THE GEOLOGICAL SURVEY OF WYOMING

Daniel N. Miller, Jr., State Geologist

MEMOIR No. 1

A REGIONAL STUDY OF ROCKS OF PRECAMBRIAN AGE IN THAT PART OF THE MEDICINE BOW MOUNTAINS LYING IN SOUTHEASTERN WYOMING— WITH A CHAPTER ON THE RELATIONSHIP BETWEEN PRECAMBRIAN AND LARAMIDE STRUCTURE.

By

R. S. Houston and others

Second printing of 2000 copies, July 1978, by Frontier Printing, Inc., Cheyenne

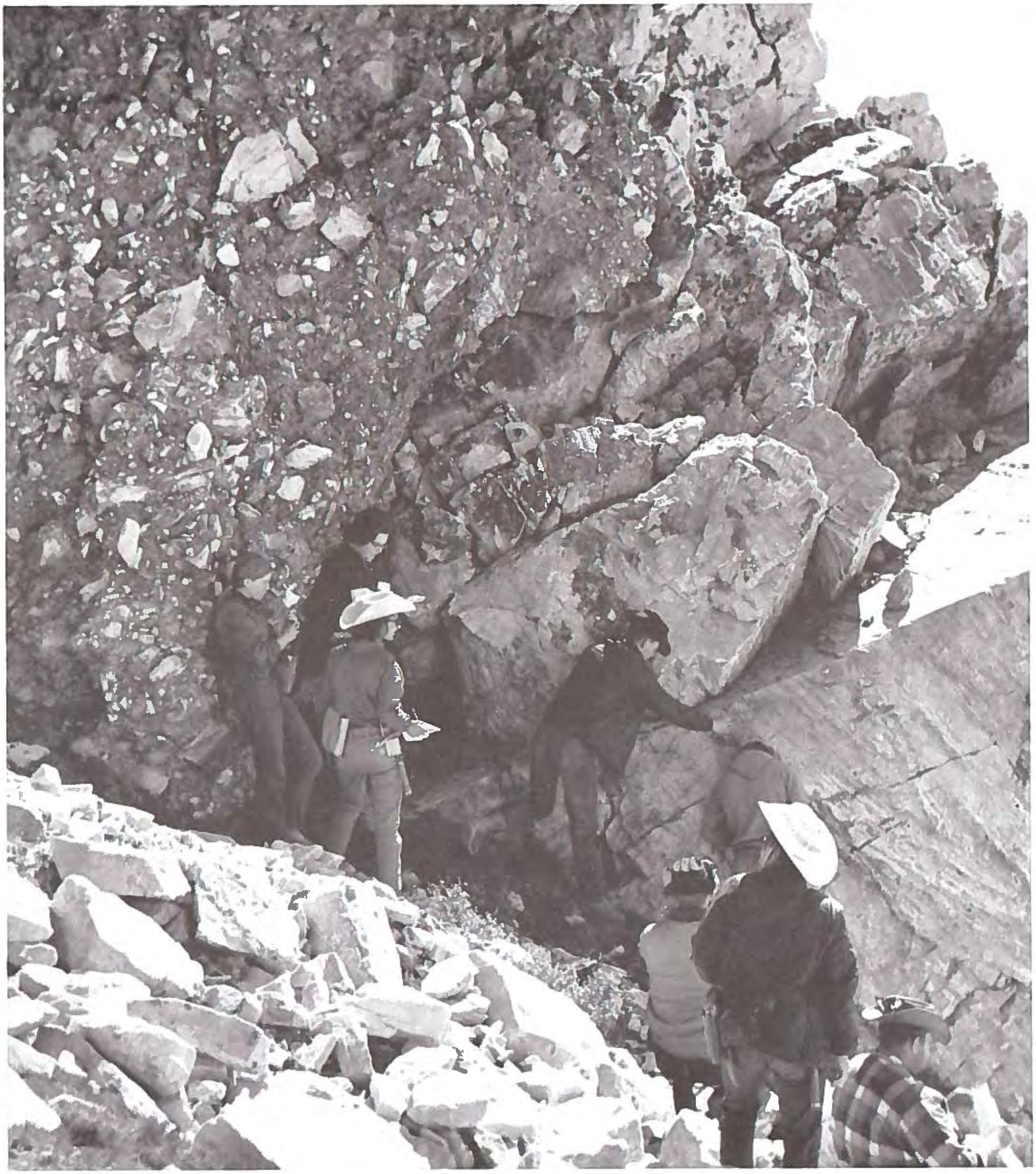
For sale by The Geological Survey of Wyoming. Write

The Geological Survey of Wyoming
P.O. Box 3008, University Station
Laramie, Wyoming 82071,

or visit the Geological Survey Building, in Laramie.

Cover. View, looking north, of the Corner Mountain thrust fault, of Laramide age. The low-angle thrust, dipping approximately seventeen degrees to the west, is expressed by a topographic break. In the foreground, S. H. Knight explains his sketch of the structure on the blackboard: rocks of Precambrian age are thrust over sedimentary rocks ranging in age from Cretaceous to Early Tertiary. Photograph taken at a locality about one mile east of Centennial, Wyoming.

THE GEOLOGICAL SURVEY OF WYOMING
MEMOIR NO. 1



Frontispiece—View of Medicine Peak quartzite taken near crest of the Snowy Range axial remnant above Lake Marie. Large scale cross-bedding can be seen in lower right where plane of dip fault is exposed. Fault breccia in upper left. View looking northeast parallel to strike of bedding in quartzite.

THE GEOLOGICAL SURVEY OF WYOMING

Daniel N. Miller, Jr., State Geologist

MEMOIR No. 1

A REGIONAL STUDY OF ROCKS OF PRECAMBRIAN AGE IN THAT PART OF THE MEDICINE BOW MOUNTAINS LYING IN SOUTHEASTERN WYOMING— WITH A CHAPTER ON THE RELATIONSHIP BETWEEN PRECAMBRIAN AND LARAMIDE STRUCTURE.

by

R. S. Houston and others

Contributions by: M. E. McCallum, J. S. King, B. B. Ruehr, W. G. Myers, C. J. Orback, J. R. King, M. O. Childers, Irwin Matus, D. R. Currey, J. C. Gries, H. L. Stensrud, E. J. Catanzaro, M. N. Swetnam, D. D. Michalek, and D. L. Blackstone, Jr.

Age determinations by: F. Allan Hills, Paul W. Gast, and Ian Swainbank.

Laramie, Wyoming

1978

PREFACE TO THE SECOND PRINTING

During the ten years since publication of this memoir, geologic studies have been made of a number of areas in the Medicine Bow Mountains and adjacent areas that have substantially advanced our understanding of the geology. In the southern Medicine Bow Mountains, quadrangle mapping on a scale of 1/24,000 has been completed of the Lake Owen, Keystone, and Albany quadrangles (Houston and Orback, 1976; McCallum and Houston, in preparation; McCallum, in preparation). These studies, which are supported by the United States Geological Survey, were undertaken to evaluate the platinum and gold potential of this part of the Medicine Bow Mountains; of particular geologic interest is the discovery that both the Lake Owen mafic complex and Mullen Creek mafic complex are layered complexes containing abundant primary structures such as graded bedding and channeling. The Lake Owen mafic complex contains leucocratic rocks and is magnetite rich; it is made up of cyclic units, each of which ranges from troctolite at its base to magnetite gabbro at its upper part (Houston and Ridgley, 1976, p. 68).

Geologic and geochemical studies have been made of the Sheep Mountain area of the eastern Medicine Bow Mountains as part of an evaluation of the Sheep Mountain Wilderness study area (Houston, Patten, and Gersic, in press), and a similar study is being undertaken in the northern Medicine Bow Mountains to evaluate the economic potential of the Snowy Range Wilderness study area. These studies are by the United States Geological Survey and the United States Bureau of Mines. The Snowy Range study is of particular interest because the initial work demonstrates that there may be a potential for quartz-pebble conglomerate-type uranium in some of the metasedimentary rocks of this area. This indication of possible uranium potential has made expanded studies necessary which are a cooperative effort of the United States Geological Survey, the Department of Geology, University of Wyoming, the Geological Survey of Wyoming, and the United States Department of Energy.

The stratigraphy and depositional environments of the metasedimentary rocks of the northern Medicine Bow Mountains are better understood as a result of these Snowy Range projects, and geologic reports by Karlstrom (1977), Sylvester (1973), and Lanthier (1978) give new information on stratigraphy and depositional environments of the metasedimentary rocks. A tentative revision of stratigraphic nomenclature for metasedimentary rocks of both the Medicine Bow Mountains and the Sierra Madre is in Houston, Karlstrom, and Graff (in press). Essentially, a supergroup is proposed that includes a Phantom Lake Group of metasedimentary and metavolcanic rocks (the lower Deep Lake Formation of Memoir One), a Deep Lake Group of metasedimentary rocks (the upper Deep Lake Formation of Memoir One), and a Libby Creek Group of metasedimentary rocks that is a slightly modified version of the Libby Creek Group of Memoir One. Readers interested in the stratigraphic succession are urged to read these papers because they contain much more detailed information than that presented in Memoir One.

Reports on the uranium-thorium bearing quartz-pebble conglomerate of the Medicine Bow Mountains and of similar conglomerate in the Sierra Madre are in Miller, Houston, Karlstrom, Hopkins, and Ficklin, 1977; Houston, Graff, Karlstrom, and Root, 1977; Graff and Houston, 1977; and Houston, Karlstrom, and Graff, in press. These reports describe the quartz-pebble conglomerate, outline its distribution, give a preliminary discussion of mineralogy, and compare the conglomerate with similar rocks of the Blind River District of Canada and the Witwatersrand of South Africa.

Several excellent reports on structure and stratigraphy of Paleozoic, Mesozoic, and Tertiary rocks of the northern and eastern margins of the Medicine Bow Mountains have been published by the Geological Survey of Wyoming (Blackstone, 1969; Blackstone, 1973; and Blackstone, 1976). These reports present new information on stratigraphy of the rocks of Tertiary age and give a detailed structural analysis of folds and faults of Laramide and younger ages. The reports are particularly useful to individuals interested in the oil and gas possibilities of the area.

Recently published specialized reports which may also be useful to individuals interested in the geology of the Medicine Bow Mountains include reports on Earliest Eocene mammalian fossils (Prichinello, 1971), Upper Cretaceous and Lower Tertiary stratigraphy (Gill, Merewether, and Cobban, 1970), and stromatolites of the Nash Formation of Precambrian age (Knight, 1968), and an open-file aeromagnetic map published by the United States Geological Survey (Ketterer, 1976).

Finally, mention should be made of new interpretations of Medicine Bow Precambrian geology that have come about as a result of the development by geophysicists and geologists of plate tectonic models that can be tentatively applied to rocks older than the Mesozoic. Current views are that the Mullen Creek–Nash Fork shear zone may be part of a suture where island arcs that developed in the area that is now Colorado were attached to the stable core of the Wyoming Province about 1700 million years ago (Hills, Houston, and Subbarayude, 1975; Camfield and Gough, 1977; Warner, 1977). Details of this model, along with new data on stratigraphy, structure and economic geology of the Medicine Bow Mountains, will be presented in a forthcoming issue of Contributions to Geology (Boyd, in prep.) to be published this year.

The value of Memoir One is in its regional perspective and descriptive geology. The geologic framework of the Medicine Bow Mountains has not been substantially altered by more detailed mapping or the application of new concepts, but this brief review shows that substantial new information has accumulated in the last decade. I believe it serves to show once again that the Medicine Bow Mountains is a remarkable area where more geologic history is compressed into a small area than in most other parts of the planet. Furthermore, it is still possible to study horn toads and cacti at the base of the mountain and ptarmigan and tundra at the top – let's hope it remains this way.

R.S. Houston
June 1978

REFERENCES

Blackstone, D.L., Jr., 1969, Structural geology of the Rex Lake Quadrangle, Laramie Basin, Wyoming: Geol. Survey of Wyoming Prelim. Rept. 11, 17 p.

Blackstone, D.L., Jr., 1973, Structural geology of the eastern half of the Morgan Quadrangle, the Strouss Hill Quadrangle, and the James Lake Quadrangle, Albany and Carbon Counties, Wyoming: Geol. Survey of Wyoming Prelim. Rept. 13, 45 p.

Blackstone, D.L., Jr., 1976, Structural geology of the Arlington-Wagonhound Creek area, Carbon County, Wyoming – A revision of previous mapping: Geol. Survey of Wyoming Prelim. Rept. 15, 15 p.

Boyd, D.W., editor, (in preparation), Wyoming Uranium Issue II: Contributions to Geology (University of Wyoming), vol. 17, no. 1.

Camfield, P.A., and Gough, E.I., 1977, A possible Proterozoic plate boundary in North America: Canadian Jour. Earth Sciences, v. 14, no. 6, p. 1229-1238.

Gill, J.R., Merewether, E.A., and Cobban, W.A., 1970, Stratigraphy and nomenclature of some Upper Cretaceous and Lower Tertiary rocks in south-central Wyoming: U.S. Geol. Survey Prof. Paper 667, 53 p.

Graff, P.J., and Houston, R.S., 1977, Radioactive conglomerate in Proterozoic (Precambrian X) metasedimentary rocks of the Sierra Madre, Wyoming: U.S. Geol. Survey open-file report 77-830, 7 p.

Hills, F.A., Houston, R.S., and Subbarayudu, G.V., 1975, Possible Proterozoic plate boundary in southern Wyoming: Geol. Soc. America, Rocky Mtn. Sec. Program with Abstracts, Boise, Idaho, p. 614.

Houston, R.S. and Orback, C.J., 1976, Geologic map of the Lake Owen Quadrangle, Albany County, Wyoming: U.S. Geol. Survey Map GQ-1304.

Houston, R.S., and Ridgley, J.L., 1976, Cyclic units in the Lake Owen mafic complex, Albany County, Wyoming: U.S. Geol. Survey Prof. Paper 1000, p. 68.

Houston, R.S., Graff, P.J., Karlstrom, K.E., and Root, F.K., 1977, Preliminary report on radioactive conglomerate of middle Precambrian age in the Sierra Madre and Medicine Bow Mountains of southeastern Wyoming: U.S. Geol. Survey open-file Rept. 77-574, 31 p.

Houston, R.S., Karlstrom, K.E., and Graff, P.J., (in preparation), Progress report on the study of radioactive quartz-pebble conglomerate of the Medicine Bow Mountains and Sierra Madre, southeastern Wyoming, with a chapt. on uranium and thorium minerals in Precambrian mataconglomerates, Medicine Bow Mountains, south-central Wyoming by G.A. Desborough and W.M. Sharp: U.S. Geol. Survey open-file report.

Houston, R.S., Patten, L.L., and Gersic, J., (in preparation) Mineral resources of the Sheep Mountain wilderness study area, Wyoming: U.S. Geol. Survey open-file report.

Karlstrom, K.E., 1977, Geology of the Proterozoic Deep Lake Group, Central Medicine Bow Mountains, Wyoming: unpublished M.S. thesis, Univ. of Wyoming, 116 p.

Ketterer, W.P., 1976, Aeromagnetic map of Keystone and vicinity, Wyoming: U.S. Geol. Survey open-file map 76-687.

Knight, S.H., 1968, Precambrian stromatolites, bioherms and reefs in the lower half of the Nash Formation, Medicine Bow Mountains, Wyoming: Contrib. to Geology, (University of Wyoming) v. 7, no. 2, p. 73-116.

Lageson, D.R., and Hausel, W.D., editors, 1978, Occurrences of Uranium in Precambrian and younger rocks of Wyoming and adjacent areas: Geol. Survey of Wyoming, Public Info. Circ. 7, 37 p.

Lanthier, L.R., 1978, Stratigraphy and structure of the lower part of the Precambrian Libby Creek Group, Central Medicine Bow Mountains, Wyoming: unpublished M.S. thesis, Univ. of Wyoming, 30 p.

Miller, W.R., Houston, R.S., Karlstrom, K.E., Hopkins, D.M., and Ficklin, W.H., 1977, Geological and geochemical investigations of uranium occurrences in the Arrastre Lake area of the Medicine Bow Mountains, Wyoming: U.S. Geol. Survey open-file rept. 77-92, 30 p.

McCallum, M.E., (in preparation), Geologic map of the Keystone Quadrangle, Albany and Carbon Counties, Wyoming: U.S. Geol. Survey Map GQ.

McCallum, M.E., and Houston, R.S. (in preparation) Geologic map of the Albany Quadrangle, Albany County, Wyoming: U.S. Geol. Survey Map GQ.

Prichinello, Kathryn, 1970, The mammalian fauna of Cooper Creek, Albany County, Wyoming: unpublished M.S. thesis, Univ. of Wyoming, 40 p.

Sylvester, G.H., 1973, Depositional environment of diamictites in the Headquarters Formation, Medicine Bow Mountains, southeastern Wyoming: unpublished M.S. thesis, Univ. of Wyoming, 83 p.

Warner, L.A., 1978, The Colorado lineament: a middle Precambrian wrench fault system: Geol. Soc. America Bull., v. 89, no. 2, p. 161-171.

CONTENTS

Page

Cover Photograph Title—Quartzite of Medicine Peak Quartzite exposed
near crest of Medicine Bow Mountains

ILLUSTRATIONS

Figure		Page

TABLE		**PAGE**

COVER PHOTO

View looking northwest showing Medicine Peak Quartzite cut by mafic dikes—central Medicine Bow Mountains.

ABSTRACT

The Medicine Bow Mountains of Wyoming are a complex uplifted mass of Precambrian rocks that is the core of a large asymmetric anticline bounded by thrust faults on the east side. The northern part of the mountain area is split into two divisions by a north plunging syncline. Sedimentary rocks of Paleozoic and Mesozoic age are along the flanks of the mountains and rocks of Tertiary age lie both along the flank and in great plates of gently dipping sedimentary rocks in the central part of the area. Glacial deposits of Pleistocene age are especially abundant in the northern part of the Medicine Bow Mountains where they originated as an ice cap near Medicine Bow Peak.

The rocks of Precambrian age are in the uplifted Medicine Bow Mountains proper, and in uplifted blocks along the northwest and southeast flanks of the mountains. The Precambrian rocks are divided into two major units by the Mullen Creek-Nash Fork shear zone that trends in a northeasterly direction throughout the central part of the mountains. North of the shear zone the older Precambrian rocks are a complex sequence of quartzo-feldspathic gneisses of the almandine-amphibolite facies. These gneisses are marked by a north to northwest trending structure and are cut by hundreds of dikes of basaltic composition that are also deformed and metamorphosed. Large and small bodies of granite, quartz diorite and gabbro, are present in the quartzo-feldspathic gneiss. Age determinations on both gneiss and granite indicate that these rocks have been affected by two major events at 2.4 b.y. and 1.5-1.6 b.y. The retention of evidence of an event at 2.4 b.y. indicates that the gneiss and granite originated prior or during a geologic event of great antiquity. They have been subsequently affected by a thermal event that ended in the 1.5-1.6 b.y. time range. The quartzo-feldspathic gneiss is overlain by less metamorphosed metasedimentry rocks in the northeastern part of the mountains. Where contacts with gneiss can be observed, the metasedimentary rocks are in conformable contact with gneiss or the contact is obscured by bodies of mafic igneous rock. Top and bottom criteria in the metasedimentary rocks indicate that these rocks are younger than the gneiss, however. The metasedimentary rocks are divisible into two units, the older Deep Lake Formation which is a series of interbedded quartzites, metavolcanic (?) rocks, and conglomerates and a younger Libby Creek Group, consisting of a basal conglomerate overlain by thick sections of quartzite, metadolomite, and slate. Contacts between the two metasedimentary units are conformable, although regional relationships suggest they are separated by an unconformity. The metasedimentary rocks have been dated and preliminary results suggest that they are older than 1.6 b.y. The metasedimentary rocks were deformed to produce a general northeast trending fold pattern, gabbroic magma was introduced, following which the metasedimentary rocks and contained gabbroic intrusions were deformed again to produce northwest trending structures. The early deformation was not severe and in the central part of the mountains where evidence of the northeast fold system is best preserved the sedimentary rocks are little deformed. Primary structures are beautifully preserved and the rocks retain the rank of greenschist facies. To the northeast and southwest, however, the metasedimentary rocks are more highly deformed. The northwest trending structure is better developed and the metasedimentary rocks and contained sills of gabbro have been raised to the rank of almandine-amphibolite facies.

South of the Mullen Creek-Nash Fork shear zone, both the structure and rock types differ strikingly from that of the Precambrian rocks to the north. Older Precambrian metamorphic rocks include hornblende gneiss, quartzo-feldspathic gneiss, biotite gneiss, sillimanite-biotite gneiss, diopside gneiss, impure marble, and quartz-biotite-andesine gneiss. Hornblende gneiss is much more abundant than in the north and makes up approximately 40% of the older Precambrian rocks. The various older Precambrian gneisses are interlayered and grade into each other along and across strike, and are believed to have been a series of flows, tuffs, and graywackes with interbeds of limestone, quartzo-feldspathic sandstone, and rare beds of aluminous shale. Sills of mafic composition are as common in the gneiss as they are north of the shear zone, but they are metamorphosed to amphibolite and are so recrystallized as to make their origin debatable. The gneisses are cut by quartz diorite (older) and various large and small

bodies of gabbroic composition. Two of the gabbroic bodies form large mafic complexes; the Mullen Creek mafic complex which is found in the west, and the Lake Owen mafic complex located in the east (Pl. 1). The gneiss, quartz diorite and some of the gabbroic bodies are themselves cut by granite of two ages. The older is a well foliated granite; the younger is a massive granite known as the Sherman Granite. Pegmatites are especially abundant in the hornblende gneiss unit. Age determinations were made on quartzo-feldspathic gneiss and older granite, but the results were not conclusive. The granite yielded two apparent ages of 1.47 and 1.7 b.y. and pegmatites that cut the gneiss give ages of 1.46 to 1.56 b.y. Although ages are not conclusive, the determinations on both granite and gneiss show that evidence of the 2.4 b.y. event is not retained in these rocks. Either these rocks are in fact younger than their counterparts to the northwest, or they have been metamorphosed to the extent that their primary ages are now thoroughly masked. Age determinations on the Sherman Granite were firmly established at 1.35 ($\pm$.50) b.y., and aside from a few late felsic dikes that cut the granite this is the youngest rock of Precambrian age in the Medicine Bow Mountains.

The structural geometry of rocks south of the shear zone is extremely complex. Most folds noted in the gneisses have axes that trend east-northeast to northeast, and it is evident from studies of the geometry of the folds and lineations on the folds that they are refolded. This evidence of refolding plus the obliteration of primary structure in the dikes, and the lack of retention of the 2.4 b.y. event in older granite and gneiss of this area combine to suggest that rocks south of the shear zone have been affected by deformation and metamorphism that was not well defined north of the fault.

There is no record of geologic events in the Medicine Bow Mountains from the time of formation of the Sherman Granite (1.35 b.y.) to deposition of the Mississippian Madison Limestone. Sedimentary rocks of Paleozoic age are typical of a stable shelf environment and are not thick. Mesozoic sedimentary rocks are thicker than Paleozoic units but still typical of a stable shelf environment and thin as compared with equivalent units in western Wyoming.

In latest Cretaceous time the Medicine Bow Mountains were uplifted and from latest Cretaceous through early Eocene complex structures developed along the margins of the mountains and thick deposits of conglomerate and arkosic sandstone were laid down in basins adjacent to the mountain uplifts. In some areas there is clear evidence that the Precambrian basement was involved in this period of deformation. The crystalline basement takes the configuration of the Laramide fold. There is no evidence of control of Laramide folding by structure of the Precambrian basement. Major faults that developed during this Laramide deformation do show a clear correlation with basement structure. It is suggested that horizontal compression was important in the development of the Laramide structure.

A REGIONAL STUDY OF ROCKS OF PRECAMBRIAN AGE IN THAT PART OF THE MEDICINE BOW MOUNTAINS LYING IN SOUTHEASTERN WYOMING—WITH A CHAPTER ON THE RELATIONSHIP BETWEEN PRECAMBRIAN AND LARAMIDE STRUCTURE.

by

R. S. Houston* and others

INTRODUCTION

LOCATION AND ACCESSIBILITY

The Medicine Bow Mountains[1] are located 30 miles west of the city of Laramie in southeastern Wyoming (Fig. 1). The mountain area is a complex of Precambrian rocks that is the core of a large asymmetric anticline bounded by west dipping thrusts on the east flank. It is bordered on the east by the north-trending Laramie basin that contains sedimentary rocks ranging in age from Paleozoic to Recent and on the west by the northwest trending Saratoga Valley that principally contains sedimentary rocks of Tertiary age. On the north the mountain area is split into two anticlinal segments that are separated by a synclinal valley largely containing rocks of Paleozoic and Mesozoic age. These two segments are similar to the main body in that the anticlines have cores of Precambrian rocks, are asymmetric, and are thrust to the east (Pl. 1). The area mapped encompasses 29 seven and one-half minute topographic quadrangles or approximately 1580 square miles.

The Medicine Bow Mountains are readily accessible to automotive vehicles from June through September, but most highways and roads are closed during the rest of the year because of snow cover. The mountain area is traversed by two main highways that cross it in an east-west direction; State Highway 130 crosses the central part and State Highway 230 crosses the southern part. In addition to these highways there are two main gravel-surfaced roads; the Keystone road that crosses the Medicine Bow Mountains in an east-west direction about midway between Highways 130 and 230 and the Sand Lake Road that makes a great loop through the northern part of the mountain area and opens this area to automotive travel (Pl. 1). There are also many roads and trails that are best traveled by a pick-up truck or a four-wheel drive vehicle.

PLAN AND PURPOSE OF THE STUDY

This study is part of a larger program initiated by the Geological Survey of Wyoming in 1957 to study the rocks of Precambrian age in southeastern Wyoming. The program was started because the rocks of Precambrian age were for the most part unmapped and represented a major gap in our knowledge of the geology of Wyoming. In addition to the new scientific knowledge to be acquired a basic knowledge of the Precambrian rocks would be useful in gaining a better understanding of the structure of the sedimentary rocks of the basins, and also the mapping program might lead directly or indirectly to the discovery of new mineral deposits.

The mapping program has been done with the full cooperation of the Geology Department of the University of Wyoming. Something over fifty percent of the field work was done by the writer and the rest by graduate students of the University. The research facilities of the University, which have been used during the laboratory portion of the study, made possible a more quantitative chemical and mineralogical approach.

---

*Professor of Geology, University of Wyoming, Laramie, Wyoming.

1. The geographic term Medicine Bow Mountains includes a mountain complex that extends south into Colorado. In this report, the term is limited to that part of the Medicine Bow Mountains north of the Wyoming-Colorado state line.

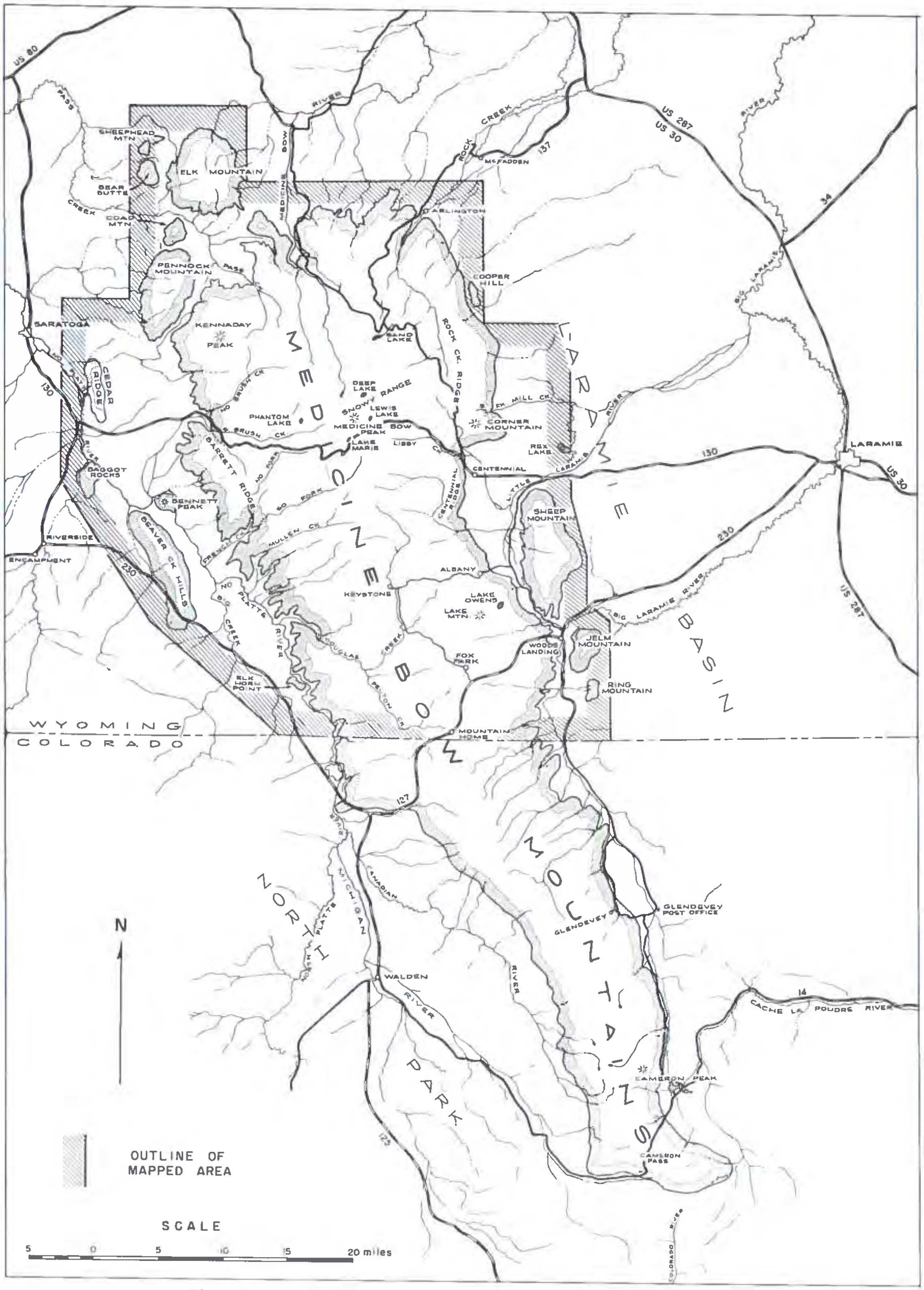


Figure I. Index map of The Medicine Bow Mountains

In the initial planning the Medicine Bow Mountains were selected as the primary area to be studied because this range contained the greatest variety of rocks of Precambrian age and the most complete section of metasedimentary rocks known in Wyoming. The post-Precambrian geologic record is well-recorded in this area and it is possible to relate Precambrian structure to structural patterns developed during all post-Precambrian geologic events.

The field work was started along the flanks of the mountains because of better exposures and less forest cover. In general, mapping along the flanks of the mountains has been done in greater detail and with a higher degree of accuracy than in the central part because the north central region of the Medicine Bow Mountains is partially covered with glacial drift and the south part is a nearly flat upland surface of low outcrop density and heavy forest cover. A rough estimate of the detail of mapping in any given area can be made by noting the density of structural observations (Pl. 1). Most of the mapping has been done on enlarged air photographs and geologic data has been transferred to seven and one-half minute topographic maps that became available during the latter stages of the program.

