-strontium ages of mica suggest ages of between 350 and 400 million years, or between Middle and Late Ordovician. These discrepancies in ages are at present unresolved. The characters of the batholith seem to be definitely those of the epizone, but it is not certain what significance is to be drawn from Wright's reference to 9 miles of eroded material. The axis of the batholith is in part strongly discordant to the fold axes of the regional structure. The discordant Boulder–San Juan (Colorado) line of plutons, described earlier, may be referred to as a much "nearer-surface" expression of similar relationship.

PLUTONS OF MESOZONE

"The granite was once hot, full of gas and molten . . . it rose along a large broad front, stretched out and expanded sideways and yielded to a force from the depth which pushed and drove it." Hans Cloos, 1953.

Introduction

It may be expected that a substantial period of time would be necessary to permit erosion to expose plutons of the mesozone. The fact that no plutons of Tertiary age in North America are known to the writer to be emplaced in the mesozone is consistent with the foregoing principle. In the western Cordillera, Jurassic to Lower Cretaceous plutons, however, are dominantly of mesozonal character.

Characteristics

The individual plutons of the mesozone are inferred normally to have the following characteristics. The degree of metamorphism of the regional country rock is not more intense than the green-schist and epidote-amphibolite facies. The argillaceous country rocks of sedimentary origin are usually slates and phyllites. The inferred temperature of the country rock at the time of intrusion is generally no higher than 400°–500°C. There is no apparent direct relationship between the plutons and volcanic rocks. The stocks and batholiths are effectively always of composite character made up of two or more units. The units in general vary systematically; the younger intrusions are more alkalic and siliceous. The characteristic plutons have *complex* emplacement relationships to the country rock—*in part discordant, in part concordant*. Locally there may be some replacement: Some may show transection of roof in archlike shape, and have steep outward flaring walls in depth. Rarely, a mass such as the Sugar Hill pluton in Vermont (Doll, 1951, p. 44) has crowded a series of uniformly north-northeast trending metasediments into a conformable steep dipping funnel. Planar foliation is often well developed, especially in the outer portions of the pluton, but commonly is local, elusive, or missing in the core. The planar structure of characteristic batholiths and stocks, especially in the border zones, is subvertical, consistent

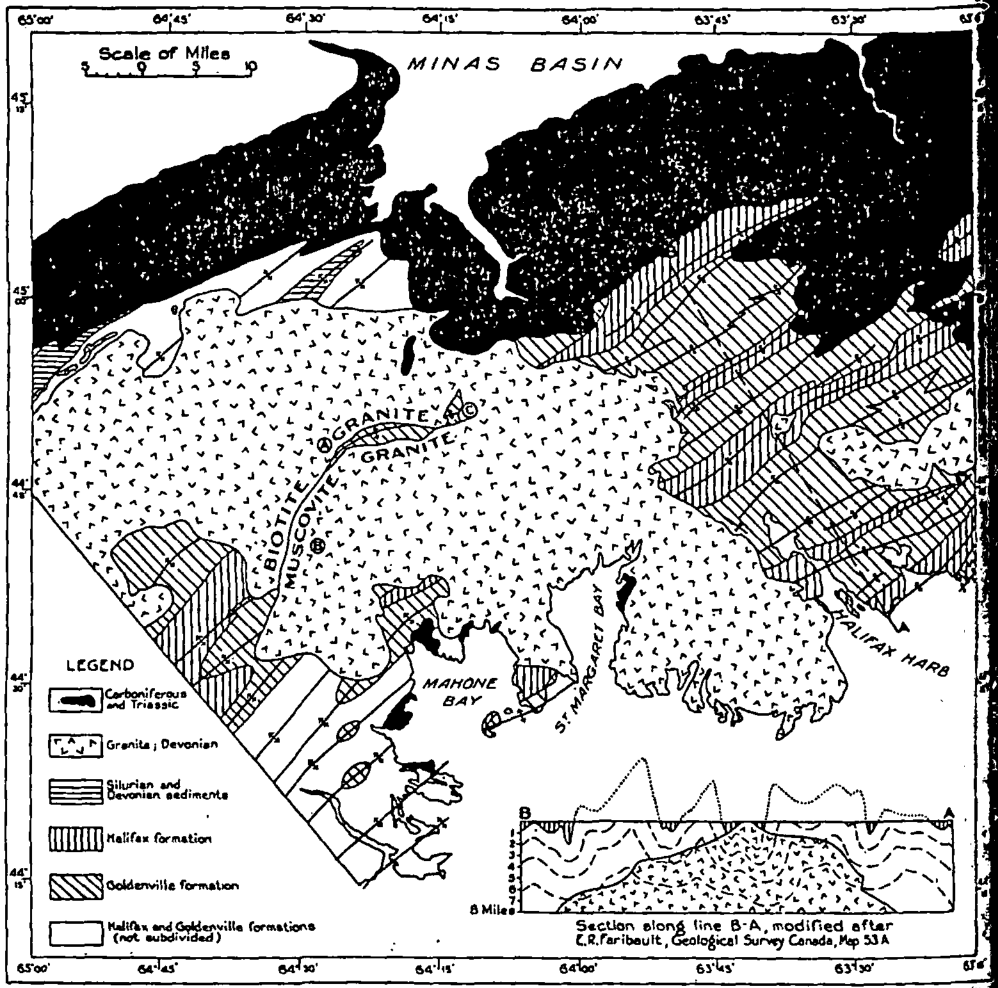

FIGURE 8.—GREAT DISCORDANT DEVONIAN (?) GRANITE BATHOLITH OF SOUTHWEST NOVA SCOTIA
Emplaced in lower part of epizone or upper part (?) of mesozone. After W. J. Wright (1931, modif after E. R. Fairbault)

ally, gentle dips in the core or parts of the co may indicate a domal roof. The planar structu of some plutons may suggest a rough funn structure either right side up or upside dow with the axis vertical or at moderate inclin tions; often the structure may be as shown i Figure 9. Younger units may crosscut the foliation of older units, or locally the planar flow structure may be independent of boundaries of units within a composite pluton (Fig. 9) and maintain its swing across their contacts. The planar structure in some plutons is in part systematically oriented at an angle to the oute border of the pluton. Linear structure ma

core of the pluton without planar structure. similation may be significant in border or of zones. Wall rocks in the contact zones often ow the development of a steep schistosity nformable with the contact and with a linear ructure more or less parallel to the dip, in- cating flowage in a subvertical direction. plift of the roof may be inferred. Minor fold es and lineation of country rock adjacent to pluton may have steeper plunges than fur- r away (Trefethen, 1944, Pl. 1). Bedded cks or dikes at large angles to the contact y be crumpled back on themselves as though formed by outward pressure from the pluton. dded rocks or older sills parallel to the con- ct may show boundinage structure due to dis- ntion.

Emplacement by reconstitution and replace- ent of country rock is commonly either absent subordinate. An occasional small pluton, wever, may be emplaced wholly by replace- ent. Chill-border facies, in the sense of apha- tic texture, are absent. Typical migmatites e commonly minor and are often absent. Peg- atites and aplites, however, may be common, pecially in border zones. They may have a dial fabric in some plutons. Miarolitic struc- re is absent. Marginal fissures with inward ps, in part lined with aplite or pegmatite, y be present locally in border zones of plu- ns. The plutons are pre-eminently those where e applicability of Cloos' system of "granite ctonics" is most rewarding.

Some batholiths may be bordered by a zone dike injection or contact agmatites. Dikes ig. 13) or dike systems (Pl. 1) as well as sills y occur in the country rocks bordering the trusives. Hutchinson (1955) has shown that the Ross Lake area, Northwest Territories, nada, pegmatite veins of an area of several uare miles are spatially related to a body of anodiorite. In the hottest zone near the gran- they are often large and emplaced by gran- zation of granodiorite layers whereas in a oler zone they were emplaced as pegmatitic ids along dilatant fracture zones, in part ith the development of zoning of the "com- ex" type of pegmatite.

Contact-metamorphic aureoles may be well veloped around stocks (Philbrick, 1936; tcher and Sinha, 1958) and small batholiths d around areas of dike complexes in roof nes (Eric and Dennis, 1958, Pl. 1). In the rger or relatively deep mesozonal batholiths gional metamorphism is associated. Phyllites canic rocks or metalimestones in the vicinity of contacts. A schistose structure is common in the country rock bordering mesozonal plutons.

Mesozoic Plutons with Both Discordant and Concordant Relations to Country Rock

White Creek batholith, British Columbia.—A late Mesozoic batholith described by Reesor (1954), from British Columbia shows clearly many significant phenomena (Fig. 9). The country rock consists of the Proterozoic Lower Purcell series regionally recrystallized to phyllites, little deformed in the eastern part of the area but compressed into isoclinal folds in the western part. The rocks belong to the greenschist metamorphic facies. The batholith is about 12 to 17 miles in diameter. There is a contact-metamorphic aureole of about 1000 feet diameter, and all the rocks of the contact zone are schistose. The batholith crosscuts a north-plunging anticline of Proterozoic rocks. Although the batholith in part truncates preexisting structures, the bordering strata are generally visibly and violently side-thrust and forced to conform with the direction of the contact. Vertical isoclinal folds occur along the north and south borders. In one area a marked cross cleavage has developed; fracture cleavage has developed in competent beds, crenulations with cleavage parallel to axial plane in phyllites, and a planar foliation across the bedding. A vertical upward motion of the magma during emplacement is indicated by the vertical to near-vertical lineation and by joints. Evidence of stoping is indicated by an almost universal irregularity of the contact of the pluton with country rock and by local stoping. Inclusions occur everywhere in varying amounts, except within the interior granite. The rocks of the batholith consist, in order of their emplacement, of biotite granodiorite, hornblende granodiorite, and porphyritic granodiorite in roughly concentric arrangement with an elongate core of quartz monzonite. The quartz monzonite grades into an aplitic facies which cuts the other rocks. Aplites and pegmatites occur in great abundance in parts of the mass, notably in the porphyritic granodiorite but not in the quartz monzonite. Inward-dipping joints are well developed throughout the batholith, and many are filled with aplite or pegmatite. They rarely extend into the sediments

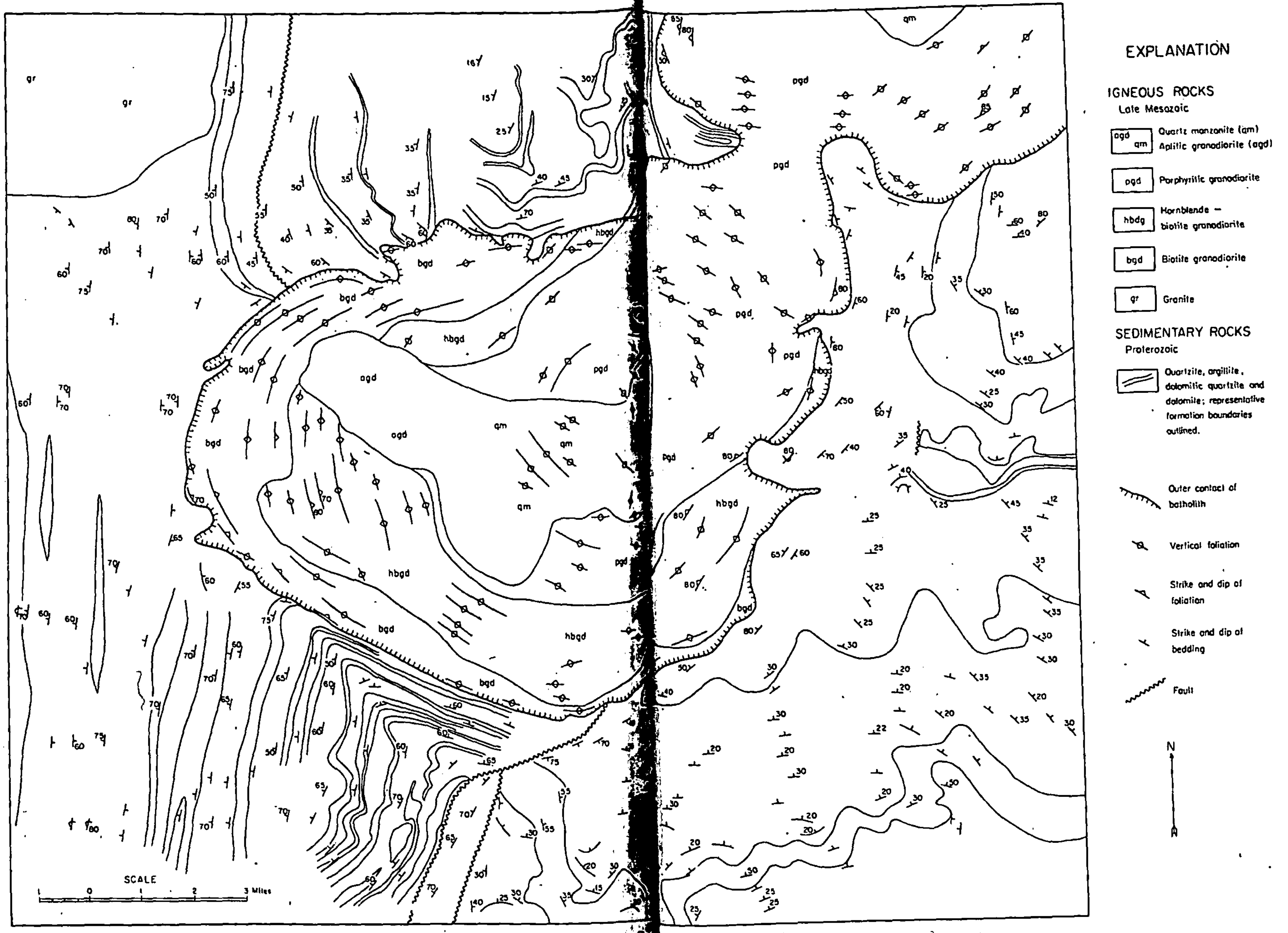

FIGURE 9.—REPRESENTATIVE C IC BATHOLITH OF MESOZONE
In part discordant, in part concordant because of crowding aside of country Creek batholith, British Columbia. Modified after J. E. Reesor (1954)

inferred to have originated at the last stage in consolidation of the interior granite. The interior planar foliation is universally vertical to subvertical, and although generally conform- of the batholith it locally

locally at the border contact of the complex it follows every irregularity. There is no linear structure in the quartz monzonite. A mafic-rich facies of the granodiorite may be due to incorporation as indicated by the presence of basic

Sierra Nevada batholith, California.—The Sierra Nevada pluton is about 300 miles long and 50–60 miles wide. It has not been fully explored, and an adequate systematic unified description of the existing data is not available.

the Yosemite area, and Cloos (1936) has published a structural study of the batholith of Yosemite National Park and vicinity and of an area northwest of Lake Tahoe. The structural study has been effectively summarized by Balk from literature later than that available to Balk will be cited here.

Hamilton (1956, p. 21–23) has given a summary upon which the following abstract is based: The granites were intruded in hundreds of separate plutons, some a few acres in extent and some covering several hundred square miles. Contacts are sharp, and gradation from granite to country rock commonly takes place in less than an inch. The uniformity of the granitic rocks over wide areas, and the gradual nature of the variations indicate that large volumes of the material could be mixed and homogenized. Some of the intrusive material seems to have been wholly liquid (as in the alaskites), whereas some was only partly liquid.

The granites intruded and produced contact metamorphism of a country rock which had already undergone moderate regional metamorphism. Structures in the metamorphic rocks are cut across sharply by the granites, and straight dilation dikes from the granites transect contorted metamorphic rocks. It is possible, however, that some of the deformation of the country rocks was the result of forces of intrusion acting before establishment of the final contacts. Early plutons intruded the metamorphic rocks; later plutons intruded earlier plutons.

The intrusive granites moved generally upward, as shown by their flow structures. Hamilton states that stoping operated, but to a degree that is only conjectural. He infers that an explanation in terms of cauldron subsidence–stoping on a huge scale could account for the emplacement of some of the plutons, particularly the smaller ones which are intrusive entirely into other plutons. However, he finds that no evidence for such emplacement has been recognized in the Sierra except that the geometry of some contacts might favor it. His conclusion is that the plutons formed from mobile magmas which moved upward and expanded outward, mostly passively but in part forcibly. Large amounts of material were incorporated into the margins by assimilation of wall rock and stoped blocks, but most of the granitic material was introduced from lower levels.

Durrell (1940), Macdonald (1941), and Mayo (1941), who have studied contact relations in the Sierra, have concluded that, whatever the means of intrusion, these means were passive. They find that contacts are mostly complexly discordant in detail and, if the wall rocks were shouldered aside, it was apparently accom-

zones about the intrusions, not with shearing along the contacts.

Mayo (1941, p. 1081) suggested that the dominant means of emplacement of Sierra plutons was "permissive intrusion, tectonically controlled," and "the structural features of the region seem in harmony with the idea that space for the intrusion was provided . . . mostly by buckling of the isoclinally folded wall rocks as a result of a north-south compression".

Durrell (1940) and Macdonald (1941) believe that the belts of country rocks within the batholith are roof pendants with relatively flat bases, remaining portions of a once-continuous roof. They agree that the contacts of the igneous bodies dip outward in most places 60°–70° and that there is a suggestion that the upper parts of the stocks, and probably all the larger intrusives, are more or less dome-shaped. Balk (1937, p. 67), on the basis of the work of Cloos, states that an eastern pluton of the batholith in the Yosemite region has what might be called a schlieren dome were it not for a small mass of structureless granite at the core, and that "each schlieren dome, or arch, is composed of a number of closely related, but petrographically different, rock types which tend to conform with the dome structure, as a system of concentric shells". This pluton also shows marginal upthrusts along the contacts of the oldest intrusion. A comparison of the geologic map by Calkins (1930) and the structure map of Cloos suggests that the flow structure in part crosses the boundary of units as in the White Creek batholith, British Columbia. The foregoing discussion deals largely with the central part of the batholith. Webb (1938, p. 315) finds that in the southern more deeply eroded part of the batholith the foliation of the septae have high-angle dips, none less than 60°.

Coast Range batholith, Alaska–British Columbia.—The upper Jurassic to Lower Cretaceous Coast Range batholith of Alaska and British Columbia has a length of at least 1250 miles with widths of 35 to 60 miles common and up to a maximum of 125 miles. A substantial part of the batholith consists of screens of schist and gneiss. There are few detailed maps of any parts of the batholith.

The batholith in a general way parallels the trend of the prebatholithic structures, but only in a general way. Phemister (1945, p. 79) notes that near Vancouver, British Columbia, "The country rocks strike slightly north of east, while the batholith has its greatest extension from south to north". Buddington (1929, p. 293)

olith as extending for many miles parallel to the strike of the adjacent formations, although locally crosscutting; he notes that for a length of 40 miles the batholithic margin strikes nearly north and transects the general structure at an angle of 15° to 40° whereas the northern part trends northwest at a slightly greater angle than the formations of the country rock. The batholith is stated to work across a synclinorium of Mesozoic rocks into Carboniferous rocks.

Smith and Stevenson (1955, p. 816–817) write concerning the emplacement of the intrusives in southern British Columbia that in many places the batholiths transect the structures of the country rock, that in some places much of the rock was pushed aside, and that trend lines around the southern end of the Coast Range batholith were probably formed by broad-scale and strong lateral pressures transmitted through the magma. The formations southwest of the batholith in southeastern Alaska are generally isoclinally overturned to the southwest, and their dip and the dip of the foliation of the border belt of the batholith is steep northeast or vertical.

Also in southeastern Alaska (Buddington, 1929, p. 181) reconnaissance indicates that the quartz diorite is predominant in the southwestern part (5 to 15 miles wide) of the batholith, and quartz monzonite is predominant in the eastern part (10–15 miles wide) with mixed rocks of generally granodioritic character in the core. Smith and Stevenson (1955, p. 811) describe the igneous intrusions in southern British Columbia as consisting of dioritic to granodioritic rocks in the western portion on Vancouver Island, predominantly granodiorite in the Coast Range, and principally granitic in central and eastern British Columbia. The foregoing relationships recall the Sierra Nevada batholith about which Durrell (1940, p. 12–13) writes that by far the most abundant plutonic rock in the area studied by him (southwestern part of batholith) is quartz diorite but that the most easterly plutonic rocks are quartz monzonite and granite, and that there is a gradual, although overlapping, progression from basic types in the west to acid types in the east.

The batholith is exposed along Tracy Arm in southeastern Alaska for a width of about 15 miles at an angle of 60° to the trend of the major structure. The intrusive rocks contain many large belts of injection gneiss and so many inclusions of country rock that it is doubtful if any area as much as 10 feet square is clear of them (Buddington, 1929, p. 69).

part of the Coast Range batholith in southeastern Alaska have characteristics definitely like those of bodies emplaced in the mesozone; the southwestern part of the main batholith and the adjoining country rock equally definitely have characters like those of the catazone. The northeastern part of the batholith appears to belong to the upper part of the mesozone or even the lower part of the epizone. A difference in degree of metamorphism of the country rock bordering the western and eastern parts of the batholith was early recognized by the Wright brothers (Wright, F. E. and C. W., 1908, p. 67) and discussed later by Schofield (*in* Schofield and Hanson, 1922, p. 65–66) and by Buddington (1928, p. 293–294). The batholith in southeastern Alaska is bordered on the southwest by a belt of medium- to high-grade metamorphic rocks. This belt has a width of about 35 miles at the southern border of Alaska; it narrows and pinches out at the north near Juneau. The belt of metamorphic rocks changes in metamorphic intensity toward the batholith from the general regional green-schist facies on the southwest through garnet-biotite and staurolite-kyanite zones to a sillimanitic facies with a migmatite zone adjacent to the batholith. There are many plutons of granodiorite or quartz diorite throughout. The rocks to the northwest along the southwest border of the batholith and along the northeast border of the batholith are slates and greenstones. It might be suggested that the belt of medium- to high-grade metamorphic rocks on the southwest border overlies an extension of the main batholith at depth and that the overlying schists were forced up from deeper zones by the upward push of underlying magma and by the accentuated upward drag of the magma mass which formed the southwestern part of the main batholith. The mechanism suggested resembles, but on a larger scale, the idea of upward drag or upward thrust of walls by rising magma proposed by Noble, Harder, and Slaughter (1949, Fig. 4). The quartz monzonite and granitic rocks of the eastern part of the batholith are younger than the quartz diorite of the western part. Granodiorite porphyry dikes of a character appropriate for the epizone are abundant locally in the country rock of the eastern border and are inferred to be related to the quartz monzonite.

Mesozoic Pseudo-igneous Plutons or Facies of Mesozone

Swedes Flat pluton, California.—Two ex-as emplaced by recrystallization and replacement in the mesozone are the Swedes Flat stock in California and the Chilliwack batholith in Washington. The roof (Pellisier granite) of the Inyo batholith, California, has also been interpreted as a product of recrystallization and replacement.

Compton (1955) has described the Swedes Flat pluton and interpreted it as the product of recrystallization and replacement. The main mass of the pluton is about 5 by 7.5 miles in diameter. Tonalite and granodiorite form the preponderant bulk of the pluton with about 4 square miles of gabbroic and dioritic rocks at the north end and another mass at the south end. Granophyre forms a local late intrusive phase. Gradational contacts between granodiorite and dark hornfels and amphibolite are striking. The latter give way to homogeneous granodiorite through a broad zone of intricately veined mixed rocks, 100–1000 feet wide against gabbro and diorite, up to 3 miles wide against metavolcanic rocks. The regular, gradational changes from the incipient webbing of these mixed rocks to inclusion-charged granitic rocks and then to granitic rocks with only vague hornblende clots suggest that the granitic rocks have passed through these stages in their development. Compton believes that a replacement origin for the homogeneous rocks is supported by their lack of flow structures or fracture pattern and by the fact that the foliate inclusions fit the projection of country-rock foliation through the pluton rather than any reasonable movement picture within it. He suggests that most of the granitization was produced by fluids moving in open channels and that the source of these fluids was probably a magma that formed the core of Swedes Flat pluton. Such a magma he infers would have lain south of, and perhaps at a lower level than, the rocks as now exposed.

Pellisier granite facies of Inyo batholith, California.—The Pellisier granite facies of the Inyo batholith, California, of middle or late Mesozoic age has been interpreted by Anderson (1937) as a roof facies of a major batholith developed *in situ* by replacement and recrystallization of both sedimentary and igneous rocks. The granite carries abundant inclusions, most of which appear to have originally been schistose or argillaceous. The replacement origin he bases upon field evidence of gradation between country rock and granite, the variability in composition of the granite, the inheritance of a structure closely resembling stratification, replacements, especially albitization. The part of the batholith mapped is 35 miles long. The Pellisier granite is in substantial part a hornblendic granite, whereas the main part of the batholith is a biotite granite. The solutions effecting the development of the Pellisier granite are inferred by Anderson to have been derived from the underlying magma which crystallized to form the main part of the batholith. Migmatitic gneisses have been formed in the border rocks of the batholith at depth.

Chilliwack batholith, Washington.—The description by Misch (1952) of the Chilliwack batholith, Washington, affords an example in which the hypothesis of granitization has been applied to explain the development of a batholith in the mesozone in the latter part of the Mesozoic. The batholith is about 35 miles long and 5–10 miles wide in Washington and extends north into British Columbia. The country rocks are a series of geosynclinal rocks, mostly phyllites, quartzites, marbles, greenstones, and greenschists except at the southern end of the batholith where, in the border zone of the batholith, the phyllites are changed to mica schist, and the greenstones and green schists to amphibolites. The granodiorite and quartz diorite of the batholith are inferred by Misch to be the result of a continued granitization process which first yields a series of granitic gneisses (the Skagit gneiss) and then large directionless granite bodies with gradual passages between. The country rocks on this hypothesis are not pushed aside along part of the contact, but in part the granodiorite is reported to have moved as a plastic crystalline mass and intruded the metamorphic rocks. Several smaller bodies appear as isolated intrusive stocks which have forced their way by pushing the metamorphic rocks aside. Part of these intrusives are in Lower Cretaceous rocks. The magma of these intrusions is assumed by Misch to have formed at depths below the level now exposed as the final climax in a long process of granitization and mobilization, and it is stated that it did not come from far away or an unknown depth. The depth of formation of the granitic gneisses and granites he estimates at three or four to ten miles.

Mesozoic Plutons with Protoclastic or Late-Stage Postconsolidation Deformation

Colville, Washington, and Cassiar, British Columbia, batholiths.—The Colville and Cassiar batholiths are discussed together as they both illustrate the results of continued magma movement after much of the complex is solid. The following summary has been made from the descriptions by Waters and Krauskopf (1941) and by Poole (1956).

The Colville batholith intrudes folded and dynamically metamorphosed sedimentary and volcanic rocks of late Paleozoic and Triassic age. The age of the batholith is thought to be late Jurassic or early Cretaceous. It is about 80 miles long. Waters and Krauskopf describe the batholith as remarkably heterogeneous both structurally and petrographically. A central mass of structureless granodiorites grades outward into a belt of foliated igneous rock which commonly shows intricate swirling of the foliation. These swirled rocks grade into a peripheral belt of variable but well-foliated migmatitic gneisses characterized by severe granulation of the constituent minerals. Over broad zones they find that this rock is a mylonite and that locally recrystallization has produced types resembling metamorphic granulites. The trend of the folds (in the country rock) as well as the direction of foliation and other structural features is predominantly northwest-southeast, and across these structural trends the batholithic border cuts with decided discordance. The wall rocks at the periphery of the Colville batholith show almost no evidence of contact metamorphism. Pegmatites are abundant near the border. From detailed consideration of relations both within the wall rock and within the adjacent intrusion, however, Waters and Krauskopf suggest that the features of the contact zone can best be attributed to the rise and emplacement of the batholith as a unit. By this interpretation they infer that the intense brecciation and usual lack of metamorphism in the wall rocks is due to a rise of the intrusive at a time when its peripheral portion (which now forms the gneissic and mylonitic facies) was nearly solid. Where the linear structure shown by swirl axes is well developed, the prevailing pitch of the axes is invariably northwest. They take this to indicate at least a slight regional control of this structure. The core of the intrusive mass is homophanous and composed of granodiorite and quartz diorite.

The following description and interpretation of certain phenomena shown by the Cassiar batholith is taken from Poole (1956). The batholith is of late Jurassic or Early Cretaceous age and is 13 miles wide by 70 miles long. The country rocks are sedimentary and volcanic rocks regionally folded and metamorphosed in the greenschist facies or of lower metamorphic grade. The Cassiar batholith is largely a biotite granodiorite and quartz monzonite. The western part of the batholith for a width of about 4 miles is a strongly foliated cataclastic gneiss, whereas the interior and the eastern part is effectively massive though foliated in places. The central and eastern part is relatively undeformed. There is evidence for shouldering aside and uplift during the emplacement of the magma although it is unlikely that all the space occupied by the batholith was made in this manner. Poole infers that the cause of the deformation of the western border zone was renewal of intrusion in the solid state such that the northeast part of the batholith and adjoining sedimentary rocks moved up and perhaps southwest relative to the southwest border zone, producing the northeast dip of the foliation. He notes that the sedimentary rocks for several miles southwest of the batholith are in isoclinal folds with a southwest dip, and it is unlikely that this would have persisted if regional deformation had acted to produce the northeast-dipping foliation of the west part of the batholith.

Precambrian Batholiths of Mesozone

General statement.—It has been noted that stocks emplaced in the epizone may be of Precambrian age, and so also may Precambrian stocks and batholiths belong to the mesozone. The Giants range (Allison, 1925) and the Vermillion (Grout, 1925) batholiths in Minnesota may be examples, and descriptions of others follow.

Noranda-Senneterre belt, Quebec.—Several batholiths of granite and granodiorite occur as intrusives in a series of intermediate and basic flows with subordinate tuffs of the Keewatin series and Timiskaming sediments in the Noranda-Senneterre belt, Quebec. The country rocks have complex structure and a low-grade regional metamorphism. Norman (1945) and Tremblay (1950) have described some of these batholiths, and a map and description of structure on which Figure 10 is based have been published by Dawson (1954).

The batholithic rocks include muscovite and muscovite-biotite granites and hornblendic varieties; the hornblendic varieties are quartz poor.

The boundary of the La Motte batholith is almost wholly conformable with the foliation of the country rock but does transgress a series of metasedimentary beds at one locality. The foliation indicates an elongate domal roof

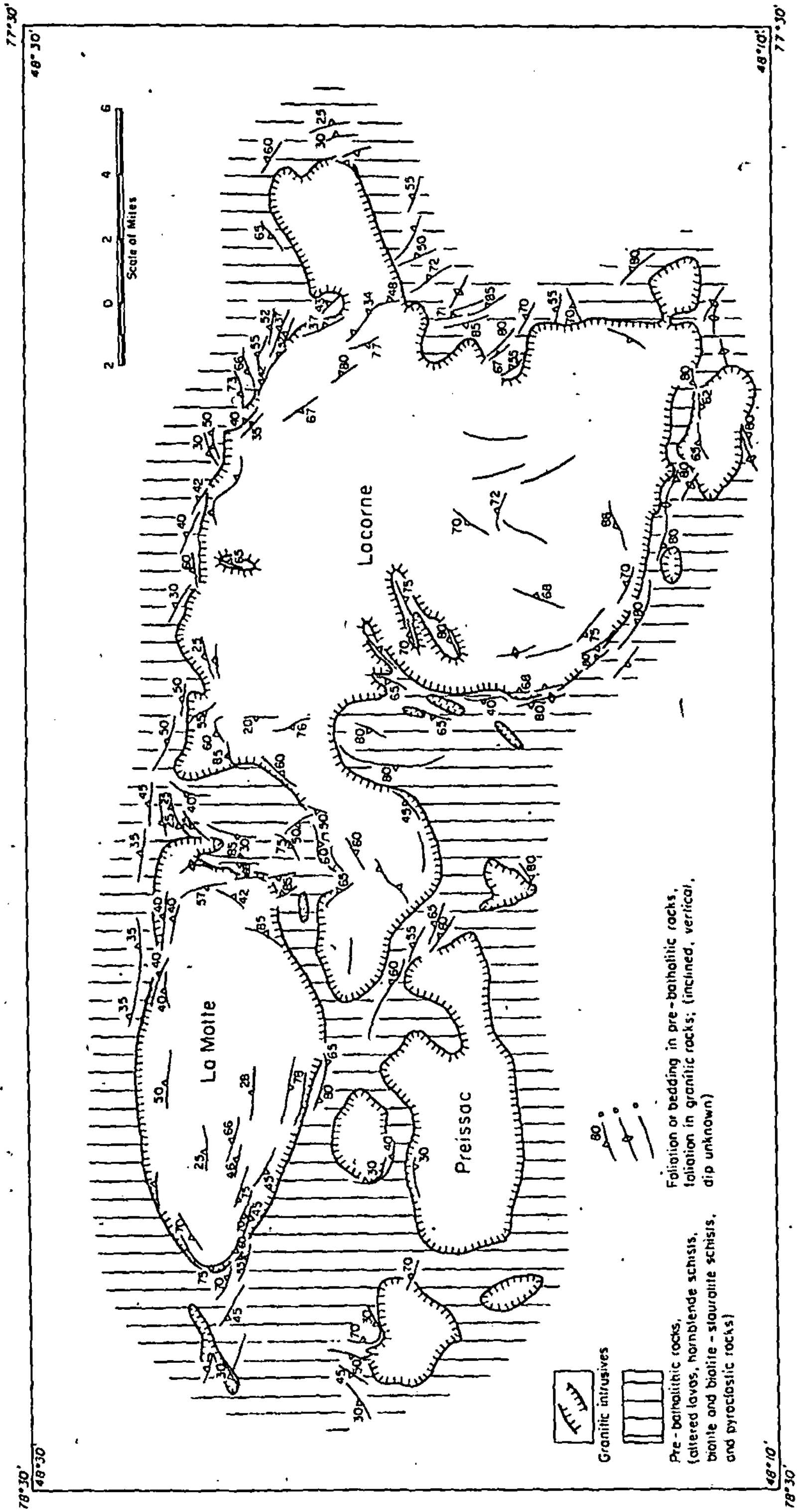

FIGURE 10.—EARLY PRECAMBRIAN BATHOLITHS OF MESOZONE

Lacorne batholiths, in part discordant, in part concordant; Noranda-Senneterre belt, Quebec; Canadian shield. Modified after K. R. Dawson (1954)

with moderately dipping walls on the north and steeply dipping borders around the rest of the pluton. Migmatization occurs very locally. Pegmatite forms nearly 50 per cent of the rock in a zone half a mile to a mile wide around the border of the pluton and at least 10 per cent of the interior of the pluton.

The Lacorne batholith has borders which are essentially concordant on the north and south but which on the east cut directly across the strike of the enclosing formations. Migmatization occurs locally near the southwest border of the pluton. Certain large inclusions have been moved from the vicinity of the walls and rotated. Thin lenses and dikes of granitic material have been injected in peripheral metasedimentary rocks. The foliation of the pluton indicates a structural dome in the northeast part of the main mass and another in the southwest part. As in the La Motte pluton the north border dips moderately, and the south border steeply. Dawson infers that the Lacorne pluton originally had steeply dipping walls and may have been slightly overturned toward the southwest.

The satellitic intrusions are interpreted as unroofed cupolas that join the main granite mass at depth.

Dawson (1958, p. 232) on the basis of a multivariate variance analysis of the mineralogical compositions concluded that the batholiths are composed essentially of a quartz monzonite that is homogeneous within the individual massifs and also within the intrusives as a whole.

The intrusions are all interpreted by the authors as of magmatic origin. The age determination (Shillibeer and Cuming, 1956) of mica from the Lacorne pluton by the K^{40}-A^{40} method gave 2500 ± 150 m.y.

Plutons north of Great Slave Lake, Northwest Territories.—Batholithic emplacement in Precambrian rocks north of Yellowknife on Great Slave Lake, Northwest Territories, Canada, has been described by Henderson (1943) and Jolliffe (1944), and the following is an abstract of their work. The country rocks away from the intrusions (Fig. 11) consist of graywacke, slate, and volcanic rocks. Pillowed lavas are so little metamorphosed that tops of flows can be determined. Chlorite and sericite are common in the graywacke and slate. The rocks are isoclinally folded with steep dips. Henderson believes that the isoclinal folds in turn have been warped into synclinal- and anticlinal-like structures with the axes of the secondary folds plunging nearly vertically. The granitic batholiths of granodiorite and quartz diorite were emplaced in the refolded isoclinal rocks with accompanying metamorphism of the formations in the vicinity into andalusite cordierite and quartz-mica schists (locally with staurolite) and hornfels. The outer boundary of metamorphism crosses the trends of the folds. The granitic batholiths are concordant, and Henderson infers that lateral pressures of invading magma may have produced deformation. The sediments and volcanic rocks dip away from the batholiths at steep angles. There is a series of small younger muscovite-biotite granite stocks that have also produced similar metamorphism of the country rocks. Many granite pegmatite veins are discordant with the foliation of the country rocks.

Supplementary Descriptions of Plutons Emplaced in Mesozone

Introduction.—Other excellent detailed descriptions of mesozonal plutons in the western Cordillera are those by Taubeneck (1957) of the Bald Mountain batholith in Oregon, by Smith (1947) of the Surf Point stock in British Columbia, by Krauskopf (1943) of the Wallowa batholith in Washington; and by Compton (1955) of the Bald Rock batholith in California. Plutons emplaced in the mesozone are also dominant in most other orogens, but descriptions of only the Snowbank stock in Minnesota and the Enchanted Rock batholith in Texas will be referred to here.

Bald Rock batholith, California.—The Bald Rock batholith, California, described in detail by Compton (1955), shows many characteristic phenomena of the mesozonal batholiths. The Bald Rock batholith is one of four plutons in a chain somewhat more than 40 miles long of small satellitic intrusions that lies 20 miles west of the main Sierra Nevada batholith. The batholith is about 9.5 miles wide, a little more than 11 miles long, and has an area of about 80 square miles. The bedding of the country rocks swings concordantly around the batholith, and Compton interprets it as formed by a forceful intrusion. However, he also notes that local crosscutting relations show that about a fourth of its area at the exposed level was gained by other means, and that large concentric outliers and a hull of injection migmatite suggest stoping. Inclusions are scarce, but gradational zoning of the batholith from a trondhjemite core through granodiorite to a heterogeneous tonalite rim suggests that

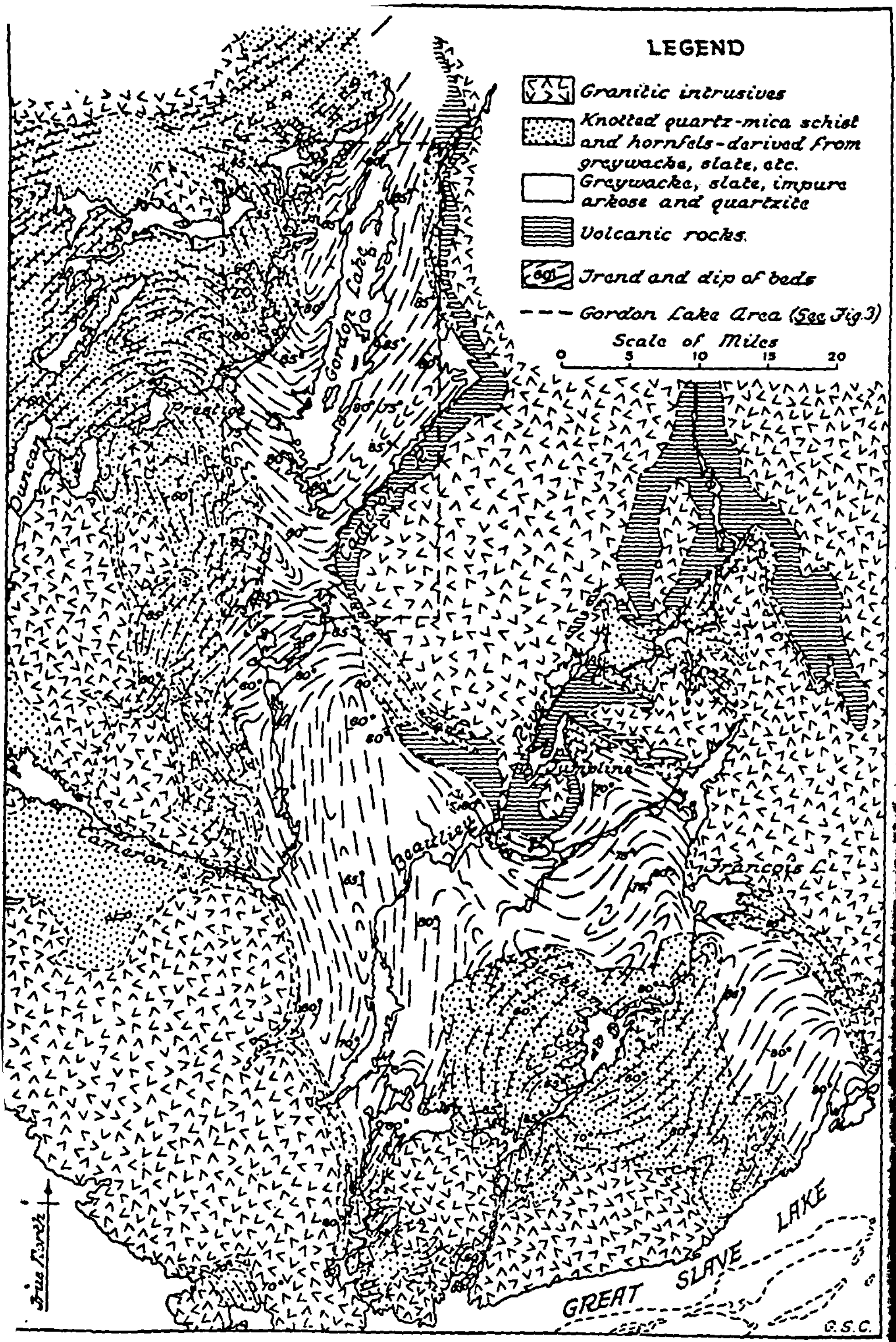

Figure 11.—Precambrian Batholiths Emplaced in Mesozone

Note discordant boundary of zone of metamorphism; from Henderson (1943) reproduced by courtesy he American Journal of Science

basic stoped rock contaminated an originally trondhjemitic magma.

Contact-metamorphic rocks of epidote amphibolite to possible pyroxene hornfels form an aureole that has an area nearly as great as the original kinetic intrusion. Squeezing and vertical stretching of wall rocks is indicated by pebbles of conglomerates and crystal lineation, most pronounced near the contact.

The zone of the contact migmatites is only a few feet to a few tens of feet wide where country-rock foliation is parallel to the contact; but it is as much as a quarter of a mile wide where the contact is sharply discordant to foliation. Most of the granitic parts of the contact migmatites occur as sharply bordered dikes that generally parallel foliation but locally cross it. The flow structure of the intrusives is more obvious near the contact because mafic minerals and inclusions are here most abundant, but its actual perfection, grain to grain, does not vary greatly from the rim to the core of the intrusion. In some places the flow surfaces parallel the gradational boundaries between rock types, in other cases they cut across them at large angles. There are several extensive, unconformable junctions of the flow lines that are probably local intrusive contacts. Notably, the flow structures dip as steeply at the core as anywhere in the intrusion. In the west half of the pluton there are thousands of thin aplite and pegmatite dikes, in the east half several large dikes and pipes of aplite and microgranitic rocks. Almost all the dikes are vertical and trend at about right angles to the flow structure. Thus they fan out on the flanks of the trondhjemite mass. The late-emplaced bodies in both halves of the batholith are interpreted as controlled by radial fractures resulting from upward pressure of an underlying mobile core. The way in which the flow layers locally cut across the gradational rock boundaries poses a considerable problem. Compton's suggestion is that immobile tonalite was forming near the contact while granodiorite was forming somewhat farther from the contact, and in some cases while trondhjemite was forming still further from the contact. He believes the flow structures and their overall pattern can be explained only by assuming that the magma was mobile during the growth of the batholith and that grain orientation took place when a zone of mobility slowly grew into the intrusion from its walls. Larsen and Poldervaart (1957) state that

> "The distribution of two distinct zircon populations in the Bald Rock batholith as well as structural relations demonstrated by Compton, are explained in terms of a parautochthonous intrusive of migma-magma, with solid phases predominant at the borders of the pluton and silicate melt predominant in the core".

Larsen and Poldervaart emphasize that xenoliths are concentrated between the trondhjemite core and granodiorite-tonalite rim but are rare in the rim itself.

The Merrimac pluton to the north, described by Hietanen (1951), shows many similar phenomena.

Snowbank stock, Minnesota.—The Precambrian Snowbank stock, Minnesota, described by Balk and Grout (1934) is a very fine example of a typical stock emplaced in the mesozone. It is an elliptical mass, 3 miles wide by 5 miles long, with planar foliation in the borders and a faint linear structure throughout, although a late granite portion is almost massive. The country rocks are strongly crowded outward with structures largely conformable with the contact, and the magma rose steeply as a cylindrical mass at an angle of about 70°.

Enchanted Rock batholith, Texas.—A pluton emplaced in quartzo-feldspathic gneisses and high-grade metamorphic schists has been described by Hutchinson (1956) as the Enchanted Rock batholith from the Precambrian of Texas. Although of Precambrian age and intruded in high-grade metamorphic rocks, the batholith has some characters of those emplaced in the mesozone. It is also exceptionally interesting in that one-third of the batholith has a phacolithic relationship to the country rock. The batholith is 9 miles wide and 15 miles long. The batholith shows predominantly a peripheral concordance of the country rock with the border of the pluton, with minor discordance. Except for the phacolithic part of the batholith the foliation is nearly vertical throughout. The lineation is also nearly vertical, and during the early stages of intrusion and crystallization the principal direction of transport is inferred to have been vertical. Marginal fissures are restricted to the outer mile-wide perimeter, dip 10°–25° inward, and are filled with pegmatite and aplite. There is one joint system with steep dip, subradial and at right angles to the planar structure, also filled with pegmatite and aplite. Chilled border rocks are 10 to 2 feet wide. Hiatal porphyritic texture prevails in chilled borders and apophyses. The phacolithic part occupies a synclinal trough plunging 35°–40°. There are four concentric zones of different granitic facies within the pluton.

Hutchinson infers that the batholith was emplaced at a late stage in the deformation by forceful injection and that not more than 5 per cent of the batholith is of replacement origin. The age given by the "Larsen" method is 815 m.y.

PLUTONS OF TRANSITIONAL MESOZONE–CATAZONE

General Discussion

In a number of regions part of the plutons have characteristics of the mesozone, and part those of the catazone, although both are of similar age. Individual plutons also have some characteristics of both the mesozone and the catazone. Such mixed associations or characteristics appear to occur particularly in plutons emplaced in rocks with an intermediate grade of metamorphism; that of the epidote-amphibolite or staurolite-kyanite subfacies. Where plutons are of similar age in the same region, yet vary from mesozonal to catazonal, it seems probable that we are dealing with local variations in the physical conditions at the site of emplacement rather than with different depth zones. In many such examples the pluton with characteristic of the catazone may be the roof facies of a mesozonal batholith.

Plutons of Wolverine Complex, British Columbia

Armstrong (1949) describes the Wolverine complex as occupying more than 1000 square miles in British Columbia. According to him it includes a series of micaceous quartz-feldspar gneisses (in part with 40 to 65 per cent quartz) and migmatites with granodiorite plutons (up to 10 square miles) formed in place by progressive injection of granitic material and gradual replacement of injected rock. Roots (1954) considers the complex was formed by a metamorphism and granitization superimposed on previously regionally metamorphosed Proterozoic and Lower Cambrian sedimentary beds whose grade of regional metamorphism increases in intensity with successively lower stratigraphic horizons, low-grade quartz-chlorite schists; crystalline limestone, slate, phyllite, chloritoid schist, and graywackes in the upper part of series; quartz-mica schists, quartzite, garnetiferous schists, and kyanite and staurolite schists in the lower part of series. The regional metamorphism he believes preceded folding, and the temperature rise was due in part to emplacement of underlying igneous or anatectic material and to sim produced by relatively gentle orogenic deform tion. Granitizing fluids developed leucogran some of which consolidated in place; some w mobilized and traveled along foliation pla and fractures in partly granitized metase ments to form sills and dikes. A stock of gra diorite (5 square miles) with sharp contac steep walls, and a flat domed roof intrudes th schists.

Plutons of Shuswap Complex, British Colum

According to Cairnes (1940) the area of th Shuswap complex may be more than 400 square miles and consists of intensely meta morphosed Precambrian beds of Belt (?) a about Shuswap Lake, but in other areas ma include Upper Paleozoic and probably Trias formations. The metamorphic complex co tains abundant pegmatites, gneisses wi aplitic injection material as an important co stituent, great bodies of granitoid gneiss, ma sive granite with many bodies of pegmati granite, and a "sill–sediment" complex crystalline schists and sill-like bodies of grani gneiss. He believes that the principal process have seemed to involve a gradual upwa seepage of this material (pegmatitic and apliti differentiates), infiltration along bedding plane replacement or partial replacement of inte vening rock matter, and the growth, *in sit* of perhaps much of the pegmatitic granite. H suggests that, in places, the continued suppl of magmatic material resulted in the comple conversion of large bodies of the original stra into massive granitoid rock, which, under the conditions of transformation, became partly plastic or molten and, where subjected to loc stresses, behaved much as a normal intrusive rock behaves in its contact relation with adjoining rock masses. Cairnes believes th granitization was effected in connection wit the emplacement of the Mesozoic batholiths. Armstrong and Roots consider the Shuswap complex the equivalent of the Wolverine complex and date the granitization as pre-Pennsylvanian or pre-Mississippian.

Williamsburg Granodiorite Pluton, Massachuse

The Williamsburg granodiorite pluton in th Williamsburg quadrangle, Massachusetts, described by Willard (1956), appears to be an example of emplacement under condition transitional between those of the mesozone and catazone. The following abstract is bas

Willard's report. The country rocks consist garnetiferous quartz-mica schists, quartzite, rble, phyllite, and amphibolite. The phyl- s carry *metacrysts* of staurolite, and the hists locally have metacrysts of staurolite d kyanite, most abundant near the granoorite. Sillimanite and tourmaline are deloped near contacts with the intrusive rocks. he amphibolite consists of hornblende, uninned calcic plagioclase, with quartz, biotite, tanite, and epidote. The rocks may barely ve attained the staurolite-kyanite facies at e time of magma emplacement. The granoorite is in large part a mixed type consisting biotite-muscovite granodiorite and of pegmae, granitic, aplitic, and granodioritic dikes d sills of many sizes. Some of it might be lled an injection gneiss. The granodiorite is truded by granite and pegmatite dikes and lls. At many places the granodiorite and associated rocks contain randomly oriented xenoliths of schist and quartzite.

The metasedimentary beds are part of a homocline on the east limb of an *anticlinorium*.

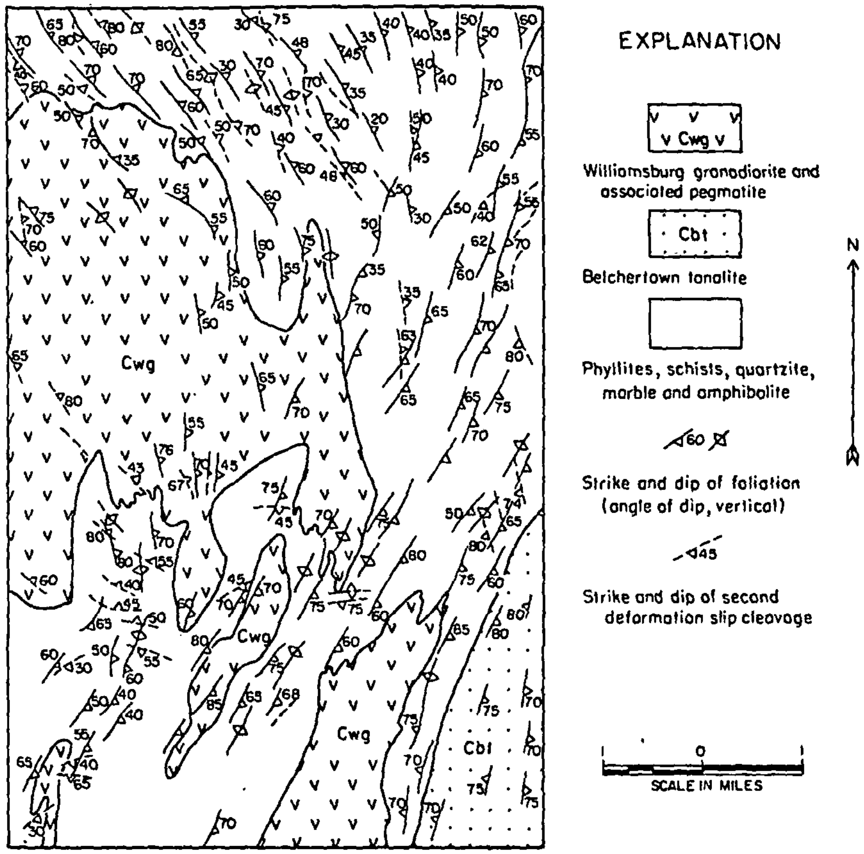

FIGURE 12.—WILLIAMSBURG GRANODIORITE PLUTON OF TRANSITIONAL MESOZONE-CATAZONE
Appalachian orogen, Massachusetts. Modified after Willard (1956)

The schistosity is approximately parallel to bedding on the limbs of folds but cuts across the bedding on the crests and in the troughs. The axial-plane schistosity is inferred by Willard to have been deflected at the north by the force of the intrusion (Fig. 12). Slip cleavage is a planar structure that coincides with the axial planes of small corrugations or microfolds in the schistosity. For the most part the slip cleavage is parallel to the exposed contact between the granodiorite and the country rock. This suggests that the schistosity planes nearest the intrusive moved up relative to those farther away. Willard believes that, in large part, the intrusive forced its way along earlier foliation planes, bending them apart, causing the de-and producing local shear couples that resulted in the slip cleavage that surrounds and dips away from it. The schistosity planes nearest the intrusive moved up relative to those farther away. A northwest-trending slip cleavage was produced by a later deformation. The present writer would also emphasize the discordant relationships between the borders of the intrusive and the country rock, as portrayed both in plan and by the section as drawn by Willard, that relate this pluton to the mesozone.

Lithonia Gneiss (and Stone Mountain Granite of Mesozone), Georgia

The Lithonia gneiss and Stone Mountain granite in Georgia afford another example of a gneiss with characteristics of the catazone inferred to be associated in time and space with a pluton apparently emplaced in the mesozone. The Rb/Sr age of the granite averages 278 m.y. and biotite from the Lithonia gneiss gives 297 m.y. (Pinson *et al.*, 1957, p. 1781).

The Stone Mountain pluton has been described by Herrmann (1954) from whose work the following summary is taken. The country rock is regionally metamorphosed and consists of schists of the staurolite-kyanite or epidote amphibolite subfacies. The muscovite granite pluton in part cuts discordantly across the structure of the country rock and in part crowds the country rock to one side. The flow structure of the granite is in part conformable to discordant contacts with the intruded gneiss. Pegmatite dikes are locally abundant in the schist on the north side of the intrusive, and aplite dikes on the south side.

An independent extensive belt of the schist was injected and replaced by syntectonic, magmatic, potassium-rich solutions which modified it to a gneiss (Lithonia gneiss) of granitic composition. Pegmatite dikes in the Lithonia gneiss are small and irregular; commonly discordant but locally partially concordant aplite forms thin veins. The relationships of the rocks are shown in Figure 13. The migmatitizing and granitizing fluids are thought by Herrmann to be co-ordinate with the magma that formed the Stone Mountain granite. The rocks surrounding the Lithonia gneiss are metamorphosed to the sillimanite-almandine amphibolite subfacies. It is thus a problem to the present writer as to whether the Lithonia gneiss is the roof portion of a mesozonal pluton or represents an earlier emplacement in the catazone.

(and Nonewaug Granite Lens of Mesozone), Connecticut

A belt of rocks in Connecticut, regionally metamorphosed in the epidote amphibolite or staurolite-kyanite facies, contains plutons whose characteristics are those of the mesozone. There are also bodies of gneiss formed by migmatization and granitization that seem best classified in the transitional mesozone-catazone.

The Paleozoic Nonewaug granite lens in this belt has been described by Gates (1954). The pluton is 9 miles long and 3 miles wide. The long axis is about N.60°E. and is across the foliation of the schists whose regional trend is north. The schists adjacent to the lens on the north and west borders have in general a foliation parallel to the border of the granite as a result of crowding aside at the time of magma intrusion. On the south border the foliation of the schists is normal to the contact. The southern part of the granite lens, however, is a complex mixture predominantly of granite, pegmatite, and granitic gneisses with subordinate feldspathized schist and schist. The foliation is variable, as in a crumpled zone. The granitic gneisses are granitized schist. The bulk of the granite mass has a layered structure dipping 35°–80° SE. and is inferred to be of magmatic origin. The granite has crosscutting apophyses and also occurs as dikes in the schists. Pegmatite veins are present in the schists. The normal schists are mica quartzites and quartz-mica schists with biotite and muscovite and accessory garnet, staurolite, and kyanite. The characteristics of the pluton seem appropriate to emplacement in the mesozone.

Stewart (1935) has described from a zone southeast of the Nonewaug pluton an extensive belt of porphyritic granitic gneiss formed in schists of similar age and grade of metamorphism by magmatic injection and by permeation and replacement by fluids whose source was in a subjacent magma. Agar (1934, p. 363–369) has also emphasized the extensive occurrence of mixed gneiss resulting from invasion of these schists by granite and pegmatite. These gneisses may be the roof portion of mesozonal batholiths, or they could belong to the upper part of the catazone.

Precambrian Phacoliths of Hanson Lake Area, Saskatchewan

It appears possible that phacolithic emplacement, although characteristic of the catazone,

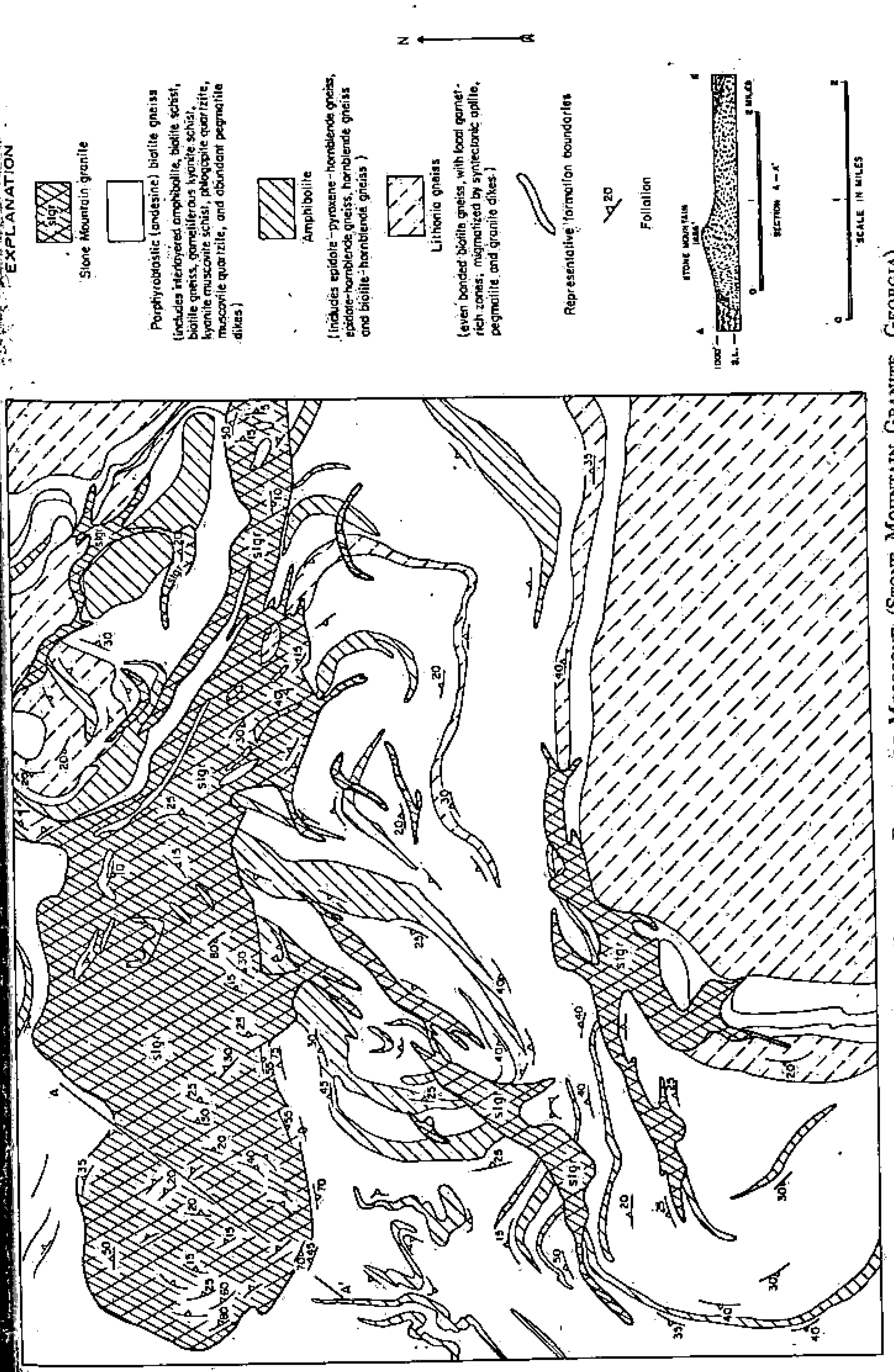

FIGURE 13.—SMALL PALEOZOIC PLUTON OF MESOZONE (STONE MOUNTAIN GRANITE, GEORGIA)

In part discordant, in part concordant, associated with penecontemporaneous migmatitic gneiss (Lithonia) of transitional mesozone-catazone. Modified
Herrmann (1954)

may also occur in the transitional mesozone-catazone. Byers (1957) has described several syntectonic phacoliths of granodiorite or quartz diorite emplaced in anticlinal structures of biotite gneiss, amphibolite, and migmatitic gneiss of the Hanson Lake area, Saskatchewan. The country rocks are described by him as regionally metamorphosed in the amphibolite facies or the garnet-staurolite zone or staurolite-kyanite subfacies; locally the sillimanite-almandite facies is attained. The anticlinal structures have steeply dipping axial planes and may be asymmetrical or isoclinal.

Syntectonic Pinckneyville Batholith, Alabama

The Pinckneyville quartz diorite batholith, Alabama, has been described by Gault (1945) as a syntectonic batholithic intrusion. It is more than 40 miles long, 8–12 miles wide, and is emplaced in phyllites, schists, quartzites, and amphibolites that have attained only an intermediate metamorphic grade.

COMPLEX HISTORY OF GREAT BATHOLITHS, LARGELY OF MESOZONE

Introduction

Great batholiths such as those of the Coast Range of Alaska and British Columbia, the Sierra Nevada of California, Southern California, and Idaho have had a most complex history. Individual units have been emplaced, usually in a systematic sequence from more mafic to more alkali-siliceous, to make up composite stocks or small composite batholiths. Such composite plutons have in turn been emplaced as a contemporaneous or successive series within a limited period of time to yield a multiple aggregate that forms part or the bulk of the batholith. Such series may in turn be repeated in periods of time separated by substantial intervals.

The ages of the members of the Sierra Nevada batholith (largely mesozone) have been determined by the Larsen method (Faul, 1954, p. 265) to range in large part between 90 and 111 m.y. and by the potassium-argon method on biotite to be between 82.4 and 95.3 m.y. in general.

The ages of several major members of the Sierra Nevada batholith in the Yosemite National Park area have been determined (Evernden, Curtis, and Lipson, 1957) by the potassium-argon method. The ages for the youngest major member is 82.4 and the oldest 95.3 million years, a range of about 13 mil years, and they believe the range is com to a few per cent at most. The average inter of time between successive intrusions is e mated as 2 million years, and each is infer to be almost completely crystalline at t time of the succeeding intrusion. The auth propose

> "that room for the batholith was made slowly in small increments by vertical uplift of the ove ing sedimentary rocks which were stripped by e sion as rapidly as they rose. Probably some of t earliest granitic intrusions were at the surface. the time the last intrusion squeezed into place",

The writer would qualify this by suggest that this mechanism was only one of sever factors in emplacement.

The ages of several outlying batholiths wi the sedimentary rocks have been determi by Curtis, Evernden, and Lipson (1958) range between 133 and 143 million years. T intrusives of the two different age groups believed by the authors to correlate with t separate major orogenic periods—one of L Jurassic and the other of early Late Cretace age.

The Coast Range and Idaho batholiths the complex of plutons of the northeast section of the Appalachian orogen all ha mesozonal plutons as the dominant elem but also many plutons emplaced in the epizo as younger members.

The ages of a granodiorite and a diorite fr the Coast Range batholith of southeast Alaska have been determined (Matzko, J and Waring, 1958, p. 538) to be 93 and 10 m.y. respectively. They note (p. 537) th Silver, Stehli, and Allen have determined mean age of 103 ± 6 m.y. for four early La Cretaceous plutonic rocks from Baja Ca fornia. The Baja California, Sierra Nevad and Coast Range intrusives would thus se in part to be of similar age. The Coast Rang intrusives, however, are called Late Juras to Early Cretaceous in this report.

Coast Range Batholith

The Coast Range intrusives in north British Columbia are stated by Kerr (193 p. 305) to comprise nine (more or less) dist intrusive phases which range in age from ea Triassic to late Early Cretaceous. Kerr d scribes the youngest member, which cuts Lo Cretaceous rocks, as a quartz monzonite p in mafic minerals, homogeneous, and miaroli with discordant relations to the country ro coarse grain persists to the sharp contacts ept for a narrow chilled edge, in places less an 1 inch. The description suggests to the esent writer emplacement in the epizone, d this is in agreement with his own observans of this rock in the Hyder district.

Kerr refers certain dome-shaped bodies of igoclase granodiorite and hornblende granoorite to a Jurassic age. The characteristics e consistent with emplacement in the mesone. He describes as still older a thick sheet granodiorite which is gneissic throughout d possibly of early Jurassic or Triassic age. nally there is an older hornblende granoorite for which he suggests a Triassic age.

The probable emplacement of part of the uartz diorite of the southwest border of the atholith in the catazone has been previously ferred to.

Mathews (1958, p. 172–177) has described rt of the southern end of the Coast Range atholith that comprises both plutons of rassic or Early Cretaceous age and two utons of post-Late Cretaceous age. The tter are homophanous. Only one of the ounger batholiths shows a faint flow structure, d that near the borders only.

There are also small epizonal plutons of ght quantitative volume in the plutonic mplex of the Coast Range. One, described Gault (1945), has developed contemporanes explosion breccias. A stock of miarolitic anite porphyry is intrusive into Tertiary yolitic volcanic rocks of similar composition Zarembo Island (Buddington, 1929, p. 275).

Idaho Batholith

The complex history of the Idaho batholith s been described by A. L. Anderson (1952). e states that the batholith is composed of screte masses of granitic rock, some of which me to place under deep-seated conditions, hers at much shallower depths. The deeply ated emplacements include two closely reted, but separately formed masses; the earlier olved while deformative stresses associated th a major orogeny were still quite intense, e other evolved during the later less intense ages. Anderson infers that these masses probly had their roots in the same source, but at the granitic bodies introduced under allower conditions came from a younger probly unrelated source. The oldest rocks of the tholith were emplaced at the close of Sierra evadan orogeny hence near the end of Jurassic me. The younger rocks appear to be associated with Laramide structures and are believed to be product of "Laramide orogeny of late Cretaceous time". He describes a "marginal facies", a gneissic quartz diorite along the western side with scattered roof masses in more central areas. The gneissic structure he takes to indicate emplacement during orogeny. An inner facies is largely quartz monzonite without gneissic structure emplaced after orogenic forces had ceased and formed under rather deep-seated conditions as was the quartz diorite. The bulk of the batholith is believed to be of these Sierra Nevadan rocks. The present writer suggests the possibility that the quartz diorite of the western part of the batholith may have been emplaced in the catazone and the quartz monzonite in the mesozone.

Anderson notes that a younger group of rocks, including diorite (of gabbrodiorite type), granodiorite, and quartz monzonite were intruded later and resemble the rocks of the Boulder batholith and its satellites of Late Cretaceous age. The diorite has chilled contacts and hypabyssal characteristics. The granodiorite and quartz monzonite have features he infers to be indicative of fairly rapid cooling and intrusion into the cold older batholithic rocks fairly close to the surface.

Larsen and Schmidt (1958) state that some coarse muscovite-bearing quartz monzonite and some very fine-grained granite of the Idaho batholith have small miarolitic cavities. They also contrast the batholith of Southern California in which the largest unit is about 200 square miles with the Idaho batholith in which several units are more than 2000 square miles each. Their statements for the Idaho batholith, however, are based on reconnaissance only, and detailed work may reveal greater complexity. They interpreted a subordinate porphyroblastic granite facies as the product of granitization of schist. Ages determined by the Larsen method on most of the rocks average 108 m.y., but one pluton gave 57 m.y.

Complex of Plutons in Appalachian Orogen, Northeast Section

A varied series of plutons occur in Paleozoic metasedimentary and metavolcanic rocks in a wide belt of the Appalachian orogen that extends north from Long Island Sound and northeast through Newfoundland. The plutons range in age from late Ordovician (?) or Taconic through Middle to Late Devonian or Acadian to post-Pennsylvanian.

Plutons of Taconic age (about 350

with features characteristic of the catazone occur in Connecticut and New York. These may include the Thomaston granite and granite gneiss (Agar, 1934, p. 363-368) and the Chelmsford granite sheet in Massachusetts (Currier, 1947). Billings (1956, p. 121) tentatively assigns the plutons of the Highlandcroft series a Late Ordovician (?) age. Detailed structural relations of these bodies are not known. They may be of transitional mesozone-catazone type.

Acadian intrusions of Middle or Late Devonian age are abundant throughout the belt. Many have the characters of plutons emplaced in the mesozone. Well-described examples include the La Poile batholith of porphyritic biotite granite in Newfoundland (Cooper, 1954, p. 26-29), the Mt. Waldo batholith of porphyritic biotite granite in Maine (Trefethen, 1944), French Pond biotite granite, New Hampshire (Billings, 1937, p. 508-509, 538; 1945, p. 57-58), binary granite plutons of the Memphremagog area, Vermont (Doll, 1951), the Scituate granite gneiss, Rhode Island (Quinn. 1951) of about 306 ± 18 m.y. (Quinn *et al.*, 1957), and the Nonewaug granite lens, Connecticut (Gates, 1954). The Winnipesaukee batholith, New Hampshire (Billings, 1956, p. 128) of age 296 ± 29 m.y. (Lyons *et al.*, 1957), Williamsburg pluton, Massachusetts (Willard, 1956), and Prospect gneiss, Connecticut (Stewart, 1935) have a complex of characteristics and relationships some of which suggest emplacement in the catazone and some the mesozone. They have been tentatively grouped as transitional mesozone-catazone. Billings (1948, p. 122) states that the New Hampshire series of plutons are syntectonic and that the Bethlehem gneiss is injected in giganitic sill-like bodies.

Many epizonal stocks and batholiths occur throughout the length of the same belt and range in age from Late Devonian (?) to post-Pennsylvanian. Examples are the Late Devonian (?) Ackeley and St. Lawrence miarolitic leucogranite batholiths of southern Newfoundland and a granite porphyry batholith of northern Newfoundland. Also, in western Newfoundland Phair (1949, p. 135) reports intrusives of epizonal characteristics that cut Lower Devonian rocks and have furnished pebbles to Mississippian conglomerates. The Mississippian (?) Quincy granite and porphy- granite stocks of Massachusetts and Cowesett granite and

270 m.y. (Quinn *et al.*, 1957). Somew younger in age—post-Pennsylvanian and abo 234 ± 23 m.y. (Quinn *et al.*, 1957)—is Narragansett Pier granite batholith and Westerly granites of Rhode Island (Nich 1956). The Youngest epizonal plutons are parently those of the White Mountain ser in New Hampshire (Billings, 1956, 129-13 that include the White Mountain batholi Ossipee stock, and other plutons associat with ring dikes, volcanic rocks, and calder structure. These plutons give ages of 186 ± 1 m.y., which would suggest Late Permian age (Lyons *et al.*, 1957), though Billings grouped them as Mississippian (?).

Plutons of the Catazone

> *"If we relax, we may easily become anarchic— the deeper geology may pass into the high lunacy"* Read, 1951

Introduction

The country rocks in which the plutons of the catazone are intruded are inferred in general to have had a temperature as high as that indicated by the grade of regional metamorphism. This may be a minimum of 450°C. Characteristically this will be at least as high as the amphibolite facies, and the country rocks may consist of amphibolites, metaquartzite, sillimanitic quartz-mica schists, marbles, granulites, orthogneisses, and paragneisses. Associated extensive migmatite zones of semiconformable veined character (phlebites) are diagnostic. There are no chill zones in the plutons. Foliation may be and commonly is well developed throughout the bodies but need not be. It may be steeply dipping throughout parallel to the elongation or periphery of the plutons. Gneissic foliation is common. There is a general conformity between country rock and intrusive. The country rock may be pulled apart as a result of extensive plastic crystalline flow during deformation preceding emplacement of the pluton and also during the emplacement.

Augen gneisses and porphyroblastic granites and granitic gneisses of replacement origin are common.

A migmatitic-like facies may develop locally as a product of metamorphic differentiation.

It has been well established that many plutons or major parts of plutons in the cata- been emplaced by recrystallization equally well established that many plutons of the catazone were emplaced predominantly by intrusion of magma. The number of plutons in the catazone, however, for which the mechanics of emplacement is at present indeterminate, problematical, or the subject of controversy probably exceeds by far the total number in the first two categories.

Batholithic bodies, in part, emplaced in the catazone have been referred to in the preceding discussion of the Coast Range batholith of Late Jurassic-Early Cretaceous age, and some Paleozoic plutons of the Appalachian orogen may belong to the catazone, but the most extensive development of bodies emplaced in the catazone is found in rocks of Precambrian age.

Batholiths of the catazone as a whole and in detail may in part crosscut the structural trends of the more rigid members. Characteristic forms for the plutons are domes, phacoliths, and conformable sheets. They are generally interpreted as syntectonic. Many of the masses, however, have irregular forms or are so large and complex as to be indeterminate without further study. Funnel and nearly vertical subcylindrical forms have also been proposed for some plutons.

Domes

Introduction.—Domical plutons are a major form of granite emplacement in the catazone. Several interpretations have been offered for their origin: (1) magmatic emplacement and therefore predominantly igneous, (2) magmatic emplacement with concordant interlayers of country rock (stromatolithic or interlayered xenolithic domes), (3) replacement domes, and (4) tectonic domes by plastic crystalline flowage or (4a) by rejuvenation and remobilization accompanying the introduction of material by fluids. Many igneous and interlayered xenolithic domes may in part have had a phacolithic mechanism of emplacement.

Igneous domes of magma emplacement.—Several igneous domes of magma emplacement are described in some detail.

SALMON LAKE BATHOLITH: The Salmon Lake batholith (Fig. 15, just southeast of Lat. 44°00′ and Long. 74°45′) in the Adirondack Mountains area of New York can be interpreted as a transitional form between a large elongate phacolith and a more nearly equidimensional dome of magma emplacement. The rock of the structure is a gneissoid hornblende granite with mesoperthite the predominant feldspar. The dips at the core of the structure are gentle, but they are steep on the limbs toward the adjoining synclines. The anticlinal structure is brought out by skeletal remnants of members of the Grenville series with its sheets of metadiorite and metagabbro. It may be noted that the lineation of practically all the granite with gneissoid (as distinct from gneissic flow) structure including the anticlinal ovoidal batholiths is subparallel to the strike of the foliation consistent with the late syntectonic emplacement and contrasting with the normal steep lineation of mesozonal batholiths.

KILLINGWORTH, BRANFORD-STONY CREEK, and CLINTON DOMES, CONNECTICUT: A cluster of granite domes described by Mikami and Digman (1957) were emplaced at depths of at least 5 miles. Two of the domes are largely of a microcline-rich granite and have a peripheral zone of migmatite, a third one is tonalite with an interlayered xenolithic facies that forms an outer facies and constitutes a mantling xenolithic dome.

The Killingworth dome consists of tonalite of magmatic origin that grades outward from a eugranitic core into a periphery interlayered with slablike amphibolite inclusions from the country rock. Amphibolite inclusions appear only in minor amount in the central area where they have a random orientation and the foliation of the tonalite flows around them. The tonalite of the peripheral interlayered xenolithic portion is granoblastic and is a slightly earlier facies deformed by continued flow of the interior. The foliation of the dome has gentle dips at the core and moderate to steep dips on the flanks except at the south where the country rock is overturned outward.

The immediate bordering country rock consists of hornblende and biotite gneisses with two-thirds of the gneisses containing more than 35 per cent hornblende. The gneisses are overlain by biotite-muscovite schists.

The Branford-Stony Creek dome has two units, a quartz monzonite as a local border facies and a younger microcline-rich granite as the main mass. The composite body is about 2 miles from the Killingworth dome and is younger. The domical foliation is inferred by Mikami and Digman to have been imposed before complete consolidation on material emplaced as magma. The dips of foliation in the Stony Creek dome are moderately steep throughout. In a few places the foliation of the country rock strikes into the granite at small angles. A few unoriented angular inclusions occur in breccialike habit in the central part of the granite. The metamorphic formations around the mass in a zone half a mile to a mile wide contain migmatite. The migmatite is interpreted as formed predominantly by forcible granitic injection with some accompanying metasomatism. There is a local development of granite augen gneiss emplaced partly by dilation and partly by replacement. The Clinton dome is similar to the Branford-Stony Brook dome.

IGNEOUS DOMES OF SOUTHERN RHODESIA: The Precambrian domes of Southern Rhodesia have been described by Macgregor (1951). He quotes (p. xxxix) with approval the following statement of Maufe,

> "It is a general rule throughout the Territory that the strike of the schists and their foliation is parallel to the edge of the batholiths and to the banding of the gneissic granite. Secondly, the batholiths, being roughly oval bodies, have curving margins, there being no general direction of strike throughout the country independent of granite batholiths. Thirdly, the schists almost always dip away from the margins of the batholiths, thus appearing to be synclinal areas."

Macgregor suggests that the large ovoidal batholiths probably originated as homogeneous magmas which ultimately consolidated as granite gneiss.

Interlayered xenolithic domes.—The term "stromatolith" was proposed by Foye (1916, p. 791) for "a rock-mass consisting of many alternating layers of igneous and sedimentary rocks in sill relationship". The types to which it was applied were granite plutons of the Haliburton-Bancroft area, Ontario (Fig. 14).

HALIBURTON-BANCROFT AREA, ONTARIO: These plutons have a minimum of 20 per cent of layers of gray gneiss and amphibolite. It was inferred that the granite magma was intruded concordantly along foliation planes of the country rock with a doming produced near the center of the intrusion with outward quaquaversal dips. Osborne (1936, p. 426–427) believes that only the border zones of the batholith described by Foye contain so many inclusions.

BLACK HILLS DOME, SOUTH DAKOTA: The following description of the Black Hills Precambrian granite domes is summarized from a report by Runner (1943). The granites of the Harney Peak area are a composite of many sills, tongues, dikes, and irregular masses of various compositions and ages. Within the area are many xenoliths of sedimentary rocks which in the central part are composed of metalimestone and amphibolite. The bedding planes and the axial planes of the isoclinal folds in the sedimentary inclusions in the interior dip outward from the central axial region and form a well-defined xenolith dome. The growth of the dome is believed by Runner to have been from the center outward by marginal intrusion on the border of a laccolithlike structure. As the structure increased in size, marginal dips steepened.

The formation of the Harney Peak dome was preceded in the area by overthrust faulting and recumbent folding. Space for the granite according to Runner was made by domal uplift, lateral spread, and replacement. He suggests that the domes of the southern Black Hills probably coalesce below the schistose sedimentary cover into a major Precambrian batholith. The foliation, he infers, has been produced by flow in the liquid state, replacement of bedding, multiple intrusion, and shearing of solid granite. Many inclusions were isolated by coalescence of parallel sills and by intersecting dikes and sills and were never engulfed in liquid magma. The age of some granite pegmatites in the Black Hills has been determined to be about 1600 m.y.

Tectonic domes and folds of plastic crystalline flowage.—Quirke and Lacey (1941) have concluded that many complex batholithiclike domes with invasive relationships shown by their diverse facies may arise from "mutual plastic invasion of the rock layers by solid flow" under conditions of deep-zone deformation, but this interpretation has not received much application.

NORTHWEST ADIRONDACK AREA: Several bodies of orthogneiss (Fig. 15) with a composition ranging from syenite to granite have been described from the northwest Adirondacks by Buddington (1948, p. 24–30). The rock of all these bodies has a granoblastic texture and evidences of complete recrystallization under conditions of high-grade metamorphism. They are inferred to have structures formed by plastic doming and anticlinal deformation (isoclinal folding at extreme) of original sheetlike or gently phacolithiclike differentiated layers of igneous rock. There is no evidence of any granitization or migmatization in connection with the remobilization and development of these domes and anticlines as in the case of the reactivated domes described by Eskola, although pressure of rising magma beneath the anticlines and domes may have been a factor. The domes are not rheomorphic in the sense that their reactivation has resulted in intrusive relationships to country rock.

Tectonic domes of remobilization with introduction of fluids.—Eskola (1949) set forth a concept of the development of domes in a second period of orogeny which has received wide acceptance. The hypothesis envisages a plutonic mass of an early orogeny later eroded and mantled with sediments. During a later orogenic cycle fluids or new granitic magma was injected into the older pluton at the same time that it was deformed into gneiss with accompanying migmatization and granitization or palingenesis. The old pluton was thus mobilized anew, and associated younger intrusive magma may display an intrusive relation to the mantle rocks.

Eskola (1949, p. 470) suggests that the domes in Maryland described by Broedel (1937) are of such an origin. Precambrian granite gneiss was reactivated in Taconic (?) time by intrusion of granite and granitization. He also suggests that these mantled domes occur in orogenic zones and have apparently been formed under the influence of horizontal thrust movements, although the doming itself is inferred to be due to vertical movements of granitic masses, most if not all of which were caused by swelling during granitization and soaking with granitic magma.

In a later paper Eskola (1952, p. 126) emphasizes that in some domes the element of later granitization is absent or only incipient.

Further discussions of the problems involved in such domes may be found in the papers by Kranck (1954) and by Balk (1946).

Phacoliths

Introduction.—The term phacolith was introduced by Harker (1909, p. 77–78) for concordant intrusive bodies introduced concurrently with folding. He states that the situation, habit, magnitude, and form of the phacolith are all determined by the circumstances of the folding itself and that the ideal type of phacolith is subject to many modifications, in accordance with the varying mechanical conditions of intrusion. Harker also suggests that originally concordant relations may be obscured, owing to the igneous rocks becoming involved in later folding. The original phacolith described by Harker is a dolerite intrusion in a relatively gentle anticline. Most intrusions to which the term has been applied since, however, are syntectonic granitic types in highly deformed rocks and may themselves have been subjected to strong post-consolidation deformation. The phacoliths are characteristically much thickened on the anticlinal plunging noses or in the plunging ends of synclines. They commonly range in size between a mile and a score of miles in length and may be up to several thousand feet in thickness.

Phacolithic intrusions emplaced in the catazone are common throughout the world and have been especially described from the Precambrian shield areas. Excellent descriptions of granite phacoliths in Africa have been published by Gevers and Frommurze (1929) and by Poldervaart and Backström (1949). Several North American examples, all of Precambrian age, are reviewed here to illustrate this mechanism of emplacement.

Phacoliths of Grenville subprovince, Canadian shield.—Phacoliths are abundant in the high-grade metamorphic Precambrian rocks of the Grenville subprovince of the Canadian shield where they have been referred to by Wilson (1925, p. 397). Osborne (1936, p. 426), and Hewitt (1953, p. 92–93), and have been described throughout the Adirondack outlier by Buddington (1929b; 1948; 1956, p. 115–117), Reed (1934); Cannon (1937), and Dietrich (1954). Some of the Adirondack phacoliths occur in marble, and all have a homogeneous and narrow range of composition (Buddington, 1957, p. 295). These relationships along with others make it highly improbable that they were emplaced by replacement but rather as magma. Nearly all those in the marble, 15 in all, have come into anticlines, most of them into anticlines parallel to the major trends, but some into anticlines or synclines that are cross-folds. Extensive phacoliths of replacement origin, however, in many places accompany those of magmatic intrusion.

Phacoliths of the New York-New Jersey highlands.—Phacoliths are also abundant in the highlands belt of Precambrian metamorphic rocks in New York and New Jersey. A synclinal phacolith has been described by Lowe (1950). The granite occurs as a synclinal sheet with a greatly thickened trough and one well-developed limb. Lowe infers that absence of secondary foliation and lack of tectonic fabric patterns in the granite indicates post-tectonic emplacement. He proposes the concept of "exchange of space" between the magma rising and the country rocks subsiding into the emptying magmatic chamber to account for the lack of evidence indicating lifting of the overlying rocks by forcible injection of the granite. The present writer has studied similar granitic plutons some miles to the southwest, and for these there is adequate deformation and recrystallization in much of the rock to justify

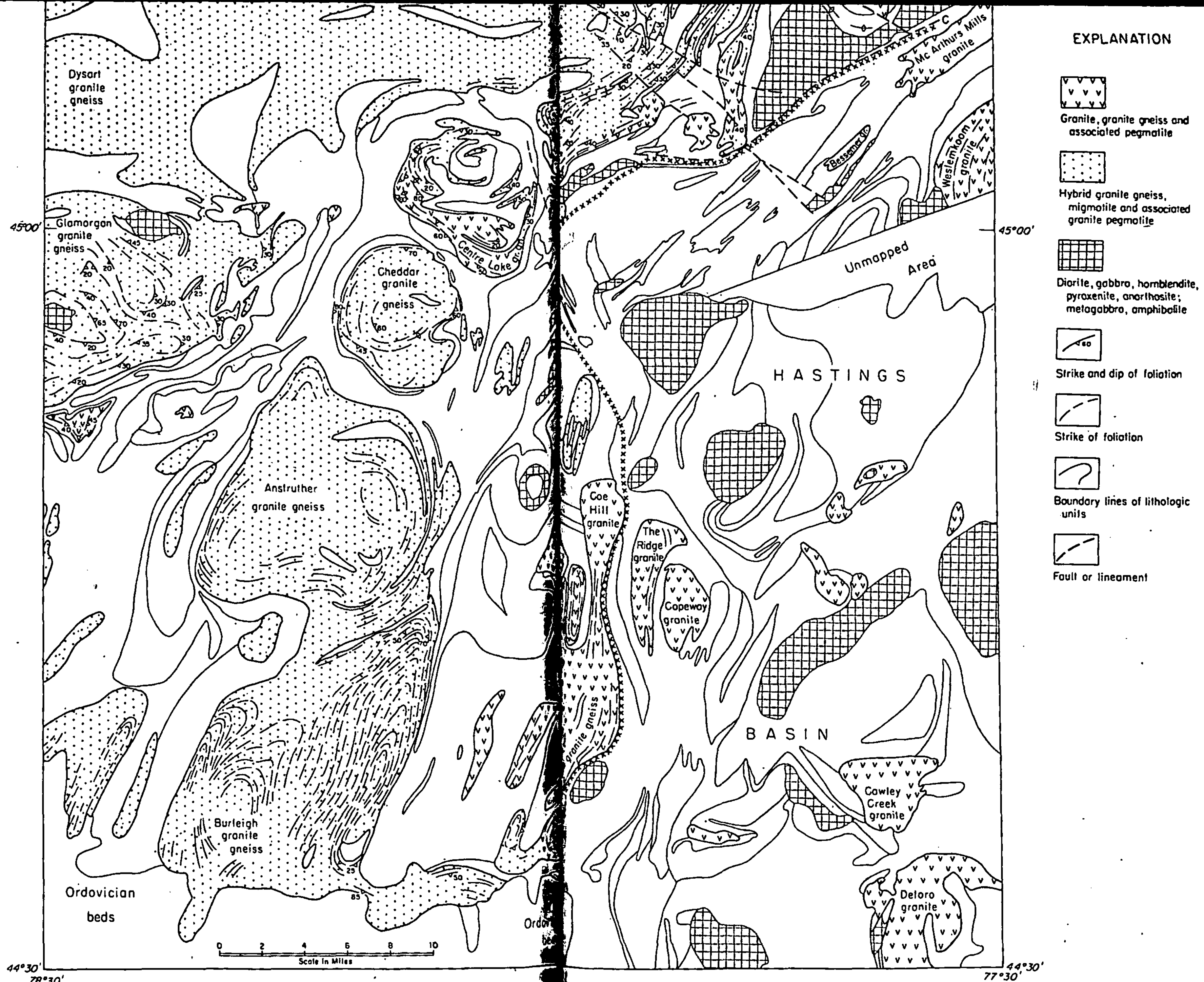

Figure 14. Geology of Part of Ha[illegible]t Area, Canadian Shield

Metasedimentary and metavolcanic rocks of Hastings Basin southeast of line of crosses are o[illegible]ate-grade metamorphism and include conglomerate, argillite, and pelitic schists, blue-limestone [illegible] greenschist facies. Metasedimentary and metavol[illegible]t of line of crosses are in high-grade metamorphic facies and include migmatites, paragneisses in [illegible] metadiorite [illegible]ite schist from basic volcanic rocks. The granites of Hastings Basin were emplaced in the [illegible]

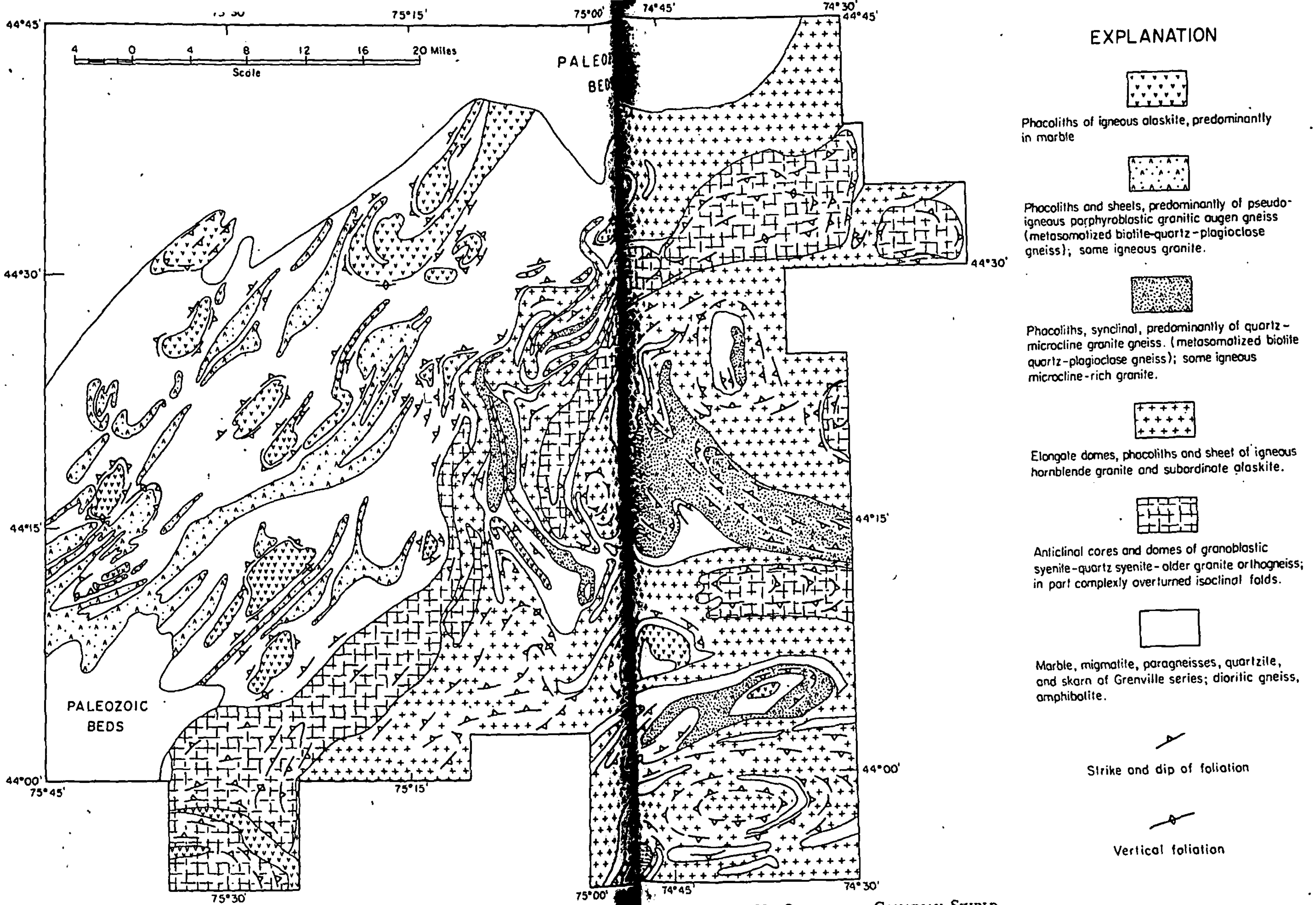

FIGURE 15.—PRECAMBRIAN PLUTONS OF CATAZONE, N ACK AREA, N. Y., OUTLIER OF CANADIAN SHIELD
Phacoliths of igneous origin, pseudo-phacoliths of metasomatic origin, co igneous domes, and tectonoplastic domes and anticlinal cores of orthogneiss

considering them as late tectonic emplacements and as phacoliths.

A phacolith of pyroxene and hornblende syenite gneiss has been described from the New Jersey highlands by Buddington (1956, Fig. 6). It occurs on a steep anticlinal fold, is more than 12 miles long on one limb with a thickness amount on the plunging nose of the anticline. Younger granite also occurs as a younger phacolith flanking the syenite gneiss (Buddington, 1956, Fig. 6) on the same anticline. In the area to the northwest of this composite phacolith Hague *et al*. (1956, p. 459) describe phacolithic Byram granite gneiss. They

the Byram gneiss and other rocks indicates that the Byram was formed either as a phacolithic intrusion or by replacement of a large metasedimentary sequence and that most of the field and microscopic evidence points to an igneous origin. They conclude that the Byram igneous intrusion coupled with partial replacement.

Phacolithic emplacement of plutons in the New Jersey highlands has also been described by Hotz (1953, p. 185–192) and by Sims (1953, p. 265–268).

Wolf Mountain phacolith, Texas.—The Pre-described by Stenzel (1936). He suggests that the granite intruded as a phacolith into the trough of a syncline that pitches on an average 16° SE. The granite is underlain by gneisses and overlain chiefly by schists. The intrusion he believes took place toward the end of the period of folding of the country rock and was accompanied by the stress that produced the folds. Stenzel infers that the feeding channels of the intrusive body are in a long shear zone, which extends along the phacolith and cuts across the schistosity of the country rock. Thus the magma, after rising in this moving shear zone, spread out into the syncline along the boundary between gneiss and schist.

Harpoliths.—The angle of pitch of the axes of phacoliths may range from gentle to 90°. The name "harpolith" was originally introduced by Cloos (1921, p. 44–47, 84–85) for intrusions of sickle-shaped form emplaced contemporaneously with the formation of cross folds in steeply folded rocks. There are numerous harpoliths in the Precambrian of the Adirondack area. The synclinal phacolith described by Dietrich (1954) could be called a harpolith.

Replacement pseudo-phacoliths.—Some phacolithiclike granitic masses are in large part the product of granitization and metasomatism, such as the Hermon pseudo-phacolith of the northwest Adirondack area, New York, and the sheet of granodioritic gneiss of the Manawan Lake area, Saskatchewan, described in sections that follow.

Ambrose and Burns (1956, p. 49–52) have inferred that the granite sheet conformably surrounding the Clare River syncline of the Grenville series in Ontario is of replacement origin, primarily because of the general conformity of long thin septae of limestone and the lack of disturbance which they infer should accompany magmatic emplacement. The present writer, however, is convinced of the possibility of essentially conformable syntectonic emplacement of magma in folded rocks.

Quirke (1929) has described a series of Precambrian intrusives from the French River area, Ontario, which he calls batholiths. His description of the structural relationships, tectonic history, and his interpretation of their origin, however, permit them to be called replacement phacoliths. The country rocks are metasedimentary rocks and migmatites. The plutons consist primarily of granitic and syenitic rocks. The major structures of these areas ac- [illegible] by a unifying its apex to the north, to which converge anticlines, synclines, and fault lines. The batholiths conform to the country gneisses of this structure. The masses of plutonic rock are in general small, less than 15 miles long and less than 5 miles wide. Quirke states that the intrusive lenses are inclined to widen along the axial region of the great syncline; indicating that intrusion and folding were closely connected in origin. The granitic rocks appear to him to be replacements of sedimentary rocks, and he cites as one line of evidence that the phenocrysts in some certainly have grown within gneisses which still are easily distinguishable as sedimentary rocks, and that these gneisses grade into masses which are so exclusively porphyritic that no trace of other structure or texture remains visible.

Phacolithic and pseudo-phacolithic emplacement, Manawan Lake area, Saskatchewan.—Complex phacolithic emplacement appears to be well exemplified in the Manawan Lake area, Saskatchewan. The area has been mapped and described by Kirkland (1956), and a part of the geologic map is shown in Figure 16. The plutons were not designated as phacoliths by Kirkland, but the structural data given are consistent with such an interpretation. The rock mapped as granodiorite gneiss is described as a strongly foliated or finely gneissic rock composed of quartz and feldspar with hornblende the most abundant mafic mineral. In many places minor amounts of nodular meta-arkose, biotite gneiss, cordierite-biotite gneiss, and hornblende gneiss also occur. The granodiorite gneiss occupies the same position on the east side of the Lake Manawan dome (L.M.) as the meta-arkose does on the west side, and the granodiorite gneiss is inferred by Kirkland to be a more highly metamorphosed granitized equivalent of the meta-arkose.

The conformable phacolithiclike core of the Lake Manawan dome consists of leucocratic porphyritic to even-grained granodiorite. The plagioclase exhibits albite and Carlsbad-albite twins. The rock is generally massive but in places weakly foliated. The granodiorite is interpreted by Kirkland as intrusive.

Subcylindrical Plutons

Wynne-Edwards (1957) has described the Westport pluton in Ontario as that of an almost vertically plunging cylinder emplaced in a vertical cylindrical fold or "vortex" that formed a natural vertical channel for the uprise of and/or granitic emanations. The lindrical fold resulted from a deformation of eviously isoclinally folded metamorphic cks of the Grenville series that caused rota- ion about a vertical axis expressed by buckling competent bands along the strike and in ertain places of flowage of incompetent layers nto "vortices". The pluton is formed in part y gabbro but in large part by monzonite that eplaces gabbro, paragneiss, and marble. The luton is about 15 square miles in area, and the oliation of the country rock is conformable ith the margin of the mass. Numerous relics country rock occur within the pluton, and heir foliation conforms to the attitude of the xternal mantle. There are several such plutons a row close together. The plutons are in- erred by Wynne-Edwards to be post-orogenic, mplaced in a dilatant zone with no indication flow structure. A syntectonic mechanism of emplacement is considered improbable because of lack of flow structure or of post-emplacement deformation.

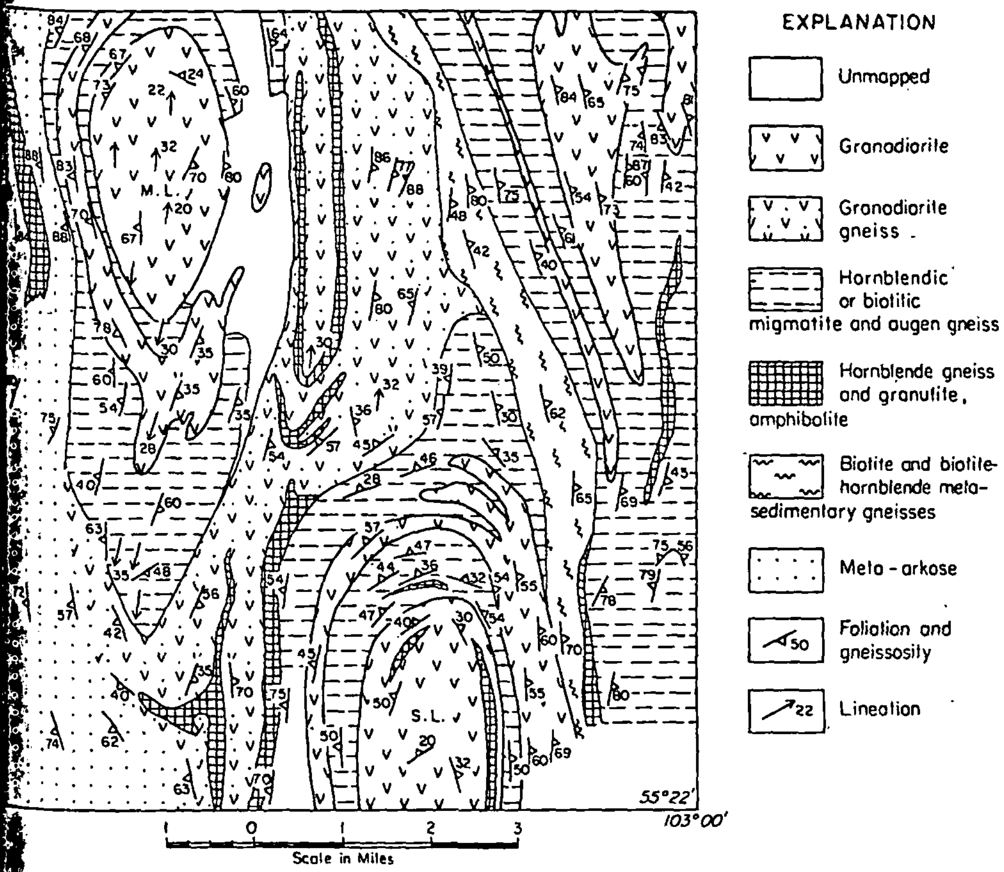

FIGURE 16.—COMPOUND GRANODIORITE PHACOLITHS (M.L. AND S.L.) OF INTRUSIVE ORIGIN AND PSEUDO-PHACOLITH OF GRANODIORITIC GNEISS (RECRYSTALLIZED AND GRANITIZED META-ARKOSE IN PLACE), CATAZONE

Modified after part of Manawan Lake area, Saskatchewan, by S. J. T. Kirkland (1956)

Emplacement in the Grenville Subprovince, Canadian Shield

General statement.—The Grenville subprovince of the Canadian shield includes a belt more than 250 miles wide and more than 1000 miles long consisting preponderantly of uniformly high-grade metamorphic rocks and igneous plutons of the catazone. The province includes the Precambrian outlier of the Adirondack area in New York. The rocks of the highlands of New Jersey are similar. Many aspects of the area are discussed in *The Grenville problem* (Thomson, 1956). Age determinations (Shillibcr and Cumming, 1956; Eckelmann and Kulp, 1957) lead to the inference that the

the Grenville series of the general Haliburton-Bancroft area, Ontario, are 800–900 m.y. and 1050 ± 20 m.y. old.

Plutons in the Grenville series of Quebec.—Osborne (1947) has had much experience with the geology of the Quebec part of the Grenville subprovince, and the following statements are based on some of his conclusions. The type locality for the Grenville series lies within a dejective zone. Within this zone the intrusives tend to be concordant with the Grenville rocks. Between the dejective zones are broad areas characterized by schistosity parallel to bedding and by gentle dips. In addition to sills in these areas there are batholiths of coarse-grained granite that cut across the structure. Osborne notes that some measure of syntexis is observable, particularly in the dejective zones, and that members of the normal sedimentary series may be missing, in which case a variety of granitic rock occurs in its place. He suggests that magma appears to have been the dominant constituent of the syntectic but that at a few localities the granitic gneisses were formed by granitization of paragneiss. In the Ottawa folded belt Osborne (1936, p. 426) finds that most of the intrusives as a whole partake of the nature of phacoliths.

Plutons of northwest Adirondack area, New York.—Several types of emplacement of plutons in the catazone are well represented in the Grenville series of the northwest Adirondack area, New York, and outlier of the Canadian shield. Representative structural relationships are shown in Figure 15.

In the northwest part of the area is a belt about 25 miles wide in which members of the Grenville series are predominant. The granitic plutons occur as alaskite phacoliths of intrusive magmatic origin, mostly in marble, and as sheets of porphyroblastic augen gneiss formed largely by metasomatism of biotite-quartz-plagioclase gneiss interbedded with marble. Most of the phacoliths have moderate plunging axes parallel to the major trend of the formations, but a few are in crests or troughs of cross-folds plunging nearly at right angles to the trend of the formations. The phacolith at the extreme northwest is emplaced in a cross-fold plunging southeast in beds isoclinally overturned to the northwest with steep dips. Several small phacoliths along the northeastern half of the southeast border of the Grenville belt are also on cross-folds, here plunging northwest. The phacoliths along the northwest and southeast portions of the Grenville belt are in zones southeast respectively and are granoblastic gneisses. In the central part of the belt, however, there are phacoliths with gneissoid mesoperthite granite.

The largest body of pseudo-igneous porphyroblastic augen gneiss in the Grenville belt is the Hermon pseudo-phacolith (Fig. 15, Lat. 44°25′ Long. 75°15′ to Lat. 44°08′ Long. 75°45′). The rock is predominantly a gneiss with augen of microcline in an even-grained gneissic groundmass. The phacolith has a maximum length of 35 miles and a width of 1-½ miles. The granitic mass is continuous at the southwest around the plunging nose of a subordinate anticline and the trough of a syncline. The granite mass transgresses the trend of a belt of biotite-quartz-plagioclase gneiss from near one side of the stratigraphically upper part of the gneiss to marble at the base. There are all gradations between porphyroblasts in the biotite-quartz-plagioclase gneiss, porphyroblastic schlieren of the country rock in the granitic mass, and uniform granitic augen gneiss. Migmatic facies of the biotite-quartz-plagioclase gneiss are also associated. The granitic gneiss is variable in composition. A syenitic facies is developed locally in mixture with amphibolite. All these phenomena have led to the interpretation that the granitic gneiss is of metasomatic origin. There is also some even-grained granite associated which is inferred to be of direct magmatic origin. There are some bodies of inequigranular granite or granite gneiss intruded as isolated sheets in marble. These may represent remobilized or partly anatectic material, but we have no critical evidence. The solutions effecting the metasomatic development of the augen gneiss are inferred to be related to magmatic masses below.

The structure of the area dominated by igneous rocks, pseudo-igneous rocks, and orthogneisses in the southeastern part (Fig. 15) has been controlled by the anticlinal folds and domes of granoblastic syenite-quartz syenite older granite orthogneiss that have served as relatively rigid buttresses. The metamorphism and deformation of these rocks along with the Grenville series of beds preceded the emplacement of the younger granitic plutons. The magmas yielding the syenite-quartz syenite older granite rocks are inferred to have been emplaced as relatively flat-lying sheets or gently dipping phacoliths of the epizone or transitional epizone-mesozone. The northeastern part of the western belt of orthogneiss is a limb of an anticline overturned to the sou[illegible]thern and eastern anticlines of orthogneiss served as floors for the phacolithic emplacement of the younger granite athough the southern part of the northwest limb of the northern belt of orthogneiss was locally cut out by the younger granite.

There are several synclinal phacoliths of quartz-microcline granitic gneiss. These represent metasomatized biotite-quartz-plagioclase gneiss (Buddington, 1957) in part and in part of quartz-microcline igneous rock of magmatic origin. Much of the metasomatic rock is sillimanitic. The biotite-quartz-plagioclase gneiss is the same type of rock that was replaced to yield the porphyroblastic augen gneiss in the Grenville belt.

The complex in the southeastern part of the area (Fig. 15) is thus composed of tectonoblastic domes and anticlinal cores of granoblastic orthogneiss of an early period; igneous domes, phacoliths and sheets; and pseudo-phacoliths of metasomatic granite gneiss.

Plutons of Haliburton-Bancroft area, Ontario.—The study of the Haliburton-Bancroft area in Ontario by Adams and Barlow (1910) has made this a classic area for the portrayal of certain aspects of batholithic emplacement. A revised geological map (Ontario Dept. of Mines, Map 1957b) and a revised interpretation of the geology by Hewitt (1956, p. 22–41) have recently been published.

The belt of high-grade metamorphic rocks northwest of the Hastings Basin (Fig. 14) includes marbles and silicated marbles, basic volcanic rocks largely altered to amphibolite schists and gneisses, metagabbro and metadiorite gneiss, paramphibolites, and paragneiss containing sillimanite and garnet. Hewitt describes the batholiths in the high-grade metamorphic terrane as mixtures of granitic material ranging from granitic gneiss to pegmatite intimately injected into and replacing paragneiss and amphibolite. Abundant inclusions and schlieren are present, and gradational hybrid facies of mixed origin are common. Most of the granitic bodies are concordant, and there is much evidence of granitization and metasomatism.

Batholithic Development of Pseudo-igneus Granite in Catazone

Introduction.—The occurrence of replacement pseudo-phacoliths and sheets in the catazone has been discussed. A few examples liths will now be considered. Many have been described from the Precambrian, especially the Canadian shield. Early papers advocating the emplacement of batholiths by replacement and recrystallization were those of Quirke (1927) and Quirke and Collins (1930). Some recent examples are those of Harrison (1949), Christie (1953), Robertson (1953), Steven (1957), and Eckelmann and Poldervaart (1957).

Harrison (1949, p. 34–39) described the rocks of the File-Tramping Lakes area, Manitoba, and concluded that granitization has locally affected all volcanic and sedimentary formations in the area and that it has taken place on a regional scale and has locally been intensive enough to produce granite. He infers further, however, that abundant evidence indicates that magmatic granite also existed in large amounts.

Robertson (1953) has described the rocks of the Batty Lake area, Manitoba. He finds that gneissic, granitelike bodies occur in the Batty Lake area as bodies of batholithic size, as stocklike and sill-like bodies, and as pegmatite bodies. Viewed in aerial photographs, the larger bodies exhibit complex to broadly sweeping folds resembling those of sedimentary formations, but in the outcrop they are granodiorites, tonalites, and granites, compositionally, with well-defined foliation and grading imperceptibly into rocks. apparently of sedimentary origin. He concludes that "granitization" commences with the development of albite-oligoclase in the sedimentary rocks, producing rocks mapped as "granitized" gneisses, and continues with the later formation of microcline to form bodies of granitoid gneiss that may, in some instances, have become mobile. Robertson suggests that the cause of regional metamorphism and "granitization" in this area is the proximity of magmatic material at depth.

The development of quartz monzonite gneiss and extensive granite pegmatite veining by metasomatism has been described in detail by Steven (1957) from the Precambrian of the Northgate district, Colorado. The country rock is hornblende gneiss, and remnants are abundant in the quartz monzonite gneiss. Metasomatism has been effected by tenuous silica- and alkali-bearing solutions. The gneiss is inferred by Steven to have locally become mobile, moved as a plastic crystalline diapirlike mass, and developed with its foliation in the form of a funnel.

area in Saskatchewan shows very well the kind of phenomena that have led to the inference of batholithic emplacement by metasomatism. The following description is based on that of Christie (1953), and two selected areas from his map are shown in Figure 17. More than 50 per cent of the Goldfields-Martin Lake map-area is underlain by a complex of granite, granite gneiss, and granitoid gneiss. Christie concludes that the granites have been emplaced mainly by a granitization or replacement process, although various small bodies such as the Mackintosh Bay granite may have been emplaced as a molten magma. The Mackintosh Bay stock is shown in the lower part of area B of Figure 17, and Christie describes it as gneissic with the appearance of having thrust aside the enclosing sedimentary strata during emplacement. The foliation of the granite near the contacts is everywhere parallel to them. Within the stock the foliation has a rough elliptical plan, and lineation indicates a plunge of about 35° SE.

Christie states that in general, although the contacts of granitic rocks and amphibolite are sharp, the contacts with quartzites are commonly gradational over tens, hundreds, or even thousands of feet. Typical coarse-grained pegmatite dikes or sills are rare except in the metasedimentary rocks north and northwest of the Mackintosh Bay granite stock.

The foliation of the granitic rocks near contacts with metasedimentary relics dips gently or moderately but tends to be steep or vertical away from them.

The evidence for emplacement of most of the granitic rocks by granitization is based by Christie largely on gradational zones with quartzite, evidence for a complex series of replacements indicated by interpretation of microtexture, and the lack of displacement of most relict structures. The latter is exemplified by the inclusions outlining a skeletal fold in the upper part of area B of Figure 17.

Quad Creek area, Beartooth Mountains, Montana.—A detailed study of the Quad Creek area of Precambrian age in the Beartooth Mountains has been published by Eckelmann and Poldervaart (1957). Age determinations of these rocks by Gast and Long (1957) based on Rb-Sr place them in general between 2730 and 2800 m.y. The following description is a summary based on the report of Eckelmann and Poldervaart. The Beartooth Mountains form an elongate range with longer axis trending northwest and consist of a core of granitic gneiss flanked by migmatites and metasediments. The historical development is believed to be (1) original deposition of an Archean sedimentary sequence; (2) emplacement of metagabbro and ultramafic intrusions, followed by folding—fold axes strike north-northeast; (3) regional metamorphism and granitization resulting in a core of granitic gneiss and mantles of migmatites and metasediments with boundaries trending northwest. The last expression of granitization was the production of pegmatites. Eckelmann and Poldervaart believe that their studies indicate in-place formation of granitic gneiss. Fold axes pass continuously and without deflection from the mantle of metasediments and migmatites across the boundary zones into the core of granitic gneiss, although the folds intersect the boundary zones at 40° to 50°. The boundary zone consists of intersecting tongues, migmatites, and granitic gneiss, and these rocks grade into each other along and across the strike. In the boundary zone more resistant rock types persist at definite horizons, continuous with skialiths of similar rocks in granitic gneiss. Throughout, foliation in granitic gneiss and banding in migmatites parallel bedding in metasediments. Growth phenomena shown by zircons of different rocks they believe also indicate autochthonous formation of granitic gneiss at temperatures probably about 500°–600°C. Eckelmann and Poldervaart conclude that granitization was effected by migrating alkaline aqueous solutions during a prolonged Archean cycle of thermal activity. The following statements are also from their description and from a personal communication from Poldervaart. The rocks of the core consist mainly of granitic gneiss, in part showing banding, with many migmatite layers and some metasediments. The zircons are a rounded type with overgrowths and outgrowths. The more homogeneous granitic facies, in particular, have some euhedral zircons. The rocks of a boundary or transition zone comprise migmatites and granitic gneiss with some metasediments. The more nearly homogeneous granitic facies have zircons like those of the core, whereas those of the more inhomogeneous areas are similar to those of the mantle. The rocks of the mantle are mostly migmatites with associated metasediments and some granitic gneiss. Rounded zircons, rounded zircons with outgrowths, and rounded zircons with overgrowths are all present. The first two are about equal in quantity. There thus appears to be a gradation in the transformation of zircons during granitization. The core of the Beartooth block consists predominantly of

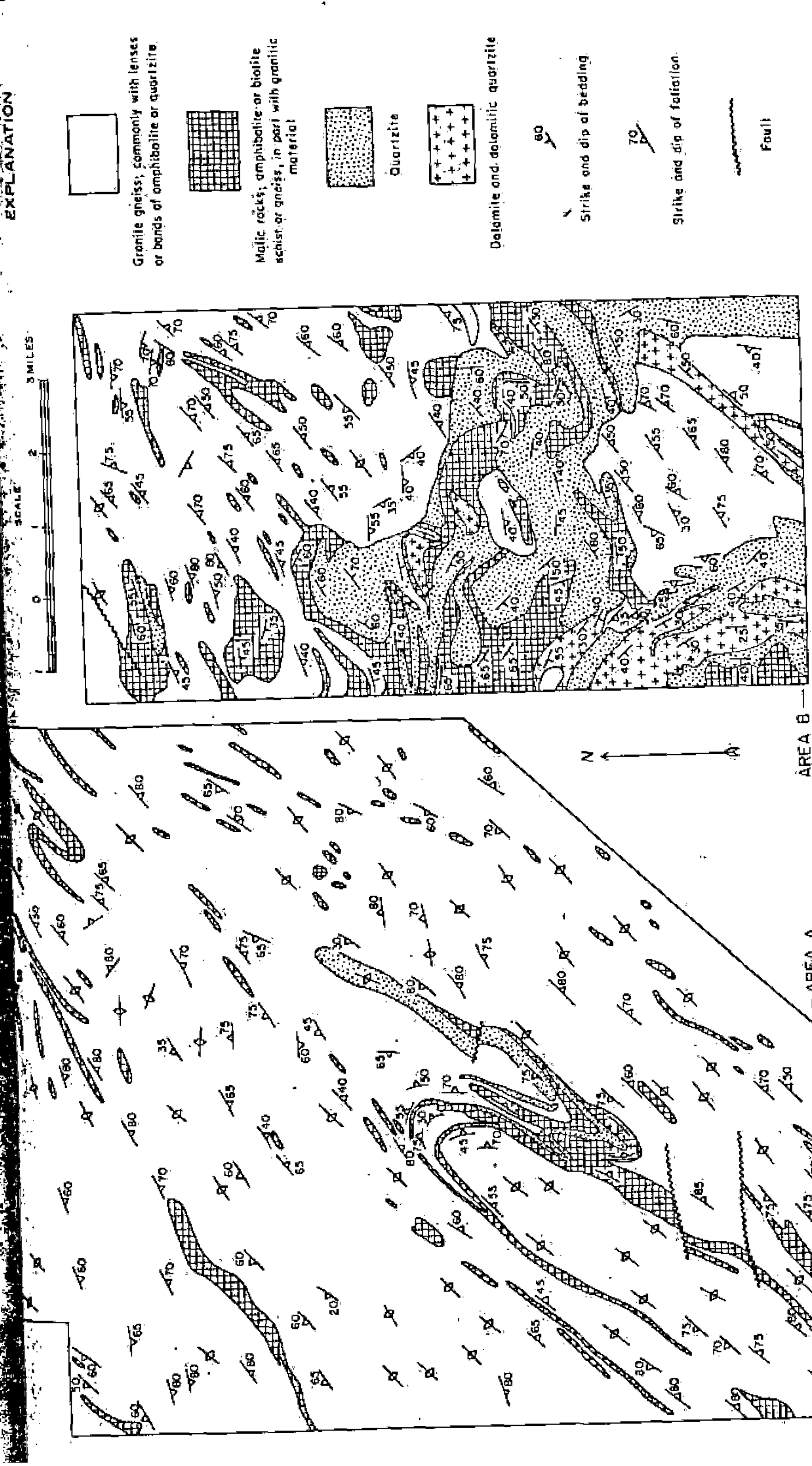

FIGURE 17.—TWO AREAS OF PRECAMBRIAN BATHOLITHS

Catazone of Canadian shield from the Goldfield-Martin Lake area, Manitoba, modified from Christie (1953). Granitic gneiss is inferred by Christie to be largely of origin by granitization with relics of folded country rock in area A and upper part of area B. Stock of intrusive granite gneiss with displacement of beds in lower part of area B.

pink leucocratic granitic gneiss, with tonalitic gneiss developed toward the migmatitic boundary zone.

The authors emphasize that field relations are critical in establishing the metasomatic hypothesis. Insofar as structure alone is concerned, however, the following alternative interpretation might be posed as a question. Could the pluton have been emplaced between synclinal leaves of country rock as magma wedges that were relatively very thick at the south and thin at the north so that a marked constriction of the magma wedges occurred along the pseudo-discordant boundary of granitic core and mixed rocks? Supplemental granitization would accompany the magma.

Complexity of Precambrian Plutonic Complexes

General.—The Precambrian plutonic complexes of any area of considerable size normally comprise a complex of granitic plutons that have been emplaced in different zones at different times. Anderson, Scholz, and Strobell (1955) have described a Precambrian complex of the Bagdad area, Arizona, where the earliest intrusive members are epizonal plutons of rhyolite and alaskite porphyry, followed by mesozonal plutons (age 1,600 m.y.) in Precambrian schists of intermediate grade of metamorphism. Kalliokoski (1952) has described similar relationships from the Weldon Bay area, Manitoba, where a Precambrian epizonal stock of fine-grained granodiorite with a porphyritic quartz latite border facies is interspersed with younger Precambrian batholithic intrusions that have developed migmatites with adjoining schists appropriate to the mesozone or catazone.

Grenville belt.—The writer has estimated that over a third of the igneous rocks (including orthogneisses) of the Adirondack area of Grenville rocks belong to the syenite-quartz syenite-granite series such as form the cores of tectono-plastic domes and anticlines (Fig. 15). These are inferred to have been emplaced originally in the deeper part of the epizone although much of the rock now belongs to the granulite metamorphic facies. Granite, perhaps 100–200 million years younger, forms 40–50 per cent of the igneous rocks and was emplaced in the catazone.

The Grenville subprovince in the Haliburton-Bancroft area (Fig. 14) includes a belt, the Hastings Basin, about 20 miles wide, within which the rocks are of a low to intermediate characteristics of the mesozone in contrast to the broad belts on each side of high-grade metamorphic rocks with plutons of the catazone. The rocks of the Hastings Basin consist in part of the Hastings series which is thought by some geologists to be younger than the Grenville series, but by other geologists to be part of the Grenville series. Hewitt (1956, p. 30) writes that

> "The Hastings Basin consists of a terrane of low to intermediate grade of metamorphism, including schists, argillites, well-bedded blue limestones, crystalline limestones, and volcanics. Sedimentary and volcanic structures such as crossbedding, grain gradation and pillows, are frequently well preserved."

The basic lavas frequently belong to the low-grade chlorite facies. Hewitt describes the Deloro granite stock as consisting of a fine- to medium-grained granite with sharp contacts and the McArthurs Mills granite stock as consisting of a massive, coarse-grained granite with irregular shape and discordant structural relations to the country rock. The intrusive plutons in the Hastings series thus have both concordant and discordant contacts, and many have a contact-metamorphic aureole characteristic of the mesozone. The present writer notes that the Marmora metasomatic iron-ore deposit is a fine-grained magnetite-pyroxene-epidote tactite of a type normal for the upper part of the mesozone and dissimilar to the characteristic magnetite-bearing skarns in the Grenville rocks of the catazone.

Colorado Front Range.—Portions of the Precambrian complexes in the Colorado Front Range have been intensively studied and afford an excellent example of their complexity.

The oldest country rocks of the area now consist of high-grade metamorphic biotite-sillimanite schists, quartz-biotite gneiss and schist, quartzite, and hornblende gneiss and amphibolite. The sillimanite-bearing rocks have more aplite and pegmatite.

Plutons of the catazone are well exemplified by the quartz monzonite gneiss (Fig. 18) bodies. Lovering and Goddard (1950, p. 23) describe these rocks as concordant and nearly everywhere parallel to the foliation, with well-developed gneissic structure, closely associated lenticular bodies of pegmatite, and intense *lit-par-lit* injection of inclusions of schist with some assimilation. It may also be noted that a few miles southeast of Central City the quartz monzonite gneiss has typical catazonal phacolithic relationship to a complex isoclinal [illegible] structure within hornblende gneiss

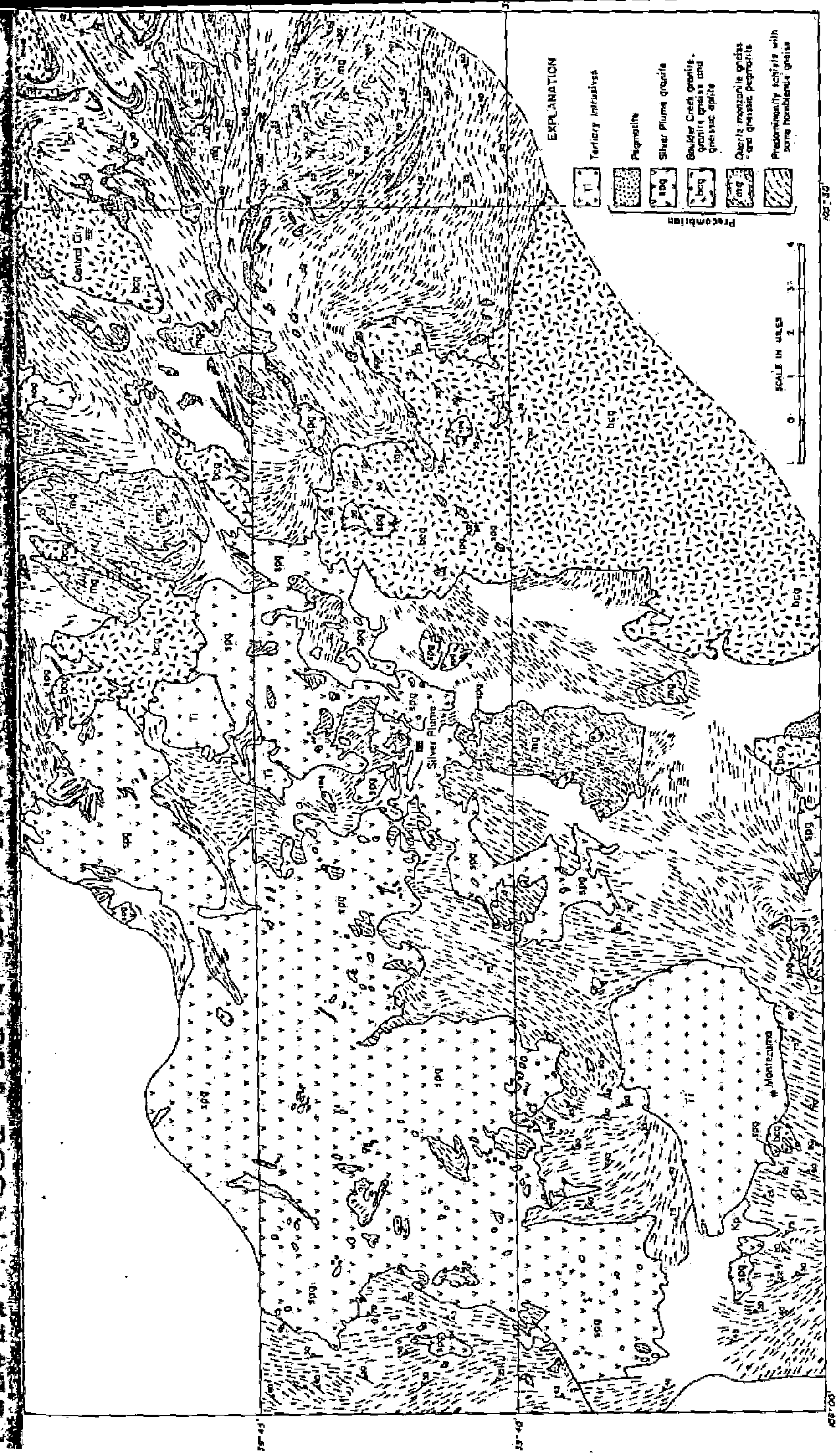

FIGURE 18.—GEOLOGY OF A PART OF THE FRONT RANGE, COLORADO

Tertiary stocks emplaced in epizone, Silver Plume granite batholith in mesozone, Boulder Creek granite batholith in transitional mesozone-catazone, and quartz monzonite gneiss stocks and phacoliths in catazone. Modified after map by Lovering and Goddard (1950)

Phacoliths also occur within the schists of the Freeland-Lamartine district to the west described by Harrison and Wells (1956, p. 54) as mostly bodies of biotite-muscovite granite that are generally concordant, although some are sharply discordant. Many of the bodies are hook-shaped and crescent-shaped in their surface exposures and are in the axial regions of folds.

Boos and Boos (1957, p. 2615–2617) state that the probably related Mt. Morrison quartz monzonite gneiss is in part saturated with ill-defined pegmatite; they suggest that it is of granitization (palingenesis and metasomatism) origin.

The Boulder Creek granite (Fig. 18) is described by Lovering and Goddard (1950, p. 25–26) as a quartz monzonite to sodic granite slightly younger and less metamorphosed than the quartz monzonite gneiss previously discussed. The granite is further described to have primary gneissic structure, less well developed but still discernible in the cores of large masses, and locally with abundant inclusions that rarely show much evidence of assimilation. Boos and Boos (1957, p. 2616–2617) state that no other granite of the Front Range has produced so many aplite dikes and sills. They ascribe the granite to a combined magmatic and metasomatic origin. The plutons of Boulder Creek granite are largely comformable but in part break across the foliation of the country rocks. Lovering and Goddard (1950, p. 52) state that the foliation and lineation suggest that the individual plutons are funnel-shaped, enlarge upward, and are accompanied by lateral thrusting. Lovering and Tweto (1953, p. 8–16) on the basis of the orientation of the primary planar foliation and linear structure infer that the batholith was emplaced by rise of magma through a central conduit from which it spread upward parallel to the linear structure that plunges 40°–60°N. The schist along the west edge of the batholith dips under the batholith, but at the north the granite dips under the schist. The foliation of the schist is generally conformable with the contact, and the schist is closely seamed with pegmatite. The present writer suggests that the plutons of this granite may have been emplaced in the transition mesozone-catazone.

The Silver Plume granite plutons (Fig. 18) are inferred by Lovering and Goddard (1950, p. 28) to be younger than the Boulder Creek granite bodies. They state that these granites to the generally concordant habit of the [illegible] intrusives. They suggest many local cent intrusion and that in some places the coale of granite masses fed from relatively [illegible] scattered conduits at depth has resulted [illegible] composite batholith. Boos and Boos ([illegible] p. 2616–2618) have suggested that the S Plume granite plutons were emplaced by [illegible] gressive magmatic stoping.

The Longs Peak-St. Vrain batholith has [illegible] correlated by Boos (1934) with the S Plume plutons. The present exposures a sent the roof portion of a batholith, and [illegible] describes it as a "pine-tree" type of emp ment effected by lateral spreading and [illegible] *par-lit* injection of the adjacent and over beds with local folding and tilting, alth the initial conduits were made by stoping [illegible] deep-seated assimilation. There is a grad shown in the walls of the cirques, 3000– feet high, from schist and gneiss at the [illegible] through almost horizontal layers of s separating thick sheets of granite, to mas granite with little schist in the lower [illegible] and floors of the cirques.

The Silver Plume granite plutons appea afford an excellent example where if [illegible] the upper portion were seen it might be terpreted as an example of emplacement i transitional mesozone-catazone, whereas [illegible] deeper parts of the pluton have character diagnostic for the normal mesozone.

A few discordant intrusive stocks of Ter age and emplaced in the epizone add t complexity of the Colorado Front Range.

Mackenzie district, Northwest Territor The batholithic complexes of the district Mackenzie, Northwest Territories, Can have been described by Henderson (1 as consisting of both Archean and Proter intrusions, each on a large scale. The Arc batholithic portion intrudes a series of [illegible] sediments and metavolcanics but is at [illegible] places overlain unconformably by a [illegible] series of Proterozoic formations with [illegible] conglomerates. These beds in turn are truded by granite batholiths. Both older [illegible] younger groups of rocks have members [illegible] belong to only a low-grade stage of [illegible] morphism. Henderson has not discussed [illegible] mechanics of emplacement of the bath The writer notes, however, that the Arc batholithic complexes in part at least ap to show (Geological Survey of Canada, [illegible] 581A) domal structure and that mig [illegible] are well developed in association with [illegible] mesozone-catazone. The Proterozoic sions, in part (Geological Survey of da, Maps 1024A and 1024B), show cross g relations to the country rock and may g to the mesozone. In part the Proterozoic ites (Henderson, 1948, p. 47–48) are ribed as associated with porphyry that ly seems to grade into the granite and to enetically related to it but in places is cut the granite with sharp contacts. Henderson s that no method has been found to nguish one granite from the other except the critical structural relationships can be rmined.

COMMENTARY

General Structural Relationships

study of the literature on the plutons of th America leads to the following com tary. Plutons emplaced in the epizone, cially the "subvolcanic" plutons with tly associated volcanic rocks, would prob y be classed with the atectonic or post onic group, those of the mesozone as -tectonic and post- or occasionally late matic, and the catazonal plutons, pre inantly at least, as syntectonic and syn matic. Also the plutons of the epizone ld belong to the "disharmonious" class as ed by Walton (1955, p. 8–11) in which e is a strong contrast between the energy d of the granite and that of the country as evidenced by contact-metamorphic cts.

he exposed plutons of the mesozone occur lly in belts of eugeosynclinal rocks although y are in general post-tectonic. Those of the one, however, occur (1) in the eugeosyn l belts; (2) in belts of miogeosynclinal ks and structure such as the Sierra Madre ntal of Mexico (*cf.* Concepcion del Oro, 6); and (3) in the Colorado, New Mexico, southern Arizona Rocky Mountain belts ounding the Colorado plateau where the ments are of mainland or intracratonic ynclinal and shelf types and the intrusions ide with belts of faulting and uplift over, Santa Rita, and Organ Mountain ons, New Mexico).. The epizonal plutons y also occur in transverse belts such as that the Boulder-San Juan in Colorado (Fig. 5) one including and extending northeast the Boulder batholith in Montana. The ozoic and Mesozoic sedimentary beds may locally be so thin that Tertiary intrusions in Precambrian rocks are exposed. Tertiary epizonal plutons of the Oregon Cascade Mountains occur in very gently warped Tertiary lavas and in the Cascade Mountains of Washington in lavas and in folded Tertiary sedimentary beds. The question arises as to whether the epizonal plutons outside the eugeosynclinal belts would pass downward into those of mesozonal type. Ewing and Press state (1957) that the thickness of the crust in the Canadian shield and central Interior Plains is 35 km, whereas it is 40–45 km in the western Great Plains and Basin and Range province and 50–55 km in the Rocky Mountain region. This permits the possibility that the Tertiary plutons of the Basin and Range and Rocky Mountain provinces are connected with deep-seated processes. A mesozonal batholith may also occur in an old orogenic eugeosynclinal structure but may have been emplaced following the development of a miogeosyncline or other type of structure in the same region.

A general discussion of the hypothesis of granite by granitization has been published by Perrin and Roubault (1949) and by Perrin (1954).

Dickson (1958, p. 35) has proposed that magma emplacement may take place by a process he calls "zone melting". This involves crystallization of the base of a column of magma to yield latent heat that is in part transferred by rise of fugitive constituents (mostly H_2O) to the top of the column where melting of the roof is effected. The differential concentration of the fugitive constituents at the top of the magma column arises in consequence of the tendency for such materials to move to zones of lower pressure and lower temperature. The quantitative role of the effectiveness of this mechanism of emplacement remains to be determined. There is the problem of adequate time and appropriate physical conditions. It may be a possible accessory factor under favorable conditions in accentuating differential incorporation of rock with low-melting constituents thereby in turn increasing the potentiality for piecemeal stoping and in increasing the intensity of conditions at the roof of mesozonal batholiths.

Reynolds (1958) has summarized in an attractive manner a stimulating hypothesis that involves a combination of granitization and diapiric rise in deeper levels with movement by fluidization and magma development in the upper levels. She writes (p. 382)

> "As a diapir rises, rocks which at a low structural level are obvious migmatites become more and more homogenized by the mechanical kneading caused by superposed movements (Wegmann), and by chemical interchange (with appropriate additions and subtractions, and recrystallization). In this way a migmatite rising in diapir style becomes gradually transformed to nebulite (Sederholm) and eventually to homogeneous granite. If recrystallization outlasts the movement then, just as in salt-diapirs, all traces of the movements will be lost,"

and p. 384

> "It is, however, only where granite diapirs have reached the zone of fracture that *evidence* of melting and the birth of acid volcanics has so far been found".

If the writer understands this hypothesis correctly it involves two assumptions for which we must await adequate support: (1) that there is time for a substantial flow of hot matter to diffuse upward through the diapir, and (2) that in the upper zone something is postulated to happen that affords energy to raise the temperature of solid material to the melting point and to supply adequate latent heat of melting to develop magma. The present writer considers it at least equally reasonable to postulate that hot matter diffusing upward from depths would help to liquify granitic material in the regionally hot catazone before that in the much cooler upper zones. The hypothesis of magma from depth to surface has advantages with respect to ease of effecting homogeneity and adaptability to explain the systematic compositional variations of the different units of composite batholiths and their structural details. Efforts to check on such hypotheses as that of Reynolds, however, would doubtless lead us to new insights.

The epizonal stocks and batholiths have been emplaced at one extreme by columnar subsidence with a 360-degree ring fracture (Ossipee pluton, New Hampshire) and at another extreme by piecemeal stoping of angular blocks (eastern part of the pluton in Northgate district, Colorado). Subsidence of blocks for most plutons probably involves both arc and intersecting angular block fracturing in varying degrees. The alternative development of columnar subsidence or of piecemeal stoping is not related to depth. One factor in aggressive piecemeal stoping as contrasted with permissive subsidence may be the predominance of a tendency for the magma to lift the roof with consequent breaking and subsidence.

Predominance of Mesozonal Batholiths in Plutonic Complexes

The tremendous total volume of batholiths with their directly associated rocks of high-grade metamorphism in plutonic complexes of orogenic belts predisposes one to think of the complexes predominantly in terms of very deep-seated erosion and of batholithic emplacement in the catazone. This, however, is only partly true, for most of the batholiths of most orogenic belts were emplaced in the mesozone (including the transition mesozone-catazone).

There is only one well-authenticated *great* belt of uniformly high-grade metamorphic rocks with associated intrusives in the whole Canadian shield. This is the belt of Grenville-type rocks that extends for over 1000 miles from Labrador to Pennsylvania and is more than 250–350 miles wide. This belt in part has rocks that belong to the granulite facies and is also characterized throughout by the presence of great anorthositic plutons. The ages of the granitic intrusives insofar as now determined range between about 900 and 1100 m.y. At least part of the belt of Precambrian rocks in the Rocky Mountains such as that in Montana, Wyoming, and Colorado may be analogous to the Grenville belt. Granitic rocks with ages of 2700–2800 m.y. and anorthosite bodies occur in this belt.

The batholiths of such provinces as the following, however, were predominantly emplaced in the mesozone: the Keewatin province with a width of more than 250 miles and with rocks intruded by granite pegmatite with ages of 2,500 m.y.; the Yellowknife belt, Northwest Territories, with pegmatites at least 1850 m.y. old; the Colorado Front Range with some granite plutons of about 1000 m.y. age; the belt about 200 miles wide across Newfoundland with plutons of Acadian age; and the zone of Jura-Cretaceous batholiths about 350 miles wide in Alaska and British Columbia. The intensity of regional metamorphism as indicated by rocks farthest from the major intrusives is prevailingly that of the greenschist facies with local zones in the staurolite-kyanite subfacies. The sillimanite facies is generally restricted to zones of rock adjacent to the major plutons or complex of plutons.

Criteria for Large-Scale Granitization

Several examples have been described in the [illegible] in which it has been postulated that large-scale granitization has resulted in batholithic masses. One of the difficulties in proving granitization is that nearly all criteria are subject to alternative interpretations. Much of the rock involved is a leucogranite of a composition approaching that of the experimentally determined ratio for the major minerals concerned at minimum temperatures with H_2O in solution. A satisfactory hypothesis as to why replacement of varied rocks in an open system by alkalic-siliceous solutions should result in leucogranite approaching the composition experimentally determined to be that of the eutectic has yet to be offered. Overgrowths and outgrowths on rounded zircons may develop in contaminated magma as well as during granitization of metasediments in place.

The problem of the proper interpretation of the significance of long thin layers of country rock occurring either as linear slabs or outlining skeletal folds is an old one. (*Cf.* Lawson, 1894, p. 296.) A skeletal fold outlined by metalimestone slabs in the Harney peak batholith is interpreted as xenolithic in intrusive granite by Runner (1943, p. 449–453), whereas a long thin metalimestone layer in granite on the border of the Clare River syncline is inferred by Ambrose and Burns (1956) to necessitate an origin as a relic in granite gneiss of granitization origin. The metasedimentary rocks outlining skeletal folds in the Goldfield-Martin Lake area are also inferred by Christie (1953) to be relics in a batholith of granite gneiss of granitization origin, although the present writer would note that some phacolithic magma emplacement, if not indicated, at least does not seem precluded by the relationships shown on the map (Fig. 17). Folded amphibolite layers and thin amphibolite layers in granite phacoliths of the Northwest Adirondacks are interpreted by Buddington (1929) as xenoliths in granite of magmatic origin, whereas skeletal folds outlined by layers of hornblende gneiss near Northgate, Colorado are explained by Steven (1957) as relics residual in quartz monzonite of granitization origin. Many recent authors emphasize a skialithic as contrasted with a xenolithic origin. Extensive thin layers of country rock do however occur in igneous sills (Eckel, *et al.*, 1949, Pl. 2), laccoliths, and stocks (Murthy, 1957, p. 94). The conditions of emplacement in the catazone are intense and syntectonic, and it seems probable that conformable emplacement would be facilitated. Metasedimentary inclusions in the form of relic folds or skeletal folds in granite gneiss are not necessarily "skialiths" but may be thought of as arising from a complex phacolithic mechanism of magma emplacement into country rock that has complex folds, boudinage structure, and formations much thickened and thinned by differential plastic flowage. It is expectable that the magma will be accompanied by some granitization, partial fluxing of the lowest-melting constituents of the country rock, and hybridization. The shredded ends of some layers may be due to being squeezed or pulled apart during plastic flowage as well as to irregular replacement. It thus seems highly probable that metasedimentary rocks as layers or skeletal folds in granite or granite gneiss may be either of xenolithic origin in magmatic granite or of skialithic origin in metasomatic granite gneiss.

It has been argued that some batholithic emplacement by granitization has been accompanied by inflation with outward deformation of the walls as the result of introduction of new material. Barth (1947, p. 181) infers that the Birkeland, Norway, batholith with conformable walls is a "petroblast" developed by the introduction of new material as a "cloud of ichor or migrating ions". Oertel (1955, p. 45) describes the Loch Doon stock in Scotland as having expanded its volume by 34 per cent and (p. 81) "Der Pluton ist durch Metamorphose unter Stoffzufuhr aus pneumatolytischen Lösungen und durch metasomatische stoffwanderung entstanden". The Loch Doon pluton has discordant contacts for much of its border. To the writer's knowledge the concept of expansion of the walls as a result of granitization has not yet been advocated for any North American pluton. This is perhaps because hundreds of examples of mineral deposits have been studied in which there has been introduction of new material, but little or no evidence for inflation in consequence of it has been reported.

Emplacement of Plutons in Epizone and Mesozone

Lava flows, acknowledged by all to be of magmatic derivation, may be considered an observable large (as contrasted with the size of an ion) base of known origin from which to extrapolate to a corresponding magmatic origin for most plutons of the epizone.

The common occurrence of homophanous structure, the common local development of miarolitic or aphanitic texture, and the inferred genetic relationship to lavas of similar composition associated in time and space all indicate that the Tertiary plutons were emplaced as magmas, largely or wholly fluid. "Far-travelled" xenoliths from depth and absence of flow structure in the enclosing rock of some plutons necessitate a fluid mode of transfer from depth. The magma of Tertiary plutons was, in the initial stages, fluid enough to yield lava flows. At later stages, after rise to higher cooler levels in the accumulated lava pile, loss of volatiles at lower pressures and partial crystallization would occur, and the magma would become viscous enough to "set" before reaching the surface *en masse*.

Many geologists have emphasized the development of a schistosity that is steeply dipping, often with subvertical lineation, in both peripheral country rock and in border facies of plutons, such as most of those of the mesozone and many of the transitional epizone-mesozone. Because of this they have inferred that the invading material had to be a highly viscous or a diapirlike mass, partly or largely crystallized, rising upward and dragging its walls. This is probably true for much of the quartz diorite facies that forms the outer part of many plutons. Such quartz diorite is reasonably interpreted as the product of incorporation of country rock by a more specifically granitic magma and might therefore be expected to be partly crystalline. Again, successive central intrusion of magma could produce inflation, deformation, and upward drag of partly to largely consolidated earlier facies. In some plutons of the mesozone, upward movement in the border zones persisted through the very last stages of consolidation and even into the solid state.

Such evidence for viscous magma in the border zones, however, does not preclude the possibility that even initial stages of magma emplacement in the mesozone may locally have been freely fluid, and it has no necessary bearing on the later intrusion of the cores.

Even conformable schistose structure in the contact zones of mesozonal plutons need not always indicate development in consequence of emplacement of viscous magma. Durrell (1940) made a systematic study of contact metamorphism in the southern Sierra Nevada and emphasizes that, in contact zones with granite, metasedimentary phyllites and schists may merely coarsen in grain but retain the original foliated structure, a mimetic inheritance. This means that if the material of a pluton of the mesozone were a fluid magma in the early stages we should not necessarily expect a hornfels to be formed from metasedimentary phyllites, and if formed it might [illegible] appear subsequently by stoping. It may [illegible] be noted that hornfels, as distinguished fr[illegible] schist, does occur in the wall rock of s[illegible] mesozonal plutons. In part, however, e[illegible] hornfels may itself be deformed at later st[illegible] of magma emplacement.

Much of the gneissoid granite does not sh[illegible] the amount of crushing and protoclastic stru[illegible]ture that would be expected if flowage occur[illegible] at a late stage of consolidation. Freely s[illegible]pended crystals in an early stage of cryst[illegible]zation may be oriented by flowage, and th[illegible] orientation may be preserved and inherited b[illegible] subsequent overgrowth and control of cryst[illegible]lization by the early fabric.

It seems probable that much, if not most, [illegible] the magma that yielded the rocks of [illegible] mesozonal plutons was predominantly liq[illegible] at the time of its emplacement.

There is good evidence that lavas a[illegible] locally some hypabyssal plutons crystalli[illegible] with some minerals characteristic of hig[illegible] temperatures than those of their pluto[illegible] equivalents (Tuttle and Keith, 1954; M[illegible] and Smith, 1956; Buddington *et al.*, 195[illegible] p. 519–522). Most of the plutons, howev[illegible] especially the larger ones, will have crystalli[illegible] in the presence of at least part of their volati[illegible] will therefore have remained partly fluid fo[illegible] long time down to temperatures lower th[illegible] those of lavas, and any initial high-temperat[illegible] minerals will have undergone recrystallizati[illegible] at lower temperatures. The plutons in th[illegible] epizone, and their inferred relationships [illegible] those of greater depth, necessitate that [illegible] fluid magma that formed them could h[illegible] risen from source to surface in a relativ[illegible] short time—a small fraction of the length [illegible] time assigned to a geologic period. Otherw[illegible] the magma would have frozen en route. [illegible]

A hypothesis is that autochthonous [illegible] parautochthonous batholiths are metaso[illegible] in origin, wholly or predominantly solid, [illegible] arose in and ascended from zones now expo[illegible] in the more deeply eroded areas. This rais[illegible] problem if we also assume, as is here done, th[illegible] most of the plutons emplaced in the ep[illegible] and in part in the mesozone were predo[illegible]nantly liquid magmatic. The foregoing hyp[illegible]theses taken together would necessitate that [illegible] plutonic mass became more liquid as it [illegible] into cooler levels and that it left a concentra[illegible] residuum of resistant type of rocks in th[illegible] source area, the catazone. Rise into low[illegible] pressure zones, chemical reactions, and incre[illegible] in volatiles in the upper part of a mag[illegible]

[illegible]ann (Kennedy, 1955) would tend to increase [illegible]ity, but it remains to be shown that these [illegible]ors are adequate to meet the requirements. [illegible] development and rise of granitic magma [illegible] of the roots of eugeosynclincal materials [illegible]d presumably leave behind a series of rocks [illegible]posed predominantly of garnet amphibo[illegible] and magnesian pyroxene-calcic plagioclase [illegible]ulite but with some other refractory [illegible]erials. This complex would be appropriately [illegible]sistent with the seismic data for rocks above [illegible] M discontinuity in the lower part of the [illegible]st not now exposed, but there is no satis[illegible]tory evidence that there is a concentration [illegible] "resistates" in the now-exposed portions of [illegible] catazone. If erosion has only locally exposed [illegible]s deeper than about 12 miles in the earth's [illegible]st, as seems probable, then we conclude that [illegible]t magma usually came from levels deeper [illegible]n most now at the surface.

The apparent dominance of lineation, insofar [illegible] flowage structures do occur, in plutons of [illegible] epizone and the dominance of planar [illegible]ation, with or without lineation, in the [illegible]tons of the mesozone correlate with the [illegible]erences in mechanics of emplacement and [illegible]erent conditions of consolidation and flow [illegible] the two zones. Lineation alone appears to [illegible] related to an early fluid stage followed by [illegible]tinued crystallization in quiet. In the [illegible]sozone continued movement during a [illegible]tracted intermediate period of crystallization [illegible]n permits the development of planar struc[illegible]e in some parts of the pluton.