In this report the results of the geologic study have been summarized on a map at the scale of one inch equals one mile (Pl. 1). Detailed maps have been prepared of key areas, but it was not considered practical to present the results of the study at a scale of two and one-half inches to the mile (the scale of Geological Survey 7½ minute Quadrangle maps) for the entire mountain range. Special reports on areas of economic interest have been published (Houston, 1961; Currey, 1965; McCallum, 1968 and McCallum & Orback, 1968).

The writer spent 19 months in the field in the summers of 1957 through 1965. Malcolm McCallum and John S. King spent five months each in the field during the summers of 1960-1961 and 1961-1962 respectively. B. B. Ruehr, W. G. Myers, C. O. Orback, J. F. King, Irwin Matus, D. R. Currey, M. N. Swetnam, John Gries and M. O. Childers each spent two months in field mapping during the course of study. The areas mapped by each individual are shown in Plate 1. The final map compilations were made by the writer, however, and because certain changes and modifications were necessary the responsibility for any errors in the final maps lies with the writer.

This is a first study of a large area of exceedingly complex rocks of Precambrian age; a study that is certainly in good part of a reconnaissance nature. We have been able to examine a small percentage of available outcrops, and to the student of metamorphic rocks this is indeed frustrating, but we feel that the major outline of the geology of the Medicine Bow Mountains has been established, and we can now turn to critical areas for both scientific and economic study.

LABORATORY PROCEDURES

Laboratory procedures have primarily involved routine petrographic study of selected specimens from mapped units. Mineral identification by universal stage and/or index determinations has been used in critical specimens only. Although some 2000 thin sections have been studied, this represents a small sample of the area (Pl. 2). Percentages of minerals in most specimens of igneous rocks and many of the metamorphic rocks have been estimated by the Chayes' method (Chayes, 1956). In metasedimentary rocks estimates of mineral percentages have been made visually. A comparison of some 100 specimens, where percentages were estimated visually and by grain counts, shows that the visual estimates were not seriously in error ($\pm 10\%$) where a mineral was present in amounts greater than five percent. In order to reduce the descriptive portion of the text, tables were used to report the petrography of the rocks wherever possible and the limits of error in mineral determination are given in these tables. The number of samples studied of each rock type and the location of samples has been given so the reader can judge the significance of conclusions drawn from the petrographic study.

Chemical analyses were made of a number of samples by the Japan Analytical Chemistry Research Institute. The results of an analysis of sample W-1, made along with the unknowns, compare favorably with the mean of all analyses reported by Stevens et. al. (1960) with all values but one within one standard deviation of the mean. The value for Na_2O was slightly high but within two standard deviations of the mean. Normative calculation for these analyses were made by Dr. Kenneth Perry.

Age determinations on 36 samples by the Rb/Sr method have been made by Allan Hills of Yale University and Paul Gast and Ian Swainbank of Lamont Geological Observatory. Both whole rock and mineral ages have been determined. The analytical procedures are described in Hills, Gast, Houston, and Swainbank, 1968.

PREVIOUS WORK

The first formal geologic study in this region was made as part of the early territorial surveys by the United States government. An excellent summary of the early work done prior to 1900 has been made by Van Hise and Leith (1909, p. 833-849). According to Van Hise, the first report that mentioned the geology of this general area was written by Captain Howard Stansbury in 1853 who states that the Laramie Mountains located east of the Medicine Bow Mountains have an extensive formation of massive red granite. Hayden included brief descriptions of the rocks in the Laramie and Medicine Bow Mountains in his reports of 1863, 1868, and 1872. The most detailed description was that of Hague (1877), who gave a relatively complete description of the metasedimentary rocks located in the north-central Medicine Bow Mountains. In 1889 Van Hise made a three day reconnaissance study of the metasedimentary rocks of the Medicine Bow Mountains and in his report of 1909, describes the metasedimentary rock sequence and states that these rocks may be folded into a syncline with its axis east of Medicine Bow Peak and an anticline with its axis west of the peak.

These early investigations were strictly reconnaissance studies and although the various rock types were often well-located with regard to cultural or physical features, none of the reports contained maps or sketches that could be used in map compilation. At the turn of the century, from 1880 to 1910, there was considerable mining activity in the Medicine Bow Mountains. This mining activity stimulated geologic investigations and there is a modest literature by the Wyoming State and Territorial Geologists describing the geology of the regions immediately adjacent to mine properties. Typical of these reports are those by the Territorial Geologist, Ricketts (1888, 1890) and by Wyoming State Geologists Beeler (1904, 1905, 1906, and 1908) and Jamison (1911) who discuss the history of certain mining operations and, in general, present optimistic views of their future potential. During the same period four areas of Precambrian rocks were mapped by geologists of the U. S. Geological Survey. The areas included the Encampment copper district in the Sierra Madre, the Esterbrook mining district in the northeastern Laramie Mountains, the Hartville district, and the southern Laramie Mountains. No mapping was done in the Medicine Bow Mountains but this work gives a general idea of the geology of the Precambrian rocks in the surrounding area. By 1910, the mining boom in southeastern Wyoming was over, and with the decline of the mining industry in the Precambrian areas, interest in the geology also ceased. With one exception, no significant geologic mapping was done in the Medicine Bow Mountains from 1910 until the early 1950's. The major exception was the work done by Eliot Blackwelder (1926) in the central Medicine Bow Mountains. Blackwelder mapped the metasedimentary rocks of the north-central part of this range and made a detailed study of the stratigraphy of these units.

During World War II interest in the mineral resources of southeastern Wyoming was revived and a number of geological investigations were made by the Geological Survey of Wyoming, the United States Geological Survey, and the United States Bureau of Mines. Reports that included geologic maps of the Precambrian rocks were those of A. F. Hagner and W. H. Newhouse. A report on cordierite deposits by Newhouse and Hagner (1949) included a detailed map of an area in the southcentral Laramie Mountains and a report by Hagner (1951) contained an outline map of the anorthosite areas of the central Laramie Mountains. During the period 1944 to 1951, W. H. Newhouse and A. F. Hagner mapped a large area in the central Laramie Mountains of approximately seventeen townships, as a project of the Wyoming Geological Survey and the United States Geological Survey. The geologic map was published in 1957 (Newhouse and Hagner, 1957) and is a geologic and structural map of the large anorthosite masses and related metasedimentary and igneous rocks of the southern Laramie Mountains. In the Medicine Bow Mountains examinations were made of pegmatites in the southern part of the mountains, (Hanley, Heinrich, and Page, 1950, p. 107-108), and a diamond drilling program was undertaken at the Rambler Copper Mine in the south central part of the mountains, (Kasteler and Frey, 1949), but no geologic mapping was done during this period.

During the early 1950's graduate students of the University of Wyoming made substantial contributions to the geology of the Precambrian by mapping areas in the northern Laramie Mountains and the Medicine Bow Mountains. In the Medicine Bow Mountains areas have been mapped by Michalek (1952), Schoen (1953), and Catanzaro (1957). The information gained from these studies has been very useful, but the areas mapped were small, and the geology of the rocks of Precambrian age in the Medicine Bow Mountains remained essentially unknown at the start of the mapping program of the Geological Survey of Wyoming in 1957.

In striking contrast to the general lack of interest in the Precambrian rocks of the Medicine Bow Mountains the rocks of Paleozoic, Mesozoic, and Tertiary age that border the mountains received considerable attention from geologists. The most notable contributions were made by R. H. Beckwith (1938, 1941, 1942), who mapped Paleozoic, Mesozoic, and Tertiary rocks in the Elk Mountain area and along the southeastern margin of the Medicine Bow Mountains, and by S. H. Knight (1953), who studied upper Cretaceous and Tertiary rocks of the Centennial area. Other contributions to the geology of post-Precambrian rocks will be discussed in chapters on the stratigraphy of these rocks.

ACKNOWLEDGEMENTS

The writers are greatly indebted to the late H. D. Thomas, State Geologist of Wyoming (1941-1967), who gave us invaluable support throughout the period of study. S. H. Knight and D. L. Blackstone, Jr., in addition to advice and editorial assistance allowed free use of the facilities of the Department of Geology, University of Wyoming. Other members of the Geology department who contributed time and assistance were P. O. McGrew who reviewed aspects of the study of rocks of Tertiary age, Brainerd Mears, who gave advice on the deposits of Pleistocene age, R. B. Parker, who assisted in the structural analysis, and D. W. Boyd who edited chapters on Paleozoic and Mesozoic stratigraphy.

The writers are especially indebted to D. L. Blackstone, Jr. and S. B. Smithson, Department of Geology, who undertook the most difficult task of editing the entire manuscript, and W. H. Wilson, Assistant State Geologist, who reviewed the manuscript and handled final preparations for publication.

E. J. Castro assisted in field work in the northcentral Medicine Bow Mountains in 1961, and Peter Davidson assisted in field work in the southcentral Medicine Bow Mountains in 1962. Mrs. Shirley Jereb and John Galey drafted parts of the maps and text figures.

Many aspects of this study result from long continued work by members of the staff of the Geology department as part of summer field studies at the University of Wyoming Science Camp. Areas which have been studied by various staff members, especially D. L. Blackstone, Jr., S. H. Knight, and R. B. Parker, include the Big Creek area, south end of Coad Mountain, and the area southwest of Elk Mountain. We do not all agree on the interpretation of these areas, however, so that the geology shown on Plate one is the writer's responsibility. The pioneering work on the Precambrian metasedimentary rocks by Eliot Blackwelder has been the key to our study of these units and the extraordinarily meticulous mapping of the Paleozoic-Mesozoic rocks in the Elk Mountain area and along the southeast border of the Mountains by R. H. Beckwith has been of tremendous value in this study. Although some changes have been made in the geology in areas mapped by Beckwith and Blackwelder these have been almost entirely changes in interpretation.

Palynological age determinations on rocks of the Lewis shale, Medicine Bow formations, and Ferris-Hanna formations were made by E. R. Woodside, consultant paleontologist, and Robert H. Tschudy, of the U. S. Geological Survey. We wish to thank Tschundy and the U. S. Geological Survey for allowing us to report these palynological ages.

Many courtesies were extended to the writers by members of the U. S. Forest Service during the mapping of areas within the National Forest, and by ranchers who allowed access to private property along the borders of the forest. We are

especially grateful to R. H. Platt, Sr., and R. H. Platt, Jr., of Encampment, Wyoming.

Rarely is a geologic report the work of a single individual or a group of individuals since we depend, in a great measure, on previous studies to facilitate mapping. This is especially true of a study of a large area such as the Medicine Bow Mountains, and the writers would like to again note earlier workers who are credited in the chapter on previous work and on plate one.

GEOGRAPHY

PHYSICAL GEOGRAPHY

Southeastern Wyoming is an area of broad anticlinal uplifts separated by basins in which the topography is clearly controlled by geologic structure. The mountain areas are anticlinal structures in which rocks of Precambrian age are exposed, and the basin areas are synclines underlain by less-resistant post-Precambrian sedimentary rocks. In the vicinity of the Medicine Bow Mountains elevations range from 7000 feet in the Laramie Basin to 12,000 feet at the crest of the mountains. Striking changes in relief occur at the mountain front where a change in elevation of 2,000 feet in a distance of less than two miles is typical. The mountains proper are marked by upland surfaces of low relief broken by occasional axial remnants such as the Snowy Range in the north-central mountains. This is six miles long and rises to an elevation of 12,000 feet, approximately 1000 feet above the surrounding surface. The upland or high-level erosion surface is best developed south of State Highway 130 where it is referred to as Libby Flats. The Libby Flats surface extends over most of the central part of the mountain range and extends south beyond the Wyoming-Colorado State line. This surface is covered locally by patches of Tertiary sedimentary rocks and Quaternary lag gravels.

The northern Medicine Bow Mountains have been extensively modified by Pleistocene glaciation (Pl. 6). The glacial ice originated in the vicinity of the Snowy Range remnant and the most prominent glaciers moved northward into the area now drained by the Medicine Bow River, eastward into the area drained by Libby Creek and the North Fork of the Little Laramie River, and westward into the area now drained by Brush Creek. Thin deposits of ground moraine are present in the area around the Snowy Range remnant and much thicker morainal deposits occur along the northern, northeastern, and northwestern margin of the range. The most spectacular deposits of glacial drift are the great arcuate terminal moraines on the north slope of the Medicine Bow Mountains and the outwash aprons in this same area which are chacterized by hundreds of small ponds and lakes.

Four north-flowing rivers drain the mountain area. The most important is the North Platte River that heads in Colorado and flows northeastward along the west slope of the mountains. Along the southwest slope of the mountains, the North Platte has cut a deep canyon into the Precambrian rocks instead of following the Saratoga Valley a few miles to the west even though this is partially filled with sedimentary rocks of Miocene and Pliocene age. Knight (1953, p. 75) considers this an example of superposition, perhaps of some master drainage established on a surface of late Miocene sedimentary rocks. Three major tributaries of the North Platte River head in the Medicine Bow Mountains. The most important is Douglas Creek, which heads in the south central part of the mountains and flows south, then southwest, and then northwest before entering the Platte (Pl. 1). The peculiar disposition of this stream is controlled at least in part by shear zones and by lithologic boundaries of the rocks of Precambrian age. The two other tributaries of the North Platte are French Creek and Brush Creek that drain the west-central and northwestern mountains respectively. Other north-trending rivers are the Medicine Bow River that drains the northern glaciated area; the Little Laramie River that heads in the southeastern part of the mountains and flows northward between the main mountain mass and Sheep Mountain, and the Big Laramie River that heads in Colorado and flows along the southeastern margin of the mountains where it is superimposed on the rocks of Precambrian age in much the same way as is the North Platte River.

HUMAN GEOGRAPHY

The Medicine Bow Mountains are within the Medicine Bow National Forest. For this reason there is relatively little private land within the

Plate 6—View to the northeast of glaciated southeast face of Snowy Range axial remnant. Protalus ramparts are in the center of the photograph. Lake Marie is right foreground.

area and no permanent settlement with the exception of the small timber village of Fox Park. Timbering is the major industry and is carried on largely during the summer months, but operations may continue throughout the year with the exception of April, May, and part of June during the spring thaw. The forest is also used for watershed control, grazing, recreation, game refuge and control, and winter sports. Recreation is gradually becoming the most important single use of the area and trout fishing heads the list. Trout fishermen will recognize Douglas Creek, French Creek, and the North Platte River as among the most celebrated trout fishing streams of North America.

FAUNA AND FLORA

Within a distance of approximately 30 miles from the center of the Laramie Basin to the crest of the Medicine Bow Mountains there are striking changes in ecology. For example, in the area northeast of Arlington one may find cacti and horned toads whereas at the top of the mountains in the vicinity of Medicine Bow Peak there are arctic tundra and ptarmigan. These changes in flora and fauna result from the 5,000 foot difference in elevation and the variation in rainfall between the basin area and mountains. The Laramie Basin is a semi-arid grass land having an average rainfall of 11 inches. With the exception of cottonwood and alder close to the streams and rivers, few trees are present. The mountains on the other hand receive 20 inches of annual rainfall and as a consequence, are heavily forested with spruce, pine, and fir, with the exception of the area above timberline near the Snowy Range. Heavy stands of aspen are present along the flanks of the mountains especially in moist areas along stream courses and near springs. There does not appear to be a general relationship between rock type and vegetation, but some of the best stands of timber are found in areas underlain by sedimentary rocks of Tertiary age, or glacial drift.

The fauna of the Medicine Bows is typical of the Rocky Mountains. Large animals include elk, mule deer, black bear and mountain lion, and typical small mammals are red squirrel, coyote, marten, grey fox, golden marmot, and beaver. Beaver are especially abundant in the southern mountains on the high level erosion surface of Libby Flats. Hundreds of beaver dams occur along the streams of this area, and almost all of these contain an abundance of small brook trout.

GENERAL GEOLOGY

Few areas in the United States contain as complete a geologic record as the Medicine Bow Mountains. Precambrian rocks range in age from older than 2.4 b.y. to 1.3 b.y. Paleozoic and Mesozoic sedimentary rocks include units of Mississippian age to late Cretaceous, and Tertiary rocks are present that are representative of every epoch. Locally all of these units are covered by continental Quaternary deposits including extensive beds of glacial drift.

In this report emphasis will be placed on descriptions of rocks of Precambrian age, but brief summations of younger rocks will be given to clarify the discussion of Precambrian-Laramide structural relationships.

The rocks of Precambrian age are in the uplifted Medicine Bow Mountains proper, and in uplifted blocks along the northwest and southeast flanks of the mountains (Pl. 1). The Precambrian rocks are divided into two major units by the Mullen Creek-Nash Fork shear zone that trends in a northeasterly direction throughout the central part of the mountains (Pl. 1). North of the fault older Precambrian gneisses and igneous rocks are exposed to the northwest. These gneisses are overlain by metasedimentary rocks that crop out in the central and northeastern part of the mountains. South of the fault a complex sequence of gneisses and igneous rocks is exposed but none of the units in the south can be correlated with those of the north. For this reason, the two major rock sequences (i.e. north and south of the fault) will be discussed separately beginning with the older gneisses north of the shear zone.

PRECAMBRIAN ROCKS NORTH OF MULLEN CREEK-NASH FORK SHEAR ZONE

QUARTZO-FELDSPATHIC GNEISS

The quartzo-feldspathic gneiss on the northwest slope of the Medicine Bow Mountains is a composite unit that is only partly subdivided on Plate 1. This unit includes biotite gneiss, augen gneiss, hornblende gneiss, quartzite, cataclasite, and poorly-foliated quartzo-feldspathic gneiss. It crops out in the area north of the Mullen Creek-Nash Fork shear zone and west of the metasedimentary rocks, and is present at Elk, Coad, and Pennock Mountains (Pl. 1).

The rock types of the quartzo-feldspathic gneiss are regarded as older Precambrian since they dip beneath the metasedimentary rocks of the Heart Formation in Tps. 14 and 15 N., R. 81 W. (Pl. 1), and beds of the Deep Lake Formation in T. 16 N., R. 81 W. Whole rock Rb-Sr isotope compositions of three samples of biotite gneiss from this unit fit a 2410 $\pm$50 m.y. isochron (Hills, Gast, Houston, and Swainbank, 1968) and therefore the biotite gneiss is older than 2.4 billion years. All of these rocks types are interlayered with one another and are deformed as a unit.

HORNBLENDE GNEISS

Hornblende gneiss is interlayered with other rock types, and lenses of the rock are found in biotite gneiss and poorly-foliated quartzo-feldspathic gneiss throughout the area. The best exposures of this rock are on Pennock Mountain where it is interlayered with quartzo-feldspathic gneiss and makes up 40% to 50% of the southwest slope of the mountain. Elsewhere it is poorly exposed because it is less resistant to erosion than other rock types. Well developed planar structure reflects compositional layering and mineral orientation. Amphibole-rich layers alternate with quartz-feldspar-rich layers, and the amphibole may show a strong preferred orientation parallel to the compositional layering. The mineralogy of the rock is given in Table 1. The origin of this rock is not known, but mineralogical evidence indicates a sedimentary parent. The abundance of quartz suggests derivation from a sedimentary rock (Heinrich, 1956, p. 255-256, Williams, Turner, and Gilbert, 1954, p. 241-243).

The texture of this rock (Pl. 7) as well as that of the biotite gneiss and poorly-foliated quartzo-feldspathic gneiss may be interpreted in several different ways. The general paragenesis of minerals is somewhat like that for minerals in gneisses of the Tenmile Range, Colorado, Koschmann (1960, p. 1357-1370). Some quartz may be early, followed by a second generation of quartz, plagioclase, and microcline. The small well-rounded quartz grains may have been large grains in the original sediment and the plagioclase, second generation quartz, and microcline recrystallized clay-sized matrix. Koschmann (1960, p. 1365) believed that spherical quartz grains occurring as inclusions in plagioclase were remnants of early quartz that was resorbed, but he believed the early quartz was sedimentary. Because these gneisses are probably polymetamorphic it seems unlikely that any feature of an original sedimentary rock would survive. The spherical quartz may be a first generation metamorphic quartz that has been resorbed or it could be a survival of quartz rounded during cataclasis.

Table 1—Mineralogy of hornblende gneiss
(abundance in volume percent, visual estimates, P=2% or less)

Mineral	A	$\overline{X}$	Comments
Plagioclase	5/5	49	Altered to sericite and clay minerals
Quartz	5/5	30	Well-rounded to subrounded small grains, large anhedral grains
Amphibole	5/5	15	Blue-green
Epidote, Sphene, Apatite	5/5	P	

A = Number of samples with mineral/total number of samples studied
$\overline{X}$ = arithmetic mean

QUARTZITE

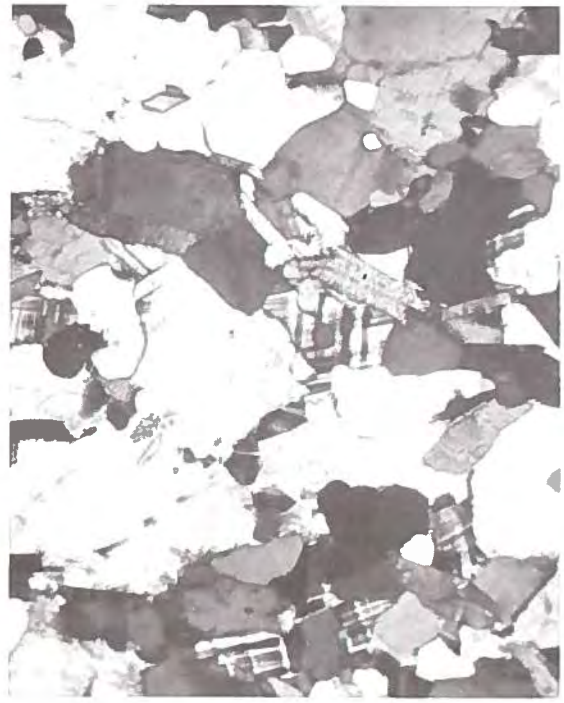
Quartzite is not abundant. An occasional thin quartz-rich layer is interbedded with the biotite gneiss in the Baggot Rocks area, and one bed of quartzite approximately 50 feet thick and 3,800 feet long is interlayered with hornblende gneiss and quartzo-feldspathic gneiss on the southwest slope of Pennock Mountain.[2] The quartzite is regarded as sedimentary although no primary structures have been noted.

BIOTITE GNEISS

The best exposures of biotite gneiss are on Cedar Ridge and on the east side of the Platte River near the confluence of the Platte and En-

2. The layer of quartzite on Pennock Mountain, secs. 9 and 16, T. 17 N., R. 82 W., is shown in purple on plate one, but is designated by the symbol Q.

A
Hornblende gneiss from Pennock Mountain showing small well-rounded quartz grains and large anhedral grains of quartz.



B
Biotite gneiss from Cedar Ridge with texture similar to hornblende gneiss. Note well-rounded quartz grains and interstitial habit of microcline.

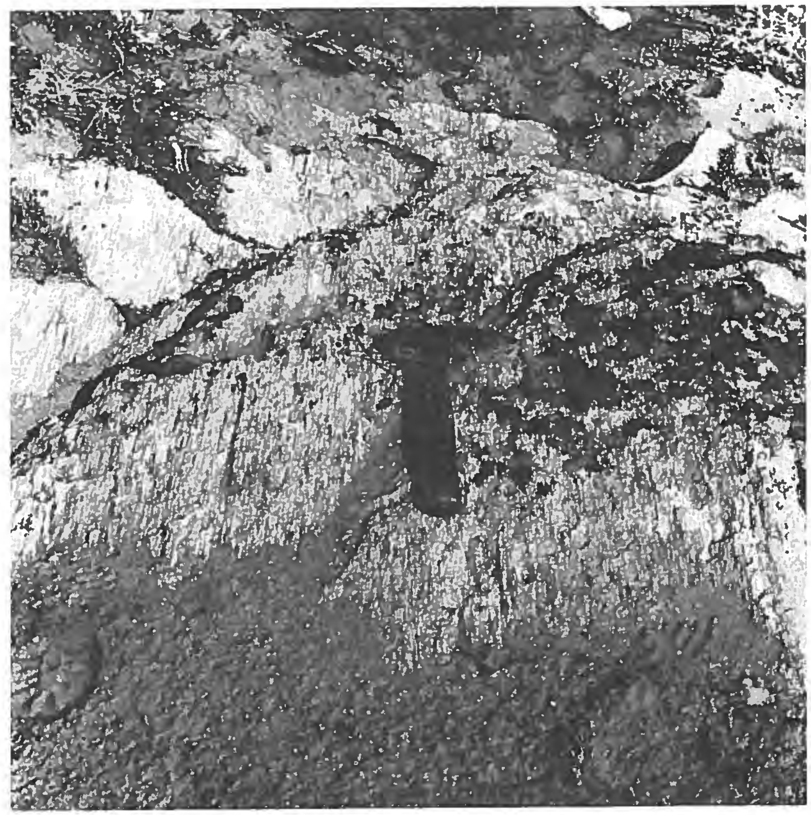
C
Quartzo-feldspathic gneiss from Baggot Rocks showing dual habit of quartz.

D
More coarsely textured quartzo-feldspathic gneiss from Cedar Ridge showing interstitial habit of microcline. Line 1 mm. long, all photos same scale.

Plate 7—Photomicrographs of hornblende gneiss, biotite gneiss, and quartzo-feldspathic gneiss from north of the Mullen Creek-Nash Fork shear zone.

A

View of typical outcrop of quartzo-feldspathic gneiss taken in area south of Beaver Creek, sec. 3, T. 15 N., R. 82 W. Lineation in center is trace of axis of minor fold.



B

Well developed lineation (biotite streaks) on plane of foliation of quartzo-feldspathic gneiss. Sec. 10, T. 15 N., R. 82 W.

Plate 8—Quartzo-feldspathic gneiss from the northwestern part of the Medicine Bow Mountains.

campment Rivers (Pl. 1). It is interlayered with other units throughout the area, but is regarded as a mappable unit only in the above localities. It is, in fact, very similar to the poorly foliated quartzo-feldspathic gneiss in the field and distinctions between the two types of rock are subjective. Outcrops consisting chiefly of biotite gneiss have a distinct slabby appearance similar to well-bedded, quartz-rich sedimentary rocks. Strong foliation in the rock is developed by biotite-rich composition layers alternating with biotite-poor layers and by a strong preferred orientation of biotite which is parallel to the composition layers.

The mineralogy of the biotite gneiss is given in Table 2. The habit of two of the minerals is of particular interest. As was true of the hornblende gneiss there are two types of quartz in the biotite gneiss; small well-rounded grains occurring as inclusions in other minerals and scattered throughout the rock and large anhedral grains similar in habit to plagioclase (Pl. 7). Microcline is commonly interstitial in irregular networks between plagioclase and quartz and some larger crystals show irregular contacts with plagioclase and contain inclusions of plagioclase suggesting replacement of plagioclase by microcline. (Pl. 7, fig. 2).

Table 2—Mineralogy of biotite gneiss
(abundance in volume percent, visual estimates)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	4/4	28	(20-40)	In two sizes, well rounded to subrounded small grains, large anhedral grains
Plagioclase	4/4	27	(17-33)	Altered to sericite and clay minerals
Microcline	4/4	27	(15-35)	In interstitial networks and as large discrete grains
Biotite	4/4	13	(7-17)	
Epidote	4/4	P		Some grains have cores of altered allanite
Apatite	3/4	P		In euhedral grains
Zircon	2/4	P		Small euhedral grains and small well rounded grains
Muscovite, sphene opaque minerals, Penninite		P		

POORLY-FOLIATED, QUARTZO-FELDSPATHIC GNEISS

This type of gneiss is most widespread and is the major rock in all areas of quartzo-feldspathic gneiss except those specifically referred to in descriptions of other units. The rock is not distinctly layered but has a poor foliation resulting from alignment of biotite and muscovite and a faint layering caused by concentration of biotite in poorly developed layers. It may show a well defined lineation (Pl. 8). This texture, plus a higher percentage of quartz, probably account for the greater resistance to erosion compared with hornblende gneiss and biotite gneiss. This differential resistance to erosion may result in an overestimation of the abundance of this rock especially in areas of poor exposures.

The mineralogy of this rock type is in Table three. Quartz and microline have the same habit as in the hornblende gneiss and biotite gneiss (Pl. 7, fig. 2). Estimates of mineral composition indicate higher quartz content, but less microcline. Whatever the original nature of the hornblende gneiss, biotite gneiss, and poorly- foliated quartzo-feldspathic gneiss these units have a strong textural similarity suggesting a common history and possibly are sedimentary rocks variously transformed by metamorphism.

Table 3—Mineralogy of poorly foliated quartzo-feldspathic gneiss
(abundance in volume percent, visual estimates)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	12/12	47	(33-67)	In many but not all samples two sizes, as in hornblende and biotite gneiss
Plagioclase	12/12	19	(10-30)	An range 15-31, average $An_{22} \pm 5\%$
Microcline	9/12	18	(0-40)	
Biotite	10/12	5	(0-10)	
Muscovite	9/12	4	(0-12)	
Epidote	10/12	4	(0-25)	
Amphibole	3/12	P		Blue-green. Some grains partly replaced by epidote
Sphene, opaque minerals, apatite, zircon, garnet		P		

AUGEN GNEISS

Augen gneiss is interlayered with the above units and is the chief rock type on Barrett Ridge. The gneiss is distinguished by large lenticular crystals of K-feldspar that are usually pink in color. The augen are not entirely potash feldspar but include large crystals of plagioclase as well as flaser of plagioclase, microcline, and quartz. The gneiss may have been a porphyritic igneous rock or an augen gneiss originating through metasomatism, but part of the unit must have been mobile at some stage in its development since stringers of the augen gneiss cut biotite gneiss locally.

The mineralogy of the augen gneiss is shown in Table 4. The rock varies in composition from a granite gneiss to a quartz monzonite gneiss.

CATACLASITE

Throughout this group of rocks there are individual layers with thickness measured in tenths of inches, and feet that are cataclastic. Where this rock type is mappable, it is indicated by short dashes on Plate one. Most of the cataclasite is conformable in structure to the foliation

Fig. 2-A

Fig. 2-B

Figure 2—Photomicrographs of quartzo-feldspathic gneiss showing irregular (replacement?) contacts between microcline and plagioclase (white). Samples from Cedar Ridge. Line ½ mm. long.

Table 4—Mineralogy of augen gneiss
(abundance in volume percent, visual estimates)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	9/9	46	(30-70)	As anhedral grains, and in veinlets parallel to foliation
Microcline	8/9	26	(0-50)	In augen, and interstially in finer grained fraction
Plagioclase	9/9	16	(10-30)	An range 21-33, Av.An 25 (±5%) in augen and as smaller grains in fine-grained fraction
Biotite	9/9	6	(2-10)	
Epidote	9/9	3	(1-5)	Some grains have cores of allanite
Muscovite	9/9	3	(1-5)	
Zircon	3/9	P		Euhedral grains, some metamict
Sphene, apatite garnet, opaque minerals, chlorite, allanite, zoisite	—	P	—	

of the quartzo-feldspathic gneiss, but there are wide zones of cataclasite in faults and this rock type will be discussed later.

METASEDIMENTARY ROCKS

Metasedimentary rocks of the central Medicine Bow Mountains were first described by Eliot Blackwelder (1926). Blackwelder studied these rocks in the area of best exposure and measured sections for most of the units he described. His units were used as the basis for mapping and for the most part his terminology is retained in this study (Table 5). Certain minor changes in terminology which will be discussed below, were made primarily to facilitate mapping.