There appears to be good evidence that some [illegible] the plutons of the mesozone had extensive [illegible]s and could not have been continuous with [illegible]tons of the epizone unless through relatively [illegible]ll connecting channels. The plutons of the [illegible]sozone and epizone locally overlap one or [illegible]re of their borders as though overriding them, [illegible]l in general the evidence for an extensive [illegible]or is lacking.

The present data do not preclude the possi[illegible]ty, however, that the plutons of the epizone, [illegible] least in part, enlarge downward and are [illegible]tinuous with plutons of the mesozone. If [illegible] emplacement of the latter were in sub[illegible]ntial part effected by crowding aside of the [illegible]lls then the possibility would exist for em[illegible]cement of magma in the epizone through [illegible]ndering of crustal blocks into magma below. [illegible] domical-planar or arch-linear structure such [illegible] occurs in some plutons of the mesozone [illegible]uld form at a late stage in consolidation and [illegible]uld not necessarily indicate a roof of original [illegible]ntry rock at time of consolidation.

The mechanics of emplacement of plutons in the epizone as a consequence of alternating magma rise with accompanying pressure effects and magma relaxation with accompanying appropriate structural collapse of roof and walls have been discussed by Anderson (1936) and by Billings (1945, p. 53–55; 1947, p. 279–288).

In the mesozone, yielding of the walls by plastic flowage may be expected to be greater with depth. This will commonly result in outward-flaring walls, distention of the roof, and the potentiality for collapse of local portions of the roof over the border zones of the underlying magma. This may in part explain the occurrence of both discordant and concordant boundaries so characteristic of plutons of the mesozone.

It is recognized that replacement may occur locally with sharp and even discordant contacts. An origin by mechanical rather than chemical processes, however, seems the best interpretation for most contacts of plutons in the epizone, and the similarity of character and probable history indicates a similar mechanical origin for most sharp discordant contacts of plutons in the deeper zones.

The writer finds no evidence for postulating a discontinuity between the compositions, textures, internal structures, or mechanics of emplacement of plutons in the epizone and those of the mesozone. There are plutons with intermediate characteristics that suggest a series of gradational changes from one to the other.

The concept of the development of migma-magma with schlierenlike structure at the source seems reasonable, but its rise and emplacement "as such" does not fit the homophanous core of the plutons of the mesozone and those of the epizone.

The North American literature of the past 25 years appears to be based on the assumption that the best theory, as of the present, for emplacement of plutons in the epizone is one of block foundering in, or some kind of stoping by, magma as a major factor. After half a century, however, the stoping hypothesis still lacks verification in the sense that we do not have desirable supporting evidence of sunken blocks in the plutons of deeper levels. Unless such blocks have been indistinguishably incorporated in magma, or reworked and metasomatized into new granites at depth, or have sunk to very great depths, we might expect to find more evidence of floored plutons than has been reported.

Three major factors of emplacement of plutons in the mesozone are inferred to be stoping or block foundering, crowding aside of the walls, and uplift of the roof. The sideward expansion may in part result in plastic flowage of country rock upward and downward. Additional factors are incorporation of country rock and metasomatism of wall rock. The importance of each factor varies with the particular example. If the tonalities of the batholith of Southern California are inferred to result from incorporation of gabbro by granodiorite magma then gabbroic material may be inferred to be equivalent to about half of the tonalite that forms 63 per cent of the area now occupied by the batholith. This would make the problem of emplacement of the batholith one of how so large a volume of the initial gabbro was introduced as well as of how so great a mass of granodiorite was emplaced. Similarly at least 10 per cent of the problem of emplacement of the Coast Range batholith of Alaska and British Columbia could be related to incorporation of mafic rocks. Most of the mafic rocks in orogens, however, are lava flows and intrusive diabase, dolerite, or gabbro sheets emplaced under conditions of the epizone previous to major deformations.

A great variety of structures associated with plutons of the mesozone have been cited as consistent with formation by upward movement of magma. Such structures within the plutons are subvertical foliation and lineation in border facies, marginal fissures with pegmatite or aplite, marginal upthrusts, radial dikes, and schlieren domes; structures in the country rock near the pluton may be increase in dip of foliation or of slaty cleavage or of the plunge of lineation, subvertical lineation and subvertical secondary fold axes of plastic flowage origin, slip cleavage roughly similar in strike to the periphery but dipping outward with the lower schist layers moving up relative to the overlying layers, and development of domical foliation in the roof.

Structures consistent with outward expansion effected by the pluton are deformation of beds into partial conformity with the periphery, intensification of folding, and in part the intensification of planar foliation in the border facies of the pluton itself.

All discordant structures need not necessarily mean stoping, for expansion effects of the pluton may result in pulling portions of the this possibility is not of sufficient magnitude t satisfy the actual relationships.

Emplacement of Plutons in Catazone

If the writer has correctly interpreted the literature, then most, perhaps all, of the plutons of the catazone were emplaced under synkinematic conditions. The roof portions of a number of batholiths of the mesozone are similar to those of plutons emplaced in the catazone, but such conformable emplacement in the mesozone does not, in part at least, appear to be related necessarily to regional tectonic forces. Such plutons may be intermediate between those typical of the mesozones and those characteristic of the catazone. The extent to which plutons of the catazone are due to metasomatism or anatexis is yet to be definitely determined. It is logical to assume that at their source granitic masses would as a result of anatexis and rise of temperature become mobile and rise as diapirs or as "migma-magma", but the extent to which source areas are now exposed is most problematical.

Marshall and Narain (1954, p. 73) have postulated that the negative gravity anomalies over granite batholiths, of a type here inferred to belong to the mesozone, are due to "granite roots", to extension of the granite pluton to depth, rather than due only to density contrasts between granite and country rock near the present surface. Presumably this could mean continuity of plutons of the mesozone with those of the catazone. Biehler and Bonini (1958) have concluded that, if reasonable assumptions are made for the probable geology and density distribution of the region of the Boulder batholith, it follows that a granite mass roughly with a plano-concave cross section and a depth not much less than and not much more than 10 miles will closely satisfy the residual negative bouguer anomaly. An additional narrow root could also be present.

Grout (1945, p. 276–278) on the basis of some experimental evidence suggests that large intrusives that rise from great depths may have only roots or a series of roots and that they may ascend along part of their route because some of the overlying rocks become so reduced in viscosity that they move aside and down along the sides in a mobile contact zone. There may be accompanying distention and sideward flow in the roof during emplacement. Such a mechanism of intrusion would the areal cross section of batholiths in the deeper part of the mesozone with the areal cross section of their feeders from the catazone. The shape of the intrusive mass would be streamlined and comparable to the shape of salt diapirs with pinched-off roots. The lower part of such bodies should have inward dips. Symmetric funnel-shaped plutons such as the Loon Lake (Fig. 14) are few, but a number of deep-seated plutons such as the Cheddar batholith (Fig. 14) and the southwest side of a part of the Coast Range batholith are asymmetric in cross section and bordered on one side by inward-dipping schists. The problem of batholithic roots in the transition zone between the mesozone and catazone and in the catazone deserve more detailed study.

The details of the interrelations of plutons of the catazone to those of the mesozone and of those of the mesozone to those of the epizone remain as problems. A tentative schematic diagram of relationships is shown in Figure 19

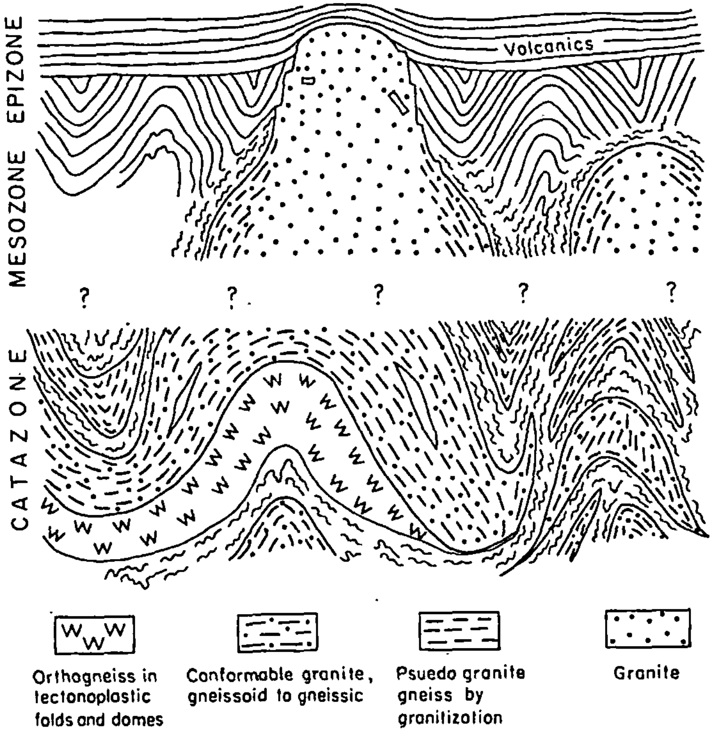

FIGURE 19.—SCHEMATIC SKETCH SHOWING POSSIBLE STRUCTURAL RELATIONSHIPS OF PLUTONS IN EPIZONE, MESOZONE, AND CATAZONE

Question is left open as to whether batholiths of mesozone enlarge downward in continuity with those of catazone or whether they have roots

seems to be a major kind in the catazone, and often phacoliths of both igneous and of metasomatic origin are reported to be associated. Many xenolithic domal batholiths may also, in part at least, result from a phacolithic mechanism of emplacement, usually igneous but in part metasomatic.

Age determinations permit the inference that the youngest plutons of the catazone with extensively developed migmatities are about 100 m.y. old and that plutons of the epizone may be at least as old as 1.65 b.y. and probably much older. Plutons of the mesozone range in age from slightly less than 100 m.y. to those of the Keewatin belt of the Canadian shield which are 2.5 b.y. or older.

Origin of Granitic Magma

"We must still entertain the hypothesis that most granites have been produced throughout geologic time by differentiation of basic (basaltic) magma, in part somewhat modified in its course from

The role of metasedimentary material as the prime source of most granitic magma is now emphasized far more than is implied in the statement of Bowen cited above. There appears to be a tendency at the present time to start with the assumption of at least two major magmas—one, the granitic derived from the sialic part of the crust, the other, the basaltic from deeper down, probably in the mantle, or beneath the continents from an eclogite root.

Graywacke forms a large part of eugeosynclinal sediments. It would with increasing melting yield successively a little true granite and then a trondhjemitic magma. The latter by reaction with associated basalts would result in a tonalitic magma. Partial melting of an illitic type of clay has been shown by Winkler (1957, p. 57–58) to yield an exceptionally potassium-rich leucogranitic magma. The hypothesis that the lowest part of the sial contains some primordial granitic material differentiated from basaltic magma however is reasonable and has not as yet been precluded, especially for the early stages of geologic history. Remelting of such primordial granite would, of course, yield granite magma directly. An andesitic magma may form by partial melting of a gabbroic or eclogitic continental substratum or essentially by incorporation of graywacke in basaltic magma.

Basaltic magma may yield basalt or gabbro directly by consolidation; dioritic-andesitic magma by assimilation of sialic material or by mixing of femic and salic magma; and ultramafic, anorthositic, monzonitic, granophyric, and other subordinate facies by differentiation. The granitic magma may yield granites directly; minor diorite by incorporation or metasomatism of mafic rock; and mobile quartz dioritic or other intermediate kinds of magma by incorporation of graywacke and earlier basalt flows or gabbro plutons. The quartz dioritic magmas may in turn yield granodiorite, quartz monzonite, and new granite magma by differentiation. The initial volume of gabbro emplaced in the mesozone and epizone must have been much larger than that now represented by exposure. Much of it foundered in later intrusive granitic magma and was incorporated to reappear in the modified facies such as quartz diorite.

Read (1951, p. 22) has made a tentative suggestion that "we seek the ultimate source of the granitising fluids in crystallising simatic material below the site of the geosyncline." Adoption of such a hypothesis has a number of significant consequences. Simatic material in crystallizing at depth slowly over a long time may be expected to undergo fractional crystallization and differentiation to yield directly magmatic monzonitic to dioritic differentiates (especially if magma is undersaturated) or magmatic granitic differentiates (if oversaturated). The released granitizing fluids, either magma, gas ions, or all three may be expected to modify the material in the lowest part of the geosyncline and add to or develop granitic magma. The effect of pressure in raising the temperature of melting is not adequate to prevent a rise of temperature and pneumatolytic fluids from fluxing granitic material at the base of the sialic part of the crust rather than higher up. The source zone for the plutons would therefore not be expected to be exposed.

Tuttle (1955) has estimated that with a geothermal gradient of 30°C. per km partial melting of a geosynclinal prism of sediments might start to yield a biotitic granitic magma with a temperature of about 640°C. at a depth of 21 km and that complete melting with about 2 per cent H_2O could occur at about 31 km. With a gradient of 40°C. per km incipient melting could occur at a depth as shallow as 15 km.

The usual order of intrusion—gabbro, quartz diorite, granodiorite, quartz monzonite, and granite—in composite plutons is one that corresponds to the order theoretically expectable as a result of magmatic differentiation or alternatively to that of decreasing temperatures. A speculative hypothesis to explain this order might be as follows. The early rise of basaltic magma directly yields gabbro plutons, diabase sheets, and basaltic lavas; basaltic magma with incorportation of sial leads to andesitic lavas and minor diorite plutons. Subsidiary effects of the development and rise of basaltic magma and its derivatives are accentuation of the rise of the isogeotherms and fluxing in the lowest part of the sial. As the isogeotherms rise through the deep sial early formed interstitial low-melting granitic fluids work upward, react with country rock in part, and result in a differentiated domal column ranging upward in composition from residual refractory materials at the base through quartz dioritic and granodioritic facies to granite at the top. Eventually the continued rise of the isogeotherms results in sufficient melting of the lower portion of the column—either the quartz dioritic or granodioritic facies—so that it rises as a whole followed successively by melting and rise of the overlying materials. Granodioritic magma may in turn react with mafic material on its upward flow to yield quartz diorite. Later granitic magma may rise through a sheath of the earlier intrusives. Other hypotheses are desirable.

The variation in ratio of different kinds of igneous rock in the different zones needs study.

The plutons of the epizone may be predominantly granodiorite, quartz monzonite, and granite. The tonalitic facies in general appear to form a larger percentage of the rocks in the plutons of the mesozone than in those of the epizone. Quartz monzonite, granite, and leucogranite or alaskite are far more common among the members of the granite family in certain belts of the catazone than in the mesozone (Daly, 1914, p. 60; Osborne, 1956). In particular, andesine-quartz diorite apparently is relatively subordinate in these belts. Andesine and labradorite anorthosite and gabbroic anorthosite of the types found in massifs and independent sheets are possibly almost exclusively in belts of the catazone. Assuming the foregoing relationships are correct, although we do need quantitative data to substantiate them, some suggestions as to their origin may be offered as bases for study. Is the restriction of the types of anorthosite mentioned to the catazone the result of the necessity for the kind of environment which by acting as a plastic envelope (the country rock is often marble) under high pressure permits the retention of the high volatile content essential to keep the equivalent magma fluid and of gabbroic anorthositic or anorthositic composition, or is it a phenomenon to be correlated with greater age, or both? A favorable home for quartz diorite is the mesozone. Is this because the quartz diorite magma originates largely through reaction of more alkalic granitic magmas with mafic rocks, thus losing fluidity and to a substantial extent not rising above the mesozone? Is the predominance of granodiorite and granite relative to quartz diorite in the epizone the consequences of equivalent magmas being the lightest types and fluid at relatively low magmatic temperatures because of volatile content and their lower melting intervals? Why the predominance of granite in certain catazonal belts? One answer might be, "They are granitization products". But the writer is not convinced that this is the whole answer because magmatic mesoperthite granite bulks large in the Adirondack belt of catazonal rocks.

Problem of Volcanic and Plutonic Associations

One of the most critical problems is whether or not our present knowledge indicates, or is consistent with, the hypothesis that there is a direct relationship between salic volcanic rocks, aphanitic or porphyritic intrusions, and granitoid plutons.

Kennedy (Kennedy and Anderson, 1938) has made a sharp distinction between *volcanic* and *plutonic* associations with respect to igneous rocks. He postulates that the *volcanic associations* include not only lava flows and directly related vent intrusions but also such intrusions as the great sill swarms of the Karroo, South Africa, the Palisades sill of New Jersey, and even the great sheets such as the Bushveld igneous complex of the Central Transvaal, South Africa, all of which are in nonorogenic areas and intimately associated with volcanic phenomena. *Plutonic* associations on the other hand he suggests appear to be limited to orogenic regions and consist almost entirely of granodiorite and granite together with smaller amounts of their associated predominantly hornblendic, basic, ultrabasic, and lamprophyric types, while typical gabbros are characteristically rare or absent. He further states that volcanic associations, on the contrary, are overwhelmingly basic and are composed mainly of basaltic magma or of rock types belonging to a basaltic line of descent. He concludes (1938, p. 30) that

> "Altogether, there does not appear to be any very direct evidence to indicate a close connection between plutonic activity and volcanicity"

and (1948, p. 2–3) that

> "there is a universal absence of lavas belonging to a period contemporaneous with the rise of batholithic intrusions to their highest levels in the crust."

A recent discussion of the problem has been given by Raguin (1957, p. 185–199). He refers to subvolcanic granite masses as truly very special and exceptional and moreover ambiguous in interpretation. He attributes to E. Suess the development of the concepts that successively deeper denudated levels reveal lava flows, hypabyssal porphyry intrusions, and grained plutonic bodies of similar composition but different texture and that all are genetically related. He also presents a discussion of an opposite philosophy that envisages the foregoing concepts as seductive but deceiving and that interprets the association of volcanic rocks and major plutons in the earth's crust to have no direct relationship. Some proposed objections to the Suess concepts are that the succession in plutonic rocks is from basic to acid but in volcanic rocks from acid to basic, irregular or recurrent; volcanic rocks may occur independent of plutonic rocks, and *vice versa*; volcanism is related to periods of fracturing of the crust, plutonism to periods of folding; plutonism occurs after vulcanism; a true volcano has never been observed to rise from a plutonic mass; and identities such as volcanic and plutonic quartz porphyries are only a phenomenon of convergence. Raguin concludes that the problem of the relationship of vulcanism to plutonism is still unsolved.

The ideas of Kennedy have been criticized by Tyrrell (1955, p. 420) who writes

> "The lava series ranging from pyroxene andesite to rhyolite, therefore, has the same chemical composition, the same geological and geographical distribution, the same tectonic environment, and appears at the same stage of the tectono-igneous cycle as the plutonic series ranging from quartz diorite to granite. If the latter belongs to the plutonic association there seems to be no escape from the conclusion that the andesite-dacite rhyolite series likewise does . . . . The writer suggests that the whole trouble with plutonic and volcanic associations is that they are wrongly named."

The many citations (21) in this review demonstrate that it is normal for volcanic rocks of equivalent composition to be associated in space, time, and tectonics with plutons emplaced in the epizone. Many granitic bodies that have been called plutonic because of their medium to coarse grain and their batholithic size are not plutons in the sense of deep-seated emplacement. Exceptionally cogent examples that provide a clear demonstration of the succession of magmatic activity from volcanism to high-level granite emplacement have been described from Northern Nigeria by Jacobson, MacLeod, and Black (1958, p. 7). Here are about 40 granite complexes in a belt about 270 miles long and up to 100 miles wide. The total area of granite is about 2000 square miles, and there is about 500 additional square miles of rhyolite of similar chemical composition and age. The rhyolite furthermore is almost wholly confined within the granite ring complexes. The granite plutons are up to 285 square miles in area. If there about 3 times as many complexes, each about 3 times the area, in the mesozone neath this belt, it would mean a contin great batholith, 270 miles long with an ave width of 65 miles.

Some of the very small plutons of the epi may be granitic, granophyric, or monzo differentiates of basaltic magma bodies loca somewhat below. But it seems essential th most of the salic lava flows and stocks and the batholiths of the epizone originated fr granitic magma of deep-seated origin.

A diversity of succession in lava sequen can reasonably be interpreted in terms of s cessive pulses of basaltic or andesitic mag from depths and magmas of the rhyo rhyodacite, dellenite, dacite group fr epizonal plutons. It is also probable that so of the epizonal dikes of granitic composi came directly from great depth rather th from epizonal plutons. In summary, the ma rocks are predominantly extruded as lavas emplaced in the epizone, whereas the fe rocks, although including lava flows and m related plutons emplaced in the epizone, are volume per cent largely emplaced in mesozone and catazone. If the granite or nated as a differentiate of gabbroic magma the mesozone or catazone, at a level above source of the gabbroic magma itself, then bod of gabbro at least 10 times as large as granite plutons should be common in mesozone or catazone of the orogens. S bodies do not occur in appropriate volume these zones as now exposed.

CONCLUSION

The foregoing survey shows that the auth who have reported on detailed struct studies of stocks and batholiths in Nor America have reasoned that granitization o occasionally has played a role in the mecha of emplacement of plutons in the epizone, been of subordinate significance (one au excepted) in the mesozone, but has played major although not necessarily a greatly pr ponderant part in the catazone. Otherwise reliance has been on magma emplacem One might also conclude from such a surv however, that the unity of agreement in interp tation has been far too great to be healthy the optimum advance of our understand Hypotheses, in general, as to the respect roles of magma and of metasomatism in the problem remains one demanding cal studies and more dependable criteria.

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REGIONAL GEOPHYSICS OF THE BASIN AND RANGE PROVINCE

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> *Of late years the most important contributions have come from the Physicists, and in their scales have been weighed the old theories of Geologists.*
>
> G. K. Gilbert (1874)

INTRODUCTION AND GEOLOGIC SETTING

Nearly one hundred years ago, Gilbert (23, 24) and other geologic pioneers introduced the idea that much of the seeming jumble of mountains and valleys in western North America was the result of far different processes than fold mountain systems such as the Appalachians or Alps. After a century of geologic and geophysical investigations in the region, it is now generally accepted that the physiography of the Basin and Range province (Figure 1) is one of sculptured and partially buried fault-bounded blocks that have been produced by the extension of the region during late Cenozoic time. Crustal blocks composed of complexly deformed, diverse pre-Cenozoic rocks and relatively undeformed, predominantly nonmarine volcanic rocks of early and middle Cenozoic age have been variously uplifted, tilted, and dropped along numerous normal faults throughout a broad region from Mexico to Canada—from as far west as California and Oregon to as far east as western Texas (e.g. Cook, 13; Gilluly, 27; Thompson, 76).

The distribution of late Cenozoic normal faults in the western United States is shown on Figure 2 (note that the regional extent of faulting is somewhat larger than the Basin and Range physiographic province of Figure 1). The recent seismicity (Figure 3) shows that small earthquakes are widespread in what Atwater (5) called a wide soft zone accommodating oblique divergence between the Pacific and North American plates. The net effect of fault movements within this region is a crustal extension oriented roughly WNW–ESE. The actual motion on individual faults is quite variable, however, and appears to be controlled by the orientation of faults with respect to this principal extension (Thompson & Burke, 78). In the northern portion of the region—across Nevada and western

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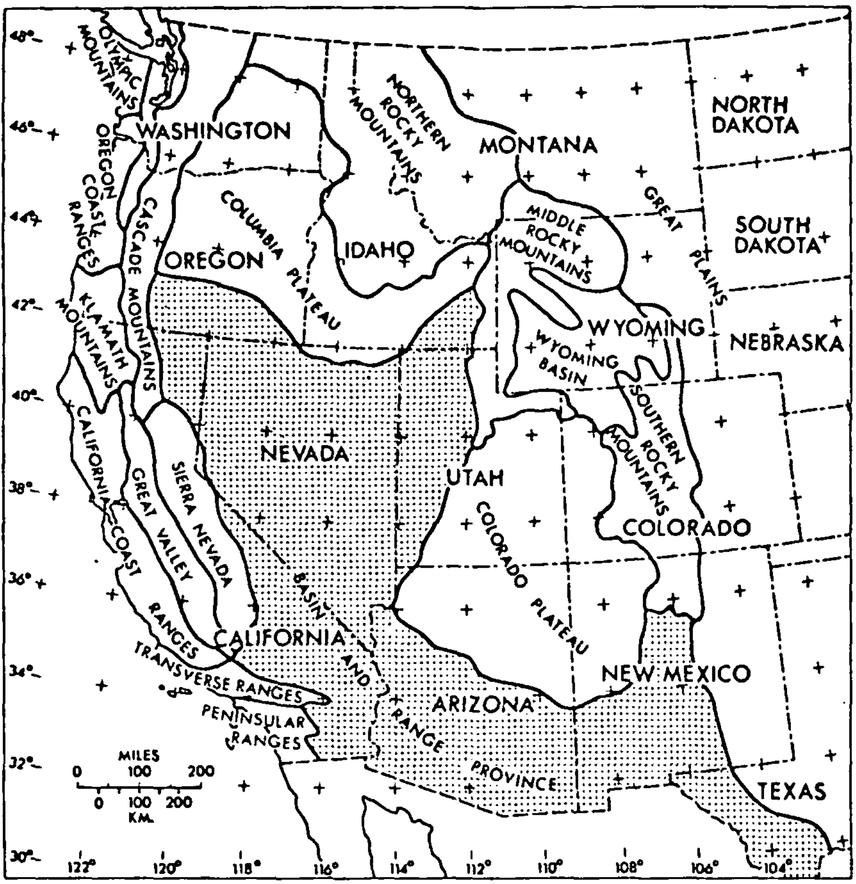


Figure 1 Physiographic provinces of the western United States (Fenneman, 21).

Utah—the domain of faulting is neatly confined between the Sierra Nevada of California and the Wasatch Mountains of north-central Utah. The relatively unfaulted Colorado Plateau separates the central portion from a zone of faulting in the Rio Grande trough in New Mexico and west Texas. Relative motion between the unextended and rather enigmatic mass of the plateau and the encircling faulted terrain is presumably accommodated by a component of right-lateral strike-slip along the southern plateau border. Faulted terrain extends southwards without interruption into Mexico and the Gulf of California. Faulting seems to die out to the north, and the manner in which relative motions are accommodated along the northern boundary remains a troublesome problem.

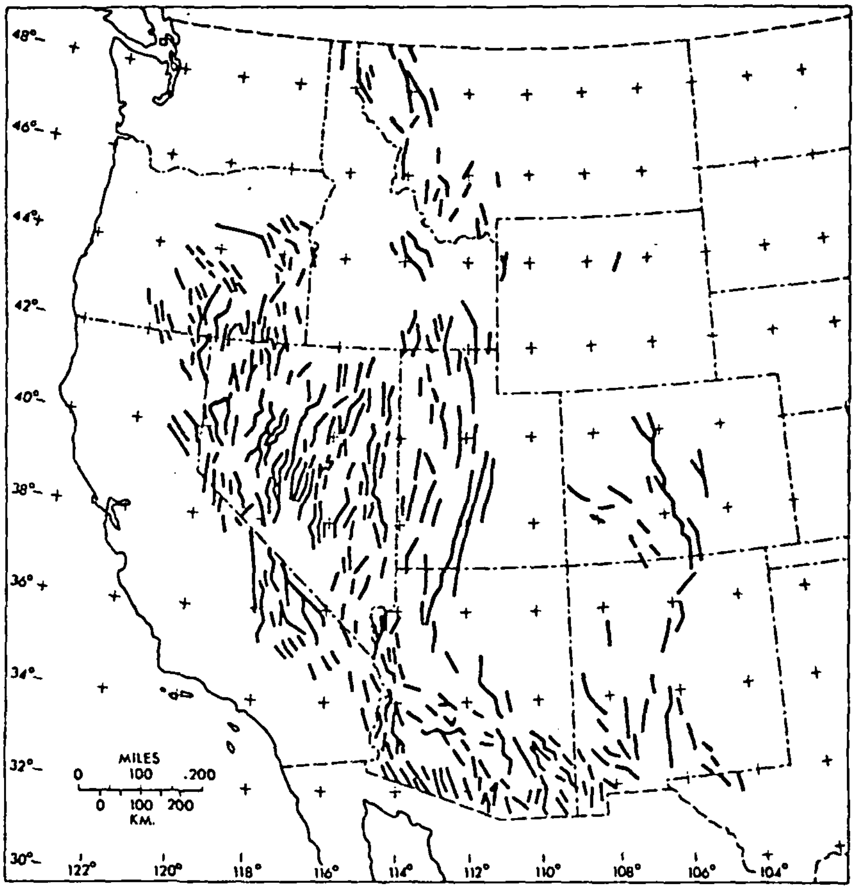


Figure 2 Predominantly normal (Basin and Range) faults of late Cenozoic age in the western United States (modified from Gilluly, 26).

Although the Basin and Range province is in many ways a unique physiographic and geologic entity, increasingly precise and reliable geophysical studies, together with advances in tectonic theory, highlight similarities between the province and other regions of past or present crustal extension. It has a high heat flow and widespread volcanism like other regions of active normal faulting, such as the Rift Valleys of Africa, the Lake Baikal depression of the USSR, the Rhine graben of Europe, the marginal basins of the western Pacific Ocean, and the worldwide system of oceanic ridges and rises. Along with the Sierra Nevada and Colorado Plateau, it forms a wide elevated region averaging 1-2 km above sea level and thus *may resemble the elevated, thermally expanded oceanic ridges (Sclater &*

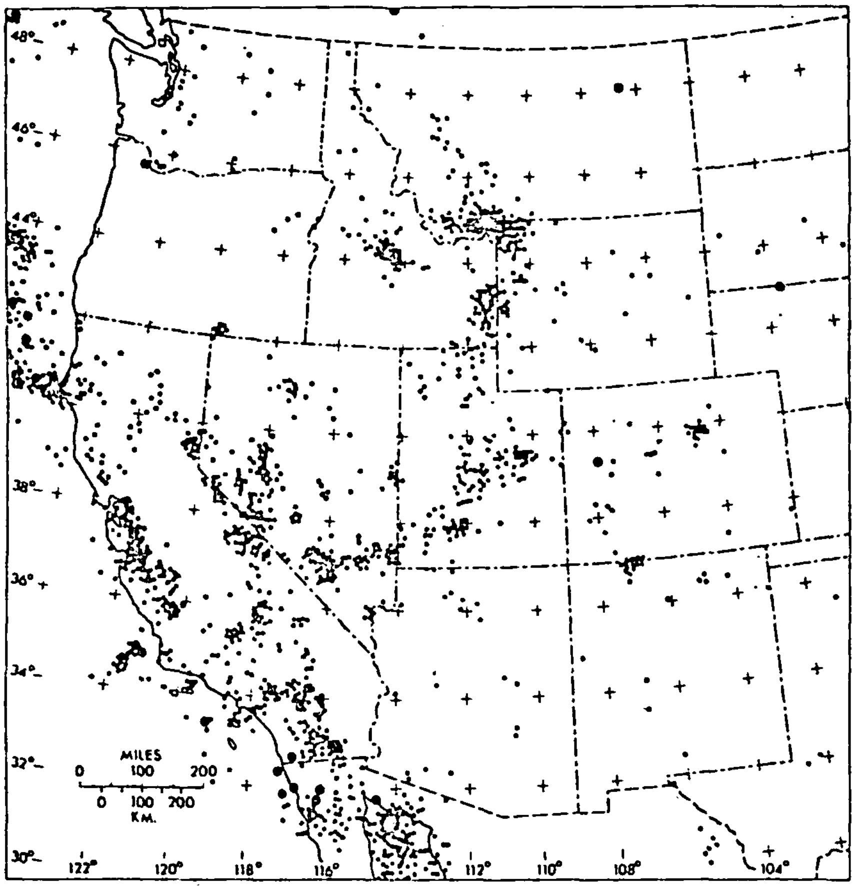


Figure 3 Earthquake epicenters in western North America for the period 1961–1970. Small dots represent earthquakes of magnitude about 3 to 5, large dots greater than 5. National Oceanographic and Atmospheric Administration epicenters replotted by J. C. Lahr and P. R. Stevenson of the US Geological Survey (personal communication, 1973).

Francheteau, 65). Also like some of these other regions, it has a thin crust and low mantle velocity.

Can regional geophysical data for the Basin and Range province, combined with interpretations of its geologic history, lead toward a better understanding of the tectonic processes that have controlled its development? To what extent have earlier geologic events in the region preordained the pattern of faulting that we now see in western North America? What constraints must be heeded in tectonic models of the region, and what aspects of the province allow these models to be compared with other portions of the global system of ever-changing lithosphere plates? We believe this last consideration to be of great importance, although it can only be touched on lightly here, because much understanding of the province derives from analogy with other regions of crustal extension. The currently most promising models relate Basin and Range structure to an earlier subducting plate at the western margin of North America, and they incorporate close physical comparisons with the marginal basins of the western Pacific.

REGIONAL CRUST AND MANTLE STRUCTURE

Crustal Thickness: Seismic Refraction

Seismic waves from explosions have provided the most reliable and detailed information on crustal thickness and indicate that the region of distinctive Basin and Range structures corresponds quite closely with a region of thin continental crust (Pakiser, 52; Prodehl, 55). Prior to the work of Tatel & Tuve (74) it was generally assumed that the crust would be thicker under this elevated region than in continental regions near sea level, a relationship that has been found in other mountain regions. It was thought that lateral variations of velocity and density in the mantle were unimportant, or at least inconvenient in seismic interpretation, and that isostatic compensation was accomplished mainly by variations in crustal thickness.

Tatel and Tuve found that the crust in northwestern Utah is an anomalously thin 29 km. Verification came from Berg et al (6), Diment et al (16), and Press (54), although these authors initially used a different definition of the crust. They found abnormally low *P*-wave velocities of 7.6 to 7.8 km/sec at shallow depth for what we have now come to identify as P_n, the wave traveling in the uppermost mantle below the *M* discontinuity.

Extensive explosion studies carried out by the US Geological Survey established the basic picture as we know it today. David H. Warren, of the USGS (personal communication, 1973) has compiled and interpreted these and other data into a contour map of crustal thickness (Figure 4). The contours are based on data of varying quality and on varying interpretation of velocity structures within the crust; nonetheless they represent a good first approximation. Almost the whole region from the Rocky Mountains westward has a thin but variable crust, roughly two thirds the thickness found in stable regions of comparable elevations. The eastern border of the Basin and Range province is marked by a fairly sharp gradient at the 35 km contour to a thicker crust under the Colorado Plateau. Southeast of the Colorado Plateau there is some indication of thinning beneath the Rio Grande trough of New Mexico and west Texas.

The crust is thicker beneath the Sierra Nevada to the west of the province [although this conclusion has been called into question by Carder (10)]. It is interesting to point out that in detail the thick crust of the Sierran region (Figure 5) extends into the Basin and Range province to the east of the Sierra Nevada. The eastward extent of thick crust does not correspond with the eastern border of the Mesozoic Sierra Nevada batholith (Figure 6), however; although a correlation of the low velocity zone with the border of the batholith is not ruled out.

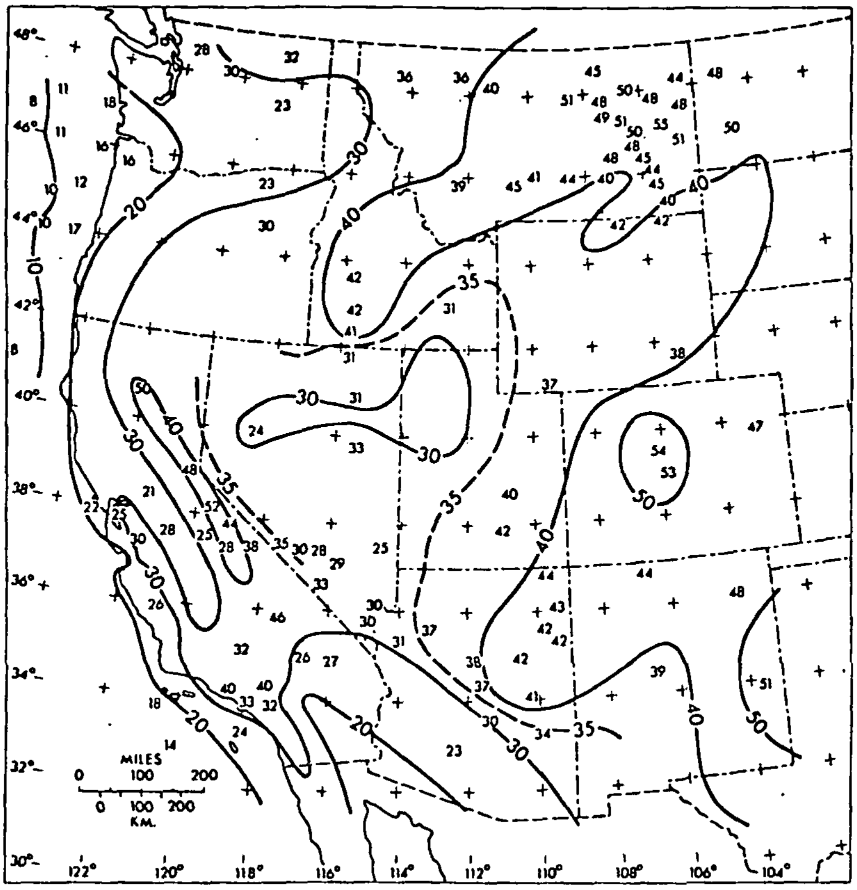


Figure 4 Contour map of crustal thickness (in kilometers) based on seismic refraction studies. Small numbers indicate individual thickness determinations. Compiled by David H. Warren from the following sources: 1, 4, 7, 11, 16, 19, 20, 22, 30, 34–39, 44, 55, 57–59, 62, 66, 67, 70, 72, 73, 82–85.

Upper Mantle Velocity and Implications From Gravity

When it was found that the crust is abnormally thin beneath the Basin and Range province and adjacent regions it was also discovered that P_n is anomalous. Its velocity of 7.7 to 7.9 km/sec is significantly less than the normal velocity of about 8.2 km/sec observed in stable regions (Pakiser, 52; Herrin & Taggart, 33; see Figure 7). Most of the Basin and Range province is characterized by the lowest P_n velocities, less than 7.8 km/sec.

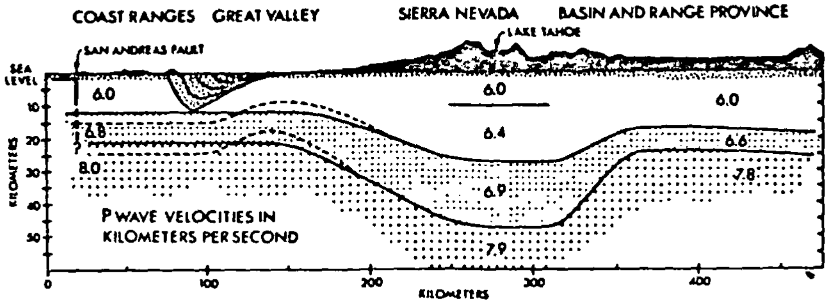


Figure 5 Crust and upper mantle structure in a section across central California and west-central Nevada as deduced from seismic-refraction studies. An alternative model beneath the Coast Ranges and Great Valley is shown by dashed lines; topography greatly vertically exaggerated (from Eaton, 20).

Gravity data supply a fundamental constraint on the amount of mass per unit area underlying any region. This information is particularly valuable because seismic refraction measurements do not by themselves allow interpretations of the thickness of the anomalous upper mantle of low P_n velocity. Gravity interpretation utilizes: 1. crustal thicknesses from seismic refraction, 2. crustal densities estimated from seismic velocities and geology, and 3. upper mantle densities estimated from P_n velocities. The gravity data then yield estimates of the thickness of anomalous mantle relative to stable regions (Thompson & Talwani, 79). The required thickness of low-density, low-velocity anomalous upper mantle is at least 20 km over much of the region.

In comparison with stable continental regions near sea level, most of the isostatic support for the high Basin-Range and adjacent regions is in the anomalous upper mantle. This material must surely be a key element in any tectonic model.

Isostatic gravity anomalies in the United States (Figure 8) show that most of the region from the west coast to the eastern limit of the Basin and Range province is deficient in mass, with an average anomaly of perhaps around -10 mgal. In this respect the region is similar to marginal basins of the western Pacific, which also tend to be isostatically negative.

The Lake Bonneville Experiment

A natural experiment in gravitational unloading of the Basin-Range crust occurred as pluvial Lake Bonneville, of late Pleistocene age, dried up, leaving the Great Salt Lake as its principal remnant. Prominent shorelines around the edge and on former islands mark the successively lower levels of Lake Bonneville. These shorelines are domed up toward the center as much as 64 m as a result of the unloading (Gilbert, 25; Crittenden, 14).

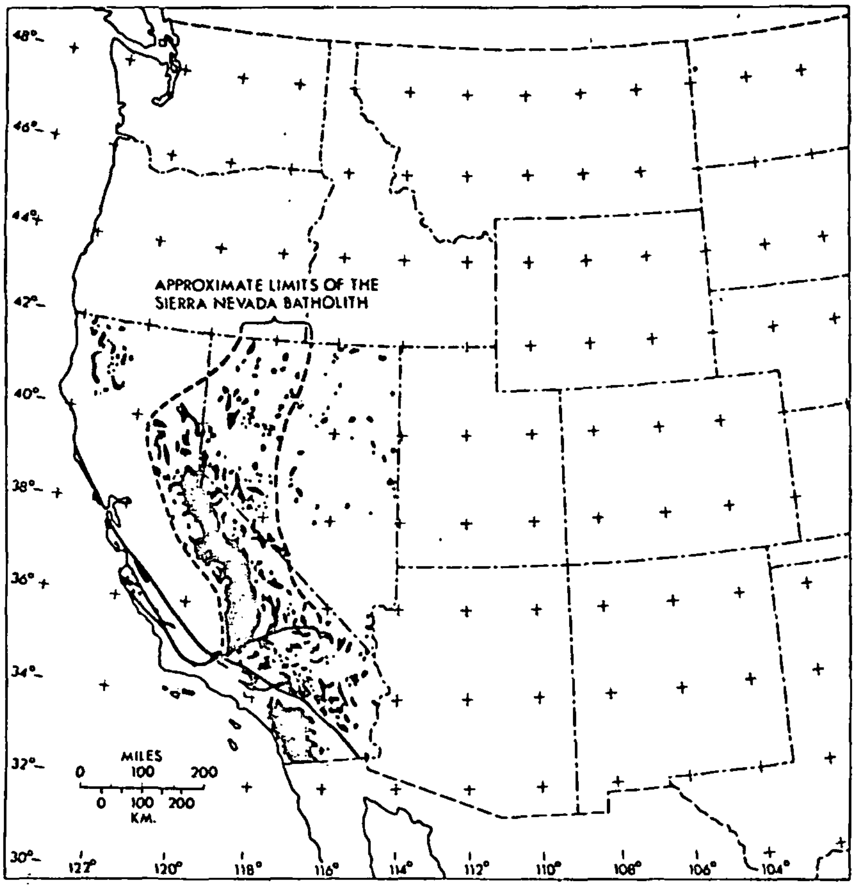


Figure 6 Distribution of granitic rocks in California and Nevada. Solid lines represent major active strike-slip faults (from Crowder et al, 15).

Using data from this natural experiment and a simple model of an elastic lithosphere floating on a fluid asthenosphere, Walcott (81) has computed the apparent flexural rigidity of the lithosphere and compared it with that of other regions subjected to various kinds of loading and unloading. Walcott's results, as shown in Table 1, illustrate that the flexural rigidity of the Basin and Range lithosphere is unusually low. He suggests that the anomaly may be explained by a "very thin lithosphere, only about 20 km thick, with hot, lower crustal material" acting as part of the asthenosphere. In contrast, the flexural rigidity of stable continental and oceanic regions suggests lithosphere thicknesses of 110 km and

Table 1 Apparent flexural rigidity of the lithosphere (from Walcott, 81)

Data	Region	Apparent flexural rigidity, Newton-meters	Characteristic time, years
Lake Bonneville	Basin and Range province	5×10^{22}	10^4
Caribou Mountains	Stable continental platform	3×10^{23}	5×10^6
Interior Plains	Stable continental platform	4×10^{23}	5×10^6
Boothia uplift	Stable continental platform	7×10^{22}	5×10^8
Lake Algonquin	Stable continental platform	6×10^{24}	10^3
Lake Agassiz	Stable continental platform	9×10^{24}	10^3
Hawaiian archipelago	Oceanic lithosphere	2×10^{23}	10^7
Island arcs	Oceanic lithosphere	2×10^{23}	10^7

75 km or more, respectively. The low P_n velocity and high heat flow (discussed in a later section) are consistent with Walcott's interpretation.

Anomalous Mantle and the Low-Velocity Zone

Several studies have indicated that the Basin-Range region has an unusually well-developed upper mantle low-velocity zone (LVZ) for both *P*- and *S*-waves. The relationship is not always clear between the accentuated LVZ (as defined by waves refracted at deeper levels in the mantle) and the anomalous upper mantle (as defined by low P_n velocity). In a study applicable to the central part of the Basin and Range province in Nevada and western Utah, Archambeau and associates (2) derived a model (Figure 9) in which the *M* discontinuity is at a depth of 28 km and the P_n velocity just below it is 7.7 km/sec. This low velocity remains nearly constant to a depth of 130 km, where it undergoes a rapid transition to 8.3 km/sec. Thus the LVZ is about 100 km thick; it begins at the top of the mantle and coincides with the anomalous upper mantle.

In comparison, the same investigators derived three models applicable to regions northeast and east of the Basin and Range province, including the Colorado Plateau (Figure 9). These models have in common a "lid" of higher velocity material (P_n about 8.0 km/sec) above the LVZ, which is only about half as thick as in the Basin-Range model.

In the foregoing discussion a single model has been assumed to represent the Colorado Plateau, and this assumption seems reasonable because of the geological uniformity of the Plateau. However, within the limited resolution of the data, P_n velocities (Figure 7) appear to vary markedly over the Plateau and would not allow a single upper mantle model. This seeming conflict invites further research.

Helmberger (31) developed a new technique for studying regional variations of the LVZ. The method makes use of the nearly constant velocity of the *PL* wave in the crustal wave guide and the regional variation in the velocity of long-period *P*-waves. Results are mapped on Figure 10 (York & Helmberger, 87) as observed time differences minus the time difference predicted from a model LVZ roughly

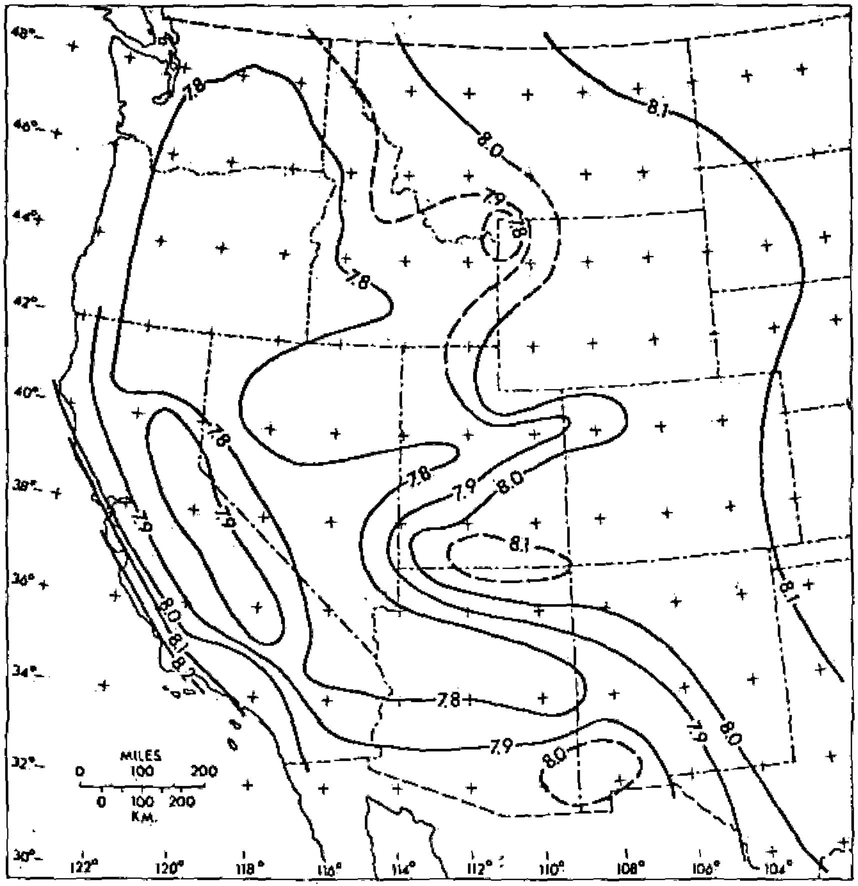


Figure 7 *Contour map of P_n (upper mantle) velocities (in kilometers per second) (from* Herrin, 32).

comparable to the Colorado Plateau model of Figure 9. Progressively more negative Δt values (delays of the long-period P-wave relative to the model) represent progressively thicker LVZ or lower upper mantle velocity. Positive values represent thinner or higher velocity LVZ relative to the model. Two main zones of thick LVZ within the -3 sec contour trend northward through eastern Nevada and western Utah and northeastward into the Rio Grande trough in New Mexico. These zones join to the southwest and continue across southern California and northern Mexico toward the continental borderland off southern California (generally considered to have Basin-Range structure) and the Gulf of California. The Colorado Plateau is strikingly outlined by the zero contour, which is expected because the reference model resembles the Colorado Plateau mantle.

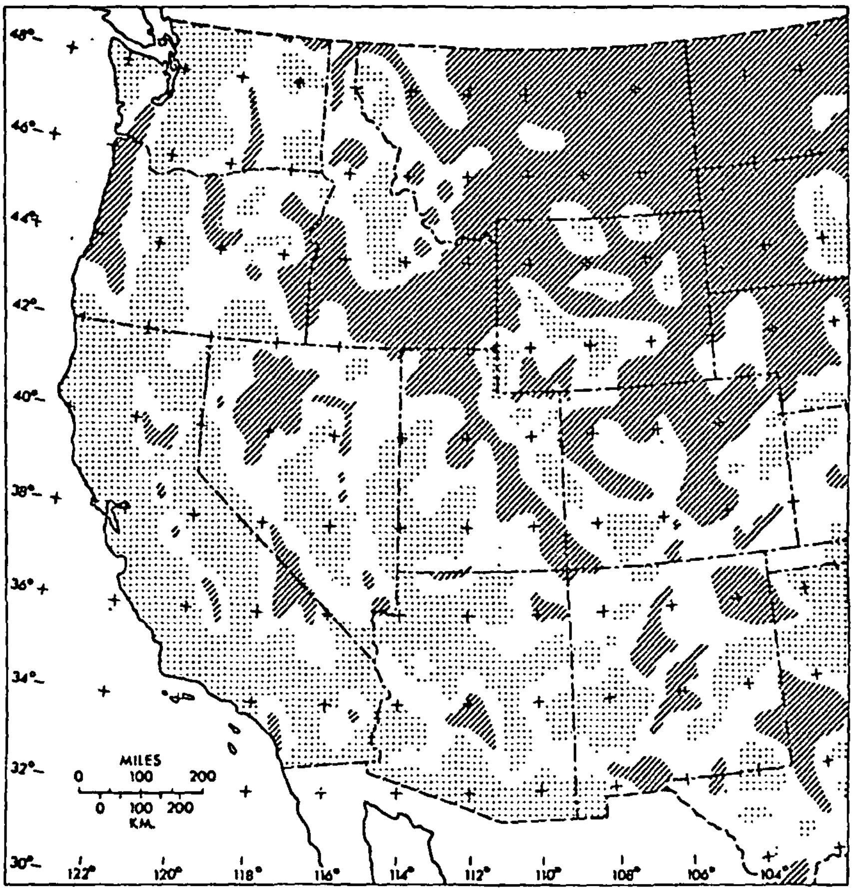


Figure 8 Regional isostatic gravity anomaly (based on Airy-Heiskanen concept with standard column 30 km). Line pattern, greater than +10 mgal; stippled pattern more negative than −10 mgal (from Woollard, 86).

In other important investigations Robinson & Kovach (56) studied upper mantle *S*-waves in the Basin and Range province, and Herrin (32) compared the Basin and Range upper mantle with that of a stable region, the Canadian Shield. Using direct measurements of the travel time gradient, Robinson and Kovach found a thin lid zone (9 km) of shear velocity 4.5 km/sec at the top of the mantle, overlying a low velocity zone with a minimum velocity at 100 km. Herrin's comparative model for the Canadian Shield contains no LVZ for *P*-waves and only a weak one for *S*-waves. The comparison is important because it emphasizes a degree of similarity between the Basin and Range and Colorado Plateau mantles relative to the stable region.

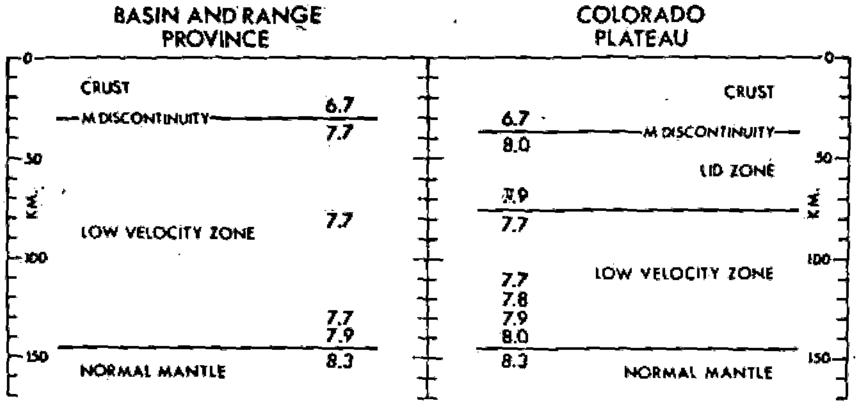


Figure 9 Generalized comparison of crust and upper mantle structure in the Basin and Range province and Colorado Plateau. *P*-wave velocities are in kilometers per second (adapted from Archambeau et al, 2).

RATE AND DIRECTION OF SPREADING

Seismological Evidence

Recent studies of focal mechanisms of many small earthquakes highlight a strikingly consistent direction of ongoing Basin and Range extension. Although recent earthquakes have been concentrated near the eastern and western borders of the province and in a belt across southern Utah and Nevada, evidence of older faulting indicates that they are a reasonable sample of this longer but much more widespread tectonic activity.

Focal solutions compiled by Scholz et al (64) show predominantly normal faulting, with the extension direction ranging approximately from east-west to northwest-southeast. The few examples of strike-slip motion are also consistent with this extension direction.

Only a few of the larger historical earthquakes were accompanied by surface ruptures large enough for the amount of offset to be directly observed, and these larger shocks (Figure 11) probably account for most of the total deformation. The main north-south zone of historical faulting in Nevada and adjacent California is nearly continuous. Horizontal extension across the faults ranges from a few centimeters to a few meters (Thompson, 75) and is greatest near the north and south ends of the zone. This wide range in extension, plus the existence of unfaulted gaps, shows that the 100-yr historical period is too short for measuring a meaningful rate of extension.

Dixie Valley, a Type Basin

Near the northern end of the zone of historical faulting, at the site of the 1954 faulting in Dixie Valley (Figure 12), two measures of long-term displacement have

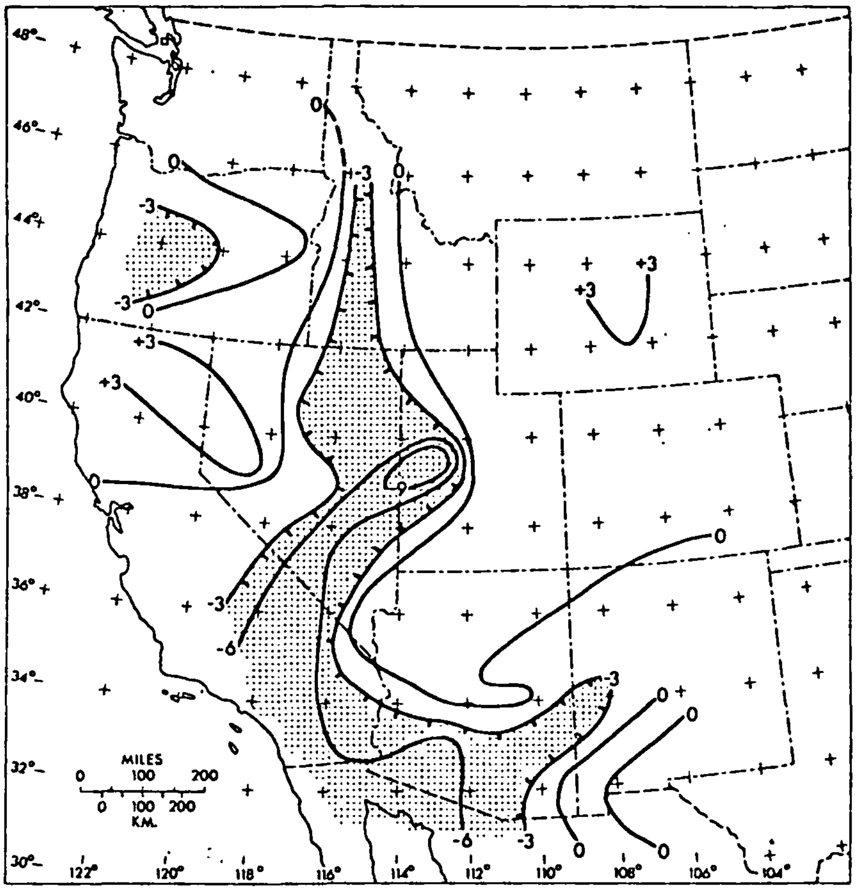


Figure 10 Relative development of upper mantle LVZ (low velocity zone), expressed as contours of time difference in seconds with respect to a model LVZ similar to that of Colorado Plateau. Stipple pattern accentuates region of pronounced (thicker or lower velocity) LVZ (from York & Helmberger, 87).

been investigated (Thompson & Burke, 78): 1. Displacements of the shoreline of a late Pleistocene lake supply a measure of extension during the last 12,000 years (Figure 13), and 2. fault displacements determined from geophysical exploration of the valley give the total amount of extension for late Cenozoic time, at least 5 km in 15 m.y. The average spreading rates are 1 mm/yr for the short interval and at least 0.4 mm/yr for the total displacement. The spreading direction we obtained from large slickenside grooves on fault planes is approximately N55°W-S55°E,

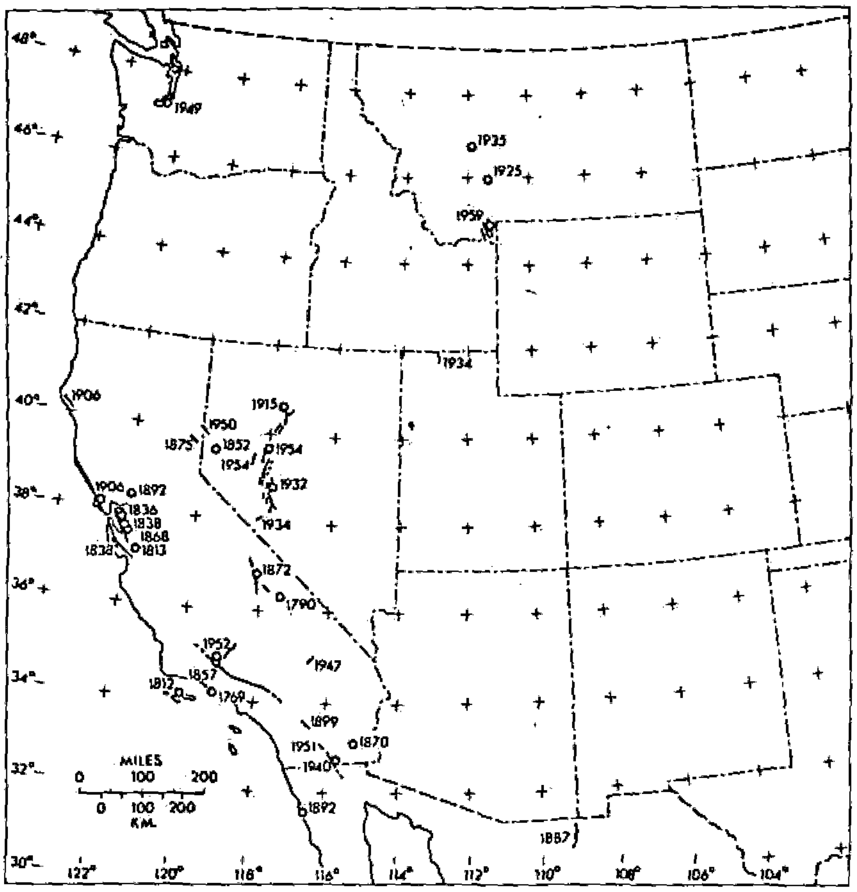
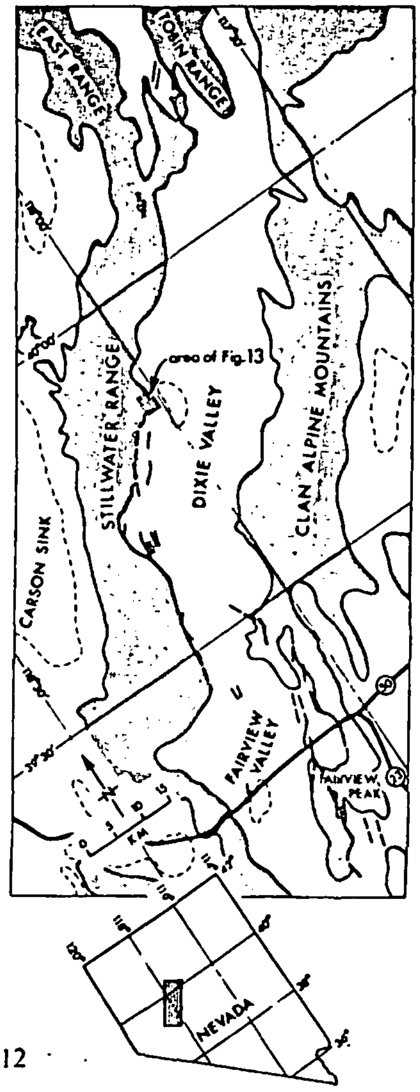


Figure 11 Historic surface offsets and epicenters for earthquakes of greater than about magnitude 7 in the western United States (from Ryall et al, 61).

which corresponds well with the range of directions obtained from earthquake focal mechanisms.

Dixie Valley is the only basin for which this much data is available. A simple extrapolation to 20 major basins across this part of the province suggests a total Basin-Range spreading of about 100 km (10% increase in crustal area) and a spreading rate of 8 mm/yr. On somewhat different assumptions, Gilluly (28) estimated that the areal expansion ranges from 4% to 12% over most of the province. Stewart (71) estimated 50 to 100 km (5% to 10%) of extension on the basis of a careful analysis of all available data. Hamilton & Myers (29) suggest that the extension may be as great as 300 km (30%). More subsurface data on many basins is needed to improve these estimates.



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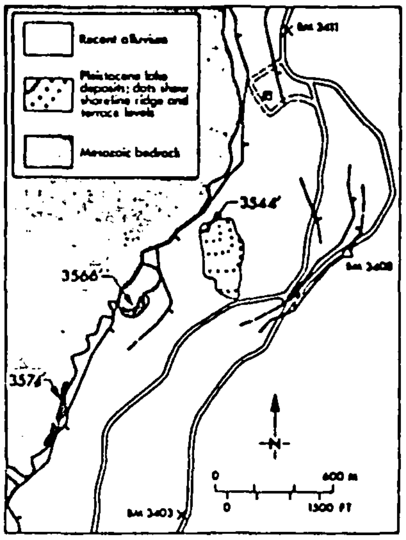

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Figure 13 (above) Map of offset lake shorelines in west-central Dixie Valley. The relative vertical spacing of beach ridges around the valley demonstrates that the highest beach ridge preserved in this area (3544 ft) marks—like the tufa cemented terrace deposits on bedrock—the highest lake stand. The age of the high shoreline is 12,000 years (from Thompson & Burke, 78).

Figure 12 (left) Dixie Valley region. Fault scarps formed or reactivated in 1903, 1915, and 1954 are shown (from Thompson & Burke, 78).

Locus and Time of Basin-Range Faulting

The present seismicity (Figure 3) is a misleading guide to even the geologically youngest faulting. Fault scarps of Quaternary age are widespread and bear little relation to the seismicity. Slemmons (69) has documented this fact for Nevada with maps of faults in three age groups covering roughly the last 100,000 years. The locus of faulting appears to have shifted randomly over the whole breadth of the province rather than having been confined to the area of recent seismic activity.

Although older normal faults are known (Burke & McKee, 9), the main onset of block faulting is marked by the widespread disruption of drainage and formation of local sedimentary basins about middle or late Miocene time. The lower Miocene ash-flow sheets which cover broad areas were deposited on surfaces of low tectonic relief (McKee, 46; Noble, 50). The inception of Basin-Range faulting over at least Nevada and adjacent California is dated at 15 to 17 m.y. (see Noble, 50, for references). It must be emphasized that after faulting began it was probably sporadic in any one area. On physiographic evidence, some areas appear to have been inactive for a long time (for example, parts of Arizona, New Mexico, and west Texas), while activity continued to the present in other areas.

THE PATTERN OF RUPTURE

Basin-Range faults are often described in a general way as high-angle normal faults striking north to northeast, but the impression conveyed by that description is highly misleading. Individual faults tend to be extremely crooked in map plan and the fault pattern is more nearly rhomboid or even rectilinear. Some mountain ranges are bounded by en echelon faults that strike diagonal to the range (eastern front of Sierra Nevada for example). Considerable warping and tilting of the blocks accompany the faulting, particularly near the ends of elongate basins.

Nowhere is the fault pattern better exhibited than in the late Cenozoic basalt flows of south-central Oregon (Figure 14), but similar patterns are common from Nevada (Figures 12 and 13) to Texas. Moreover, the roughly rhomboid map pattern of faulting is characteristic of other regions of present or past crustal extension, such as the African Rifts, the Rhine graben, the Oslo graben, and the Triassic basins of eastern North America.

No well-founded explanation for the complex rupture pattern is known. Alternative hypotheses include changes in the stress system with time, influence of older structures, and anisotropy in mechanical properties of the crust. Another possibility is that the pattern is roughly analogous to the near-orthogonal pattern formed by oceanic ridges and transform faults, a pattern which has been explained as offering minimum resistance to plate separation (Lachenbruch & Thompson, 42). Oldenburg & Brune (51) dramatically reproduced the near-orthogonal oceanic pattern in a laboratory model with a thin crust of wax forming on molten wax, and Duffield (18) observed similar patterns forming on the solidified crust of a convecting lava lake.

The simple application of the minimum resistance theory to the Basin and Range province would suggest a series of northeast-trending grabens (normal to the spreading direction) and northwest-trending transform faults. The actual mechanics are more complex, and the faults commonly are hybrid, having components of

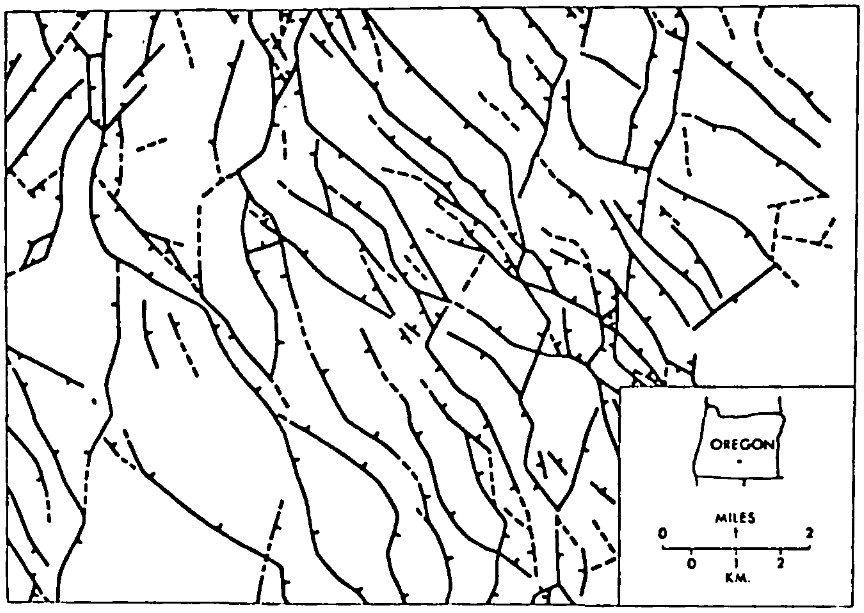

Figure 14 Rhomboid pattern of rupture expressed by late Cenozoic normal faults in south-central Oregon. Barbs on downthrown side of faults; faults dashed where inferred (from southeast portion of plate 3 of Donath, 17).

both dip- and strike-slip. The pattern is not simple and the question of the rupture pattern is far from resolved.

In addition to the problem of the pattern of faulting, the question of whether the normal faults systematically flatten with depth has been much debated, in part because such changes would imply greater regional extension. The seismic focal mechanisms lend no support to the notion of major decreases in dip, however, and serious geometric problems would ensue at the ends of basins if such decreases did occur. Therefore the low dipping to subhorizontal normal faults that have been observed in surface exposures and mine workings seem best ascribed to gravitational sliding and tilting in response to deeper primary faulting. The problem has been explored by Stewart (71) and Moore (48). Armstrong (3) interprets low-angle faults in eastern Nevada as gravitational sliding features of late Cenozoic age.

HEAT FLOW AND CRUSTAL TEMPERATURE

Regional Variation of Heat Flow

A region of anomalously high heat flow comprises the entire Basin and Range province and extends across the Columbia Plateau and part of the Rocky Mountain province (Figure 15). Heat flow values greater than 2 HFU [heat flow

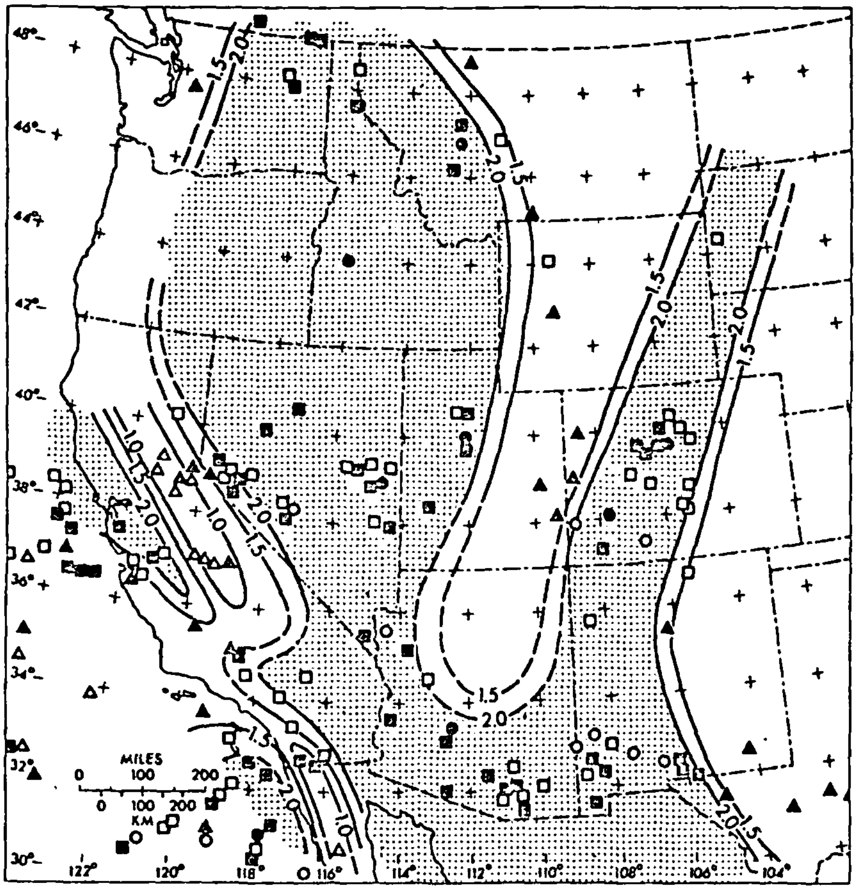

Figure 15 Contour map of heat flow. Contours in Heat Flow Units (μcal cm^{-2} sec^{-1}); dashed where extended on the basis of meager data. Data points shown as open triangles are measured heat flows in the range 0 to 0.99; solid triangles, 1.0 to 1.49; open squares, 1.5 to 1.99; solid squares, 2.0 to 2.49; open circles, 2.5 to 2.99; solid circles, 3.0 and larger (from Roy et al, 60).

units (HFU), μcal cm^{-2} sec^{-1}] characterize this broad region, in contrast to normal average values of about 1.5 HFU.

Although the Colorado Plateau is at least partly an area of normal heat flow, the distribution of measurements is inadequate to explore its boundaries with the Basin and Range province. The boundary with the Sierra Nevada appears to be surprisingly sharp.

Another compilation of the regional heat flow, by Sass and associates (63),

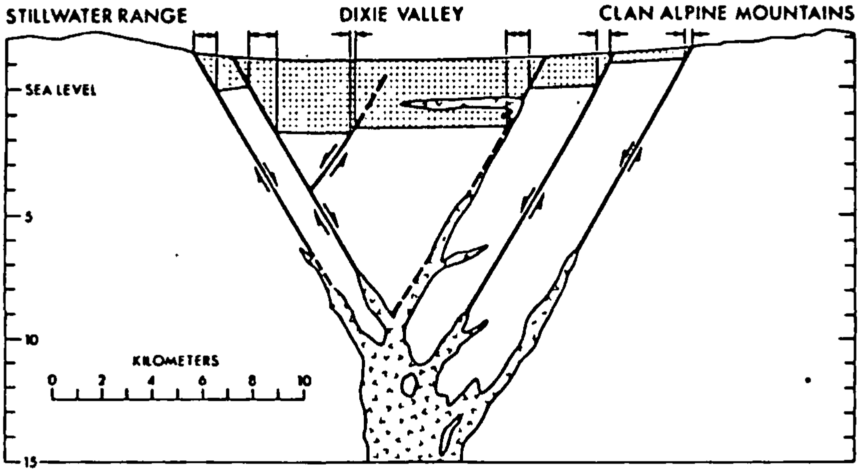

Figure 16 Cross section of Dixie Valley, Nevada. The subsurface structure to the depth of the sedimentary fill (stippled) is based on geophysical exploration. Dike at depth is hypothesized to accommodate surface extension, as shown by arrows (based on Burke, 8; Thompson, 76).

although more conservatively contoured, contains important additional details. One cluster of consistently high values (mostly above 3 HFU), the "Battle Mountain high" in northern Nevada, is interpreted as a transient effect of fairly recent crustal intrusion. To the south in Nevada, a cluster of values less than 1.5 HFU, the "Eureka low," is thought to be the result of unusual deep circulation of ground water. These examples emphasize the importance of nonconductive heat transfer. We point out that spreading of the grabens may be accompanied by intrusion of dikes at depths of a few kilometers (Thompson, 76), and these intrusions may be important in the heat transfer (Figure 16). The Battle Mountain high is on the projection of the active zone of spreading (historic fault breaks) at its north end (Figure 11).

The thermal transition to the Sierra Nevada may occur within a lateral distance of only 10 or 20 km (Sass et al, 63). If this proves to be the case, it will require shallow heat sources and will strengthen the hypothesis of intrusions beneath the grabens. Furthermore, present evidence suggests that the heat flow boundary with the Sierra Nevada follows in detail the irregular boundary of the normal faulting and not the generalized physiographic or topographic boundary.

Heat Production and the Linear Heat Flow Relation

A surprising and remarkably simple relationship has been found between heat flow and the heat production of surface rocks in plutonic areas; that is, within areas such as the Sierra Nevada and Basin and Range province, the heat flow varies linearly with the radioactive heat production at the surface (Roy et al, 60). This relationship is best explained by an exponential decrease of heat production with depth in the crust, combined with an additional flow of heat from the mantle (Lachenbruch, 41). The flow from the mantle—called the reduced heat flow—amounts to 1.4±0.2 HFU in the Basin and Range province, compared to 0.8±0.1 HFU in the United States east of the Rocky Mountains and only 0.4 HFU in the Sierra Nevada (Roy et al, 60).

Crustal temperature profiles for the three heat-flow provinces have been calculated by Lachenbruch (41), based on the exponential model. Temperatures at a depth of 30 km in the Basin and Range province range from 700–1000°C (depending on surface heat flow or heat production), as compared to 400–600°C in eastern United States. Temperatures in the Basin-Range crust may thus reach the melting range for granite, and temperatures in the upper mantle may reach melting for basalt. These high temperatures, combined with widespread late Cenozoic volcanism, form a basis for the generally accepted hypothesis that partial melting is responsible for the thin lithosphere and for the shallow, accentuated low velocity zone (asthenosphere) of the Basin and Range province.

The conductive model will need to be modified if much heat is carried into the crust by intrusions beneath spreading centers, as we have suggested (Figure 16).

Hot-spots and Mantle Plumes?

The Yellowstone volcanic region in northwestern Wyoming may represent a hot-spot above an upwelling convective plume in the mantle (Morgan, 49). According to Morgan's theory the North American lithosphere as a whole is moving west-southwest with respect to the mantle. The trail of the persistent Yellowstone hot-spot across its mantle plume would be marked by the older volcanics west-southwest of Yellowstone (in the Snake River part of the Columbia Plateau province). Other possible hot-spots have been suggested within the Basin and Range province.

A significant point about the theory should be kept in mind regarding the origin of the fault-block structures. If the Yellowstone plume is a driving mechanism for the structures and the lithosphere is moving westward across it, the locus of Basin-Range tectonic activity should be migrating eastward; and we know of no strong evidence for an eastward march of tectonic activity. Westward movement of the lithosphere at a rate on the order of 1 cm/yr would have produced a movement of 150 km in the 15 m.y. since the inception of Basin-Range faulting.

MAGNETIC AND ELECTRICAL ANOMALIES

Anomalies in the regional magnetic field and in electrical conductivity generally support other evidence of a hot upper mantle in the Basin and Range province, but the resolution of lateral variations has so far been very limited.

Zietz (88) showed that from the Sierra Nevada to the Rocky Mountains, magnetic anomalies are subdued in amplitude, and that long-wavelength anomalies are absent. This fact suggests that the lower crust and mantle may be above the Curie temperature (578° for magnetite). Surprisingly, the magnetic field over the Colorado Plateau does not appear to differ significantly from that over the Basin and Range province, in contrast to results from other kinds of studies.

Porath & Gough (53) explored variations in mantle electrical conductivity from the eastern and southern Basin and Range province to the Great Plains by measuring geomagnetic fluctuations. The anomalies are well represented by variations in depth to a half-space of conductivity 0.2 $(\text{ohm m})^{-1}$. The top of this conductor is inferred to correspond approximately with the 1500° isotherm. Depths to the surface of the conductor are 190 km under the Basin and Range province and 350 km under the Colorado Plateau, with a ridge of depth 120 km at the boundary. The depth under the Rio Grande trough is 120 km, that under the southern Rocky Mountains is 150 km, and that under the Great Plains is 350 km. Although such models are naturally not unique, they strengthen the interpretations of regional heat-flow variations and add another dimension to the unusual properties of the Basin and Range province.

PETROLOGIC RELATIONS

Three important relationships among the rocks deserve special emphasis:

1. Prior to Basin and Range faulting, lower and middle Cenozoic volcanoes erupted largely intermediate-composition rocks that become more alkalic toward the continental interior (Lipman et al, 43). This pattern is similar to volcanics now being erupted around the Pacific margin in association with convergent plate margins.

2. A major change to fundamentally basaltic volcanism (including bimodal mafic-silicic associations) took place during late Cenozoic time at about the inception of Basin-Range faulting (Christiansen & Lipman, 12). The transition to this new volcanism began in the southeastern part of the region and moved northwestward. The time of transition may be correlated with the initial intersection of the East Pacific Rise with the continental-margin trench system, an intersection which Atwater (5) also interprets as having progressed northwestward.

3. The composition of the crust and upper mantle as it existed beneath the Colorado Plateau prior to Basin-Range faulting has been ingeniously reconstructed from crystalline rock fragments in a breccia-filled diatreme, which is about 30 m.y. old (McGetchin & Silver, 45). The crust contained about 31% intermediate and acidic igneous rocks, 66% basic metaigneous rocks, and 3% eclogite. The upper mantle to a depth of about 100 km contained about 75% peridotite and pyroxenite and 25% eclogite. It is especially interesting that the mantle 30 m.y. ago contained this much eclogite, because eclogite is capable of converting into gabbro with a volume expansion of about 10% in response to a rise in temperature or decrease in pressure.

Eclogite may be a key to an understanding of late Cenozoic uplift of the broad region that includes the Sierra Nevada, Basin and Range province, and Colorado Plateau. The expansion of eclogite in only 60 km of mantle could produce an uplift of 1.5 km (60 × 25% × 10%). The former eclogite may now be represented by gabbro dispersed in the mantle low velocity zone, or by crustal additions of basic metaigneous rock, or by basaltic volcanics.

SYNTHESIS AND TECTONIC MODEL

The regional geophysical data put many useful constraints on speculations about the fundamental tectonic processes of the Basin and Range province. Among these data the heat flow is central; the volcanism, thin crust, low mantle velocity, accentuated low velocity zone, generally high elevation, subdued magnetic anomalies, high electrical conductivity, and great breadth of the seismically active zone can logically be associated with high temperatures and high heat flow.

The gravity data—coupled with the estimated extension—supply an interesting constraint that does not seem to have been widely recognized (Thompson, 77). If a 30 km crustal plate were simply attenuated by a horizontal extension of 10%, a negative isostatic anomaly of more than 300 mgal would be produced. If the attenuated plate were only 10 km thick the anomaly would still be 100 mgal. Because the regional isostatic anomalies average no more than about 10 mgal, the gravity emphatically indicates that the circuits of mass flow must be closed. Near-surface crustal spreading is almost perfectly matched by lateral backflow in the mantle.

If we imagine a vertical fence surrounding the Basin and Range province and extending through the crust and mantle, the integrated flux of mass through the fence must be zero, despite the outward flow by extension in the upper crust. We now need to find out how the deep lateral inflow takes place. Is the lateral flow in the low velocity zone? Is it a deeper mantle flow associated with narrow upwelling convective plumes, analogous to a thunderhead in the atmosphere? Is the flow related to former subduction of an oceanic lithospheric plate at the continental margin? At present these questions lead rather quickly into speculation.

The regional geophysical characteristics, geologic history, and petrology rather strongly suggest a link with plate-tectonic interactions at the western edge of the continent going back to early Cenozoic time. Analogies with spreading marginal basins of the western Pacific are especially promising (Karig, 40; Matsuda & Uyeda, 47; Scholz et al, 64; Sleep & Toksöz, 68; Thompson, 77; Uyeda & Miyashiro, 80).

The general idea is that in a broad belt on the continental side of an arc-trench system, a descending lithospheric plate either generates magma along its upper surface or creates a convecting subcell by viscous drag. The rising magma or convection current helps to move the arc away from the continent, creating a spreading marginal basin. The situation is somewhat different along the central coast of North America in that subduction ceased when a spreading ridge reached the trench in middle Cenozoic time. But because the descending young lithosphere would still be very hot when the ridge reached the trench, and because conductive heat transfer is very slow, it is easy to imagine that the thermal effects of a past subduction process are still being felt in the Basin and Range province.

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Springs of Western United States

NOAA

TABLE 1

Paul J. Grim AUG 10 '77

UNIVERSITY OF UTAH
RESEARCH INSTITUTE
EARTH SCIENCE LAB.

AREA
US west
Springs

ARIZONA

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
	AZ	1	21	78	26		0	31.880	109.204	SOUTHWESTERN RESEARCH STATION
	AZ	1	20	89	32		1	31.595	110.963	AGUA CALIENTE SPRING
	AZ	1	18	89	32		1	31.941	113.010	AGUAJITA SPRING
	AZ	1	14	121	50		1	32.998	109.901	INDIAN HOT SPRINGS
●	AZ	2		179	82		1	32.971	109.350	GILLARD HOT SPRINGS
	AZ	1	12	90	32		1	32.982	110.376	(WARM SPRING)
	AZ	2	19	130	55		1	32.334	110.238	HOOKERS HOT SPRINGS
	AZ	3		89	32		1	32.281	110.730	AGUA CALIENTE
	AZ	1	11	104	39		1	32.984	113.324	AGUA CALIENTE SPRINGS
	AZ	1		139	60		1	32.741	114.068	RADIUM HOT SPRINGS
	AZ	1		120	48		1	33.400	109.152	HANNAH HOT SPRING
●	AZ	2		138	59		1	33.080	109.303	(HOT SPRING)
	AZ	3	13			W	1	33.068	109.975	TOM NIECE SPRING
	AZ	4		120	48		0	33.059	109.441	
●	AZ	5	17	166	75		0	33.055	109.295	CLIFTON HOT SPRINGS
	AZ	6	15			W	1	33.051	109.558	QUAIL SPRING
●	AZ	7	16	96	36		1	33.046	109.440	EAGLE CREEK HOT SPRINGS
	AZ	8		91	33		0	33.008	109.778	GRAPEVINE SPRING (7)
	AZ	9		91	33		0	33.000	109.801	
	AZ	1	9	78	25		0	33.831	110.574	SALT BANKS
	AZ	2	10	65	18		1	33.832	110.285	SODA SPRING
	AZ	3				W	1	33.780	110.317	(WARM SPRING)
	AZ	4		83	28		0	33.748	110.235	
	AZ	5		84	29		1	33.436	110.213	(WARM SPRINGS)
	AZ	6		91	33		0	33.209	110.818	PIONEER SPRING
	AZ	7		98	37		0	33.198	110.668	
	AZ	8		96	36		1	33.170	110.528	COOLIDGE DAM SPRING (10)
	AZ	9		85	30		1	33.152	110.639	MESCAL WARM SPRING
	AZ	10				W	0	33.016	110.030	
	AZ	1		118	48		1	33.669	111.165	ROOSEVELT DAM SPRING (10)
●	AZ	1	8	121	50		1	33.986	112.359	CASTLE HOT SPRINGS
	AZ	1				W	1	34.621	109.575	STINKING SPRINGS
	AZ	2	7	74	23		1	34.433	109.401	SALADO SPRINGS
	AZ	1	5	75	24		1	34.649	111.744	SODA SPRING
●	AZ	2	6	96	36		1	34.357	111.710	VERDE HOT SPRINGS
	AZ	3		85	30		1	34.076	111.708	(SPRING (HOT))
	AZ	1		98	37		1	34.695	113.572	COFER HOT SPRING
	AZ	1	4	80	27		1	34.895	114.310	OATMAN WARM SPRINGS
	AZ	1				H	1	35.984	114.742	(HOT SPRING)
	AZ	2	2	94	35		1	35.960	114.725	RINGBOLT CANYON SPRING (10)
	AZ	1		85	30		0	36.509	111.852	COLORADO POOL (10)
	AZ	1	1	100	38		1	36.418	113.956	PAKOON SPRINGS
	AZ	2	3	89	32		1	36.195	113.081	LAVA WARM SPRINGS

Circ. 736	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
								CALIFORNIA		
	CA	1	180	101	38		1	32.953	116.305	AGUA CALIENTE SPRINGS
	CA	2	181	101	38		1	32.517	116.189	JACUMBA SPRINGS
	CA	1		91	32		1	33.602	114.728	NICHOLLS WARM SPRINGS
	CA	1	176	84	28		1	33.512	115.827	DOS PALMAS SPRING
●	CA	2	176A	179	82		1	33.423	115.685	HOT MINERAL SPA (PILGER ESTATE HOT SPRS.)
	CA	3		75	23		1	33.371	115.637	FRINK SPRING
	CA	4				B	1	33.217	115.580	MUD POT
	CA	5				B	1	33.212	115.591	MUD POTS
	CA	6				B	1	33.201	115.577	MUD POTS
	CA	1	172A	112	44		1	33.969	115.942	HIGHLAND SPRINGS
	CA	2	174A	116	46		1	33.962	116.507	DESERT HOT SPRINGS
	CA	3	173	117	47		1	33.835	116.988	GILMAN (SAN JACINTO, RELIEF) HOT SPRINGS
	CA	4	175	100	37		1	33.823	116.542	PALM SPRINGS (AGUA CALIENTE SPRING)
	CA	5	174	111	43		1	33.800	116.927	SOBODA (RITCHEY) HOT SPRINGS
	CA	6				H	0	33.540	116.742	
	CA	7	182	90	32		1	33.407	116.035	FISH SPRINGS
●	CA	8	179	147	64		1	33.284	116.631	WARNER (LAS AGUAS CALIENTE) HOT SPRING
	CA	1	172	109	42		1	33.896	117.060	EDEN HOT SPRINGS
	CA	2				H	0	33.865	117.103	
	CA	3	171	100	37		1	33.837	117.145	LAKEVIEW (PILARES) HOT SPRINGS
	CA	4	167	131	54		1	33.753	117.495	GLEN IVY (TEMESCAL) HOT SPRINGS
	CA	5	168	125	51		1	33.680	117.320	WRENDEN (BUNDYS) AND ELSINORE HOT SPRINGS
	CA	6	165	96	35		0	33.664	117.913	FAIRVIEW HOT SPRING
	CA	7	156	123	50		1	33.589	117.511	SAN JUAN (CAPISTRANO) HOT SPRINGS
	CA	8	170	132	55		1	33.558	117.154	MURRIETTA HOT SPRINGS
	CA	9		115	46		1	33.551	117.167	TEMECULA HOT SPRINGS
	CA	10	177	85	29		0	33.436	117.324	DE LUZ WARM SPRINGS
	CA	11	178	92	33		1	33.362	117.012	AGUA TIBIA SPRING
	CA	1		77	24		0	33.802	118.393	
	CA	2		114	45		0	33.731	118.348	WHITES POINT HOT SPRINGS
	CA	1	157	77	24		0	34.824	116.587	NEWBERRY SPRING
	CA	2	154	88	31		1	34.271	116.838	PAN HOT SPRINGS
	CA	1	160	100	37		1	34.340	117.173	(WARM SPRING) AND (HOT SPR)
	CA	2	158	92	33		1	34.228	117.483	TYLERS BATH SPRING
●	CA	3	162	94	34		1	34.185	117.262	ARROWHEAD SPRINGS AND WATERMAN HOT SPRING
	CA	4	153	90	32		1	34.124	117.078	
	CA	5	161	120	48		1	34.121	117.232	HARLEM HOT SPRINGS
	CA	1	112	92	33		1	34.605	118.561	ELIZABETH LAKE CANYON WARM SPRING
●	CA	2	111	191	88		1	34.595	118.999	SESPE HOT SPRINGS
	CA	3	112A	85	29		1	34.108	118.791	SEMINOLE HOT SPRINGS
	CA	1	110	108	42		0	34.583	119.044	WILLETT HOT SPRINGS
	CA	2	105	90	32		0	34.542	119.451	
	CA	3		133	56		1	34.538	119.560	AGUA CALIENTE SPRING
	CA	4	104	90	32		1	34.537	119.513	LITTLE CALIENTE SPRING
	CA	5	102	110	43		1	34.537	119.879	SAN MARCOS HOT SPRING

Circ. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
			FAHRENHEIT	CELSIUS					
CA	6	109	102	38		1	34.507	119.291	WHEELERS (HOT) SPRINGS
CA	7	107	123	50		1	34.500	119.341	VICKERS AND STINGLEYS HOT SPRINGS
CA	8	108	109	42		1	34.482	119.302	MATILIJA HOT SPRINGS
CA	9		93	33		0	34.481	119.883	TECOLATE TUNNEL
CA	10	103	112	44		1	34.461	119.637	MONTECITO HOT SPRINGS AND ARSENIC SPRINGS
CA	1	101	108	42		1	34.503	120.219	LAS CRUCES HOT SPRINGS
CA	1				W	1	35.978	116.273	SHOSHONE SPRING
CA	2	145	80	26		1	35.946	116.189	CHAPPO SPRING (YEOMAN HOT SPRINGS)
CA	3				W	1	35.888	116.258	
CA	4	147	80	26		1	35.879	116.153	RESTING SPRING
CA	5	146	108	42		1	35.871	116.230	TECOPA HOT SPRINGS
CA	6	154	82	27		1	35.682	116.422	SARATOGA SPRING
CA	7	155	102	38		1	35.143	115.813	PARADISE SPRING
CA	8	156	75	23		1	35.141	116.101	SODA SPRINGS
CA	1	143	80	26		0	35.934	117.913	
CA	2		92	33		1	35.842	117.380	BAINTER SPRING
CA	1	137	126	52		1	35.881	118.670	CALIFORNIA (DEER CREEK) HOT SPRINGS
CA	2	148	113	44		1	35.726	118.408	(HOT SPRINGS)
CA	3	149	131	54		1	35.620	118.473	SCOVERN (NEILLS) HOT SPRINGS
CA	4	150	119	48		1	35.575	118.534	MIRACLE AND HOBO (CLEAR CREEK) HOT SPRINGS
CA	5	151	112	44		1	35.558	118.612	DELONEGHA SPRINGS
CA	6				H	1	35.536	118.651	(SPRING (HOT))
CA	7	152	115	46		1	35.528	118.665	DEMOCRAT SPRINGS
CA	8	153	100	37		1	35.434	118.478	YATES (WILLIAMS) HOT SPRINGS
CA	1	95	110	43		0	35.658	120.694	PASO DE ROBLES MUD BATH SPRINGS
CA	2	97	92	33		1	35.582	120.666	SANTA YSABEL SPRINGS (SULPHUR SPR.)
CA	3	99	95	34		0	35.269	120.851	PECHO WARM SPRINGS
CA	4	98A	100	37		1	35.185	120.714	SYCAMORE (SAN LUIS HOT) SPRINGS
CA	5		135	57		1	35.181	120.702	HIDDEN VALLEY HOT SPRINGS
CA	6	100	99	37		1	35.122	120.542	NEWSOM'S (ARROYO GRANDE) SPRINGS
CA	1	140A	93	33		1	36.675	116.923	KEENE WONDER SPRINGS
CA	1				W	1	36.830	117.736	UPPER WARM SPRING
CA	2		120	48		1	36.814	117.763	PALM SPRING
CA	3	139	110	43		1	36.806	117.771	LOWER (BURRO) WARM SPRINGS
CA	4				H	1	36.330	117.948	DIRTY SOCKS (HOT) SPRING
CA	5	144	80	26		1	36.122	117.217	WARM SULPHUR SPRINGS
● CA	6	142			B	1	36.045	117.769	COSO HOT SPRINGS
CA	7	141A	206	96		1	36.036	117.802	DEVILS KITCHEN
CA	8	141	203	94		1	36.031	117.833	(FUMAROLE)
CA	1	135	123	50		1	36.477	118.404	KERN (JORDAN) HOT SPRING
CA	2	136	100	37		1	36.210	118.176	SODA SPRINGS (MONACHE MEADOWS)
CA	3	134	77	24		1	36.149	118.656	MOOREHOUSE SPRINGS
CA	1		81	27		0	36.765	120.901	
● CA	2	132	119	48		1	36.703	120.860	MERCEY HOT SPRINGS
CA	3		75	23		0	36.640	120.684	
CA	4	133	112	44		1	36.144	120.555	COALINGA MINERAL (FRESNO HOT) SPRINGS
CA	1		77	24		1	36.938	121.565	SARGENT ESTATE
CA	2		100	37		0	35.619	121.846	SULPHUR HOT SPRING

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS				
	CA	3	90	114	45	0	36.333	121.840	
	CA	4	92	98	36	1	36.333	121.367	PARAISO HOT SPRINGS
	CA	5	91	144	62	1	36.234	121.546	TASSAJARA HOT SPRINGS
	CA	6	93	122	49	1	36.122	121.638	SLATES (BIG SUR) HOT SPRINGS
	CA	7	94	98	36	1	36.084	121.584	DOLANS HOT SPRING
	CA	1	140	100	37	1	37.029	117.383	GRAPEVINE SPRINGS
	CA	1				H 1	37.977	118.926	(HOT SPRING)
	CA	2	127A	70	21	0	37.884	118.487	BERTRAND RANCH
●	CA	3	127	134	57	1	37.802	118.532	BENTON HOT SPRINGS
	CA	4		75	24	1	37.719	118.735	(SPRINGS)
	CA	5		67	20	1	37.708	118.813	(HOT SPRING)
	CA	6	122	179	82	1	37.692	118.839	LITTLE HOT CREEK SPRING
	CA	7		125	52	0	37.677	118.790	DEHY HOT SPRING AND OTHERS
	CA	8		127	53	0	37.667	118.781	
	CA	9		199	93	1	37.665	118.828	HOT CREEK SPRINGS
	CA	10		165	75	0	37.664	118.802	THE TUB AND OTHERS
	CA	11		120	49	0	37.657	118.764	
	CA	12		170	77	0	37.656	118.834	HOT CREEK SPRINGS
	CA	13		71	22	0	37.653	118.925	
	CA	14	125	129	54	1	37.648	118.806	(HOT SPRINGS)
	CA	15	123	199	93	1	37.648	118.914	CASA DIABLO HOT SPRINGS AND GEYSER
	CA	16	124	154	68	1	37.647	118.859	CASA DIABLO HOT POOL
	CA	17		85	30	0	37.639	118.756	
	CA	18		73	23	0	37.638	118.887	CHANCE SPRING
	CA	19		62	17	0	37.638	118.866	
	CA	20		127	53	0	37.635	118.720	
	CA	21	126	94	35	1	37.630	118.808	WHITMORE HOT SPRINGS
	CA	22		73	23	0	37.607	118.808	
	CA	23		62	17	0	37.557	118.705	
	CA	24	138	138	58	1	37.253	118.373	KEOUGH HOT SPRINGS
●	CA	25	131	109	43	1	37.234	118.881	BLANEY MEADOWS HOT SPRINGS
●	CA	1				H 0	37.997	119.045	PAOHA ISLAND FUMAROLE
●	CA	2	120	202	95	0	37.997	119.020	PAOHA ISLAND SPRINGS
	CA	3				H 0	37.622	119.028	FUMAROLES
●	CA	4	128	120	48	1	37.618	119.074	REDS MEADOW HOT SPRINGS
	CA	5	129	110	43	1	37.532	119.025	FISH CREEK HOT SPRINGS
	CA	6		95	34	0	37.414	119.140	
●	CA	7	130	111	44	1	37.327	119.018	MONO HOT SPRINGS
	CA	1	86	96	35	1	37.847	121.635	BYRON HOT SPRINGS
	CA	2	87	80	26	0	37.503	121.904	ALAMEDA WARM (MISSION SAN JOSE HOT) SPRS.
	CA	3	88	84	28	1	37.398	121.794	MINERAL (ALUM ROCK PARK) SPRINGS
	CA	4	89	106	41	1	37.108	121.478	GILROY HOT SPRINGS
	CA	1	85	81	27	0	37.924	122.045	SULPHUR SPRINGS
	CA	2	84	90	32	0	37.880	122.627	ROCKY POINT SPRINGS
	CA	1	121	91	32	1	38.033	118.902	MONO BASIN WARM SPRINGS
●	CA	1	113	145	63	1	38.699	119.845	GROVERS HOT SPRINGS
●	CA	2	114	143	62	1	38.348	119.400	FALES HOT SPRINGS
●	CA	3	116	157	70	1	38.244	119.205	TRAVERTINE HOT SPRINGS

CIRC. 736	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
●	CA	4	115	147	64		1	38.237	119.325	BUCKEYE HOT SPRING
	CA	5	117	113	44		1	38.223	119.217	THE HOT SPRINGS
	CA	6	118	100	37		1	38.210	119.151	WARM SPRINGS FLAT
	CA	7	119	100	37		1	38.203	119.118	(WARM SPRING)
●	CA	8		145	63		1	38.046	119.080	BLACK POINT HOT SPRING
	CA	1	113A	75	23		1	38.192	120.827	VALLEY SPRINGS
	CA	1				W	0	38.994	122.745	
	CA	2				W	0	38.982	122.734	
	CA	3				H	0	38.979	122.659	(FUMAROLE)
	CA	4				W	0	38.963	122.724	
	CA	5				W	0	38.956	122.699	
	CA	6				W	0	38.950	122.654	
	CA	7	52	82	27		1	38.936	122.907	HIGHLAND SPRINGS
	CA	8	54	76	24		1	38.916	122.799	CARLSBAD SPRING
●	CA	9		76	24		1	38.892	122.533	BAKER SODA SPRING
●	CA	10	59	125	52		1	38.873	122.689	SEIGLER SPRINGS
	CA	11	58	113	44		1	38.858	122.671	HOWARD SPRINGS
●	CA	12		71	22		0	38.833	122.357	ONE SHOT MINING CO.
	CA	13	72			B	1	38.802	122.810	THE GEYSERS
	CA	14	73	120	48		0	38.788	122.777	SULPHUR CREEK
	CA	15	64	120	48		1	38.785	122.655	HARBIN SPRINGS
	CA	16	63	128	53		1	38.773	122.705	ANDERSON SPRINGS
	CA	17	74	160	71		0	38.772	122.755	LITTLE GEYSERS
	CA	18	62	163	72		1	38.768	122.717	CASTLE ROCK SPRINGS
●	CA	19	80	91	33		1	38.655	122.484	AETNA SPRINGS
●	CA	20		66	19		1	38.652	122.355	WALTER (WALTERS MINERAL) SPRINGS
●	CA	21	81	172	77		0	38.580	122.575	CALISTOGA HOT SPRINGS
●	CA	22	75	87	31		1	38.550	122.720	MARK WEST SPRINGS
●	CA	23	83	78	26		1	38.519	122.256	NAPA ROCK (PRIESTS) AND PHILLIPS SODA SPR
	CA	24	82	90	32		0	38.490	122.498	ST. HELENA WHITE SULPHUR SPRING
●	CA	25	76	87	31		1	38.392	122.552	LOS GUILICOS (MORTONS) WARM SPRINGS
●	CA	26		60	16		1	38.391	122.279	(JACKSONS) NAPA SODA SPRINGS
	CA	27	77	73	22		0	38.391	122.570	MCEWEN RANCH WARM SPRINGS
	CA	28	78	70	21		0	38.350	122.515	SONOMA (ELDRIDGE) STATE HOME WARM SPRINGS
	CA	29		76	24		1	38.339	122.259	(NAPA) VICHY SPRINGS
	CA	30		83	28		0	38.324	122.268	
	CA	31	79	115	46		1	38.320	122.483	AGUA CALIENTE AND FETTERS HOT SPRINGS
	CA	32	79	112	44		1	38.311	122.471	BOYES (OHMS) HOT SPRINGS
●	CA	1		60	16		1	38.910	123.307	(OLD) ORNBAUN SPRINGS
●	CA	2	47	111	44		0	38.874	123.518	POINT ARENA HOT SPRINGS
	CA	3	70	100	37		0	38.800	123.170	HOODS (FAIRMONT) HOT SPRINGS
●	CA	4	71	134	57		1	38.691	123.024	SKAGGS HOT SPRINGS
	CA	1		108	42		0	39.927	120.578	DOYLE HOT SPRINGS
	CA	2	41A	163	72		1	39.755	120.359	MARBLE HOT SPRINGS
	CA	3	42	86	29		0	39.728	120.547	MCLEAN SULPHUR (WARM) SPRING
	CA	4	43	111	43		1	39.572	120.348	CAMPBELL HOT SPRINGS
	CA	5				H	0	39.426	120.026	(STEAM VAPOR)
●	CA	6	44	139	60		0	39.226	120.010	BROCKWAY (CARNELIAN) HOT SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	TEMPERATURE CELSIUS	WARM, HOT, BOILING LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	CA	7	44A	75	23	1	39.015	120.338	WENTWORTH SPRINGS
●	CA	1		76	25	1	39.430	122.538	SALT SPRING AND SULPHUR SPRING
●	CA	2		62	17	0	39.413	122.977	SODA SPRING
●	CA	3	48A	78	26	1	39.348	122.668	FOUTS SRINGS (RED EYE AND CHAMPAGNE SPRS.)
●	CA	4	48	105	41	1	39.292	122.821	CRABTREE HOT SPRINGS
●	CA	5		62	17	1	39.255	122.526	COOKS SPRINGS
	CA	6	49	92	33	1	39.196	122.714	NEWMAN SPRINGS
●	CA	7		60	16	1	39.174	122.980	SARATOGA SPRINGS
	CA	8	51	74	23	1	39.171	122.511	COMPLEXION SPRING
●	CA	9	65	78	26	1	39.085	122.459	DEADSHOT SPRING
	CA	10	51A	70	21	0	39.075	122.580	CHALK MOUNTAIN
	CA	11		70	21	0	39.064	122.595	
	CA	12	69	153	67	1	39.057	122.475	ELGIN MINE
●	CA	13	68	139	60	1	39.037	122.420	WILBUR SPRINGS AND OTHERS
	CA	14	66	120	48	1	39.032	122.432	BLANK SPRING
	CA	15		85	30	1	39.022	122.444	ABBOT MINE
	CA	16		80	26	0	39.019	122.581	
	CA	17	55	87	30	1	39.008	122.789	BIG SODA (SODA BAY) SPRING
●	CA	18	57	156	69	1	39.002	122.664	SULPHUR BANK
	CA	1	45A	70	21	1	39.699	123.482	(SPRING)
	CA	2		80	26	1	39.663	123.584	JACKSON VALLEY MUD SPRINGS
●	CA	3	45	103	40	1	39.227	123.362	ORRS HOT SPRINGS
●	CA	4	46	89	32	1	39.165	123.159	VICHY SPRINGS
	CA	1	29A	70	21	1	40.580	120.255	TIPTON SPRINGS
	CA	2		72	22	1	40.568	120.325	SELLICKS SPRINGS
	CA	3		204	95	1	40.364	120.243	(HOT SPRING)
●	CA	4	30	205	96	1	40.355	120.257	WENDEL HOT SPRINGS
●	CA	5	31	204	95	1	40.302	120.195	AMEDEE HOT SPRINGS
	CA	6	32	100	37	1	40.245	120.005	HIGH ROCK RANCH SPRING
	CA	7	39	122	49	1	40.142	120.935	INDIAN VALLEY (KRUGER) HOT SPRINGS
	CA	1	25	150	65	0	40.457	121.545	MILL CREEK SPRINGS
	CA	2	27	199	93	1	40.455	121.501	BUMPASS HELL (BUMPASS HOT SPRINGS)
	CA	3	26	199	93	1	40.447	121.536	SULPHUR WORKS (SUPAN S (TOPHET HOT) SPRS.)
	CA	4	36	148	64	1	40.444	121.409	DRAKESBAD (DRAKE HOT SPRINGS)
	CA	5	34	205	96	1	40.440	121.434	DEVILS KITCHEN
	CA	6	35	83	28	0	40.440	121.420	HOT SPRINGS VALLEY
	CA	7	37	190	87	1	40.434	121.399	BOILING SPRINGS LAKE
	CA	8	38	205	96	1	40.421	121.375	TERMINAL GEYSER
	CA	9		203	94	1	40.393	121.507	GROWLER HOT SPRING
●	CA	10	33	202	95	1	40.382	121.513	MORGAN HOT SPRING
	CA	11	40	100	38	1	40.019	121.036	(SPRING)
●	CA	1		67	20	0	40.673	122.634	SALT SPRING
●	CA	2	45B	85	30	1	40.238	122.110	TUSCAN (LICK) SPRINGS
	CA	3		101	38	1	40.223	122.747	STINKING SPRINGS
	CA	1				W 1	41.959	120.936	(WARM SPRING)
	CA	2	12	109	43	0	41.873	120.163	PETERSON RANCH
	CA	3		114	46	1	41.860	120.158	FORT BIDWELL HOT SPRINGS
	CA	4	4	78	25	1	41.828	120.917	POTHOLE SPRING

Circ. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
			FAHRENHEIT	CELSIUS					
CA	5	13	67	20		1	41.726	120.081	BOYD HOT SPRING
● CA	6	14	206	97		1	41.670	120.206	LAKE CITY (SURPRISE VALLEY) HOT SPRINGS
CA	7	16	188	87		1	41.615	120.102	SEYFERTH HOT SPRINGS
CA	8	17	150	66		1	41.600	120.088	LEONARDS HOT SPRINGS
CA	9	18	208	98		1	41.534	120.078	HOT SPRINGS (SURPRISE VALLEY AND BENMAC)
CA	10	6	91	33		1	41.492	120.700	HOT CREEK RANCH AND ESSEX SPRINGS
CA	11	7	81	27		1	41.484	120.764	(SPRING)
CA	12	10	72	22		1	41.452	120.515	(SPRING)
● CA	13	8	204	96		1	41.450	120.834	KELLY HOT SPRING
CA	14	20	138	59		1	41.266	120.080	MENLO HOT SPRINGS
CA	15		75	23		1	41.252	120.521	(WARM SPRING)
CA	16	21	109	43		0	41.219	120.068	SQUAW BATHS
CA	17				W	1	41.196	120.475	(SPRINGS)
CA	18		170	77		0	41.190	120.383	WEST VALLEY RESERVOIR HOT SPRING
CA	19	22	70	21		0	41.167	120.032	BARE RANCH
CA	20		70	21		1	41.160	120.403	(WARM SPRING)
CA	1	3A	191	88		1	41.607	121.523	HOT SPOT
CA	2	11	170	77		1	41.229	121.405	LITTLE HOT SPRING
CA	3	28	174	79		1	41.143	121.110	BASSETT HOT SPRINGS
CA	4	29	174	79		1	41.126	121.028	KELLOG (STONEBREAKER) HOT SPRINGS
● CA	5	23	136	58		1	41.036	121.924	HUNT (KOSK CREEK) HOT SPRING
● CA	6	24	179	82		1	41.025	121.924	BIG BEND HOT SPRINGS
CA	7				H	1	41.012	121.274	(HOT SPRINGS)
CA	1	2	152	66		1	41.973	122.202	KLAMATH HOT SPRINGS
CA	2	2A	76	24		1	41.919	122.369	BOGUS SODA SPRINGS
CA	3	3	184	84		1	41.407	122.197	(HOT SPRING ON MOUNT SHASTA)
CA	1	1	84	28		1	41.659	123.319	SULPHUR SPRINGS

COLORADO

CO	1	44	67	20		1	37.294	105.784	DEXTER WARM SPRING
CO	1	33	85	30		1	37.751	106.317	SHAWS WARM SPRING
● CO	2	31	134	57		0	37.754	106.828	WAGON WHEEL GAP HOT SPRINGS
CO	3	32	103	40		1	37.511	105.945	RAINBOW HOT SPRINGS
CO	4	41	80	27		1	37.033	106.805	STINKING SPRINGS
CO	1		89	32		1	37.741	107.034	ANTELOPE WARM SPRING
CO	2		85	30		0	37.718	107.060	BIRDSIE WARM SPRING (2)
CO	3	34	91	33		0	37.453	107.803	PINKERTON HOT SPRINGS
CO	4	35	111	44		0	37.400	107.849	TRIPP HOT SPRINGS
CO	5	36	96	36		0	37.391	107.846	TRIMBLE HOT SPRING
● CO	6	39	136	58		1	37.263	107.011	PAGOSA SPRINGS
CO	1	29	111	44		1	37.771	108.091	DUNTON HOT SPRING
CO	2		82	28		0	37.762	108.113	GEYSER WARM SPRING
CO	3		114	46		0	37.752	108.131	PARADISE HOT SPRING
CO	4	30	111	44		0	37.689	108.031	RICO HOT SPRINGS
CO	1	22	91	33		0	38.487	105.912	WELLSVILLE WARM SPRING
CO	2		82	28		0	38.479	105.891	SWISSVALE WARM SPRINGS (2)

Circ. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
	CO	3		96	36		1	38.459	105.200	FREMONT NATATORIUM HOT SPRING
	CO	4	22A	103	40		0	38.433	105.261	CANON CITY HOT SPRING
	CO	5		64	18		0	38.305	105.979	FULLINWIDER WARM SPRING
	CO	6	24	98	37		1	38.192	105.816	VALLEY VIEW HOT SPRINGS
●	CO	7	23	139	60		1	38.168	105.924	MINERAL HOT SPRINGS
	CO	1	13	78	26		1	38.836	106.825	CEMENT CREEK WARM SPRING
	CO	2	12	80	27		1	38.816	105.873	RANGER WARM SPRING
●	CO	3	19	136	58		1	38.812	106.226	COTTONWOOD HOT SPRINGS
●	CO	4	20	132	56		1	38.733	106.162	MOUNT PRINCETON HOT SPRINGS
	CO	5		181	83		1	38.732	106.178	HORTENSE HOT SPRING
	CO	6		76	25		0	38.653	106.056	BROWN S CANYON WARM SPRING (2)
	CO	7		73	23		0	38.634	106.072	BROWN S GROTTO WARM SPRING (2)
●	CO	8	14	175	80		1	38.512	106.507	WAUNITA HOT SPRINGS
●	CO	9	21	159	71		1	38.498	106.076	PONCHA HOT SPRINGS
●	CO	1	15	105	41		1	38.272	107.100	CEBOLLA (POWDERHORN) HOT SPRINGS
●	CO	2	27	127	53		1	38.133	107.736	ORVIS (RIDGWAY) HOT SPRING
	CO	3	28	156	69		0	38.021	107.672	OURAY HOT SPRINGS
	CO	1	26	91	33		0	38.014	108.054	LEMON HOT SPRING
	CO	1	4	78	26		1	39.932	105.277	ELDORADO SPRINGS
●	CO	2	5	114	46		1	39.742	105.510	IDAHO SPRINGS
	CO	3	17	125	52		1	39.017	105.793	HARTSEL HOT SPRINGS
	CO	1	16	76	25		1	39.164	106.062	RHODES WARM SPRING
	CO	2	9	100	38		1	39.012	106.891	CONUNDRUM HOT SPRING
	CO	1	7	89	32		0	39.624	107.106	DOTSERO WARM SPRINGS
	CO	2		118	48		0	39.552	107.412	SOUTH CANYON HOT SPRINGS
●	CO	3	6	123	51		1	39.548	107.320	GLENWOOD SPRINGS
●	CO	4	8	134	57		0	39.233	107.228	AVALANCHE SPRINGS
	CO	5		132	56		1	39.227	107.224	PENNY HOT SPRINGS
●	CO	1	2	147	64		1	40.559	106.849	ROUTT HOT SPRINGS
●	CO	2	2A	102	39		1	40.486	106.831	STEAMBOAT SPRINGS
	CO	3	3	111	44		1	40.073	106.113	HOT SULPHUR SPRINGS
	CO	1	1	100	38		1	40.467	107.952	JUNIPER HOT SPRINGS
								IDAHO		
	ID	1	192	107	42		1	42.910	111.556	(WARM SPRING)
	ID	2		107	42		0	42.891	111.702	
	ID	3	193	87	31		1	42.656	111.607	SODA SPRINGS
	ID	4	191	125	51		1	42.562	111.801	
	ID	5				H	1	42.336	111.719	(HOT SPRINGS)
●	ID	6		168	76		1	42.305	111.708	MAPLE GROVE HOT SPRINGS
●	ID	7		170	77		1	42.135	111.930	WAYLAND HOT SPRINGS
	ID	8				H	1	42.118	111.929	SQUAW HOT SPRINGS
	ID	9	196	118	48		1	42.114	111.264	BEAR LAKE HOT SPRINGS
	ID	1	189	89	32		1	42.725	112.873	INDIAN SPRINGS
	ID	2	190	112	45		1	42.620	112.014	LAVA HOT SPRINGS
	ID	3		100	37		1	42.542	112.903	(WARM SPRINGS)

				TEMPERATURE						
CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	ID	4	191A	109	43		1	42.389	112.089	DOWNATA HOT SPRINGS
	ID	5		75	24		0	42.341	112.434	
	ID	6		76	25		0	42.175	112.239	
	ID	7	194	76	25		1	42.153	112.349	PLEASANTVIEW WARM SPRINGS
	ID	8	195	80	27		1	42.053	112.242	WOODRUFF HOT SPRINGS
	ID	1	186	98	37		0	42.452	113.520	
	ID	2	183	84	29		0	42.236	113.579	SEARS SPRING (9)
	ID	3				W	1	42.226	113.780	(WARM SPRING)
●	ID	4	182	116	47		1	42.170	113.857	OAKLEY WARM (HOT) SPRING
	ID	5	184	204	95		1	42.107	113.390	(WELL (HOT)) (BRIDGE SPRING) (FRAZIER H S
	ID	6		96	36		0	42.099	113.635	DURFEE SPRING (9)
	ID	1	173	129	54		0	42.703	114.857	
	ID	2	174	129	54		1	42.587	114.859	HOT SULPHUR (MIRACLE HOT) SPRINGS
	ID	3	175	130	55		1	42.686	114.826	BANBURY HOT SPRING
	ID	4	178	100	37		0	42.413	114.161	ARTESIAN CITY HOT SPRINGS
	ID	5		96	36		1	42.347	114.509	NAT-SOO-PAH WARM SPRING
	ID	6	177	90	32		0	42.309	114.323	
	ID	7	181	69	20		0	42.015	114.237	THOROUGHBRED SPRINGS
	ID	8		114	46		0	42.005	114.504	MAGIC HOT SPRINGS
	ID	1	163	105	41		1	42.799	115.744	BRUNEAU HOT SPRINGS
	ID	2	165	114	46		1	42.799	115.721	BAT AND PENCE (TRAMMEL≠S) HOT SPRINGS
	ID	3	164	111	44		1	42.796	115.732	
	ID	4	169	111	44		0	42.779	115.716	
	ID	5	166	105	41		0	42.767	115.725	
	ID	6	167	102	39		1	42.762	115.739	INDIAN BATHTUB HOT SPRINGS
	ID	7	169A	156	69		0	42.337	115.646	INDIAN HOT SPRINGS
●	ID	8	169B	123	51		1	42.030	115.367	MURPHY HOT SPRINGS
	ID	1		76	25		0	42.866	116.372	
	ID	1	153	111	44		1	43.792	111.435	PINCOCK (GREEN CANYON) HOT SPRINGS
	ID	2				W	1	43.658	111.717	ELKHORN WARM SPRING
	ID	3				W	1	43.652	111.698	HAWLEY WARM SPRING
	ID	4	152	120	48		1	43.643	111.686	HEISE HOT SPRING
	ID	5	154	76	25		0	43.427	111.405	
	ID	1	157	89	32		1	43.112	112.169	YANDELL SPRINGS
	ID	2				W	1	43.040	112.006	(WARM SPRING)
	ID	1		111	44		0	43.364	113.781	
	ID	2	147	125	52		0	43.325	113.916	CONDIE HOT SPRINGS
	ID	1	102	120	49		1	43.988	114.798	PIERSON HOT SPRINGS
	ID	2	103	123	51		1	43.984	114.486	(SPRING (HOT))
	ID	3	140	102	38		1	43.802	114.586	RUSSIAN JOHN HOT SPRING
	ID	4	141	99	37		1	43.779	114.539	EASLEY HOT SPRINGS
	ID	5	132			W	1	43.703	114.739	(SPRINGS (HOT))
●	ID	6	142	157	70		1	43.583	114.410	GUYER HOT SPRINGS
	ID	7	133			H	1	43.646	114.814	SKILLERN HOT SPRINGS
	ID	8	143			H	1	43.639	114.488	WARFIELD HOT SPRING
	ID	9	134	145	63		1	43.606	114.947	LIGHTFOOT HOT SPRINGS
	ID	10	135	105	41		1	43.580	114.834	PREIS HOT SPRING
●	ID	11	136	177	81		1	43.561	114.791	WORSWICK HOT SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
●	ID	12	144	116	47		1	43.556	114.413	CLARENDON HOT SPRINGS
●	ID	13	145	145	63		1	43.504	114.356	HAILEY HOT SPRINGS
●	ID	14	138	129	54		1	43.422	114.628	ELK CREEK HOT SPRING
●	ID	15	137	150	66		1	43.382	114.932	WALDROP HOT SPRING
●	ID	16	139	159	71		0	43.292	114.906	BARRON≠S HOT SPRING
●	ID	17	170	148	65		1	43.051	114.952	WHITE ARROW HOT SPRINGS
●	ID	18	171	80	27		1	43.050	114.933	HOT SULPHUR LAKE
	ID	1	82	115	46		1	43.818	115.863	(WARM SPRINGS)
	ID	2	123	130	54		0	43.814	115.045	
	ID	3	120	130	54		0	43.802	115.393	GRANITE CREEK SPRINGS
●	ID	4	119	148	65		0	43.789	115.434	DUTCH FRANK SPRINGS
	ID	5	117			W	0	43.779	115.481	POOL CREEK HOT SPRING
●	ID	6	116	168	76		1	43.754	115.570	NEINMEYER HOT SPRINGS
	ID	7	115			H	1	43.736	115.580	VAUGHN HOT SPRING
	ID	8	113	120	49		1	43.726	115.602	LOFTUS HOT SPRING
	ID	9	112			H	1	43.721	115.615	SMITH CABIN HOT SPRINGS
	ID	10	110	141	61		1	43.697	115.655	SHEEP CREEK BRIDGE HOT SPRING
	ID	11	84	154	67		1	43.669	115.698	TWIN SPRINGS
	ID	12	126	122	49		1	43.638	115.127	(HOT SPRING)
	ID	13	124			W	1	43.613	115.250	(SPRINGS)
	ID	14	125			W	1	43.613	115.158	
	ID	15	127	120	49		1	43.603	115.069	(HOT SPRING)
●	ID	16	129	132	56		1	43.553	115.273	PARADISE HOT SPRINGS
	ID	17	128	138	59		0	43.542	115.285	BRIDGE HOT SPRINGS (8)
	ID	18	131	157	70		1	43.154	115.515	(HOT SPRINGS)
●	ID	19	131A	130	55		1	43.113	115.305	LATTY HOT SPRING
	ID	20				W	1	43.005	115.125	(WARM SPRING)
	ID	21	131B	125	51		0	43.002	115.171	
●	ID	1	66	130	55		1	43.951	116.353	ROYSTONE HOT SPRINGS
	ID	2	159	128	53		1	43.425	116.718	GIVENS HOT SPRINGS
	ID	1		62	17		1	44.150	111.104	LILY PAD LAKE
	ID	2	151			W	0	44.135	111.303	
●	ID	3		105	41		0	44.091	111.459	ASHTON WARM SPRINGS
	ID	1	148	84	29		1	44.253	112.639	(WARM SPRINGS)
	ID	2	150	121	50		1	44.144	112.546	LIDY HOT SPRINGS
	ID	1	65	87	31		1	44.613	113.365	(WARM SPRINGS)
	ID	2	108	84	29		1	44.267	113.450	BARNEY HOT SPRINGS
	ID	1	54			H	1	44.951	114.706	(WARM SPRING)
	ID	2		115	46		1	44.836	114.790	HOSPITAL HOT SPRING
	ID	3	55	121	49		1	44.799	114.806	(HOT SPRING)
	ID	4	49	130	54		1	44.784	114.854	COX HOT SPRINGS
	ID	5		149	64		1	44.730	114.995	SUNFLOWER HOT SPRINGS
	ID	6	63	114	46		1	44.722	114.017	(HOT SPRING)
	ID	7	57			W	1	44.661	114.650	(HOT SPRINGS)
	ID	8	56	190	87		1	44.645	114.738	(HOT SPRINGS)
	ID	9	58	121	50		1	44.626	114.598	SHOWER BATH SPRINGS
	ID	10	105	123	50		1	44.524	114.175	BEARDSLEY HOT SPRINGS
	ID	11	88			W	0	44.511	114.880	

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				TEMPERATURE						
	STATE	COMPILER NUMBER	WARING NUMBER	FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	ID	12	107			W	1	44.383	114.089	(WARM SPRING)
●	ID	13	93	168	76		1	44.269	114.754	SUNBEAM HOT SPRINGS
	ID	14	91			H	0	44.264	114.844	
	ID	15	92			W	0	44.260	114.797	
	ID	16	100	105	41		1	44.254	114.443	SULLIVAN HOT SPRINGS
	ID	17	90			H	0	44.246	114.883	
	ID	18	95	135	57		1	44.243	114.671	ROBINSON BAR RANCH HOT SPRINGS
	ID	19	94	105	41		1	44.222	114.931	STANLEY HOT SPRING
	ID	20	99	121	50		1	44.167	114.625	SLATE CREEK HOT SPRING
	ID	21	101	105	40		1	44.101	114.863	(WARM SPRING)
	ID	1	26	94	35		0	44.900	115.496	
	ID	2	25	116	47		0	44.850	115.691	
	ID	3		157	69		1	44.829	115.213	KWISKWIS HOT SPRING
	ID	4	48	162	72		1	44.812	115.121	(HOT SPRING)
	ID	5		191	88		0	44.796	115.128	
	ID	6	29			H	0	44.754	115.685	
	ID	7		116	46		0	44.731	115.020	
	ID	8	47	138	58		1	44.717	115.016	
	ID	9	46	116	46		1	44.719	115.207	(HOT SPRINGS)
	ID	10	28	120	49		1	44.675	115.944	(HOT SPRING)
●	ID	11	31	138	59		0	44.537	115.689	MOLLY≠S HOT SPRING
	ID	12	30	121	50		0	44.626	115.756	TRAIL CREEK HOT SPRING (9)
	ID	13	44			H	1	44.627	115.196	SHEEPEATER HOT SPRINGS
	ID	14	86			W	0	44.571	115.071	
●	ID	15	32	188	87		1	44.565	115.697	VULCAN HOT SPRINGS
	ID	16	33			H	0	44.547	115.697	
	ID	17	43	110	43		0	44.544	115.312	
	ID	18	42			H	0	44.524	115.280	
	ID	19	41			W	0	44.443	115.234	
	ID	20	39			H	1	44.429	115.763	BULL CREEK HOT SPRINGS
	ID	21	87			W	0	44.399	115.017	
	ID	22	35			H	1	44.399	115.820	(HOT SPRINGS)
	ID	23	36	100	37		1	44.391	115.834	(HOT SPRINGS)
	ID	24	37	90	32		1	44.380	115.843	(HOT SPRINGS)
●	ID	25	38	186	86		1	44.363	115.856	BOILING SPRINGS
	ID	26	40	90	32		1	44.328	115.801	(HOT SPRING)
	ID	27		120	49		1	44.252	115.891	(HOT SPRINGS)
	ID	28	89			H	0	44.203	115.045	
	ID	29	81	100	37		0	44.167	115.188	SACAJAWEA HOT SPRINGS
	ID	30	73	168	76		1	44.153	115.995	(HOT SPRINGS)
●	ID	31	80	184	85		0	44.153	115.310	BONNEVILLE HOT SPRINGS
	ID	32	77	141	61		0	44.075	115.788	
	ID	33	78			W	0	44.074	115.554	HAVEN LODGE (9)
●	ID	34	79	148	65		0	44.071	115.541	KIRKHAM HOT SPRINGS
	ID	35		103	40		1	44.060	115.816	(HOT SPRING)
	ID	36	74	123	51		0	44.055	115.906	
	ID	37	75			H	0	44.051	115.826	
●	ID	38	76	130	55		0	44.046	115.855	

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
●	ID	1	17	109	43		0	44.971	116.199	KRIGBAUM HOT SPRINGS
●	ID	2		132	56		1	44.853	116.444	STARKEY HOT SPRINGS
●	ID	3	19	149	65		1	44.679	115.231	WHITE LICKS HOT SPRING
	ID	4	18			H	1	44.667	116.305	(HOT SPRINGS)
●	ID	5	27	93	34		1	44.641	116.045	HOT CREEK SPRINGS
●	ID	6		157	70		0	44.582	116.631	
●	ID	7		121	50		1	44.531	116.751	
	ID	8	34			H	0	44.515	116.055	
	ID	9		100	38		0	44.510	116.035	
●	ID	10		159	71		0	44.415	116.030	CABARTON HOT SPRINGS
●	ID	11		197	92		0	44.306	116.744	CRANE CREEK
	ID	12		165	74		0	44.212	116.710	COVE CREEK
	ID	13	72			W	0	44.185	116.115	
●	ID	14		175	80		1	44.090	116.047	
●	ID	1		71	22		1	44.305	117.047	WEISER WARM (HOT) SPRINGS
	ID	1	53	175	80		0	45.313	113.839	
	ID	2	59	112	45		1	45.093	113.837	SALMON HOT SPRINGS
●	ID	3	60	125	52		1	45.010	113.608	SHARKEY HOT SPRING
	ID	1	9			H	1	45.857	114.939	(WARM SPRINGS)
	ID	2	50	109	43		1	45.502	114.460	(HOT SPRINGS)
	ID	3	51			W	1	45.343	114.462	OWL CREEK HOT SPRINGS
●	ID	4	52	199	93		1	45.307	114.335	BIG CREEK HOT SPRINGS
●	ID	1	10	130	55		1	45.788	115.199	RED RIVER HOT SPRINGS
	ID	2	11	141	60		1	45.512	115.045	BARTH HOT SPRINGS
●	ID	3	14	112	45		1	45.277	115.914	BURGDORF HOT SPRINGS
	ID	4	22			H	0	45.169	115.808	
	ID	5	23			W	0	45.068	115.827	
	ID	6	24	136	57		0	45.029	115.688	
	ID	1	12	110	43		0	45.429	116.008	
●	ID	2	13	116	47		1	45.415	116.172	RIGGINS HOT SPRINGS
	ID	3	15			H	0	45.151	116.291	
●	ID	4	16	148	65		1	45.039	116.292	ZIM≠S RESORT (YOGHANN) HOT SPRINGS
	ID	1	2	105	41		1	46.465	114.935	COLGATE WARM SPRINGS
	ID	2	3	118	48		1	46.463	114.871	JERRY JOHNSON HOT SPRINGS
	ID	3	7			W	0	46.223	114.709	
	ID	1	1	118	48		1	46.462	115.035	WEIR CREEK HOT SPRINGS
	ID	2	5			H	1	46.316	115.257	STANLEY HOT SPRINGS
	ID	3	6			H	1	46.138	115.090	STUART HOT SPRINGS
	ID	4	8			H	1	46.005	115.021	MARTEN HOT SPRINGS
							MONTANA			
	MT	1				H	1	44.985	111.615	WOLF CREEK HOT SPRINGS
	MT	2				H	1	44.798	111.145	(HOT SPRINGS)
	MT	1	36	139	60		1	45.757	110.256	HUNTERS HOT SPRINGS
	MT	2		76	25		0	45.706	110.975	BRIDGER CANYON SPRINGS (4)
	MT	3	40	70	21		1	45.552	110.142	ANDERSON SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	MT	4	37	107	42		1	45.337	110.692	CHICO HOT SPRINGS
	MT	5	38	148	65		1	45.088	110.771	LADUKE (CORWIN) HOT SPRING
	MT	6	39	89	32		1	45.036	110.666	BEAR CREEK SPRING
	MT	1	35	121	50		1	45.659	111.185	BOZEMAN HOT SPRINGS
	MT	2				W	1	45.603	111.900	(WARM SPRINGS)
	MT	3	31	121	50		1	45.591	111.899	POTOSI HOT SPRINGS
●	MT	4	32	125	52		1	45.573	111.684	NORRIS HOT SPRINGS
●	MT	1	20	141	61		1	45.896	112.233	PIPESTONE HOT SPRINGS
●	MT	2	30	161	72		0	45.690	112.282	BARKELS (SILVER STAR) HOT SPRINGS
	MT	3		127	53		1	45.461	112.474	BILTMORE HOT SPRINGS
	MT	4	27			H	1	45.422	112.693	ZIEGLER HOT SPRINGS
	MT	5	33	109	43		1	45.170	112.158	PULLER HOT SPRINGS
	MT	6	28	72	22		1	45.111	112.713	(WARM SPRING AREA)
	MT	7	29	72	22		1	45.104	112.751	BROWN SPRINGS
	MT	1	14	125	51		1	45.747	113.936	GALLOGLY SPRING
	MT	2	26	120	49		1	45.457	113.105	ELKHORN HOT SPRINGS
●	MT	3	25	136	58		0	45.366	113.410	JARDINE (JACKSON, BIG HOLE) HOT SPRINGS
	MT	1		112	45		1	45.845	114.033	MEDICINE HOT SPRINGS
	MT	2				H	1	45.695	114.379	(HOT SPRING)
	MT	3	13			H	1	45.695	114.359	(HOT SPRINGS)
●	MT	1	24	134	57		1	46.545	110.903	WHITE SULPHUR SPRINGS
●	MT	1	18	138	59		0	46.447	111.985	ALHAMBRA HOT SPRINGS
	MT	1	5	75	24		0	46.505	112.777	
●	MT	2	7	148	65		0	46.598	112.109	HELENA (BROADWATER) HOT SPRINGS
●	MT	3	19	168	76		1	46.199	112.095	BOULDER HOT SPRINGS
	MT	4	15	170	77		1	46.180	112.788	WARM SPRINGS (STATE HOSPITAL)
●	MT	5	17	165	74		1	46.042	112.811	GREGSON HOT SPRINGS
	MT	1	4	111	44		1	46.729	114.530	LOLO (GRANITE) HOT SPRINGS
	MT	2	12	125	52		1	46.104	114.002	SLEEPING CHILD HOT SPRINGS
	MT	1	8	85	30		0	47.996	108.454	BIG WARM SPRINGS
	MT	1	10	69	21		1	47.221	109.474	(WARM SPRING)
	MT	1	6	84	28		1	47.531	112.858	MEDICINE SPRINGS
	MT	1	1	112	45		1	47.607	114.663	CAMAS HOT SPRINGS
	MT	2	3	114	45		1	47.328	114.789	

NEVADA

	NV	1				H	1	35.982	114.748	(HOT SPRINGS)
	NV	2				H	1	35.964	114.743	(HOT SPRING)
	NV	3				H	1	35.944	114.733	(HOT SPRING)
	NV	1		90	32		1	36.722	114.717	MUDDY SPRING
	NV	2	150	90	32		1	36.710	114.714	(WARM SPRINGS)
	NV	1	151	78	25		0	36.564	115.669	INDIAN SPRINGS
	NV	2	141	75	23		0	36.157	115.903	MANSE SPRINGS
	NV	1	138	110	43		1	36.970	116.718	HICKS HOT SPRINGS
	NV	2		97	36		1	36.963	116.725	
	NV	3		88	31		1	36.954	116.726	

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STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
NV	4		82	27		1	36.493	116.342	FAIRBANKS SPRING
NV	5		82	27		0	36.488	116.148	
NV	6		84	28		1	36.479	116.323	ROGERS SPRING
NV	7	139	82	27		1	36.465	116.325	LONGSTREET SPRING
NV	8		94	34		1	36.464	116.314	
NV	9		94	34		1	36.430	116.304	
NV	10		92	33		1	36.427	116.289	DEVILS HOLE
NV	11		91	32		1	36.421	116.323	CRYSTAL POOL
NV	12		80	26		1	36.406	116.304	
NV	13		93	33		1	36.401	116.272	POINT OF ROCKS SPRINGS
NV	14		82	27		1	36.391	116.278	JACK RABBIT SPRING
NV	15		83	28		1	36.375	116.276	BIG SPRING
NV	1	144A	70	21		0	37.861	114.311	DELMUES SPRING
NV	2	145	88	31		1	37.807	114.384	PANACA SPRING
NV	3	144	70	21		1	37.783	114.532	BENNETT SPRINGS
NV	4	146	128	53		0	37.622	114.513	CALIENTE HOT SPRING
NV	1	147	90	32		0	37.597	115.214	HIKO SPRING
NV	2	148	90	32		1	37.531	115.234	CRYSTAL SPRINGS
NV	3	149	97	36		1	37.462	115.188	ASH SPRINGS
NV	1		80	26		0	37.948	117.804	
NV	2		76	24		0	37.913	117.843	
NV	3				H	0	37.852	117.653	
NV	4		110	43		1	37.824	117.484	PEARL HOT SPRINGS
NV	5	112	140	59		1	37.826	117.341	ALKALI HOT SPRING
NV	6	111	160	71		1	37.759	117.631	(HOT SPRING)
NV	1	142	70	21		1	38.688	114.624	GEYSER RANCH SPRINGS
NV	2	142	70	21		1	38.672	114.625	GEYSER RANCH SPRINGS
NV	3	142	70	21		1	38.658	114.630	GEYSER RANCH SPRINGS
NV	1	103A	128	53		1	38.950	115.229	WILLIAMS HOT SPRINGS
NV	2	122			W	1	38.951	115.702	BIG WARM SPRING
NV	3				W	1	38.937	115.695	LITTLE WARM SPRING
NV	4	104	72	22		1	38.924	115.081	PRESTON SPRINGS
NV	5	135	70	21		1	38.622	115.051	EMIGRANT SPRINGS
NV	6	134	100	37		1	38.594	115.140	MOORMAN SPRING
NV	7	128	82	27		1	38.562	115.530	BLUE EAGLE SPRINGS
NV	8	129	73	22		1	38.555	115.532	KATE SPRING
NV	9	126	99	37		1	38.555	115.771	BIG (LOCKES) SPRING
NV	10		99	37		1	38.553	115.761	HAY CORRAL SPRING
NV	11	127	160	71		1	38.465	115.791	CHIMNEY SPRINGS
NV	12	136	75	23		1	38.442	115.014	BUTTERFIELD SPRINGS
NV	13		75	23		1	38.423	115.024	FLAG SPRINGS
NV	14		85	29		1	38.400	115.858	STORM SPRING
NV	15	137	92	33		1	38.381	115.154	HOT CREEK SPRING
NV	16	134A	92	33		1	38.350	115.181	MOON RIVER SPRING
NV	1	121			W	1	38.758	116.435	FISH SPRINGS
NV	2				W	1	38.694	116.436	UPPER WARM SPRING AND (WARM SPRING)
NV	3		92	33		1	38.533	116.464	UPPER WARM SPRING
NV	4		180	82		1	38.520	116.365	(HOT SPRING)

Circ. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	TEMPERATURE CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	NV	5				W	1	38.338	116.660	(WARM SPRING)
	NV	6				W	1	38.253	116.828	(WARM SPRING)
●	NV	7	125	141	61		1	38.187	116.373	WARM SPRINGS
	NV	1	116	80	26		1	38.955	117.049	CHARNOCK (BIG BLUE) SPRINGS
●	NV	2	118	206	97		1	38.820	117.181	DARROUGHS HOT SPRINGS
	NV	3	123			W	0	38.458	117.303	INDIAN SPRINGS
	NV	1	108			W	1	38.965	118.690	DOUBLE SPRING
	NV	2	113	144	62		1	38.921	118.200	WEDELL HOT SPRINGS
	NV	3	109	101	38		0	38.494	118.967	
	NV	4				H	0	38.392	118.105	MINA HOT SPRING (7)
	NV	5	110	101	38		1	38.342	118.106	
●	NV	1	60	159	71		1	38.978	119.833	WALLEYS HOT SPRINGS
●	NV	2	61	141	61		1	38.901	119.413	NEVADA (HINDS) HOT SPRINGS
	NV	3				H	1	38.860	119.175	(HOT SPRING)
	NV	4		184	84		1	38.769	119.174	WILSON HOT SPRING
	NV	1		83	28		0	39.917	114.669	
	NV	2	95	149	65		0	39.891	114.898	JOHN SALVI≠S HOT SPRING (7)
	NV	3	98	174	79		0	39.667	114.809	MONTE NEVA HOT SPRINGS
	NV	4	100	76	24		0	39.547	114.917	
	NV	5	101	83	28		1	39.416	114.780	MCGILL SPRING
	NV	6	102	95	34		1	39.284	114.865	LACKAWANNA SPRINGS
	NV	1	102A	70	21		1	39.817	115.611	MOORES RANCH SPRINGS
	NV	2				H	1	39.422	115.683	SULPHUR SPRING
	NV	3				W	1	39.072	115.635	BIG BLUE SPRING
	NV	1	91A	87	30		1	39.988	116.043	SIRI RANCH SPRING
	NV	2	91B	106	41		1	39.943	116.074	SHIPLEY HOT SPRING
	NV	3	92	181	82		1	39.941	116.681	(HOT SPRINGS)
●	NV	4	93	161	72		1	39.905	116.591	WALTI HOT SPRINGS
	NV	5				H	1	39.893	116.650	LITTLE HOT SPRINGS
	NV	6	91C	74	23		1	39.836	116.067	SULPHUR SPRING
	NV	7	93A	103	40		1	39.558	116.365	(HOT SPRINGS)
	NV	8				W	1	39.529	116.389	(WARM SPRING)
●	NV	9	93B	152	67		1	39.404	116.344	BARTHOLOMAE (CLOBE) HOT SPRINGS
●	NV	10	86	161	72		1	39.328	116.858	SPENCER HOT SPRINGS
	NV	11	119	112	45		1	39.079	116.639	POTT≠S RANCH HOT SPRING (10)
	NV	12		138	59		1	39.030	116.666	DIANAS PUNCH BOWL
	NV	1	70	204	95		0	39.950	117.935	
●	NV	2	85	185	86		1	39.309	117.553	(HOT SPRINGS)
	NV	3	114			H	1	39.026	117.136	(HOT SPRING)
	NV	1				H	0	39.863	118.012	
●	NV	2	71A	161	72		1	39.795	118.064	DIXIE HOT SPRINGS
	NV	3		70	21		1	39.629	118.176	MUD SPRINGS
	NV	4	73	193	90		1	39.565	118.853	
	NV	5	74	178	81		0	39.336	118.630	BORAX SPRING
	NV	6	75			H	1	39.281	118.419	SAND SPRINGS
●	NV	7	74A	190	88		1	39.208	118.721	LEE HOT SPRINGS
	NV	1	52	120	48		0	39.944	119.509	
	NV	2	53			W	1	39.846	119.713	(SPRING) (WARM)

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
●	NV	3	72	208	98		1	39.787	119.009	BRADYS HOT SPRINGS
	NV	4		186	86		0	39.597	119.110	
	NV	5		91	33		0	39.511	119.908	LAWTON HOT SPRINGS
	NV	6	55A			H	1	39.486	119.804	MOANA SPRINGS
	NV	7	55B	94	34		0	39.421	119.767	HUFFAKER SPRINGS
●	NV	8		204	96		1	39.396	119.748	STEAMBOAT SPRINGS
●	NV	9	56	204	96		1	39.384	119.742	STEAMBOAT SPRINGS
	NV	10	57	114	46		0	39.283	119.843	BOWERS MANSION HOT SPRING
	NV	11	59	120	48		1	39.195	119.749	CARSON HOT SPRINGS
	NV	12				H	1	39.167	119.154	(HOT)
●	NV	13	62	206	97		1	39.161	119.183	WABUSKA HOT SPRINGS
	NV	14	59A			W	0	39.159	119.737	NEVADA STATE PRISON
	NV	15				H	1	39.056	119.743	(HOT SPRING)
	NV	16				H	1	39.055	119.809	HOBO HOT SPRING
	NV	1	30C	73	22		1	40.967	114.509	BIG SPRINGS
	NV	2		65	18		1	40.957	114.748	
	NV	3		86	29		1	40.947	114.751	(SPRINGS)
	NV	4	94	92	33		1	40.086	114.643	COLLAR AND ELBOW SPRING
	NV	1				W	0	40.971	115.012	
●	NV	2	32	92	33		1	40.820	115.774	HOT HOLE (ELKO HOT SPRINGS)
	NV	3	33			W	1	40.782	115.361	(WARM SPRING)
	NV	4	34	65	18		1	40.755	115.034	(WARM SPRINGS)
●	NV	5		199	93		1	40.585	115.284	SULPHUR HOT (HOT SULPHUR) SPRINGS
	NV	6	34A	149	65		1	40.253	115.407	(SPRINGS (HOT))
	NV	1	76	83	28		0	40.917	116.906	IZZENHOOD RANCH SPRINGS
	NV	2	31	98	36		1	40.762	116.040	(HOT SPRINGS)
	NV	3	77			W	1	40.745	116.686	WHITE HOUSE SPRING
●	NV	4		174	79		0	40.697	116.130	
	NV	5				W	1	40.682	116.155	(WARM SPRING)
	NV	6		130	55		1	40.671	116.837	(HOT SPRINGS)
	NV	7	88	136	57		0	40.603	116.466	HORSESHOE RANCH SPRINGS
●	NV	8	77A	208	98		1	40.559	116.591	BEOWAWE HOT SPRINGS (THE GEYSERS)
	NV	9				H	0	40.429	116.503	
●	NV	10	88A	129	54		1	40.400	116.517	HOT SPRINGS POINT
	NV	11	89	78	26		1	40.324	116.058	(HOT SPRINGS)
	NV	12		186	85		1	40.318	116.434	(HOT SPRING)
	NV	13	90	102	38		0	40.298	116.057	
	NV	14	99A	152	66		1	40.222	116.065	BRUFFEY≠S HOT SPRINGS
	NV	15	82	102	38		1	40.189	116.805	(SPRING)
	NV	16	91	78	25		1	40.079	116.035	FLYNN RANCH SPRINGS
	NV	1	18	93	33		1	40.990	117.767	(SPRING)
	NV	2	19	165	74		1	40.962	117.491	(HOT SPRING)
●	NV	3	19	114	46		1	40.952	117.491	GOLCONDA HOT SPRING
●	NV	4	19A	135	58		1	40.923	117.108	HOT POT (BLOSSOM HOT SPRINGS)
	NV	5		82	27		1	40.872	117.936	(SPRING)
	NV	6	19F	94	34		1	40.831	117.306	BROOKS SPRING
●	NV	7	19G	184	85		1	40.761	117.491	(HOT SPRINGS)
●	NV	8	64	204	96		1	40.602	117.646	LEACH HOT SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	TEMPERATURE CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	NV	9				H	0	40.416	117.415	
●	NV	10	66	170	77		1	40.405	117.885	KYLE HOT SPRINGS
●	NV	11	78	174	79		1	40.366	117.324	BUFFALO VALLEY HOT SPRINGS
	NV	12	79	110	43		1	40.314	117.067	MOUND SPRINGS
	NV	13	80	127	53		1	40.199	117.104	(HOT SPRINGS)
	NV	14		128	53		1	40.192	117.110	(HOT SPRINGS)
	NV	15		129	54		1	40.182	117.101	
●	NV	16		84	29		1	40.176	117.494	(HOT SPRING)
●	NV	17	68	199	93		1	40.087	117.725	SOU HOT SPRINGS
	NV	18		120	48		1	40.081	117.605	MCCOY SPRINGS
	NV	19		103	40		1	40.035	117.596	(HOT SPRINGS)
	NV	20				H	1	40.031	117.644	
	NV	21	69	175	79		1	40.005	117.720	HYDER HOT SPRINGS
	NV	1	16	134	57		0	40.971	119.007	
●	NV	2		193	90		0	40.950	119.003	
●	NV	3	37	175	80		1	40.858	119.328	FLY RANCH HOT SPRINGS
	NV	4		84	29		0	40.826	119.532	
●	NV	5	63	186	86		1	40.770	119.113	BUTTE SPRINGS (TREGO HOT SPRINGS)
	NV	6	41	83	28		0	40.687	119.594	WALL SPRING
●	NV	7	38	193	90		1	40.664	119.365	GREAT BOILING SPRING (GERLACH HOT SPRINGS
	NV	8	39	166	75		0	40.650	119.375	MUD SPRINGS
	NV	9	50	186	86		1	40.260	119.380	BOILING SPRING
	NV	10		82	27		1	40.220	119.201	
●	NV	11	49	208	98		1	40.145	119.683	(HOT SPRINGS)
	NV	12	48	73	22		1	40.104	119.887	FISH SPRINGS
	NV	1	23	85	30		1	41.978	114.378	(HOT SPRING)
	NV	2				W	1	41.968	114.573	(WARM SPRINGS)
	NV	3	24	109	43		1	41.927	114.071	NILE SPRING
	NV	4	25	110	43		1	41.883	114.116	GAMBLES HOLE
●	NV	5	228	139	60		0	41.791	114.735	MINERAL (SAN JACINTO) HOT SPRINGS
	NV	6		84	28		0	41.567	114.427	
	NV	7				H	0	41.424	114.750	
	NV	8		138	58		1	41.412	114.677	(SPRINGS (HOT))
	NV	9	27	112	45		1	41.382	114.165	(SPRING (HOT))
	NV	10		82	27		0	41.363	114.072	
●	NV	11	30A	141	61		1	41.183	114.991	(HOT SPRINGS)
●	NV	12	30	121	50		1	41.160	114.986	HOT SULPHUR (SULPHUR) SPRINGS
	NV	13				H	1	41.146	114.991	(HOT SPRINGS)
	NV	14				W	1	41.068	114.993	(SPRING)
	NV	1	22	104	39		1	41.877	115.628	(HOT SPRINGS)
	NV	2	21	110	43		1	41.776	115.924	RIZZI RANCH HOT SPRING (7)
	NV	3		129	54		1	41.647	115.768	(HOT SPRINGS)
	NV	4	30D	154	67		1	41.574	115.184	HOT CREEK SPRINGS
	NV	5	28	122	49		1	41.260	115.306	(HOT SPRINGS)
	NV	6				H	1	41.192	115.290	(HOT SPRING)
	NV	7	29			H	0	41.174	115.280	(HOT SPRINGS)
●	NV	1		193	90		1	41.469	116.148	HOT SULPHUR SPRINGS (TUSCARORA)
	NV	2				H	1	41.150	116.734	HOT LAKE