Table 5—Terminology for metasedimentary rocks

Blackwelder (1926)	This Report	
French Slate	French Slate	Libby Creek Group
Towner Greenstone	Towner Greenstone	
Ranger Marble Anderson Phyllite Nash Marble	Nash Fork Formation	
Sugarloaf Metaquartzite	Sugarloaf Quartzite	
Lookout Schist	Lookout Schist	
Upper Medicine Peak Metaquartzite Lower Medicine Peak Metaquartzite	Medicine Peak Quartzite	
Heart Metagraywacke	Heart Formation	
Headquarters Schist	Headquarters Schist	
Deep Lake Metaquartzite	Deep Lake Formation	

CHANGES IN TERMINOLOGY

Deep Lake Formation

The Deep Lake Metaquartzite of Blackwelder was the oldest metasedimentary unit he described, and was named for Deep Lake located in secs. 32 and 33, T. 17 N., R 79 W. Blackwelder studied this quartzite in the Gold Hill area, T. 16 N., R. 80

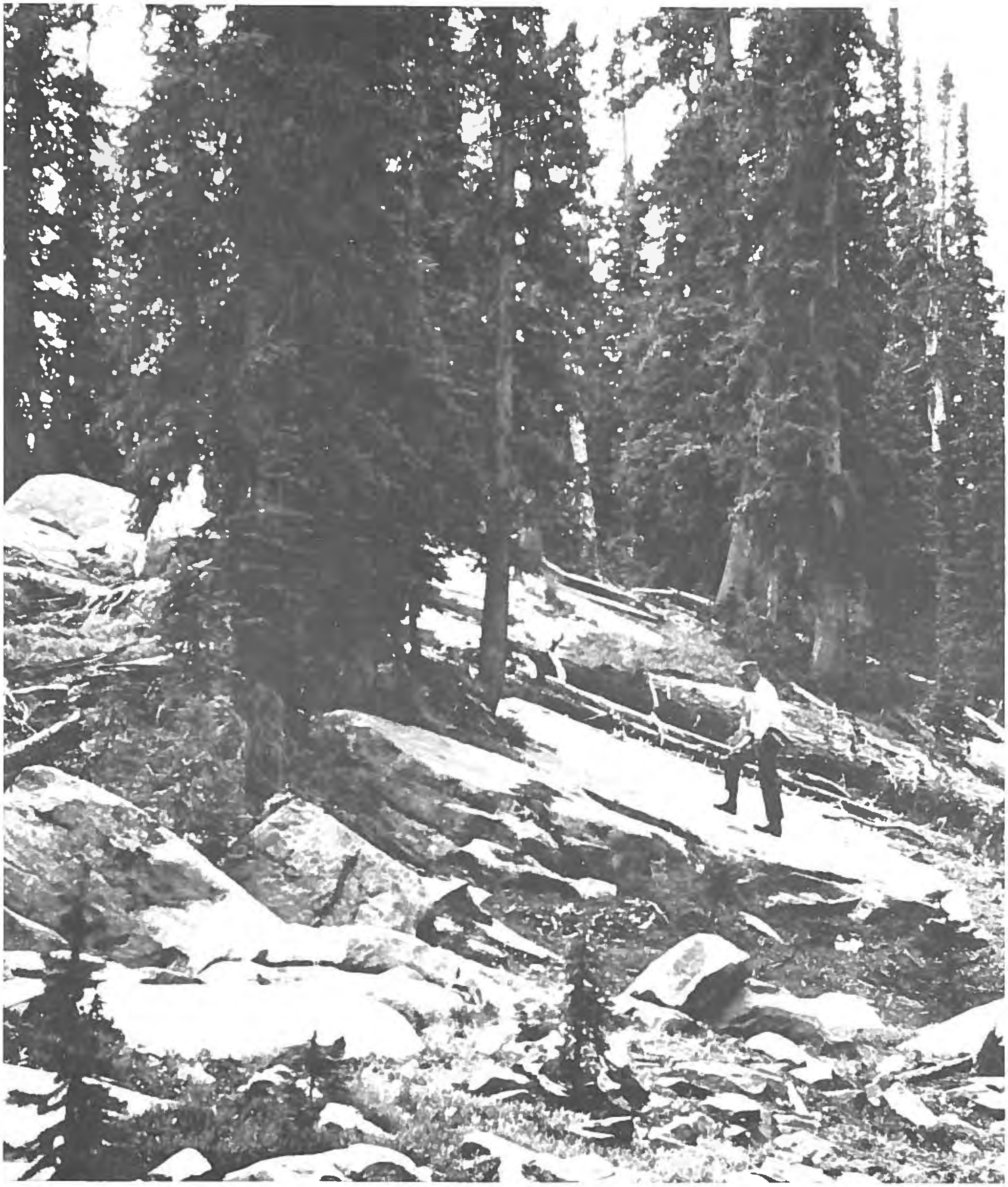


Plate 9—Outcrops of quartzite of the Deep Lake Formation in area north of Trail Creek, sec. 18, T. 17 N., R. 79 W. Student standing on bedding plane surface.

W., and recognized that it probably extended as far north as Windy Hill, sec. 18, T. 17 N., R. 79 W. He measured a section 995 feet thick on the northwest side of the South Fork of Rock Creek composed primarily of quartzite and metaconglomerate which he regarded as typical of the unit. Blackwelder recognized that the unit was thicker than 995 feet and he estimated a total thickness in excess of 2000 feet.

The Deep Lake Metaquartzite is very poorly exposed in the central Medicine Bow Mountains where it is covered by glacial deposits. To the northeast it is better exposed but is more highly deformed and metamorphosed, cut by numerous mafic igneous bodies, and top and bottom criteria are rare. In the course of this study no well-exposed section of the Deep Lake Metaquartzite could be found although a poorly exposed section was measured in the vicinity of Gold Hill that exceeded 10,000 feet in thickness.

This unit was studied by mapping individual exposures and using top and bottom criteria to determine structure. The rocks mapped consisted of quartzite, metaconglomerate, chlorite schist, amygdular metabasalt, metatuff (?), and siliceous marble. On plate one the unit is subdivided into quartzite, siliceous marble, metaconglomerate, and various metamorphic rocks classed as metavolcanic (?). In view of these facts it is here proposed to modify the nomenclature of Blackwelder and use the term Deep Lake Formation for this complex group of rocks.

Libby Creek Group

The Deep Lake Formation is overlain by a group of metasedimentary rocks herein designated as the Libby Creek Group. The name Libby Creek Group is chosen for these rocks because Libby Creek heads in the center of the outcrop area. The Libby Creek Group includes all metasedimentary rocks younger than the Deep Lake Formation and includes all of Blackwelder's units from the Headquarters Schist (oldest) to the French Slate (youngest), Table 5.

Some minor changes in terminology for the rocks of the Libby Creek Group are shown in Table 5.

1. The term meta is dropped, to conform to modern useage, where it applies to quartzite.

2. The Heart Metagraywacke is changed to Heart Formation because this is a complex unit containing rocks types in addition to graywacke.

3. The Medicine Peak Quartzite is mapped as a unit because the marker beds used to separate upper and lower parts could not be traced to the southwest.

Nash Fork Formation

A major change in terminology for rocks of the Libby Creek Group is to group Blackwelder's three-fold unit, Nash Marble (oldest), Anderson Phyllite, and Ranger Marble (youngest) into one formation. It is here proposed that these units be called the Nash Fork Formation for Nash Fork Creek that heads in Brooklyn Lake near the center of the outcrop area of this formation. The change in Blackwelder's nomenclature is suggested because the Anderson Phyllite is a lenticular unit that cannot be traced beyond the area of the type sections. Phyllite similar to the Anderson Phyllite is at different stratigrahpic levels in the Nash Fork Formation in other areas where this unit crops out (Pl. 1), and without the phyllite as a marker horizon it is difficult to distinguish the Nash from the Ranger Marble. Thus for mapping purposes these units are called the Nash Fork Formation, and where possible marble and phyllite are shown as separate lithologies on the geologic map (Pl. 1).[3]

DEEP LAKE FORMATION

The rocks of the Deep Lake Formation crop out in the central Medicine Bow Mountains roughly from a locality on North French Creek, sec. 31, T. 16 N., R. 80 W. northeast to Arlington, sec. 30, T. 19 N., R. 78 W. Quartzite of this formation is in contact with the Headquarters Schist along the southeast margin of outcrop. Isolated beds of quartzite that are considered part of this formation are as far west as Middle Cedar Creek, sec. 25, T. 17 N., R. 82 W. (Pl. 1). Near Arlington quartzite of the Deep Lake Formation is in contact with hornblende gneiss and gneissic granite. The gneiss and granite at Arlington is probably basement gneiss and cross-bedding in the quartzite at this locality shows that the quartzite is younger than the gneiss. In other areas the metasedimentary rocks of the Deep Lake Formation are sepa-

3. On the geologic map, Plate one, two of the terms used in this report are labeled improperly. Nash Formation for Nash Fork Formation and Libby Group for Libby Creek Group.

rated from basement gneiss by metamorphosed mafic igneous rocks.

The metasedimentary rocks of the Deep Lake Formation do not crop out to the southwest where quartzite of the Heart Formation of the Libby Creek Group lies on older basement gneiss. This suggests that units of the Libby Creek Group, with the possible exception of the Headquarters Schist, lie unconformably on rocks of the Deep Lake Formation. Thus rocks of the Deep Lake Formation are older than the basement gneiss and younger than rocks of Libby Creek Group.

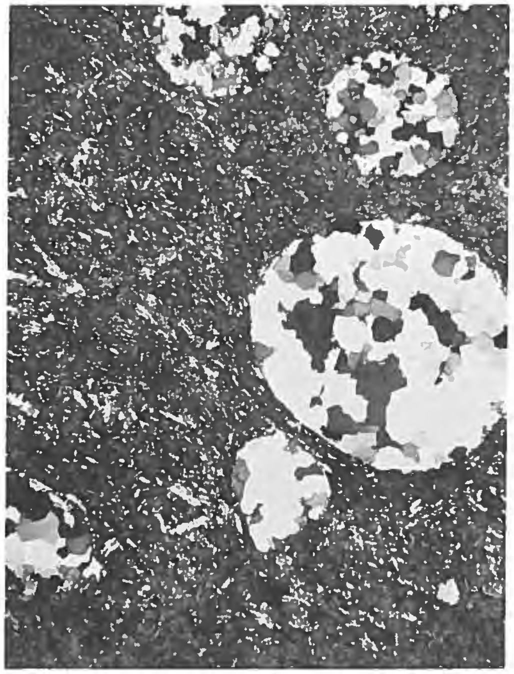
The difficulty in establishing positive age relationships especially with the gneissic rocks is probably the result of late deformation and metamorphism that has affected both units causing homogenization at contacts and also late emplacement of basaltic magma along contacts. This is best shown in T. 16 N., R. 81 W. where quartzite of the Deep Lake Formation is separated from the gneiss by orthoamphibolite. The quartzite strikes northeast and dips southeast with top to the southeast. The strike of foliation in the amphibolite (metamorphosed basalt) is parallel to the strike of bedding in the quartzite and the strike of foliation in the gneiss (basement?) is also parallel to the strike of bedding. However, the strike of foliation in the quartz monzonite gneiss changes to the northwest, away from the contact area (Pl. 1). The original quartz monzonite gneiss-quartzite contact may have been an angular unconformity.

The best exposures of the Deep Lake Formation are in T. 16 N., R. 80 W., northeast of the ghost town of Gold Hill (Pl. 1). Near Sand Lake the metasedimentary rocks are largely covered with glacial drift, and to the northeast the units are more deformed and are cut by many large bodies of mafic igneous rock. The increase in deformation to the north and accompanying increasing rank of metamorphism has largely destroyed primary structures of the rocks so top and bottom criteria are not common enough to allow satisfactory stratigraphic study. For this reason the stratigraphy of the formation is known only in the upper 10,600 feet from poorly exposed sections measured in the vicinity of Gold Hill. The lower measured section is in the crest of a plunging anticline located in the Phantom Lake area north and west of Gold Hill, T. 16 N., R. 80 W. The base of the section is at the confluence of Arrastre Creek and South Brush Creek where the most abundant rocks are quartzite, sericitic quartzite, and metabasalt (Table 6). These beds are overlain by metabasalt and fine-grained sericitic quartzite with relatively thin beds of conglomerate and thick beds of chlorite schist in the upper part.

The lower 400 feet of the measured section may be complicated by faults and/or unconformities. The lowermost bed of quartzite of unknown thickness is separated from interbedded sericitic quartzites, metabasalt, and amphibolite by an amphibolite body. This lowermost bed of quartzite is similar in lithology to quartzites in the upper part of the Deep Lake Formation and therefore may in fault contact with the interbedded sericitic quartzite, metabasalt, and amphibolite exposed at the confluence of Arrastre Creek and South Brush Creek. The body of amphibolite that separates these units may be altered igneous rock that was emplaced along a fault. East of the confluence of Arrastre and South Brush Creek a small body of sheared gneiss is exposed that could have been brought up along a fault, but this rock could also be a glacial erratic. In any event the extremely poor exposures in this area make various interpretations possible so the measured section is at best an approximation of lithology and thickness of the various units.

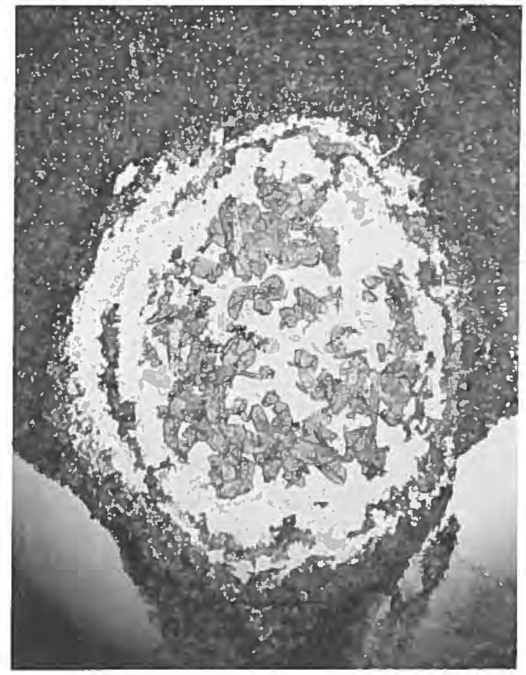
Two sections have been measured in units overlying this Phantom Lake section. One was measured north of Gold Hill in secs 10 and 11, T. 16 N., R. 80 W., and another was measured by Blackwelder (1926, p. 626-627) along the South Fork of Rock Creek. These upper beds are largely coarse-grained quartzite showing large scale cross-bedding interlayered with 2 to 3 foot beds of conglomerate containing well rounded pebbles of chert and quartz. Near the middle of this upper section, 10 to 25-foot beds of conglomerate contain pebbles and cobbles of granite (Table 6).

Fine-grained metabasaltic rocks similar to the rocks in the lower part of the measured section crop out in the Crater Lake area; in the area north of Sand Lake; and in the vicinity of Colberg, secs. 10 and 15, T. 18 N., R. 79 W. These units are altered to amphibolite (Table 7) but may have relict amygdules composed largely of quartz with lesser clinozoisite, biotite and amphibole (Pl. 10). The metabasalt is interbedded with chlorite schist (Table 6) that has a fragmental texture especially



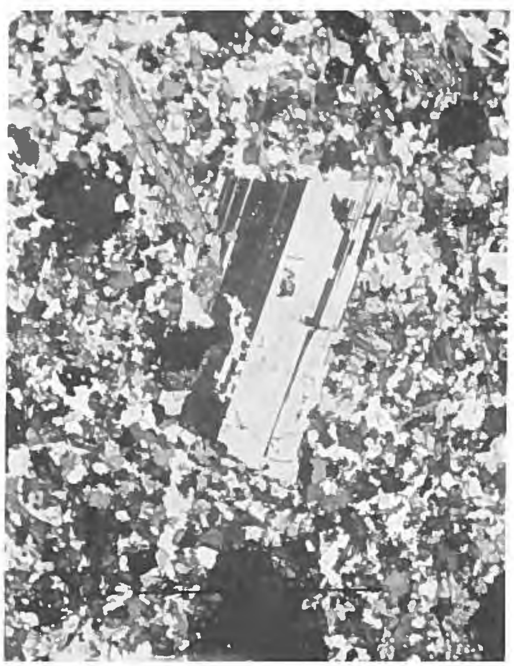
A

Amygdular basalt showing amygdules filled with quartz in the fine-grained matrix of secondary amphibole and plagioclase. Sample from near Colberg, sec. 10, T. 18 N., R. 79 W. Line 2 mm. long.



B

Amygdular basalt showing amygdules filled with quartz and clinozoisite in matrix of amphibolite. Sample from north of Arrastre Lake, sec. 10, T. 16 N., R. 80 W. Line 2 mm. long.



C

Relict phenocryst of plagioclase in recrystallized groundmass of amphibole, quartz, and plagioclase. Sample from near Colberg, sec. 10, T. 18 N., R. 79 W. Line 1 mm. long.

D

Irregular masses of chlorite schist in coarse-grained matrix of biotite-chlorite-quartz. Large euhedral crystal of quartz in upper left. Probable tuff from sec. 32, T. 18 N., R. 79 W. Line 1 mm. long.

Plate 10—Photomicrographs of metavolcanic (?) rocks of the Deep Lake Formation.

in the area north of Sand Lake. The matrix of the schist is fine-grained quartz and chlorite with lesser amounts of muscovite and amphibole. Large angular fragments of quartz and plagioclase and irregularly shaped masses of chlorite and chlorite-amphibole schist are set in this matrix (Pl. 10). This fragmental rock is regarded as metavolcanic(?). Both at Crater Lake and in the area north of Sand Lake exceedingly fine-grained, dense, quartz-rich rocks are interlayered with the chlorite schist and basalt. These rocks are composed of very fine-grained quartz and sericite and have metacrysts of calcite, chlorite, and sphene. Their origin is uncertain, but they may have been a felsic volcanic rock. Northeast of Crater Lake and south of Rock Mountain there is an occurrence of graphitic schist interlayered with metabasalt. This schist contains about 75 percent graphite, 15 percent quartz, and lesser amount of sericite and hematite. This unit appears to be unique to these two localities.

Table 6—DEEP LAKE FORMATION

(Upper part of section)

Gold Hill Section. Measured perpendicular to strike north of Gold Hill in secs. 10 and 11, T. 16 N., R. 80 W. Poorly exposed. Thickness approximate.

Lithology	Thickness (in feet)
Quartzite, white to orange, beds of pebbly quartzite up to 100 feet thick, pebbles well rounded, consist of quartz and chert	350
Norite sill, black to purple, massive, locally faintly foliated	1700
Quartzite, white to orange, coarse grained, large-scale cross-bedding	750
Quartzite, white to orange, coarse to medium grained, large scale cross-bedding, beds of conglomerate 2 to 3 feet thick. Contains well-rounded pebbles of quartz and slate	1200

(Lower quartzite lies on chlorite schist of Phantom Lake section.)

South Fork Rock Creek Section. Blackwelder, 1926, p. 626-627. Well-exposed.

Lithology	Thickness (in feet)
Quartzite, white to orange, local beds of pebbly quartzite, pebbles well rounded, consist of quartz and clay pellets	790
Quartzite, white to orange, beds of metaconglomerate 10 to 25 feet thick. Pebbles and cobbles in conglomerate are mostly granite	205

Base unknown

(Lower part of Section, base unknown)

Phantom Lake Section. Located north and west of Gold Hill, T. 16 N., R. 80 W., base of section at the confluence of Arrastre Creek and South Brush Creek. Section incomplete. Basal part measured on nose of plunging anticline. Poorly exposed. Thickness approximate.

Lithology	Thickness (in feet)
Chlorite schist, greenish-grey, fine-grained with interbedded pebbly quartzite	130
Metabasalt, greenish black, fine-grained, massive to faintly foliated	850
Conglomerate, greenish-grey, conglomerate to conglomeratic schist, containing pebbles and cobbles of granite and amphibolite	200
Metabasalt, greenish-black, fine-grained, faintly foliated	120
Metabasalt, greenish-black, amygdular, faintly foliated	180
Metabasalt, greenish-black, fine-grained, altered to amphibolite, in part. Very poor exposures	3,200
Quartzite, light green, fine-grained	350
Metabasalt, greenish-black, fine-grained	880
Quartzite, white, fine-grained	520
Amphibolite, dark green, coarse-grained, well foliated. May be metamorphosed sill	1,070
Quartzite, white to pink, fine-grained, feldspathic	130
Quartzite, white, sugary textured, massive at bottom, even bedded at top with layers ½" to 1" thick	575
Amphibolite, dark greenish-blue, medium-grained, poorly foliated	51
Quartzite, faint red color, fine-grained, contains 1 to 6 foot layers of sericitic quartzite in upper part	84
Amphibolite, dark greenish-blue, medium-grained, faintly foliated	60
Quartzite, silvery-white, sericitic quartzite, grades to complexly folded sericite schist at top	195
Metabasalt, dark green, massive in center, sheared and well-foliated along contact with quartzite	15
Quartzite, pink to white, sericitic, fine-grained	24
Amphibolite, dark blue to dark green, poorly foliated, may be sill	?
Quartzite, white to grey, medium to coarse-grained	?
Base not exposed	
Total thickness lower part (approximate minimum)	8,634±
Total thickness of measured section	12,634±
Thickness less sills	9,864±

Table 7—Mineralogy of schists and metabasalt of the Deep Lake Formation

(abundance in volume percent, visual estimates)

Mineral	A	$\overline{X}$	Range	Comments
			Chlorite-biotite schist	
Quartz	8/8	37	(15-54)	
Chlorite	8/8	27	(5-45)	
Muscovite	4/8	9	(0-30)	
Plagioclase	4/8	7	(0-30)	Albite and sodic oligoclase (may be clastic)
Biotite	5/8	7	(2-30)	
Opaque minerals	5/8	P		Largely magnetite and ilmenite (altered to leucoxene)
Carbonate	2/8	P		Some large crystals in matrix
Amphibole	1/8			Large metacrysts partly replaced by chlorite
Penninite, epidote, garnet, apatite, sphene, rutile, hematite		P		
			Fragmental chlorite schist (metatuff ?)	
Quartz	4/4	35	(27-52)	
Chlorite	4/4	26	(20-38)	
Muscovite	4/4	13	(2-30)	
Plagioclase	3/4	7	(0-20)	
Opaque minerals	3/4	3	(0-5)	Altered magnetite and ilmenite
Biotite, penninite, clinochlore, amphibole, zoisite, siderite, sphene, garnet, epidote		P		
			Metabasalt	
Amphibole	10/10	60	(41-70)	In part, actinolite. Uralite, in part.
Quartz	9/10	16	(0-31)	
Plagioclase	5/10	8	(0-25)	Metamorphic plagioclase oligoclase-andesine. Labradorite where less altered
Chlorite	6/10	6	(0-18)	
Sphene	6/10	3	(0-7)	Euhedral
Zoisite, opaque minerals, apatite, biotite, carbonate, talc, epidote		P		

Perhaps the most distinctive rock type in the Deep Lake Formation is a thick metaconglomerate (Fig. 3) found interlayered with chlorite schist, and metavolcanic rocks north of Sand Lake, along Deep Creek south of Crater Lake, in the vicinity of Colberg, and along South Brush Creek west of the confluence of Little Brush Creek and South Brush Creek. Unfortunately the stratigraphic position of this unit is not known as no conglomerate of comparable thickness was located in the measured stratigraphic section. The matrix of the conglomerate contains approximately 51% quartz, 21% chlorite, 12% muscovite, 7% amphibole, 4% opaque minerals (largely magnetite, 2% carbonate, 2% plagioclase, with accessory garnet, apatite, tourmaline, and biotite. Pebbles and cobbles make up 30% to 50% of the rock and consist chiefly of granite and chlorite schist. Other pebbles are porphyritic rhyolite, chlorite-biotite phyllite, chlorite quartzite, sericite quartzite, chlorite-sericite schist, sericite-martite schist, and amphibolite. Many of these rock types, notably the porphyritic rhyolite and sericite-martite schist, are not present in the Medicine Bow Mountains or elsewhere in mapped areas of Precambrian rocks of southeastern Wyoming. These rock fragments may be the only record of a group of metavolcanic and metasedimentary rocks older than the Deep Lake Formation.

Figure 3—Metaconglomerate of the Deep Lake Formation. Outcrop near the head of Three Mile Creek Canyon, T. 18 N., R. 79 W. (Photo by J. S. King.)

Three types of quartzite in the Deep Lake Formation are: (1) coarse-grained, pebbly quartzite, (2) sericite quartzite, (3) feldspathic quartzite (Table 8). These types of quartzite are interbedded with one another, but in general, the feldspathic and sericitic quartzites are more abundant in the lower part of the section, and the coarse-grained, pebbly quartzite is more common in the upper part.

The coarse-grained, pebbly quartzite is an orange-weathering rock with abundant quartz grains ½ to 2 mm. in diameter. It is marked by large scale cross-bedding having a thickness of 2 to 5 feet in the cross-bedded unit. Beds of pebble conglomerate ranging from 2 inches to 100 feet in thickness are interbedded with the quartzite. Pebble conglomerate beds 2 to 8 inches in thickness are most common. These conglomerate beds contain granules and pebbles of quartz, black chert, and grey-brown slate, and are usually well-sorted. Although pebble conglomerate may be found throughout the formation, it is especially conspicuous near the top. The upper 50 to 100 feet of the formation always contains thin beds of pebble conglomerate and in secs. 14 and 11, T. 16 N., R. 80 W., west of Dipper Lake the conglomerate reaches a thickness of over 100 feet. In some areas, notably in the area northeast of Sheep Lake, the coarse-grained quartzite is not so well bedded as elsewhere but contains fine-grained layers an inch or two in thickness that have well-developed cross-bedding. Thin laminae of gray slate are interlayered with the quartzite in a zone approximately 3,500 feet below the top. These beds are well exposed in sec. 27, T. 17 N., R. 79 W. in the area west of Lindsey Lake. The mineralogy of the coarse-grained quartzite is shown in Table 8.

Table 8—Mineralogy of quartzite of the Deep Lake Formation
(abundance in volume percent, visual estimates)

Mineral	A	$\bar{X}$	Range	Comments
			Coarse-grained, pebby, quartzite	
Quartz	16 16	68	(50-81)	
Muscovite	15/16	14	(5-22)	
Pagioclase	15/16	7	(0-22)	Oligoclase in part Probably clastic
Biotite	10/16	5	(0-20)	
Microcline chlorite, opaque minerals, zircon tourmaline, carbonate, garnet, sphene, epidote		P		Zircon, tourmaline, garnet sphene and epidote may be detrital. Some magnetite grains in euhedra
			Sericitic quartzite	
Quartz	5/5	64	(45-75)	Large and small grains, inter grown but not sutured
Sericite	5/5	26	(22-30)	In layers and interstitial
Chlorite	3/5	5	(0-22)	
Biotite, plagioclase, microcline, opaque minerals		P		
			Feldspathic quartzite	
Quartz	6/6	59	(45-68)	
Microcline	6/6	12	(5-17)	
Plagioclase	6/6	9	(3-15)	In one sample, An_{12}
Muscovite	6/6	11	(2-15)	
Biotite	4/6	3	(0-10)	
Chlorite sphene clinozoisite amphibole epidote, garnet, opaque minerals				Some accessory grains may be detrital

The sericitic quartzite is fine-grained and similar in composition to the other quartzite except for a larger proportion of sericite (Table 8). Fine-grained units with a lesser proportion of sericite are bluish. Sericite-rich types grade toward sericite schist and have a distinct schistose appearance. This quartzite is generally without recognizable bedding.

Feldspathic quartzite is found in the lower part of the section on Arrastre Creek and is more common in the northeastern part of the area near Arlington. It is a massive non-bedded type that weathers to a distinct pink color. The mineralogy of this rock type is shown on Table 8.

Siliceous marble, interbedded with quartzite and chlorite schist, crop out in four localities; on upper Trail Creek, sec. 26, T. 17 N., R. 79 W.; on Rock Creek, secs. 5 and 6, T. 17 N., R. 78 W., sec. 32, T. 18 N., R. 78 W.; on Cooper Hill, sec. 27, T. 18 N., R. 78 W.; and in a small outcrop in the east central portion of sec. 33, T. 18 N., R. 78 W. The unit is not well-exposed nor continuously exposed in any of these areas, and one cannot be certain that the same siliceous marble is present in each of the localities. The siliceous marble on Trail Creek is probably near the upper part of the Deep Lake Formation—2500 feet below the top. The siliceous marble is light brown and medium grained, and is marked by quartz-rich layers up to several inches in thickness (Fig. 4). The quartz-rich layers contain about 50% quartz, 15% biotite, 10% epidote, 15% plagioclase, and 15% muscovite, and the carbonate-rich layers contain about 60% carbonate, 20% quartz, and 20% plagioclase. The carbonate is a calcite-dolomite mixture with late veinlets of calcite.

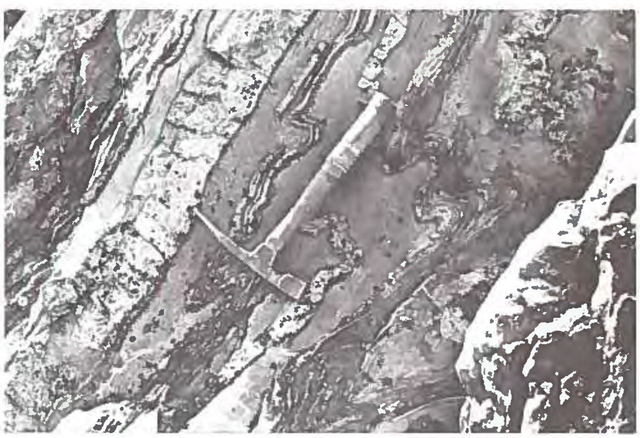


Figure 4—Siliceous marble of the Deep Lake Formation showing minor folds in the silica-rich layers. Outcrop in Rock Creek canyon, sec. 5, T. 17 N., R. 78 W. (Photo by J. S. King).

The conditions of deposition of rocks of the Deep Lake Formation are not known, primarily because there has not been detailed study of primary textures and structures that are present in the rocks. Some quartzite with well developed large scale cross-bedding and interlayered slate layers may be shallow-water marine. Certainly the presence of slate pebbles in the quartzite suggests periodic exposure of shale layers to dessication.

LIBBY CREEK GROUP

Rocks of the Libby Creek Group (Table 5) lie on quartzite of the Deep Lake Formation in the central Medicine Bow Mountains. The basal unit, the Headquarters Schist is in contact with quartzite of the Deep Lake Formation from the vicinity of South Twin Lake, sec. 22, T. 16 N., R. 80 W. to Rock Creek Knoll, sec. 35, T. 17 N., R. 79 W. The contact between the quartzite of the Deep Lake Formation is conformable with rocks of the Headquarters Schist throughout this area except in the vicinity of Rock Creek Knoll where reversals in dips in beds of the Headquarters Schist are probably related to faulting.

To the southwest, however, in T. 15 N., R. 81 W., the Headquarters Schist and rocks of the Deep Lake Formation are absent and the Heart Formation lies directly on the quartzo-feldspathic gneiss of the Precambrian basement. The Headquarters Schist may simply thin to the southwest, but exposures are such that this cannot be verified. Thin beds of conglomerate, that resemble units in the Headquarters Schist, are in the lower part of the Heart Formation in exposures along Mullen Creek at the southwest limit of outcrop. This suggests that the unit mapped as Heart Formation in the southwest includes a wedge-edge of the Headquarters Schist. As noted earlier the fact that rocks of the Libby Creek Group lie on basement gneiss beyond the outcrop limit of the metasedimentary rocks of the Deep Lake Formation suggests that there is an unconformity between these two groups of metasedimentary rocks.

Metasedimentary rocks of the Libby Creek Group are in a major syncline plunging to the northeast (Pl. 1). The axis of the syncline is in the southwest and the southeast limb of the syncline is cut out by the Mullen Creek-Nash Fork shear zone. This syncline extends to the northeast limit of outcrop of Precambrian rocks in the Medicine Bow Mountains (Pl. 1). In the central Medicine Bow Mountains, the type area of Blackwelder, the total thickness of rocks of the Libby Creek Group is about 22,200 feet. Top and bottom criteria, largely cross-bedding in quartzite, observed by the writer, Childers (1957), and Blackwelder (1926) show top to the southeast and support the contention that this is a normal stratigraphic succession.

The metasedimentary rocks of the Libby Creek Group are less deformed and are of lower metamorphic grade in the center of the outcrop area, T. 16 N., R. 79 and 80 N. Primary textures and structures are best preserved in this area. The rank of metamorphism and intensity of deformation increases to the southwest, and to a lesser extent to the northeast.

Headquarters Schist

The Headquarters Schist consists largely of interbedded phyllite and metaconglomerate. It also contains thin beds of quartzite like the sericitic quartzite of the Deep Lake Formation and beds of siliceous dolomite. In many areas incompetent beds of this formation are complexly folded and, in general, beds of the Headquarters Schist lack continuity. For example, the basal unit near South Twin Lakes, sec. 15, T. 16 N., R. 80 W. is a metaconglomerate approximately 200 feet thick, but in sec. 1, T. 16 N., R. 80 W. the basal unit is a fragmental chlorite schist regarded by Blackwelder (1926, p. 628) as a metamorphosed basic pyroclastic rock, and the basal unit at Reservoir Lake, sec. 6, T. 16 N., R. 79 W., is a conglomeratic phyllite containing a bed of siliceous dolomite up to 100 feet thick. The conglomeratic layers in the unit vary in thickness and are discontinuous: perhaps the best way to describe the unit is to consider it a phyllite with discontinuous layers of metaconglomerate and sporadic beds of quartzite and siliceous dolomite.

In the type area near South Twin Lake Blackwelder (1926, p. 628) measured 2,770 feet of interbedded phyllite, conglomerate, and quartzite. The unit may vary considerably in thickness, however, probably partly due to drag folding of incompetent units and partly due to variation in original thickness of the beds. For example the unit shows a greater map thickness near South Twin Lake, in sec. 12, T. 16 N., R. 80 W., and, near Rock Creek Knoll. It is much thinner near Reservoir Lake, sec. 6, T. 16 N., R. 79 W., on Rock Creek Ridge, sec. 21, T. 17 N., R. 78 W., and as noted above it may pinch out to the southwest where it is represented by a few thin beds of conglomerate in the Heart Formation. Part of this variation in thickness is probably a result of faulting (i.e. near Reservoir Lake), but it is the writer's opinion that much of it is depositional.

The phyllite is a green to brownish-green-muscovite-chlorite phyllite containing metacrysts of biotite, and it is finely laminated (Fig. 5 and Table 9). Laminations are developed by alternations of muscovite-chlorite rich layers with quartz rich layers (Fig. 5).

Table 9—Mineralogy of muscovite-chlorite phyllite (abundance in volume percent, visual estimates)

Mineral	A	$\bar{X}$	Comments
Quartz	2/2	31	
Muscovite	2/2	28	
Chlorite	2/2	18	
Opaque Minerals	2/2	8	
Biotite	2/2	4	In large metacrysts
Plagioclase	1/2	P	
Garnet	1/2	P	Anisotropic
Sphene	1/2	P	
Tourmaline	1/2	P	No = Green

(Sample from South Twin Lakes area, Central Medicine Bow Mountains.)

Chemically the phyllite is similar in composition to the average shale (Table 10). It does not have a high sodium to potassium ratio that characterizes some glacial clays and tills (Table 10). It is, however, similar in composition to argillite of the Precambrian Fern Creek Formation of Dickinson County, Michigan, that is considered a Pellodite or consolidated varved clay of glacial origin (Pettijohn, 1942, 1957, p. 345.)

Table 10—Chemical composition of muscovite-chlorite phyllite compared with varved shales of glacial origin, and average shale. Analyses B, C, D, E, F, and G from tables 61 and 62, Pettijohn, 1957, p. 344-345.

(Weight %)	A	B	C	D	E	F	G
SiO_2	58.21	59.20	50.33	52.00	62.74	66.87	58.10
TiO_2	0.81	1.20	1.13	—	—	0.47	0.65
Al_2O_3	17.34	16.14	19.17	16.11	16.94	15.36	15.40
Fe_2O_3	4.86	4.36	6.50	4.69	5.07	2.81	4.02
FeO	4.63	3.24	2.52		1.59	1.89	2.45
MnO	0.03	0.09	0.13	—	—	0.05	—
MgO	2.47	3.14	3.77	4.10	3.05	2.40	2.44
CaO	0.13	2.52	1.43	8.26	1.39	0.34	3.11
Na_2O	1.54	3.82	1.78	2.76	6.07	1.21	1.30
K_2O	5.38	1.97	4.03	1.74		6.60	3.24
P_2O_5	n.d.	0.17	0.14	—	—	0.23	0.17
CO_2	0.06	—	—	—a	—	0.28	2.63
H_2O+	4.12	1.16	4.87	—	3.20[b]	1.35	5.00
H_2O-	0.57	1.15	3.74	9.64[b]	0.36	none	
SO_3	n.d.	—	—	0.09	—	—	0.64
C	n.d.	1.94	0.41	—	—	0.04	0.80
Total		100.10	99.95	99.39	100.41	99.93	99.95

A. Muscovite-chlorite phyllite, South Twin Lakes area, Headquarters Schist, Medicine Bow Mountains. Analyst, Shirore Imae, Japan Analytical Chemistry Research Institute.

B. Summer silt, late-glacial varved sediment, Leppakosi, Finland, L. Lakka, Analyst (Eskola, 1932).

C. Winter clay, same as B.

D. Varved clay, north end Lake Timiskaming (Miller, 1905, p. 27).

E. Argillite, Cobalt series (Precambrian) Cobalt District, Ontario (Miller, 1905, p 42).

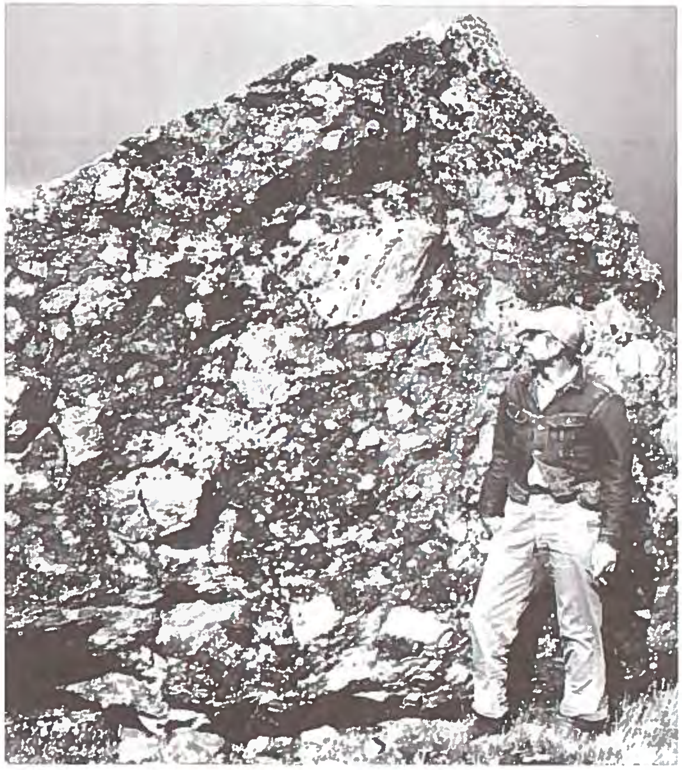
F. Argillite, Fern Creek Formation (Precambrian), Dickinson County, Michigan, B Bruun, Analyst.

G. Average shale (Clarke, 1942) p. 24)

a = Included in ignition loss

b = Loss in ignition.

Metaconglomerates of the Headquarters Schist are characterized by poor sorting (Pl. 11) and have every gradation in size of constituents up to boulders five feet in diameter. The clasts are angular to subrounded. The conglomerates also



A

Metaconglomerate (tillite ?) showing large angular boulders of quartzite and poor sorting of clases in the conglomerate. Glacial erratic near North Twin Lakes, central Medicine Bow Mountains.

B

Metaconglomerate (tillite) showing open framework, poor sorting, angularity of clasts. Outcrop near South Twin Lakes.

Plate 11—Metaconglomerate of Headquarters Schist.

A

Muscovite chlorite phyllite showing laminations. Knife for scale. Outcrop northwest of South Twin Lake.

B

Photomicrograph of muscovite-chlorite phyllite showing layering. Note orientation of micaceous mineral at nearly ninety degrees from trend of lamination, and metacrysts of biotite on left of photograph. Sample from South Twin Lake exposure. Line 2 mm. long.

Figure 5—Muscovite-chlorite phyllite of Headquarters Schist.

show great variation in texture from rocks composed largely of pebbles and cobbles to phyllites containing rare cobbles and boulders. In general, the conglomerates have an open framework and larger clasts are not in contact (Pl. 11). The composition of pebbles and boulders in the conglomerates is variable but the most abundant are granite, quartz monzonite, and quartzite followed by fine-grained felsic igneous rocks, chlorite schist, and phyllite.

Blackwelder regarded the conglomerates of this unit as tillites and the laminated phyllites as varved clay. The poor sorting, sub-angular shape of the boulders, and large size of certain boulders is certainly suggestive of glacial origin for the conglomerate. Also the presence of pebbles and cobbles randomly distributed through some of the phyllites is even more suggestive of glacial origin. In recent years (Dott, 1961), it has become common practice to question the glacial origin of deposits of this type particularly since it has been recognized that mass movements during deformation followed by periodic resedimentation by turbidity currents could produce deposits quite similar to the conglomerate described above.

Heart Formation

The Heart Formation is composed chiefly of quartzite, but has interbeds of chlorite schist, phyllite, biotite schist, muscovite schist, calc-schist, biotite-staurolite schist, and rare layers of conglomerate. Some of the dark colored quartzites especially in the type area near Heart Lake

Table 11—Heart Formation

(Heart Metagraywacke of Blackwelder)

Composite section pieced together from different areas at northeast end of Snowy Range. Type section of Blackwelder, 1926, p. 631-632. Very poorly exposed.

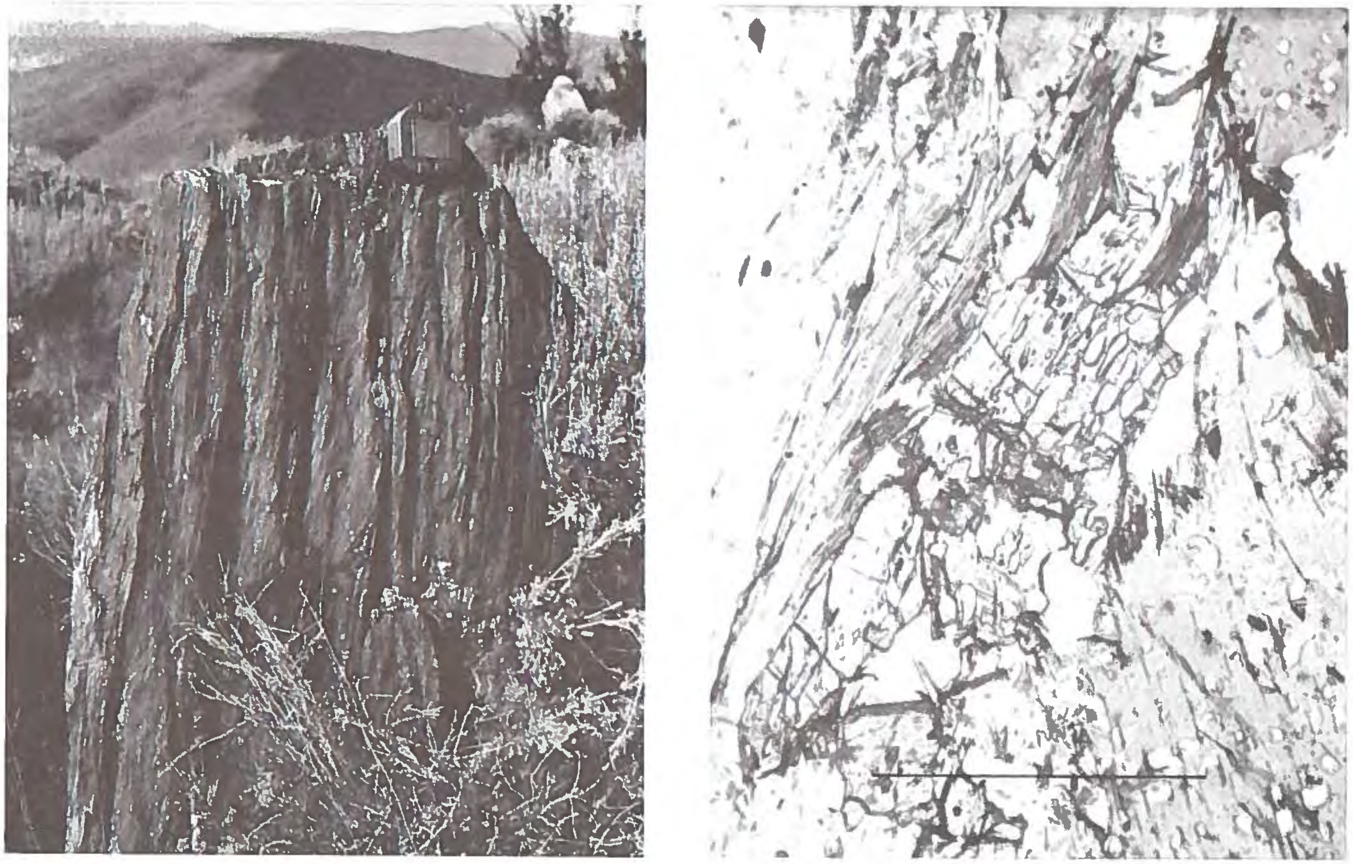
(Conformable contact with Medicine Peak Quartzite)

Lithology	Thickness (in feet)
Quartzite gray to pale olive, muscovite, biotite, and feldspar present, beds of gray sericite schist	250
Quartzite gray white, schistose	100
Quartzite, white, massive	125
Quartzite, olive gray, thin beds of schistose quartzite	90
Quartzite, gray, laminated, (metagraywacke of Blackwelder)	20
Quartzite, olive gray to smoky, laminated (metagraywacke of Blackwelder)	15
Quartzite dark-gray, slaty, with pyrite cubes (metagraywacke of Blackwelder)	100
Quartzite dark gray, spotted (metagraywacke of Blackwelder)	10
Schist dark chloritic, interlayered with schistose quartzite (metagraywacke of Blackwelder) (very poorly exposed)	270
Quartzite smoky gray, spotted, good cross-bedding (metagraywacke of Blackwelder)	30
Phyllite, interbedded with thin-bedded quartzite (metagraywacke of Blackwelder), (very poorly exposed)	35
Quartzite, smoky gray, spatulate current marks (metagraywacke of Blackwelder)	15
Quartzite light gray, thin schistose layers (metagraywacke of Blackwelder)	100
Quartzite, gray white, massive (metagraywacke of Blackwelder)	20
Quartzite dark gray, white spots $\frac{1}{4}$ to $1\frac{1}{2}$ inches in diameter, contains feldspar, muscovite and biotite	75
Quartzite, white, massive, interlayered with slaty quartzite (metagraywacke of Blackwelder) also beds of gray laminated quartzite, spotted quartzite, and slate	250
Quartzite, greenish gray, thin bedded, spotted, dense, (metagraywacke of Blackwelder)	10
In conformable contact with beds of Headquarters Schist	
Total Thickness	1,515

Mullen Creek Section at southwest limit of outcrop, T. 14 N., R. 81W. Poorly exposed. Approximate thickness.

(Conformable contact with Medicine Peak Quartzite)

Lithology	Thickness (in feet)
Schist, gray, crenulated, composed of muscovite, biotite and staurolite	130
Quartzite, white, sericite	130
Schist, dark gray, biotitic, with pebbles of granite scattered through part of unit	155
Amphibolite, greenish black, foliated	70
Quartzite, gray, sericitic, (very poor exposures)	1200
Amphibolite, greenish black, well-foliated	225
Quartzite, grayish white, sericitic, laminated	220
Marble, gray, massive to laminated (very poorly exposed)	135
Quartzite, white to gray, sericitic, faint lamination (very poorly exposed)	340
Amphibolite greenish black, foliated	120
Quartzite, gray, sericitic (very poorly exposed)	340
Quartzite, gray, with layers of sericite schist, crenulated schist layers	120
Quartzite, gray, sericitic faint lamination (very poorly exposed)	190
Amphibolite greenish black, very well foliated, some layers folded	85
Quartzite, gray, sericitic	50
In conformable contact with quartzofeldspathic gneiss Gneiss highly biotitic at this locality	
Total thickness	3,510
less bodies of amphibolite	3,000



A B

Figure 6—Staurolite schist of the Heart Formation. A. Complex crenulations of more than one generation in schist from outcrop on north side of Mullen Creek. B. Photomicrograph of same rock showing large sieved metacryst of staurolite. Biotite in lower right.

are rich in sericite and chlorite that may have been clay-sized matrix originally. These rocks can be classed as graywacke (Williams, Turner, and Gilbert, 1954, p. 293-294) and were viewed as such by Blackwelder. They do not contain rock fragments, however, and the writer prefers to consider them as metamorphosed impure, perhaps first cycle, sandstones. In any event, the Heart Formation contains a variety of rock types other than graywacke so the nomenclature of Blackwelder is modified to Heart Formation.

The lithology of the unit in the type area appears in Table 11. The interbedded schists are of the greenschist facies of regional metamorphism in this area, but to the southwest where the Formation is exposed between Mullen Creek and French Creek the metamorphic rank is higher and such rocks as biotite-staurolite schist (Fig 6, Table 11) are interbedded with quartzite. The Heart Formation is also thicker in the southwest (Table 11) and as noted above may include rocks in the stratigraphic position of the Headquarters Schist.

The mineralogy of some typical rocks of the Heart Formation is presented in Table 12, where rocks both from the type area and the exposures to the southwest are compared.

Plate 12—Medicine Peak Quartzite exposed on east face of Snowy Range axial remnant. Outcrop near top of Medicine Bow Mountains north of Lake Marie.

Table 12—Mineralogy of rocks of the Heart Formation
(visual estimate volume percent)

Mineralogy	A	$\bar{X}$	Range	Comments
Sericite Quartzite (near North Twin Lakes)				
Quartz	1 1	46		
Muscovite	1 1	25		
Chlorite	1 1	15		
Plagioclase	1/1	7		An_{31}, probably detrital
Biotite	1 1	3		
Tourmaline	1 1	P		
Sphene	1 1	P		
Rutile	1 1	P		
Muscovite schist (near North Twin Lakes)				
Muscovite	2 2	32		
Quartz	2 2	35		
Chlorite	2 2	13		Some large crystals along cleavage planes
Biotite	1 2	13		
Opaque minerals	2 2	3		Some altered to leucoxene
Tourmaline	1 2	P		No = green
Apatite	2 2	P		
Sphene	1 2	P		
Biotite schist (southwest, near Mullen Creek)				
Quartz	3 3	38	(30-45)	
Biotite	3 3	28	(20-35)	
Muscovite	3 3	20	(7-28)	Some irregular masses may be after Al_2SiO_5 mineral
Chlorite	1/3	5	(0-15)	Probably late
Garnet	2 3	3	(0-5)	Inclusions in garnet suggestion of rotation
Kyanite	1 3	3	(0-10)	In metacrysts
Tourmaline	2 3	P		Zoned, No = green
Opaque minerals	2 3	P		
Sphene	1 3	P		
Calc-schist (southwest, near French Creek)				
Carbonate	1 1	67		
Quartz	1 1	15		Grains well rounded
Amphibole	1 1	7		Pale green
Phlogopite	1 1	5		
Microcline	1 1	3		
Plagioclase	1 1	P		An_{32} =
Opaque minerals	1 1	P		
Epidote	1 1	P		
Chlorite	1 1	P		
Biotite-staurolite schist (southwest, near Mullen Creek)				
Quartz	1 1	41		
Muscovite	1 1	35		
Biotite	1 1	7		
Staurolite	1 1	6		Large metacrysts
Garnet	1 1	5		
Opaque minerals	1 1	4		Magnetite altered, in part, to hematite
Kyanite	1 1	P		
Tourmaline	1 1	P		

The environment of deposition of rocks of the Heart Formation is not known. As was the case for the Deep Lake Formation a more detailed study must be made of primary structures especially in the central Medicine Bow Mountains before satisfactory conclusions can be made concerning conditions of deposition.

Medicine Peak Quartzite

This great resistant body of quartzite, nearly 6,000 feet thick, that underlies the crest of the Medicine Bow Mountains is the most prominent Precambrian rock type of the Medicine Bow Mountains and very likely the State of Wyoming (Pl. 12). The quartzite conformably underlies the Lookout Schist and conformably overlies rocks of the Heart Formation. It crops out in a northeast trending belt from Mullen Creek in the southwest to Rock Creek Ridge in the northeast (Pl. 1).

Originally Blackwelder subdivided the quartzite into lower and upper units and the lithology of these units is shown in Table 13. The description and thickness given for the lower unit is that for a section along French Creek near the southwestern limit of the quartzite. The description and thickness of the upper unit is for a well exposed section that crops out north of Silver Lake. As previously noted distinction between the upper and lower units of the quartzite is not made in this report because the transition between these units is gradual so that the contact is picked with difficulty.

Table 13—Medicine Peak Quartzite

Upper part of section measured by Childers, with brunton and tape. Well exposed section north of Silver Lake. Conformable contact with Lookout Schist.

Lithology	Thickness (in feet)
Quartzite, blue, coarse-grained, many thin beds of conglomerate containing well rounded quartzite pebbles up to ¾ inch in diameter	480
Quartzite, brown to buff, schistose, fractured, may contain small veins of quartz up to 10 inches thick, generally conglomeratic with quartzite pebbles up to ¼ inch in diameter	740
Quartzite, white, coarse-grained, some beds of conglomerate with pebbles ranging from ½ inch to 1 inch in diameter. Conglomerate shows graded bedding	370
Quartzite, buff, very coarse-grained, thin beds of conglomerate	400
Quartzite, white, coarse-grained, poorly exposed	200
Quartzite, white, coarse-grained, scattered cobbles	230
Quartzite, tan to white, medium-grained, some conglomerate, some lenses up to 15 feet long and 3 feet wide not recrystallized	280
Quartzite, white, medium-grained, some green kyanitic beds	820
Conglomerate, white, pebbles ranging from 1/16 to 1/4 inch in diameter	430
(Lower part of section from mapped area, Sec. 35, T. 15 N., R. 81 W., thickness approximate)	
Quartzite, light blue, light green, and white, coarse-grained, locally massive, local kyanite-rich areas	620
Quartzite, blue to blue-gray, coarse-grained, kyanitic, well-developed cross-bedding, veins of rose quartz containing kyanite crystals	730
Quartzite, dark blue, dark green and dark gray, medium to coarse-grained	350
(Conformable contact with Heart Formation)	
Total Thickness	5,650

The lower part of the Medicine Peak Quartzite is a distinctive bluish to deep green, kyanitic quartzite. The lower 1000 feet is blue-grey, coarse-grained quartzite with well-developed cross-bedding. The cross-bedding is distinguishable because of dark-purple to reddish coloration along bedding planes. The thickness of the cross-bedded unit ranges from three inches to two feet (Fig. 10). The cross-bedding of this unit is unusual because it is deformed. Examples of the deformation in the cross-bedded unit are shown on Figure 7. The deformation of the cross-bedding may be a type of interformational recumbent folding (McKee, et al., 1962) which it resembles very closely (Fig. 8), but the most highly deformed cross-bedding is in the southwest where there is evidence of refolding (Houston and Parker, 1963). The blue quartzite grades upward into a green kyanitic quartzite, which is also coarse-grained.

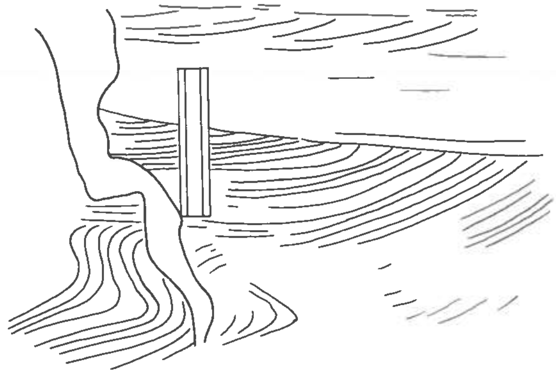


Figure 7—Deformed cross-bedding in the Medicine Peak Quartzite. Sketch from photograph taken on north side of French Creek, sec. 27, T. 15 N., R. 87 W. Six inch ruler for scale.

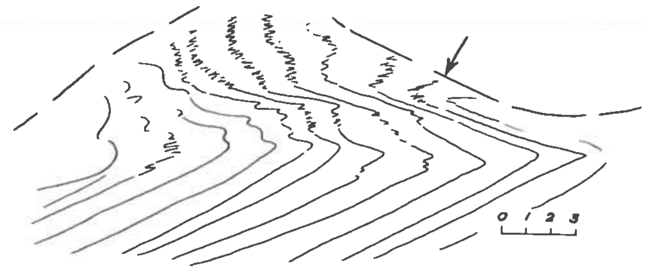


Figure 8—Deformed cross-bedding in unconsolidated sandstone generated by lateral push with board from right to left as shown by arrow Scale in inches. After McKee and others (1962, p. D155-D160).

This quartzite grades into the white, finer grained quartzites of the upper Medicine Peak Quartzite.

The mineralogy of the lower Medicine Peak Quartzite is presented in Table 14. The quartzite is remarkably pure, averaging 83 percent quartz and containing kyanite and muscovite as the only important accessory minerals. The kyanite is scattered through the rock and is also found in quartz-kyanite veinlets that cut the quartzite but are themselves deformed in some areas (Fig. 9).

Table 14—Mineralogy of lower part of Medicine Peak Quartzite (visual estimates, thin section)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	8/8	83	(75-97)	Strained, in single crystals, and veinlets
Kyanite	7/8	10	(0-15)	Altered, in part, to muscovite
Muscovite	8/8	4	(2-8)	
Tourmaline	6/8	P	—	No = green
Opaque minerals	5/8	P	—	Magnetite
Epidote	1/8	P	—	Well Rounded

Samples from Southwest part of area between French Creek and Mullen Creek

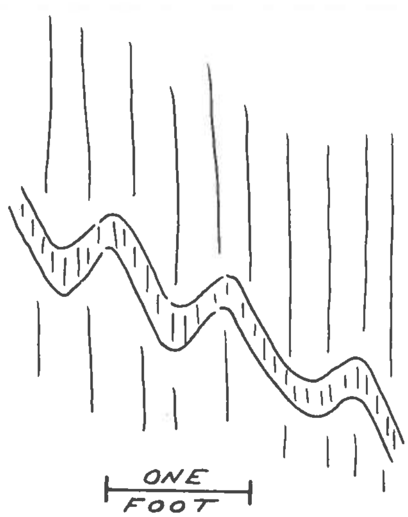


Figure 9—Field sketch of deformed quartz-kyanite veinlet in Medicine Peak Quartzite. Cleavage in veinlet parallels bedding (?) of quartzite. Veinlet composed largely of quartz with crystals of kyanite along margin. Sec. 35, T. 16 N., R. 81 W.

The upper part of the Medicine Peak Quartzite is a white medium to coarse-grained quartzite that contains only small amounts of kyanite. The unit is marked by thin beds of conglomerate pebbles ranging from 1/16 to 1 inch in diameter. The pebbles are well-rounded quartzite, quartz, and rare red jasper. Large scale cross-bedding is common in the upper 500 feet of the quartzite with a thickness of the cross-bedded unit of two to three feet (frontispiece). This cross-bedding is not as well defined as that in the lower part of the Medicine Peak Quartzite, because the bedding planes are not defined by thin laminae of hematite as in the lower member. In some localities the uppermost 60 to 100 feet of the Medicine Peak Quartzite is a conspicuous bright green to emerald green color. A prominent outcrop of green quartzite is in sec. 9, T. 16 N., R. 79 W., due west of Big Telephone Lake, and this quartzite is also well developed in sec. 31, T. 17 N., R. 78 W.

The mineralogy of the upper part of the Medicine Peak Quartzite is shown in Table 15. The major difference between rocks of the upper and

Figure 10—Photograph of kyanite-rich lower part of Medicine Peak Quartzite showing cross-bedding. Outcrops south of French Creek near southwest extremity of metasedimentary rocks.

lower units is in the kyanite content. Kyanite was not noted in any of the sections studied of the upper part. Another interesting aspect of the mineralogy of the upper Medicine Peak Quartzite is the relative abundance of zircon in some samples. In one sample zircon made up nearly one per cent of the rock and was in very large grains up to 0.3 mm. in diameter. The uppermost green quartzite does not differ mineralogically from the rest of the upper Medicine Peak, but it does contain a much larger percentage of muscovite, reaching approximately 17 percent.

Table 15—Mineralogy of upper part of Medicine Peak Quartzite (visual estimates, thin section)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	7/7	88	(85-91)	
Muscovite	7/7	6	(1-10)	
Opaque minerals	6/7	5	(0-7)	Large euhedral crystals of magnetite, also in veinlets altered to hematite
Zircon	7/7	P		Well rounded, generally, some grains, 0.3 mm dia
Tourmaline	1/7	P		No blue
Sphene	4/7	P		

Samples from area north of Silver Lake

LOOKOUT SCHIST

The Lookout Schist is a laminated quartz-muscovite schist that conformably overlies the Medicine Peak Quartzite and conformably underlies the Sugarloaf Quartzite. It is found throughout the central Medicine Bow Mountains and extends to the eastern limit of the outcrop of Precambrian rocks, but it thins to a wedge-edge to the southwest in sec. 4, T. 15 N., R. 80 W. (Pl. 1). The unit is an incompetent layer between the two massive bodies of quartzite and is complexly folded in many areas and is variable in thickness. It ranges in thickness from zero to over 1300 feet.

The Lookout Schist contains interbeds of quartzite and chlorite-muscovite schist. Sills of metadiabase are common (Table 16). Many of the beds of quartzite have small-scale cross-bedding. The most distinctive rock is the laminated schist (Pl. 13) that has alternating layers that are muscovite-poor and muscovite-rich. The muscovite- rich layers are 1/64 inch and the muscovite-poor layers are 1/8 inch thick. In the area west of Telephone Lakes the laminated schist is relatively undeformed and shows a remarkable preservation of convolute lamination (Pl. 13). The convolute lamination is confined to this unit and is not found in interbedded quartzite. The origin of such structures is not known, but as suggested by Potter and Pettijohn, (1963, p. 155), it may result from load deformation during sedimentation.

Table 16—Lookout Schist

Section measured in sec. 9, T. 16 N., R. 79 W., west of Telephone Lake, taped section. Probably fault contact with Nash Fork Formation, Sugarloaf Quartzite missing.