CIRC. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	TEMPERATURE CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
●NV	1	11	136	58		1	41.424	117.391	THE HOT SPRINGS
NV	2				H	0	41.064	117.085	
NV	3	19D	70	21		1	41.031	117.326	
NV	1				W	1	41.932	118.807	(WARM SPRINGS)
●NV	2	2	129	54		1	41.921	118.804	BOG HOT SPRINGS
●NV	3		177	81		1	41.918	118.707	BALTAZOR HOT SPRING
NV	4		112	44		1	41.752	118.840	(WARM SPRINGS)
NV	5		108	42		1	41.726	118.919	WEST SPRING
NV	6		132	56		1	41.718	118.523	
●NV	7		132	56		1	41.717	118.504	HOWARD HOT SPRING
NV	8		80	26		1	41.703	118.298	(SPRING)
●NV	9		150	66		1	41.565	118.562	DYKE HOT SPRINGS
NV	10		104	39		1	41.526	118.570	(SPRING)
●NV	11		199	93		0	41.369	118.791	PINTO HOT SPRINGS (EAST)
●NV	12		195	91		0	41.366	118.810	PINTO HOT SPRINGS (WEST)
NV	13		73	23		1	41.253	118.934	CAIN SPRINGS
NV	14		170	76		1	41.050	118.715	MACFARLAND HOT SPRING
NV	1				W	1	41.790	119.108	(WARM SPRING)
NV	2				W	1	41.745	119.795	(WARM SPRINGS)
NV	3	35A	83	28		1	41.730	119.787	HILL≠S WARM SPRING
NV	4	35C	71	21		1	41.590	119.864	TWIN SPRINGS
NV	5	8	84	29		1	41.395	119.169	SOLDIER MEADOW
NV	6		112	45		0	41.393	119.193	SOLDIER MEADOW
NV	7		129	54		1	41.382	119.189	SOLDIER MEADOW
NV	8		136	58		0	41.370	119.189	SOLDIER MEADOW
NV	9		132	56		0	41.358	119.192	SOLDIER MEADOW
NV	10		109	43		1	41.356	119.220	SOLDIER MEADOW
NV	11		73	23		0	41.355	119.114	
NV	12		127	53		0	41.344	119.192	SOLDIER MEADOW
NV	13		78	26		0	41.337	119.170	SOLDIER MEADOW
NV	14		71	22		0	41.333	119.221	SOLDIER MEADOW
NV	15		102	39		1	41.332	119.199	SOLDIER MEADOW
●NV	16		75	24		1	41.319	119.207	SOLDIER MEADOW
NV	17	36			H	0	41.175	119.957	
NV	18		152	67		0	41.146	119.020	
NV	19		80	26		0	41.137	119.135	
NV	20		121	50		1	41.113	119.004	
●NV	21	12	175	80		1	41.049	119.029	DOUBLE HOT SPRINGS
NV	22		114	46		0	41.029	119.018	
NV	23		161	72		0	41.019	119.015	
NV	24		172	78		0	41.011	119.012	
NV	25		204	96		0	41.003	119.007	

NEW MEXICO

●NM	1	38	125	52		1	32.501	106.926	RADIUM SPRINGS
NM	1		92	33		1	32.795	107.276	DERRY WARM SPRINGS

CIRC. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE						
			FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
NM	2	34	144	62		1	32.753	107.834	MIMBRES HOT SPRINGS
NM	3		66	18		1	32.703	107.759	GOAT SPRING
NM	4		72	22		0	32.693	107.811	
NM	5	36	131	54		1	32.554	107.994	FAYWOOD HOT SPRINGS
NM	1		77	24		0	32.975	108.631	
NM	2		78	25		1	32.884	108.357	ALLEN SPRINGS
NM	3		68	19		1	32.867	108.698	
NM	4		72	22		1	32.816	108.412	ASH SPRING
NM	5	35			W	0	32.639	108.124	APACHE TEJO WARM SPRING
NM	6				W	1	32.562	108.027	KENNECOTT WARM SPRINGS (6)
NM	1		69	20		1	32.899	109.035	
NM	2		69	20		1	32.830	109.044	GOAT CAMP SPRING
NM	1		70	21		1	33.813	106.971	
NM	1		66	18		1	33.911	107.159	SAWMILL SPRING
NM	2	24	83	28		1	33.572	107.600	OJO CALIENTE
NM	3		87	30		1	33.279	107.563	(WARM SPRINGS)
NM	4	37	109	42		0	33.129	107.254	TRUTH OR CONSEQUENCES (LAS PALOMAS) H.SPG
NM	1		70	21		0	33.898	108.501	ARAGON SPRINGS
NM	2		98	36		1	33.829	108.797	(UPPER) FRISLO HOT SPRINGS
NM	3				H	0	33.305	108.247	
NM	4		91	32		0	33.304	108.330	THE MEADOWS (WARM SPRING) (6)
NM	5		81	27		0	33.283	108.264	(NO NAME SEEP) (6)
NM	6		94	34		0	33.261	108.233	(NO NAME SPRING) (6)
● NM	7	25	121	49		1	33.244	108.880	LOWER FRISCO (SAN FRANCISCO) HOT SPRINGS
NM	8				H	1	33.237	108.880	(HOT SPRINGS)
NM	9	27	150	65		0	33.218	108.237	(NO NAME SPRING) (6)
● NM	10	30	154	68		1	33.199	108.206	GILA HOT SPRINGS
NM	11	31	126	52		0	33.192	108.180	LYONS HUNTING LODGE HOT SPRINGS (6)
NM	12	32	113	44		1	33.162	108.209	(SPRING (HOT))
NM	13	29			H	0	33.113	108.486	
NM	1		92	33		0	33.708	109.025	FRIEBORN CANYON SPRING
NM	1		69	20		0	34.995	106.454	CLEAR WATER SPRING
NM	2		70	21		0	34.264	106.883	
NM	3		68	19		1	34.116	106.980	OJITOS SPRINGS
NM	4		70	21		1	34.049	106.939	COOK SPRING
NM	5	23	92	33		1	34.038	106.940	SOCORRO SPRING AND SEDILLO SPRING
NM	6		79	26		1	34.032	106.777	OJO DE LAS CANAS
NM	1		68	19		0	34.916	107.143	
NM	2	22	73	22		1	34.903	107.085	EL OJO ESCONDIDO
NM	3		73	22		0	34.886	107.090	
NM	4		80	26		1	34.854	107.088	(SPRING (SALT)) LAGUNA PUEBLO SPRINGS (4)
NM	5	2	75	24		1	34.847	107.091	(SPRING (SALT))
NM	6	3	82	27		1	34.833	107.091	(SPRING (SALT)) LAGUNA PUEBLO SEEPS (4)
NM	7		68	19		0	34.815	107.388	
NM	8		86	29		1	34.808	107.091	(SPRINGS (SALT))
NM	9		80	26		0	34.791	107.091	
NM	10		78	25		1	34.769	107.085	(SPRING (SALT))
NM	11		71	21		1	34.698	107.129	(SPRING)

CIRC. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
			FAHRENHEIT	CELSIUS					
NM	12		72	22		0	34.326	107.095	
NM	1	21	71	21		0	34.912	108.951	OJO CALIENTE
NM	2		72	22		0	34.158	108.319	
NM	1	20	137	58		1	35.654	105.292	MONTEZUMA (LAS VEGAS) HOT SPRINGS
NM	1	11	101	38		1	35.971	106.561	SAN ANTONIO WARM SPRING
NM	2	10	120	48		1	35.939	105.644	SAN ANTONIO (MURRAY) HOT SPRING
NM	3	12	158	69		1	35.907	106.614	SULPHUR SPRINGS
NM	4		111	43		1	35.848	106.629	SPENCE HOT SPRING
NM	5	14	90	32		1	35.821	106.628	MCCAULEY HOT SPRING
NM	6	13	118	47		1	35.793	106.685	SODA DAM HOT SPRINGS
●NM	7	15	163	73		1	35.769	106.691	JEMEZ SPRINGS
NM	8	16	70	21		1	35.601	106.860	PENASCO (PHILLIP S) SPRINGS
NM	9	17	123	50		0	35.592	106.753	INDIAN SPRINGS
NM	10	19	68	19		1	35.548	106.827	SAN YSIDRO WARM SPRINGS
NM	11	18	85	29		0	35.540	106.854	SAN YSIDRO HOT SPRINGS
NM	12		68	19		0	35.308	106.471	
NM	1		72	22		1	35.060	107.133	ALAMOS SPRING
NM	1		99	37		0	36.523	105.713	(NO NAME SPRING) (6)
NM	2		100	37		1	36.508	105.722	MANBY (MAMBY S,AMERICAN) HOT SPRINGS
NM	3		95	34		1	36.324	105.606	PONCE DE LEON SPRINGS (HOT SPRING)
NM	1		83	28		1	36.368	106.059	STATUE SPRING (6)
NM	2	8	113	44		1	36.305	106.053	OJO CALIENTE (JOSEPH S HOT SPRINGS)
NM	3				H	0	36.246	105.826	AGUA CALIENTE (6)
							OREGON		
OR	1	85	120	49		0	42.977	117.061	CANTER S HOT SPRING
OR	2	85C	93	34		0	42.534	117.181	
OR	3	86	125	52		1	42.078	117.760	
OR	1	65	80	27		1	42.891	118.897	HOGHOUSE HOT SPRINGS
OR	2	66	82	28		0	42.837	118.859	
OR	3	67	89	32		1	42.823	118.901	(WARM SPRINGS)
●OR	4		206	97		1	42.677	118.345	MICKEY SPRINGS
●OR	5	68	168	76		1	42.544	118.530	ALVORD HOT SPRINGS
OR	6	69	204	96		1	42.340	118.599	(HOT SPRINGS)
●OR	7	70	96	36		1	42.328	118.602	BORAX LAKE (HOT LAKE)
OR	8	71	100	37		1	42.254	118.311	
OR	9	72	125	52		0	42.190	118.383	
OR	1	48	114	46		0	42.543	119.672	
OR	2	48A	103	40		1	42.501	119.693	ANTELOPE HOT SPRINGS
OR	3	49			H	1	42.472	119.710	HART MOUNTAIN HOT SPRINGS
OR	4		71	22		1	42.308	119.876	MOSS RANCH
●OR	5	49A	154	68		1	42.300	119.778	FISHER HOT SPRINGS
OR	6	49B	82	28		1	42.290	119.869	MOSS RANCH
●OR	7	49C	172	78		1	42.230	119.884	CRUMP (CHARLES CRUMP S SPRING)
OR	8	49D	197	92		0	42.220	119.877	WARNER VALLEY RANCH
OR	9	50	159	71		0	42.175	119.858	ADEL HOT SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
				FAHRENHEIT	CELSIUS					
	OR	10	51	159	71		0	42.078	119.882	HOUSTON HOT SPRINGS
	OR	11	50A	112	45		1	42.077	119.933	HALLINAN
	OR	12		159	71		0	42.073	119.922	HALLINAN
	OR	1	37	66	19		0	42.998	120.770	ANA RIVER SPRING
	OR	2		64	18		0	42.997	120.653	
	OR	3	38	67	20		0	42.990	120.728	BUCKHORN CREEK SPRINGS
	OR	4	40A	66	19		0	42.980	120.780	
	OR	5		66	19		0	42.960	120.655	
	OR	6	40	66	19		0	42.952	120.639	THOUSAND SPRINGS
	OR	7	40E	67	20		0	42.929	120.645	LOST CABIN SPRING
	OR	8	41	75	24		0	42.928	120.801	PARDON WARM SPRING
●	OR	9	42	109	43		1	42.725	120.647	SUMMER LAKE HOT SPRING
	OR	10	44E	67	20		1	42.380	120.331	BEAN HOT SPRING
	OR	11	44D	67	20		1	42.324	120.327	WHITE ROCK RANCH HOT SPRING
	OR	12	31			H	1	42.266	120.991	ROBINSON SPRING
●	OR	13	45	204	96		1	42.220	120.367	HUNTERS HOT SPRINGS
	OR	14	46	161	72		1	42.160	120.343	LEITHEAD HOT SPRINGS (JOYLAND PLUNGE)
	OR	15	47	190	88		1	42.153	120.346	BARRY RANCH HOT SPRINGS
	OR	1		94	34		0	42.420	121.950	
	OR	2		70	21		0	42.185	121.823	
	OR	3		70	21		0	42.181	121.807	HARDBOARD SPRING (3)
	OR	4	28	165	74		0	42.174	121.617	OLENE GAP HOT SPRINGS
	OR	5	28A	66	19		0	42.168	121.574	HIGH BROS (TAYLOR) WARM SPRING (3)
	OR	6	28B	78	26		0	42.158	121.629	CRYSTAL SPRING
	OR	7	30	75	24		1	42.132	121.218	WILKERSON S HOT SPRING
	OR	8	29	141	61		1	42.115	121.287	BIG HOT SPRING (OREGON HOT SPRINGS)
	OR	1	25	103	40		0	42.211	122.729	JACKSON HOT SPRINGS
●	OR	1	77	206	97		0	43.983	117.232	VALE HOT SPRINGS
●	OR	2	76	157	70		0	43.894	117.503	
	OR	3	79	143	62		1	43.763	117.160	MITCHELL BUTTE HOT SPRING
	OR	4	80	195	90		1	43.738	117.178	DEER BUTTE HOT SPRING
	OR	5				H	1	43.728	117.203	SNIVELY HOT SPRING
	OR	6	82	71	22		0	43.701	117.191	S BLACK WILLOW SPRING
	OR	7		105	41		1	43.299	117.386	(HOT SPRING)
	OR	8	84A			H	0	43.212	117.514	
●	OR	1	74	139	60		1	43.943	118.138	BEULAH HOT SPRINGS
	OR	2				W	1	43.775	118.043	(WARM SPRING)
	OR	3	51A	71	22		0	43.662	118.738	
	OR	4	54	143	62		0	43.643	118.255	
●	OR	5	84	145	63		0	43.467	118.200	(HOT SPRINGS)
●	OR	6	53	172	78		1	43.440	118.641	CRANE HOT SPRINGS
	OR	7	55	107	42		1	43.393	118.307	(HOT SPRINGS)
	OR	1		84	29		1	43.950	119.635	(WARM SPRINGS)
	OR	2	52	82	28		1	43.538	119.084	MILLPOND SPRING
	OR	3	52B			W	1	43.531	119.079	GOODMAN SPRING
	OR	4	52D	71	22		1	43.501	119.093	ROADLAND SPRING
	OR	5	52E	69	21		0	43.481	119.077	BAKER SPRING
	OR	6		71	22		0	43.274	119.313	DOUBLE O RANCH

CIRC. 726

STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
			FAHRENHEIT	CELSIUS					
OR	7	58	73	23		0	43.272	119.346	DOUBLE O SPRING
OR	8	50	73	23		0	43.267	119.295	BASQUE SPRING
OR	9	63A	69	21		0	43.260	119.019	DUNN SPRING
OR	10	61	71	22		0	43.261	119.279	JOHNSON SPRING
OR	11	62	67	20		0	43.249	119.257	HUGHET SPRING
OR	12		69	21		0	43.235	119.057	
OR	13	62D	108	42		0	43.209	113.141	
OR	14		92	33		0	43.205	119.060	
OR	15	62D	108	42		0	43.196	119.131	
OR	16	64	154	68		1	43.177	119.060	
OR	1		87	30		0	43.877	120.026	
OR	1	32	69	21		0	43.731	121.255	PAULINA SPRINGS
OR	2	33	141	61		0	43.717	121.203	EAST LAKE HOT SPRINGS
OR	1	21	98	37		1	43.809	122.305	WALL CREEK HOT SPRINGS
OR	2	22	163	73		1	43.710	122.292	MCCREDIE SPRINGS
OR	3	23	114	46		1	43.689	122.375	KITSON SPRINGS
OR	4	26			W	0	43.451	122.143	
OR	5	24	105	40		0	43.294	122.367	UMPQUA WARM SPRINGS
● OR	1	17A	136	58		1	44.930	117.938	RADIUM HOT SPRINGS
OR	2	17B	80	27		1	44.778	117.809	SAM O SPRING
OR	3	17	80	27		0	44.550	117.425	NELSON SPRING
OR	4	73			H	0	44.201	117.466	JAMIESON HOT SPRINGS
OR	5	78	164	73		0	44.041	117.022	MALHEUR BUTTE SPRINGS
● OR	6	75	188	87		1	44.022	117.462	NEAL HOT SPRINGS
OR	1	14	120	49		0	44.653	118.831	HOT SULPHUR SPRING
OR	2		69	21		1	44.373	118.739	LIMEKILN SPRING (2)
● OR	3	16	136	58		1	44.356	118.575	BLUE MOUNTAIN HOT SPRINGS
OR	4	15	102	39		1	44.286	118.964	JOAQUIN MILLER RESORT
● OR	1	13	105	41		1	44.894	119.142	RITTER HOT SPRINGS
OR	2				H	1	44.448	119.109	MT VERNON HOT SPRINGS
OR	3		71	22		0	44.071	119.427	BRISBOIS RANCH SPRINGS (2)
● OR	4		114	46		0	44.002	119.655	WEBERG HOT SPRING
OR	1	7	125	52		1	44.866	121.224	(SPRINGS)
● OR	2	7	125	52		1	44.860	121.201	KAHNEETA HOT SPRINGS
● OR	3	6	197	92		1	44.785	121.972	BREITENBUSH HOT SPRINGS
OR	1	5	136	58		1	44.934	122.175	BAGBY HOT SPRINGS
● OR	2	18	159	71		1	44.193	122.049	BELKNAP HOT SPRINGS
OR	3	19	174	79		1	44.154	122.100	FOLEY SPRINGS
OR	4	20	111	44		0	44.103	122.240	COUGAR RESERVOIR HOT SPRING
OR	1	10			W	0	45.503	117.971	
OR	2		84	29		0	45.295	117.805	COVE SPRINGS
● OR	3	11	175	80		1	45.243	117.965	HOT LAKE
OR	4				H	1	45.202	117.910	
● OR	5	12	139	60		1	45.019	117.624	MEDICAL HOT SPRINGS
OR	1		93	34		1	45.741	118.232	BINGHAM SPRINGS
OR	2	8	141	61		1	45.150	118.661	LEHMAN HOT SPRINGS
OR	3	9	100	38		1	45.125	118.732	HIDAWAY SPRINGS
OR	4		82	28		1	45.060	118.456	WARM MINERAL SPRING

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	TEMPERATURE CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	OR	1					1	45.631	119.704	
●	OR	1	1	193	90		0	45.373	121.697	MOUNT HOOD FUMAROLE
●	OR	2	2	80	27		1	45.293	121.731	MOUNT HOOD WARM SPRINGS
●	OR	1	4	186	86		1	45.021	122.011	AUSTIN (CAREY) HOT SPRINGS
								SOUTH DAKOTA		
	SD	1	4	72	22		0	43.526	103.376	MARTIN VALLEY (BUFFALO GAP) SPRINGS
	SD	2	2	90	32		1	43.447	103.509	(SPRINGS)
	SD	3	1	89	31		1	43.438	103.483	HOT SPRINGS
	SD	4	3	68	19		1	43.333	103.551	CASCADE SPRINGS
								TEXAS		
	TX	1		105	40		1	29.183	102.994	(HOT SPRINGS)
	TX	1	3	114	45		1	30.036	104.598	HOT SPRINGS
	TX	1	1	100	37		1	30.859	105.342	RED BULL SPRING
	TX	2	2	126	52		1	30.822	105.311	INDIAN HOT SPRINGS
								UTAH		
	UT	1	57	91	32		1	37.700	110.421	(WARM SPRING)
	UT	1		98	36		0	37.329	113.687	VEYO HOT SPRING
	UT	2	54	107	42		1	37.189	113.269	DIXIE (LAVERKIN) HOT SPRINGS
	UT	1	37	71	22		0	38.998	111.870	REDMOND SPRINGS (LAKE)
●	UT	1	28	105	41		1	38.864	112.508	MEADOW HOT SPRINGS
	UT	2		100	37		1	38.849	112.492	HATTON (BLACK ROCK) HOT SPRINGS
	UT	3		73	23		0	38.776	112.096	RICHFIELD WARM (HOT) SPRINGS
	UT	4		168	76		1	38.638	112.100	RED HILL HOT SPRING
●	UT	5	48	147	64		1	38.632	112.106	MONROE (COOPER) HOT SPRINGS
●	UT	6	49	147	64		1	38.611	112.199	JOSEPH HOT SPRINGS
	UT	7	47	77	24		1	38.602	112.108	JOHNSON WARM SPRING
	UT	8				W	0	38.588	112.554	(WARM VAPOR)
●	UT	9	51	190	88		0	38.496	112.857	ROOSEVELT (MCKEANS) HOT SPRINGS
	UT	10	53	96	36		1	38.211	112.909	RADIUM (DOTSONS) WARM SPRINGS
●	UT	1	52	193	90		1	38.173	113.201	THERMO HOT SPRINGS
	UT	1	17	70	21		1	39.955	111.855	GOSHEN WARM SPRINGS
	UT	2	31	73	22		0	39.246	111.644	LIVINGSTON WARM SPRINGS
	UT	3	35	72	22		1	39.180	111.693	STERLING (NINEMILE) WARM SPRING
	UT	1		74	23		1	39.614	112.805	FUMAROLE BUTTE
●	UT	2	24	188	87		1	39.611	112.727	BAKER (ABRAHAM, CRATER) HOT SPRINGS
	UT	1	20	168	75		1	39.905	113.428	WILSON HOT SPRINGS
	UT	2	21	82	27		1	39.884	113.408	BIG SPRING (NORTH SPRINGS)
	UT	3	22	82	27		1	39.841	113.394	FISH SPRINGS
	UT	4	25	82	27		0	39.456	113.998	GANDY WARM SPRINGS

CIRC. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	 CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
	UT	1	19A	86	29		1	40.465	109.219	SPLIT MOUNTAIN WARM SPRINGS
	UT	1	11	132	55		1	40.815	111.915	BECKS HOT SPRINGS
	UT	2	12	105	40		0	40.790	111.895	WASATCH HOT SPRINGS
	UT	3		115	46		1	40.532	111.478	WARM DITCH SPRING
	UT	4	14	100	38		1	40.528	111.490	MIDWAY HOT SPRINGS (SCHNEITTER S HOT POTS
	UT	5	14B	104	39		1	40.524	111.468	MIDWAY HOT SPRINGS (BUHLER S SPRINGS)
	UT	6	14A	115	46		1	40.516	111.475	MIDWAY HOT SPRINGS (LUKE S HOT POTS)
●	UT	7	13	136	58		0	40.487	111.911	CRYSTAL HOT SPRINGS
	UT	8				W	0	40.353	111.895	CRATER HOT SPRINGS
	UT	9	15	111	43		1	40.348	111.908	SARATOGA HOT SPRINGS
	UT	10		77	24		0	40.240	111.865	(WARM SPRINGS)
	UT	11		75	23		0	40.232	111.868	(WARM SPRINGS)
	UT	12		86	29		0	40.177	111.801	
	UT	13	16	89	31		0	40.144	111.807	LINCOLN POINT WARM SPRINGS
	UT	14	19	67	20		1	40.118	111.341	DIAMOND FORK WARM SPRINGS
	UT	15	18	111	44		1	40.041	111.530	CASTILLA SPRINGS
	UT	1	9			W	1	40.733	112.647	BIG WARM SPRINGS
	UT	2				W	1	40.667	112.677	BURNT SPRINGS
	UT	3	10	90	32		1	40.644	112.523	GRANTSVILLE WARM SPRINGS
	UT	4					1	40.638	112.695	MUSKRAT SPRING
	UT	5				W	1	40.513	112.712	HORSESHOE SPRINGS
	UT	6				W	1	40.559	112.741	IOSEPA (DESERET) SPRINGS
	UT	7	10A	80	26		1	40.397	112.418	MORGANS WARM SPRINGS
	UT	8	10B	80	27		1	40.388	112.424	RUSSELLS WARM SPRINGS
	UT	1	8	135	57		0	41.232	111.929	OGDEN HOT SPRINGS
	UT	2		77	24		1	41.035	111.658	COMO WARM SPRINGS
	UT	1	3	110	43		1	41.855	112.157	UDDY HOT SPRINGS
	UT	2	2	86	29		1	41.832	112.455	BLUE CREEK SPRING (BLUE WARM SPRINGS)
	UT	3		80	26		0	41.836	112.056	CUTLER WARM SPRINGS
	UT	4		75	23		0	41.722	112.266	BOTHWELL WARM SPRINGS
	UT	5	4	134	57		1	41.659	112.086	CRYSTAL (MADSENS) SPRINGS
	UT	6	4A	90	32		0	41.584	112.240	LITTLE MOUNTAIN WARM SPRING
	UT	7		124	51		1	41.579	112.232	STINKING SPRINGS
	UT	8	6	137	58		1	41.339	112.032	UTAH (BEAR RIVER) HOT SPRINGS
	UT	9	5	84	28		1	41.233	112.415	(SPRING)
●	UT	10		139	60		1	41.138	112.169	HOOPER HOT SPRINGS
	UT	1	1	80	26		1	41.756	113.602	(WARM SPRING)
	UT	2		107	41		1	41.687	113.986	(SPRING (HOT))

WASHINGTON

	WA	1		100	37		1	45.823	121.116	KLICKITAT SPRINGS
	WA	2		120	49		1	45.728	121.800	ST MARTINS HOT SPRINGS
	WA	3				H	1	45.723	121.927	GRAYS HOT SPRINGS
	WA	4	16	89	32		1	45.658	121.962	MOFFETTS HOT SPRINGS
●	WA	1	12	90	32		1	46.452	120.959	SODA SPRINGS
●	WA	1		69	21		1	46.752	121.814	LONGMIRE

Circ. 726	STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
●	WA	2	11	120	49		1	46.738	121.562	OHANAPECOSH HOT SPRINGS
	WA	3		55	13		0	46.703	121.483	SUMMIT CREEK (SODA)
	WA	4	12B			H	1	46.202	121.492	MOUNT ADAMS CRATER
	WA	5	14	76	24		1	46.005	121.191	(SPRINGS)
	WA	1	12A	190	87		0	46.215	122.188	MT ST HELENS FUMAROLE
●	WA	1	6	100	38		1	47.892	121.342	GARLAND MINERAL SPRINGS
	WA	2	7	122	49		0	47.710	121.152	SCENIC (GREAT NORTHERN) HOT SPRINGS
	WA	3	8	127	52		1	47.484	121.391	GOLDMEYER HOT SPRINGS
	WA	4	9	122	49		1	47.201	121.536	(HOT SPRINGS)
●	WA	1	3	125	52		1	47.977	123.682	OLYMPIC HOT SPRINGS
●	WA	2	2	132	56		1	47.969	123.864	SOL DUC HOT SPRINGS
●	WA	1	1	107	42		1	48.763	121.667	BAKER HOT SPRING
●	WA	2	4	98	37		1	48.254	121.170	SULPHUR HOT SPRINGS
●	WA	3		139	60		0	48.172	121.039	GAMMA HOT SPRING
●	WA	4	5	109	43		1	48.118	121.192	KENNEDY HOT SPRING

WYOMING

	WY	1	115	120	48		1	41.449	106.802	SARATOGA HOT SPRINGS
	WY	1	116	69	21		1	42.249	104.779	IMMIGRANTS WASH TUB (WARM SPRINGS)
	WY	1	114	85	30		1	42.663	105.396	DOUGLAS WARM SPRINGS (3)
	WY	1	113	129	54		0	42.545	106.723	ALCOVA HOT SPRINGS
	WY	1	112	75	24		1	42.703	107.103	HORSE CREEK SPRINGS
	WY	1		60	16		0	42.813	108.033	CONANT (3)
	WY	2	110	89	32		1	42.491	108.171	(WARM SPRINGS)
	WY	1	105	102	39		1	42.747	109.618	STEELE HOT SPRINGS
●	WY	1	103	143	62		0	42.827	110.997	AUBURN WARM SPRINGS
	WY	2		114	46		0	42.817	110.993	JOHNSON SPRINGS (3)
	WY	3		60	16		0	42.395	110.507	BIG FALL CREEK (3)
	WY	1	111	132	56		1	43.654	108.195	THERMOPOLIS (BIG HORN) HOT SPRINGS
	WY	2		71	22		1	43.582	108.209	WIND RIVER CANYON (3)
	WY	3	108	111	44		1	43.009	108.837	WASHAKIE MINERAL HOT SPRINGS
	WY	1	106	84	29		1	43.560	109.729	WARM SPRING (3)
	WY	2	107	76	25		0	43.520	109.670	LITTLE WARM SPRING (3)
	WY	1		121	50		0	43.963	110.693	JACKSON LAKE HOT SPRINGS
	WY	2		112	45		0	43.910	110.189	NORTH BUFFALO FORK (3)
	WY	3		80	27		1	43.637	110.614	KELLY WARM SPRING
	WY	4		64	18		0	43.624	110.605	TETON VALLEY WARM SPRING
	WY	5		80	27		1	43.545	110.739	ABERCROMBIE WARM SPRING
	WY	6		85	30		0	43.471	110.833	BOYLES HILL SPRING (3)
	WY	7	102	112	45		1	43.367	110.444	GRANITE HOT SPRINGS AND GRANITE FALLS (3)
	WY	8	101	98	37		1	43.296	110.774	ASTORIA MINERAL HOT SPRINGS
	WY	9	104	85	30		1	43.282	110.020	KENDALL WARM SPRING
	WY	1	98	67	20		0	44.735	108.188	LITTLE SHEEP MOUNTAIN WARM SPRINGS
	WY	2	99	69	21		0	44.612	108.136	SHEEP MOUNTAIN WARM SPRINGS
	WY	1	97	96	36		1	44.512	109.115	DEMARIS (CODY) HOT SPRINGS
	WY	2				W	0	44.493	109.204	BUFFALO BILL RESERVOIR (3)

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STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE						
			FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
WY	1	1	136	58		1	44.981	110.688	HOT RIVER
WY	2	2	165	74		1	44.967	110.708	MAMMOTH HOT SPRINGS
WY	3	65A	156	68		1	44.906	110.395	CALCITE SPRINGS
WY	4	66			H	1	44.841	110.167	(HOT SPRINGS)
WY	5	3			H	1	44.825	110.673	(GAS VENTS)
WY	6	67			H	1	44.823	110.114	WAHB SPRINGS
WY	7	7	195	91		1	44.796	110.725	AMPHITHEATER SPRINGS
WY	8	7A	198	92		1	44.787	110.735	CLEARWATER SPRINGS AND SEMI-CENTENNIAL GE
WY	9	8	156	68		1	44.779	110.697	WHITEROCK SPRINGS
WY	10				H	1	44.779	110.737	(STEAM VENTS)
WY	11	68			H	1	44.771	110.266	RAINBOW HOT SPRINGS
WY	12				H	1	44.767	110.303	(HOT SPRINGS)
WY	13	50	198	92		1	44.762	110.430	WASHBURN HOT SPRINGS AND SULPHUR CREEK H
WY	14	73			H	1	44.757	110.305	COFFEE POT HOT SPRINGS
WY	15	9	184	84		1	44.757	110.733	BIJAH SPRING
WY	16	69			H	1	44.751	110.254	(HOT SPRING)
WY	17	52			H	1	44.750	110.409	(HOT SPRINGS)
WY	18	9A			H	1	44.749	110.712	FRYING PAN SPRING AND (GAS VENT)
WY	19	75	195	91		1	44.746	110.225	HOT SPRING BASIN GROUP
WY	20				H	1	44.740	110.699	(GAS VENTS)
WY	21	72	198	92		1	44.740	110.329	WHISTLER GEYSER AND JOSEPHS COAT SPRINGS
WY	22	75			H	1	44.741	110.256	HOT SPRINGS BASIN GROUP
WY	23	76			H	1	44.738	110.035	(HOT SPRINGS)
WY	24	11	195	91		1	44.726	110.706	NORRIS GEYSER BASIN
WY	25	78			H	1	44.721	110.355	(HOT SPRINGS)
WY	26	11	188	87		1	44.716	110.702	ECHINUS GEYSER (NORRIS GEYSER BASIN)
WY	27	55			H	1	44.714	110.557	(GAS VENTS)
WY	28	54			H	1	44.712	110.475	FOREST SPRINGS
WY	29	77			H	1	44.702	110.034	(HOT SPRINGS)
WY	30	79			H	1	44.696	110.378	(HOT SPRINGS)
WY	31	12	190	87		1	44.696	110.758	SYLVAN SPRINGS
WY	32	13	198	92		1	44.695	110.722	GIBBON HILL GEYSER
WY	33	14	199	92		1	44.689	110.738	ARTISTS PAINTPOTS
WY	34	81			H	1	44.687	110.327	(HOT SPRINGS)
WY	35	15	201	94		1	44.685	110.727	GEYSER SPRINGS GROUP
WY	36	16	197	91		1	44.683	110.753	MONUMENT GEYSER BASIN
WY	37	16A	197	91		1	44.675	110.750	BERYL SPRING
WY	38	83	180	82		1	44.673	110.295	PONUNTPA SPRINGS GROUP
WY	39				H	1	44.673	110.237	(HOT SPRINGS)
WY	40	56			H	1	44.656	110.570	VIOLET SPRINGS
WY	41	61	194	89		1	44.648	110.489	SULPHUR SPRINGS
WY	42	87			H	1	44.634	110.231	THE MUDKETTLES AND THE MUSHPOTS
WY	43	61B	193	90		1	44.622	110.436	SULPHUR CAULDRON, MUD VOLCANO, AND OTHERS
WY	44	57			H	1	44.615	110.617	HIGHLAND HOT SPRINGS
WY	45	59			H	1	44.604	110.617	(HOT SPRINGS)
WY	46				H	1	44.598	110.630	(GAS VENT)
WY	47				H	1	44.598	110.235	(HOT SPRINGS)
WY	48	92			H	1	44.598	110.211	PELICAN SPRINGS

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STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE		WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
			FAHRENHEIT	CELSIUS					
WY	49	89	196	91		1	44.594	110.319	(HOT SPRINGS)
WY	50				H	1	44.592	110.619	(GAS VENTS)
WY	51	90			H	1	44.587	110.332	EBRO SPRINGS
WY	52	91			H	1	44.582	110.312	VERMILLION SPRINGS
WY	53	19	204	95		1	44.571	110.805	MORNING MIST SPRINGS AND QUAGMIRE GROUP
WY	54	27			H	1	44.570	110.691	(HOT SPRINGS)
WY	55				H	1	44.568	110.851	FLAT CONE SPRING
WY	56	18	203	94		1	44.565	110.831	RIVER GROUP SPRINGS
WY	57	17	160	71		1	44.563	110.871	RED TERRACE SPRING AND QUEENS LAUNDRY
WY	58				H	1	44.557	110.847	
WY	59		202	95		1	44.556	110.813	FOUNTAIN GROUP
WY	60	93			H	1	44.554	110.301	BEACH HOT SPRINGS
WY	61	21	199	93		1	44.547	110.807	FOUNTAIN PAINT POT
WY	62	94	197	92		1	44.545	110.257	TURBID SPRINGS
WY	63	20	202	94		1	44.543	110.859	FAIRY SPRINGS
WY	64		201	94		1	44.543	110.790	HOT LAKE AND FIREHOLE LAKE
WY	65	22	204	95		1	44.534	110.798	WHITE DOME GEYSER AND GREAT FOUNTAIN GEY.
WY	66	95	198	92		1	44.529	110.296	STEAMBOAT SPRINGS
WY	67	24	202	94		1	44.529	110.878	IMPERIAL GEYSER AND SPRAY GEYSER
WY	68	25	199	93		1	44.523	110.837	EXCELSIOR GEYSER CRATER AND GRAND PRISMAT
WY	69	26A	201	93		1	44.520	110.814	(HOT SPRINGS) (RABBIT CREEK AREA)
WY	70	96	190	87		1	44.519	110.275	BUTTE SPRINGS
WY	71	29	202	95		1	44.483	110.854	SAPPHIRE POOL AND OTHERS
WY	72		192	89		1	44.473	110.863	HILLSIDE SPRINGS
WY	73	30A	205	96		1	44.472	110.844	MORNING GLORY POOL, GROTTO GEYSER, AND OT
WY	74	34	199	92		1	44.458	110.825	OLD FAITHFUL GEYSER AND OTHERS
WY	75	33A	200	93		1	44.461	110.839	CASTLE GEYSER (BLACK SAND BASIN)
WY	76	33	158	69		1	44.458	110.852	EMERALD POOL AND OTHERS
WY	77				H	1	44.436	110.974	(HOT SPRINGS)
WY	78	63	200	93		1	44.432	110.578	POTTS HOT SPRING BASIN
WY	79				H	1	44.432	110.809	(HOT SPRINGS)
WY	80	35			H	1	44.423	110.952	SMOKE JUMPER HOT SPRINGS
WY	81	37			H	1	44.418	110.802	LONE STAR GEYSER
WY	82	64	200	93		1	44.415	110.569	WEST THUMB GEYSER BASIN
WY	83				H	1	44.407	110.821	(HOT SPRINGS)
WY	84	36			H	1	44.403	110.937	(HOT SPRINGS)
WY	85	38	202	95		1	44.351	110.796	SHOSHONE GEYSER BASIN
WY	86	42	199	93		1	44.313	110.656	(HOT SPRINGS)
WY	87	44A	201	94		1	44.305	110.530	(HOT SPRINGS)
WY	88				H	1	44.289	110.504	(HOT SPRINGS)
WY	89	40			H	1	44.291	110.888	THREE RIVER JUNCTION SPRINGS
WY	90	39			H	1	44.282	110.899	BECHLER RIVER HOT SPRINGS
WY	91	41			H	1	44.283	110.879	TENDOY FALLS SPRINGS
WY	92	45	195	91		1	44.279	110.506	RUSTIC GEYSER
WY	93	43	154	67		1	44.274	110.640	(HOT SPRINGS)
WY	94	48			H	1	44.203	110.485	(HOT SPRINGS)
WY	95		141	61		1	44.180	110.727	(HOT SPRINGS)
WY	96	47			B	1	44.165	110.588	SNAKE RIVER (SNAKE) HOT SPRINGS (5)

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STATE	COMPILER NUMBER	WARING NUMBER	TEMPERATURE FAHRENHEIT	CELSIUS	WARM, HOT, BOILING	LOCATED ON MAP	LATITUDE (N)	LONGITUDE (W)	NAME OF SPRING
WY	97				H	1	44.155	110.698	(HOT SPRINGS)
WY	98	46			B	1	44.141	110.656	SOUTH ENTRANCE HOT SPRINGS (5)
● WY	99	100	159	71		1	44.113	110.684	HUCKLEBERRY (FLAGG RANCH) HOT SPRINGS (5)
WY	1				H	1	44.245	111.022	(HOT SPRINGS)

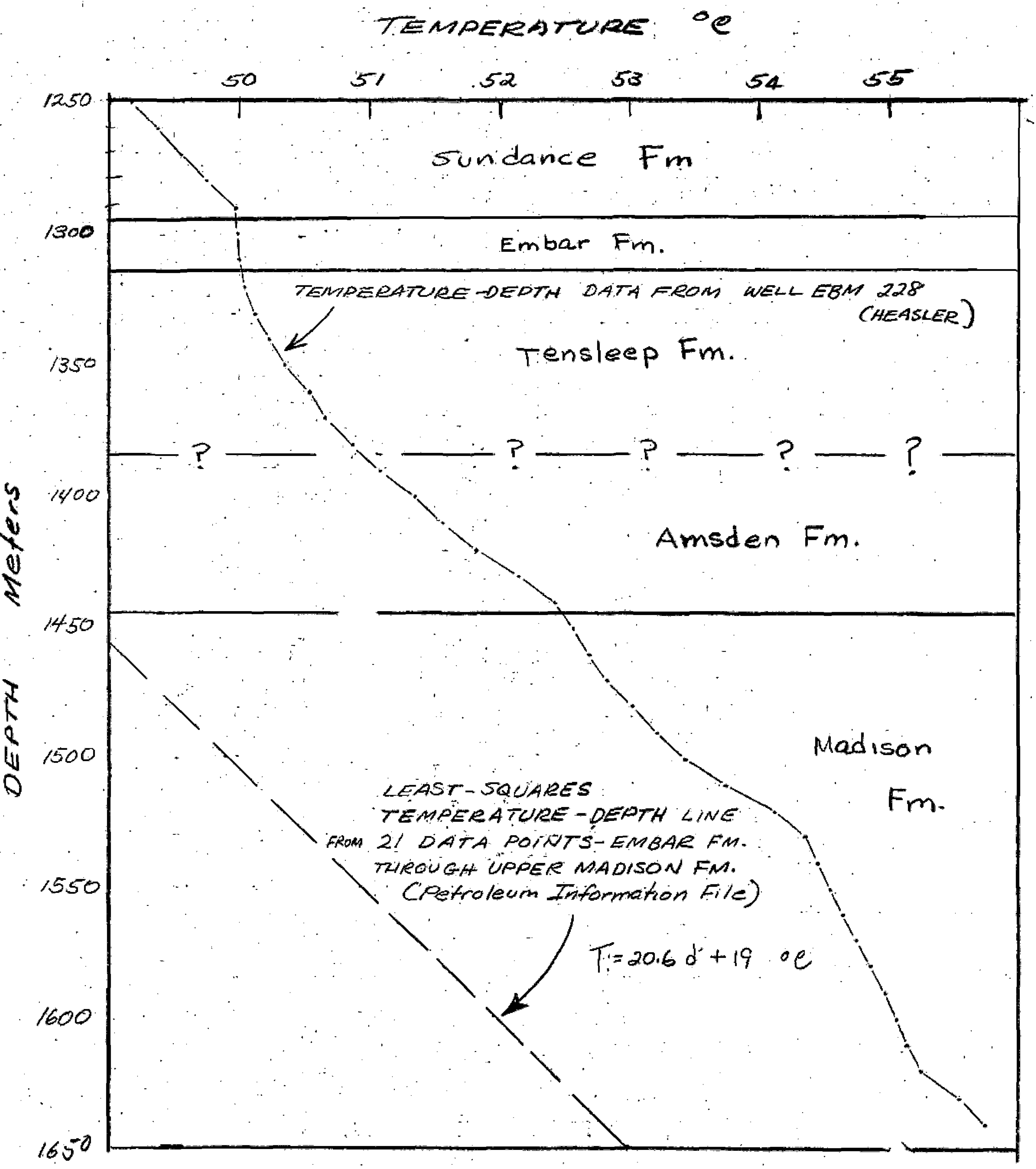
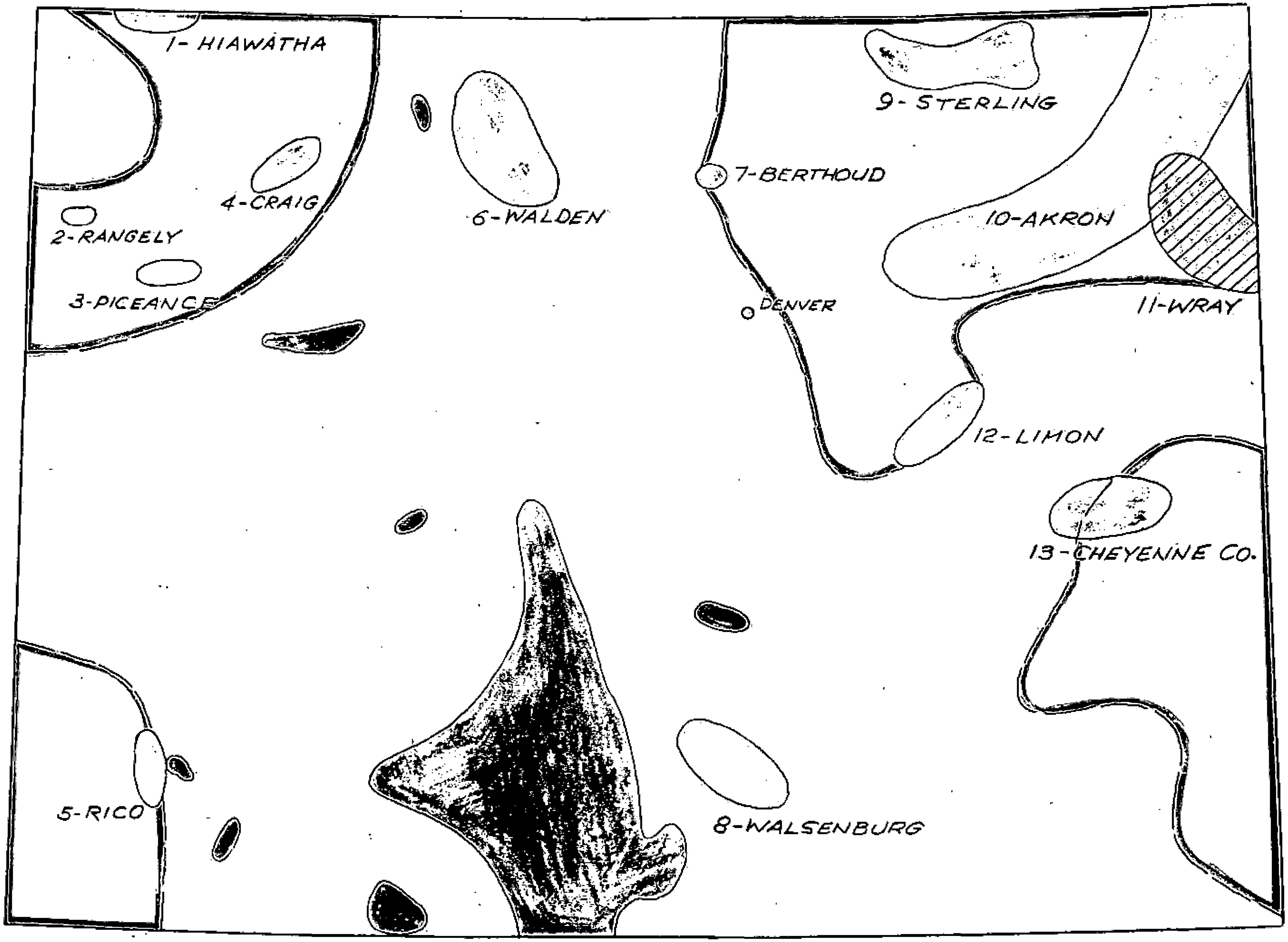


FIGURE 5. Temperature/depth and stratigraphic relationships in the Elk Basin oil field, Wyoming and Montana

Figure 7. COLORADO



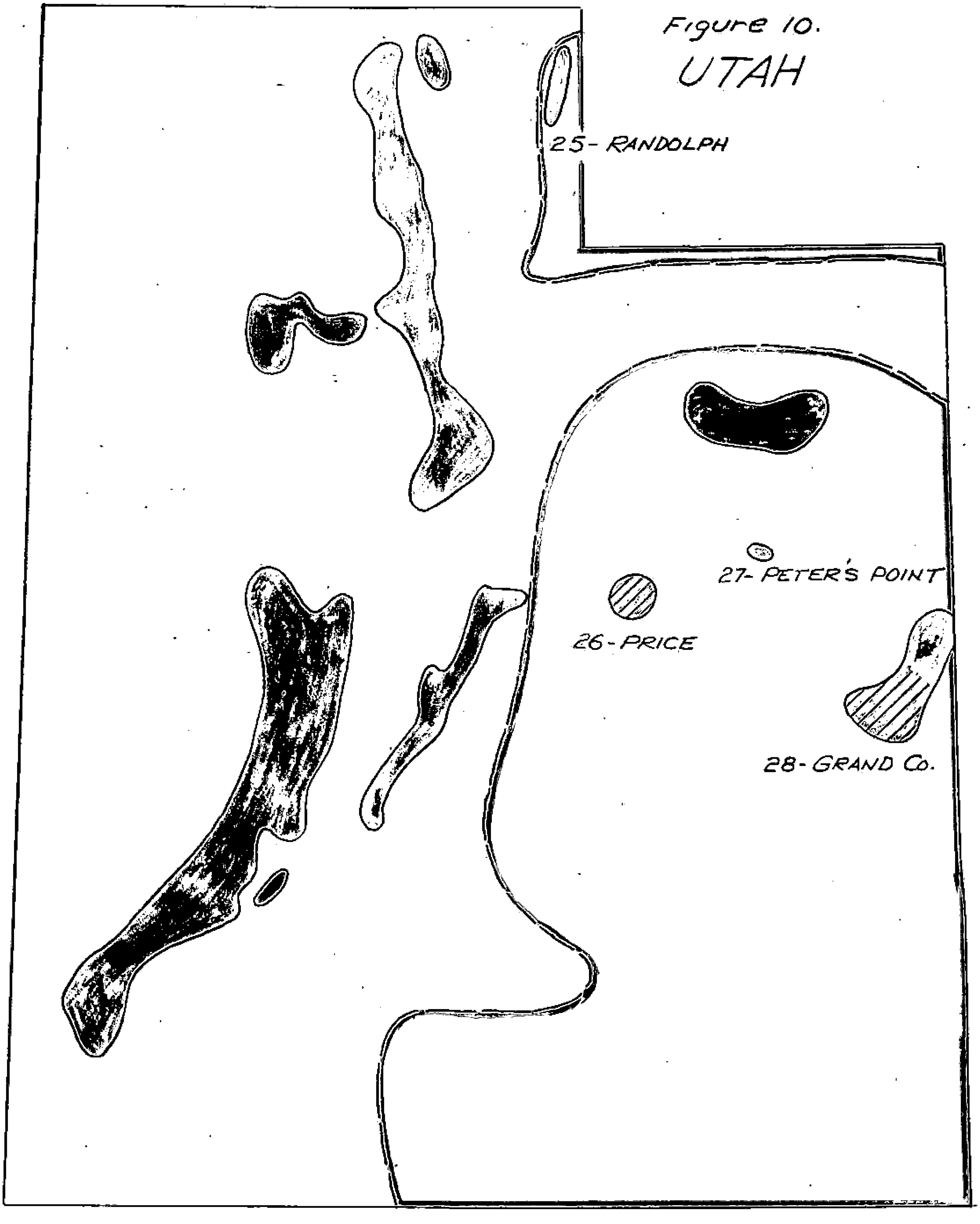


Figure 10.
UTAH
25- RANDOLPH
27- PETER'S POINT
26- PRICE
28- GRAND Co.

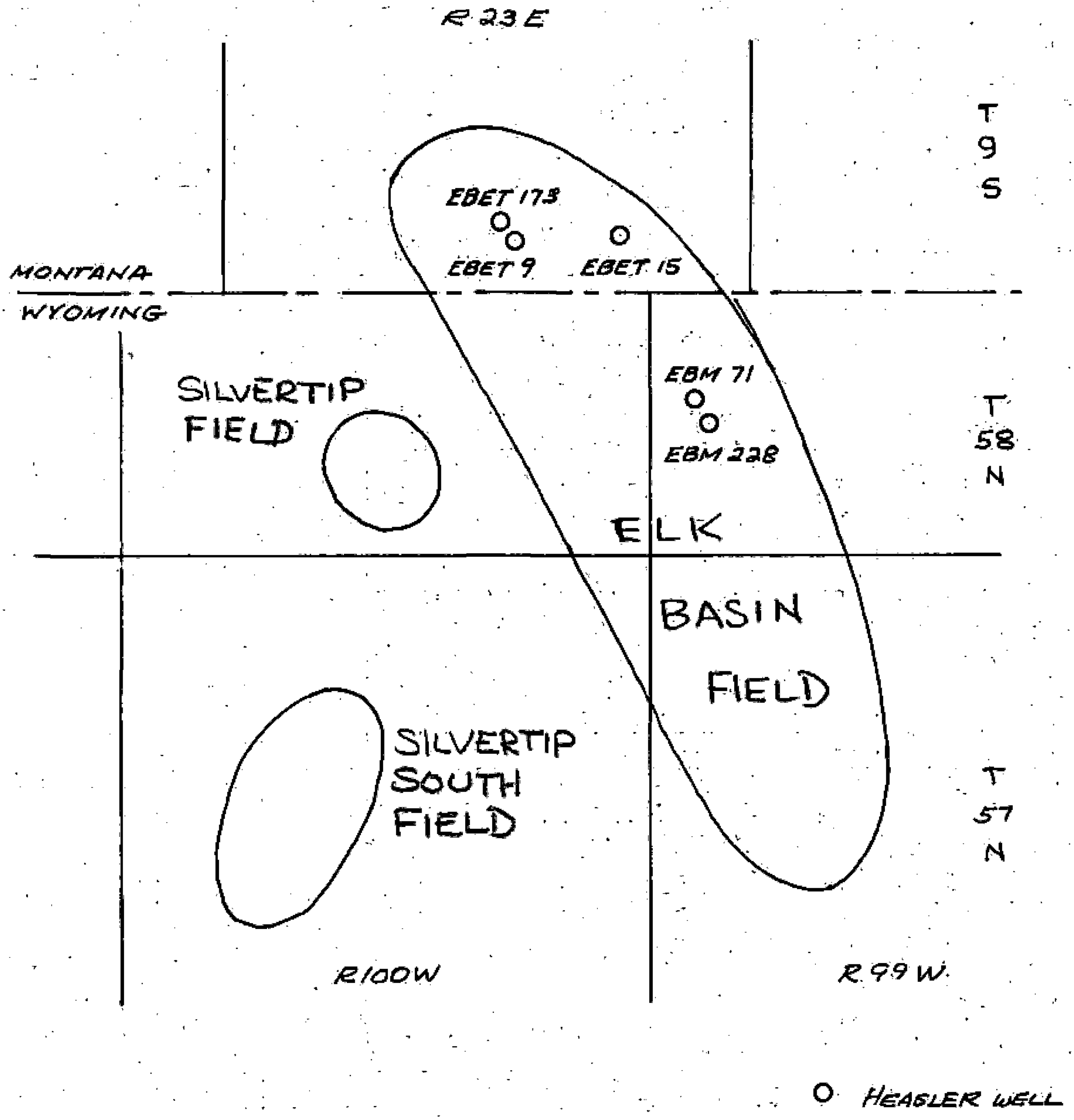


Figure 2. Locations of wells tested by Heasler and outlines of oil fields which supplied Petroleum Information data.

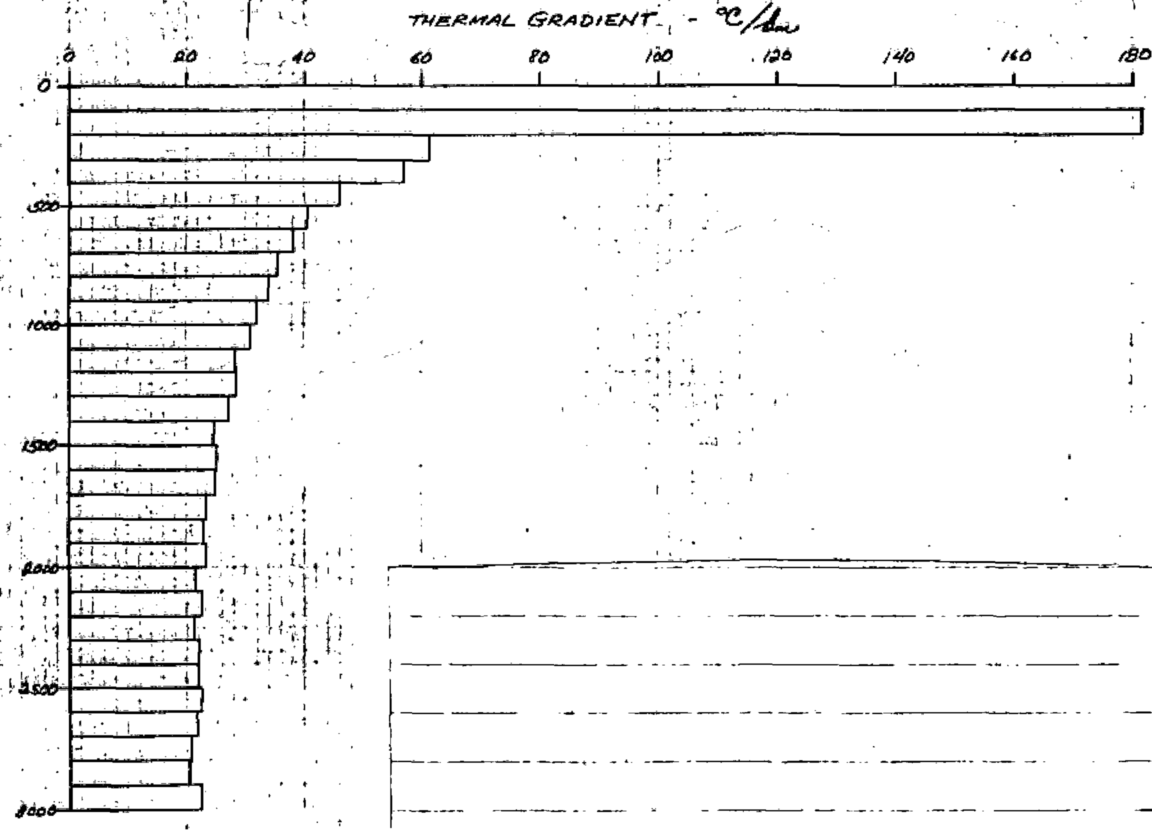


Figure 5. THERMAL GRADIENTS IN UTAH TO THE SURFACE, USING MEAN ANNUAL ATMOSPHERIC TEMPERATURES. AVERAGE GRADIENTS OVER 100 m INTERVALS PLOTTED AGAINST DEPTH

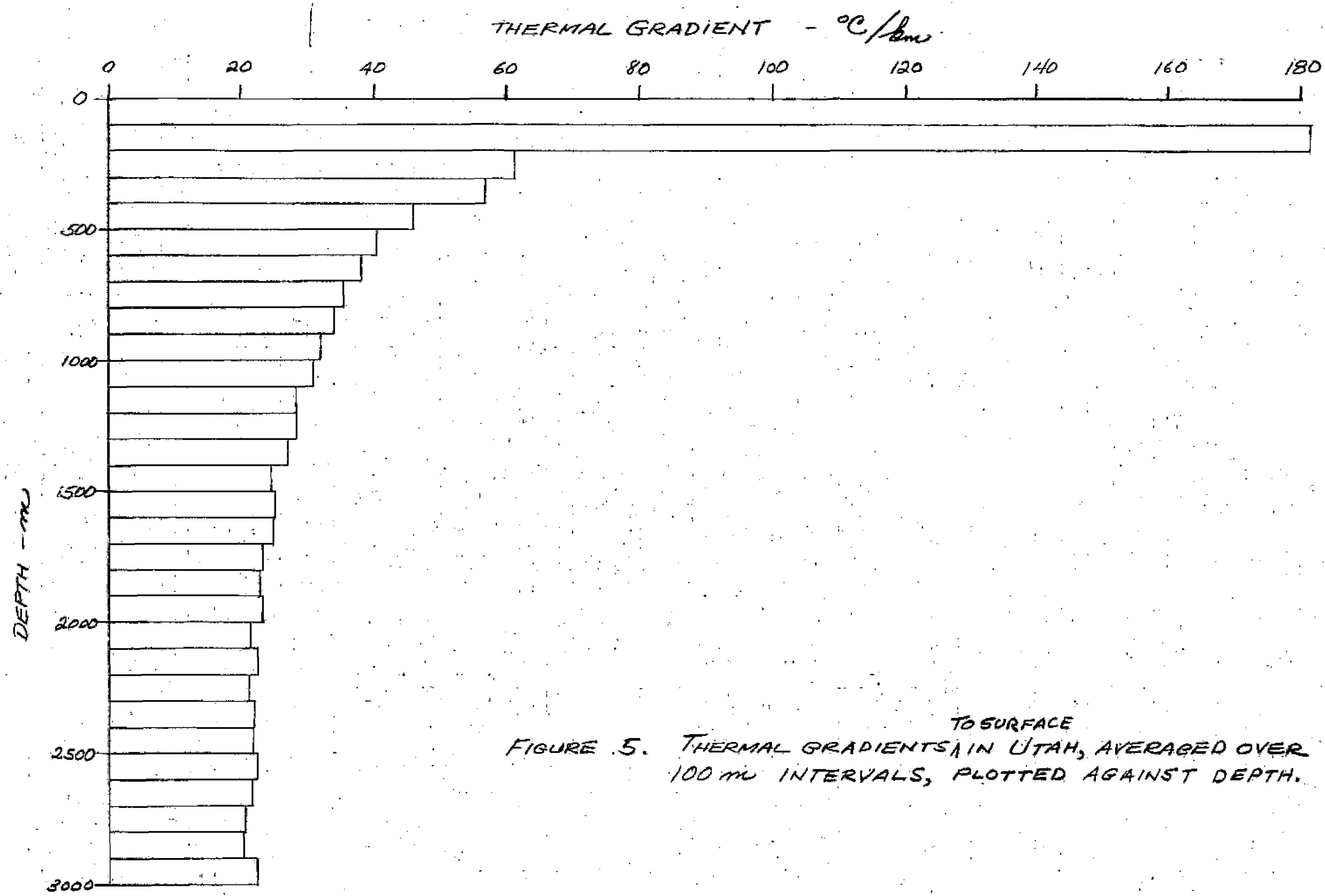
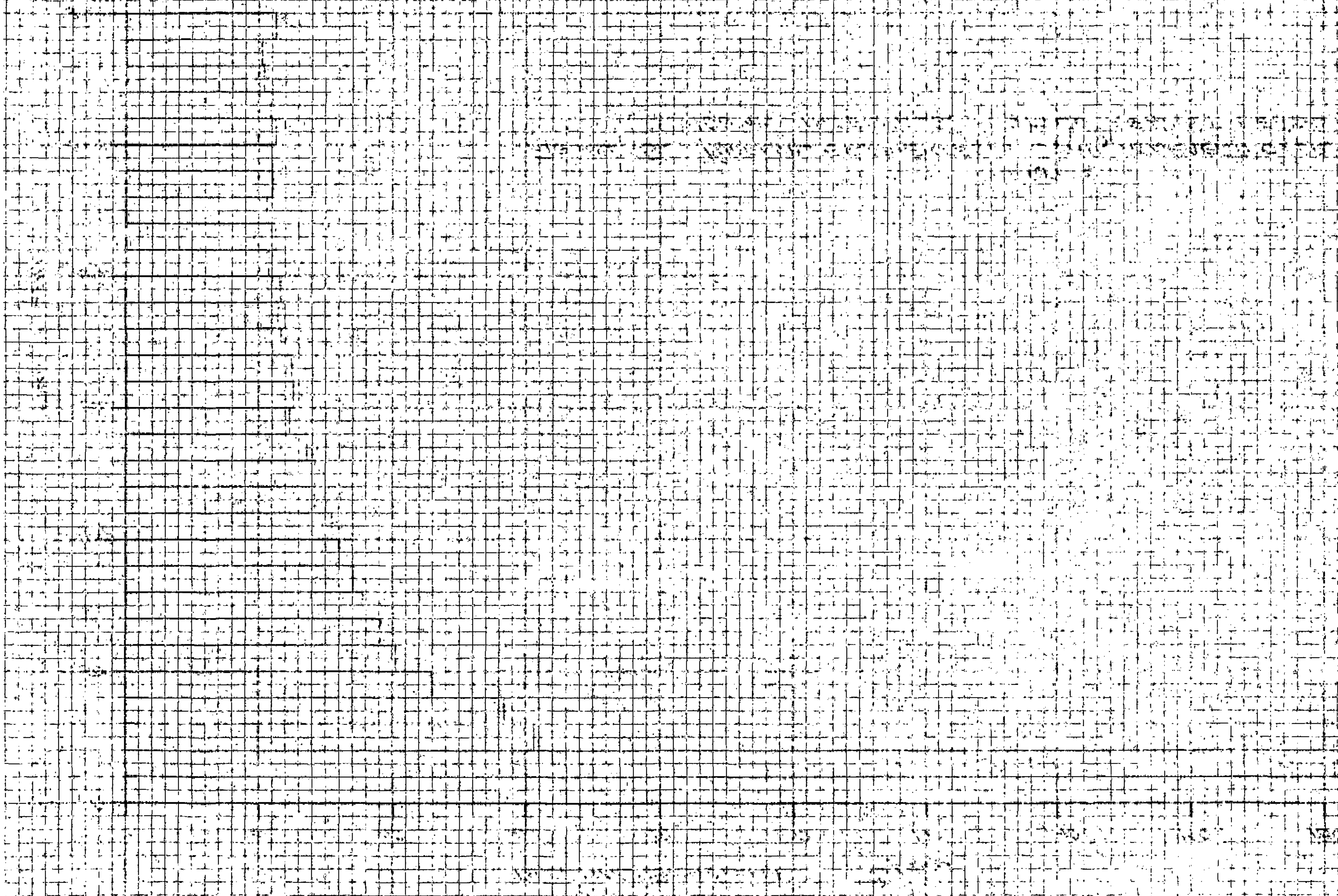


FIGURE 5. THERMAL GRADIENTS TO SURFACE IN UTAH, AVERAGED OVER 100 m INTERVALS, PLOTTED AGAINST DEPTH.



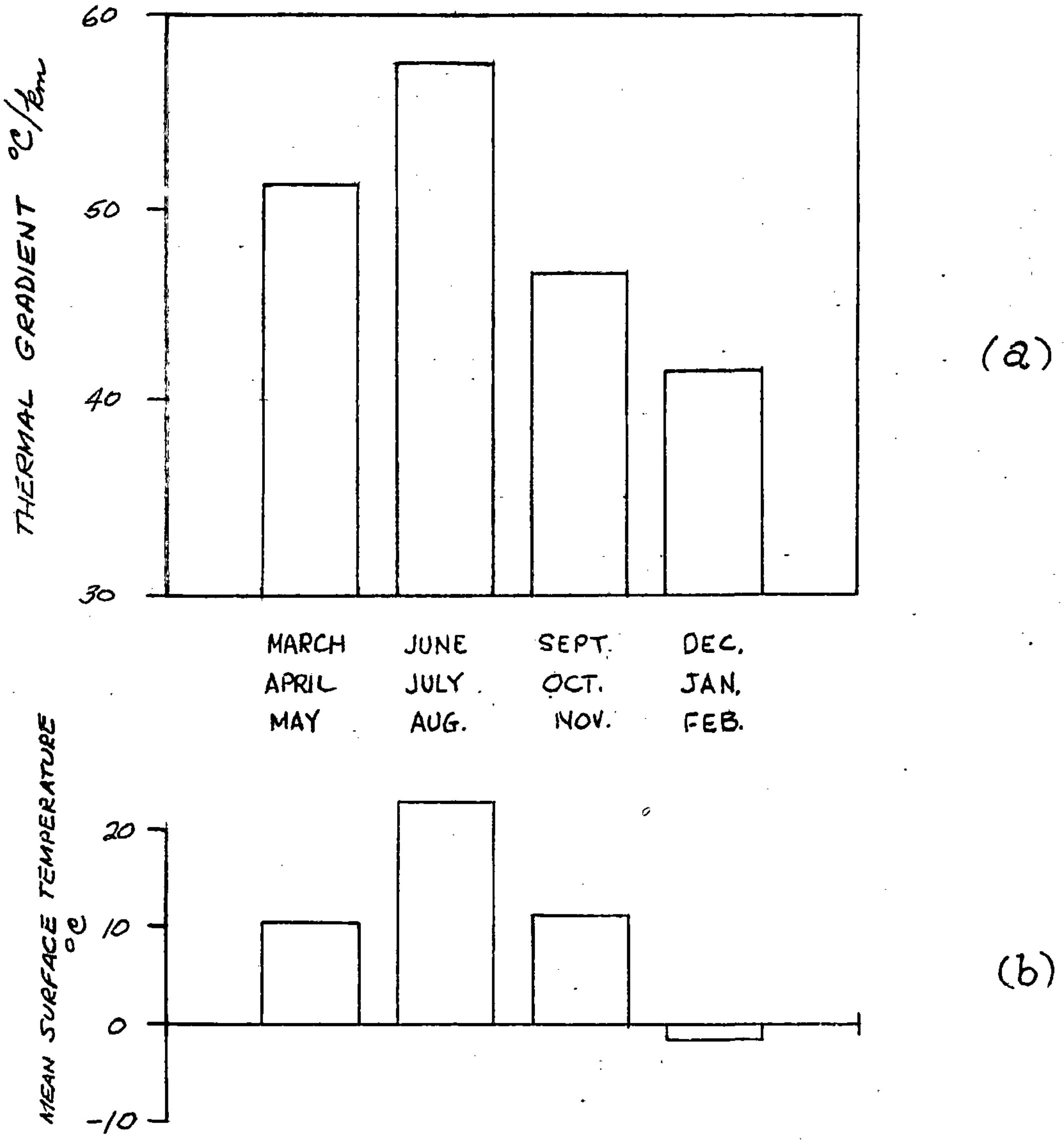


Figure 6. (a) THERMAL GRADIENTS IN UTAH FROM DEPTHS OF 500 m OR LESS TO THE SURFACE, USING MEAN ANNUAL ATMOSPHERIC TEMPERATURES. AVERAGE GRADIENTS BY SEASONS. (b) MEAN SEASONAL ATMOSPHERIC TEMPERATURES.

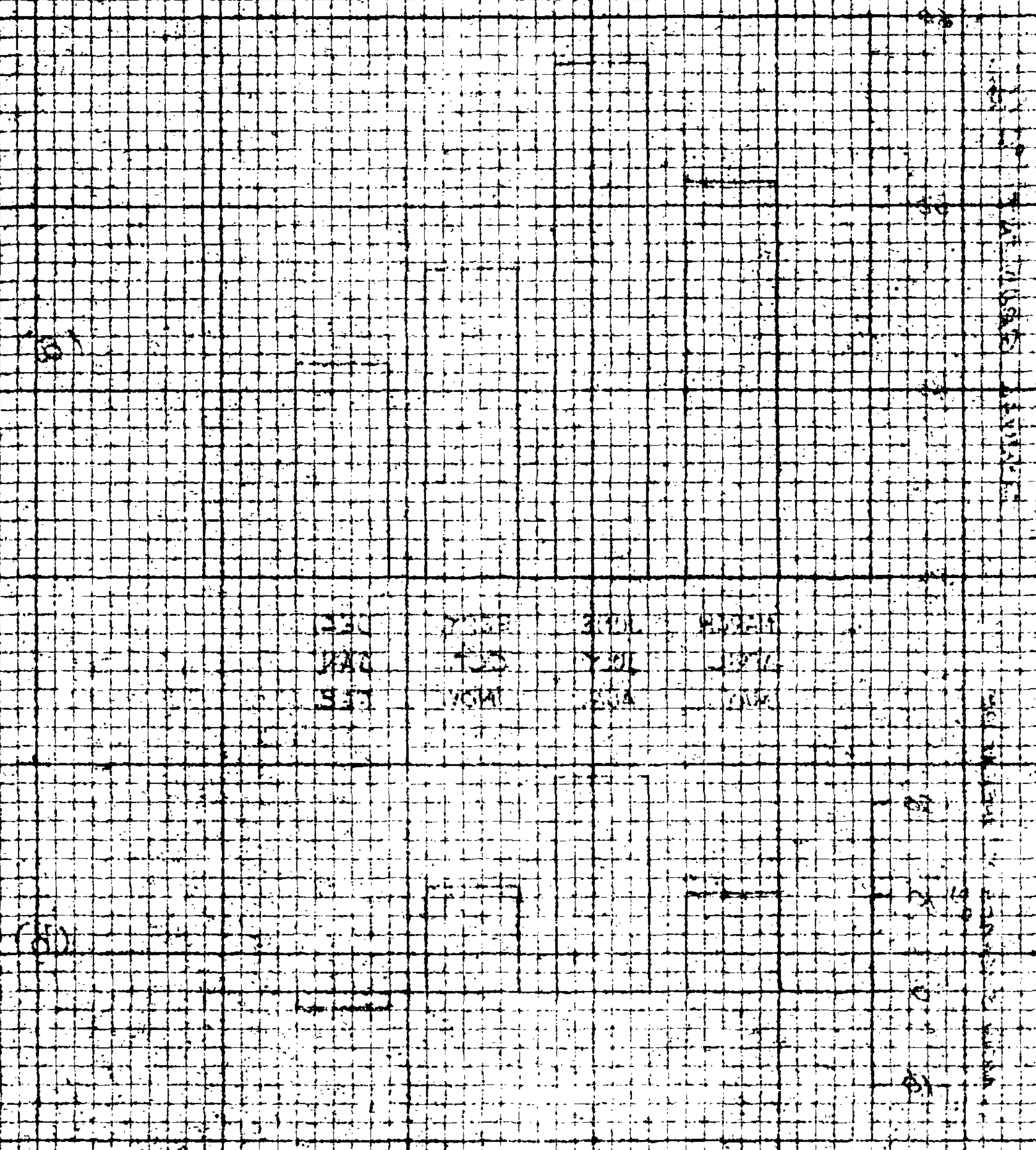


Figure 1. (a) Average thermal gradients in Utah from the surface to depths of 600 m or less. (b) Mean seasonal surface temperatures.

Figure (a) Thermal gradients in Utah from depths of 600 m. or less to the surface. [illegible] annual atmospheric temperature average [illegible] (b) [illegible] temperatures.