Lithology	Thickness (in feet)
Quartzite, blue, medium to fine-grained, cross-bedded	8
Schist, greenish gray, chlorite	53
Quartzite, blue, fine-grained, cross-bedded	22
Diabase, green, foliated, altered to green-schist, in part	77
Schist, gray, laminated, quartz muscovite rich, convolute laminations	151
Amphibolite, dark green, fine grained	43
Schist, gray, laminated, quartz muscovite rich, layers of white quartzite 5 feet thick	71
Quartzite, white, cross-bedded	31
Schist, gray, laminated quartz muscovite rich, exceptionally well-developed convolute structure	730
(conformable contact with Medicine Peak Quartzite)	
Total Thickness	1,186

The laminated schist is complex mineralogically (Table 17) and contains a large number of heavy minerals such as tourmaline, zircon, apatite, and sphene. Some beds contain up to 6 percent siderite, and Blackwelder reported a layer of dolomite in a section that he examined. The chlorite-muscovite schist contains approximately 45% muscovite, 35% chlorite, 3% quartz, 15% tremolite and minor amounts of sphene and zircon. The presence of tremolite suggests that these units may originally have been calcareous slates.

A
Convolute laminations in laminated quartz-muscovite schist. Outcrop west of Telephone Lake, sec. 8, T. 16 N., R. 79 W.

B
Secondary folds in quartz-muscovite schist showing development of axial plane cleavage. Outcrop in SW¼ sec. 2, T. 16 N., R. 79 W.

Plate 13—Primary and secondary fold in Lookout Schist.

A
Sugarloaf Mountain northwest of Lewis Lake in central Medicine Bow Mountains.

B
Ripple marks in Sugarloaf Quartzite exposed in road cut on north side of Highway 130 west of Lake Marie.

Plate 14—Sugarloaf Quartzite

Table 17—Mineralogy of laminated, quartz-muscovite schist
(visual estimates, thin section)

Minerals	A	$\bar{X}$	Range	Comments
Quartz	6/6	53	(38-66)	
Muscovite	6/6	17	(7-35)	
Feldspar	6/6	9	(6-15)	Mostly oligoclase, some samples contain small amount microcline, detrital (?)
Chlorite	4/6	7	(0-15)	
Opaque minerals	6/6	5	(2-8)	Magnetite, euhedral crystals common
Zircon	6/6	P		Mostly well rounded, some large grains, some euhedral grains
Tourmaline	5/6	P		
Biotite	2/6	P		
Sphene	2/6	P		Well rounded
Apatite	3/6	P		
Siderite	1/6	P		
Epidote	1/6	P		Well rounded
Garnet	1/6	P		
Amphibole	1/6	P		

Samples from area west of Telephone Lake

Ten Haaf (1956) and Sanders (1960) have suggested that convolute lamination is a characteristic structure of turbidite sequences, but the absence of graded bedding, and the presence of small-scale cross-bedding in associated quartzite suggests that the Lookout Schist is not a turbidite sequence. Dott and Howard (1962) have pointed to convolute laminations in non-turbidite sequences in the Western United States that are mostly near-shore marine, shallow-water sequences. One of these sequences near Mack Arch, Oregon, is very similar lithologically to the Lookout Schist (Dott and Howard, 1962, p. 116-117). The weight of the evidence suggests that the Lookout Schist is a shallow-water marine unit perhaps deposited in a near-shore environment.

SUGARLOAF QUARTZITE

The Lookout Schist becomes more quartz-rich near the top and grades into the Sugarloaf Quartzite, a white remarkably homogeneous quartzite which is glassy in appearance on fresh surface. Blackwelder (1926, p. 636) made a most apt observation on this rock stating that isolated pieces of it resemble smooth pieces of Castile soap. This is especially true of pieces of quartzite that have been transported and rounded somewhat.

The Sugarloaf Quartzite is characterized by homogeneity. The bulk of the unit is white non-bedded quartzite, but there are some beds of pale blue quartzite, thin layers of schist, and near the top of the unit at the contact with the Nash Formation, beds of metadolomite may be present. Blackwelder reports (1926, p. 635) a few thin beds of quartz pebble conglomerate near the top. Both large and small scale cross-bedding is present, and some bedding planes have oscillation ripple marks (Pl. 14). The quartzite is composed almost entirely of quartz and averages over 95 percent in samples examined. Small amounts of sericite and a few opaque minerals are the only additional minerals observed.

The Sugarloaf Quartzite must have been an orthoquartzite originally and, as such, a texturally and mineralogically mature sandstone. Such a sandstone may be a product of reworking during marine transgression, or it may result from constant reworking along a stable shoreline.

NASH FORK FORMATION

The Nash Fork Formation as herein defined includes the Nash Marble, Anderson Phyllite, and Ranger Marble of Blackwelder (Table 5). As noted previously, this change in terminology is made because black phyllites and slates similar to the Anderson Phyllite occur at different stratigraphic horizons in the Nash Marble. The Nash Marble and Ranger Marble are difficult to distinguish in outcrops to the southwest. Also the marbles of Blackwelder prove to be metadolomites and although the term marble can be used to describe metamorphic rocks consisting entirely of dolomite—metadolomite seems desirable to avoid semantic confusion.

In his description of these units Blackwelder regarded the Anderson Phyllite as a unit consisting chiefly of black slates and phyllites that occurred stratigraphically above the Nash Marble and below the Ranger Marble. This is generally true in the type area of the central Medicine Bow Mountains, but to the southwest black phyllite is at the base of the units. South of Sugarloaf Mountain there are two distinct layers of phyllite separated by metadolomite but on Rock Creek Ridge there is only one thin bed of black phyllite in the lower part of the unit (Pl. 1). It seems best therefore to regard the Nash Fork Formation as a metadolomite with lenticular bodies of calc-biotite phyllite and black slate.

The metadolomites are not highly deformed and show a great range in composition, texture and structure that is considered primary. The metadolomite ranges from massive metadolomite to beds that have thin inter-layers of black slate and calc-biotite phyllite, to very distinctive laminated rocks that have alternating layers of dolomite and quartz to sedimentary breccia layers, to metadolomite partially replaced by silica. Primary structure has been observed in the metadolomite including cross-bedding and current marks, but the most interesting and diagnostic feature of the rock is the dome-like structure found on different scales and at different stratigraphic levels throughout the units. These structures were regarded as algal in origin by C. E. Walcott (Blackwelder, 1926, p. 639) who considered them to belong to the genus **Collenia.** Fenton and Fenton (1939, p. 92-95) made a more detailed study of the algae and named them **Hadrophycus immanis.** The **Hadrophycus immanus** or "mighty sea plant" of the Fentons included both dome-shaped bodies and horizontally stratified units; both consisting of alternating layers of quartz and dolomite. Subsequently Hensley (1955, p. 1676) noted a difference in texture between the horizontally laminated type and the domed type and suggested that the term **H. immanis** be used to include only the domed variety.

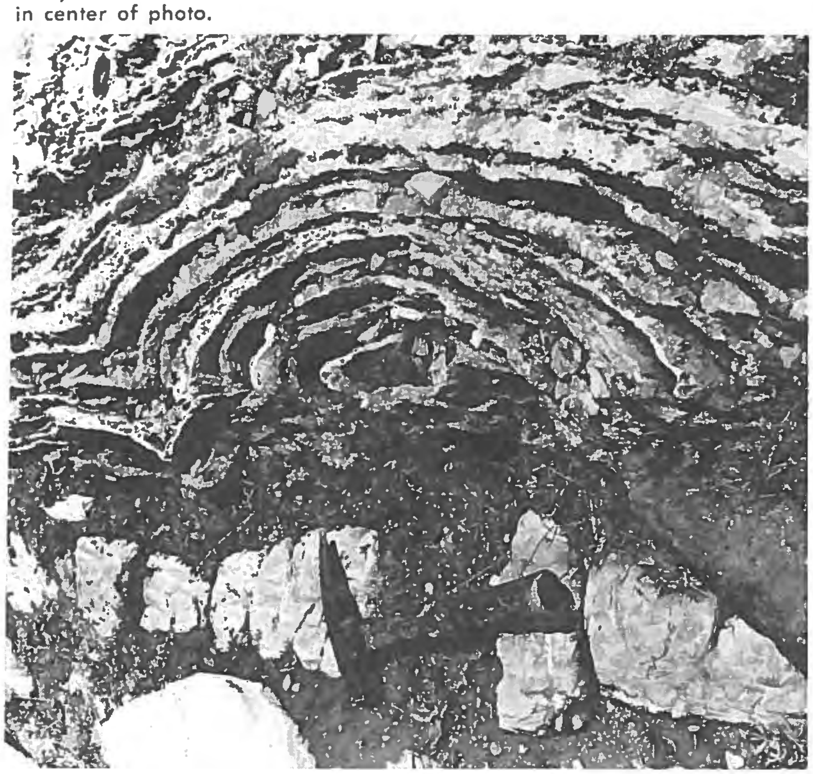
There seems to be agreement among paleontologists who have examined these structures that they are algal deposits of some sort although precise classification may be difficult. Some of these units are a series of linked domes L L H (laterally linked hemispheres) of Logan, et al, 1964, found in individual layers of the metadolomite (Pl. 15). The domes vary in width from a few inches to several feet, but individual layers always contain domes of the same width. The dome layers are interstratified with laminated metadolomite consisting of alternating layers of quartz and metadolomite as well as some layers of pure dolomite ranging in thickness from six inches to tens of feet. Commonly the layers of pure dolomite lie at the base of the dome structure (Pl. 15), and domes seem to have developed on the metadolomite layer. As noted by Hensley (1955, p. 1076), the domes are made up of laminae of alternating layers of fine-grained quartz and quartz-dolomite intergrowths. The quartz-dolomite layers have quartz rods or pillars connecting the laminae. The interbedded laminated metadolomite is also made up of quartz layers alternating with quartz-dolomite mixtures but does not have the pillars (Pl. 15). If all the dome-shaped masses and laminated metadolomite is algal, approximately one-third of the meta-dolomite is of algal origin.

In a recent detailed study of the "algal" structures Knight and Keefer (1966, p. 1-11) have noted three types of structures which they class as Type 1A, Type 1B, and Type II. Type 1A is spheroids and hemispheroids that vary greatly in size and are marked by distinctive rod-like structures oriented normal to the layering. Type 1B is similar to 1A in gross structure, but instead of rod-like structures oriented perpendicular to layering, 1B is marked by irregular open tube which are nearly parallel to the layering. Type II consists of silicified layers separated by layers of metadolomite and does not have the tubes or rod-like structure. Type II structures are illustrated in Plate 15. The most interesting feature of the "algal" structures noted by Knight and Keefer was the spheroid shape that many of the structures have rather than an expected dome-like or hemispheroid form. Some of the spheroidal heads developed around angular rock fragments, not unlike giant oolites. Knight and Keefer (1966, p. 10) do not specify an algal origin for the structures, but do state that the structures are organo-sedimentary features of some type.

The phyllite and slate interbedded with the metadolomites are of two major types; a black slate similar to the French Slate and a calc-biotite phyllite with metacrysts of biotite, sphene, and pyrite (Table 18). The black slate is distinctly laminated with alternating layers rich in muscovite and layers rich in quartz. The phyllite, on the contrary, is well-foliated as a result of distinct mineral parallelism of micaceous minerals but does not have the well-developed laminae. Some beds in the phyllite-slate unit are rich in iron. These beds, which contain numerous pyrite cubes and are graphitic, fall into the sulfide facies of iron formation of James (1954). Rocks of this composition mark the base of the original Anderson Phyllite of Blackwelder (Table 5), in secs. 30 and 31, T. 16 N., R. 79 W., along upper French Creek, and in the rather thin layers of phyllite on Rock Creek Ridge. The phyllite and slate often have thin layers of carbonate interbedded with them, and it is interesting to note that the layers as well as the interstitial carbonate in these rocks is calcite rather than dolomite as in the related rocks.

A

Algal structure in Nash Fork Formation from exposures southeast of Libby Lake in central Medicine Bow Mountains. Note dome-like structures in center of photo.



B

Close-up of dome-like structure from same outcrop as A.

Plate 15—Nash Fork Formation.

Table 18—Mineralogy of rocks of the Nash Fork Formation
(visual estimates, thin section)

Mineral	A	$\bar{X}$	Comments
		Metadolomite (Nash Marble of Blackwelder)	
Carbonate	1/1	85	Stain test shows all carbonate is dolomite
Quartz	1/1	10	Scattered grains, angular
Muscovite	1/1	5	As thin laminae
		Metadolomite (Ranger Marble of Blackwelder)	
Carbonate	1/1	84	Stain test shows all carbonate is dolomite
Quartz	1/1	15	In one area, quartz is interstitial surrounding large masses of carbonate
Muscovite	1/1	P	
		Siliceous metadolomite (near Towner Lake)	
Carbonate	1/1	35	
Quartz	1/1	35	Two types, irregular grains in carbonate, and fine-grained associated with talc in layers
Talc	1/1	20	In laminae with fine-grained quartz
Plagioclase	1/1	5	Detrital (?) oligoclase
Opaque Minerals	1/1	P	Magnetite and hematite
Apatite	1/1	P	
		Calc-biotite phyllite (between Towner and Telephone Lakes)	
Quartz	1/1	20	
Chlorite	1/1	20	
Carbonate	1/1	15	Stain tests show all carbonate calcite
Amphibole	1/1	15	Tremolite
Biotite	1/1	10	As large metacrysts
Zoisite	1/1	10	
Sphene	1/1	5	As large metacrysts
Pyrite	1/1	P	
		Slate (between Towner and Telephone Lakes)	
Quartz	1/1	55	
Muscovite	1/1	33	
Opaque minerals	1/1	7	Small grains, both pyrite and magnetite, locally altered to hematite
Chlorite	1/1	5	

Two major clues to the environment of deposition of the rocks of the Nash Fork Formation are the "algal" structure of the metadolomite and the pyritic black phyllite of the sulfide facies of iron formation. The presence of "algal" beds interbedded with cross-bedded limestone and shale suggest deposition in relatively shallow, well-oxygenated waters that periodically received an influx of fine material that temporarily stopped the growth of algae. The black pyritic phyllite, on the other hand, requires reducing conditions for its formation, and James (1961, p. 46) has suggested a restricted basin separated from the open ocean by barriers that would inhibit circulation and allow the reducing conditions to develop. This seems to pose a mutually exclusive environment for the rock types known to be interbedded with each other. Perhaps the algal mats grew in intertidal zones and the reducing environments developed in adjacent restricted bays.

Another interesting aspect of these units is the possible clues to climate that the rocks yield. Algal mats have a wide climatic range, but they are best developed in tropical to subtropical oceans. Students of iron formations have suggested that the iron in solution must come from adjacent lowlands that have been weathered under tropical to subtropical conditions (James, 1954, 1961). If these clues are valid there must have been a significant change in climatic conditions from the time of deposition of the "glacial" (?) deposits of the Headquarters Schist.

TOWNER GREENSTONE

The Towner Greenstone is a dark chlorite amphibolite with both massive amphibolite and schistose amphibolite varieties. No contacts have been observed with the underlying formation, but Childers (1957, p. 27) noted a conformable, sharp contact with the overlying French Slate in the northeast wall of Silver Creek Canyon. The green-

Table 19—Mineralogy of rocks of Towner Greenstone
(visual estimates, thin section)

Mineral	A	$\bar{X}$	Comments
		Amphibolite (near Towner Lake)	
Amphibole	1/1	57	Actinolite
Quartz	1/1	15	
Carbonate	1/1	15	
Opaque Minerals	1/1	12	
Epidote	1/1	P	
		Chlorite amphibolite (near Towner Lake)	
Amphibole	1/1	73	
Chlorite	1/1	15	
Epidote	1/1	7	
Plagioclase	1/1	5	Albite

stone ranges in thickness from 600 feet to 1600 feet reaching its maximum thickness near the type locality at Towner Lake.

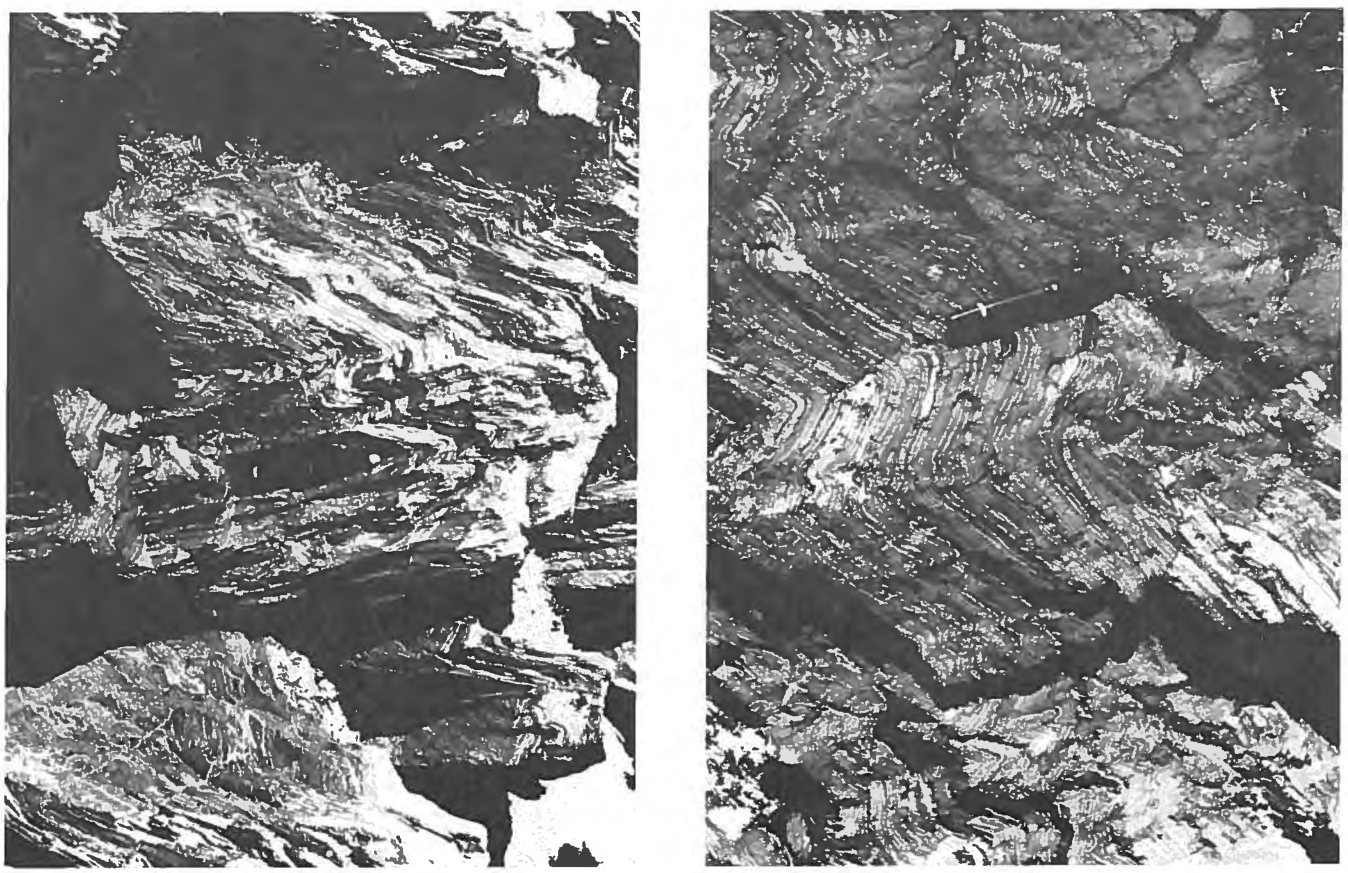
According to Childers, the basal portion of the Towner Greenstone contains many small lenses of poorly cemented sandstone about one foot in length and 2 and one-half inches in thickness, which are often calcareous. All outcrops observed by the writer were either massive amphibolite or schistose type rich in chlorite. The mineralogy of two samples of the massive type is shown in Table 19. The chlorite amphibolite of Table 19 is coarse-grained and somewhat resembles an altered mafic igneous rock, but the texture is not diagnostic.

The Towner Greenstone is certainly an enigmatic rock since it contains no diagnostic textures or structures, but the presence of small lenses of sandstone in the lower part suggest it was deposited on the surface as a flow or series of mafic tuffs or both.

FRENCH SLATE

The French Slate is the uppermost, youngest unit of the Libby Creek Group. It is a black, pyritic slate that extends from the axes of the syncline in the southwest to the northeastern limit of exposure of the metasedimentary rocks (Pl. 1). According to Blackwelder, it is about 2,000 feet in thickness in the type area, but the true thickness is probably not exposed because the upper beds of the unit are in contact with the Mullen Creek-Nash Fork shear zone along their entire outcrop. Beds in the axes of the syncline are so poorly exposed and cut by igneous intrusions that little can be determined as to their original character.

The French Slate consists of interbedded muscovite-chlorite slates and phyllites and includes some highly siliceous slates as well as one or two beds of quartzite (Pl. 1) near the top. Metacrysts of pyrite are scattered through the slates and phyllites, but are never as abundant as



A B

Figure 11—Complex crenulations in French Slate showing chevron form of some folds. Outcrops on north side of highway 130 west of area where Nash Fork Creek crosses road.

in the iron-rich units of the Nash Fork Formation. The phyllites have discontinuous laminae rich in muscovite and chlorite in a matrix of quartz. They commonly have metacrysts of biotite and pyrite that grew perpendicular to and across the laminae. The slate has well-developed continuous laminae consisting of layers rich in muscovite and chlorite (20% muscovite, 15% opaque minerals, 15% quartz, and 10% chlorite) alternating with layers rich in quartz (85% quartz and 15% muscovite and chlorite). The slate has very complex crenulations (Figs. 11 and 12) of more than one generation, perhaps developed during the formation of the syncline and as a result of movement along the fault.

The rocks of the French Slate are regarded as metamorphosed equivalents of normal marine shales and as such indicate a period of marine transgression following deposition of the rocks of the Nash Fork Formation.

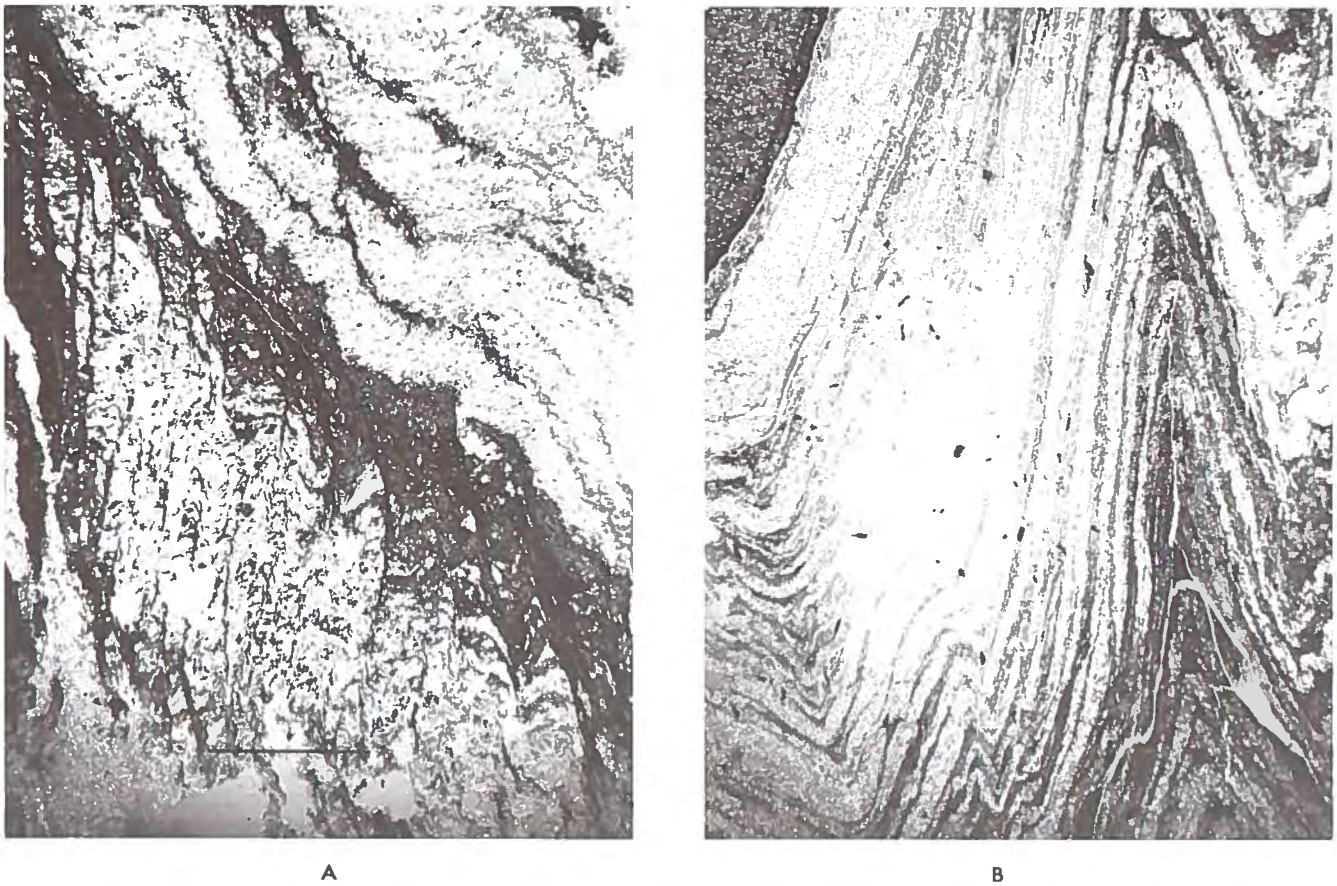


Figure 12—Photomicrographs of folding bedding in French Slate. A. Cleavage developed at angle to bedding. B. White areas are where biotite metacrysts were present that were lost during sectioning. Uncrossed nicols. Line 2 mm. long, both photos same scale.

IGNEOUS ROCK OF PRECAMBRIAN AGE NORTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

A variety of igneous rocks crop out in the area north of the Mullen Creek-Nash Fork shear zone. These igneous rocks probably range in age from older than 2.4 b.y. to rocks equivalent in age to the Sherman Granite of 1.35 b.y. They have been emplaced at different times during the long Precambrian history and many have been recrystallized and probably remobilized during different geologic events. The result of late geologic events is to mask earlier ones and make age relationships between these various rocks obscure, but an attempt is made to review these rocks in order of age in the following descriptions.

Certain of these units, Coad Mountain augen gneiss, quartz diorite, granite at Baggot Rocks, multiple sills on Elk Hollow Creek, quartz monzonite of Elk Mountain and quartz monzonite of

Lincoln Park may or may not be igneous rocks if one defines igneous rocks as having crystallized from liquid magma. Some units (sills on Elk Hollow Creek, Coad Mountain Augen Gneiss, granite at Baggot) are classed as igneous because contacts are cross-cutting locally suggesting that the unit was mobile at some stage in its history. Others (quartz diorite, quartz monzonite of Elk Mountain, and quartz monzonite of Lincoln Park) are classed as igneous because they are more massive texturally than the quartzo-feldspathic gneiss and may contain inclusions of county rock near contacts. All of these units have been deformed and probably metamorphosed at some later stage in their history further masking their earlier history. The classification of these rocks as igneous must therefore be considered tentative. We must also emphasize that these rocks show considerable variation in composition and that they are classified on the basis of minimal petrographic study.

COAD MOUNTAIN AUGEN GNEISS

The Coad Mountain Augen Gneiss is a coarse-grained augen biotite gneiss with large porphyroblasts of microcline that have an average diameter of one-half inch and reach several inches in diameter. It crops out in a large body on the south end of Coad Mountain and in a small body on the south end of Elk Mountain in the northwest Medicine Bow Mountains (Pl. 1). At Coad Mountain the gneiss is in contact with biotite gneiss on the north, and some contacts are gradational and others cross-cutting with respect to the biotite gneiss country rock. At gradational contacts, biotite gneiss grades into augen gneiss by increase in percentage of potash feldspar which occurs as augen in the biotite gneiss. At other contacts, augen gneiss stringers extend into the biotite gneiss, is in sharp contact with biotite gneiss or is in pods and lenses in the biotite gneiss.

In texture and mineralogy (Table 20) this rock is similar to augen gneisses described as part of the quartzo-feldspathic gneiss unit. It bears a close resemblance to augen gneiss in the Barrett Ridge area. If it is related to these units it may be one of the most ancient igneous rocks of the area; perhaps metamorphosed and recrystallized more than once, or it may have simply been a potassic shale, highly metamorphosed and partially mobilized.

Table 20—Mineralogy of Coad Mountain Augen Gneiss
(mode in volume percent)

Mineral	A	%	Comments
Plagioclase	1/1	34	An_{30}($\pm$5%), reverse zoning, outer zone fresh with sharp extinction
Quartz	1/1	30	
Microcline	1/1	18	As large lenticular bodies (augen), some rimmed by plagioclase, usually entirely microcline, but some composite with quartz and plagioclase. Few grains in matrix.
Biotite	1/1	10	May be intergrown with chlorite
Epidote	1/1	3	Some allanite xtals with rims of epidote
Sphene	1/1	2	
Apatite	1/1	1	
Allanite, chlorite, zircon		—	Both allanite and zircon metmict

QUARTZ DIORITE

Reconnaissance mapping shows that there are two large bodies of quartz diorite in the northwestern part of the Medicine Bow Mountains; one makes up the major part of Bear Butte and another crops out along the western side of Pennock Mountain (Pl. 1). These intrusions cut biotite gneiss and quartzo-feldspathic gneiss and are in turn cut by granite, pegmatite, and metadiabase. The quartz diorite is for the most part strongly foliated, but there are areas especially along the southwestern part of Bear Butte where the rock is less deformed and retains much of its original character. These two bodies of quartz diorite are early intrusive rocks, perhaps the first emplaced, but temporal relationships to large gabbroic bodies are not known, and some orthoamphibolite may be earlier.

The quartz diorite body on Bear Butte grades from slightly foliated dark grey quartz diorite on the southwest side of the Butte to strongly foliated quartz diorite gneiss on the north. On the north it is cut by numeous conformable bodies of grey granite that is also strongly foliated. Some areas in the quartz diorite contain abundant dark colored inclusions of amphibolite. Conformable less-metamorphosed bodies of metadiabase are contained in the quartz diorite, but contacts between the metadiabase and quartz diorite were not observed.

The mineralogy of the quartz diorite is shown in Table 21. The samples studied were stained with sodium cobaltinitrite to test for potash feldspar, but none was noted even in more strongly foliated samples.

Table 21—Mineralogy of quartz diorite at Bear Butte
(mode in volume percent)

Mineral	A	%	Comments
Plagioclase	1/1	35	An_{32}(±5%), normal zoning
Amphibole	1/1	33	Nx = brown, Ny = green, Nz = blue-green
Quartz	1/1	29	
Epidote	1/1	1	After amphibole and as discrete crystals
Chlorite	1/1	—	After amphibole
Sphene, Opaque minerals, Apatite, Allanite	1/1	—	

The mineralogy of the quartz diorite at Pennock Mountain appears in Table 22. Samples from this body were also stained as a test for potash feldspar, but none was noted. The samples examined from Bear Butte and Pennock Mountain are similar mineralogically, but the Pennock Mountain sample contains a higher percentage of mafic minerals.

Table 22—Mineralogy of quartz diorite at Pennock Mountain
(mode in volume percent)

Mineral	A	%	Comments
Quartz	1/1	38	
Plagioclase	1/1	33	An_{30}(±5%), may be two generations
Amphibole	1/1	19	Ny = green, Nx = brown, Nz = blue-green
Biotite	1/1	8	Brown, many zircon inclusions with pleochroic haloes
Sphene	1/1	—	
Apatite	1/1	—	
Zircon	1/1	—	

The quartz diorite on Pennock Mountain is a conformable northwest trending body that is cut by aplitic granite and is covered by Quaternary deposits to the southeast (Pl. 1). It may grade into quartz monzonite locally. The quartz diorite is more massive in the central part, but along its borders it is mainly a series of multiple sills of quartz diorite that are all strongly foliated. The quartz diorite cuts quartzo-feldspathic gneiss on the northeast, but it is cut by a white aplitic granite on the northwest and southwest and contains numerous dikes and sills of the granite and associated pegmatite. The quartz diorite contains dikes and sills of metadiabase, and although contacts between these dikes and sills of metadiabase and quartz diorite were not observed, the dikes cut across strike of the foliation of the quartz diorite in many places. Inclusions of amphibolite are common and are generally oriented parallel to the foliation of the quartz diorite.

GRANITE AND QUARTZ MONZONITE IN QUARTZO-FELDSPATHIC GNEISS

There are four units of felsic rock in the quartzo-feldspathic gneiss. These are the "granite" at Baggot Rocks, the multiple sills on Elk Hollow Creek, the quartz monzonite of Lincoln Park, and the quartz monzonite on Elk Mountain.

The "granite" at Baggot Rocks crops out in a group of hills south of the confluence of the Encampment and North Platte Rivers called Baggot Rocks. It is an irregularly-shaped phacolithic body located in the axis of a north-plunging antiform in quartzo-feldspathic gneiss (Pl. 1). It conforms generally in strike to foliation of the quartzo-feldspathic gneiss, but is locally cross-cutting especially along the southwest border. It ranges from massive "granite" to faintly foliated "granite" to strongly sheared, well-foliated "granite." Contacts with quartzo-feldspathic gneiss may be sharp or gradational with sharp contacts common on the southwest and gradational contacts common on the east. The foliation of the "granite" is commonly conformable to contacts. Both the "granite" and quartzo-feldspathic gneiss are cut by variously metamorphosed mafic dikes. In the highly sheared part of the "granite," these dikes are drawn into boudins (Fig. 34), and along the west side of the "granite" the contacts between the "granite" and dikes show the classic relationship described by Sederholm (1923) in which the dikes cut the "granite," but close examination of contacts shows that the "granite" also intrudes the dikes (Pl. 16). Although we may deal with dikes of two ages or even "granite" of two ages, the less complicated explanation is that one "granite" has somehow moved after emplacement of dikes.

Table 23—Mineralogy of "granite" at Baggott Rocks
(mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Plagioclase	3/3	32	(22-42)	An[illegible] reverse zoning
Quartz	3/3	32	(26-42)	Some as graphic intergrowth with plagioclase
Microcline	3/3	25	(11-35)	Some grains show patch perthite, Large augen in one sample
Biotite	3/3	7	(5-9)	
Epidote	1/3	1	(0-3)	
Opaque Minerals	2/3	<1		Largely magnetite
Amphibole	1/3	<1		Nz = blue-green
Muscovite, Sphene Zircon	—	—		

A rock analysis (Table 24) indicates a part of the "granite" at Baggot Rocks is similar to average alkali granite except that it is slightly richer

A

Note dike cutting quartzo-feldspathic gneiss with off-shoot beginning in lower left and continuing to upper left of photo.

B

Close-up of dike showing granite stringers cutting dike. Note that granite shows both diffuse and sharp contacts with gneiss.

Plate 16—Dike cut by granite at Baggot Rocks from exposures on west side of railroad cut, sec. 16, T. 15 N., R. 83 W.

in iron oxide and calcium oxide and contains somewhat less silica. It is thus transitional to quartz monzonite as suggested by the mode. Table 23 shows that the rocks mapped as granite show considerable compositional variations.

Table 24—Chemical analysis and norm of "granite" at Baggot Rocks compared with average plutonic alkali granite (Nockolds)

Weight %	1	2		Norm
SiO_2	70.66	73.86	Ap	0.0
TiO_2	0.41	0.20	Il	0.78
Al_2O_3	13.45	13.75	Or	30.28
Fe_2O_3	1.88	0.78	Ab	29.16
FeO	1.62	1.13	An	6.05
MgO	0.73	0.26	Mt	2.73
CaO	1.67	0.72	Wo	0.93
Na_2O	3.45	3.51	Fs	0.75
K_2O	5.13	5.13	En	1.81
H_2O	1.36	0.47	Qz	26.50
P_2O_5	—	0.14		

1. Granite from southwest part of Baggot Rocks, sec. 16, T. 15 N., R. 82 W., analyst Tadashi Asari, Japan Analytical Research Institute.
2. Average alkali granite (Nockolds, 1954)

"Granitic" rocks at Baggot Rocks show such variation in texture, composition, and in field relationships to country rock that it was most difficult to decide which units should be mapped as gneiss and which units should be mapped as granite. Certainly in the southwest where the "granite" clearly cross-cuts country rock most observers would consider the unit as a true granite, but to the northeast where contacts with gneiss are gradational or where "granite" may grade into gneiss through zones of migmatite there may be considerable debate concerning the classification of the unit and placement of contacts. These problems in classification of the unit may be related to a complex history of the granitic rocks. The "granite" may have been emplaced prior to a period of shearing and prior to the emplacement of mafic dikes, and portions of the granite may have been remobilized after dike emplacement. Age determinations on this "granite" support a concept of two periods of metamorphism. Seven whole-rock samples of the "granite" define a 2400 m.y. Rb/Sr isochron and zircon separated from the "granite" gives a uranium-lead discordia solution of approximately 2400 m.y. (Hills et al., 1968. However, mineral isocrons based on apatite, plagioclase, microcline, and epidote indicate a later event of 1,500 to 1,600 m.y. These age determinations suggest a time of formation of the granite at 2.4 b.y. followed by a second event in the 1.5-1.6 b.y. range.

The multiple sills of Elk Hollow Creek are a series of sills of red granite located on the north-northwest plunging nose of an antiform in quartzo-feldspathic gneiss. The sills are located in secs. 2, 4, 5, 8, 9, and 10 of T. 16 N., R. 82 W., along Elk Hollow Creek. A large body of gabbro is found in the northern part of this area, but its relationship to the sills has not been determined. These sills have not been examined in detail because of trespass problems, but they are probably similar to the granite at Baggot Rocks. These sills are shown in the same color scheme as late granite dikes on Plate one, but as noted above are probably much older than the dikes.

The quartz monzonite of Lincoln Park is a faintly foliated body that crops out in the area northwest and southeast of Lincoln Park (Pl. 1). Isolated outcrops suggest that layers of biotite gneiss are in the unit, but no contacts with any of the metamorphic rocks were noted. This unit may simply be a more massive phase of the quartzo-feldspathic gneiss, but it was distinct enough from the gneiss in the field to be mapped as a separate unit. It is discussed with the igneous rocks but there are no good clues to the origin of the unit.

Three samples of this unit were examined in thin section and all contained highly altered feldspars making classification uncertain. Major minerals include plagioclase (probably calcic oligoclase) and quartz. Microcline was abundant in only one sample. Accessory minerals are muscovite and biotite with lesser amounts of sphene, epidote, chlorite, and zircon. The unit may range in composition from quartz monzonite to granodiorite. All samples showed evidence of cataclasis.

The greater part of Elk Mountain in the extreme northwestern part of the Medicine Bow Mountains is a pink quartz monzonite somewhat similar in appearance to the rocks at Baggot Rocks and Elk Hollow Creek. This body has not been studied in detail but according to King it cuts quartzo-feldspathic gneiss and is cut by mafic dikes.

A

Dike about three feet wide, note light colored zone at contact.

B

Close-up showing fine-grained foliated contact, coarser grain size of gneiss at contact. Pencil shows direction of strike of foliation of gneiss.

Plate 17—Small dikes cutting quartzo-feldspathic gneiss. Exposures along irrigation ditch north of French Creek, sec. 33, T. 15 N., R. 81 W.

DIKES OF MAFIC COMPOSITION

Dikes of mafic composition are ubiquituous in the area north of the Mullen Creek-Nash Fork shear zone. They cut rocks of all ages from the quartzo-feldspathic gneiss to metasedimentary rocks of the Libby Creek Group. Temporal relationships between dikes have not been well established but the oldest are probably the complexly folded generally concordant dikes in the quartzo-feldspathic gneiss.

Dikes of the Quartzo-feldspathic Gneiss

If one considers the dikes of the quartzo-feldspathic gneiss strictly from a structural aspect, they can be classed in two groups: dikes generally concordant in strike to the foliation of the gneiss and apparently folded with the gneiss, and dikes that cross-cut the structural trend of the gneiss at large angles. Although the cross-cutting dikes are younger, we cannot eliminate the possibility that some concordant dikes are the same age as those that cut the structural trend of the gneiss at large angles.

Nearly Concordant Dikes

The nearly concordant dikes are orthoamphibolite and although an igneous texture is preserved in many of them, all are altered to amphibolite, and some are so strongly deformed that the rocks have a gneissic texture and may be locally migmatized. The field relations of these dikes is regarded as significant because they might yield the only clues to timing of events in rocks such as the quartzo-feldspathic gneiss (Sederholm, 1923, Poldervaart, 1953).

Field evidence suggests that the dikes were emplaced after formation of the quartzo-feldspathic gneiss. This evidence is listed below.

1. Contacts between gneiss and many of the dikes show that the strike of the dike is not concordant to the foliation of the gneiss, although this angular discordance may be small (Pl. 17).

2. Some dikes have both discordant and concordant contacts (Fig. 34).

3. The dikes are not as highly metamorphosed as the gneiss although all rocks belong to the almandine amphibolite facies of regional metamorphism, the dikes commonly have primary structures preserved, whereas none have been recognized in the gneiss.

4. In many instances dikes have fine-grained borders (Pl. 17) suggesting chilled contacts. The border zones are foliated, however, and could result from differential movement during metamorphism.

5. Misoriented inclusions have been noted along borders of a few dikes and indicate forceful emplacement of dikes in pre-existing gneiss or in corporations of pre-existing gneiss in the dikes during differential movement.

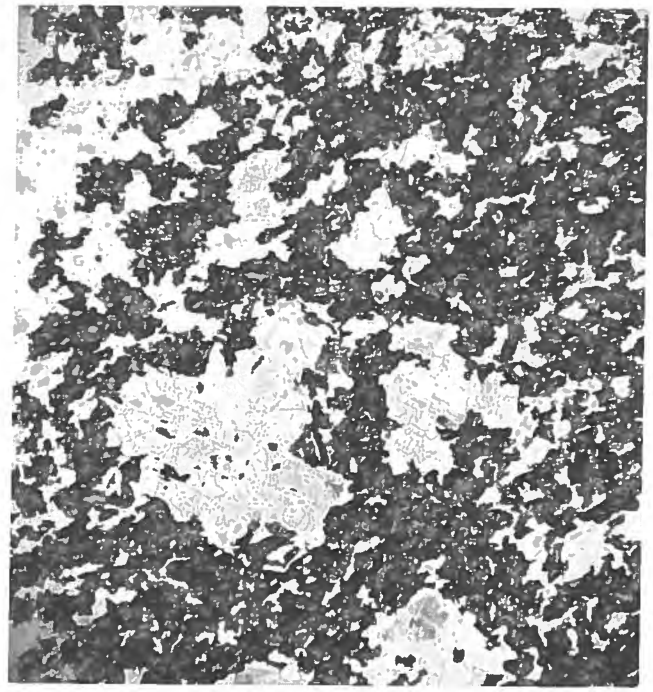
6. Gneiss at the contact of some dikes is coarser-grained, less well-foliated, and lighter in color (Pl. 17) which suggests contact metamorphism of pre-existing gneiss. Although some of this evidence may be subject to more than one interpretation, it suggests that the quartzo-feldspathic gneiss formed during an early metamorphic event, basaltic magma was introduced in fractures, and the gneiss and contained dikes were deformed in a later event.

In the above discussion, emphasis is placed on the fact that there is considerable variation in the degree of deformation and recrystallization of individual dikes. Some, for example, preserve most of their original texture and mineralogy and others are completely recrystallized and even drawn into boudins (Fig. 34). This variation in character of the dikes is not a systematic one from one part of the area to the other, but the more severely deformed dikes are in zones of more intense deformation that may be found in any part of the area. The field relationships that suggest a late origin for the dikes are found in the less deformed and metamorphosed areas. Sutton and Watson (1951, p. 241-295) have described dikes of the Loch Torridon and Scourie area of Scotland that have a history similar to that proposed for the dikes in the quartzo-feldspathic gneiss. The significance of the dikes regarding structural history will be discussed later.

The petrography of the dikes reflects the degree of deformation and metamorphism. For example, dikes on the west side of Baggot Rocks, T. 15 N., R. 83 W., are not highly deformed and are foliated only along the borders whereas those on the east side are more deformed and largely altered to amphibolite. The core of the dike on the west side is unaltered. It is a medium-grained gabbro consisting of labradorite (37), hypersthene (19), augite (28), with minor amounts of hornblende, quartz, biotite, spinel, and ilmenite

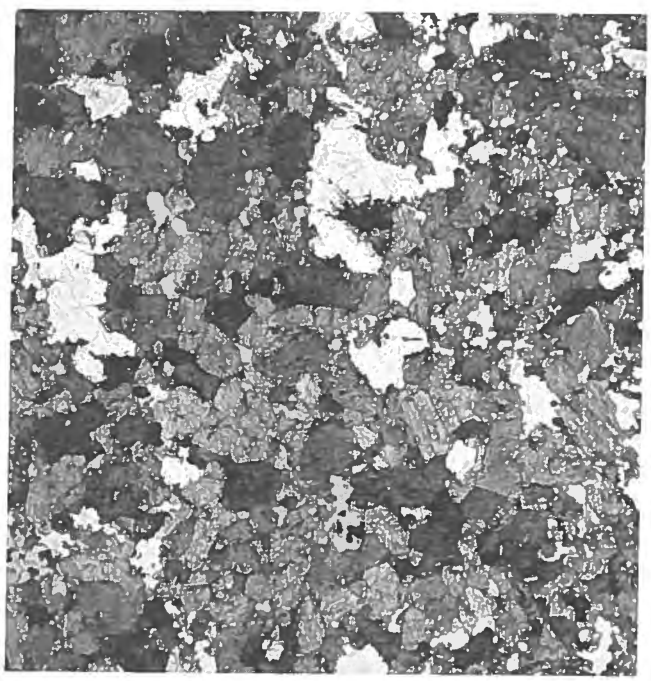
A

Relatively unaltered igneous rock showing fresh pyroxenes and plagioclase. Texture preserved. Line 2 mm. long, all photos same scale.



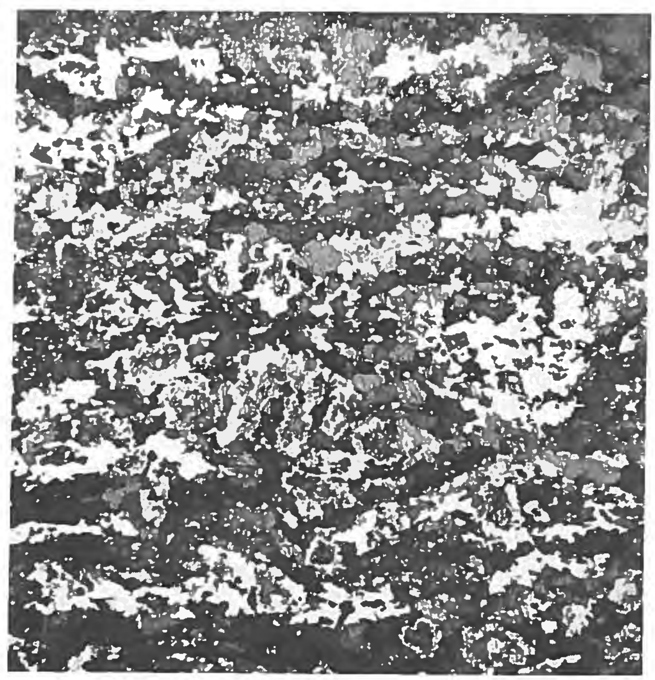
B

Pyroxene altered to fine-grained aggregates of amphibole and quartz. Plagioclase altered but still not recrystallized. Texture largely preserved.



C

Rock altered to aggregate of amphibole, plagioclase, quartz, and sphene. All minerals metamorphic, texture destroyed.



D

Development of distinct layers of tectonic origin. Amphibole-rich layers alternate with quartz-feldspar rich layers. All minerals metamorphic.

Plate 18—Photomicrographs showing transition from relatively unaltered igneous rock to layered amphibolite in dikes of the Baggot Rocks area. Uncrossed nicols.

(Table 25). Hornblende develops as a reaction between plagioclase and pyroxene (Pl. 18). In other areas, these less altered dikes which are diabase retain typical diabasic texture and are composed largely of augite and labradorite. The more altered dike on the east side of Baggot Rocks is an amphibolite containing dark green to brown amphibole (46), andesine (26), quartz (16) with minor amounts of ilmenite and spinel (Table 25). The amphibole is in aggregates of small grains. Although the rocks seem largely recrystallized there are altered areas of high relief in the masses of amphibole that may be pyroxene remnants and a few larger crystals of plagioclase have clouded cores that may be remnants of the original plagioclase (Pl. 18). Other rocks examined under the microscope are amphibolites containing from 56 to 62% amphibole, 20 to 27% andesine, 8 to 13% quartz, and minor ilmenite, sphene, apatite, garnet, epidote and spinel. Ilmenite may be altered to wormy masses of opaque minerals (leucoxene) and sphene. The rock is completely recrystallized and has poorly developed layering in places.

Table 25—Mineralogy of dikes of mafic composition (mode in volume percent)

Mineral	A	$\bar{X}$	Comments
Nearly concordant gabbro dike, not highly metamorphosed, quartzo-feldspathic gneiss, west side of Baggot Rocks			
Plagioclase	1/1	37	An[illegible], normal zoning, amphibole rims at pyroxene contacts
Clinopyroxene	1/1	28	Pigeonite augite, 2 V = 45° (est.)
Orthopyroxene	1/1	19	Hypersthene
Amphibole	1/1	6	After pyroxene, as reaction rims between plagioclase and pyroxene, pale green
Opaque minerals	1/1	4	Ilmenite, partly
Quartz	1/1	3	
Biotite	1/1	tr.	Red-brown
Spinel	1/1	tr.	Red
Nearly concordant amphibolite, foliated, local preservation of igneous texture, quartzo-feldspathic gneiss, east side of Baggot Rocks			
Amphibole	1/1	46	In aggregates after pyroxene, dark green to brown
Plagioclase	1/1	26	An[illegible]±, clear probably recrystallized
Quartz	1/1	16	
Opaque minerals	1/1	6	Ilmenite, in part
Pyroxene	1/1	5	Remnants in masses of amphibole
Spinel	1/1	1	
Nearly concordant amphibolite, foliated, no preservation of igneous texture, quartzo-feldspathic gneiss, Beaver Hills			
Amphibole	2/2	59	2 V = 65 (est.), Ext. 30
Plagioclase	2/2	20	An_{28}
Quartz	2/2	14	
Opaque minerals	1/2	3	Ilmenite, wormy intergrowth with sphene
Garnet	1/2	2	
Sphene	2/2	1+	
Epidote	1/2	1	
Apatite	1/2	1	
Spinel	1/2	tr.	Blood red

Dikes That Cross-Cut Gneiss at Large Angles

The second group of dikes in the quartzo-feldspathic gneiss according to the structural classification proposed above cross-cuts the structure of the quartzo-feldspathic gneiss at large angles. These dikes appear less metamorphosed than most of the conformable dikes but do not differ significantly in composition. All that were examined petrographically are partially altered to amphibolite. The rocks retain a diabasic texture and pyroxene outlines may be preserved but under the microscope all were visibly altered, in part, to bluish green amphibole. The original plagioclase (labradorite) is retained although in some specimens it is clouded and in others it is saussuritized. In some areas in the dikes and in some entire dikes, the rock is completely recrystallized and altered to orthoamphibolite of the almandine-amphibolite facies.

Mafic dikes cutting the metasedimentary rocks

Mafic dikes are rare in the metasedimentary rocks of the Deep Lake Formation, but are common in rocks of the Libby Creek Group. Mafic rocks that cut the Deep Lake Formation are chiefly large bodies of norite or gabbro, which will be reviewed later, but a few small sills about 50 feet in thickness crop out along the Medicine Bow River south of Stillwater Park and some highly altered orthoamphibolite in hornblende gneiss is found near Arlington (Table 28). The sills in outcrops along the medicine Bow River are orthoamphibolite with the diabasic texture preserved very similarly to those described above. The orthoamphibolite in hornblende gneiss and also in associated granite gneiss is strongly foliated and completely recrystallized and consists chiefly of blue-green amphibole, andesine, and quartz with minor amounts of epidote, sphene, and garnet. One of these bodies contains large stretched masses that appear to be altered plagioclase phenocrysts. Some large unaltered plagioclase grains remain, but most grains are altered to masses of quartz, secondary plagioclase, clinozoisite, and carbonate.

Dikes cutting the rocks of the Libby Creek Group can be divided into two groups: (1) those that cut quartzite, (2) those that cut carbonate-rich rocks. The dikes cutting quartzite are best exposed in the vicinity of Medicine Bow Peak where there are a number of dikes in joint planes and faults in the Medicine Peak Quartzite. These dikes are probably metadiabase and although all examined are altered to some degree the texture is preserved. A sample from a road cut along Highway 130, sec. 25, T. 16 N., R. 80 W. contains an unidentifiable pyroxene, altered to masses of amphibole, quartz, minor chlorite and biotite; labradorite, unaltered except for scattered crystals of epidote, and opaque minerals largely altered to sphene. The dikes in carbonate-rich rocks are found within various units of the Nash Fork Formation. Some of these are highly altered green rocks which may be conformable or cross-cutting with respect to the bedding of the units of the Nash Fork Formation. Most of these dikes are probably mafic in composition, but some seem to have more felsic borders, and others may be intermediate in composition. A typical sample of these green dikes contains amphibole (36%), epidote (16%), albite (15%), chlorite (12%), quartz (10%), sphene (7%), and minor amounts of biotite and carbonate. Along contacts with metadolomite the dikes have assimilated carbonate and may have large crystals of pyrite and garnet and irregular-shaped masses and rosettes of tremolite and actinolite. The original composition of these dikes is not known as their composition has obviously been modified by assimilation of carbonate. They now contain minerals typical of the green-schist facies of regional metamorphism.

LARGER INTRUSIVE BODIES OF GABBROIC COMPOSITION

Coarse-grained intrusive bodies of gabbroic composition are common in both the quartzo-feldspathic gneiss and the metasedimentary rocks. These range from small plugs to large intrusive bodies several miles in diameter. As in the case of the mafic dikes the intrusions in the quartzo-feldspathic gneiss will be reviewed first.

Gabbroic Intrusives of the Quartzo-feldspathic Gneiss

A series of small semicircular bodies of gabbro in T. 15 N., R. 81 W., crop out on the east side of Baggot Rocks and in isolated intrusives elsewhere in the area. These bodies are recrystallized and foliated along contacts with quartzo-feldspathic gneiss, but only slightly altered in the central portion. They range from coarse-grained diabase consisting primarily of augite and labradorite with accessory ilmenite to norites consisting chiefly of orthopyroxene and labradorite (Table 26). Rims of amphibole are developed around pyroxene and in some cases around grains of ilmenite. In some samples, pyroxene is partially replaced by amphibole and epidote. Quartz and garnet are present in more highly altered samples.

Table 26—Mineralogy of small mafic intrusive bodies (mode in volume percent)

Mineral	A	$\bar{X}$	Comments
			Medium-grain gabbro, east side of Baggot Rocks
Plagioclase	1/1	47	An[illegible], normal zoning
Clinopyroxene	1/1	38	2 V = 45 (est.), pigeonitic augite
Amphibole	1/1	8	After pyroxene and as rims around ilmenite
Opaque minerals	1/1	6	Ilmenite
Garnet	1/1	tr.	
			Norite, Corral Creek area
Plagioclase	2/2	48	An[illegible], relatively unaltered
Pyroxene	2/2	44	Generally partly altered to blue-green amphibole, but largely hypersthene
Quartz	2/2	4	
Opaque minerals	2/2	2	

There are two larger mafic intrusions in the quartzo-feldspathic gneiss which will be referred to informally as the Beaver Hills gabbro and the Bennett Peak gabbro.

The Beaver Hills gabbro, which is a partly conformable and partly cross-cutting mafic intrusion in the south-central part of T. 15 N., R. 82 W., and the north-central part of T. 14 N., R. 82 W., is a hook-shaped body approximately two and one-half miles long and 3/4 mile wide at the southern end (Pl. 1). It cuts the quartzo-feldspathic gneiss, but is, in turn, cut by pegmatite. It is transitional from coarse-grained gabbro to slightly altered medium-grained highly sheared amphibolite. The least altered gabbro consists of labradorite and pyroxene with accessory red-brown biotite and ilmenite. The biotite is commonly associated with ilmenite and may occur as rims around it. Reaction rims between pyroxene and plagioclase and ilmenite and plagioclase are well developed. A typical reaction rim consists of green amphibole in contact with pyroxene and garnet in contact with plagioclase although the garnet layer is not always present. The pyroxenes are all altered to some degree, but both orthopyroxene and clinopyroxene are believed to be present. There are some areas in the rocks that may have been olivine originally that consist of aggregates of opaque minerals and garnet and have reaction rims with plagioclase. In more altered gabbro, the pyroxene is completely converted to aggregates of green amphibole and minor amounts of quartz, but plagioclase remains unaltered.

The Bennett Peak gabbro is a large irregular-shaped, cross-cutting intrusion located north of Bennett Peak, in T. 15 N., R. 82 W. (Pl. 1). It is a massive slightly altered gabbro with areas

A B

Figure 13—Photomicrographs of olivine gabbro from Bennett Peak. A. Olivine gabbro showing well-developed reaction rims around olivine where it is in contact with plagioclase, uncrossed nicol. Line 2 mm. long. B. Enlargement from same slide showing reaction rims between olivine and plagioclase but rims lacking at contact of orthopyroxene (black), crossed nicols. Line 1 mm. long.

especially near the contacts with quartzo-feldspathic gneiss that are altered to amphibolite but still retain the original texture. Dike-like offshoots of the gabbro are somewhat finer-grained than central portions of the gabbro. Along its western border the gabbro is cut by dikes of diabase, and along its eastern border it is cut by a series of simple pegmatites. The Bennett Peak gabbro is similar to the Beaver Hills gabbro since it is composed chiefly of pyroxene and labradorite with accessory red-brown biotite and ilmenite (Table 27). It differs in that it is less altered, both augite and hypersthene are present, and the rock contains olivine locally. Reaction rims similar to those described above are also present, but are better developed. Rims around olivine consist of orthopyroxene in contact with the olivine which grade to amphibole and garnet.

The reaction rims are well-developed where olivine is in contact with plagioclase, but may be absent where olivine is in contact with orthopyroxene (Fig. 13). The rims are probably synantectic developed by reaction between minerals, i.e. plagioclase-olivine, (Sederholm, 1916) during regional metamorphism.

Table 27—Mineralogy of Bennett Peak gabbro (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Plagioclase	3/3	46	(28-66)	An [illegible] reaction rims with olivine, in sequence olivine, pyroxene, amphibole, plagioclase
Pyroxene	3/3	40	(34-58)	Orthopyroxene and clinopyroxene, hypersthene and augite, orthopyroxene, in part, as rims around olivine
Amphibole	3/3	6	(4-7)	As green uralite after pyroxene, as rims around opaque mineral, olivine, pyroxene
Opaque minerals	3/3	4	(2-7)	Reaction rim with plagioclase, outer rim. Discrete grains
Olivine	1/3	3	(0-9)	Most grains have reaction rims, as above. Some grains completely altered to pyroxene leaving rim without core
Biotite	2/3	1	(0-2)	Red brown, mostly as rims around opaque minerals
Garnet	1/3			As part of reaction rim in some cases in sequence plagioclase, garnet, amphibole, pyroxene
Micropegmatite	1/3			Intergrowth of quartz K feldspar one grain noted in thin section

A conformable body of highly altered mafic igneous rocks approximately three miles long crops out near the crest of Barrett Ridge in the southwestern part of T. 16 N., R. 81 W., and the northeastern part of T. 15 N., R. 81 W. This rock is composed of a mixture of talc and amphibole with minor quartz probably derived from pyroxene. It may be a highly metamorphosed ultramafic igneous rock.

Gabbroic Intrusions of the Metasedimentary Rocks

A series of large coarse-grained mafic intrusive bodies cuts both the rocks of the Deep Lake Formation and rocks of the Libby Creek Group. These igneous rocks conform in a general way to the structure of the enclosing rocks, but are commonly cross-cutting in detail. Offshoots of some larger intrusive bodies extend into faults adjacent to them or may cut through metasedimentary rocks that do not show obvious offset (Pl. 1). Most of these larger mafic intrusives were introduced after a period of folding and faulting of the metasedimentary rocks. On a regional basis, the intrusive bodies in the central part of the metasedimentary complex are the least altered and deformed whereas similar intrusive rocks are more highly altered and approach true amphibolites to the northeast and southeast.

Mafic intrusive rocks of the Deep Lake Formation will be discussed first. Because on a regional basis most of these larger bodies are conformable in strike and dip to that of adjacent metasedimentary rocks, they will be called sills. For convenience in discussion, the mafic intrusions will be given informal names and divided into six groups: the Gold Hill sill, the Brush Creek sill, the Campbell Lake sills, the Windy Hill sills, the Carlson Creek sills, and the Arlington sills.

The Gold Hill sill is in the upper part of the Deep Lake Formation and is exposed from the vicinity of the confluence of Arrastre Creek and South Brush Creek to an area south of Deep Lake, a distance of seven and one-half miles (Pl. 1). Another sill that may be in the same stratigraphic position is exposed along the head of Trail Creek and again in Rock Creek Canyon, sec. 6, T. 17 N., R. 78 W. The least metamorphosed portion of the sill is in the area east of Arrastre Lake where the rock is norite consisting of labradorite, hypersthene, and augite with accessory red-brown biotite and ilmenite (Table 28). The pyroxene is altered slightly to a mixture of chlorite and amphibole. Rocks cropping out near the confluence of Arrastre Creek and South Brush Creek and along Rock Creek Canyon are altered to amphibolite but retain the original texture. Samples of these rocks indicate that the pyroxene is altered to aggregates of blue-green amphibole, and plagioclase is altered to aggregates of clinozoisite, quartz, and biotite.

The Brush Creek sill is located northwest of the confluence of Little Brush Creek and South Brush Creek (Pl. 1). It is composed of labradorite, augite and hypersthene with accessory red-brown biotite and ilmenite much like the Gold Hill sill, but augite is more abundant so the rock is classed as gabbro (Table 28). Sections were made of samples from the base of the sill and the top to see if there was a variation in rock type, but none was noted.

The Campbell Lake sills are located east and north of Campbell Lake in the southern part of T. 17 N., R. 80 W. (Pl. 1). These sills are moderately altered quartz gabbro. The rock contains augite, normally zoned labradorite, and quartz with accessory red-brown biotite and opaque minerals (Table 28). Amphibole is present as reaction rims between plagioclase and augite.

A series of sills called the Windy Hill sills is located northwest and south of Windy Hill and north of the Medicine Bow River near its head at Deep Lake (Pl. 1). All are quartz gabbros similar to the intrusive bodies of Campbell Lake (Table 28). The sills located near Windy Hill and on the Medicine Bow River near Deep Lake are the least altered, and that body located northwest of Windy Hill is more highly altered but still retains its igneous texture.