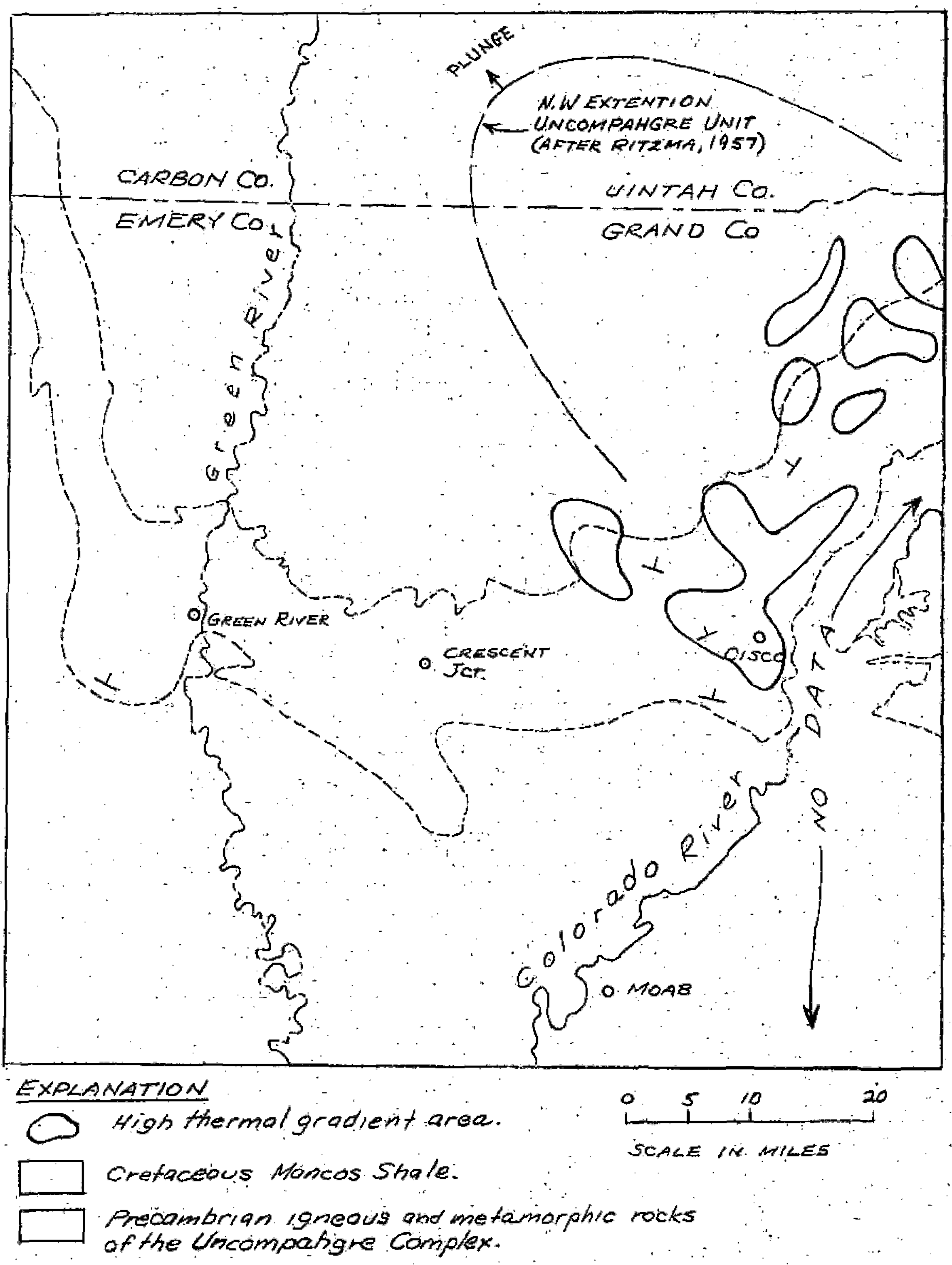


Figure 11. Detail of anomalous thermal gradient areas in Grand County, Utah.

Figures 7, 8, 9, and 10

COLORADO, MONTANA, NEW MEXICO, AND UTAH

COMPARISONS OF THERMAL GRADIENTS IN OIL WELLS WITH AREAS CONSIDERED GENERALLY FAVORABLE FOR THE RECOVERY OF THERMAL WATERS

EXPLANATION

○	AREA GENERALLY FAVORABLE FOR THE RECOVERY OF THERMAL WATERS (USGS CIRCULAR 790).
○	AREA OF ANOMALOUS THERMAL GRADIENTS.
▨	AREA OF ANOMALOUS THERMAL GRADIENTS CALCULATED TO DEPTHS PREDOMINANTLY LESS THAN ONE KILOMETER.
◌	REGIONS OF BEST WELL DENSITY.

Figure 7

COLORADO

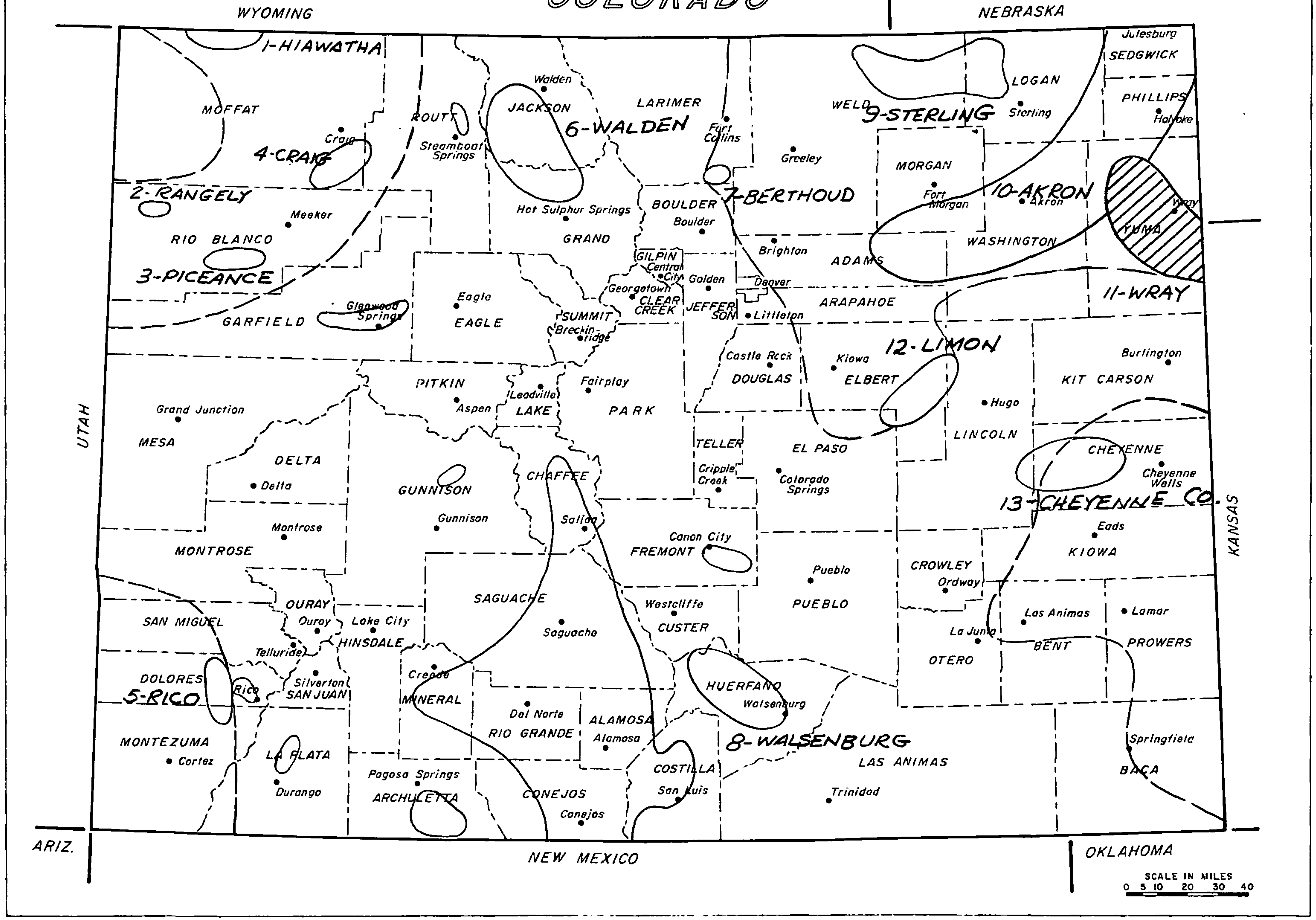
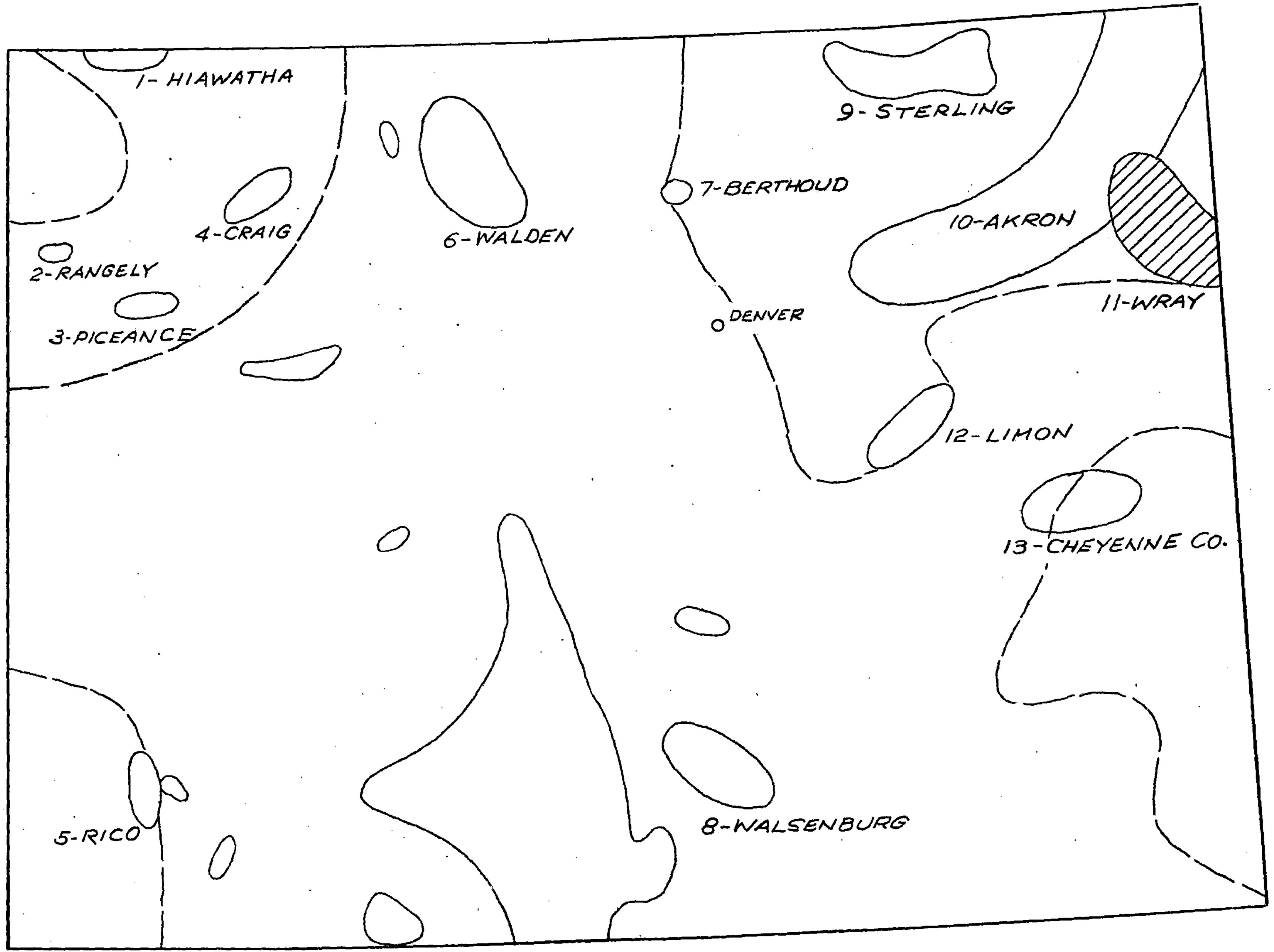
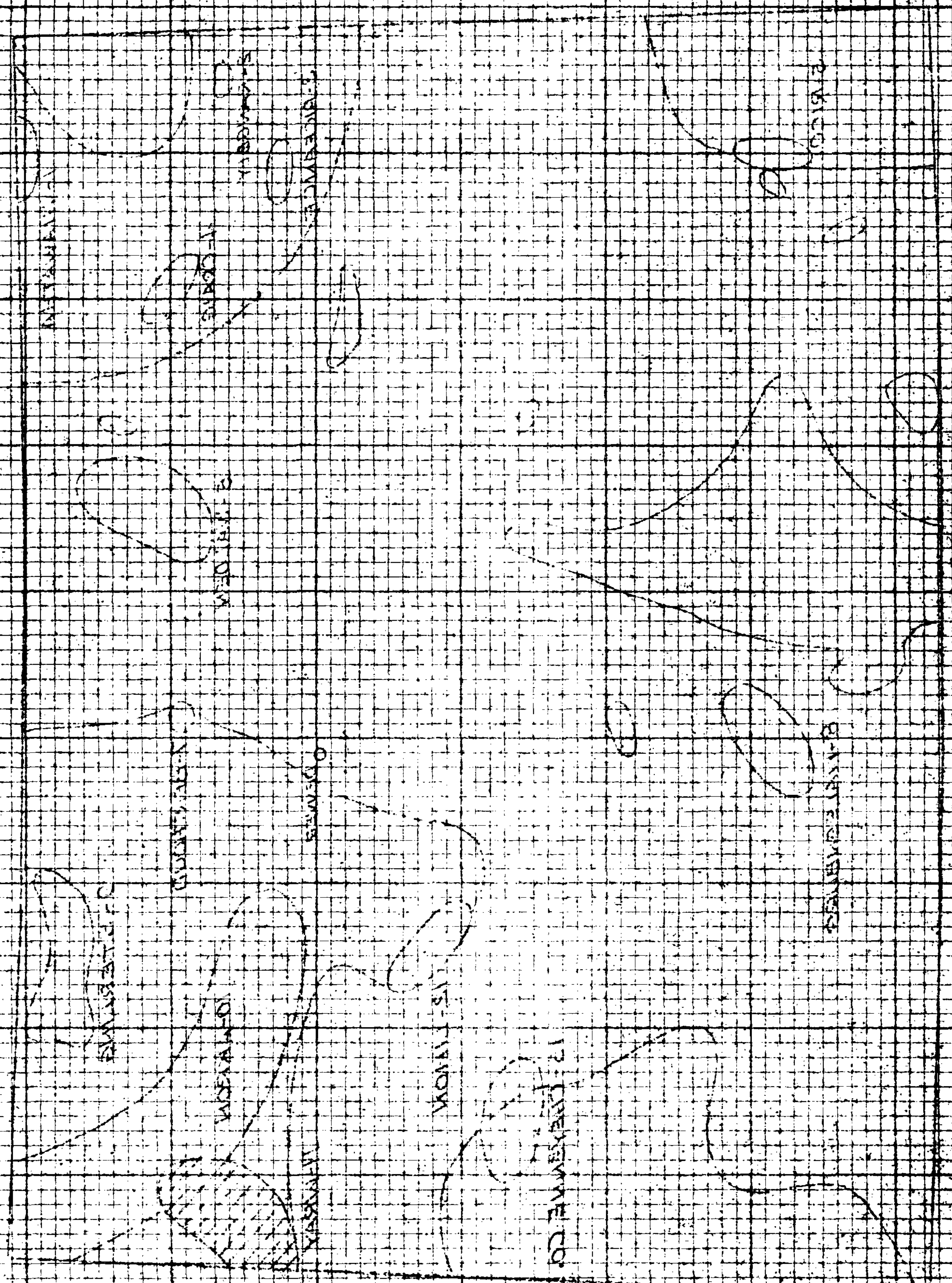


Figure 7. COLORADO





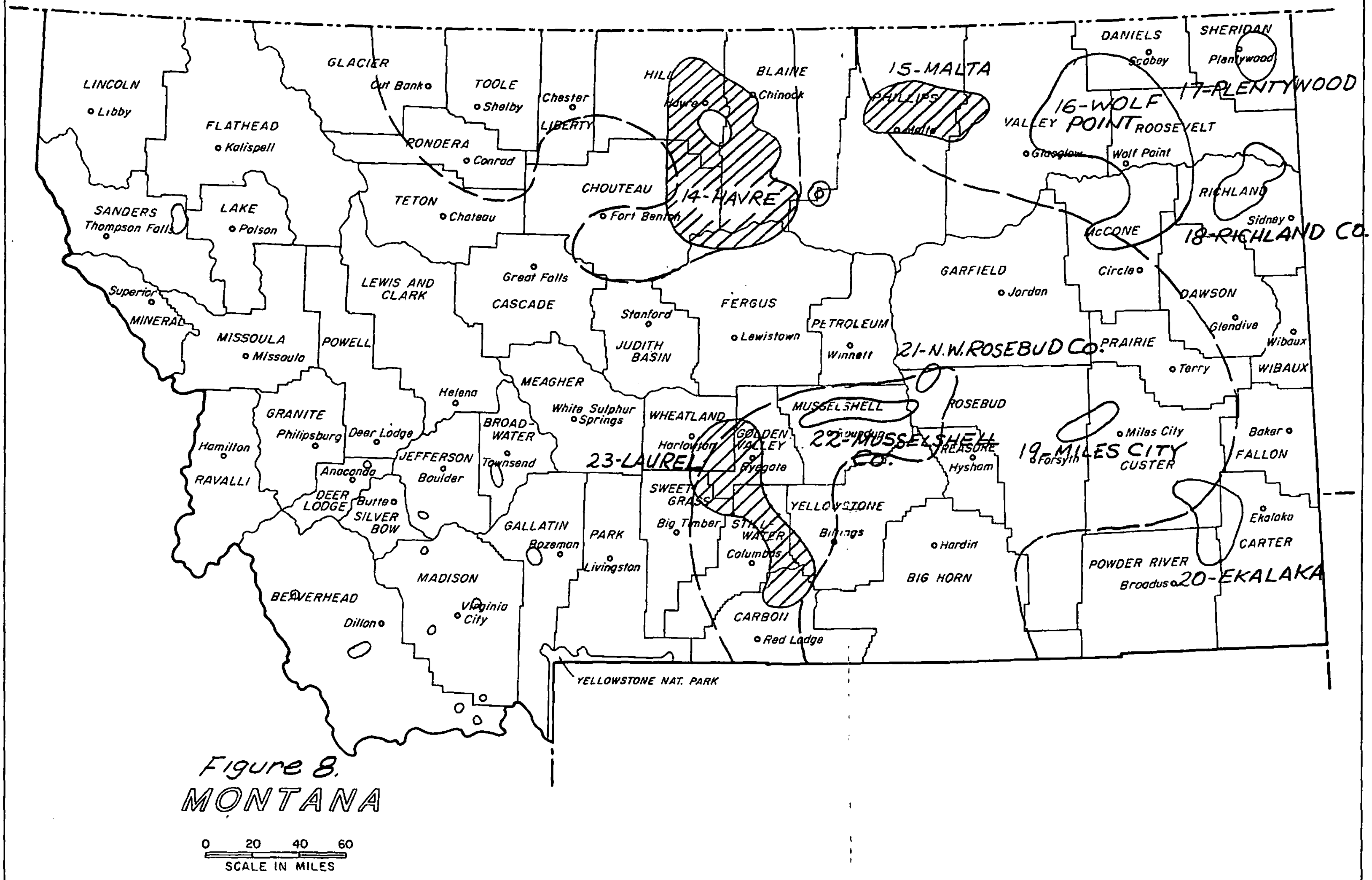
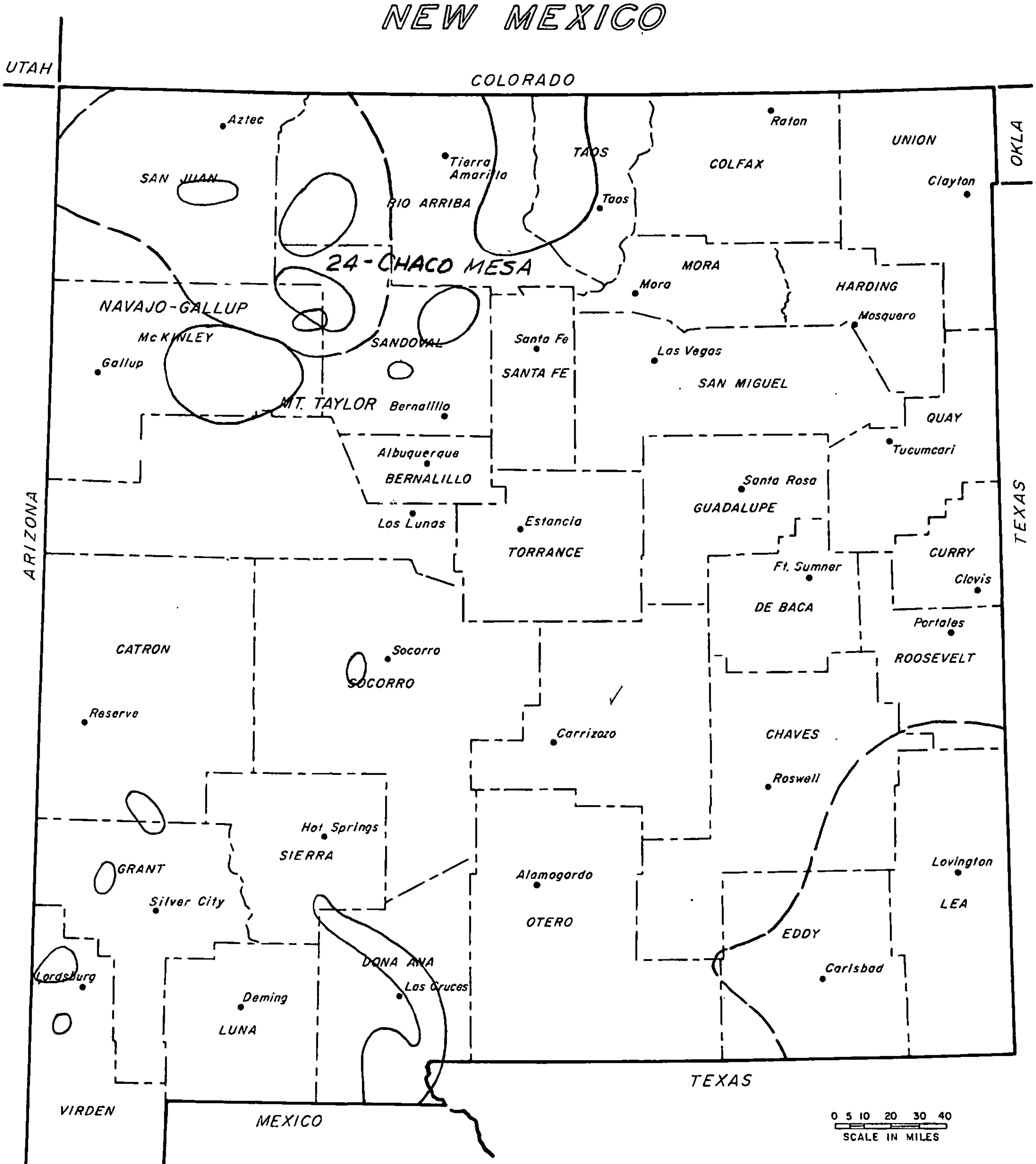


Figure 8.
MONTANA

Figure 9.

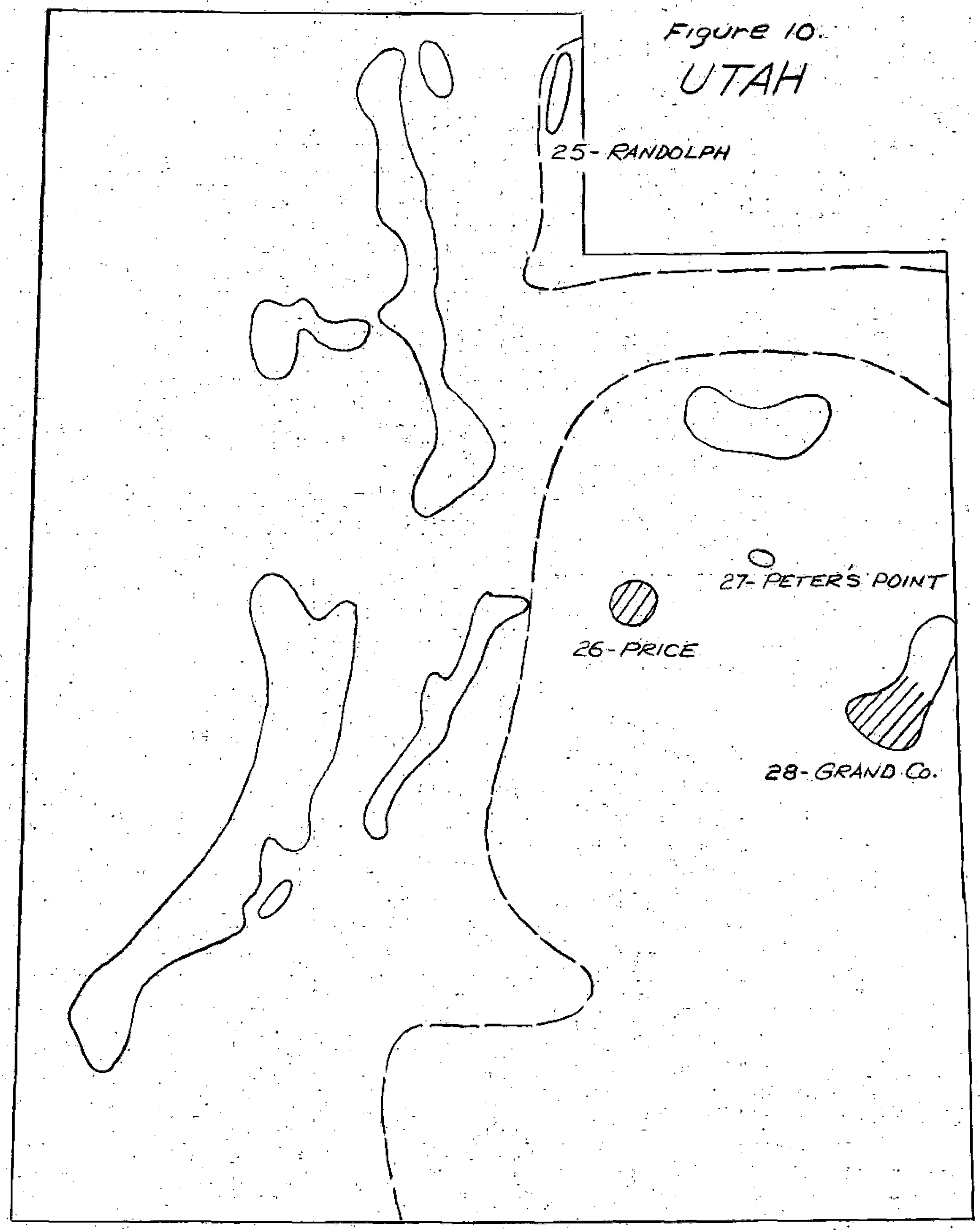
NEW MEXICO

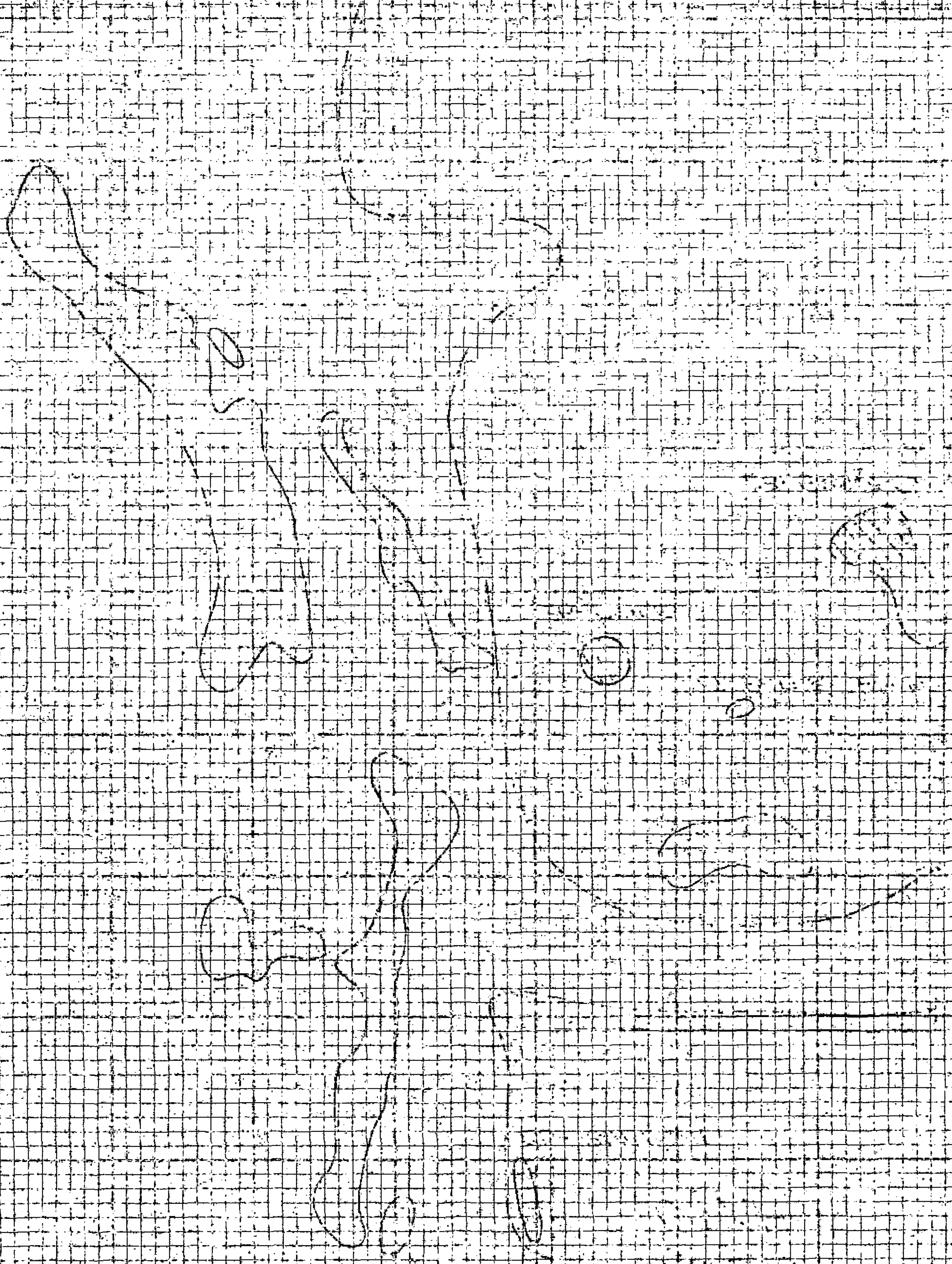


REVISE BASE MAP

Figure 10.

UTAH





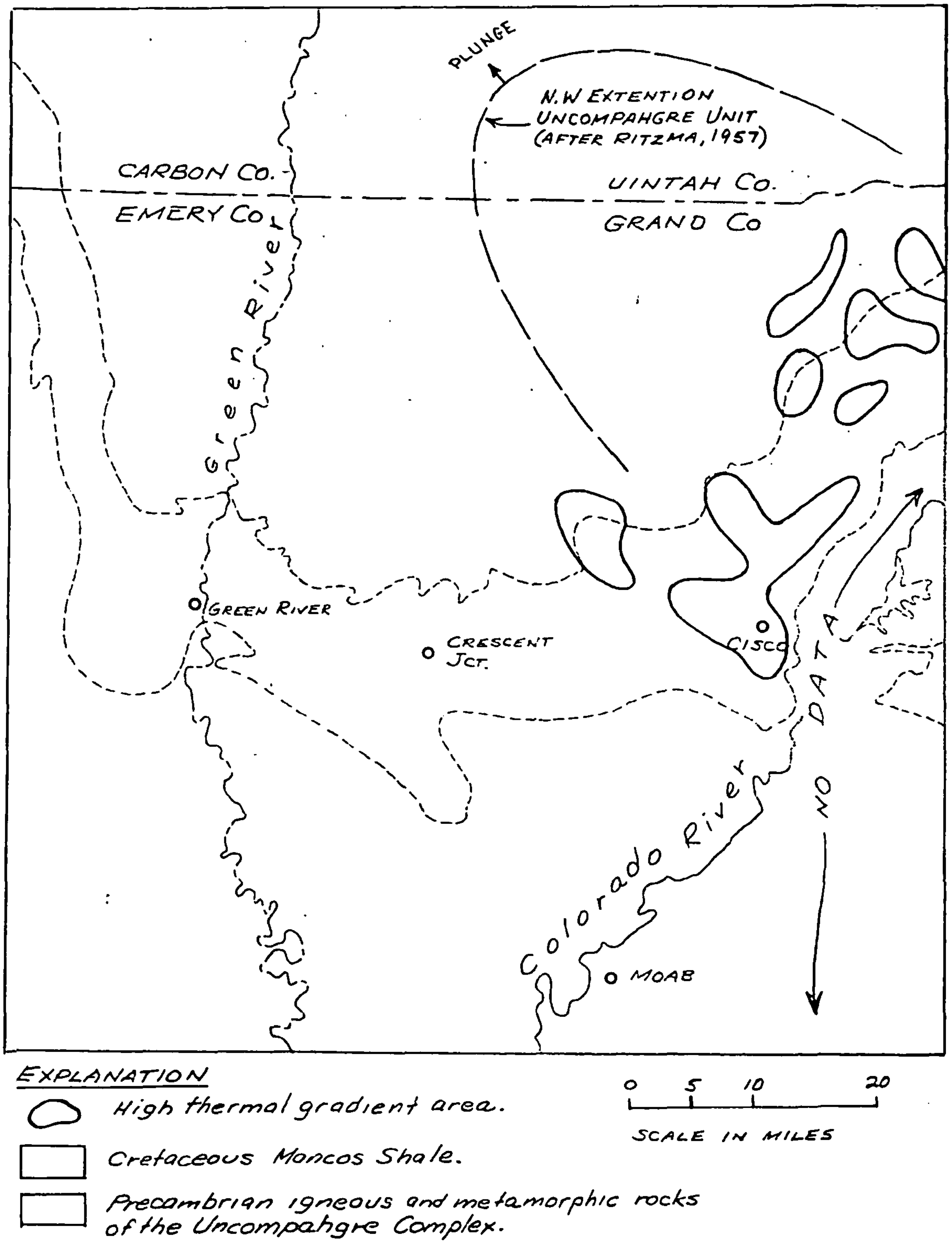


Figure 11. Detail of anomalous thermal gradient areas in Grand County, Utah.

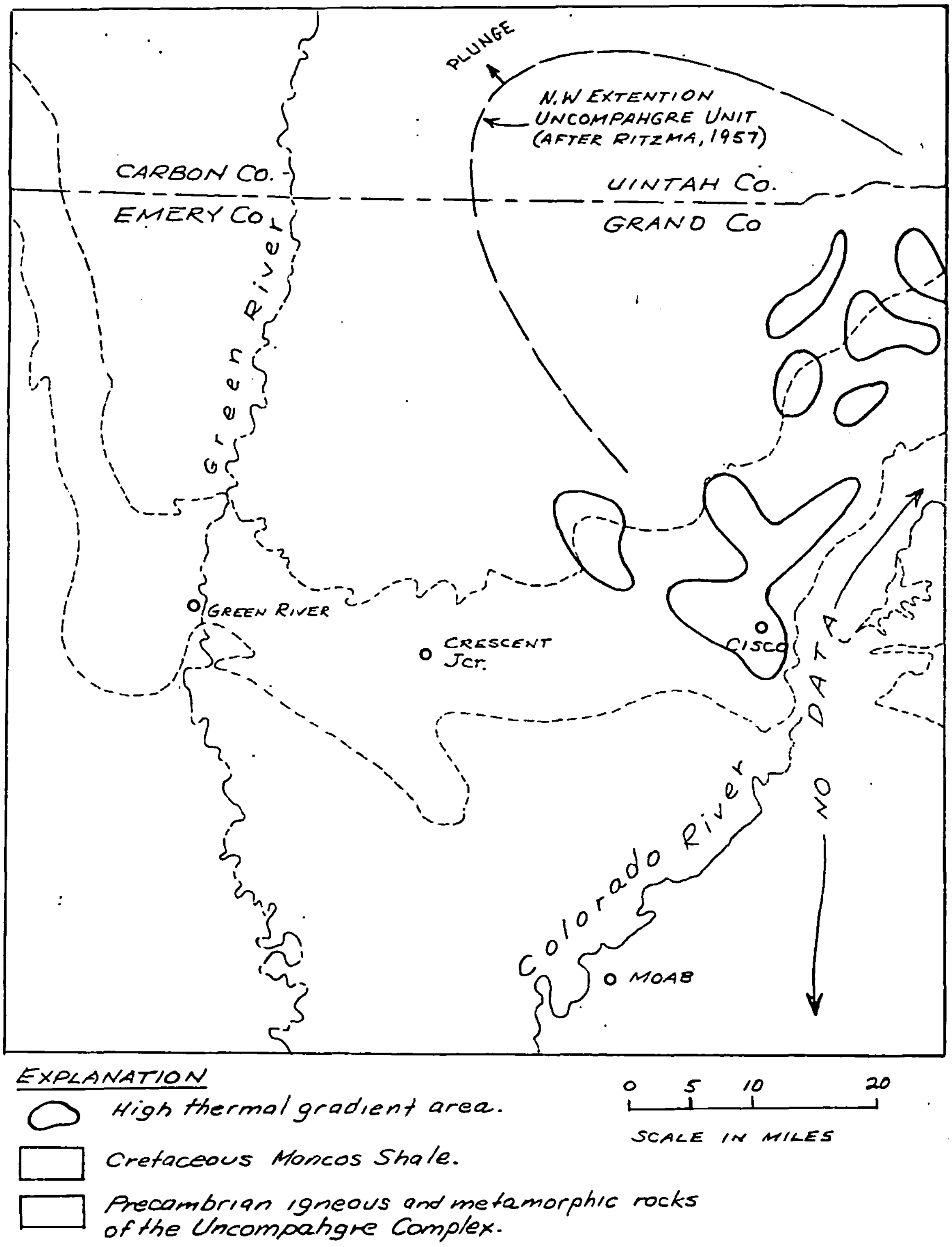


Figure 11. Detail of anomalous thermal gradient areas in Grand County, Utah.

TABLE 2 AREAS OF PREDOMINANTLY HIGH THERMAL GRADIENTS (>35°C/km) - continued

ALTERNATE NOS.	AREA NUMBER AND NAME	NUMBER OF WELLS		DEPTHS TO ANOMALOUS BHT'S m			MAXIMUM GRADIENT °C/km	STRATIGRAPHIC LOCATIONS OF ANOMALOUS BHT'S	COMMENTS
		TOTAL	ANOMALOUS	MINIMUM	MAXIMUM	AVERAGE			
	M O N T A N A (FIGURE 8)								
14	1. HAVRE	83	62	88	716	357	171	CRETACEOUS	WEST OF LITTLE ROCKY MOUNTAINS. SHALLOW WELLS OVER BEAR PAW ARCH AND TO THE NORTH. TERTIARY INTRUSIVE AND EXTRUSNE ROCKS PRESENT.
15	2. MALTA	12	12	290	410	353	101	CRETACEOUS	SHALLOW WELLS. CORRESPONDS CLOSELY WITH APEX OF BOWDOIN DOME. LARGE EXPOSURES OF CRETACEOUS COLORADO SH.
16	3. WOLF POINT	93	78	1,158	2,844	2,064	66	PRINCIPALLY DEVONIAN AND MISSISSIPPIAN	CORRESPONDS, IN PART, WITH DOMING WEST OF THE WILLISTON BASIN.
17	4. PLENTYWOOD	12	11	2,066	3,267	2,342	43	ORDOVICIAN, DEVONIAN, AND MISSISSIPPIAN	WESTERN WILLISTON BASIN
18	5. RICHLAND CO.	24	20	2,570	3,773	2,952	47	ORDOVICIAN, DEVONIAN AND MISSISSIPPIAN	WESTERN WILLISTON BASIN
19	6. MILES CITY	7	7	1,401	1,556	1,449	40	CRETACEOUS KOOTENAI AND MUDDY FORMATIONS	EAST OF PORCUPINE DOME. GENTLY EAST-DIPPING STRATA.
20	7. EKALAKA	7	7	1,255	1,444	1,357	53	CRETACEOUS GREENHORN AN MUDDY FORMATIONS	NORTH END OF BLACK HILLS UPLIFT.
21	8. NW ROSEBUD CO.	16	12	1,472	1,742	1,549	46	MISSISSIPPIAN AND PENNSYLVANIAN	OVER SUMATRA SYNCLINE AND SUMATRA ANTICLINE.
22	9. MUSSELSHELL CO	53	43	787	1,790	1,244	53	PRINCIPALLY MISSISSIPPIAN AND PENNSYLVANIAN	INCLUDES PARTS OF BULL MOUNTAIN BASIN, POLE CREEK ANTICLINE, AND WILLOW CREEK SYNCLINE.
23	10 LAUREL	23	22	202	994	513	90	PRINCIPALLY CRETACEOUS	SHALLOW WELLS. EXTENDS ACROSS LAKE BASIN TO THE FROMBERG FAULT ZONE. ON THE SOUTH.
	N E W M E X I C O (FIGURE 9)								
24	1. CHACO MESA	15	10	1,198	1,949	1,565	41	JURASSIC AND CRETACEOUS	NEAR LITTLE BLUE MESA - AREA CONSIDERED GENERALLY FAVORABLE FOR THE RECOVERY OF THERMAL WATERS (USGS CIRC. 790).

TABLE 2 AREAS OF PREDOMINANTLY HIGH THERMAL GRADIENTS (>35°C/km) continued

AREA NUMBER AND NAME	NUMBER OF WELLS		DEPTH TO ANOMALOUS BHT's m			MAXIMUM GRADIENT °C/km	STRATIGRAPHIC LOCATIONS OF ANOMALOUS BHT's	COMMENTS
	TOTAL	ANOMALOUS	MINIMUM	MAXIMUM	AVERAGE			
			U T A H (FIGURE 10)					
25 1. RANDOLPH	4	3	105	1,416	714	228	JURASSIC TWIN CREEK FM.	WEST FRONT OF CRAWFORD MOUNTAINS IN THE GREEN RIVER BASIN. BHT OF 65.6°C AT 1,416m ONE MILE SOUTH OF RANDOLPH*
26 2. PRICE	4	4	370	608	480	48	PROBABLY TERTIARY	SHALLOW WELLS. MAXIMUM BHT, 40°C AT 608 m.*
27 3 PETER'S POINT	11	7	405	998	789	137	TERTIARY	UINTA BASIN. ALL WELLS IN PETER'S POINT OIL FIELD.
28 4. GRAND CO	180	136	154	1,404	659	98	TRIASSIC, JURASSIC AND CRETACEOUS	ASSOCIATED WITH MANCOS SH. OUTCROP AND UNCOMPAHGRE UPLIFT. BHT OF 72.8°C AT 867m TWELVE MILES WEST OF CISCO.*

*TEMPERATURES AND DEPTHS FROM THE UTAH DIVISION OF OIL, GAS, AND MINING.

Whipple

September 21, 1979

MEMORANDUM

TO: ESL Staff

FROM: Mike Wright

SUBJECT: Management of the Geochemical Laboratory and Staff

As you all know, Bob Bamford plans to leave ESL in order to pursue private consulting. Therefore, effective 1 October 1979, Joe Moore will assume responsibility for management of the Geochemical Laboratory facility and staff. Reporting directly to Joe will be the current lab staff, including Odin Christensen, Regina Capuano, Dave Cole, Ruth Kroneman, and Tina Serling.

Joe will be responsible for implementing new programs planned for FY80, for continuation of the current programs, for ensuring the quality of the analytical and other work produced by the lab and for assigning priorities. Requests for lab staff assistance and for analytical work should be communicated to Joe.

After 1 October, Bob will be working as a consultant to ESL. He will be working at the lab full time until about the middle of October, when the initial writing for the Roosevelt Hot Springs report will be completed. At that time, Dave Cole will be taking over Bob's office space, and Bob's work for ESL will be on an as needed basis until The Geysers work is complete. Odin Christensen will assume primary responsibility for the Roosevelt and Geysers studies on behalf of ESL after 1 October and will interface with Bob on Bob's continuing contribution.

We have very much appreciated the excellent work which Bob has directed while at ESL. Industry has recognized this work as being highly interesting and significant. Those of you who have contributed to the success of the geochemical research efforts to date can be justly proud.

Mike

Mike Wright
Associate Director

MW:srm

ANACONDA

From
DENNIS L. NIELSON

(Uranium)
Cathles - Econ Geology - Dec. ⇒ 1978
Jan^{1979} - Hydrothermal
convection system generated by
radioactive decay.

Stuckless et.al, 1977,
J. Res. U.S.G.S.
v. 5, no. 1., 61-81

PI Master
Idaflya pp 1693- 1696 film 009
no new ones

WED
FHP DR HECT 3 pm

Chart

~~wells [illegible]~~

Areas which appear to be associated with known thermal waters

Areas of no thermal waters without associated anom thermal gradient areas

Areas associated with abundant low-thermal conductivity ~~[illegible]~~ rocks in outcrop.

Areas associated with doming

areas associated with outcropping granite rocks

COLORADO ANOMALOUS AREAS

AREA	WELLS			DEPTH TO ANOMALOUS BHT			MAX. GRAD. °C/km	STRAT. LOCATION ANOM. BHTS	Comments
	ANOM	TOTAL	% ANOM	MIN	MAX	AV			
1. HIAWATHA	3	5	60	957	1807	1490	48	TERT + K WASATCH TO MESAVERDE	
2. RANGELY	11	14	78	902	1673	1151	50	K TO Upper TR DAKOTA,	
3. PICEANCE	4	5	80	1281	2725	1728	47	T-WASATCH & K MESAVERDE	
4. CRAIG	5	8	62	926	1478	1196	43	CRETACEOUS PRINC. DKOT	
5. RICO	4	4	100	1238	2025	1735	54	MISSIPPIAN & Pennsylvanian	Dunton-Rico Hot Spgs. area
6. WALDEN	6	9	67	270	1974	1364	109	TERTIARY-CRETACEOUS	
7. BERTHOUD	3	4	75	934	1107	998	45	CRETACEOUS	At town of Berthoud
8. WALSENBURG	8	9	89	512	1885	1087	59	TERT. TO JURASSIC	
9. STERLING	17	23	74	1496	2361	1855	48	DAKOTA Ss	
10. AKRON	51	83	61	1050	2089	1568	46	CRETACEOUS DAKOTA Ss.	
11. WRAY	17	25	68	494	924	716	55	CRETACEOUS NIOBRARA Fm	SEPARATED FROM AKRON ON BASES OF DISTINCTIVE DEPTHS AND FORMATIONS TESTED
12. LIMON	7	7	100	944	1906	1650	47	CRETACEOUS princ. DKOT	
13. CHEYENNE Co.	5	5	100	979	1489	1245	44	PERMIAN & PENNSYLVANIAN	

NEW MEXICO

LITTLE BLUE MESA

AREA	WELLS			DEPTHS TO ANOMALOUS BHT'S			MAX GRAD °C/km	STRAT. LOCATIONS ANOMALOUS BHT'S	COMMENTS
	ANOM	TOTAL	%	MIN.	MAX	AV			
CHACO MESA	10	15		1198	1949	1565	41	JURASSIC and CRETACEOUS	LITTLE BLUE MESA THERMAL WATERS AREA USGS CIRC. 790

MONTANA ANOM. AREAS

AREA	WELLS			DEPTHS TO ANOM BHTS			MAX GRAD °C/km	ANOM. BHT'S STRAT. LOCATIONS	COMMENTS
	ANOM	TOTAL	%	MIN	MAX	AV			
1. HAVRE	62	83	75	88	716	357	171	CRETACEOUS	West of Little Rockies
2. MALTA	12	12	100	290	410	353	101	CRETACEOUS	
WOLF POINT	78	93		1158	2844	2064	66	princ. Miss + Dev	1-Jurassic 1-Cretaceous
PLENTYWOOD	11	12		2066	3267	2342	43	ORDOVICIAN, DEVONIAN + MISSISSIPPIAN	
RICHLAND CO.	20	24		2570	3773	2952	47	ORDOVICIAN THRU MISSISSIPPIAN	
MILES CITY	7	7		1401	1556	1449	40	CRETACEOUS KOOTENAI and	MUDDY FM's.
EKALAKA	7	7		1255	1444	1357	53	CRETACEOUS GREENHORN and	MUDDY FM's.
NW ROSEBUD CO	12	16		1472	1742	1549	46	MISSISSIPPIAN and PENNSYLVANIAN	
MUSSELSHELL CO	43	53		787	1790	1244	53	MOST FROM MISSISSIPPIAN and PENNSYLVANIAN	2-Cretaceous 1-Jurassic
LAUREL	22	23		202	994	513	90	MOST CRETACEOUS	1- Pennsylvanian (deepest)

ANOM AREAS
UTAH

AREA	WELLS		% ANOM	DEPTH TO ANOM BHTs (meters)			MAX ANOM GRADS	LOCATION OF ANOM BHTS	Comments
	ANOM	TOTAL		MIN	MAX	AV			
RANDOLPH	3	4	75	105	1416	714.	228.	Jurassic Twin Creek Fm	W FRONT CRAWFORD MTNS
PRICE	4	4	100	370	608	480	48.	UNKN prob Tert(?)	
PETERS PT.	7	11	64	405	998	798	137	TERT	Peter's Point oil Field
GRAND Co	136	180	75	154	1404	659	98	K, J, and TR	ASSOCIATED w Km + PE

Comments

RANDOLPH – West front Crawford mtns. Well 1 mile south of Randolph has a BHT of 65.6 °C at 1416 m.*

PRICE - Shallow wells. Maximum BHT is 40°C at 608 m. This location is 4 miles SW of Price*

PETERS POINT. Area 20 miles northeast of East Carbon City. (Good PI well here - will check DOGM files for it - missed on initial check

GRAND COUNTY – Associated with Mancos Shale outcrop and the Uncompahgre Uplift. Well 12 miles west of Cisco has a BHT of 72.8°C at 867 m.* Water reported

*From files of Utah Division of Oil, Gas, and Mining

UTAH

ISOLATED HIGH GRADS
BHT's > 120°F

PI

	County	#	°F	ft	°C	m	°C/km	°C MAST	FM
①	Uintah	54	200	5458	93.3	1664	49.7	10.6	404 DRCK
②	Duchesne	82	185	5602	850	1707	45.6	7.2	652GRRV
③	Uintah	91	142	2930	61.1	893	60.4	7.2	"
④	Uintah	94	145	1286	62.8	392	141.8	7.2	659TRTR
⑤	San Juan	154	125	2610	51.7	796	51.6	10.6	419CTLR

should be able to find PI wells in DOGM
(not necessarily same BHT)

DOGM

	County-No.	°F	ft	°C	m	°C MAST	°C/km
⑥	BOX ELDER-1	142	3500	61.1	1067	9.4	48.4
⑦	SUMMIT-11	160	4124	71.1	1257	6.1	51.7
⑧	TOOELE-2	193	5343	89.4	1628	10.0	48.8
⑨	UINTAH-88	128	524	53.3	160	7.2	288.1

① S. EDGE County.
② 12 MI S. OF MYTON
③ 2 MI W. OF LITTLE BONANZA
④ 16 MI S. OF OURAY
⑤ 4 MI NE OF LA SAL
⑥ ROSSEL PT GR. SALT LAKE
⑦ 20 MI W. OF MANILA
⑧ 4 MI SE OF BURMESTER
⑨ 12 MI W. OF RAINBOW

Wells to Check Dogm

PI-96 Peters' Point Carbon Co.
Reserve Oil 14 Peters' Point
sec ~~13~~7, 13S-17E missed 1st search

PI-104 Grand Co.
Atlantic Richfield 2-2 Arco-State
sec 2, T 16S - R 24E
No 14 DOGM – NOTHING REMARKABLE 28.8°C/km

PI-54 Uintah Co.
Superior Oil R-14 MCU
sec 27, T 15S - R19E missed 1st search

PI-82 Duchesne Co.
Pure Oil 1 Castle Draw
sec 10 9S - 17E

PI-91 Uintah Co.
Chorney Oil 1-18 South Red Wash-Fed
sec 18, T9S-24E

PI-94 Uintah Co
Mapco Inc 2-8 Hope Unit-Fed
sec 8 11S - 21E

PI-154 San Juan Co.
Union Oil of Calif
sec 33, T28S - R25E

1 Pine Ridge USA

Anomalous Areas Colo, Mont, NM, Ut.

AV. DEPTH m	AV GRAD °C/km
1490	36.6
1151	40.3
1728	38.9
1196	36.7
1735	45.8
1364	42.4
998	41.0
1087	41.2
1855	37.2
1568	36.0
716	39.9
1650	38.4
1245	41.0
357	47.0
353	54.7
2064	38.9
2342	36.6
2952	37.0
1449	37.3
1357	40.3
1549	37.5
1244	38.2
513	52.0
1565	37.0
714	86.7
480	44.7
789	45.3
659	42.9

— Colo

mont

— mont

— N M

UT

NW ROSEBUD, MONT MAST 45°F

TOTAL

#	°F	ft	°C/km	
188			39.2	
192			37.6	
198	141	5780	30.2	
201			35.8	
202			43.0	
203			37.5	
211			43.4	
212			35.7	378
213	135	4834	33.9	
214	138	5493	30.8	
215			44.0	
216			42.3	
218			46.5	
219			37.0	
220			45.0	
210	138	4960	34.2	392

av of all
37.5 °C/km

→ Combine with Hill Co to form

MONT ~~[illegible]~~ Havre MAST 42.3° F
5.7°C

CHOUTEAU - HILL - BLAINE Cos
& FERGUS COS.

#	°F	ft	°C	m	°C/km	FM
275	92	1316				604 EGLE
338	80	616				
376	76	1062				
377	91	554		168		
410	87	1242				
411	70	980				
414	89	1435				
415	75	1084				
444B	100	1796				604 EGLE
446	89	1654				604 EGLE
447	71	991				604 EGLE
448	100	815				604 JDRV
451	98	1984				604 EGLE
452	75	1004				604 EGLE
457	98	1906				604 JDRV
* 460	126	2350	52.2	716	64.9	604 EGLE
* 461	121	2261	49.4	689	63.4	"
466 ~~467~~	66 ~~78~~	777				"
467	84	1364				602 SKCK
468	65	290		88		604 JDRV
469	58	551				

west of Little Rockies
whose springs have
temperatures 22-32°C
72-90°F

Only highest gradients
listed

15 Mi to NORTH - 1 well

* 523	144	1087	62.2	331	170.7	604 EGL

not attractive

* not near population centers

MAST 5.7°C

MONT

Eastern Phillips Co

	#	°F	ft	°C	m	°C/km	FM
	528	65	872				604 BWDN
	531	68	1067				"
8 Mi N. Beaverton	619	95	950	35.0	290	101.0	604 BWDN
	620	76	1237				"
	640	72	1198				~~6~~ "
	641	100	1210	37.8	369	87.0	604 PLPS
	642	92	1332				604 BWDN

most attractive

33 MILES SOUTH LONE WELL

2 Mi N. Sun Prairie	379	245 (?)	5465	118.3°C	1666m	67.6°C/km	306 DPRW

not included in Malta Area

See attached sheet

MALTA

MONT

MAST 5.5°C

SHERIDAN Co

#	°F	ft	°C	m	°C/km	FM
692	263	10718	128.3	3267	37.6	203 RDRV
693	210	7108	98.9	2166		353 CRLS
732	180	6778	82.2	2066		"
733	184	6996	84.4	2132		354 RCLF
735	185	7060	85.0	2151		"
737	185	7183	85.0	2189		"
741	185	6983	85.0	2128		352 DWYR
785	168	6717	75.6	2047		354 RCLF
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]

6 mi W. of Westby ON RR

See PLENTYWOOD

Highest grade only

[redacted]

MONT

RICHLAND CO

MAST 5.5°C

	#	°F	ft	°C	m	°C/km	FM
9Mi WEST OF NOHLY	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	354 RCLF
	428	283	10460	139.4	3188	42.0	306 NSKU

highest only

MAST 41.9 °F
5.5 °C

MONT

CRESCENT SHAPE - NE MONT (96 wells)

#	°F	ft	°C	m	°C/km	FM
472	194	5804	90.0	1769	47.8	353CRLS
473	230	7398	110.0	2255	46.3	306NSKU
489	252	7422	122.2	2262	51.6	"
489	196	5355	91.1	1632	52.4	353KBBY
501	200	5904	93.3	1799	48.8	353CRLS
504	236	7684	113.3	2342	46.0	306 NSKU
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	353 KBBY
541	205	6428	96.1	1959	46.2	352MSNC

10 mi North Poplar

highest gradients only - listed

[redacted]

SEE WOLF POINT

MONT

NW Miles City - Custer Co.

MAST 44.9°F
7.2°C

#	°F	ft.	°C	m	°C/km	FM
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	602 MDDY
79	137	4752	58.3	1448 ✓	35.3 ✓	"
80	135	4659	57.2	1420	35.2 ✓	"
81	150	5106	65.6	1556	37.5 ✓	602 KOTN
94	144	4670	62.2	1423	38.6 ✓	602 MDDY
95	148	4788	64.4	1459 ✓	39.2 ✓	"
96	134	4598	56.7	1401	35.3 ✓	"

av 1449

All grads in area

MONT

MAST 7.2°C

45

NW EKALAKA, CARTER Co.

13 MI NW EKALAKA

#	°F	ft	°C	m		FM
23	138.	✓ 4485	58.9	1367		602 MDDY
24	135	✓ 4465	57.2	1361		"
25	139	✓ 4738	59.4	1444		"
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	603 GRNR
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	602 MDDY
897	165	✓ 4119	73.9	1255	53.1	"
915	126	4191		1277		"

143.4 4452 ft av 9502

av 1357

All grads in area

GRNR - Greenhorn fm

MONT.

NW

GARFIELD-ROSEBUD COS.

MAST 7.2 °C

	#	°F	ft	°C	m	°C/km	FM
	202	160	4878				402 TYLR
	~~205~~	158	4921				354 SSVD
	211	160	4824				"
16 mi N. Sumatra	218	168	4829	75.6	1472	46.5	402 TYLR
	220	165	4852	73.9	1479	45.0	"

highest grade only

MONT

MUSSELSHELL CO.

MAST 47°F
(ROUNDUP) 8.3°C

	#	°F	ft	°C	m	°C/km	FM
10 MI NW ROUNDUP	85	218	5874	103.3	1790	53.1 ✓	402 AMSD
13 MI NE ROUNDUP	117	140	3574	60.0	1089	47.5	402 TYLR
17 MI NE ROUNDUP	126	136	3450	57.8	1051	47.1	402 AMSD
2 MI NE MELSTONE	152	184	4984	84.4	1519	50.1 ✓	354 SSVD
2 MI E MELSTONE	153	169	5019	76.1	1530	44.3 ✓	"

highest grads only

MAST USE 47.0°F

MONT

SOUTHWESTERN MOST ANOMALOUS AREA

	#	°F	ft	°C	m	°C/km	FM
20 Mi W. BILLINGS	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	603FRNR
9Mi N RAPELJE	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	602CCRT
13 Mi SW RYEGATE	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	602LKOT
3 Mi S. ROTHIEMAY	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	402 AMSD

highest grade. only with minimum BHT of 120°F

MONT
central Hill Co.

MAST 42.3°F
5.7°C

#	°F	ft	°C	m	°C/km	FM
636	70	1115				604EGLE
637	70	840				604 JDRV
669	66	1065				604 EGLE
670	72	470	22.2	143	115.4	604CLGT
712	72	996				604EGLE
713	74	1268	23.3	386	45.6	〃

Highest temp 23.3°C

all BHT's less than 1 km

Combined with page 1
to form Havre area

MONTANA

ISOLATED HIGH GRADS

BHTs' > 120°F

County	#	°F	ft	°C	m	°C/km	MAST	FM
BLAINE	674 ①	126	1223	52.2	373	124.7	5.7	604 EGLE
CARTER	897 ②	165	4119	73.9	1255	53.1	7.2	602 MDDY
CARTER	939 ③	171	3289	77.2	1002	69.9	7.2	〃
POWDER R	1052 ④	182	4360	83.3	1329	57.3	7.2	〃

① 16 Mi NE OF CHINOOK

② 11 Mi W OF POWDERVILLE

③ 5 Mi SW OF CAPITEL

④ 3 Mi SW OF RANCHCREEK

379 & ~~[illegible]~~ 523 listed with anomalous areas

MONT

HAVRE AREA

MAST 42.3

		°F	ft	°C	m		°C/km	FM
	274	80	1385					603NBRR
	275	92	1316					604EGLE
	276	78	1382					EGLE
	306	65	427.					"
	307	65	574.					"
	336	65	1028.					"
	338	80	616.					604JVRD
	371	65	911.					"
	372	64	885.					EGLE
10691 ✓	373	85	2167					603 CRLL
	376	76	1062					EGLE
	377	91	554		168	elev 336C		"
	378	71	1143					"
	410	87	1242					"
	411	70	980					604 VRGL
	412	87	1679					EGLE
	413	89	1447					"
	414	89	1435					"
	415	75	1084					603CRLL
12422 ✓	444B	100	1796					604EGLE
	445	91	1514					"
	446	89	1654					EGLE
	447	71	991.					"
	448	100	815.					604 JDR
	451	98	1984.					EGLE
	452	75	1004.					"
	457	98	1906.					JDRV

	#	°F	ft	°C	m		°C/km	FM
✓	460	126	2350	52.2	716		64.9	604EGLE
	461	121	2261	49.4	689		63.4	"
OMIT	~~462~~	~~78~~	NO ~~DEPTH~~			~~OMIT~~		
	466	66	777					EGLE
14256 ✓	467	84	1364					602SKCK
	468	65	290		88	~~elev~~ 3400	78.3	604JDRV
	469	58	551					"
	480	64	580					604 JDRV
	481	90	1778					CRLL
	482	79	1321.					604VRGL
	484	90	1791.					EGLE
	485	61	851.					"
	519	70	1166.					EGLE
	520	89	1675.					604 JDRV
11367 ✓	521	97	2063.					EGLE
	523	144	1087.	62.2	331	elev 3532	170.7	EGLE
	526	75	1097.					"
	575	69	1270.					"
	578	69	1300.					"
	579	79	1028.					604VRGL
	581	67	1005.					EGLE
	608	65	906.					"
	609	71	995.					"
	610.	67	1040.					"
11791 ✓	614	70	914.					"
	615	75	1140.					"
	616	70	914.					"

(12)

#	°F	ft	°C	m		°C/km	FM
634	64	1083.					EGLE
635	67	1295.					"
636	70	1115.					604EGLE
637	70	840.					604JDRV
638	65	1052.					EGLE
669	66	1065.					EGLE
670	72	470.	22.2	143		115.4	604CLGT
712	72	996.					EGLE
713	74	1268.	23.3	386		45.6	"
~~715~~	~~131~~	~~3934~~					~~562STTH~~

12152

72,679

av 1172 ft = 357 m

MONT

MAST 42.3°F

MALTA

#	°F	ft	°C	m	°C/km	FM
527	68 ✓	1096			47.4	604BWDN
528	65 ✓	878				604 BWDN
530	68 ✓	1281				"
531	68 ✓	1067				BWDN
619	95 ✓	950	35.0	290	101.0	BWDN
620	76 ✓	1237				BWDN
621	71 ✓	1313				604 PLPS
640	72 ✓	1198				BWDN
641	100 ✓	1210	37.8	369	87.0	604 PLPS
642	92 ✓	1332		406		BWDN
643	64 ✓	1004				BWDN
644	86 ✓	1347				604PLPS

77.08

13907

av 1158.9 ft 353 m

353 m

7.8
93

MONT
WOLF POINT

#	°F	ft	°C	m	°C/km	FM
267	233	8360				309DVNN
278	164	5604				353KBB
279	168	6145.				354 RCLF
280	176	6303.				353CRLS
281	165	5854.				353KBBY
282	169	5851.				〃
284	158	5758.				〃
285	180	6254.				〃
287	180	6200.				〃
288	182	6464.				〃
289	184	6842				CRLS
290	176	6887				352MSN
292	200	7209.				CRLS
309	170	5760.				552 PIPR
311.	168	6050.				KBBY
312	189	6463.				354RCL
313	204	7184.				352MSN
314	194	6818.				CRLS
315	179	6463.				KBBY
316	194	6947.				CRLS
343	210	8510.				NSKU
380	218	7317.				CRLS
381	244	9330.				306DPR
417	196	7773.				NSKU
418	185	6539.				351MDS
472	194	5804.	90.0	1769	47.8	353CRL
473	230	7398.	110.0	2255	46.3	306NSK
487	252	7422.	122.2	2262	51.6	〃

62793 ✓

66623 ✓

n4.5

53555

	#	°F	ft	°C	m	°C/km	FM
	488	190	7376.				306 NSKU
	489	196	5355.	91.1	1632	52.4	353 KBBY
72824 ✓	497	155	4976				353 MIDL
	498	134	3799.				602 DKOT
	499	199	7490.				NSKU
	500	168	6312.				CRLS
	501	200	5904.	93.3	1799	48.8	353CRLS
	502	177	6250.				352MSNC
	503	172	5995.				353 CRLS
	504	236	7684.	113.3	2342	46.0	306 NSKU
	505	182	6326.				CRLS
	507	168	6147.				CRLS
60,883 ✓	508	185	6093.				CRLS
	509	232	5210.	111.1	1588	66.5	353KBBY
	535	138	4606.				353CRLS
	536	173	5552.				352MSN
	537	192	7258.				NSKU
	538	208	7252.				"
	539	160	5893.				351 MDSN
	540	180	5906.				"
	541	205	6428.	96.1	1959	46.2	352MSNC
	543	191	7728.				306 DPR
61926 ✓	544	190	7672.				NSKU
	545	191	7446.				"
	547	198	7614.				"
	548	202	7788.				"
	549	197	7600.				"

7740
7704

	#	°F	ft	°C	m		°C/km	FM
	550	204	7790.					306NSKU
	551	208	7564.					"
	552	200	7632.					"
	553	200	7704.					"
	554	202	7742					"
76552 ✓	558	188	7530.					"
	559	200	7628.					"
	582	160	5548.					353KBBY
	584	188	7563.					NSKU
	585	205	7731.					"
	586	199	7577.					"
	590	238	7664.					"
	591	175	6402.					353CRL
	592	204	7710.					306NSK
	593	196	7636.					"
72989 ✓	595	190	7668.					"
	596	192	7668.					"
	623	184	6001.					352MSN
	624	202	7709.					"
	677	150	5534.					353CRL
	678	181	5886					"
	728	186	5796.					"
	729	210	7279					NSKU
53541 ✓								

528131

av 6770' = 2064 m

MONT

MAST 55°C

PLENTYWOOD

#	°F	ft	°C	m	°C/km	FM
650	240	10252	115.6	3125	35.2	✓302 WPGS
692	263	10718	128.3	3267	37.6	✓203 RDRV
693	210	7108	98.9	2166	43.1	✓353 CRLS
694	186	7364	85.6	2244	35.7	✓354 RCLF
732	180	6778	82.2	2066	37.1	✓ CRLS
733	184	6996	84.4	2132	37.0	✓354 RCLF
735	185	7060	85.0	2151	37.0	✓ "
736	182	7200	83.3	2194	35.5	✓351 MDSN
737	185	7183	85.0	2189	36.3	✓ RCLF
741	185	6983	85.0	2128	37.4	✓352 DWYR
784	178	6878	81.1	2096	36.1	✓354 RCLF
734	165	7230			31.0	
				25,758	~~4080~~	

av 2342 m

ord, Dev & Miss

Note: 6 Mi W of Westby on RR
4 Mi. NE of Plentywood area.

#						
786	210	6634	98.9	2022	46.2	354 RCLF

RICHLAND CO MONT

MAST 5.5°C

20/24

#	°F	ft	°C	m		°C/km	FM
318	250	10815	121.1	3296.		35.1	302DSNB
346	256	10110	124.4	3082.		38.6	306NSKU
347	265	11018	129.4	3358.		36.9	302WPGS
382	212	8433	100.0	2570.		36.8	352MSNC
383	220	8719	104.4	2658.		37.2	11
384	256	9995	124.4	3046.		39.0	306 NSKU
385	225	9335	107.2	2845.		35.7	352MSNC
386	220	9038	104.4	2755.		35.9	353CRLS
419	224	8860	106.7	2700.		37.5	354RCLF
420	230	8995	110.0	2742.	29052	38.1	352MSNC
421	260	10470	126.7	3191.		38.0	306NSK
422	284	12378	140.0	3773.		35.6	203RDR
423	220	8774	104.4	2674.		37.0	354 RCL
424	255	10306	123.9	3141.		37.7	306 NSKU
427	270	8762	132.2	2671.		47.4	354 RCL
428	283	10460	139.4	3188.		42.0	306 NSK
430	266	11494	130.0	3503		35.5	259SLR
475	218	8757	103.3	2669.		35.6	353CRL
490	248	8494	120.0	2589.		44.2	354 RCL
492	214	8492	101.1	2588.		36.9	351 MDS

29987

59039

av 2952

302 Dev

306 Dev
351 miss
352 miss

353 miss
354 miss
203 Ord
259 Sil

MONT.

MAST 45°F
7.2°C

NW ROSEBUD Co.