The Carlson Creek sills are a series of conformable mafic bodies located both west and east of Carlson Creek near the point where the creek crosses the road to Arlington (Pl. 1). These mafic intrusions crop out across Rock Creek and may be found as for north as sec. 6, T. 18 N., R. 78 W. The sills are emplaced in metabasaltic rocks of the Deep Lake Formation and in areas where the units have been converted to amphibolite it is difficult to tell the intrusive rocks from the metabasalt. All of the rocks examined were altered to some degree, but the least altered rocks were in the vicinity of Carlson Creek where the major rock type was norite (Table 28). To the northeast across Rock Creek all of these units are altered to amphibolite.

A series of sills southwest of Arlington (Pl. 1) called the Arlington sills are all altered to amphibolite. Some, especially the larger bodies in the southcentral part of T. 19 N., R. 79 W., retain a suggestion of an igneous texture but all studied are completely recrystallized and consist chiefly of blue-green amphibole, andesine, and quartz (Table 28).

Table 28—Mineralogy of mafic sills in the Deep Lake Formation (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Gold Hill Sill (least altered sample)				
Plagioclase	1/1	56	—	An_{54}
Pyroxene	1/1	33	—	Hypersthene and augite, altered in part, to amphibole
Biotite	1/1	4	—	
Opaque Minerals	1/1	2	—	
Amphibole	1/1	1	—	Blue-green, after pyroxene
Chlorite	1/1	2	—	Rims some pyroxene crystals
Quartz	1/1	2	—	Secondary (?)
Brush Creek Sill				
Plagioclase	2/2	43	—	An[illegible], some grains clouded
Pyroxene	2/2	50	—	Hypersthene and augite, partly altered to amphibole
Biotite	2/2	3	—	Red brown, as rims around opaque minerals
Opaque Minerals	2/2	2	—	
Amphibole	1/2	1	—	After pyroxene
Clinozoisite	1/2	1	—	Very fine-grained, at contacts between plagioclase and pyroxene
Campbell Lake Sills				
Plagioclase	2/2	40	—	An[illegible], normal zoning
Pyroxene	2/2	44	—	2 V = 40 (est.), pigeonitic augite
Quartz	2/2	7	—	Primary
Amphibole	2/2	4	—	As rims between pyroxene and plagioclase, and alteration of pyroxene
Biotite	2/2	3	—	Red brown
Opaque Minerals	2/2	2	—	
Windy Hill Sills				
Plagioclase	4/4	35	(26-44)	An[illegible]. Some grains altered to clinozoisite
Pyroxene	2/4	22	(0-45)	2V = 40 (est.), pigeonitic augite, herringbone structure
Amphibole	4/4	29	(4-57)	Primarily as alteration of pyroxene, but in some less-altered rocks as rims between plagioclase and pyroxene
Quartz	4/4	7	(4-10)	Primary, interstitial
Opaque Minerals	4/4	4	(2-5)	Ilmenite and some grains of pyrrhotite (?)
Clinozoisite	2/4	2	(0-6)	After plagioclase
Biotite	2/4	1	(0-3)	Red-brown
Chlorite	1/4	—	—	After pyroxene
Apatite	1/4	—	—	
Potash Feldspar	1/4	—	—	Microcline, interstitial, primary
Carlson Creek Sills				
Plagioclase	4/5	22	(0-42)	An[illegible], normal zoning, some grains altered to carbonate and chlorite
Pyroxene	2/5	20	(0-56)	In less-altered rocks, both augite and hypersthene can be identified. In some rocks pyroxene completely converted to amphibole and chlorite
Amphibole	4/5	30	(0-71)	Blue-green, fibrous, after pyroxene
Quartz	2/5	3	(0-8)	Secondary quartz, in highly altered rocks only
Opaque Minerals	5/5	4	(1-7)	
Carbonate	2/5	13	(0-58)	In highly altered rocks after plagioclase mostly
Chlorite	2/5	4	—	After pyroxene in altered rocks
Sphene	2/5	tr.	—	As discrete crystals and rims around opaque minerals
Biotite	1/5	tr.	—	
Clinozoisite	1/5	2	—	After plagioclase
Sills Southwest of Arlington (after J. S. King)				
Amphibole	10/10	65.7	(48-79)	Blue-green, fibrous
Plagioclase	10/10	10.9	(2-18)	An[illegible]
Epidote	10/10	12.9	(2-27)	Assocated with plagioclase (clinozoisite molecule 75-90)
Quartz	10/10	7.8	(4-15)	
Sphene	9/10	0.8	(0-2)	
Chlorite	6/10	0.7	(0-4)	
Opaque Minerals	8/10	0.7	(0-2)	Hematite, in part
Biotite	5/10	0.5	(0-4)	

There are several bodies of coarse-grained mafic igneous rock in the metasedimentary rocks of the Libby Creek Group and numerous smaller bodies especially in the Heart Formation at the southwest extremity of the outcrop of the metasedimentary rocks. The larger intrusions are all located along French Creek and its tributaries (Pl. 1). These mafic intrusions have not been examined in detail. In the north, mafic bodies in the Sugarloaf quartzite, sec. 25, T. 16 N., R. 80 W. are largely metagabbro. South of these units mafic bodies in T. 15 N., R. 80 W., are largely converted to amphibolite.

There are many sills in the Heart Formation near its contact with quartzo-feldspathic gneiss, and all are altered to amphibolite. The most highly deformed and metamorphosed sills are found along the southern extremity of outcrops of the Heart Formation where the sills are distinctly gneissic in texture.

NOTES ON CHEMISTRY AND PETROLOGY OF MAFIC IGNEOUS ROCKS NORTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

Chemical analyses have been made of three samples of mafic igneous rock from locations north of the shear zone. These were made for comparison with amphibolite south of the shear zone, but the results are of interest because they give us some clues regarding magma types. Analysis 2, Table 29, is of an orthoamphibolite that cuts quartzo-feldspathic gneiss in the area southwest of the Beaver Hills, sec. 3, T. 13 N., R. 82 W. Relict or primary minerals have not been recognized in this rock, but some samples have a suggestion of a diabasic texture. The norm of this rock is similar to the mode (Table 25) of a dike that cuts the quartzo-feldspathic gneiss in the Baggot Rocks area. The dike at Baggot Rocks is the least metamorphosed of the dikes studied from this area, and the only one where a reasonable estimate of primary minerals could be made. Analysis 1, Table 29, is of a metadiabase cutting Medicine Peak Quartzite in sec. 27, T. 16 N., R. 80 W., and analysis 3, Table 29, is of metagabbro from the Gold Hill sill (Table 28). These two rocks could be younger than the dike cutting quartzo-feldspathic gneiss but do not show striking differences chemically. The norm for the metagabbro shows a relatively high percentage of quartz that did not appear in the mode (Table 28). Most metagabbro sills from this area are quartz gabbros, however (Table 28).

Table 29—Chemical analyses and norms of metaigneous rocks and orthoamphibolite, north of Mullen Creek-Nash Fork shear zone. Analyses by Tadashi Asair, Japan Analytical Chemistry Research Institute

	1	2	3		1	2	3
SiO_2	48.91	50.20	51.73	qu	2.09	2.46	14.73
TiO_2	1.95	0.76	0.73	or	9.50	2.31	2.29
Al_2O_3	14.51	12.24	15.64	ab	23.08	11.71	6.69
Fe_2O_3	4.39	2.61	2.55	an	22.51	25.90	32.57
FeO	8.81	7.84	7.98	di	12.07	27.23	19.33
MnO	0.14	0.16	0.15	hy	16.20	22.30	18.13
CaO	7.76	12.04	11.26	mt	6.35	3.80	3.68
MgO	5.12	9.82	6.51	il	3.71	1.40	1.39
Na_2O	2.74	1.39	2.07	ap	0.36	—	—
K_2O	1.61	0.40	0.39	H_2O	3.28	1.94	1.30
P_2O_5	0.16	0.06	0.06	CO_2	0.47	0.99	0.01
H_2O+	2.73	1.56	0.77				
H_2O-	0.55	0.38	0.53				
CO_2	0.47	0.99	0.01				

1 Diabase dike cutting Medicine Peak Quartzite, sec. 27, T. 16 N., R. 80 W.

2 Orthoamphibolite (diabase texture locally preserved), sec. 3, T. 13 N., R. 82 W.

3 Gabbro from Gold Hill sill, sec. 15, T. 16 N., R. 80 W.

In Table 30 a comparison is made of the average composition of these mafic dikes with esimates of parent magma of dikes of the Beartooth Mountains, of Wyoming and Montana, and with average tholeiite and average alkaline olivine basalt. The rocks show a strong similarity to theolitic type basalt and Beartooth Rocks chemically. This is compatible with the mineralogy of the rocks because most are two pyroxene bearing and olivine free. The presence of significant amounts of quartz in the norm as well as appreciable hypersthene is further support for a classification as tholeiite.

Table 30—Comparison of chemical composition of mafic rocks north of the shear zone with dikes of the Beartooth Mountains, Wyoming and Montana, and average tholeiite and Alkaline olivine basalt. (calculated water-free, to 100 percent by weight)

	1	2	3	4	5
SiO_2	52.1	51.0	51.0	51.8	46.1
TiO_2	1.2	1.4	0.9	0.7	2.6
Al_2O_3	14.6	15.6	15.3	15.5	14.8
Fe_2O_3	3.9	1.1	2.2	1.9	3.2
FeO	8.5	9.8	8.9	9.3	8.8
MnO	0.2	0.2	0.2	0.2	0.2
MgO	7.4	7.0	8.6	7.4	9.4
CaO	9.7	10.5	9.7	9.6	10.8
Na_2O	2.2	2.2	2.4	2.4	2.7
K_2O	0.8	1.0	0.7	1.1	1.0
P_2O_5	0.1	0.2	0.1	0.1	0.4

1 Average composition of 3 mafic intrusions north of Mullen Creek-Nash Fork Shear zone, Medicine Bow Mountains, Wyoming.

2 Average tholeiite (Green and Poldervaart, 1955, p. 185 in Prinz, 1964, table 8)

3 Parent magma, average of 6 analyses. Archean Metadolerites, Beartooth Mountains, Wyoming and Montana, (Prinz, 1964, Table 8.)

4 Parent magma, late Precambrian dolerites, Beartooth Mountains, Wyoming and Montana, (Prinz, 1964, Table 8)

5 Average alkaline olivine basalt (Nockolds, 1954, p. 102, in Prinz, 1964, table 8)

"GRANITIC" BODIES OF THE METASEDIMENTARY ROCKS

In the northeastern Medicine Bow Mountains near Arlington a body of granite gneiss is in contact with hornblende gneiss on the east and quartzite of the Deep Lake Formation on the west (Pl. 1). The gneiss appears to be emplaced in the axis of a fold in the metasedimentary rocks, but this is uncertain because of overlap by rocks of Paleozoic age to the northwest. The granite gneiss grades into hornblende gneiss on the east and is in sharp contact with quartzite on the west. It shows rather wide compositional variations (Table 31), but it is essentially a muscovite granite gneiss. At some localities, a dark biotite-rich, finely foliated gneiss occurs interlayered with granite gneiss. In contrast to most felsic rocks of the area the plagioclase is chiefly albite, but shows considerable variation in composition from An_5 to An_{20}. Microcline is found in interstitial aggregates and as large porphyroblasts. This gneiss may or may not be igneous in origin. It may be part of the older basement upon which sandstone (now quartzite) of the Deep Lake Formation was deposited. A tentative minimum age assignment of 2.26 ± .08 b.y. by Hills et. al. (1968) supports an assignment of the gneiss to basement.

Table 31—Mineralogy of granite gneiss at Arlington (after J. S. King, in part) (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	10 10	38.8	25-51	
Plagioclase	10 10	36.9	22-71	An_{5-20} in large grains, sieved with inclusions, and as interstitial aggregates
K-Feldspar	7 10	13.7	0-38	In large grains and interstitial aggregates, some grains perthitic, myrmekite in associated plagioclase
Muscovite	10 10	4.9	1-12	Some grains oriented perpendicular to foliation
Biotite	5 10	3.3	0-13	
Epidote	4 10	0.6	0-3	
Carbonate	1 10	0.6	0-6	
Opaque minerals	7 10	0.2	0-1	Magnetite, in part, altered to hematite
Chlorite	2 10	0.2	0-1	
Garnet, sphene, apatite	—	—	—	

Gaps "Granite"

Two small bodies of felsic igneous rock cut metasedimentary rocks of the Libby Creek Group. One is located in sec. 8, T. 16 N., R. 79 W., the other in sec. 17, T. 15 N., R. 80 W., both are introduced along faults.

The northern most felsic body is a heart-shaped intrusive emplaced in a fault zone (Pl. 1). It is here named the Gaps "Granite" for Gap Lake just north of the granite outcrop. It is in contact with the Medicine Peak Quartzite on the east and

Lookout Schist on the west. Contacts with quartzite and schist are clearly cross-cutting. Numerous narrow stringers cut through brecciated country rock, and the granite contains abundant misoriented inclusions of quartzite along its border. The Gaps "Granite" has been sheared after emplacement especially in zones along the original fault plane. In these zones of cataclasis, there are misoriented pieces of granite.

All samples of this rock studied in thin section were highly altered. Essential minerals are quartz and plagioclase and some completely sericitized K-feldspar may be present. Muscovite makes up about 15 percent of the rock and accessory minerals are opaque minerals and apatite.

SMALL FELSIC BODIES AND PEGMATITE OF THE QUARTZO-FELDSPATHIC GNEISS

Small bodies ranging in width from a few inches to a few feet of granite, aplite, or pegmatite are common in the quartzo-feldspathic gneiss. Most aplite forms in dikes or sills and appears to fill fractures in the gneiss (Figs. 14 and 15). The granite and pegmatite has varied occurrences from lenticular masses or irregular masses with diffuse borders (Fig. 33) to networks of cross-cutting veinlets in the gneiss (Fig. 33). A typical example of the relationship of one of these small bodies of granite to hornblende gneiss and quartzo-feldspathic gneiss appears in Figure 16. The quartzo-feldspathic gneiss shows replacement like contacts with the hornblende gneiss and contains numerous conformable stringers and remnants of hornblende gneiss. One stringer of quartzo-feldspathic gneiss cuts the hornblende gneiss. The quartzo-feldspathic gneiss may be metasomatic and formed by replacement of hornblende gneiss. The stringers of quartzo-feldspathic gneiss cutting the hornblende gneiss suggest mobilization. The granite is emplaced in the axis of the fold and cross-cuts all rock types. One might note the close parallel in general habit of this small body of granite to the larger mass of granite at Baggot Rocks.

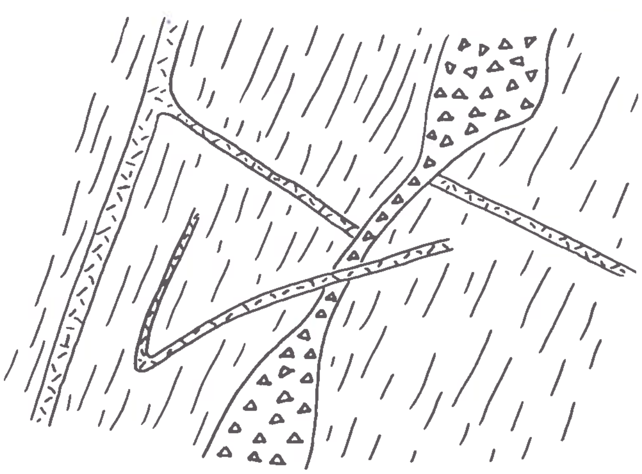


Figure 14—Quartzo-feldspathic gneiss (lined) cut by aplite (stippled) and pegmatite (triangle). Contacts are sharp except that aplite fades into gneiss along strike. Maximum width of pegmatite about 3 feet. From field sketch on west side of Platte River, sec. 26, T. 15 N., R. 82 W.

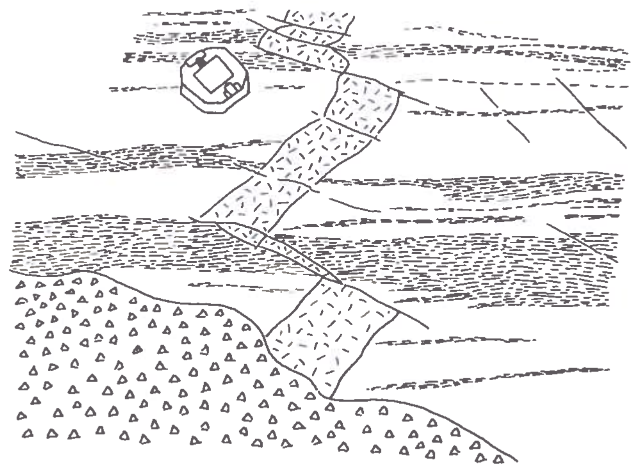


Figure 15—Quartzo-feldspathic gneiss (white) with interlayered biotite gneiss (lined), cut by aplite (stippled) and pegmatite (triangle). From photograph of outcrop in sec. 6, T. 15 N., R. 82 W.

The relationship between pegmatite and aplite is illustrated in figures 14 and 15. These figures show the cross-cutting relationships that the aplite and pegmatite have with respect to the foliation and compositional layering of the quartzo-feldspathic gneiss and also show that aplite may be either older or younger than pegmatite. These conflicting time relationships may indicate formation of these bodies at about the same time.

These small felsic bodies in the gneiss are found in most areas studied, but are rarely concentrated enough to be referred to as migmatite. They are most abundant in zones subject to cataclasis such as on the east side of Baggot Rocks. They apparently formed during metamorphism and metasomatism. Since most of the bodies are small unconnected pods, lenses and stringers, they may simply be mobilized portions of the gneiss.

DISCRETE BODIES OF PEGMATITE

There are many discrete bodies of pegmatite in the gneiss either as lenticular conformable bodies or cross-cutting masses. They are larger

Figure 16—Field sketch of hornblende gneiss (lined), quartzo-feldspathic gneiss (white) and granite (stippled). Contacts between hornblende gneiss and quartzo-feldspathic gneiss are gradational, and small stringers of hornblende gneiss are in quartzo-feldspathic gneiss. Granite contacts are sharp, but grade into quartzo-feldspathic gneiss as shown in lower part of sketch. Outcrop about three feet wide. Sketch from isolated exposure a few hundred feet south of road in sec. 19, T. 15 N., R. 82 W.

than the units described above, some as much as 10 feet wide and several hundred feet long. They are abundant in the axes of antiforms such as in the Encampment district (Fig. 41) and the Baggot Rocks area (Pl. 1). They may be concentrated in zones of penetrative movement (Fig. 33). These pegmatites conform generally to the strike of the foliation of the gneiss but may cross-cut both the structure of the gneiss and the contained ortho-amphibolite. Some areas in the gneiss have so many pegmatites that they can be referred to as pegmatized areas. These pegmatized areas may contain as much as 50% pegmatite, and are shown on Plate One by a concentration of red dots.

All pegmatites noted in the gneiss are of the simple unzoned type and consist of quartz, microcline, and plagioclase. The microcline is generally pink and the plagioclase is white. Microcline and quartz normally are much more abundant than plagioclase. Some pegmatites may contain small amounts of biotite and muscovite, and pegmatites in the Encampment district, southeastern part of T. 14 N., and northeastern part of T. 13 N., R. 82 W., contain crystals of kyanite, sillimanite, and corundum (Hagner, 1944, p. 19), but this variation in mineralogy is the exception rather than the rule.

Pegmatites are also present in the large gabbroic intrusions, the Beaver Hills gabbro and the Bennett Peak gabbro. Those pegmatites that cut the Bennett Peak gabbro must be post-kinematic because this intrusion is virtually undeformed.

PEGMATITE OF THE METASEDIMENTARY ROCKS

Pegmatites are uncommon in the metasedimentary rocks. A few small pegmatites have been noted in the gabbroic intrusions that cut the rocks of the Deep Lake Formation, and one large pegmatite cuts the Gold Hill sill in the area north of Magnolia Lake. This pegmatite is poorly exposed, but a shaft sunk on the pegmatite or along the contact between the pegmatite and norite contains pegmatite consisting of microcline, quartz, and plagioclase with accessory muscovite and biotite.

GRANITE DIKES

Cross-cutting dikes of granitic composition are common in the quartzo-feldspathic gneiss in the area northwest of the Mullen Creek-Nash Fork shear zone. These dikes cut the northwest-trending foliation of the quartzo-feldspathic gneiss at large angles (perpendicular in many cases) and generally parallel the trend of the shear zone (Fig. 41). They have probably been introduced in fractures developed during some period of movement on the fault and are therefore fault related. The best exposures of the dikes are in outcrops along State Highway 230, sec. 1, T. 13 N., R. 82 W. (Pl. 1). These dikes are probably among the last igneous rocks emplaced because they cut the gneissic structure and late metadiabase dikes as well.

The essential minerals of the dikes are quartz, microcline, and albite with biotite as a major accessory mineral and epidote, ilmenite, muscovite, rutile, sphene, garnet, and zircon in minor amounts (Table 32). Every dike examined has been subjected to cataclasis (Pl. 26). The cataclasis developed parallel to the trend of the dikes so that all dikes have a strong cataclastic foliation at a large angle to the foliation of the associated gneiss.

Table 32—Mineralogy of granite dikes
(mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	3/3	41	40-43	
K-Feldspar	3/3	29	24-37	Some grains have perthite, both large grains and interstitial small grains
Plagioclase	3/3	20	12-39	Albite
Muscovite	3/3	7	1-15	
Biotite	3/3	2	0.3-4	Reddish-brown
Sphene, epidote opaque minerals, garnet, zircon	—	—	—	Zircon in large euhedral crystals, no pleochroic haloes in biotite

PRECAMBRIAN ROCKS SOUTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

HORNBLENDE GNEISS AND RELATED ROCKS

South of the Mullen Creek-Nash Fork shear zone, there is a complex of hornblende gneisses of great mineralogic and textural variation that have been mapped as a unit because the rocks are interlayered with one another and grade into each other. The hornblende gneiss unit includes fine-layered hornblende gneiss, coarse-layered hornblende gneiss, amphibolite, biotite gneiss, sillimanite biotite gneiss, calc-biotite gneiss, diopside gneiss, and marble. The main outcrops of these units are in a roughly east-west belt in T. 13 N., a north-east trending belt in the south largely in T. 12 N., and another northeast-trending belt of rocks southwest of Centennial (Pl. 1). Similar units are present at the north end of Sheep Mountain and in smaller outcrops scattered throughout the area usually as small bodies in quartzo-feldspathic gneiss. These rocks are dark and well foliated generally and are less resistant to erosion than other rocks of this area. Because of their ease of erosion, the actual outcrop area of the units is probably underestimated especially in the areas of poor outcrops.

FINE-LAYERED HORNBLENDE GNEISS

The fine-layered hornblende gneiss is a laminated rock with alternating dark amphibole-rich layers and light amphibole-poor layers (Pl. 19). The layers are of variable thickness ranging from a few millimeters to three to five centimeters, and commonly layers of different thickness are associated with one another. These layered hornblende gneisses grade into more massive rocks richer in amphibole that are distinctly foliated but have less well-developed layering. In many outcrops the layered rocks show very complex folding of the passive type (Pl. 19) which may be on large and small scales but can usually be seen in a single hand specimen. Some of the best exposures of this rock type are found in the vicinity of the Big Creek Mine, sec. 9, T. 13 N., R. 78 W., along upper Pelton Creek, sec. 7, T. 12 N., R. 78 W., secs. 12 and 13, T. 12 N., R. 79 W., and along the U-shaped southern part of Douglas Creek (Pl. 1).

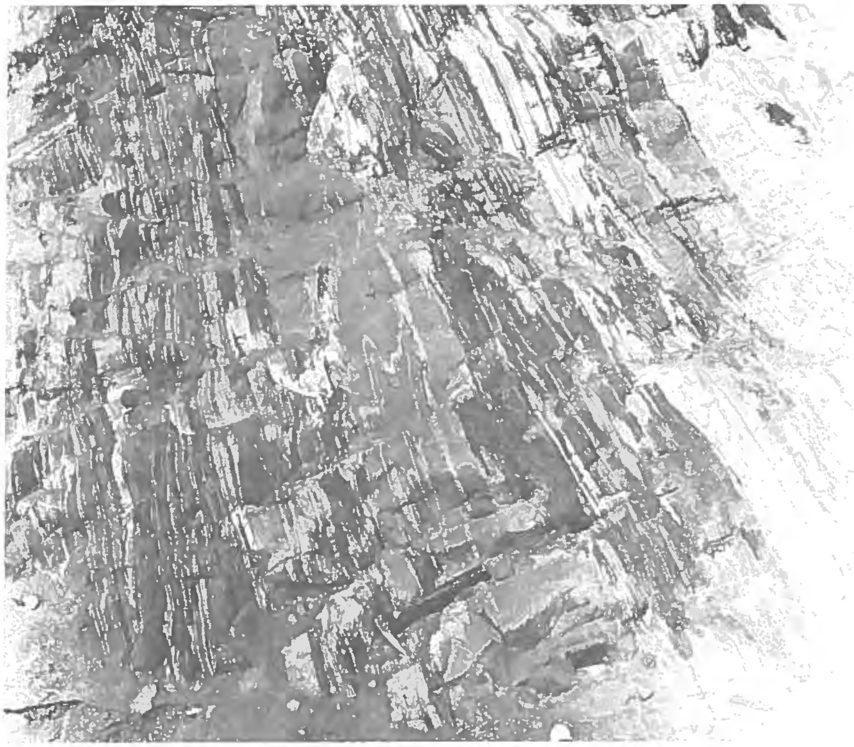
The average hornblende gneiss (Table 33) is a plagioclase-amphibole rock with quartz and epidote as major accessory minerals. The less well-layered types contain less quartz, and the anorthite content of plagioclase is higher.

Table 33—Mineralogy of fine-layered hornblende gneiss
(visual estimates, in percent)

Mineral	A	$\bar{X}$	Range	Comments
Amphibole	15/15	33	15-58	Ny = green, Nx = tan, Nz = blue-green, $z \wedge c$ = 15-19, 2V = 70-80, neg., v<r
Plagioclase	15/15	28	5-59	Av. Comp. An[illegible], Range An[illegible] - An[illegible]
Quartz	15/15	27	8-64	
Epidote	12/15	6	0-25	As discrete grains and by alteration of amphibole, rarely in veinlets
Sphene	12/15	2	0-5	
Biotite, chlorite, clinozoisite, microcline, garnet, zircon, opaque minerals				Microcline interstitial, along intergrain boundaries. Zircon small euhedral xtals with pleochoric haloes in amphibole. Chlorite after amphibole and interlayered with biotite

COARSE-LAYERED HORNBLENDE GNEISS

In contrast to the hornblende gneiss previously discussed, these units have a layering on the scale of feet, with individual layers of gneiss ranging from 2 feet to 50 feet in thickness (Fig. 17). These layered rocks range from units in which the principal rock type is hornblende gneiss (in some areas more nearly amphibolite) to units



A

Photo of outcrops on north side of Highway 230, sec. 19, T. 13 N., R. 77 W.

B

Photo of outcrop on west side of Big Creek, sec. 9, T. 13 N., R. 81 W.

Plate 19—Fine-layered hornblende gneiss showing passive folds developed by deformation of compositional layering in the gneiss.

having approximately equal amounts of each rock type to units consisting chiefly of quartzo-feldspathic gneiss. Most of these units have been shown as hornblende gneiss on the geologic map (Pl. 1), but some have been included with the quartzo - feldspathic gneiss where hornblende gneiss is present in minor amounts. Individual layers are in sharp contact with each other and these conformable contacts may be traced for considerable distances on the surface. In general, contacts between these layers and the attitude of foliation is also conformable, but certain broader structural relationships, such as, in the fold on Big Creek, suggest that this relationship may not hold in major structures (Pl. 3). The best examples of the coarsely-layered hornblende gneiss are found in exposures along Big Creek north of the point where it crosses Highway 230, and in exposures along road cuts on Highway 230 west of Woods Landing, secs. 18 and 19, T. 13 N., R. 77 W.

The hornblende gneiss of the layered units is similar to that previously described and may be finely layered and may even have passive folds as in units composed chiefly of hornblende gneiss The mineralogy of the hornblende gneiss or dark layers is shown in Table 34 and in general the is similar to that previously described and may be finely layered and may even have passive folds as in units composed chiefly of hornblende gneiss. The mineralogy of the hornblende gneiss or dark layers in shown in Table 34 and in general the composition is like that of the hornblende gneiss

Figure 17—Conformable pegmatite in unit containing lay ers of hornblende gneiss (dark) and quartzo-feldspathic gneiss (light). Photo of road cut on north side of Highway 230, a few hundred feet north of point where Big Creek crosses highway.

above. Some hornblende-rich layers grade into amphibolite.

The mineralogy of the quartzo-feldspathic or light colored layers is quite different and shows a significant increase in quartz, biotite, and microcline (Table 34). The light colored layers also have a lower anorthite content of plagioclase An_{31} as compared to An_{42}, and ranges of plagioclase composition (Table 34) do not overlap. The microcline in the light colored layer is largely found in intergranular networks.

Table 34—Mineralogy of coarse-layered hornblende gneiss (visual estimates, volume percent)

Mineral	A	$\bar{X}$	Range	Comments
		Dark Colored Layer		
Amphibole	6/6	42	30-55	Ny = green, Nx = tan, Nz = blue-green $z \wedge c$ = 20 2 V = 60 (est.)
Plagioclase	6/6	36	20-50	Av. comp. An_{42}, Range An_{34} - An_{54}
Quartz	6/6	13	4-25	Inclusions in amphibole and as interstitial grains, small grains well rounded
Sphene	4/6	P		Scattered grains, some euhedral
Epidote	5/6	P		Largely after amphibole
Apatite	5/6	P		In euhedral crystals
Biotite, Opaque minerals, chlorite, garnet		P		Opaque minerals largely magnetite, in part, altered to hematite
		Light Colored Layers		
Quartz	3/3	47	36-55	Largely anhedral, some small subrounded
Plagioclase	3/3	34	27-40	Av comp An_{31}, Range An_{30}-An_{32}
Biotite	3/3	7	3-15	Dark brown
Microcline	3/3	4	1-7	Mostly interstitial, magnetite at plagioclase contacts
Epidote, Sphene, Zircon, Apatite, Garnet, Amphibole, Opaque minerals		P		Amphibole similar to that in dark colored layer, more altered to epidote, exceptionally large crystals of sphene and magnetite

BIOTITE GNEISS

Biotite gneiss is interlayered with hornblende gneiss and grades into hornblende gneiss both along strike and across contacts. It is a common rock type in the area west of Deerhorn Point, secs. 14 and 23, T. 13 N., R. 81 W., along Pelton Creek in secs. 14 and 22, T. 12 N., R. 79 W., and in a belt about two and one-half miles long along the north edge of Platte Ridge, secs. 13, 14, 15, and 24, T. 13 N., R. 80 W. The rock is not as distinctly layered as the hornblende gneiss but has a well-developed foliation as a result of a strong preferred orientation of biotite.

The biotite gneiss is somewhat similar in composition (Table 35) to the light colored layers of the coarsely layered hornblende gneiss but is richer in microcline and muscovite. As in the light colored layers of the coarsely layered hornblende gneiss, the microcline is found in intergranular networks, but in the biotite gneiss there are sporadic large microcline porphyroblasts.