#	°F	ft	°C	m	°C/km	FM
188	168	5714		1742	39.2	402 TYLR
192	149	5048 ✓			37.6	"
201	140	4842 ✓			35.8	"
202	160	4878 ✓			43.0	402 TYLR
203	146	4908 ✓			37.5	"
215	180	5590 ✓			44.0	"
216	174	5561 ✓			42.3	"
211	160	4824 ✓			43.4	354SSVD
212	141	4905 ✓			35.7	"
218	168	4829	75.6	1472	46.5	402TYLR
219	147	5020			37.0	354 OTT
220	165	4852	73.9	1479	45.0	402TYLR

60971

av 5081 = 1549 m

Pennsyl – 402 TYLR – Tyler fm
miss 354 SSVD
" 354 OTTR

MONT

MUSSELSHELL CO

MAST 47°F
8.3°C

#	°F	ft	°C	m		°C/km	FM
✓85	218	5874.	103.3	1790		53.1	402 AMSD
87	161	5758.	71.7			36.1	402TYLR
✓110	115	2581.	46.1	787		48.0	"
✓112	110	3032.	43.3			37.9	402 TYLR
✓113	111	3170.	43.9			36.8	354 SSVD
115	120	3720	48.9			35.8	402TYLR
✓116	118	3216.	47.8			40.2	"
✓117	140	3574	60.0	1089		47.4	"
118	132	3584.	55.6				354 SSVD
119	121	3755	49.4				TYLR
120	136	3812	57.8		38253 ✓		SSVD
121	122	3721.	50.0				TYLR
✓122	124	3516.	51.1			39.9	SSVD
✓126	136	3450.	57.8	1051'		47.0	402AMSD
127	137	4002	58.3				TYLR
128	132	3955	55.6				"
129	132	3953	55.6				"
130	131	3953	55.0				"
131	131	3962	55.0				SSVD
132	131	4000	55.0				TYLR
134	141	4260.	60.6		38324 ✓		"
135	132	4325	55.6				SSVD
136	120	3770.	48.9				552 PIP
137	138	4684	58.9				SSVD
138	146	4376	63.3				"
139	132	4252	55.6				"

#	°F	ft	°C	m		°C/km	FM
140	125	4236					402 TYLR
141	136	4214					"
142	138	4336					354 HETH
144	160	4678					402 AMSD
					43131 ✓		
145	154	4982					TYLR
147	139	4988					SSVD
149	156	4786					TYLR
✓152	184	4984	84.4	1519		50.1	SSVD
✓153	169	5019	76.1	1530		44.3	"
156	166	5302					TYLR
157	160	5306					"
167	107	2773					602 CCRT
168	124	3768					TYLR
169	134	3638					"
					45546 ✓		
170	110	2835					602 CCRT
172	124	3753					SSVD
173	125	3656					TYLR
					10244 ✓		
		175498					

43 av 4081 ft = 1244 m

402	Penn
354	miss
552	Jurassic
602	Cretace..

23
21

MONT - LAUREL

MAST 47.0°F

#	°F	ft	°C	m		°C/km	FM
3	73	664		202		71.4 ✓	604 CLGT
4	127	2618	52.8	798		55.7 ✓	603 FRNR
14	65	809				40.6 ✓	604 CLGT
22A	80	932				64.5 ✓	604 JDRV
→ 22B	95	2534				34.5 ✓	603 FRNR
27	90	2172				36.1 ✓	604 EGLE
28	81	1170				53.0 ✓	603 FRNR
29	130	2287	54.4	697		66.1 ✓	602 CCRT
30	112	2871				41.3 ✓	602 CCRT
31	88	1389			17446	53.8 ✓	602 MWRY
32	88	1715				43.6 ✓	603 FRNR
33	99	1187				79.8 ✓	"
39	120	3238	48.9	987		41.1 ✓	602 LKOT
46	88	1707				43.8 ✓	603 BGFK
63	85	1831				37.8 ✓	602 LKOT
73	98	2200				42.2 ✓	602 MWRY
→ ~~74~~	~~80~~	~~1908~~				~~31.5~~	~~553 MRSN~~
84	130	3260	54.4	994		46.4 ✓	402 AMSD
902	78	828				68.2 ✓	602 MDDY
90.3	70	710				59.0 ✓	"
909	102	1116			17792	89.8 ✓	604 TPCK
910	72	973				46.8 ✓	603 MSSR
920	70	834				50.3 ✓	604 BRPW
2	95	2855 37045			1807	30.6	

~~af~~ 1,683.86' = 513 m

53 ~~NW Rosebud~~ - MONT MAST 47°F

TOTAL

MUSSELSHELL

			°c/hr		
85			53.1		
87			36.1		
110			48.0		
✓ 111	98	2902	32.0		
112			37.9		
113			36.8		
115			35.8		
116			40.2		
117			47.4		
(53) 118	132	3584	43.2	410.5 1	
119	121	3755	35.9		
120	136	3812	42.6		
121	122	3721	36.7		
122	124	3516	39.9		
123	108	3496	31.8		
124	91	3764	21.3		
125	83	1980	33.1		
126			47.0		
127	137	4002	41.0		
128	132	3955	39.2	3685 2	
129	132	3953	39.2		397.9
130	131	3953	38.7		410.5
131	131	3962	38.6		368.5
132	131	4000	38.3		373.1
133	114	4012	30.4		366.0
134	141	4260	40.2		108.1
135	132	4325	35.8		38.19 avg for all

			°c/hr	
136	120	3770	35.3	
137	138	4684	35.4	
138	146	4376	41.2	3 3731
139	132	4252	36.4	
140	125	4236	33.6	
141	136	4214	38.5	
142	138	4336	38.2	
143	121	4400	30.6	
144	160	4678	44.0	
145	154	4982	39.1	
146	126	4722	30.5	
147	139	4988	33.6	
149	156	4786	41.5	4 3660
152			50.1	
153			44.3	
156	166	5302	40.9	
157	160	5306	38.8	
166	110	3603	31.9	
167	107	2773	39.4	
168	124	3768	37.2	
169	134	3638	43.6	
170	110	2835	40.5	
171	110	3678	31.2	5 3979
172	124	3753	37.4	
173	125	3656	38.9	
174	118	4069	31.8	360

MAST 5.5°C
41.9°F

RICHLAND Co.
Mont

#	°F	ft	°C/km			
318			35.1			
344	200	8507	33.9			
346			38.6			
347			36.9			
348	265	11948	34.0			
382			36.8			
383			37.2			
384			39.0			
385			35.7			
386			35.9	36.31,	363	1
419			37.5			
420			38.1			
24 421			38.0			
422			35.6			
423			37.0			
424			37.7			
425	210	9522	32.2			
427			42.0			
428			42.0			
429	266	12459	32.8	37.29 2	372	9
430			35.5			
475			35.6			
490			44.2			
492			36.9	38.0 2.4	152	2
					37.	0
					av	

COLO
STERLING

MAST 47°F

#	°F	ft	°C/km	
518			39.8	✓
519			36.7	✓
520	132	5252	29.5	✓
521			38.4	✓
522			37.0	✓
523			35.9	✓
524	138	5562	29.8	✓
525			361	✓
526			48.2	✓
527	178	7076	33.7	✓
528			38.9	✓
529			41.7	✓
544			450	✓
545			393	✓
546			374	✓
547	180	7336	33.0	✓
562			35.4	✓
563			42.5	✓
564	138	5274	31.4	✓
565	146	5507	32.8	✓
566			38.5	✓
570			35.3	✓
571			39.5	✓
23			av 37.2	

Colo-Akron

MAST 50°F

				°C/hr		
39				43.6		
40				36.4		
41				40.8		
42				35.7		
43	144	4945		34.6		
44	144	5110		33.5		
45	120	5260		24.2		
46				37.2		
47	150	~~0~~5314		34.3		
48	156	5928		32.6	10	35.29
49				35.9		
176	119	3956		31.8		
177				44.0		
178	130	4374		33.3		
179	136	4604		34.0		
180				41.6		
181				38.0		
182				36.4		
~~183~~						
184	134	4794		31.9		
185	133	4832		31.3	20	35
186	139	4990		32.5		
187	124	5300		25.4		
188				35.3		
189	148	5164		34.6		
190				34.6		
191				36.1		

22

Akron Cont

27

#				Π		
380	124	3927		34.3		
381	128	4144		34.3		
382				39.7		
383				36.2	30	34.30
384				39.1		
385	125	4126		33.1		
386				36.7		
387				38.1		
388				42.1		
389				39.6		
390	124	4632		29.1		
391				35.6		
419				37.7		
~~449~~	~~126~~	~~2167~~				
~~450~~	~~150~~	~~2874~~				
451				37.4	40	36.8
479				41.5		
542				46.2		
560				40.7		
603	135	4802		32.3		
604				35.7		
605				34.6		
606				35.2		
607	164	6118		34.0		
608	144	6318		27.1		
609				36.3	50	36.3
610	165	6470		32.4		

Akron Cont

27

				P	
611				34.7	
612	170	6420		34.1	
613	157	6543		29.8	
614	155	6422		29.8	
615	168	6401		33.6	
616	146	6318		27.7	
617				35.4	
618	168	6735		31.9	
649				36.8	60 32.62
650	134	4426		34.6	
651				34.7	
652				35.4	
653				40.1	
654	149	4841		37.3	
655				37.2	
656	144	5474		31.3	
657	125	5457		25.0	
658	145	5138		33.7	
659				35.1	70 34.44
660	140	5778		28.4	
661	167	6289		34.0	
662	168	6217		34.6	
663				36.1	
664				35.0	
665				35.7	
666				37.6	
667				35.2	

Akron Cont

#				r	
668				34.8	
669	180	6507		36.4	80 (34.78
728	152	4730		39.3	
729	152	4678		39.7	
730	148	4712		37.9	
~~733~~					
734				36.3	
735				40.3	
736	140	5410		29.9	
737				36.7	
738				36.0	
739				35.9	
740	168	6228		34.5	90 (36.65
741	150	6226		29.3	

14

35	29	✓
35	82	✓
34	30	✓
36	85	✓
36	36	✓
32	62	✓
34	44	✓
34	78	✓
36	65	✓

35.23

27
27
27
14
95

Wray

			oc/hm	
35			42.3	✓
36	100	3844	23.7	✓
37			41.9	✓
38			35.8	✓
174			35.1	✓
175	102	2934	32.3	✓
378			37.2	✓
379			41.0	✓
417			36.1	✓
418	107	3045	34.1	✓
449			50.1	✓
450			50.7	✓
601	94	2520	31.8	✓
602	91	2400	31.1	✓
640			38.4	✓
641			50.4	✓
642			55.0	✓
643	82	1878	31.0	✓
644			51.8	✓
645	87	1972	34.2	✓
646			47.5	✓
647	92	2351	32.6	✓
648	^		42.7	✓
713	99	1782	50.1	✓
714	90	1753	41.6	✓

wolf Point MAST 41.9

	°F	ft	Γ°C/km		
267	233	8360	41.7		
278	164	5604	39.7		
279	168	6145	37.4		
280	176	6303	38.8		
281	165	5854	38.3		
282	169	5851	39.6		
✓ 283	160	6784	31.7		
284	158	5758	36.8		
285	180	6254	40.2		
✓ 286	138	6438	27.2	10	37.14
287	180	6200	40.6		
288	182	6464	39.5		
289	184	6842	37.8		
290	176	6887	35.5		
✓ 291	170	7035	33.2		
292	200	7209	40.0		
~~[illegible]~~					
309	170	5760	40.5		
✓ 310	152	6527	30.7		
311	168	6050	38.0		
312	189	6463	41.5	20	37.73
313	204	7184	41.1		
314	194	6818	40.7		
315	179	6463	38.7		
316	194	6947	39.9		
380	218	7317	43.9		
381	244	9330	39.5		

	WP		°C/km		
417	196	7773			
418	185	6539	36.1		
472			47.8		
473			46.3		
487			51.6	— 30	42.56
488	190	7376	36.6		
489			52.4		
497	155	4976	41.4		
498	134	3799	44.2		
499	199	7490	38.2		
500	168	6312	36.4		
501			48.8		
502	177	6250	39.4		
503	172	5995	39.6		
504			46.0	— 40	42.30
505	182	6326	40.4		
✓ 506	171	10100	23.3		
507	168	6147	37.4		
508	185	6093	42.8		
509			66.5		
535	138	4603	38.0		
536	173	5552	43.0		
537	192	7258	37.7		
538	208	7252	41.7		
539	160	5893	36.5	— 50	40.73
540	180	5906	42.6		
541			46.2		
✓ 542	179	7249	34.5		
543	191	7728	35.2		

	wP				
544	190	7672	35.2		
545	191	7446	36.5		
✓546	182	7572	33.7		
547	198	7614	37.4		
548	202	7788	37.5		
549	197	7600	37.2	— 60	37.60
550	204	7790	37.9		
551	208	7564	40.0		
552	200	7632	37.8		
553	200	7704	37.4	—	
554	202	7742	37.7		
✓555	158	6659	31.8		
✓556	180	7624	33.0		
2 ✓557	180	7640	32.9		
558	188	7530	35.4		
559	200	7628	37.8	— 70	36.17
582	160	5548	38.8		
✓583	144	5417	34.4		
584	188	7563	35.2		
585	205	7731	38.4		
586	199	7571	37.8		
✓587	166	7576	29.8		
✓588	174	7519	32.0		
✓589	190	7772	34.7		
590	238	7664	46.6		
591	175	6402	37.9	— 80	36.56
592	204	7710	38.3		

	WP		°C/hr		
593	196	7636	36.8		
✓594	185	7594	34.3		
595	190	7668	35.2		
596	192	7668	35.7		
623	184	6001	43.2		
624	202	7709	37.8		
✓678	181	5886	43.1		
728	186	5796	45.3		
729	210	9279	42.1	— 90	39.18

9
26
29
27
89

37.14
37.73
42.56
42.30
40.73
37.60
36.17
36.56
39.18

34997

av 38.88

HAVRE – TOTAL WELLS

MAST 42.3

	274	80	1385		
	275	92	1316		
	276	78	1382		
	306	65	427		
	307	65	574		
	336	65	1028		
✓	337	68	1417		
	338	80	616		
	371	65	911		
	372	64 72.2	885	994.1	54.82 °C/km
	373	85	2167		
✓	374	74	1689		
✓	375	72	1685		
	376	76	1062		
	377	91	554		
	378	71	1143		
	410	87	1242		
	411	70	980		
	412	87	1679		
	413	89 80.2	1447	1364.8	50.62 °C/km
	414	89	1435		
	415	75	1084		
✓	443	70	1560		
✓	444 A	77	1882		
	444 B	100	1796		
	445	91	1514		
	446	89	1654		

	447	71	991		
	448	100	815		
✓	449	71 83.3	2185	2491.6	50.10 °C/km
✓	450	69	1581		
	451	98	1984		
	452	75	1004		
✓	453	65	1202		
✓	454	87	2546		
✓	455	76	2466		
✓	456	81	2457		
	457	98	1906		
✓	458	83	2704		
✓	459	74 80.6	2165	2001.5	34.88 °C/km
	460	126	2350		
	461	121	2261		
	~~462~~	~~omit~~			
✓	463	72	1952		
✓	464	70	2415		
✓	465	84	2304		
	466	66	777		
	467	84	1364		
	468	65	290		
	469	58	551		
	480	64 81.0	580	1484.4	47.5~~2~~5 °C/km
	481	90	1778		
	482	79	1321		
✓	483	65	2420		

Hawn

484	90	1791	
485	61	851	
✓ 494	60	1595	
✓ 519	70	1166	
520	89	1675	
521	97	2063	
523	144 84.5	1087 1574.7	6 48.85 °C/km
526	75	1097	
575	69	1270	
578	69	1300	
579	79	1028	
✓ 580	77	2515	
581	67	1005	
608	65	906	
609	71	995	
610	67	1040	
614	70 70.9	914 1202.0	7 43.19 °C/km
615	75	1140	
616	70	914	
✓ 617	57	1230	
634	64	1083	
635	67	1295	
636	70	1115	
637	70	840	
638	65	105.2	
669	66	1065	
670	72 67.6	470 1020.4	8 45.19
712	72	496	
713	74 73	1268 [illegible]	49.43

54.82
50.62
50.10
34.88
47.52
48.85
43.19
45.19

3 7517

av	46.90 ×80	375
	49.43 ×2	98

3850

av 46.96

av all

NEW MEXICO

MAST 49.7

		°F	ft	°C/km Γ	
	9	154	4618	41.2	✓
	14	—	5582	36.0	✓
	15		5474	38.0	✓
	16		5162	37.5	✓
	17	202	8521	32.6	
	22	—	6395	35.4	✓
	23		3929	38.2	✓
	24	115	3349	35.6	✓
	25		5518	38.4	✓
	27	168	6333	34.0	
	28		4154	40.5	✓
X	29	83	1548	39.2	✓
	30	—	4448	39.4	✓
	31	160	5999	33.5	
	32		6059	35.6	✓
	15				

ave grad = 37.0

av anom
depth 4686.3

1428

NEW MEXICO

SOUTHERN SAN JUAN BASIN

LITTLE BLUE MESA

MAST 9.8 °C
49.7°F

all PI

#	°F	ft	m	°C/km	FM
9	154	4618		41.2	553 ENRD
14	160	5582.		36.0	"
15	164	5474.		38.0	553 MRSN
16	156	5162.		37.5	553 ENRD
22	174	6395.	1949	35.4	602 DKOT
23	132	3929	1198	38.2	603 GLLP
25	166	5518.		38.4	553 ENRD
28	142	4154.		40.5	603 GLLP
30	146	4448		39.4	"
32	168	6059.		35.6	553 MRSN

51339

av 5134' = 1565 m

most attractive (?)

Little Blue Mesa area

553 Jurassic
602 Cretaceous
603 "

NEW MEXICO

ISOLATED HIGH GRADS.

BHT's > 112°F

County	#	°F	ft	°C	m	°C/km	°C MAST	FM
① Bernalillo	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	13.7	603 MNCS
② San Juan	110	113	1530	45.0	466	75.5	9.8	603GLLP
③ Rio Arriba	106	253	MAXIMUM* 8400	122.8	2560	44.1	9.8	?
④ Eddy	[redacted]	195	4379	90.6	1335	56.5	15.2	453DLWR
	~~#29~~							[illegible]
⑤ Lea	[redacted]	174	4192	78.9	1278	49.8	15.2	453QUEN

① 20 Mi W. of ALBUQUERQUE
② 23 Mi NW. of FARMINGTON
③ 16 Mi W. of NAVAJO
④ 12 Mi NE of CARLSBAD
⑤ 12 Mi N. of HOBBS

[redacted]

*Depth entry omitted in PI file
8400' is total depth after plug back

COLO

WALSENBURG

MAST-USE 50°F.
10°C

all PI

#	°F	ft	°C	m	°C/km	FM
						603 CDLL
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	~~406 TOPK~~
~~1056~~	~~125~~	~~3326~~	~~51.7~~	~~1014~~		602 DKOT
1048	170	6184	76.7	1885	35.4	"
1052	158	5427	70.0	1654	36.3	553 ENRD
~~1041~~	~~140~~	~~3434~~	~~60.0~~	~~1047~~		553 ENRD
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	

8 mi NW WALSENBURG

UPPER REACH HUERFANO R,

10 mi NW GARDNER

[redacted]

COLORADO

✓ 041	140	3434	60.0	1047	47.8 ✓	553 ENRD
044	108	2418	42.2	737	43.7 ✓	603 CDLL
✓ 047	119	2903	48.3	885	43.3 ✓	603 CDLL
✓ 048	170	6184	76.7	1885	35.4 ✓	602 DKOT
049	112	3063	44.4	934	36.8 ✓	603 CDLL
050	190	7734	87.8	2357	33.0 ✓	602 DKOT
051	118	3419	47.8	1042	36.3 ✓	602 DKOT
✓ 052	158	5427	70.0	1654	36.3 ✓	602 DKOT
055	104	1680	40.0	512	58.6	652 RTON

COLO

MAST 43.8 °F
6.6°C

NORTHERN MOFFAT CO

HIAWATHA

all PI

#	°F	ft	°C	m	°C/km	FM
596	168	5595		1705	40.5	651 FRUN
AT HIAWATHA → [redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	652 WSTC
600	163	5929		1807	36.6	604 MVRD

[redacted]

596
599
600

av grad 36.6

582	150	5774

Γ

597	30.5
598	27.8

COCO

WESTERN CHEYENNE Co

MAST
USE 50°F
10°C

all PI

	#	°F	ft	°C	m	°C/km	FM
	906	155	4886	68.3	1489	39.1 ✓	405 MSSR
6MI NE WILD HORSE	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	452 LYNS
	922	145	4303	628	1312	40.2 ✓	406 VRGL
7MI SE WILD HORSE	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	451 WFMP
	933	146	4118	63.3	1255	42.5 ✓	"

6225

OK

[redacted]

Pennsylvanian	405	MSSR	Missouri
Permian	452	LYNS	LYONS
Pennsylvanian	406	VRGL	VIRGIL
Permian	451	WFMP	WOLFCAMP

COLO LIMON

EASTERN ELBERT CO

MAST USE 50°F 10°C

all PI

5 mi West Cedar Pt. →

#	°F	ft	°C	m	°C/km	FM
862	162	5634				602 DKOT
~~863~~	~~159~~	~~5340~~				"
870	130	3099	54.4	944	47.0	604 SSSX
871	167	5954				602 DKOT
✓ 869	162	5280				602 DKOT
~~868~~	~~161~~	~~6120~~				451 CHSE
✓ 878	168	5493				602 DKOT
887	170	6167				"
888	176	6254				"

most attractive

	°F	ft	°C	m	°C/km	
862	162	5634			36.2 ✓	602DKOT
869	162	5280			38.7 ✓	602 DKO
870	130	3099	54.4	944	47.0 ✓	604 SSS
871	167	5954			35.8 ✓	602 DKO
878	168	5493			39.1 ✓	602 DK
887	170	6167			35.5	602 DK
888	176	6254		1906	36.7	602 DK

37881

av

5412

COLO

LOGAN & WELD COS.
NW STERLING

MAST 47°F
8.3°C

All PI

#	°F	~~ft~~	°C	~~m~~	°C/km	FM
526	228	6855	108.9	~~2089~~	48.1	602DKOT
544	182	5466	83.3	1666	45.0	"
~~545~~	~~168~~	~~5621~~				"

Two largest grads listed

STERLING anomolous wells

#	°F	ft	°C	m	°C/km	FM
518	154	4909	67.8	1496	39.8	602DKOTD
519	150	5128	65.6	1563	36.7	DKOTJ
521	158	5275	70.0	1608	38.4	DKOTJ
522	153	5225	67.2	1592	37.0	DKOTJ
523	160	5735	71.1	1748	35.9	DKOTD
525	158	5611	70.0	1710	36.1	DKOTD
526	228	6855	108.9	2089	48.2	DKOTJ
528	200	7169	93.3	2185	38.9	DKOT
529	208	7034	97.8	2144	41.7	DKOT
544	182	5466	83.3	1666	45.0	DKOT
545	168	5621	75.6	1713	39.3	DKOT
546	200	7446	93.3	2270	37.4	UNKN
563	170	5274	76.7	1608	42.5	DKOT
566	174	6021	78.9	1835	38.5	DKO
570	197	7747	91.7	2361	35.3	DKO
571	210	7530	98.9	2295	39.5	DKO
562	152	5406	66.7	1648	35.4	DKOT

BERTHOUD

COLO

MAST 47°F
8.3°C

SE LARIMER CO.

	#	°F	ft	m	°C/km	FM
7 mi S. LOVELAND	[redacted]	124	3123	952	44.9	603 TMPS
	436	120	3064	934	43.4	603 NBRR
	437	136	3632	1107	44.6	602 DKOT

[redacted]ttractive (?)

✓	435			
✓	436			
✓	437			
	438	110	3720	30.9

COLO

~~South~~ OF WALDEN
~~(N. Hot Sulphur Sprgs)~~

WALDEN

~~TOO HIGH~~
~~BUT USE~~

MAST 47.0°F
= 8.3°C

all PI

#	°F	ft	°C	m	°C/km	FM
405	160	5600	71.1	1707	36.8	654 MOCN
457	105	2789	40.6	850	38.0	603 FRNR
491	164	5722	73.3	1744	37.3	?
492	172	6475	77.8	1974	35.2	603 NBRR
~~553~~	~~166~~	~~6228~~			~~34.8~~	~~553 MRSN~~

✓405						
406	153	5383	67.2	1641	35.9	602 LKOT
✓457						
474	184	7390	84.4	2252	33.8	604 PIRR
✓491						
✓492						
493	158	6475	70.0	1974	31.2	603 NBRR
535	100	887	37.8	270	10.9.2	603 FRNR
536	92	3348	33.3	1020	24.5	602 DKOT

COCO

SOUTH OF CRAIG

MAST 6.6 °C

all PI CRAIG

#	°F	ft	°C	m	°C/km	FM
441	112	3040	44.4	926	40.8	602 DKOT
~~458~~	158	4848	70	1478	42.9	"
460	139	4284	59.4	1306	40.4	"
477	122	4001	50.0	1212	35.8	604 MRPS

best (?)

	439	100	3848	37.8	1173	26.6	
	440	116	3475	46.7	1059	37.9	?
✓	441						
✓	458						
	459	140	5015	60.0	1528	34.9	
✓	460						
✓	477						
	478	194	7910	90.0	2411	34.6	

COLO

RANGELY

MAST 6.6°C

all PI.

#	°F	ft	°C	m	°C/km	FM
362 ✓	138	3778	58.9	1152	45.4 ~~53.4~~ ✓	602 DKOT
363 ✓	136	3692	57.8	1125	45.5 ✓	602 DKOT
371 ✓	108	3290	42.2	1003	35.2 ✓	553 MRSN
372 ✓	120	2958	48.9	902	46.9 ✓	602 DKOT
[redacted] ✓	138	3674	58.9	1120	46.7 ✓	602 DKOT
373 ✓	121	3108	49.4	947	45.2 ✓	602 CDMN ~~602 DKOT~~
~~766~~	~~252~~	~~9912~~	~~122.2~~	~~3021~~	~~38.3~~	~~553 MRSN~~
167 ✓	134	4108	56.7	1252	40.0 ✓	553 MRSN
169 ✓	136	3984	57.8	1214	42.2 ✓	602 DKOT
170. ✓	153	5489	67.2	1673	36.2 ✓	503 SRMP
171. ✓	138	3802	58.9	1159	45.1 ✓	602 DKOT
168 ✓	145	3672	62.7	1119	50.1 ✓	?
172 ✓	148	6602	64.4	2012	28.7 ✓	419 WEBR
361 ✓	147	6517	63.9	1986	28.8 ✓	419 WEBR
360 ✓	148	6646	64.4	2025	28.5 ✓	419 WEBR

ISOLATED WELL 33 mi NW RANGELY

#	°F	ft	°C	m	°C/km	
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]

[redacted]

COLO

NEAR RICO

MAST 43.8 °F
= 6.6 °C

All PI

#	°F	ft	°C	m	°C/km	FM
13	144	4063	62.2	1238	44.9	404 DRCK
16	214	6644	101.1	2025	46.7	359 MSSP
20	170	6168	76.7	1880	37.3	404 DRCK
21	220	5893	104.4	1796	54.4	409 HRMS

Dunton - Rico Hot Spgs area

geothermometer 58°C
surface 28-46°C

10¹
58
43

COLO

MAST
USE 50° F
10°C

LARGEST AREA
NE COLO

NORTHEAST AREA

All PI

#	°F	ft	°C	m	°C/km	FM
39	146	4010	63.3	1222	43.6	602 DKOT
449	126	2767	52.2	843	50.0	603 NBRR
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	603 NBRR
[redacted]	[redacted] 144	3896	395.6 62.2	1188	[redacted] 43.9	602 DKOT
542	147	3830	63.9	1167	46.2	602 DKOT

3 mi west Wages →

5 mi west Yuma

[redacted]

~~Some shallow wells near Wray (< 1 km)~~
BHT's up to 108°F

DIVIDE INTO 2 AREAS
AKRON – DEEP
WRAY – < 1 km
↑
attached

COCO

PICEANCE CREEK
S-CENTRAL RIO BLANCO CO

43.8 °F

all PI

MAST 6.6°C

#	°F	ft	°C	m	°C/km	FM
702	146	5204	63.3	1586	35.7	652 WSTC
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	652 WSTC
706	142	4204	61.1	1281	42.5	"

[redacted]

700	217	8940	102.8	2725	35.3	604 MVRD
701	132	4664	55.6	1422	34.4	652WSTC
✓ 702						
✓ 705						
✓ 706						

COLORADO

ISOLATED HIGH GRADS.
BHT's >120°F

County	#	°F	ft	°C	m	MAST °C	°C/km	FM
Mesa	[redacted] ①	138	3749	58.9	1143	6.6	45.8	604 CRCR
Rio Blanco	[redacted] ②	206	6373	96.7	1942	6.6	46.4	553 MRSN
Rio Blanco	[redacted] ③	159	2772	70.6	845	6.6	75.7	604 CSLG
~~Garfield~~	~~823~~ ④							
Garfield	[redacted] ⑤	140	2749	60.0	838	6.6	63.7	604 MVRD

① 26 Mi E. GRAND JCT.

② S. Western Rio Blanco Co

③ "

~~④~~

⑤ 32 Mi N. Grand Jct

4/13 listed with Rangely

[redacted] - [redacted] - AKRON AREA - DEEP

DIVIDE INTO 2

COLORADO

NORTHEAST AREA

n=66

MAST 50°F
10°C

#	°F	ft	°C	m	°C/km	FM
35	106	[redacted]	41.1		[redacted]	[redacted]
37	106	[redacted]	41.1		[redacted]	[redacted]
38	98	[redacted]	36.7		[redacted]	[redacted]
39	146	4010 ✓	63.3		43.6	602DKOT
40	146	4802	63.6		36.4	"
41	160	4914	71.1		40.8	"
42	148	5004	64.4		35.7	"
46	158	5285	70.0		37.2	"
49	164	5793	73.3		35.9	"
174	105	[redacted]	40.6		[redacted]	[redacted]
P. 177	144 ? (744)	3896	? 395.6		44.0 (324.7)	DKOT
180	154	4558	67.8		41.6	"
181	149	4744	65.0		38.0	"
182	147	4862	63.9		36.4	"
188	150	5159	65.6		35.3	"
190	160	5791	71.1		34.6 X	"
191	169	6014	76.1		36.1	"
378	110	[redacted]	43.3		[redacted]	[redacted]
379	115	[redacted]	46.1		[redacted]	[redacted]
382	140	4130	60.0		39.7	DKOT
383	130	4025	54.4		36.2	"
384	136	4006	57.8		39.1	"
386	139	4420	59.4		36.7	"
387	144	4500	62.2		38.1	"
388	150	4333	65.6		42.1	"
389	149	4561	65.0		39.6	"
391	146	4908	63.3		35.6	"

#	°F	ft	°C	°C/km	FM
417	110	[redacted]	43.3	[redacted]	[redacted]
419	132	3960	55.6	37.7	DKOT
449	126	[redacted]	52.2	[redacted]	[redacted]
450	130	[redacted]	54.4	[redacted]	[redacted]
451	142	4482	61.1	37.4	DKOT
479	138	3868	58.9	41.5	"
542	147	3830	63.9	46.2	"
560	127	3446	52.8	40.7	"
604	155	5358	68.3	35.7	"
605	158	5683	70.0	34.6	"
606	157	5530	69.4	35.2	"
609	175	6274	79.4	36.3	"
611	171	6350	77.2	34.7	"
617	183	6853	83.9	35.4	"
640	96	[redacted]	35.6	[redacted]	[redacted]
641	98	[redacted]	36.7	[redacted]	[redacted]
642	100	[redacted]	37.8	[redacted]	[redacted]
644	96	[redacted]	35.6	[redacted]	[redacted]
646	108	[redacted]	42.2	[redacted]	[redacted]
648	105	[redacted]	40.6	[redacted]	[redacted]
649	139	4410	59.4	36.8	DKOT
? 651	140	4721	60.0	34.7	"
? 652	142	4732	61.1	35.4	"
663	155	4770	68.3	40.1	"
655	162	5491	72.2	37.2	"
659	162	5812	72.2	35.1	"
663	173	6217	78.3	36.1	"

#	°F	ft	°C	°C/km	FM
664	171	6292	77.2	35.0	DKOT
665	173	6277	78.3	35.7	"
666	177	6156	80.6	37.6	"
667	178	6634	81.1	35.2	"
668	175	6540	79.4	34.8	"
713	99	[redacted]	37.2	[redacted]	[redacted]
714	90	[redacted]	32.2	[redacted]	[redacted]
734	155	5268	68.3	36.3	DKOT
735	162	5065	72.2	40.3	"
737	172	6053	77.8	36.7	"
738	169	6032	76.1	36.0	"
739	170	6090	76.7	35.9	"

66 —

[redacted]

av 2350' 716m Wray

av 5,143 1568m Akron

UTAH MAST 10.6°C

Northeast Grand Co.

PI	°F	ft	°C	m	10.6 °C/km	24.2 °C/km
#104	180	3235	82.2	986	72.6	58.8

in 604 CSLG
Cretaceous Castlegate
gas production

DOGM

#	°F	ft	°C	m	10.6 °C/km	24.2 °C/km
90	121	2347				
109	120	2350				
181	120	2530				
171	163	2844	72.8	867	71.7	56.ⓐ
378	125	3155				
146	128	3208				
120	150	3603				
71	169	4607				

10 mi NW Thompson

not confidential

Jock Campbell
United Energy Corp SWSW Sec 29, 20S-22E

°F	ft
163	2844

SAME

28-42 GPM

Confidential [illegible]

UTAH
near Price

MAST 51.°F = 10.6°C

4 mi. SW PRICE → 20

all	DOGM					MAST 24.2
	°F	ft	°C	m	°C/km	
19	90	1600	32.2	488	44.3	16.4
20	104	1994	40.0	608	48.4	26.0
21	80	1490	26.7	454	35.5	5.5
22	85	1214	29.4	370	50.8	14.0

most attractive

UTAH

Rich Co Anomaly MAST - 43°F = 6.1°C

	PI	°F	ft	°C	m	°C/km	FM
	8	98	2040	36.7	622	49.2✓	553 TCRK
	142	86	346	30.0	105	227.6✓	553 TCRK
T11N-R7E	143 (See below)	—	—				?
	144	~~161~~ 157	~~9466~~ 7418	69.4	2261	28.0	503 CHNL

DOGM

1 this is same as 143 above

at town of Randolph

°F	ft	°C	m	°C/km
[redacted]	[redacted]	[redacted]	[redacted]	[redacted]

553 TCRK = Twin Creek - Upper Jurassic

503 CHNL Chinle - Upper Triassic

all west front Crawford Mtns
no water production noted.

[redacted]

X [redacted]ST
E.1/10. [redacted] AST

UTAH
northeast Carbon Co.

USE MAST 10.6
MAST 45°F = 7.2°C

		°F.	ft	°C	m	Γ °c/km	MAST 24.2	FM
	PI							
NEAR X	(?)95	155	4919	68.3	1499	38.4 ~~40.8~~	29.4	652 WSTC
	96	145	1329	62.8	405	128.9 ~~137.2~~	← 95.3	659 TRTR
	DOGM							
—	~~366~~	~~87~~	~~3040~~					
☒ ✓	367	115	2294	46.1	699	50.8	←	
x ✓	368	110	3275	43.3	998	32.8	←	check
—	~~369~~	~~92~~	~~3535~~					
x ✓	370	112	3052	44.4	930	36.4	←	
x ✓	371	116	3097	46.7	944	38.2	←	
NOT A ✓	372	no temp						
NOT A ✓	373	110	4984	43.3	1519	21.5		
✓	376	102	4753	38.9	1449	19.5		
x	377	132	4982	55.6	1518	29.6		
x ✓	378	125	3155	51.7	962	42.7	←	
● ✓	379	100	2128	37.8	649	41.9	←	
	~~380~~	~~105~~	~~22 6~~					11 TOTAL
x	[illegible]							
x	360	121	5630	49.4	1716	22.6		
x.	363	150	6395	65.7	1949	28.2		
x	364	304	17259	~~151.1~~	5260	26.7		
ⓧ ✓	365	134	5288	56.7	1612	28.6		

most attractive

>150°C

UTAH

ISOLATED HIGH GRADS.

BHT's > 120°F

PI

County	#	°F	ft	°C	m	°C/km	°C MAST	FM
① Uintah	54	200	5458	93.3	1664	49.7	10.6	404 DRCK
② Duchesne	82	185	5602	85.0	1707	45.6	7.2	652GRRV
③ Uintah	91	142	2930	61.1	893	60.4	7.2	"
④ Uintah	94	145	1286	62.8	392	141.8	7.2	659 TRTR
⑤ San Juan	154	125	2610	51.7	796	51.6	10.6	419 CTLR

DOGM

	County-No.	°F	ft	°C	m	°C MAST	°C/km
⑥	BOX ELDER-1	142	3500	61.1	1067	9.4	48.4
⑦	SUMMIT-11	160	4124	71.1	1257	6.1	51.7
⑧	TOOELE-2	193	5343	89.4	1628	10.0	48.8
⑨	UINTAH-88	128	524	53.3	160	7.2	288.1

① S. EDGE County.

② 12 MI S. OF MYTON

③ 2 MI W. OF LITTLE BONANZA

④ 16 MI S. OF OURAY

⑤ 4 MI NE OF LA SAL

⑥ ROSSEL PT GR. SALT LAKE

⑦ 20 MI W. OF MANILA

⑧ 4 MI SE OF BURMESTER

⑨ 12 MI W. OF RAINBOW

PETERS POINT MAST 45° F, 7.2°C

WELLS IN ANOM. AREA (using low MAST)

#	°F	ft	°C	m	°C/km
PI-96	145	1329		405	137.1 ✓ LOOK UP DOGM?
366	87	3040			25.2 ✓
367	115	2294			55.6 ✓
368	110	3275			36.1 ✓
369	92	3535			24.2 ✓
370	112	3052			40.0 ✓
371	116	3097			41.7 ✓
373	110	4984		1519	23.8 ✓
376	102	4753			21.8 ✓
378	125	3155			46.2 ✓
379	100	2128			47.1 ✓
	av	3149		960	

Northeast Carbon County

IDAHO

ALL WELLS IN PI FILE

MAST 10°C

County	#	°F	ft	°C	m	°C/km	FM
Payette	①	382	7406	194	2257	81.5	?
Elmore	②	372	8914	189	2717	65.9	109? PCMB

① near Vales HS KGRA

② near Mtn. Home

NEVADA

ISOLATED HIGH GRADS

	County	#	°F	ft	°C	m	°C/km	MAST °C	FM
①	Nye	2	164	3796	73.3	1157	54.4	10.4	302SMNS
②	Nye	5	169	4853	76.1	1479	44.4	10.4	000VLCS
③	NYE	14	214	6598	101.1	2011	45.1	10.4	659 TRTR
④	WHT. PINE	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	[redacted]	7.7	302 SMNS
⑤	"	17	181	5190	82.8	1582	47.5	7.7	000 UNKN

① RR VALLEY
② RR VALLEY
③ RR VALLEY – 10 MI S. OF CURRANT
④ 3 MI. SE OF EAST ELY
5 7 MI. N. OF McGILL

Author: Applegate ?

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Internal use only

AREA
USwest
GtBasin
Gthm

SUBREGION DESCRIPTIONS

SUBREGION I: SNAKE RIVER PLAIN

Subregional Setting

The Snake River Plain and the Yellowstone volcanic field constitute a major young volcanic province extending in a broad arc from the Idaho-Oregon state line eastward across Idaho to Yellowstone National Park and vicinity (Figure 1). The common geologic element in the region is the volcanic activity as indicated by the young basaltic rocks at Craters of the Moon in the eastern Snake River Plain and the massive young rhyolitic volcanic deposits and associated basalts of the eastern Snake River Plain and Yellowstone.

The Snake River Plain is an area of generally low relief which is covered by basaltic lava flows interbedded with young, flat-lying river and lake deposited sedimentary rocks. The plain contains a large proportion of the state of Idaho's irrigated agricultural lands, and most of its population centers.

The northeastern portion of the plain is terminated by a generally circular, forested, silicic volcanic feature, the island Park Caldera. The island Park Caldera borders the higher plateaus and mountains of Yellowstone National Park immediately to the east. Radiometric dating indicates that the volcanism has progressed eastward along the plain toward Yellowstone.

The Eastern plain has been characterized as a downwarp, and geophysical surveys indicate 3 to 5 Kws of sedimentary and volcanic fill within the trough. The primary recharge area for the Snake Plains Aquifer is the high snowfall region in the island park area. The outflow area is at Thousand Springs near Buhl in the canyon of the Snake River. Scattered young silicic volcanic centers are present both within the Eastern plain and marginal to it within the Blackfoot Volcanic Field.

The Western plain has been described as a rift valley. Volcanism in the Western plain is older than that of the Eastern plain and no specific prospects based on young rhyolitic volcanism have been identified.

Economy of the Subregion

The agricultural emphasis in both the eastern and western Snake River Plain is on potatos and sugar beets. Associated food processing industries are a major agribusiness element of the subregion. Alfalfa and grain are also of agricultural significance in the subregion. The Island Park area supports an active timber industry.

The most significant mineral industry in the region is also agriculture-related. Extensive phosphate deposits occur in the region marginal to the Snake River Plain in southeastern Idaho and significant phosphate processing industry is centered at Pocatello.

Energy Production and Consumption

At present, wnergy generation within the subregion is dominantly hydroelectric, with extensive generation capacity developed by the Idaho Power Company from reservoirs along the Snake River, and imported electric power from the Columbia River system provided by the Bonneville Power Administration. Idaho Power Company and rural electric coops are seriously considering alternative electrical generation alternatives as the hydroelectric generating capacity of the Snake River system has been developed to near-capacity. Coal-fired plants have been proposed to meet projected requirements but specific projects have been rejected on environmental grounds.

The major supplier of energy for direct heat uses in the subregion is the Intermountain Gas Company. The company is agressively investigating geothermal markets, many of which (as at Boise) have traditionally used geothermal energy for direct heat applications.

Geothermal Potential

Confirmed geothermal resources of the area are of low and moderate temperature, and are presently being exploited for a variety of direct heat uses. Known geothermal resources within the subprovince are primarily associated with normal faulting along the margins of the Snake River plain.

Young silicic volcanic rock within the Eastern plain and the high temperature geothermal systems at Yellowstone National Park indicate the potential for igneous-related high temperature resources in the Eastern plain. The high flow rate Snake Plains Aquifer is responsible for obscuring the high heat flow that may be associated with such hidden resources.

Moderate temperature hydrothermal systems are indicated by geochemical thermometry for the Western Snake River plain. High temperatures (~200°C) have been reported from deep exploration wells in the area, but no high temperature fluid production has been confirmed.

The fracture zones paralleling the northwestern trend of the plain contain at least moderate temperature geothermal resources along both the northern and southern flanks of the Western Snake River plain. The Bruneau-Grandview area adjacent to the plain on the south is notable in this respect. At least a 12 by 60 mile area contains hot fluids at depths of 1,000 to 3,000 feet. The region may ext end from Twin Falls near the easternmost portion of the Western plain more or less continuously to the Oregon state line near Vale.

Strategy

Based on existing resource data, near-term geothermal utilization within the subregion will be primarily a continuation and expansion of direct heat applications of the low and moderate temperature hydrothermal resources. The Technical Initiatives Program (TIP) insures that potential users are made more fully aware of the geothermal potential existing throughout the subregion and, together with the PON program for direct heat application, will maximize the replacement of fossil fuel used to create low grade energy with geothermal resources. The Midterm electrical generation goal of 8,000-9,500 MWe from high temperature resources will be largely dependent on the successful exploration for the hidden resources of the deep Snake River plain. Impediments to industry exploration of the subregion without government support include both the masking of the deep thermal situation by the cold Snake Plains Aquifer and the drilling difficulties anticipated in drilling through the volcanic sequence. A combination of ongoing USGS and DOE-supported assessment, combined with cost-shared drilling with industry, is expected to establish the high temperature geothermal potential of the Snake River plain by 1982.

The midterm utilization of the extensive moderate temperature resources for electrical power generation will depend on the successful development of moderate temperature electrical generation technology at Raft River.

SUBREGION II - NORTHERN ROCKY MOUNTAINS

Subregional Setting

The Northern Rocky Mountain Subregion (Figure 1) is a mountainous area characterized by rugged topography, extensive forests and low population density. The subregion is here defined as those portions of Montana and Northern Idaho characterized by the presence of batholiths and folded mountain ranges exclusive of the young volcanic provinces of the Yellowstone and Snake River plain.

The combined Idaho and Boulder Batholiths comprise much of central Idaho and southwestern Montana. The area geologically is dominated by batholithic complexes of intermediate to silicic rocks of Cretaceous age (90MYBP). The batholiths are similar in composition and age to the Sierra Nevada Batholith.

The southern and western margins of the Idaho Batholith contain faulted sediment basins such as Little Camas Prairie, which provide ideal reservoir conditions with approximately 2 KM of sedimentary fill.

In addition to the batholithic ranges, the subregion contains north-trending, folded, sedimentary mountains such as the Sawtooth Range in central Idaho.

Economic Parameters

As the subregion is dominantly dorested land, the economy is based on forext products, tourism, agriculture and hard-rock mining. Population density is low and much of the subregion is composed of National forests, wilderness areas and primative areas.

Energy Production and Consumpsion

Electrical generation within the subregion is a mix of hydroelectric power imported from the BPA and Snake River systems, coal fired generation at the Jim Bridger Plant near Green River, Wyoming, and smaller local hydro and coal fired plants. Domestic and Canadian natural gas provides the most important energy source for direct heat applications throughout the subregion. Concern over natural gas availability has motivated widespread interest in alternative energy sources for direct heat applications.

Geothermal Potential

The Northern Rocky Mountain Subregion has a widespread potential for the discovery and development of moderate and low temperature geothermal resources. There is no geochemical or geologic evidence for the existence of very high temperatures suitable for electrical power generation. Widespread moderate and low temperature resources are localized by the presence of fractures and fault zones. Heat flow throughout at least the Idaho Batholith and Boulder Batholith is known to be high and virtually any recently active fault zone within this portion of the subregion has a potential for providing a moderate temperature resource through deep convective circulation. Hot springs and shallow, moderate temperature, hydrothermal resources are particularly common along structural zones within the Idaho Batholith.

Strategy

The geothermal program in this subregion will emphasize the stimulation of the development of moderate temperature resources for direct heat applications. Individual communities and industries within this natural gas-dependent area will provide the primary targets for development of the geothermal resources for direct heat applications. The program will be implemented largely through the TIPS project and a continuation and expansion of the PON program. The wide-spread occurence of hard rock mining throughout the subregion presents both an opportunity for expanded utilization of natural hot waters in mineral beneficiation and an institutional question concerning the status of geothermal rights versus mineral rights.

SUBREGION III - WASATCH FRONT

Subregional Setting

The Wasatch Fault Zone and its northern continuation as the Teton Fault Zone through southeastern Idaho to the southern border of Yellowstone volcanic field contains a disproportionate percentage of the subregion's population and land suitable for agricultural purposes. The Wasatch Fault Zone marks a sharp boundary between the Basin and Range Province to the west and the Wasatch Range to the east. The western margin of the Wasatch Front Subregion generally corresponds to a zone of seismic activity known as the Intermountain Seismic Belt, which continues on northward in the Northern Rocky Mountains past Yellowstone through western Montana to the Canadian border near Glacier National Park. The Unita uplift, the Wind River Range and the associated overthrust belts are also included within this subregion.

Economy of the Subregion

The Wasatch Front Subregion is generally lightly populated forest land with forest products and ranching dominating its rural economy. A narrow strip along the western margin of the subregion, the Wasatch and Teton Fault Zones, contains most of the area's major population and trade centers including Salt

Lake City, a major intermountain commerce and transportation center. This same area also includes much of the subregion's crop land. The subregion contains the watershed for several major drainages, but the populated portions of the area remain relatively water short. Water constitutes one of the major restrictions on the economic growth in the subregion.

Energy Production and Consumption

The subregion is a net importer of energy. Both coal and petroleum are imported from the Colorado Plateau immediately to the east of the subregion in the state of Utah. Utah Power and Light, the principal electrical utility for the southern half of the region currently purchases power from BPA and has tentatively contracted with Phillip for geothermal steam from the Roosevelt field, in south central Utah.

Geothermal Potential

The techonically active margins of the subregion, which border the Basin and Range Subregion, contain low and moderate temperature geothermal resources suitable for direct heat applications. The general absence of young volcanism within the subprovince is negative evidence concerning the potential for the discovery of the +200^{o}C fluid resources suitable for electrical generation in the near term. The fortunate coincidence of the area's population centers with widespread low to moderate temperature resources associated with the Wasatch Fault Zone provides a major opportunity for direct heat applications.

Oil and gas exploration of the overthrust belt of southeastern Idaho has provided direct confirmation of the presence of moderate temperature resources. Water near the boiling point has been produced from carbonate aquifers at depths of less than 2 km at several points in Teton Valley.

Strategy

The near term geothermal program for the Wasatch Front subregion will emphasize the acceleration of the development of low and moderate temperature resources for direct heat applications. This will be accomplished by means of the inventory of these resources in cooperation with state agencies in Utah and Idaho and the geothermal program of the U. S. Geological Survey. Following the initial inventory of the resources which will be completed in FY 78, the program will emphasize site specific studies and projects under the TIPS Project designed to bring the resource to the attention of the potential users. A significant impact on new energy requirements for low grade heat will be possible through an ambitious program which addresses markets in the private and public sector.

SUBREGION IV - COLORADO PLATEAU

Subregional Setting

The Colorado Plateau and the young volcanic ranges which occur around its margin with the Basin and Rande Province are here considered as a single subregion. The Colorado Plateau is roughly a circular area bounded on the west and south by the Basin and Range Province and on the north and east by the Wasatch Front and southern Rocky Mountains. Topographically, the plateau is divided into a number of individual uplifts and basins which range in elevation from 5,000 to 11,000 feet. The margins of the plateau contain a number of relatively young volcanic ranges with a significant geothermal potential. These include the Mineral Range in southwestern Utah, the San Francisco Peaks near Flagstaff, Arizona, the White Mountains in southcentral Arizona, the Zuni Uplift in northwestern New Mexico and the San Juan Range of Southwestern Colorado.

Economy of the Subregion

The subregion contains an abundance of mineral resources, including coal, oil and gas, uranium, and precious metals. The area is, in general, sparcely populated with scattered commerce centers, such as Flagstaff, serving large geographic areas. A number of cities have prospered in the Four Corners region as a result of oil and gas production and uranium exploration and development. Coal mining is a major activity within the Uinta region and the Kaiparowits Plateau field. Agriculture in the form of truck farming and orchards is an important local source of income in the valleys marginal to the plateau. Much of the land is semi-arid and supports sheep and cattle ranching.

Energy Production and Consumption

The subregion is a net exporter of energy as a result of the coal generation plant at Four Corners. The oil and gas fields of the subregion include the Uinta Basin, and the numerous fields in the Paradox Basin and the San Juan Basin. Energy consumption within the region is low, but the region presents major opportunities for the growth of energy-intensive industries co-located with the coal deposits and geothermal resources.

Geothermal Potential

The interior of the Colorado Plateau is generally thought to be a relatively low heat flow province. The margins of the plateau, however, contain major, confirmed, high-temperature geothermal systems associated with young silicic volcanic centers. The Roosevelt-Cove Fort-Sulfurdale-Thermo KGRAs in southcentral Utah constitute a major electrical generation resource which is being actively developed by industry with DOE support.

Young volcanic centers in central and eastern Arizona include the San Francisco Peaks and the White Mountains. These regions have not been explored by deep drilling but appear promising on the basis of their geologic setting, regional heat flow measurements and limited geochemical data. The Zuni Uplift in northeastern New Mexico is interesting but its potential is less substantiated.

Strategy

The young volcanic fields marginal to the Colorado Plateau constitute a high priority target within the total region. Acceleration of the rate of development of the known fields in southern Utah in order to meet the 1985 goal of 100 MWe and the year 2000 goal of 2600 MWe will be accomplished primarily by means of the industry-coupled drilling program which was initiated in FY 77. Further drilling in adjacent KGRAs will be encouraged during subsequent solicitation programs. The rate of development of these reservoirs will also be accelerated by means of case studies of the data set provided by the industrial participants in the program.

The utilization of by-product fluids produced by electrical generation at these fields will be encouraged through an expanded region-wide PON program for direct heat applications.

Oil and gas exploration within the Colorado Plateau will be carefully monitored for abnormal gradients encountered during the oil and gas exploration. The margins of the plateau will receive particular emphasis in state-USGS cooperative origrans with Arizona, Utah, New Mexico and Colorado. These programs are designed to target reservoirs suitable for direct heat applications, which the TIPS Project will help make available to potential users.

SUBREGION V - BASIN AND RANGE

Subregional Setting

The Basin and Range subregion is a major physiographic province which includes most of Nevada, southwestern Arizona, western Utah, southwestern New Mexico and a small portion of southern Idaho. The subregion includes block-faulted basins and ranges which are generally north-south trending. ~~throughout the region.~~ The subregion is arid to semiarid and characteristically is composed of desert lands and closed drainages. Although basaltic and rhyolicic lavas dated 6 to 20 million years before present are common throughout the province, there are very few young rhyolitic centers. The region has a higher than normal heat flow, and hot springs and wells, particularly in northcentral and eastern Nevada.

Economics of the Subregion

Mining and ranching provide the major regional source of income. Tourism, forestry and agriculture are locally important. The availability of water throughout the region is restricted and water requirements for any new industrial or population growth must be carefully considered.

Energy Production and Consumption

Due to the low population density, the Subregion is not a large consumer of energy. The minerals industry, however, does require large quantities of energy for mineral beneficiation at the numerous smelters dispersed throughout the province.
The region could be a major supplier of electrical power to California.

Geothermal Potential

The subregion has a widespread moderate temperature resource which is almost universally present along fracture zones within the region. Geochemically predicted base reservoir temperatures of 150° to 200°C are relatively common and temperatures as high as 240°C are predicted for some fields. Although the Basin and and Range Province is characterized by its high heat flow, the thermal gradients measured throughout the region are by no means uniform. High gradients are especially common in a region of northcentral and ~~northeastern~~ Nevada known as the Battle Mountain High. This area of unusually high heat flow does not appear to be associated with any known igneous heat source, but rather is an area of abnormally high gradient superimposed on the regional high.
The area of the Battle Mountain High continues to be the object of considerable industry interest in exploration for electrical

generating capacity. The westernmost portion and its boundary with the Sierran Front seems to possess the high temperature geothermal potential. Prospects with confirmed high temperatures include Steamboat Springs, Brady Hot Springs and Grey's Peak.

Strategy

In view of the interest displayed by industry in the electrical generating capacity of resources of the northern Basin and Range, this region has been targeted for the second initiative of the industry coupled program, beginning in 1978. Significant questions remain as to the nature of the heat source driving the numerous moderate and possibly high temperature systems. The industry coupled program will be designed to both stimulate the drilling and development of the numerous systems in the area and also to acquire detailed subsurface data which will be valuable in accelerating the industries rate of successful discoveries. A "modest" program of 10 W and moderate temperature reservoir identification is planned in cooperation with the USGS and State agencies. This program will seek as its main thrust to replace existing energy consumption in the region for mineral beneficiation at sites where mineral processing and the geothermal resources are co-located.

SUBREGION VI

RIO GRANDE RIFT - SOUTHERN ROCKY MOUNTAINS SUBREGION

The major feature of this subregion is the Rio Grande Rift, a structural depression located just west of the Sangre De Cristo Range of Northern New Mexico, which estends southward through Central New Mexico to the Texas border at El Paso. Also included in the Subregion are the Southern Rocky Mountains, which extend from the Laramie Range in Southern Wyoming to the Sangre De Cristo Range. The region is mountainous with elevations to 14,000 feet. ~~Intermountain basins called parks separate the individual ranges.~~ The Subregion is bounded on the east by the Great Plains and on the west by the Colorado Plateau and Wyoming Basin (Figure 7).

Economics of the Subregion

The economy of the subregion has a strong agriculture and forest products base. Tourism has become an increasingly important industry and environmental sensitivities are especially high. Ranching, forestry and only a limited additional agricultural activity is permitted by the topography and climate of the region.

Energy Production and Consumption

The area generally lacks energy intensive industries and is neither a major producer or consumer of electricity. Electrical power generation from high temperature geothermal resources could serve the needs of growing metropolitan areas such as Albuquerque, or could be exported to California. Many of the individual cities and town within the subregion are dependent on natural gas for direct heat applications and their service has been threatened during past winters by natural gas shortages. These urban areas constitute the major new term market for geothermal energy within the region.

Geothermal Potential

The subregion has a demonstrated high temperature reservoir which is being developed by Union Oil Company at the Valles Caldera near Los Alamos in northern New Mexico. The Caldera lies along the Rio Grande Rift on the margin of the Colorado Plateau. The presence of high temperature geothermal reservoirs at other sites along the Rift have been postulated but not confirmed. The Rift does constitute a favorable region for high temperature geothermal system discoveries. The remainder of the subprovince, particularly the more northern ranges, do not appear to have a high temperature potential. Known geothermal occurences through the San Luis Valley in southern Colorado and near Alenwood Springs in northern Colorado confirm that at least a moderate temperature resource is present throughout this region.

Strategy

A pre-commercial study of the high temperature potential of the Rio Grande Rift will be conducted during 1979 and 1980. Based on the success of this survey, an industry-coupled program will be initiated in 1981 which will be designed to stimulate industry exploration for high temperature systems within the Rio Grande Rift. The State Coop program and the PON program for direct heat applications will be employed in order to stimulate the development of low and moderate temperature geothermal resources in the major population centers of the region.

SUBREGION VII

GREAT PLAINS

Subregional Setting

The Great Plains subregion is a major physiographic province lying east of the Rocky Mountains. For the purposes of this program the Wyoming Basin is included within the Great Plains subregion. The Great Plains are underlain by eastward dipping sedimentary rocks of tertiary age. A number of individual mountain ranges, including the Black Hills of South Dakota, are present within the subregion. The Williston Basin is a large sedimentary basin centered to the northeast of the Black Hills in Montana, North Dakota and Northwestern South Dakota.

The principal deep fresh water aquifer throughout much of the subregion is the Madison Limestone.

Economy of the Subregion

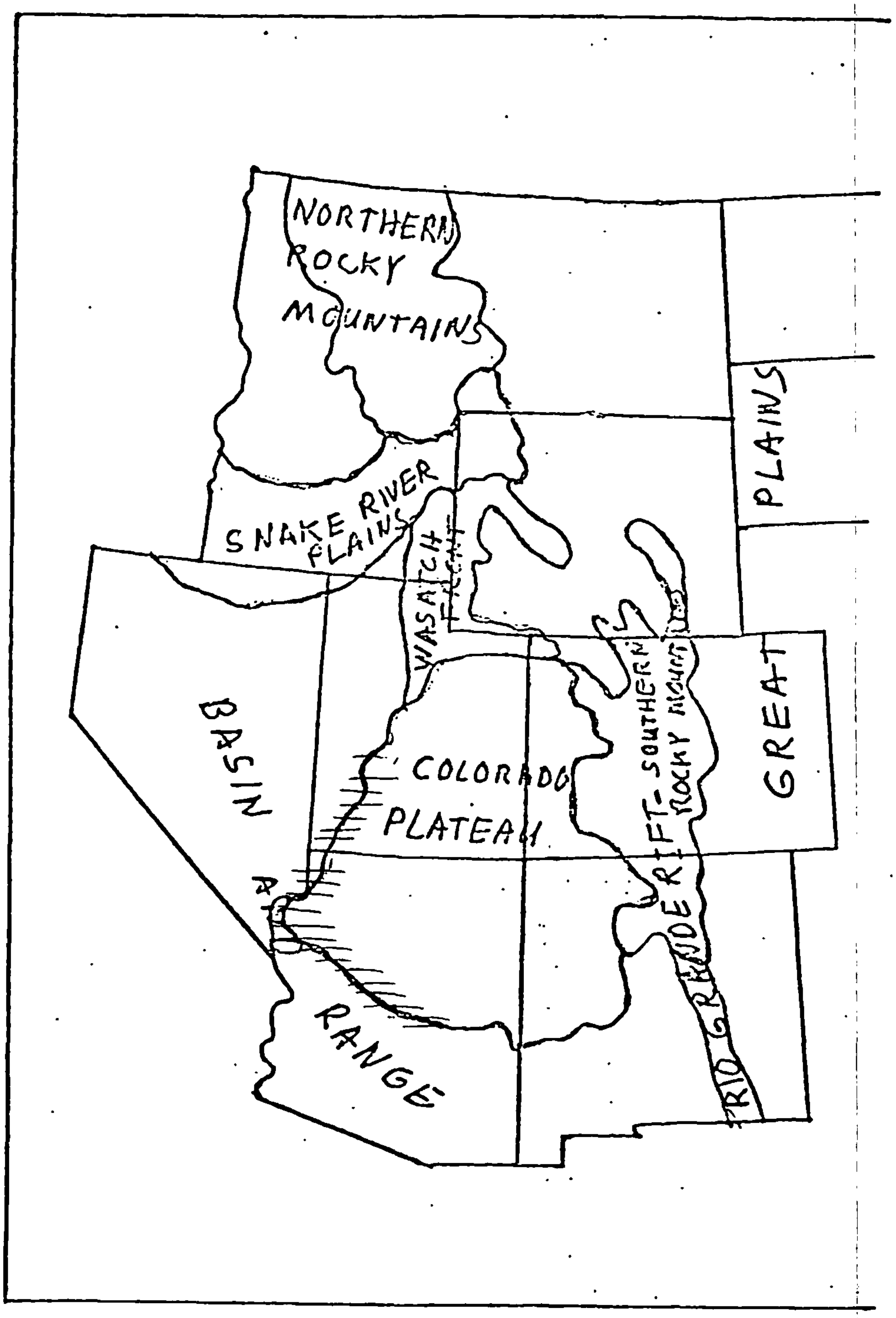
The economy of the area is dominantly agricultural, with most of the subregion being utilized for grain production and ranching. Oil production from the Williston Basin in North Dakota, the Powder River, Big Horn and Wind River Basins in Wyoming, and gas and oil production from several fields in Montana have constituted major non-agriculture economic activity of the subregion. Montana and Wyoming contain significant bituminous to subbituminous coal fields which are undergoing accelerated development and will significantly impact the region's economy. Coal processing will compete with other demands for ground and surface water in the Subregion.

Energy Production and Consumption

The region is a net exporter of energy and fuel as a result of its low population density and abundant energy resources. In view of the region's abundant coal deposits, coal generation of electricity may be water-limited rather than resource-limited.

Geothermal Potential

The Great Plains subregion contains no identified igneous point sources and the geologic environment does not suggest the presence of high temperature geothermal systems. Heat flow throughout most of the region is normal or near-normal and, as a result, moderate temperature convective systems are not common. The subregion does contain widespread occurrences of hot water in the Madison Aquifer, which has been locally utilized for direct heat applications. A significant development for space heating is presently underway in South Dakota under the PON program. Water near the boiling point is produced from the Madison Formation near Casper and Sheridan, Wyoming, making these urban areas potential users of geothermal energy for direct heat applications.


NORTHERN
ROCKY
MOUNTAINS
SNAKE RIVER
PLAINS
WASATCH
FRONT
PLAINS
GREAT
SOUTHERN
ROCKY MOUNTAINS
COLORADO
PLATEAU
BASIN
A
RANGE
RIO GRANDE RIFT

CHARACTERISTICS OF THE JURASSIC TWIN CREEK LIMESTONE IN IDAHO, WYOMING, AND UTAH[1]

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INTRODUCTION

This paper includes a summary of the lithologic nd stratigraphic characteristics of the seven members f the Twin Creek limestone, the descriptions of some ypical sections in western Wyoming and southeastern daho, and three lines of columnar sections. The last wo items present much information not published pre-iously. The summary descriptions represent a con-ensed version of those published in the Wyoming Geo-ogical Association Guidebook for 1950 (Imlay, 1950a) ut include some additional information. Only brief ention is made of the correlation of the members of e Twin Creek limestone, as that subject has been dis-ussed fully in the Bulletin of the Geological Society of merica (Imlay, 1952a). Likewise, the origin of the arious kinds of sediments comprising the Twin Creek mestone has been discussed amply in a report pub-shed by the National Research Council (Imlay, 950b).

ISTRIBUTION AND GENERAL FEATURES

The Twin Creek limestone occurs in an area of ex-nsive thrust faulting along the Idaho-Wyoming bor-r and in north-central Utah, extending from the uthern end of the Teton Mountains west of Jackson, yo., southward to the south end of the central Wa-tch Range near Thistle, Utah. It also occurs east of e area of thrust faulting in the western part of the nta Mountains as far east as Lake Fork (Thomas and uger, 1946, p. 1275-1277). Within the area of rusting, it thickens westward from about 800 feet to 300 feet. The thickest measured section is at Thomas rk Canyon, about 22 miles north-northwest of Coke-le, Wyo., but the section on Stump Creek, Idaho, out 8 miles northwest of Auburn, Wyo., is nearly as ck. The Twin Creek consists mainly of medium- to ht-gray limestone, of which most is shaly and athers into long splinters. However, the formation o contains two persistent red members in its lower rd, one cliff-forming limestone member at the top its lower third, and one sandy member at its top.

DESCRIPTIONS OF THE MEMBERS

Member A at the base of the Twin Creek limestone kens westward in an irregular manner from an average of 75 feet in western Wyoming to about 400 feet in the Blackfoot Mountains in Idaho. Its thickness may vary markedly within distances of less than a mile. It is absent in the Uinta Mountains and locally absent in the Wasatch Range of Utah.

The member is characterized by soft, brownish-red siltstone that contains interbeds, or units, of brecciated or honeycombed limestone. In western Wyoming and locally in eastern Idaho the lower part of the member contains a conspicuous unit of brecciated gray to yellow limestone that ranges from 10 to 50 feet in thickness. This unit is a jumble of sharply angular blocks, generally includes a little red siltstone, and shows faint stratification. The position of this brecciated limestone unit is occupied by thick masses of gypsum in the southeast corner of the Jackson Quadrangle in the E½ sec. 36, T. 36 N., R. 115 W. Locally, gypsum has been found in the lower part of the member near the head of Crow Creek, Caribou County, Idaho, in sec. 10, T. 11 S., R. 45 E. (Mansfield, 1927, p. 96). In many sections the middle and upper parts of the member contain one or more beds or thin units of yellow honeycombed or brecciated limestone that are generally inconspicuous. In southeastern Idaho the middle part of the member contains a unit of dense limestone that is siliceous and bears nodules and lenses of brownish-gray chert. This unit is about 70 feet thick on Stump Creek in the S½ secs. 27 and 28, T. 6 S., R. 45 E., Caribou County, and at least 140 feet thick on Williams Creek in the SE¼ sec. 12, T. 2 S., R. 39 E., Bingham County. A similar chert-bearing limestone generally only 1 or 2 feet thick occurs near the middle of the member in several sections near the Idaho-Wyoming border. Most sections contain minor amounts of brownish-red fine-grained sandstone interbedded with the red siltstone. Yellowish-white sandstone occurs locally at or near the base of the member. The basal beds of the member may consist of red siltstone, of soft yellowish sandstone, or of brecciated limestone, and they invariably rest sharply on the hard quartzitic Nugget sandstone. The upper contact of the member is marked by an equally sharp change from soft red siltstone to sandy or massive oolitic limestone.

Member B thickens westward from 25 to nearly 300 feet. In western Wyoming this member consists mainly of medium- to thin-bedded, grayish-black to dark brownish-gray limestone. Its basal unit generally

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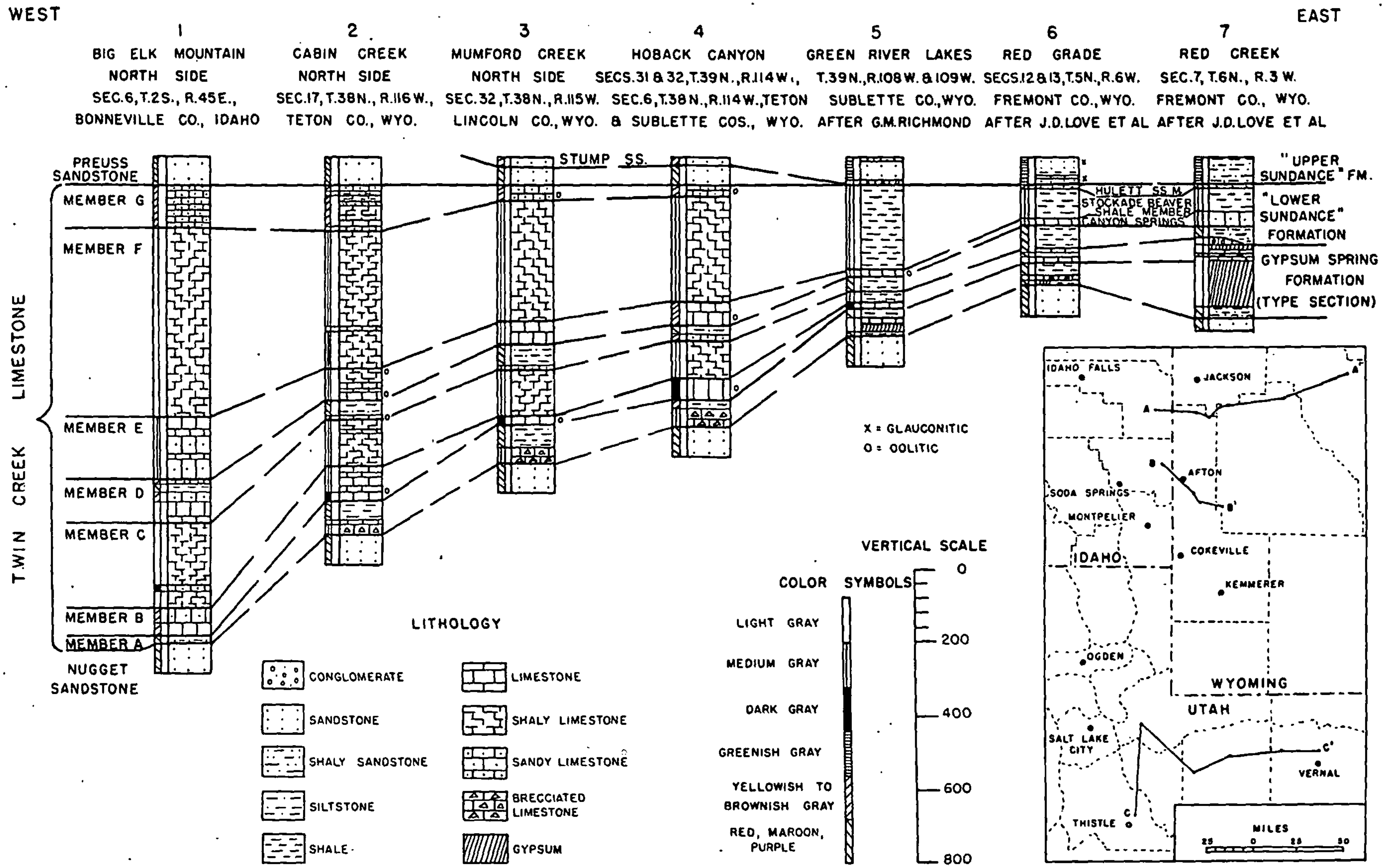
[1] ublication authorized by the Director, U. S. Geological Survey.

consists of 5 to 15 feet or more of dark oolite that contains a few sand grains and some pyrite. Thinner oolitic beds occur higher in the formation in some sections. In southeastern Idaho the basal 20 to 60 feet generally consists of brownish, sandy, crossbedded limestone that may contain tiny pebbles of red, green, and gray siliceous material. Similar sandy limestones also occur at higher levels. Some of the sandy beds are glauconitic. Oolitic beds are generally present above the basal unit of sandy limestone. A light-green to white volcanic tuff (Mansfield, 1927, p. 97) from 5 to 10 feet thick occurs within the member in the general area between Cokeville and Afton, Wyo. The member has furnished a large fauna of mollusks. Gryphaea planoconvexa Whitfield is one of its most common and most characteristic fossils. The ammonites Stemmatoceras and Chondroceras (Defonticeras) have been found in most sections in the upper part of the member and prove its middle Bajocian (earlier Middle Jurassic) age. The member persists southward into north-central Utah as least as far as Thistle. In the Uinta Mountains it is recognizable as far east as Lake Fork but is absent on the Whiterocks River. In the Jackson Hole area it thins eastward and becomes shaly, but the basal oolitic unit persists.

Member C thickens westward from 50 feet in western Wyoming to 350 feet in Idaho and consists mainly of medium-gray shaly limestone that is very soft basally but becomes harder upward, contains some thin beds near its top, and grades into the overlying silty beds of member D. A few thin beds near the top are generally composed mainly of crinoidal fragments. It weathers characteristically into light-gray splintery fragments. Its basal contact is transitional within a few inches in most sections. Its lower two-thirds has furnished Gryphaea planoconvexa Whitfield. The ammonites Stemmatoceras and Chondroceras were found about 10 feet below the top of the member on the North Fork of Stump Creek in the Freedom Quadrangle, Idaho. These ammonites show that the member is of early Middle Jurassic age. Member C is recognizable lithologically in northern Utah as far south as Thistle and as far east in the Uinta Mountains as the Whiterocks River. It thins eastward rapidly in the Jackson Hole area and is only about 50 feet thick at Lower Slide Lake on the Gros Ventre River.

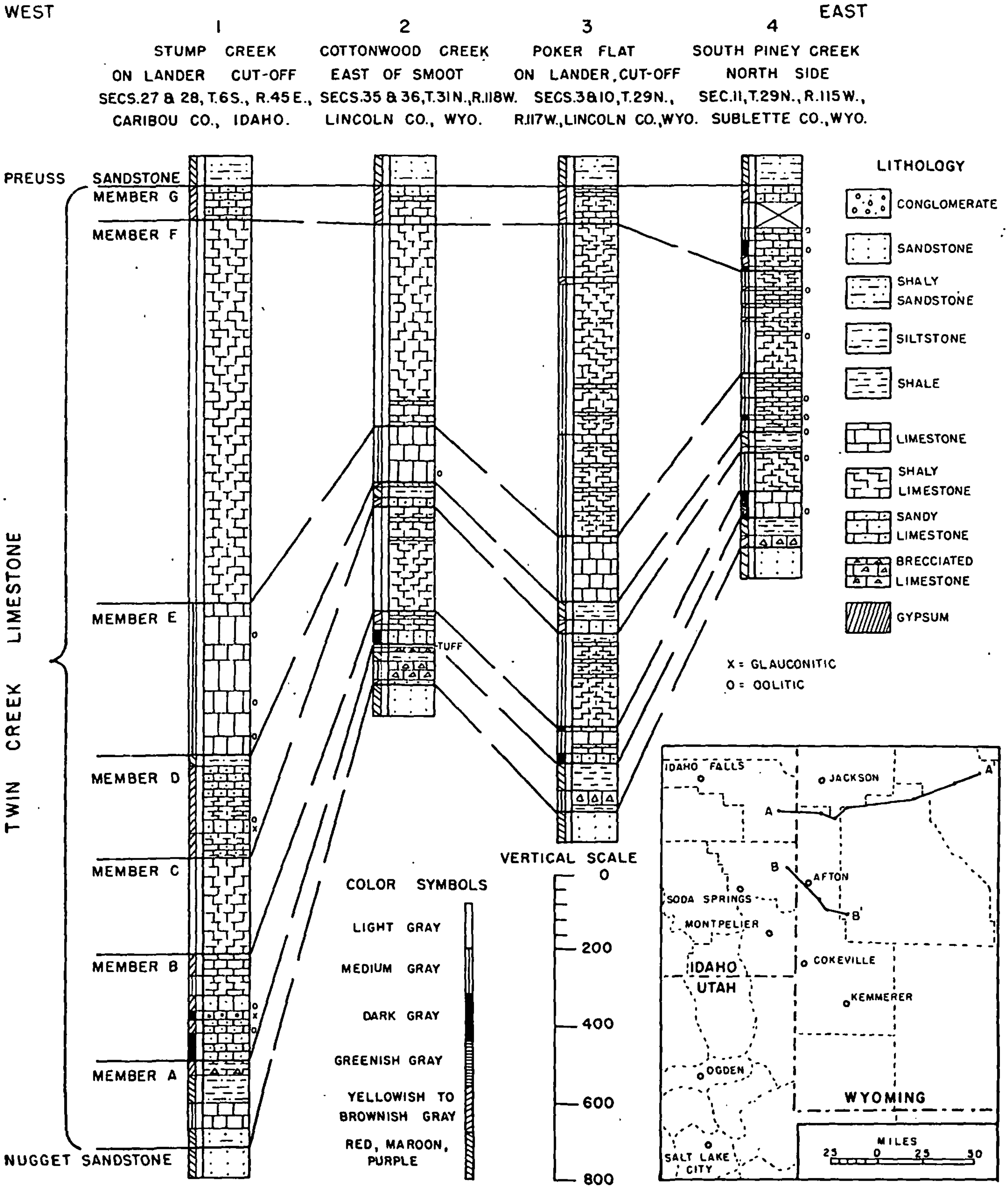
Member D thickens generally westward from 35 feet in western Wyoming to 270 feet in Idaho but varies considerably in thickness within short distances. It consists of interbedded soft red, green, or yellow siltstone, silty to finely sandy yellowish limestone, and greenish-gray silty shale. The limestones vary from shaly to thick-bedded, frequently show crossbedding, and contain marine fossils. Red siltstone dominates over limestone in the easternmost sections in Wyoming, but westward the member becomes more calcareous, sandier, and loses its red units. In Idaho is consists mostly of yellowish limestone whose sandy members are cliff-forming. In northern Utah, member D at most places consists of a unit of yellowish sandy limestone overlain by a unit of soft red siltstone. The base of member D is generally marked by a unit of silty to sandy limestone that is transitional into the underlying member. The top of member D makes a sharp contact with the overlying cliff-forming limestone at the base of member E.

Member E thickens westward from about 60 feet in western Wyoming to 400 feet in Idaho. It consists mostly of medium-gray to brownish-gray, medium-bedded, cliff-forming limestone but includes many thin beds in its middle and upper parts. Most of the beds are dense, but oolitic beds occur throughout. Generally the basal bed is massive and oolitic. In Idaho, along Preuss Creek and Stump Creek, some of the limestones are slightly sandy. Member E is the main ridge-former in the Twin Creek limestone and could be mapped easily if detailed mapping of the Twin Creek is ever found desirable. Its basal contact is sharp. It grades into the overlying member through a unit of thin-bedded to shaly limestone, and the boundary must generally be chosen arbitrarily within an interval of 30 to 50 feet. It has furnished very few fossils. Some of the beds contain crinoid parts and Camptonectes. Gryphaea nebrascensis Meek and Hayden was found near the top of the member on Sliderock Creek and on Cottonwood Creek east of Smoot, Wyo. Member E is recognizable lithologically in northern Utah as far south as Thistle and at least as far east as the Whiterocks River in the Uinta Mountains. The lowest few feet of limestone in the Carmel formation north of Vernal in sec. 26, T. 3 S., R. 21 E., is probably the easternmost limit of the member. Between the Whiterocks and Duchesne Rivers the upper part of the member contains a thin but conspicuous unit of grayish-white, thin-bedded, nearly lithographic limestone. This limestone is overlain at Lake Fork by a few feet of sandy limestone. Near Manila on the north side of the Uinta Mountains, member E consists entirely of slightly sandy oolitic limestone. Equivalent beds at Lower Slide Lake on the Gros Ventre River are about 57 feet thick and include, from base to top, 20 feet of medium-bedded oolitic limestone, 30 feet of shale with thin interbeds of limestone, and 7 feet of oolitic limestone. About 8 feet above the base of the shale were obtained the ammonites Arcticoceras and Cadoceras. These genera are common in the basal part of the Rierdon formation in Montana and in equivalent beds in north-central Wyoming.



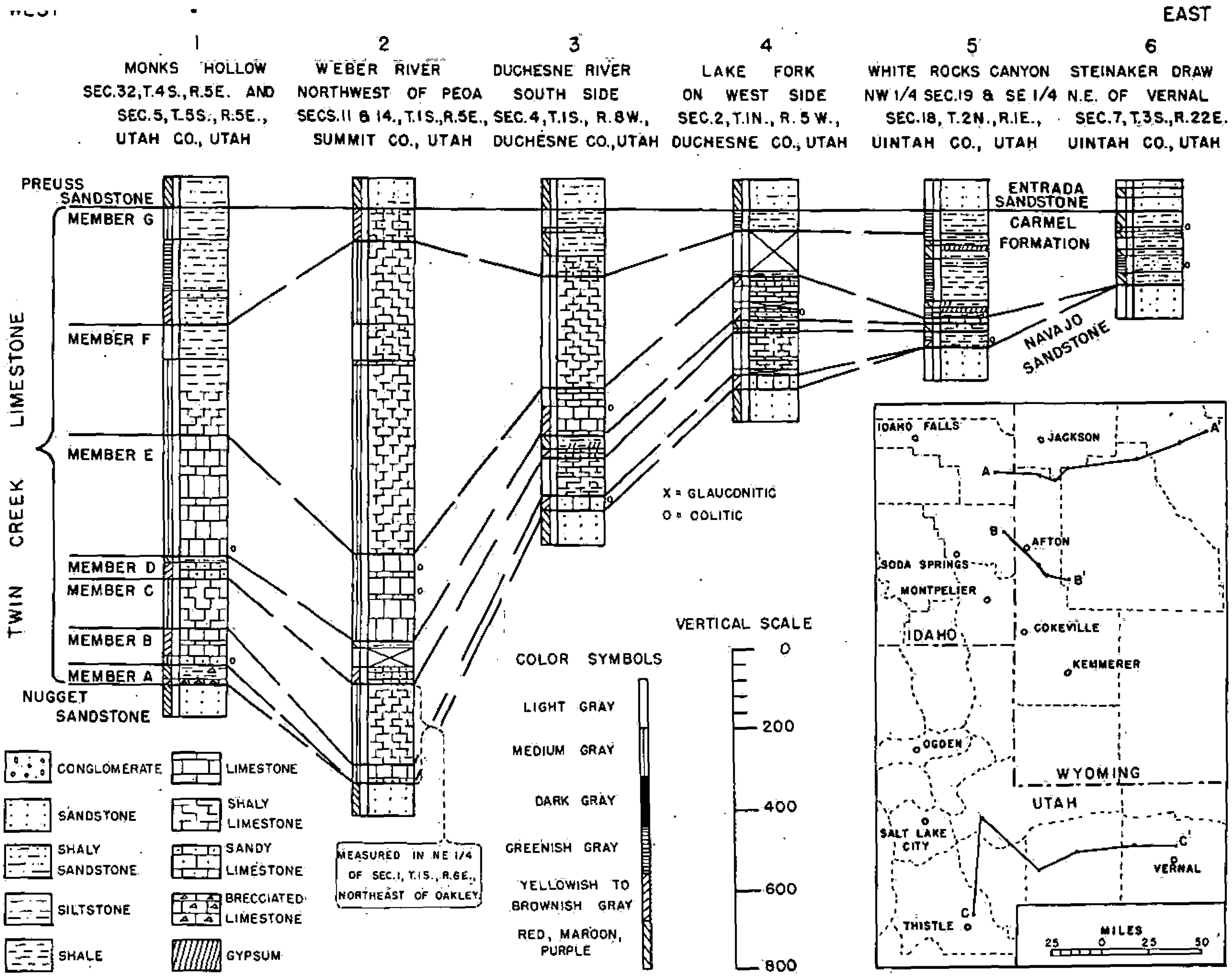
COLUMNAR SECTIONS ALONG LINE A–A'

FIGURE 1.



COLUMNAR SECTIONS ALONG LINE B-B'

FIGURE 2.



COLUMNAR SECTIONS ALONG LINE C–C'

FIGURE 3.

Member F thickens westward from about 250 feet in western Wyoming to 1,600 feet or more in Idaho. It is by far the thickest and most conspicuous part of the Twin Creek limestone, forming extensive bare slopes of light-gray color that are visible for great distances. It consists mainly of soft, dense, light-gray shaly limestone that weathers generally into lighter-colored splintery fragments. At wide intervals the member contains hard, thin beds that bear many fragments of crinoids and echinoids, fairly well preserved **Camptonectes**, a few oysters and belemnites, and rarely such pelecypods as **Pinna**, **Astarte**, **Isocyprina**, and **Trigonia**. In the middle and upper parts of the member, some of the thin limestone beds are oolitic, and others are silty to sandy and ripple-marked. Eastward the member becomes less calcareous, and a few of the units weather into chunky rather than splintery fragments. Associated with these chunky beds are some thin nodular limestones that may contain an abundance of **Gryphaea nebrascensis** Meek and Hayden. Such fossiliferous units are common in the section on Greys River in the Afton Quadrangle and on Cabin Creek and Fall Creek in the Jackson Quadrangle. The member is overlain transitionally by the silty to sandy beds of member G, and the boundary must be selected arbitrarily in most sections. Member F is recognizable lithologically in northern Utah as far south as Thistle but becomes much shalier southward, and some units are calcareous shales rather than limestones (Baker, et al., 1947). In the Uinta Mountains, member F is typically developed as far east as Lake Fork. At the Whiterocks River, and eastward, the beds occupying the stratigraphic position of member F consist mostly of redbeds and gypsum that are customarily included in the Carmel formation. Eastward, in the Jackson Hole area beyond the Darby-Absaroka line of overthrusting, the beds equivalent to member F consist of thinner, medium-gray, calcareous shales that near their base in some sections contain a few thin beds of nodular limestone. The nodular limestones contain a great variety of mollusks, including the ammonites **Cadoceras** and **Xenocephalites**. The shales are especially characterized by an abundance of **Gryphaea nebrascensis** Meek and Hayden, which contrasts with the rarity of the species in the underlying oolitic limestones that contain **Arcticoceras**. These shales are lithologically and stratigraphically identical with the Rierdon formation of Montana. Eastward, in the Wind River Basin, they pass into the Stockade Beaver shale member of the Sundance formation.

Member G ranges in thickness from about 25 feet to at least 288 feet, is highly variable in thickness, and within the area of thrust faulting does not thicken appreciably in any direction. It consists mostly of yellowish to greenish, lavender, or pinkish, silty to finely sandy, ripple-marked, thin-bedded limestone, and some shaly limestone. Some units consist of medium-bedded limestone that is generally oolitic or sandy. Some of the sandy units are crossbedded. Many beds are a coquina of crinoid and echinoid fragments, and their upper surfaces are commonly matted with shells of **Camptonectes**. The upper part of the member is generally harder and thicker-bedded than the lower part and in places forms low cliffs. Westward, in Idaho, the member becomes sandier and consists of interbedded units of ripple-marked sandy limestone and glauconitic, thin- to thick-bedded sandstone. At Thomas Fork Canyon and Wolverine Canyon the member is more than half sandstone and at Preuss Creek is mostly sandstone (Imlay, 1952b, p. 1740). The sandy units are lithologically similar to the Stump sandstone. Member G is overlain transitionally by red siltstone or sandstone at the base of the Preuss sandstone. At the top, in most sections, is a transitional zone that is generally less than 10 feet thick. In some sections, as on South Piney Creek and on Cabin Creek in Wyoming, the transitional zone is much thicker. In such sections the highest limestone is arbitrarily placed in the Twin Creek, because a marine limestone is apt to be more persistent than a red unit. Member G is recognizable in Utah as far south as Thistle and as far east as Lake Fork. East of the Duchesne River the member consists mostly of greenish-gray siltstone and sandstone. In the Jackson Hole area it disappears eastward and is absent at Lower Slide Lake on the Gros Ventre River. Member G is similar lithologically and stratigraphically to the Hulett sandstone member of the Sundance formation in central and eastern Wyoming, western South Dakota, western North Dakota, and southeastern Montana.

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LOCAL SECTIONS

Lower part of Twin Creek limestone about 1½ miles east of Bear Lake on road to Pegram in NW¼ sec. 29 and NE¼ sec. 30, T. 15 S., R. 45 E., Bear Lake County, Idaho:

TWIN CREEK LIMESTONE

	Feet
Member C:	
24. Limestone, shaly, soft, medium-gray, weathers light-gray. Not measured; at least several hundred feet exposed.	?
Member B:	
23. Limestone, medium-bedded, slightly sandy, crossbedded, medium yellowish-gray	30
22. Limestone, greenish-gray	1
21. Limestone, thin-bedded, yellowish- to pinkish-gray	20
20. Limestone, medium- to thick-bedded, finely sandy, crossbedded, medium-gray, weathers light brownish-gray, traces of oysters	5
19. Covered	11
18. Limestone in beds 8 to 12 inches thick, very sandy, a few small pebbles of red and yellow chert, crossbedded, dark brownish-gray, weathers light-brown, many oyster fragments in top bed	6
17. Sandstone, thin-bedded, crossbedded, fine-grained, light-gray.	5
16. Limestone, full of grit and small pebbles consisting of red, yellow, gray and black chert and white quartz, many shell fragments.	1
15. Covered.	44
14. Limestone, thin-bedded, sandy, partly crossbedded, a 6-inch coquina bed about 5 feet below top, medium yellowish-gray, weathers grayish-yellow	37
13. Limestone, thin-bedded, sandy, medium-gray, weathers same.	32
12. Limestone, thin- to medium-bedded, medium yellowish-gray, weathers brownish-gray	14
11. Limestone, sandy, hard, crossbedded, contains oysters, bryozoans, and crinoid fragments	1½
10. Limestone, finely sandy, partly crossbedded, medium-gray, weathers light yellowish-gray	22
Member A:	
9. Mostly covered. Some red siltstone occurs within 32 feet of top	130
8. Siltstone, light-red, soft	20
7. Siltstone, olive-green, soft	7
6. Siltstone, light-red, soft	6
5. Limestone, light yellowish-gray, nodular	1
4. Siltstone, light-red to light-green	30
3. Limestone, medium- to thin-bedded, nodular and porous but not brecciated, light yellowish-gray to olive-gray, weathers yellowish-gray	18
2. Limestone, thin-bedded, laminated, medium dark-gray, weathers light-gray	2½
1. Siltstone, reddish-brown, soft	32

UGGET SANDSTONE.

Twin Creek limestone on north side of Preuss Creek E½ sec. 15, T. 11 S., R. 45 E., Bear Lake County, ho:

EUSS SANDSTONE.

IN CREEK LIMESTONE:

	Feet
ember G:	
25. Sandstone, thin-bedded; some sandy limestone, pinkish, weathers dull pinkish-gray, contact with Preuss sandstone transitional within 10 feet	71
ember F:	
24. Limestone, shaly, soft, some thin beds, medium-gray, weathers light-gray. Cannot be measured because of strong folding but at least	1,500+
ember E:	
23. Limestone, medium- to thin-bedded, dense, medium-gray.	95
22. Limestone, massive, slightly oolitic, some shell fragments, forms top of cliff	10
21. Limestone, thin- to medium-bedded, slightly sandy, medium yellowish-gray	48
20. Covered.	26
19. Limestone, thick-bedded, cliff-forming, medium-gray	61
18. Limestone, massive, oolitic, medium-gray	6
Member D:	
17. Siltstone, brownish-red, soft	6
16. Limestone, thin-bedded to shaly, silty to sandy, light yellowish-gray.	33
Member C:	
15. Limestone, shaly, soft, light-gray, a few thin beds	271
Member B:	
14. Limestone, medium- to thin-bedded, medium-gray, forms low cliff	20
13. Limestone, thin-bedded, light-gray	21
12. Limestone, thin- to medium-bedded, sandy, some grains of grit size, brownish-gray	30
11. Tuff, dense, light-green to white	5
10. Covered.	22
9. Limestone, thin- to medium-bedded, sandy, brownish-gray, becomes less sandy toward base, some beds coquinoid.	111
8. Limestone, medium-bedded, gray	20
Member A:	
7. Covered.	15
6. Siltstone, brownish-red, soft	37
5. Limestone, brecciated, gray	4
4. Sandstone, thin-bedded, brownish-red, very fine-grained.	33
3. Siltstone, brownish-red, soft	30
2. Limestone, brecciated, gray	7
1. Siltstone, brownish-red, soft	3
Approximate thickness	2,485+

NUGGET SANDSTONE (not measured).

Twin Creek limestone along old Lander Trail south of Stump Creek in S½ secs. 27 and 28, T. 6 S., R. 45 E., Caribou County, Idaho (thicknesses approximate):

PREUSS SANDSTONE.

TWIN CREEK LIMESTONE:

	Feet
Member G:	
22. Limestone, thin-bedded, slightly sandy, yellowish- to pinkish-gray, overlain by soft red siltstone at base of Preuss sandstone	90
Member F:	
21. Limestone, shaly, soft, light-gray	1,000±
Member E:	
20. Limestone, medium- to thick-bedded, slightly sandy, some beds oolitic, contains some comminuted shells, medium-gray to light yellowish-brown	400
Member D:	
19. Siltstone, brownish-red, soft	30
18. Limestone, thin-bedded, sandy, yellowish-gray	60
17. Limestone, cliff-forming, finely sandy, light yellowish-gray, contains many small Gryphaea	20
16. Limestone, thin-bedded to shaly, silty, yellowish-gray	60
15. Limestone, cliff-forming, sandy, glauconitic, partly oolitic, greenish- to pinkish-yellow	40
14. Limestone, thin-bedded to shaly, interbedded with calcareous siltstone, some beds sandy, yellowish-gray	60
Member C:	
13. Limestone, shaly, soft, light-gray, becoming harder upwards, contains abundant Gryphaea planoconvexa Whitfield in its lower two-thirds	250
Member B:	
12. Limestone, thin-bedded, slightly sandy, yellowish-gray, weathers light yellowish-gray, upper 10 feet contains many Gryphaea planoconvexa Whitfield	60

TABLE 1.

THICKNESS IN FEET OF THE MEMBERS OF THE TWIN CREEK LIMESTONE AND SOME EQUIVALENT FORMATIONS IN WYOMING, IDAHO, AND UTAH

	A	B	C	D	E	F	G	Total
Mosquito Pass, Wyo.: N½ sec. 34, T. 41 N., R. 118 W.	80	75	95	45	65	395	25	780
Lower Slide Lake, Wyo.: Sec. 4, T. 42 N., R. 114 W.	46	56	50	38	57	163	0	410
Wolverine Canyon, Idaho: E½ sec. 28 & W½ sec 27, T. 1 S., R. 39 E.	NOT EXPOSED					1500+	172	
Fall Creek, Idaho: Sec. 18, T. 1 N., R. 43 E.	96	200	338	77	160	628	131	1630
Big Elk Mountain, Idaho: SW¼ sec. 6, T. 2 S., R. 45 E.	20	74	228	123	172	520	120	1257
Fall Creek, Wyo.: NE¼ sec. 20, T. 39 N., R. 116 W.	63	55	151	40	69	477	48	903
Cabin Creek, Wyo.: S½ sec. 17, T. 38 N., R. 116 W.	97	97	140	40	89	370	127	960
Mumford Creek, Wyo.: SE¼ sec. 32, T. 38 N., R. 115 W.	107	24	125	71	65	330	41	763
Hoback Canyon, Wyo.: Secs. 31 & 32, T. 39 N., R. 114 W.; sec. 6, T. 38 N., R. 114 W.	76	60	104	43	65	290	28	666
Stump Creek, Idaho: S½ secs. 27 & 28, T. 6 S., R. 45 E.	223	281	250	270	400	1000	90	2514
Greys River, Wyo.: Sec. 4, T. 33 N., R. 116 W.	110	45	240	64	155	475	89	1178
Cottonwood Creek, Wyo.: W½ sec. 36 and E½ sec. 35, T. 31 N., R. 118 W.	102	87	275	66	146	530	100	1306
Poker Flat, Wyo.: Secs. 3 & 10, T. 29 N., R. 117 W.	125	88	247	83	175	813	102	1633
South Piney Creek, Wyo.: Sec. 11, T. 29 N., R. 115 W.	82	70	103	49	157	262	232	955
Preuss Creek, Idaho: E½ sec. 15, T. 11 S., R. 45 E.	129	229	271	39	246	1500+	71	2485
Thomas Fork Canyon, Wyo.: Secs. 19 & 20, T. 28 N., R. 119 W., sec. 24, T. 28 N., R. 120 W.	40+	188	315	168	305	1625	111	2752+
Ferney Gulch, Wyo.: Secs. 1 & 2, T. 27 N., R. 117½ W., sec. 1, T. 27 N., R. 118 W.	140	91	252	115	400	575	86	1659
Devils Hole, North Fork, Wyo.: Sec. 15, T. 27 N., R. 117 W.	75	79	245	69	218	735	102	1523
LaBarge Creek, Wyo.: NW¼ sec. 16 & NE¼ sec. 17, T. 27 N., R. 115 W.	53	75	208	59	339	249	128	1111
Sliderock Creek, Wyo.: Sec. 10, T. 25 N., R. 118 W.	150	85	275	75	154	1089	186	2014
Fontenelle Creek, South Fork, Wyo.: NW¼ sec. 33, T. 26 N., R. 116 W.	77	68	184	35	212	487	177	1240
Leed Canyon, Wyo.: Secs. 1 & 2, T. 22 N., R. 119 W.	76	95	260	108	182	1118	102	1941
Manila, Wyo. (4 miles south of): SW¼ sec. 6, T. 2 N., R. 20 E.	0	0	24	7	23	227	50	331
Weber River near Pesa, Utah: SW¼ sec. 11 & NW¼ sec. 14, T. 1 S., R. 5 E.	0	47+	125	107	220	776	82	1357+
Duchesne River, Utah: SW¼ sec. 4, T. 1 S., R. 8 W.	0	42	91	68	104	280	165	750
Lake Fork, Utah: Sec. 2, T. 1 N., R. 5 W.	0	32	109	30	109	114	49	443
Whiterocks River, Utah: NW¼ sec. 19 & SE¼ sec. 18, T. 2 N., R. 1 E.	0	0	40	21	17	182	85	345
Monks Hollow, Utah: Sec. 32, T. 4 S., R. 5 E., & sec. 5, T. 5 S., R. 5 E.	49	92	123	57	305	275	288	1189
Thistle, Utah: W½ sec 33, T. 8 S., R. 4 E.	9	71	183	41	345	?	?	

11. Limestone, shaly, medium-gray, weathers light yellowish-gray, contains Stemmatoceras	50
10. Limestone, massive, dense, medium-gray, weathers white	6
9. Limestone, thin- to medium-bedded, slightly sandy, oolitic, brownish-gray, rather soft	35
8. Limestone, sandy, glauconitic, crossbedded, contains small pebbles of gray and red chert, many oyster and crinoid fragments, dark-gray, forms low cliffs	25
7. Limestone, sandy, thin-bedded, dark yellowish-gray	35
6. Limestone, sandy, oolitic, medium- to thick-bedded, contains many oysters on bedding surfaces, dark-gray	30
5. Limestone, sandy, thin-bedded, contains many oyster and crinoid fragments, dark-gray	40
Member A:	
4. Siltstone, light-gray to pink, interbedded with yellowish thin-bedded limestone that is locally brecciated and honeycombed.	41
3. Siltstone, brownish-red, soft, poorly exposed; some beds of honeycombed limestone near base	65
2. Limestone, medium- to thin-bedded, medium-gray, weathers light-gray; contains considerable brownish- to reddish-gray chert as nodules, short, thin lenses, and as granules; crinoid fragments abundant; some beds sandy and crossbedded	70
1. Sandstone, fine-grained, and siltstone, brownish-red, poorly exposed.	47
Total thickness of Twin Creek	2,514±

Twin Creek limestone on north side of Big Elk Mountain between junction of Elk Creek and Bear Creek in the SW¼ sec. 6, T. 2 S., R. 45 E., Bonneville County, Idaho:

PREUSS SANDSTONE.
TWIN CREEK LIMESTONE:

	Feet
Member G:	
12. Limestone, thin-bedded, silty to finely sandy, yellowish-gray, ripple-marked, locally crossbedded, upper 16 feet contains interbeds of pink siltstone	120
Member F:	
11. Limestone, shaly, soft, breaks into splintery fragments, light-gray, has thin beds of nodular limestone every 10 to 15 feet. Gryphaea nebrascensis Meek and Hayden noted at 250 and 410 feet above base	520
Member E:	
10. Limestone, medium- to thin-bedded, cliff-forming, medium-gray, dense to granular, some beds slightly sandy and showing weak crossbedding. One 4-foot bed of oolitic limestone occurs about 72 feet above base.	172
Member D:	
9. Limestone, shaly to thin-bedded, sandy, yellowish	15
8. Siltstone, light brownish-red, soft	24
7. Limestone, thick-bedded, sandy, yellowish	25
Limestone, medium- to thin-bedded, becoming thicker-bedded upward, medium-gray to yellowish-gray.	59
Member C:	
5. Limestone, shaly, soft, light-gray, weathers into pencil-like fragments.	168
4. Limestone, medium-bedded, dark-gray, slightly sandy, contains many crinoid fragments	12
3. Limestone, shaly, light-gray, poorly exposed	48
Member B:	
2. Limestone, medium-bedded, yellowish-gray, becomes sandy in upper part	74
Member A:	
1. Sandstone, fine-grained, and siltstone, brownish-red to mottled gray and red; contains some beds of honeycombed limestone at top, poorly exposed	20
Total thickness of Twin Creek	1,257

Twin Creek limestone along Fall Creek in Irwin Quadrangle, measured from center to southwest corne of sec. 18, T. 1 N., R. 43 E., Bonneville County, Idaho:

PREUSS SANDSTONE.
TWIN CREEK LIMESTONE:

	Feet
Member G:	
23. Sandstone, thin-bedded (½ inch to 4 inches thick), light yellowish- to olive-gray, some glauconite	60
22. Limestone, medium- to thin-bedded (1 inch to 12 inches thick), medium yellowish-gray, mostly oolitic, some dense, silty to finely sandy, weathers yellowish-gray, some beds full of crinoid columnals and arm fragments, some glauconite	3
21. Limestone, shaly, light yellowish-gray	27
20. Limestone, same as unit 22	41
Member F:	
19. Covered.	80
18. Limestone, shaly, medium-gray, weathers light-gray, chunky to splintery	453
17. Limestone, thin-bedded to shaly, medium-gray	95
Member E:	
16. Limestone, medium- to thick-bedded, medium-gray, oolitic to dense, weathers medium-gray	160
Member D:	
15. Siltstone, red, soft	45
14. Limestone, medium-bedded, silty, oolitic in lower part, medium to yellowish-gray, weathers medium-gray, upper part dense, slightly sandy throughout but mostly sandy toward top	32
Member C:	
13. Limestone, thin-bedded to shaly, medium-gray, weathers light gray	117
12. Limestone, medium-bedded (6 to 8 inches), medium-gray, weathers light-gray	32
11. Limestone, shaly, medium-gray, weathers light-gray, chunky, becomes harder toward top	189
Member B:	
10. Limestone, thin- to medium-bedded, medium- to light-gray, weathers light-gray, contains Gryphaea planoconvexa Whitfield.	13
9. Limestone, very sandy, crossbedded, brownish-gray, weathers same, forms low cliff	11
8. Limestone, brownish-gray, slightly sandy, medium- to thick-bedded, weathers medium brownish-gray	32
7. Limestone, medium-bedded, medium-gray to yellowish-gray.	16
6. Limestone, silty, thick-bedded (6 to 24 inches thick), light brownish-gray, weathers medium brownish-gray, traces of crinoid columnals	21
5. Limestone, oolitic, medium-gray	10
4. Limestone, dense, thick-bedded, light-gray	10
3. Covered.	45
2. Limestone, medium gray to grayish-black, dense, medium- to thick-bedded, weathers dark-gray	42
Member A:	
1. Siltstone, mostly brownish-red, upper 20 feet purplish, soft; rests sharply on Nugget sandstone	96
Total thickness of Twin Creek	1,630

NUGGET SANDSTONE.

Twin Creek limestone and Preuss sandstone on Cabi Creek, Jackson Quadrangle, in S½ sec. 17, T. 38 N., F 116 W., Teton County, Wyo.:

STUMP SANDSTONE (not measured).
PREUSS SANDSTONE:

	Fee
29. Sandstone, dull-red to pink, thin-bedded to shaly, fine-grained, rather soft, contains a few hard, thin beds overlain sharply by glauconitic sandstone of Stump.	62
28. Sandstone, massive, fine-grained, hard, light pinkish-gray, weathers darker	6

27. Sandstone, dull-red, rather soft, a few hard layers	20
26. Sandstone, massive, very fine-grained, dull-pink, cliff-forming.	21
IN CREEK LIMESTONE:	
Member G:	
25. Limestone, sandy, crossbedded, beds 1 to 3 feet thick, light yellowish-gray, cliff-forming	17
24. Siltstone, shaly, mostly brownish-red, some yellowish-gray.	11
23. Siltstone, shaly, calcareous, ribboned yellow and gray.	16
22. Limestone, thick-bedded, finely sandy, shows some crossbedding, light yellowish-gray, cliff-forming	4
21. Limestone, shaly to thin-bedded, silty, light yellowish-gray.	48
20. Limestone, medium-bedded, consists mainly of crinoid and echinoid fragments, medium-gray	1 to 6
19. Limestone, shaly, medium-gray, weathers into light-gray splinters.	16
18. Limestone, medium-bedded, sandy, medium yellowish-gray, forms ledge	5
17. Limestone, thin-bedded to shaly, silty, yellowish-gray	
Member F:	
16. Limestone, shaly, medium- to light-gray, weathers into light-gray splinters	253
15. Limestone, silty, yellowish	10
14. Limestone, shaly, fissile to splintery, medium-gray	32
13. Limestone, shaly, chunky, medium-gray	75
Member E:	
12. Limestone, medium- to thin-bedded, partly oolitic, medium-gray.	11
11. Limestone, thin-bedded to shaly, poorly exposed	53
10. Limestone, medium- to thin-bedded, oolitic to dense, medium-gray.	25
Member D:	
9. Siltstone, red, soft, upper contact sharp	40
Member C:	
8. Limestone, thin- to medium-bedded, silty, some beds oolitic, medium yellowish-gray	10
7. Limestone, shaly, soft at base, forms low ledges at top, medium-gray.	130
Member B:	
6. Limestone, thin-bedded to shaly, medium-gray, Gryphaea planoconvexa Whitfield found at top	70
5. Limestone, oolitic, medium-bedded, slightly sandy, dark-gray.	11
4. Limestone, medium-bedded, dense, medium-gray	16
Member A:	
3. Siltstone, red, soft, poorly exposed	55
2. Limestone, medium-bedded, granular, light-gray, contains some chert	10
1. Limestone, brecciated, medium-gray, lower 2 feet yellow.	32
Total thickness of Twin Creek	960

GGET SANDSTONE.

Incomplete section of Twin Creek limestone on north nk of Williams Creek in SE¼ sec. 12, T. 2 S., R. 39 E., ngham County, Idaho:

	Feet
IN CREEK LIMESTONE:	
Member B (?):	
10. Limestone, oolitic, massive, sandy	15
9. Limestone, sandy, crossbedded, partly oolitic	60
Member A:	
8. Covered. Some float of soft red sandstone	100±
7. Limestone, medium- to thin-bedded, dense, medium- to dark-gray, siliceous, contains brownish chert nodules.	20
6. Limestone, yellowish-gray, brecciated or honey-combed.	20
5. Limestone, medium- to thin-bedded, dense, medium- to dark-gray, siliceous, contains some brownish chert nodules.	140
4. Limestone, brecciated, light yellowish-gray	8
3. Limestone, finely sandy, light yellowish-gray, interbedded with brownish-red siltstone	6
2. Siltstone, soft, dull-red, yellow, green, some interbedded honeycombed limestone	20
1. Siltstone, soft, pink to dull-red, some light-green or yellow, mostly non-calcareous, contains some beds of dull-red to yellow, fine-grained, non-calcareous sandstone; about 30 feet below top occurs 2 feet of dense, shaly yellow limestone	100

NUGGET SANDSTONE.

Twin Creek limestone equivalents north of Lower Slide Lake on Gros Ventre River in sec. 4, T. 42 N., R. 114 W., Teton County Wyo.:

	Feet
PREUSS SANDSTONE (?) (may be basal Stump):	
24. Siltstone, red.	3
23. Sandstone, light-gray.	5
TWIN CREEK LIMESTONE EQUIVALENTS:	
Member F:	
22. Shale, calcareous, medium-gray, weathers light-gray, one thin bed of nodular limestone in lower foot, several thin beds of fossiliferous limestone from 35 to 40 feet above base include Cadoceras and Xenocephalites. Gryphaea nebrascensis abundant throughout.	163
Member E:	
21. Limestone, oolitic, thick-bedded at top and bottom, thin-bedded in middle, medium yellowish-gray	7
20. Shale, calcareous, medium-gray, weathers light-gray. Gryphaea nebrascensis obtained 10 feet below top (lowest occurrence noted)	20
19. Shale, calcareous, medium-gray, and thin beds of soft, brownish-gray limestone, weathers light-gray. Eight feet above base occur Arcticoceras, Cadoceras, and many pelecypods	10
18. Limestone, oolitic, massive, medium-gray, weathers same.	3½
17. Limestone, medium- to thin-bedded, slightly oolitic, crumbly, medium yellowish-gray, weathers medium-gray, very fossiliferous	4½
16. Limestone, medium- to thick-bedded, beds 6 to 12 inches thick, oolitic, hard, medium yellowish-gray, weathers medium-gray, traces of fossils	12
Member D:	
15. Limestone, shaly, soft, yellowish-gray	3½
14. Limestone, shaly, soft, olive-green to yellowish-gray	1½
13. Siltstone, brownish-red, soft, makes sharp contact with underlying unit, thickens westward in ½ mile to 43 feet	33
Member C:	
12. Limestone, shaly, medium-gray, contains a few thin beds of coquinoid limestone and locally a hard bed of coquina at top	23
11. Limestone, shaly, soft, medium-gray, weathers same.	27
Member B:	
10. Limestone, mostly shaly, fairly soft, some beds from 4 to 10 inches thick at intervals of 4 to 8 feet, dark-gray to grayish-black, weathers dark-gray; 20 feet above base occurs Chondroceras; Gryphaea planoconvexa Whitfield occurs throughout	35
9. Limestone, medium- to thin-bedded, mostly dense, partly oolitic, upper 2 feet slightly sandy and pyritic, medium yellowish-gray, weathers medium-gray	16
8. Limestone, shaly, soft, medium-gray	4
7. Shale, soft, yellowish-gray	1
Member A:	
6. Siltstone, brownish-red, soft	15
5. Limestone, pinkish-yellow, weathers pinkish to yellow, forms top of cliff	8
4. Limestone, brecciated, gray to yellow, angular fragments as much as a foot in diameter but most fragments smaller, forms cliff	16
3. Limestone, brecciated, silty, purplish to yellow and gray.	2±
2. Limestone, silty, soft, yellow to pinkish	1 to 2
1. Siltstone, brownish-red, soft, rests sharply on Nugget sandstone.	3
Total thickness of Twin Creek	410

MISSISSIPPIAN STRATIGRAPHY IN THE UTAH-IDAHO-WYOMING AREA

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INTRODUCTION

Mississippian seas were widespread in the Rocky ountain region, and throughout most of Mississippian me broad seaways extended from Alaska to Mexico. ithin these seas thick sequences of limestone, sand- one, and shale were deposited, with carbonate rocks redominating in volume and extent. Eardley (1949, 665) has given the name "the Madison basin" to long narrow zone extending from western Montana, rough southeastern Idaho, and into southern Nevada, hich received over 4,000 feet of sediments in Lower ississippian time. A somewhat smaller area in west- n Utah and central Idaho which sank over 6,000 feet d received over 6,000 feet of sediments in the Mis- ssippian is called "the Brazer basin."

The area covered by the excursion (Ogden, Utah, Jackson, Wyoming) (Figure 1) lies roughly along hat Kay (1951, p. 10, 14) has called "the Wasatch ne." This hypothetical line roughly marks the posi- n of the monoclinal flexure from the craton on the st into the miogeosyncline on the west. The area of miogeosyncline west of the Wasatch line (defined a line of disappearance of the Lower Cambrian and 2,000 foot isopach of the entire Cambrian) has n termed the Millard Belt (Kay, 1947, 1951). Thus tectonic pattern in the Cordilleran region was set Cambrian time and the same pattern was followed oughout the Paleozoic. Kay (1951, p. 14) says, stems from the Ordovician through the Jurassic are erally more fully represented and thicker in the e areas in which lower Cambrian is present and whole Cambrian thicker."

The route of the trip throughout most of its extent far enough west so that practically the maximum ion of Mississippian is seen. Thus, thick fossiliferous issippian limestones in the Logan-Ogden area pass ward into drab sandstones, with the fossiliferous stones dropping out as the craton is approached. Brush Creek in the Uinta Mountains, Williams 3, p. 609) has stated that the Madison formation sists essentially of light-drab sandstones and silt- s, with tongues of red beds and a thick member of formational breccia." Also Wanless, et al. (1946) reported progressive westward thickening of the ssippian from 1,080 feet in the Gros Ventre e of western Wyoming, to 1,800 feet in the River Range of eastern Idaho; further westward in Idaho the Mississippian passes into a thick geosynclinal black shale sequence. Two formations make up the Mississippian column in most of the area: the Madison limestone (Kinderhookian) below, overlain unconformably by the Brazer limestone (Meramecian and Chesterian) above. A thin unit of shale, the Leatham formation (lower Kinderhookian), is known to conformably underlie the Madison in the Logan, Utah area; however the extent of this shale outside of the Logan area has not yet been determined by field studies.

The Mississippian generally rests unconformably upon the Devonian Jefferson limestone or an equivalent of the Upper Devonian Three Forks formation. In much of the area the Upper Mississippian Brazer limestone is unconformably overlain by the Pennsylvanian Wells formation, but from Teton Pass eastward the Darwin sandstone member of the Amsden formation (Lower Pennsylvanian age) overlies the Brazer.

The first report of Carboniferous rocks in Utah was by the Stansbury expedition in 1849 (Stansbury, 1852). Hayden, Peale, and others made observations in the Logan area. King (1876, p. 478-80) first named the strata, calling them the "Wahsatch" limestones of "Devonian" and "Carboniferous" age, however, Richardson (1913) revealed that King's "Wahsatch" included rocks of Ordovician to Mississippian age. The term Wasatch has not since been used to refer to Paleozoic rocks.

The type sections of two of the Mississippian formations lie near the route of the field trip. Richardson in 1913 named the Brazer limestone from exposures in Brazer Canyon in the Crawford Mountains, 6 miles northeast of Randolph, Utah; and Holland (1952, p. 1719) named the Leatham formation for the exposure on the north wall of Leatham Hollow about 8 air-line miles southeast of Logan, Utah. The Madison was named by Peale (1893) who failed to designate a type locality for the Madison limestone but did imply that the unit was named for the Madison River in the Three Forks, Montana area. Sloss and Hamblin (1942) and Holland (1952) have discussed the complex history of the name Madison and have described in detail the section directly north of Logan, Montana, designating this the type section of the Madison.

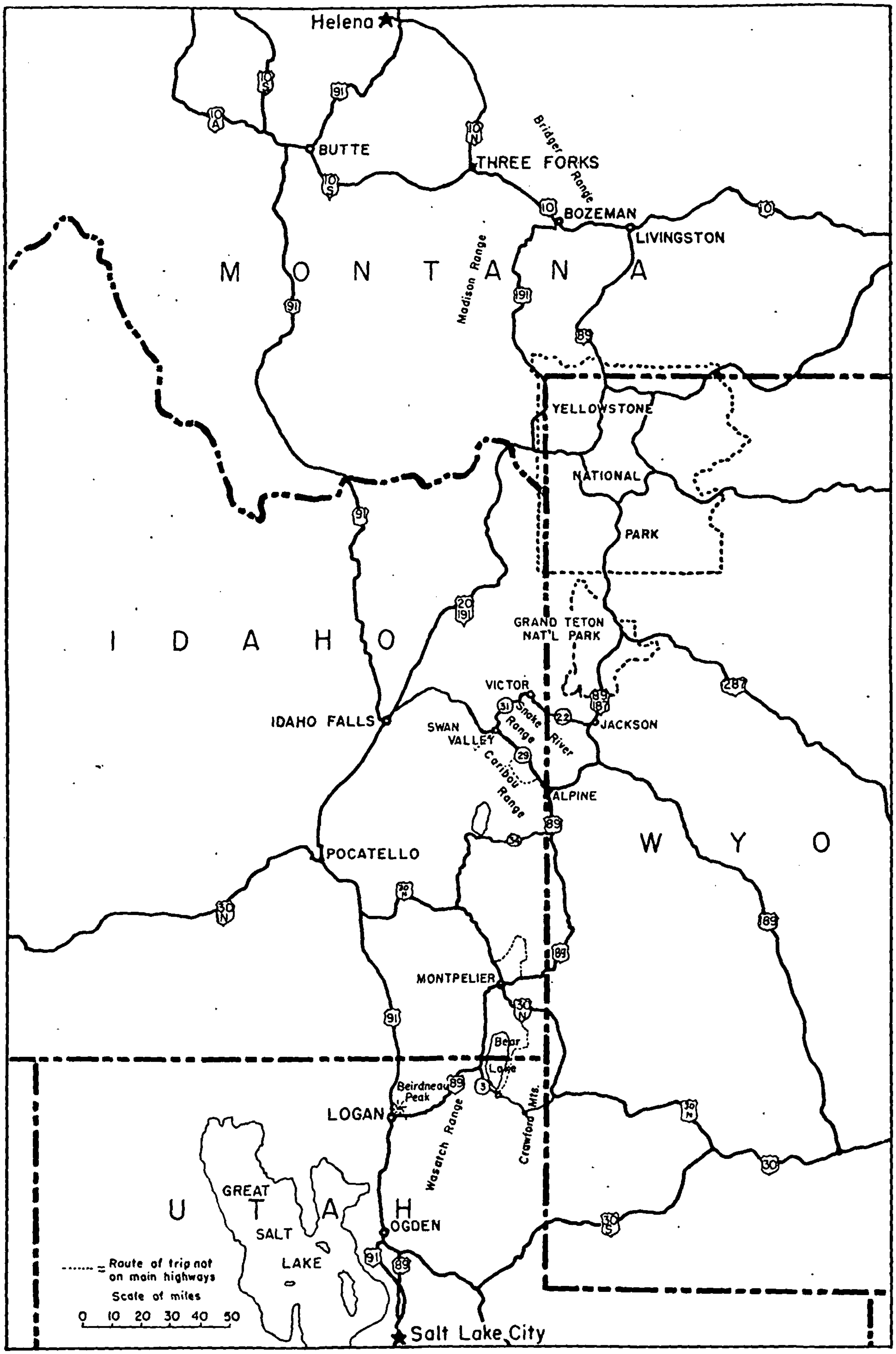

FIGURE 1.—GENERAL LOCATION MAP

Mansfield (1927) prepared a comprehensive report on the geology of southeastern Idaho and included in this report a description of the Carboniferous and Triassic fossils by Girty (1927, p. 411-46). Richardson (1941) published a report on the geology of the Randolph Quadrangle (next quadrangle east of the Logan Quadrangle) and included a geologic map. Williams (1943) has described numerous sections of Carboniferous formations in the Uinta and Wasatch Mountains. Eardley (1944) studied the geology of the north-central Wasatch Mountains, and Williams and Yolton (1945) described in detail Brazer and Wells sections near Dry Lake, southwest of Logan, Utah. Parks (1949 and 1951) zoned the Brazer on the basis of its coral fauna and Williams (1948) summarized much work in the Logan Quadrangle and presented an excellent report on the stratigraphy, structure, and historical geology, with a geologic map and detailed cross-sections.

Kirkham (1924) discussed the geology and mapped a large portion of the Caribou Range southwest of Swan Valley, Idaho.

In the mountain ranges about the Jackson Hole region the writer has relied principally upon reports by Horberg (1938), Horberg, Nelson, and Church (1949), Thomas (1948), Wanless and others (1945, 1946).

Ogden to Montpelier — East of Ogden the Mississippian crops out in Ogden Canyon. There the dark-gray, thin-bedded typical Madison is about 600 feet thick, and the Brazer is only about half as thick (1,100 feet) as in the Logan area.

Mississippian strata are not again encountered until Wellsville Mountain and the Pisgah Hills southwest of Logan. Approaching Dry Lake from the south the Leatham and Madison are not exposed along U. S. Highway 91. A very thick section of Brazer is, however, exposed along a road cut of an old portion of U. S. Highway 91 where it turns eastward across the Pisgah Hills toward Sardine Canyon. Williams and Yolton (1945) have reported 3,700 feet exposed in the Dry Lake section but this is over 1,000 feet more than was measured by Parks (1949). The former have listed over 130 species from the most typical strata of

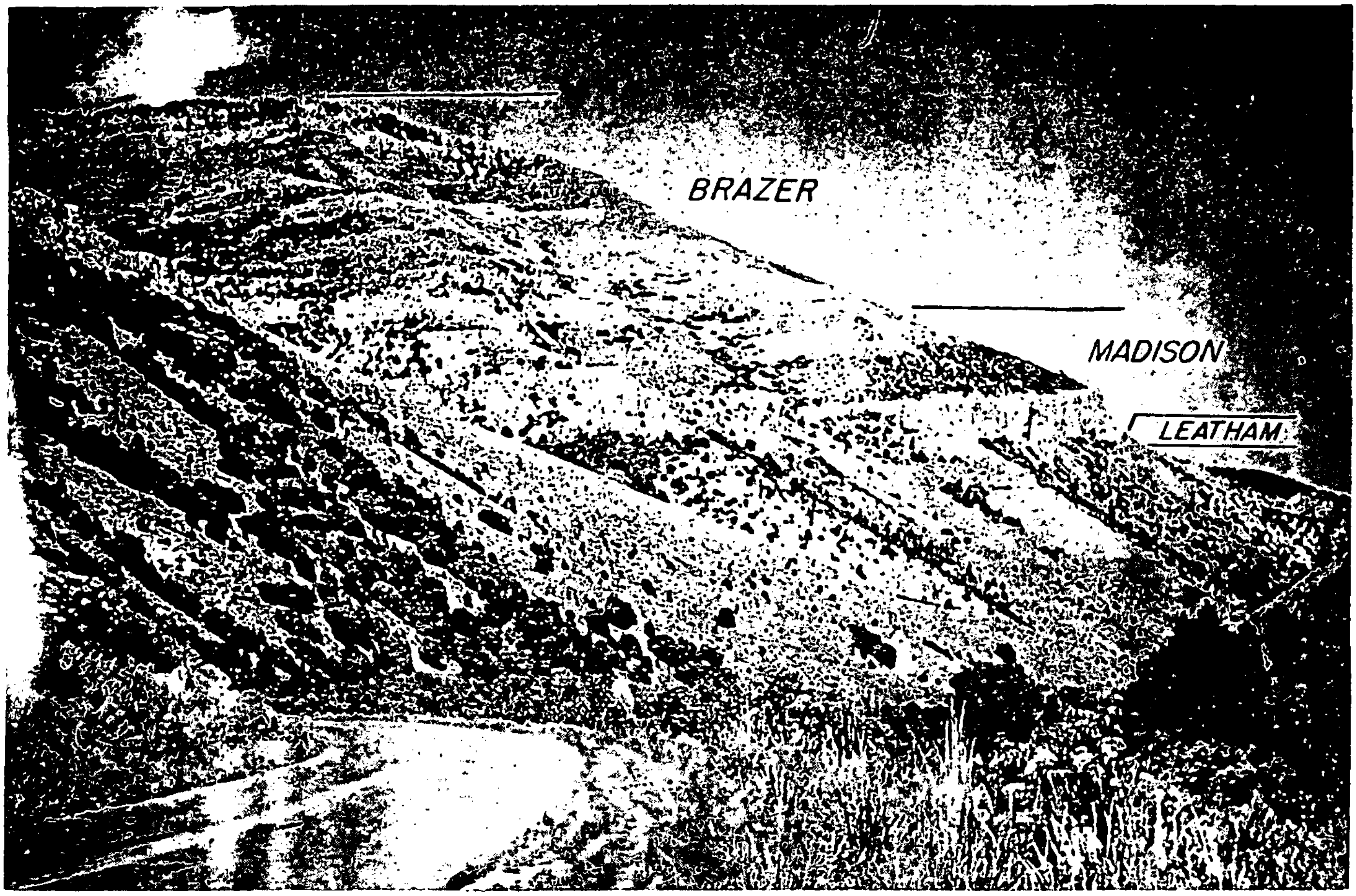

FIGURE 2.—Mississippian section on the east slope of Beirdneau Peak viewed from U. S. Highway 89 about 5 miles east of mouth of Logan Canyon, Utah.

the formation, the thin- to medium-bedded, dark-gray to grayish-black cherty limestones of the middle Brazer. Caninia, other large tetracorals, **Lithostrotion whitneyi** Meek, **Spirifer brazerianus** Girty, and Chonetes are abundant and frequently excellently preserved; many specimens are silicified and suitable for acid etching.

Entering Logan Canyon east of Logan one can see the cliffs of nearly flat-lying resistant Mississippian limestones near the axis of the Logan syncline. About 5 miles from the mouth of the canyon an excellent view of the entire Mississippian section is obtained from the road (Figure 2).

There on the east slope of Beirdneau Peak the Upper Devonian "Contact Ledge" can be seen as a thin zone of resistant limestone marking the top of the Devonian section. Above this, a slope is formed on the Lower Mississippian Leatham formation (about 75 feet thick). The Leatham consists of shales, sandy shales, and dark reddish-gray, nodular limestones characterized by abundant nodules 1 to 2 inches in diameter containing **Rhipidomella missouriensis** (Swallow) and Syringothyris. At the type section in Leatham Hollow, about 9 miles to the south, the base is marked by a 3-inch conglomeratic limestone, bearing angular chert nodules, limestone pebbles, and an occasional fragmental fish tooth.

Above the Leatham the Madison limestone rises in a sheer cliff, locally known as the "Chinese Wall". This part of the Madison is about 250 feet thick, and is composed of dark-gray, fine-crystalline limestone rhythmically interbedded with thin shaly limestone beds. A long steep slope rises to the base of a second cliff of the Madison, which may be termed the "Upper Chinese Wall". This middle slope of the Madison is formed on dark-gray, fine-crystalline to sublithographic, thinly-bedded limestone rhythmically interbedded with ⅛-inch beds of grayish-orange, soft, silty to argillaceous limestone. The lithologic character of most of the "Upper Chinese Wall" resembles that of the slope below, but this part appears to be more resistant to weathering and erosion. At several levels, benches or reentrants are weathered into the cliff, so that this upper cliff is not as well-defined as the lower cliff of the Madison. This thin-bedded limestone is the lithologic and faunal equivalent of the Lodgepole limestone of the Logan, Montana area. Osagian elements are in general lacking from the fauna. Whether never deposited, or removed by erosion, there does not seem to be an equivalent of the thick-bedded Mission Canyon portion of the Madison of Montana present in this area. The fauna of the Madison is characterized by tetracorals of smaller size than those of the Brazer, abundant **Syringopora**, **Spirifer** cf. S. centronatus, and abundant gastropods and cactocrinids.

Locally the base of the Brazer is marked by a phosphatic shale member which seems to have been deposited on the eroded upper Madison surface. Williams (1939, 1943, p. 595) has reported this basal phosphatic shale in Blacksmith Fork Canyon but it is missing in Leatham Hollow, 2 miles to the north. Williams (1943, p. 611) mentions the variety of lithologic types in the various exposures of Brazer, but says that each section generally contains some intercalated limestone and sandstone and generally some pure thick-bedded limestones (note cliffs near the top of Beirdneau Peak, Figure 2). The Wells is not exposed on the north side of Logan Canyon but appears in an incomplete section atop Logan and Millville Peaks, the high peaks just east of the town of Logan and south of lower Logan Canyon.

Steeply tilted Madison beds crop out in several small areas south and east of Laketown in the Randolph Quadrangle and then disappear under the cover of the Wasatch formation. The Brazer also outcrops about a mile east of Laketown with a bed of phosphate rock near the base.

From Sage Creek Junction the escarpment of the Crawford Mountains can be seen to the southeast. The Brazer forms this scarp, and here the Madison forms the upper slopes and the crest of the mountains. The 200 feet (or at least the upper portion) of the "thin-bedded impure earthy-gray limestone, which weathers to yellowish and reddish tints" reported by Richardson (1941, p. 20) to underlie conformably the Madison in the Crawford Mountains probably represents the Leatham formation. The Brazer type section in this area has been restudied by Williams (1943, p. 610) who states that neither the top nor the bottom is exposed, and that the limestones are dolomitized. Mississippian fossils are rare and poorly preserved in the area.

Montpelier Through Georgetown Canyon and Return. — The large fault block that rises northeast of Montpelier is composed of Madison limestone. Brazer limestone is present on the west slope of the hills just east of town, but in this area the Mississippian is faulted and the section is incomplete.

The Mississippian is next seen in a broad strip along the west side of Crow Creek; the outcrop of the Brazer limestone is crossed at the entrance to Wells Canyon. At the mouth of Wells Canyon a partial section (the base is covered by hill wash on the east) of Brazer was reported by Mansfield (1927, p. 63) as 1,130 feet thick. The section dips westward into the Webster syncline of Mansfield which is marked by the Pruess Range. One-foot to three-foot beds of dark-gray limestone mark the lower part of the section here, with whitish sandstones and light sandstones exposed above. In this section shaly, cherty, limestone marks the top of the Brazer, underneath sandstone of the basal Wells formation.

Brazer beds of essentially the same lithology form the crest of anticlinal Snowdrift Mountain, and the route passes through it along the South Fork of Deer Creek.

Although Madison is not exposed in the Crow Creek Quadrangle, it crops out at a number of places in the Slug Creek Quadrangle to the west. The high ridge west of Georgetown Canyon is formed by a large portion of Madison brought up by faulting. The Brazer is lower on the canyon walls and the trip crosses a narrow slice, dipping 75° west, brought up by a thrust subordinate to the main overthrust.

A spectacular portal or gateway at the mouth of Georgetown Canyon is formed by ledges of Madison limestone.

Alpine, Idaho, to Jackson, Wyoming. — The Madison and Brazer limestones make up the rugged mountains along the northeast edge of the Snake River Valley from Alpine to Swan Valley, Idaho.

West of the Snake River only one section of Madison is crossed by the route and this lies at the entrance to Fall Creek Canyon. Here in the Fall Creek Quadrangle, however, the Brazer is well exposed along the west side of the Snake River fault, and a complete section is obtained in Fall Creek Canyon. South of Fall Creek the Brazer runs along the axis of the Snake River anticline (Kirkham, 1924). In this area the Madison and Brazer are each about 1,000 feet thick. Each are cliff-makers but the Brazer is again the more massive, and although each is dominantly made up of dark-gray, fine- to coarse-crystalline limestone, the *Brazer again weathers to the lighter color, being light-*gray or almost white. As in the Utah area the faunas of both formations are dominated by rugose corals, those in the Brazer being much larger, generally 3 to 8 inches long.

Along Pine Creek, west of the fault, typical Madison and Brazer are exposed. In the west part of the canyon the beds dip about 25° to the west, but farther eastward dips as high as 75° are encountered near the main fault.

East of Victor, Carboniferous limestones rise from under the cover of Mesozoic and Cenozoic sediments of the Teton Basin to form the gently-dipping western lope of the Tetons. Horberg (1938, p. 16) states that "the tabular inter-stream areas and most of the important sedimentary peaks (Mt. Hunt, 10,775 feet, Rendezvous Peak, 10,924 feet, and Fossil Mt., 10,553 feet) are formed of these [Madison and Brazer] limestones." The Madison and Brazer form the bulk of the mountain just north of Teton Pass. Here the Brazer has thinned considerably and is subordinate to the Madison. In general the distinguishing characteristics of the limestones are the same in this area as in areas to the south.

Thomas (1948) states that Bachrach (1946) has recognized the Brazer over a wide area in the Hoback and Gros Ventre Mountains with the Darwin sandstone everywhere present above the Brazer. Thomas considers the Darwin as basal Pennsylvanian in this area.

The Gros Ventre Buttes northwest of Jackson are similar to each other in structure and composition. They represent normal fault blocks of gently-dipping Paleozoic strata and younger lava flows tilted westward along their eastern scarp slopes. The Madison crops out on the southeast corner of each of the buttes and Horberg (1938, p. 42) reports that a tunnel dug west of Jackson on East Gros Ventre Butte has penetrated the talus and exposed the slickensided, polished surface of Madison limestone forming the footwall in contact with breccia and talus on the east.

Two smaller buttes south of the main Gros Ventre buttes expose Madison limestone (and other Paleozoics) as remnants of the southwest-dipping Jackson thrust plane.

ACKNOWLEDGMENTS

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REGIONAL STRATIGRAPHY OF THE DEVONIAN SYSTEM IN NORTHEASTERN UTAH, SOUTHEASTERN IDAHO, AND WESTERN WYOMING

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STRATIGRAPHY

Figure 3 shows a cross-section of Devonian rocks extending from southwesternmost Montana and immediately adjacent Idaho eastward to western Wyoming (Cody area, Wind River Range, and Teton Range) and thus southwestward to northeastern Utah.

In northern and northeastern Utah relatively rapid thinning with correspondingly rapid changes in vertical stratigraphic sequence characterize rocks of Devonian age. The thickest portion of the Devonian, as exposed in Logan Canyon (Section 8, Figure 3), is composed of three stratigraphic units described in detail by Williams (1948). The units are, from bottom to top, the Water Canyon formation, a light-gray to almost white, coarsely-crystalline, sandy, dolomitic limestone, the Hyrum member of the Jefferson formation, a drab, dark-gray, medium- to coarsely-crystalline secondary dolomite, and the Beirdneau member of the Jefferson formation, a light-tan to buff unit composed of platy beds of siltstone and dolomite with partings of tan shale being prominent throughout the sequence (Figure 1). Local zones of sedimentary breccia are prominent in some parts of the section. The Beirdneau member exposed in Logan Canyon is some 400 feet thick. Holland (1952), describing what is apparently the uppermost portion of this unit, exposed a few miles south of Logan Canyon in Leatham Hollow, has measured a thickness of some 70′ of rocks of similar lithology to which he assigns a Kinderhookian age and which he correlates with the Sappington sandstone of southwestern Montana. In Leatham Hollow this unit, named the Leatham formation by Holland, rests disconformably on the dark-gray limestones and dolomites of the Jefferson formation (presumably the Hyrum member of the Jefferson of Williams in Logan Canyon). Absence of so great a thickness, in view of its presence a few miles to the north seems somewhat anomalous. However, since the writers have not visited the Leatham Hollow area they do not presume to offer an explanation of the apparent anomaly.

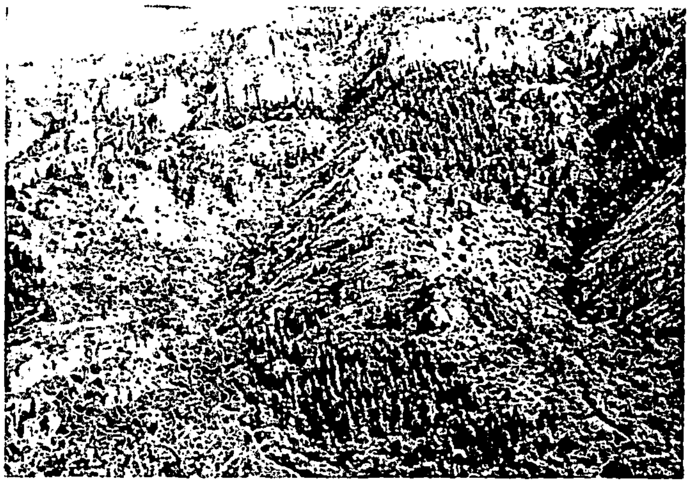


FIGURE 1.—South wall Logan Canyon east of Logan, Utah. Massive cliffs on upper wall are Madison limestone. Slope-making sequence to lower, less well developed cliffs is Beirdneau member, Jefferson formation. Lower cliffs mark upper portion of Hyrum member, Jefferson formation.

The Water Canyon formation at the base of the sequence is not recognized elsewhere in the area of the field trip. However to the south and west, in central Utah, correlable units are exposed. The Hyrum member of the Jefferson is likely correlative with the Jefperson formation which lies at the base of the Devonian section in areas to the north, east, and south. The Beirdneau member of the Jefferson formation may be chronologically a close correlative of the Three Forks formation, recognized elsewhere in the region. However, the lithologies are similar only in that both represent a change in late Devonian time from purely carbonate deposition to that of a more clastic nature.

To the east (Laketown-Randolph area, section 7, Figure 3) two units of Devonian age are recognized. The Jefferson formation, a dark- to medium-gray, medium- to coarsely-crystalline unit composed of varying beds of limestone and secondary dolomite is conformably overlain by the Three Forks formation which consists of a shaley, very thinly-bedded, olive-gray limestone typically making a topographic saddle in areas of dipping beds and almost always characterized by a red soil weathering zone at the outcrop surface (Figure 2). To the south at Durst Mountain (Section 9, Figure 3) east of Ogden, Utah, the Devonian is represented by the Three Forks formation which is composed predominantly of limey and sandy shales which are olive to buff in color and which show the red soil weathering zone typical of the Three Forks at other localities.

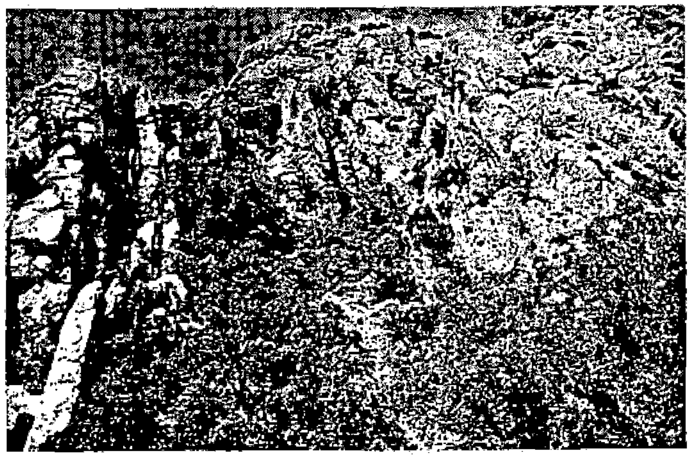


FIGURE 2.—North wall Laketown Canyon, east of Laketown, Utah. Saddle in ridge crest to left of center is typical of Three Forks formation. Units to right are Mississippian limestones. Strata to left of saddle are uppermost Jefferson formation.

Still farther to the south in the vicinity of Oakley, Utah (Section 10, Figure 3) Devonian rocks are represented by a very thin section of tan calcareous shales and siltsones which rests disconformably on a quartzite of questionably Cambrian age and which apparently grade upward into rocks of Mississippian age. A formation name has not been assigned to these rocks, but presumably they are genetically related to the Three Forks formation mentioned to the north, representing a slightly variant shelf environment. A fauna of Hackberry age is present in these beds. This section is the most eastward exposure of Devonian rocks so far noted in Utah and, in view of the thinness and clastic nature of the sediments, presumably represents deposition not too far removed from the zero edge to the east.

In Wyoming, the Devonian interval is termed the "Darby formation", but approximate lithologic equivalents of the Three Forks-Jefferson may be differentiated. The upper part of the Darby sequence is characterized by conspicuous amounts of clastics (sand, silt, and argillaceous material) interbedded with the carbonates and may be correlated with the Three Forks. The remaining lower relatively pure carbonate beds are considered Jefferson equivalents. Eastward in the shelf area of Wyoming, this Jefferson interval thins and is not recognizable near the eastern zero edge. The Three Forks beds appear to be transgressive eastward in Wyoming.

In southwestern Montana and adjoining Idaho, a two-fold division of the Devonian is recognized. The two units are roughly correlable with the Three Forks and Jefferson of central Montana. In the latter area, a pre-Jefferson basal clastic unit is also recognized and the Jefferson is divisible into an upper dolomite member and a lower limestone member. In sections 1 and 2 the lowest Devonian beds appear to be lithologic equivalents of the dolomite member of the Jefferson, and the lower limestone member is not lithologically distinguishable. However, this limestone member becomes recognizable a short distance to the north. The Three Forks interval of southwestern Montana is made up of shale and argillaceous dolomite beds containing a thin, varicolored, solution-brecciated zone at the base. Light-brown or orange-weathered, fine, sandy beds are developed in the upper part of this clastic sequence, directly below the Madison strata. These sandy beds may be lithologic equivalents of the Sappington sandstone in the Logan area of Montana, where they have been included in the Devonian by Sloss and Laird (1947). Recently, Holland (1952) has assigned a Kinderhookian age to the Sappington sandstone developed at Logan, Montana, and correlates the zone with his Leatham formation of northeastern Utah mentioned above. The exact relationship of the Three Forks beds with the Darby of western Wyoming requires further study, but the upper part of the Darby appears to be related to the Three Forks in that it also contains prominent amounts of clastics. The Darby typically shows a shelf sequence made up predominantly of secondary dolomites with variable amounts of normal marine limestone. The sandstone has variable amounts of carbonate cement and passes laterally into sandy carbonates and pure carbonates. Green or gray clay shales are also developed in thin beds or partings.

ISOPACH AND FACIES MAP

Isopach Pattern

The isopach pattern (Figure 4) suggests the existence within the area of the map of two different tectonic environments. Stable shelf conditions in the Wasatch Range—western Uinta Range area and in the area of west-central Wyoming and eastern Idaho are indicated by the rather broad extent in both areas of relatively thin Devonian sediments. The thickness of the Devonian section in these shelf areas averages between 150 and 300 feet. A gradual westward thickening from the eastern zero edge is well portrayed, particularly in the Wyoming Shelf (Andrichuk, 1951). The shelf areas are bounded on the west by an irregularly trending axis of rapid increase in thickness. This zone presumably represents the tectonic "hinge" between the shelf areas to the east and the more negative geosynclinal areas to the west. The Wyoming Shelf is likewise bounded on the north, off the area of the map, by a west to east axis of rapid thickening trending across southwestern Montana. Devonian sediments to the north in Montana increase to a thickness of about 1,000 feet, while to the west and southwest of the Wyoming Shelf, in Idaho, and the northeastern edge of Utah, thicknesses in excess of 3,000 feet are known.

The smaller shelf in north-central Utah is separated from the Wyoming shelf by the geosynclinal embay-

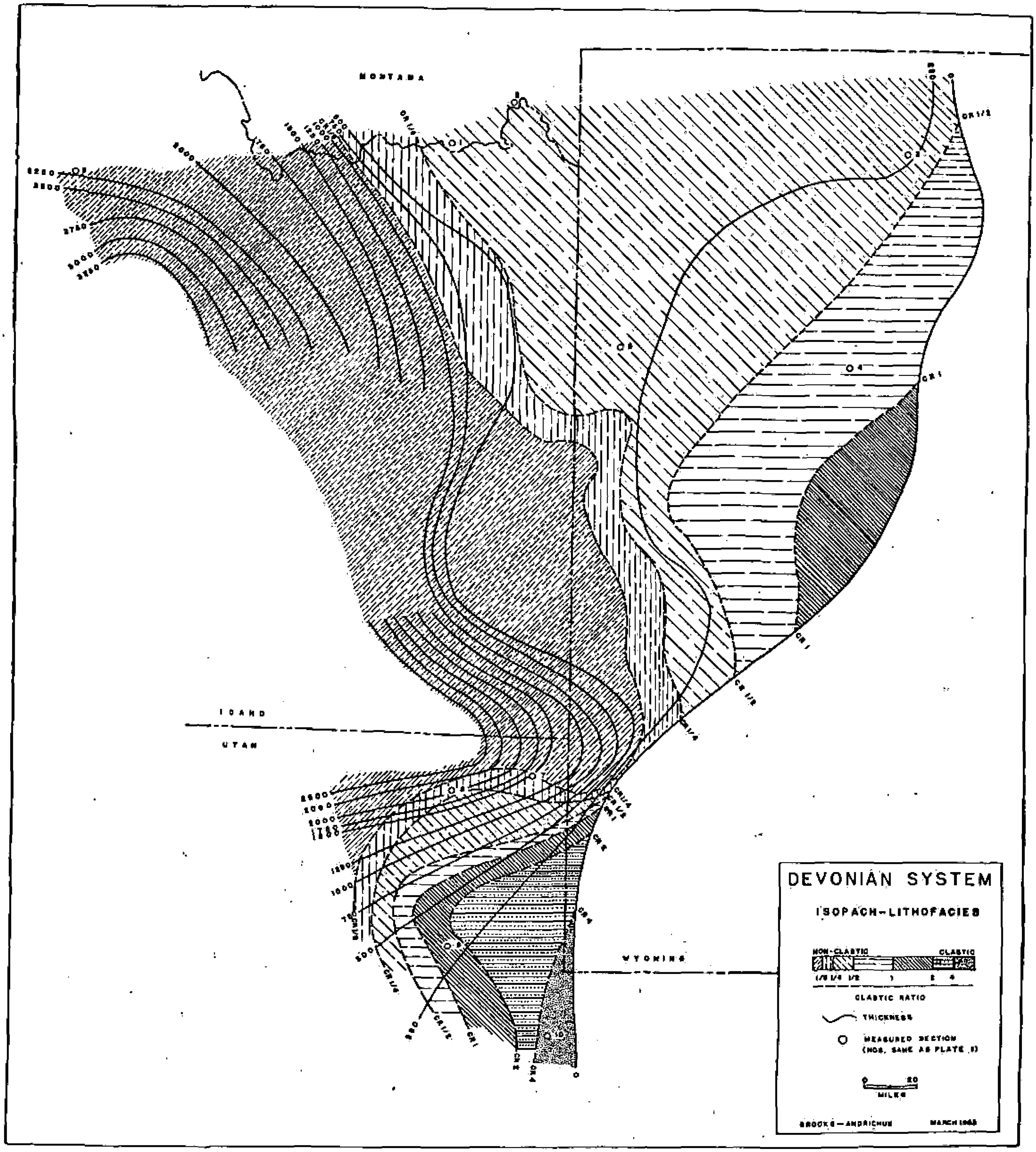
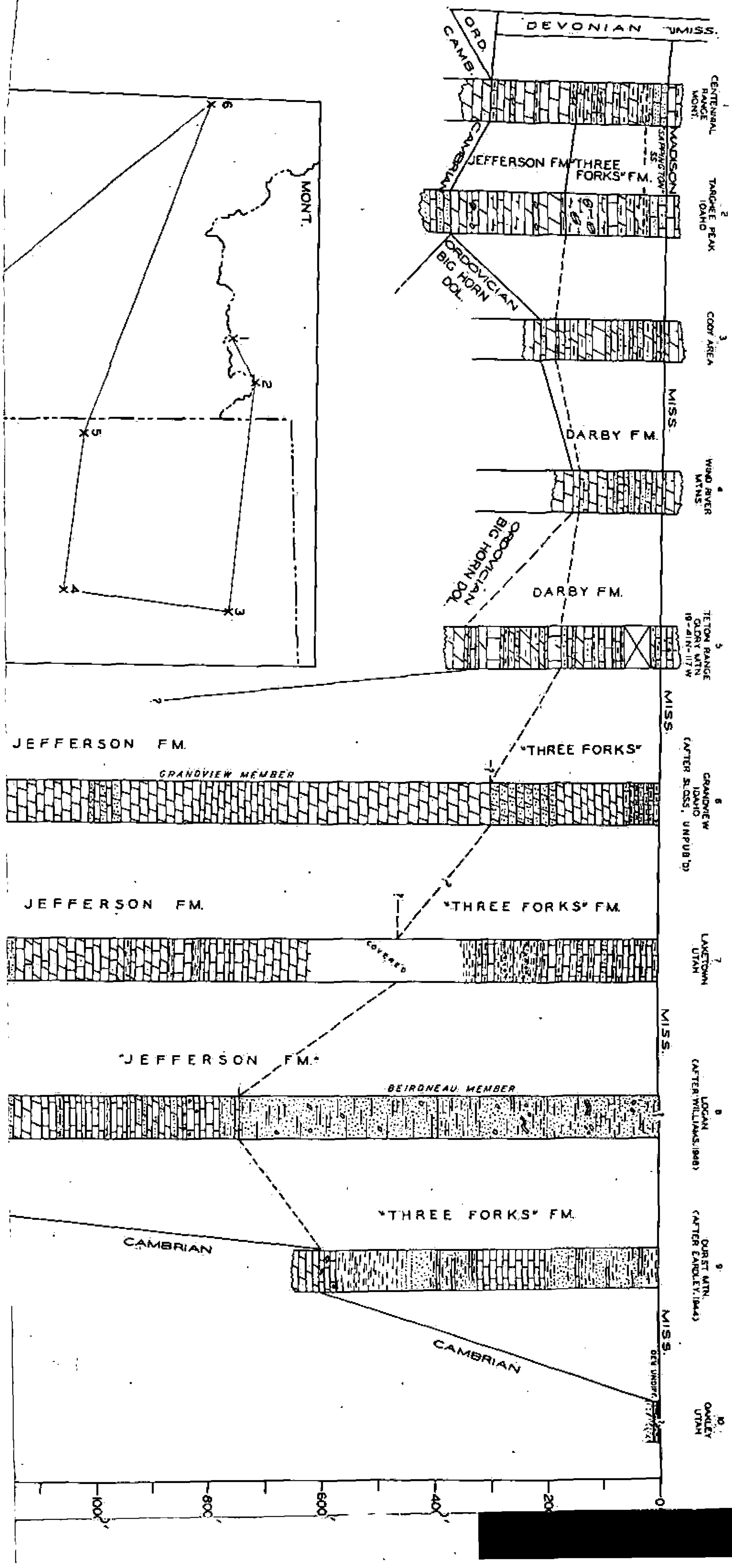


FIGURE 4.



ORD.
CAMB.
DEVONIAN
MISS.
1 CENTENNIAL RANGE MONT.
MADISON
"SAPPINGTON" SS.
CAMBRIAN
JEFFERSON FM.
"THREE FORKS" FM.
2 TARGHEE PEAK IDAHO
ORDOVICIAN BIG HORN DOL.
3 CODY AREA
MISS.
DARBY FM.
4 WIND RIVER MTNS.
ORDOVICIAN BIG HORN DOL.
DARBY FM.
5 TETON RANGE GLORY MTN 19-41N-117W
MONT.
6
1
2
5
4
3
MISS.
JEFFERSON FM.
"THREE FORKS"
GRANDVIEW MEMBER
6 GRANDVIEW IDAHO (AFTER SLOSS, UNPUB'D)
JEFFERSON FM.
"THREE FORKS" FM.
COVERED
7 LAKETOWN UTAH
MISS.
"JEFFERSON FM."
BEIRDNEAU MEMBER
8 LOGAN (AFTER WILLIAMS, 1948)
"THREE FORKS" FM.
CAMBRIAN
9 DURST MTN. (AFTER EARDLEY, 1944)
MISS.
CAMBRIAN
DEV UNDIFF.
10 OAKLEY UTAH
0
200
400
600
800
1000'

ment in northeastern Utah. Thickening from the very thin sediments of the Utah shelf, to the geosynclinal embayment to the north is moderately rapid, with thicknesses of approximately 2,500 feet being attained. However, to the west of the Utah shelf thickening toward the geosyncline is more gradual, and geosynclinal sediments in western Utah average about 1,500 feet.

Lithofacies Pattern

Clastics are important constituents of the section in both shelf areas mentioned above, and exhibit a gradual decrease in proportion westward from the zero edge. Near the present eastern limit of Devonian occurrence, the clastics locally may be quantitatively more important than the carbonates. In the shelf area of Wyoming they generally constitute at least 20 per cent of the total section. Non-clastics form over 80 per cent of the total section in the adjoining basinal areas to the west and southwest. In Wyoming the clastics become much coarser near the eastern zero edge. In the Utah shelf, which is considerably narrower than that in Wyoming, clastics near the eastern edge dominate the section, constituting at least 75 per cent of the sediments present. This condition rapidly changes to the west until in the exposures at the western edge of the shelf in the Wasatch Range carbonates constitute the greater part of the section.

TECTONIC AND ENVIRONMENTAL INTERPRETATION

The Utah and Wyoming shelf areas behaved as relatively positive areas on which sedimentation commenced somewhat later than in the adjoining negative areas to the west and north. Devonian clastic deposits of the Utah shelf represent deposition under stable conditions. The relatively fine sands and silts and clays are of a clean character and suggest deposition under conditions of stability with reworking of the depositional interface for considerable lengths of time before lithification was completed. Carbonates at the western edge of the Utah shelf likewise represent stable conditions, being of normal marine limestone type. Similarly, to the northeast, in the Wyoming shelf area the clastics and carbonates of the Darby represent shelf-type deposits laid down under relatively near-shore conditions of considerable stability. Quartzose sands, showing lateral intergradations with carbonates, green clay shales, and well developed secondary dolomites all indicate slow deposition on a slowly subsiding platform, permitting winnowing out of fine clastic material and dolomitization of the limestones. The deposits are characterized by evidences of disconformities, especially in the eastern areas. The increasing clastic content and coarsening of clastics of the east indicate that the present zero edge was also the approximate eastern depositional limit. The adjacent landmass to the east was apparently sufficiently positive to furnish the coarse and fine clastics which are prominent at the sight of deposition.

The locus of rapid change in thickness bounding the generally western edge of both the Wyoming and Utah shelves represents a tectonic hinge which, consequently, also forms the eastern boundary of the irregularly trending geosynclinal belt. The fact that carbonate sediments are predominant in this more negative belt indicates that the belt lay a considerable distance from the land area to the east from which the sediments of the region were likely derived. Even in the area of the southeastern Idaho embayment, which extends considerably closer to the sediment source area than do other parts of the geosynclinal element, the section is composed predominantly of non-clastic material, although in this area clastics do become more noticeable in the section (for example in the Logan, Utah area, Section 8, Figure 3).

Patterns shown in Idaho are constructed on the basis of available published information (Mansfield, 1927, Ross, 1934, 1937; Umpelby, 1913, 1917; Umpleby, et al., 1930) and by use of unpublished material received from L. L. Sloss and used by permission of Phillips Petroleum Company.

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