Table 35—Mineralogy of biotite gneiss
(visual estimates, volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	5 5	35	30-40	
Plagioclase	4 5	25	0-45	Av comp. An_{23}, (±5%), Range An_{21} - An_{30}
Microcline	4 5	14	0-32	Largely interstitial, myrmekite common
Biotite	5 5	12	5-35	Tan to greenish-brown
Muscovite	2 5	4	0-25	
Amphibole, opaque minerals, garnet, sphene, zircon, epidote, apatite, chlorite perovskite	—	P	—	Amphibole, Nx = tan, Ny = green, Nz = blue-green, neg., zircon both well-rounded and euhedral types

SILLIMANITE-BIOTITE GNEISS

Sillimanite-biotite gneiss is interlayered with hornblende gneiss and biotite-garnet gneiss along upper Illinois Creek about one mile north of Douglas Lookout, secs. 26 and 27, T. 13 N., R. 79 W. The rock is a dark-grey well foliated gneiss with alternating layers rich in biotite and layers rich in quartz. Sillimanite is found in large groups of fibrous crystals and is scattered through the rocks as single crystals. The mineralogy of the rocks appears in Table 36.

Table 36—Mineralogy of sillimanite-biotite gneiss
(visual estimates, volume percent)

Mineral	A	%	Comments
Quartz	1 1	36	
Biotite	1 1	20	Brown with faint reddish tint
Sillimanite	1 1	17	In large masses of fibrous crystals, and in scattered single crystals
Muscovite	1 1	10	Some crystals perpendicular to foliation
Plagioclase	1 1	8	
Garnet	1 1	4	
Opaque minerals	1 1	3	Magnetite
Apatite, tourmaline	P	P	Tourmaline as large green crystals

DIOPSIDE GNEISS, CALC-BIOTITE GNEISS AND MARBLE

Diopside gneiss, calc-biotite gneiss and marble are interlayered in three localities; Centennial Ridge, north end of Sheep Mountain, and in the area north of Smith North Creek, T. 13 N., R. 79 W. They are also interlayered with hornblende gneiss, biotite gneiss, quartz-biotite-andesine gneiss, and especially on Centennial Ridge, quartzo-feldspathic gneiss. These rock types are not only interlayered on a large scale (layers tens of feet in thickness) but also on a small scale (millimeters in thickness).

The diopside gneiss consists primarily of green diopside with lesser amounts of hornblende, calcite, quartz, and plagioclase. Epidote, sphene, apatite, magnetite, and garnet may be present as accessory minerals (Table 37). The marbles are not by any means pure calcite, but consist principally of calcite, quartz, and plagioclase, with variable amounts of amphibole, diopside, epidote, sphene, biotite, magnetite, apatite, garnet, zircon, and microcline. Both of these rock types have fine layering with layers rich in calcite-amphibole with minor diopside; calcite-quartz with minor plagioclase; diopside-quartz with minor amphibole; diopside-epidote; garnet-epidote quartz; quartz-plagioclase with minor amphibole; and epidote-amphibole with minor quartz.

Table 37—Mineralogy of diopside gneiss
(visual estimates, volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Clinopyroxene	5 5	29	20-40	Diopside, pale green, $z \wedge c$ = 38-42 and diopside augite, $z \wedge c$ = 43-47
Amphibole	5 5	29	15-40	In part after pyroxene, in part, as discrete crystals not associated with pyroxene Nx = tan, Ny = dark green, Nz = dark blue-green, 2V = 45-50 (est.)
Plagioclase	5 5	14	4-32	$An_{[illegible]}$ - $An_{[illegible]}$ (± 5%) may show reverse zoning
Quartz	5 5	13	5-17	
Carbonate	4 4	10	0-20	Calcite by stain test
Epidote	3 5	4	0-8	Some as alteration of plagioclase and as discrete grain
Garnet, sphene, opaque minerals, apatite		P		Garnet probably grossularite, N = 1.743 (± 0.003)

The calc-biotite gneiss, which is more abundant in the Smith North Creek area than elsewhere, is composed of calcite, quartz, and biotite with lesser amounts of muscovite, epidote, sphene, plagioclase, pyrite, and tourmaline (Table 38). It is interlayered with quartz-biotite andesine gneiss and hornblende gneiss.

Table 38—Mineralogy of calc-biotite gneiss
(visual estimates, volume percent)

Minerals	A	%	Comments
Carbonate	1 1	36	Calcite by stain test
Quartz	1/1	25	
Biotite	1/1	15	
Muscovite	1/1	10	
Epidote	1/1	12	
Sphene	1/1	5	
Plagioclase	1 1	4	$An_{[illegible]}$ (±5%)
Opaque minerals	—	P	Opaque minerals largely cubic xtals pyrite
Tourmaline			Tourmaline Blue green

ORIGIN OF THE HORNBLENDE GNEISS

Hornblende gneiss and amphibolite may originate by metamorphism and or metasomatism of a variety of rock types. These could be mafic igneous rocks including intrusive rocks, flows, and tuffaceous rocks; calcareous or dolomitic shales, tuffaceous shales, and impure graywackes. The problem usually reduces to an attempt to decide if the gneiss or amphibolite is para- or ortho-, i.e., of sedimentary or igneous origin. Many techniques have been used to solve this problem including field methods (Buddington, 1939; Faessler, 1948; Poldervaart and Van Backstrom, 1949), mineralogy (Wilcox and Poldervaart, 1958; Walker, et. al., 1960), chemistry (Evans and Leake, 1960; Lapadu-Harques, 1953) and trace elements (Engel and Engel, 1951; Evans and Leake, 1960, Wilcox and Poldervaart, 1958). These various methods are most successful when used in the study of lower rank metmorphic rocks and as noted by Walker, et. al. (1960, p. 149-177) there is a general convergence in the chemistry and mineralogy of these rocks with increasing metamorphic rank. The best evidence bearing on origin of the hornblende gneiss and amphibolite is field relationships and recognition of primary textures and structures. Unfortunately primary textures and structures have not been recognized in any of the units studied.

Hornblende gneiss, diopside gneiss, calc-biotite gneiss and marbles are interlayered with each other on Centennial Ridge, in the area east of Devils Gate Creek, and north end of Sheep Mountain. There is no question that the marbles in these areas are sedimentary and it seems reasonable to regard the interlayered gneiss as calcarous shales. Also in the area north of Smith North Creek the calcareous gneisses grade along strike into quartz-biotite-andesine gneiss that is regarded as a subgraywacke. In the upper Illinois Creek area where hornblende gneiss is interlayered with biotite-garnet gneiss and sillimanite-biotite gneiss, the sillimanite gneiss is more than likely derived from shale, and therefore we may consider the hornblende gneiss as an interbedded, perhaps more calcareous, shale. We cannot eliminate the possibility that the hornblende gneiss is simply an interlayered igneous rock, but some additional support for a sedimentary origin can be had from mineralogy. It has been suggested (Williams, Turner, and Gilbert, 1954, p. 241-243; Heinrich, 1956, p. 255-256) that hornblende gneisses and amphibolites of sedimentary origin have an excess of quartz, and a greater abundance of biotite, microcline, and garnet than those of igneous origin. Diopside and tourmaline in the rocks are regarded as especially diagnostic of sedimentary rocks. The hornblende gneiss and amphibolite of igneous origin, on the other hand, is viewed as having a simpler mineralogy dominated by amphibole and plagioclase in approximately equal amounts and having epidote, quartz, biotite, and sphene in minor amounts. A review of table 33 shows that the hornblende gneiss has mineralogy that best fits the sedimentary type.

Two lines of evidence suggest that a portion of the hornblende gneiss is sedimentary in origin, but can we extrapolate and state that all of the gneiss is derived from sedimentary rocks? This would be a simple but unwarranted solution to the problem; for many of the more massive types of hornblende gneiss approach amphibolite and have mineral composition that fits a rock derived from an igneous source better than a sedimentary source. The fine-layered hornblende gneiss has a more sedimentary aspect because of the increase in quartz and the layering that could be regarded as an inherited sedimentary structure; however, the layering is certainly tectonic, in part. There are excellent examples of metadiabase, basalt, and gabbro that are transitional to layered hornblende gneiss in the area of quartzo-feldspathic gneiss in the northwestern part of the Medicine Bow Mountains as well as many examples in the geologic literature (Sederholm, 1923; Evans and Leake, 1960). Perhaps it is best to admit that the origin of all types of hornblende gneiss is uncertain, and that there may be both para- and ortho-types in the area.

The coarsely layered hornblende gneiss poses a different problem. The writer knows of no example of layers of this magnitude developed during metamorphism (i.e., metamorphic differentiation). Also as previously noted, the felsic and mafic layers differ not only in the proportion of minerals but also in mineral species and in the composition of individual species. Perhaps the original rocks were interlayered mafic igneous rocks and felsic igneous rocks, but it is just as feasible to propose interlayered graywackes and feldspathic sandstone. The origin of these layered rocks will be discussed in more detail later.

In summary, good evidence suggests that a part of the hornblende gneiss unit is sedimentary in origin, but the origin of the greater part of the unit is uncertain. The writer prefers to regard the unit as a series of rocks of mafic composition including flows, tuffs and graywackes with interbeds of limestone, quartzo-feldspathic sandstone, and rare beds of aluminous shale. Such rocks are common in the eugeosynclinal facies.

NOTES ON CHEMISTRY OF HORNBLENDE GNEISS AND AMPHIBOLITE

Three samples of typical hornblende gneiss and amphibolite have been analyzed chemically. Samples one and two, Table 39, are fine-layered hornblende gneiss; sample one from outcrops along Big Creek, sec. 9, T. 13 N., R. 81 W., and sample two comes from an outcrop along Illinois Creek where fine-layered hornblende gneiss is interlayered with sillimanite-biotite gneiss. If mineralogy and texture could be considered valid criteria, these rocks would be viewed as para-amphibolite since they are layered and quartz-rich.

Table 39—Chemical analyses of three hornblende-rich rocks of the southern Medicine Bow Mountains compared with average tholeiitic basalt, average alkaline olivine basalt, mafic intrusive rocks of the northern Medicine Bow Mountains, average graywacke and Precambrian graywacke.

Weight %	1	2	3	4	5	6	7	8
SiO_2	60.10	62.33	64.67	66.75	47.35	46.77	50.28	50.83
TiO_2	1.08	0.67	0.57	0.63	0.91	3.00	1.15	2.03
Al_2O_3	15.60	14.64	13.41	13.54	16.44	14.65	14.13	14.07
Fe_2O_3	1.93	2.58	1.24	1.60	4.50	3.71	3.18	2.88
FeO	4.31	4.47	4.53	3.54	7.79	7.94	8.21	9.06
MnO	0.06	0.09	0.13	0.12	0.20	0.15	0.15	0.18
CaO	7.97	6.81	3.04	2.54	10.79	12.42	9.35	10.42
MgO	2.79	3.04	3.23	2.15	5.49	6.82	7.15	6.34
Na_2O	3.59	2.51	2.99	2.93	3.16	2.59	2.07	2.23
K_2O	0.89	1.19	2.02	1.99	1.04	1.07	0.80	0.82
P_2O_5	0.47	0.17	0.14	0.16	—	0.37	0.09	0.23
H_2O+	1.23	0.99	1.94	2.42	2.14	0.51	1.69	0.91
H_2O-	0.37	0.28	0.20	0.55	0.60		0.49	
CO_2	0.03	0.06	2.15	1.24	0.03	—	0.49	—
Cr	0.005	0.0026	—		nd	—	0.050	—
Ni	0.005	0.007	—		nd	—	0.026	—
Sr	0.032	0.010	—		nd	—	0.007	—

1. Fine-layered hornblende gneiss, outcrop on west side of Big Creek, sec. 9, T. 13 N., R. 81 W., Medicine Bow Mountains, analyst, Tamiya Asari, Japan Analyitical Chemistry Research Institute.
2. Fine-layered hornblende gneiss from outcrop along Illinois Creek, sec. 26, T. 13 N., R. 79 W., Medicine Bow Mountains, Analyst, Tamiya Asari.
3. Average of 12 Precambrian graywackes, Pettijohn, 1963, p. 57
4. Average of 61 graywackes, Pettijohn, 1963, p. 57.
5. Fine-layered amphibolite from outcrop on north side of of Stovepipe Gulch, sec. 19, T. 13 N., R. 80 W., Medicine Bow Mountains, Analyst, Tamiya Asari.
6. Average alkaline olvine basalt, Nockolds, 1954.
7. Average of 3 mafic igneous ricks, northern Medicine Bow Mountains
8. Average tholeiitic bassalt, Nockolds, 1954.

Chemically the rocks are more like graywacke than mafic igneous rocks (Table 39). They are, however, rich in CaO and poor in K_2O as compared to average graywackes and Precambrian graywackes (table 39). They are probably not derived from calcareous shales because all show a high Na/K ratio whereas the average shale is potash rich. Figure 18 is a MgO, FeO, CaO diagram showing general areas where plots of para-amphibolites and related rocks and mafic igneous rocks and ortho-amphibolite fall (Walker, et. al., 1960, Fig. 18). The plots of mafic igneous rocks and ortho-amphibolite fall in the circular area, and plots of para-amphibolite, and related rocks fall in the oblong area towards the CaO end. There is a considerable area of overlap in composition in the center of the diagram. The two samples of fine-layered hornblende gneiss are within the area of para-amphibolite, but one is within the area of overlap and the other is near it. Plots of average tholeiitic magma, average alkaline olivine basalt, average mafic igneous rocks of the northern Medicine Bow Mountains, and even average graywacke and average Precambrian graywacke fall within the composition area of mafic igneous rocks. This would seem to illustrate the futility of trying to make distinctions between these rocks using these three oxides, but, at least, the rocks could be sedimentary.

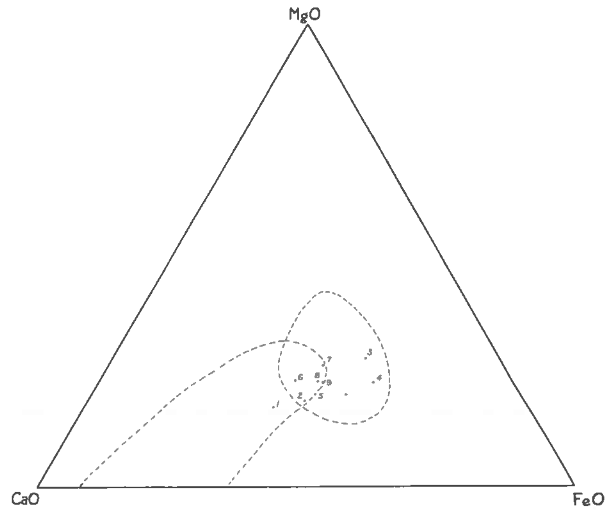


Figure 18—Compositional diagram showing plots of rock types in Table 39. Weight percentages and total iron as FeO. Oblong area in lower left is outline of compositional area for para-amphibolite and circular area for ortho-amphibolite as indicated by Walker, Joplin, Lovering, and Green, 1960, p. 163, fig. 6. Number nine is composition of amphibolite in layered complex of southwest.

The TiO_2 content of amphibolites has been used to make distinctions based on the generalization that para-amphibolite should have a TiO_2 content of 0.6 to 1.0 and ortho-amphibolite 2 to 3 percent by weight (Wilcox and Poldervaart, 1958, p. 1352; Walker, et. al., 1960, p. 165-167). The two samples under consideration have a low TiO_2 content characteristic of para-amphibolite, but unfortunately the average TiO_2 content of mafic igneous rocks north of the Mullen Creek-Nash Fork shear zone is also low (Table 29). Therefore, in this area, TiO_2 does not appear useful.

Analyses of minor elements chromium, nickel, and strontium have been made (Table 39) since chromium and nickel may be present in higher concentration in ortho-amphibolite (Engel and

Engel, 1951) and Sr may be higher in para-amphibolite (Wilcox and Poldervaart, 1958). Comparison is made with mafic igneous rocks of the Medicine Bow Mountains since these variations are of local value only. The values for Cr, Ni, and Sr in sample one fit a para-amphibolite concept, as compared to values for the average mafic igneous rock north of the fault. However, there is a range in chromium content from 0.005 to 0.091 percent in the igneous rocks making values for chromium not significant. In nickel content, all igneous rocks show a high value of 0.025 or more so the low value for nickel is suggestive of para-amphibolite. Strontium values in the igneous rocks range from 0.011 to 0.004; so for sample one, at least, the high strontium value is also suggestive of para-amphibolite. In sample two there is an overlap in minor element content for both Sr and Cr, but the low value for nickel may suggest para-amphibolite.

Sample five (Table 39) is distinctly different from the other two rocks under discussion. Although the rock does show fine-layering, the layering is not well developed and the rock is closer to a true amphibolite in mineralogy and general appearance. Chemically it is closer to mafic igneous rocks than graywacke (Table 39). It falls within the area of igneous rocks in Figure 18 where MgO, CaO, and FeO content is considered, but it is very near the area of overlap between para- and ortho-amphibolite. The TiO_2 content is low suggesting para-amphibolite, but near the average TiO_2 content of igneous rocks of this area (Table 39). This rock is within that compositional range where it is most difficult to make a distinction between ortho- and para-amphibolite on a chemical basis. Its origin is unknown.

The chemical analyses, as is now generally accepted, do not allow one to make distinctions between amphibolites and hornblende gneiss of sedimentary or igneous origin. In this case, they do lend support to a hypothesis that the fine-layered hornblende gneiss is sedimentary in origin.

QUARTZ-BIOTITE-ANDESINE GNEISS

The quartz-biotite-andesine gneiss is a dark-gray rock with a distinct layering developed by alternation of layers rich in biotite and epidote with layers rich in quartz. It is best exposed in a synform located in the south-central Medicine Bow Mountains near Keystone (Pl. 1). Here the gneiss has a sedimentary aspect in the field and resembles finely bedded quartzite. It grades into hornblende gneiss to the south and into quartzo-feldspathic gneiss on the north. Near the Keystone Mine there is a discontinuous layer in the gneiss that contains elongate aggregates of amphibole, 1 to 3 centimeters long. Other rocks of similar composition to quartz-biotite-andesine gneiss are present in east-trending belts north of the villages of Fox Park and Woods Landing, in the southern part of Jelm Mountain, in the extreme southeastern corner of the Medicine Bow Mountains proper and along the northeast slope of Sheep Mountain (Pl. 1). Contacts between these units and hornblende gneiss and quartzo-feldspathic gneiss are gradational both perpendicular to and along the strike, and some of the units are hornblende-rich locally. These rocks, mapped as quartz-biotite-andesine gneiss in different areas, may or may not be genetically equivalent. They are distinguished from other gneisses and mapped as a unit primarily because they are comparatively quartz-rich.

The quartz-biotite-andesine gneiss is quartz-rich (Table 40) as compared with related rocks, and some samples contain small well rounded quartz grains that may be original as well as larger anhedral grains with an interlocking texture. These textures are somewhat similar to those of the quartzo-feldspathic gneiss of the northwestern Medicine Bow Mountains. Textures of this sort may develop from rocks that were originally bimodal in grain size and the finer grained constituents recrystallized into interlocking coarser-grained aggregates whereas the original larger grains retained their original shape. Another interesting mineralogic feature of this rock is the presence of small grains of calcite that may be primary scattered through the rock.

Table 40—Mineralogy of quartz-biotite-andesine gneiss (visual estimates, volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	5/5	48	35-67	In two sizes, small well-rounded grains and large anhedral grains
Plagioclase	5/5	18	4-35	An_{32-36} (±5%)
Biotite	5/5	13	5-18	
Epidote	5/5	6	3-8	May include some clinozoisite
Microcline	2/5	4	0-15	Interstitial, in part
Opaque minerals	4/5	3	0-8	Magnetite and pyrite
Muscovite	2/5	3	0-8	
Carbonate, garnet, sphene, amphibole, zircon, apatite, chlorite		P		Some epidote has cores of allanite

QUARTZO-FELDSPATHIC GNEISS SOUTH OF THE SHEAR ZONE

The term quartzo-feldspathic gneiss is used for a variety of light grey to pink rocks that vary from layered gneisses to augen gneiss to faintly foliated rather massive rocks. These rocks are the most abundant in the southern area and are found throughout this part of the mountain interlayered with hornblende gneiss and amphibolite and in large masses, some of which have a granitoid aspect. The larger bodies of gneiss may have coarse-grained pegmatitic phases. This unit includes rocks of diverse origin, but the rocks commonly grade into each other both perpendicular to and along the strike. Where quartzo-feldspathic gneiss and hornblende gneiss is interlayered and quartzo-feldspathic gneiss dominates, the unit was mapped as quartzo-feldspathic gneiss.

The gneiss south of the shear zone is similar in many respects to that of the north, but the textural similarity in rocks is less evident, and large areas of quartzo-feldspathic gneiss are quite massive with an igneous aspect. In discussing the quartzo-feldspathic gneiss of this area, a distinction can be made between quartzo-feldspathic gneiss that is intimately interlayered with hornblende gneiss and the more massive gneiss that is relatively free of hornblende inclusions. The quartzo-feldspathic gneiss interlayered with hornblende gneiss is part of an extremely complex rock unit. In most areas, the interlayering may be a tectonic interlayering developed by deformation of layered supracrustal rocks (Dietrich, 1960, p. 99-120), but now disrupted to the degree that nothing can be determined about the original structure or layering of the rocks. This type of interlayering is present on Centennial Ridge, T. 15 N., R. 78 W., and can be seen in outcrops along Highway 230 in T. 13 N., R. 77 and 78 W. (Pl. 19). Some felsic layers in the hornblende gneiss are simply foliated bodies of granite and aplite, but these layers can usually be recognized in the field by local cross-cutting structure and by lack of continuity of individual layers. In other areas such as in outcrops along upper Illinois Creek, T. 13 N., R. 79 W., and in local areas east of Big Creek, sec. 9 and 10, T. 13 N., R. 81 W., the layering is so regular and individual layers show such continuity that it is possible a primary layering is preserved. It is this type of layered unit that will be discussed below. It must be emphasized that even here most of the layering may be tectonic.

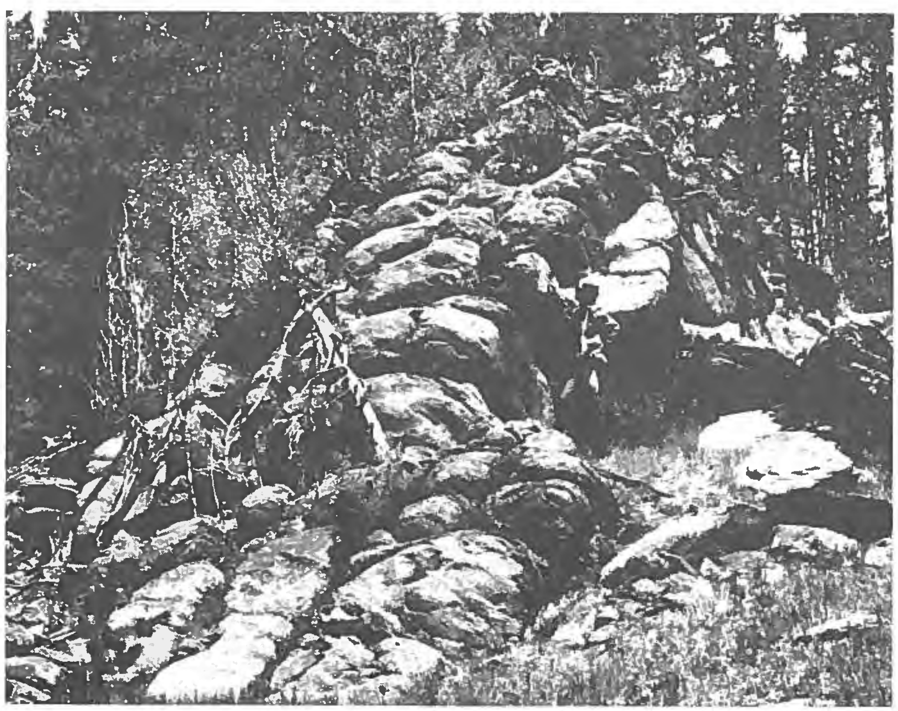
The quartzo-feldspathic gneiss in this type of layered unit is the light layers within the coarse-layered hornblende gneiss reviewed in the discussion of the hornblende gneiss unit. These light layers differ from the hornblende gneiss, not only in the proportion of minerals but also in the composition of plagioclase (more sodic) and percentage of microcline (greater in the felsic layers) (Table 34). Although there are remnants of hornblende gneiss within the light layers and local gradational contacts between these two rocks, most contacts are sharp, and it seems best to view these rocks as having original compositional differences; rather than suggesting that the felsic layers were derived from the hornblende gneiss by metasomatism or through metamorphic differentiation.

We must note that the hornblende-rich layers may differ in composition and structure. Some layers are true amphibolite whereas others are more like the fine-layered hornblende gneiss. The coarse-layered unit is probably a very complex rock unit containing hornblende-rich layers that were originally some type of sedimentary rock and hornblende-rich layers (amphibolite) that may have been dikes, sills, or flows. Perhaps these rocks were originally interlayered feldspathic sandstone and calcareous shale or graywacke that either had flows of mafic composition interbedded with them or had mafic magma introduced as sills or dikes to form the more massive amphibolite. In any event the quartzo-feldspathic gneiss may be a recrystallized sediment rather than a unit derived from the hornblende gneiss by metasomatism. The amphibolite layers will be discussed in more detail later.

The more massive bodies of quartzo-feldspathic gneiss have a number of characteristics that suggest they may be of metasomatic origin. There is, for example, a general transition from layered hornblende gneiss and quartzo-feldspathic gneiss to more massive quartzo-feldspathic gneiss in a number of areas. In the two antiform structures in the Big Creek area (Pl. 3) the layering is less distinct in the cores of the antiforms and the rock is largely quartzo-feldspathic gneiss which is less well foliated and richer in microcline than the gneiss interlayered with hornblende gneiss. In

A

Relatively close spaced joints parallel to foliation results in outcrop pattern resembling tombstones.



B

Less well-foliated gneiss with wider spacing of joints; rock granitoid in aspect.

Plate 20—Outcrops of more massive quartzo-feldspathic gneiss from the southern Medicine Bow Mountains.

areas where hornblende gneiss is the dominant rock type, gradations between the hornblende gneiss and the quartzo-feldspathic gneiss are common at contacts. Along upper Illinois Creek and Park Run Creek, sec. 26 and 27, T. 13 N., R. 79 W., hornblende gneiss grades into augen hornblende gneiss with large porphyroblasts of microcline. The latter rock in turn grades into quartzo-feldspathic gneiss with augen and finally grades into a more massive quartzo-feldspathic gneiss through a zone about one mile wide. There are also areas where hornblende gneiss and the layered units are gradational into quartzo-feldspathic gneiss, but the massive quartzo-feldspathic gneiss is distinctive enough in texture and mineralogy to be mapped as a separate unit. Several bodies of this type have been mapped as older granite on plate one. These include a body in the Deerhorn Point area, sec. 19, T. 13 N., R. 80 W., a body east of Fox Park, sec. 23, T. 13 N., R. 78 W., and two bodies on Jelm Mountain (Pl. 1). In general, the structure of these rocks (foliation developed by mineral orientation or faint compositional layering) is the same as that of the country rock.

Mineralogically the massive types of quartzo-feldspathic gneiss are somewhat similar to the quartzo-feldspathic gneiss interlayered with hornblende gneiss, but they are richer in microcline, and have a more sodic plagioclase (table 41). The microcline is commonly interstitial, and has a habit suggestive of development by replacement along grain boundaries.

Table 41—Mineralogy of massive quartzo-feldspathic gneiss (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Quartz	11/11	40.5	32-50	Some samples have two types—small well-rounded and large anhedral grains, graphic intergrowths with K-feldspar in 2 slides.
Plagioclase	11/11	39.0	30-45	Av. comp. An$_{24}$ (±5%), Range An$_{22}$-An$_{31}$. 5 samples show reverse zoning
K-Feldspar	9/11	8.4	0-18	Microcline, interstitial and as discrete grains, myrmekite at plagioclase contacts
Biotite	8/11	4.6	0-12	
Epidote	9/11	1.5	0-5	After Amphibole, in part
Amphibole	3/11	1.4	0-11	Nx = tan, Ny = green, Nz = blue-green
Muscovite	4/11	1.0	0-5	
Opaque minerals	8/11	0.9	0-5	Magnetite and ilmenite
Chlorite	4/11	0.6	0-3	After biotite, in part
Apatite	3/11	0.5	0-1	
Sphene	7/11	0.4	0-1	
Garnet	4/11	0.4	0-4	
Rutile, zircon, garnet, allanite	—	—	—	Allanite may have rims of epidote and some large xtals of allanite were noted

In order to examine the mineralogical and textural transformations from the hornblende gneiss to the massive quartzo-feldspathic gneiss more carefully a series of thin sections were cut along the margin of the body of quartzo-feldspathic gneiss in the Deerhorn Point area. Here the transition goes from fine-layered hornblende gneiss to faintly layered hornblende gneiss to quartzo-feldspathic gneiss with distinct foliation to a more massive quartzo-feldspathic gneiss. We must emphasize that although this appears to be a transitional sequence from mafic to felsic rocks, contacts may be sharp between some mafic and felsic layers. Thus we cannot eliminate the possibility that we deal with original differences in a complex sequence of rocks or even tectonic layering rather than "metasomatic" transition.

Table 42—Mineralogical changes at gradational contacts between hornblende gneiss and quartzo-feldspathic gneiss, Deerhorn Point area. (mode in volume percent)

Sample Number	62-68	62-66	62-65
	Hornblende gneiss, layered, hornblende rich	Hornblende gneiss, well-developed layering	Quartzo-feldspathic gneiss, faint layering
Amphibole	56.4% 2V = 55 (Est.)	16.4% 2V = low	2.6% 2V = low replaced by chlorite and epidote, in part
Plagioclase	34.2% An$_{31}$(±5%)	37.4% An$_{29}$(±5%) reverse zoning	64.26%
K-Feldspar	0.0%	28.1% Interstitial and as large anhedral grains—may interfinger with plagioclase	5.2% small interstitial crystals
Quartz	4.4% Small subrounded grains, scattered through rock and as inclusions in amphibole	11.9% Small subrounded grains, interstitial quartz, few large crystals	27.3% Small well-rounded grains and large masses of anhedral crystals
Epidote	3.3% After amphibole and in veinlets with chlorite	1.9% As discrete crystals	0.4%
Opaque minerals		1.3%	0.2%
Chlorite	1.8% After amphibole	—	—
Sphene	—	2.1% Discrete crystals	0.1% Discrete crystals
Apatite	—	1.0%	—
Zircon	—	—	tr

The primary changes in mineralogy are in the percentages of amphibole, quartz, and microcline. Microcline shows a somewhat erratic increase whereas amphibole decreases in a systematic fashion and quartz increases in a systematic fashion from hornblende gneiss to quartzo-feldspathic gneiss. The anorthite content of the plagioclase changes from An$_{31}$ in the fine-layered hornblende gneiss to An$_{29}$ in the intermediate rocks and is albite in the massive quartzo-feldspathic gneiss. A similar chemical change may take place in the amphibole. The amphibole is strongly pleochroic ranging from bluish-green to green-brown, and the color scheme remains about the same in the different rocks, but the 2V in the hornblende-rich rocks is approximately 55 degrees, and is smaller in the quartzo-feldspathic gneiss (Table 42). This suggests a sodium enrichment in both plagioclase and amphibole from the hornblende gneiss to the quartzo-feldspathic gneiss, and a variable increase in potash, but higher in the felsic rocks. The textural relationships between minerals has been considered to try to determine if there is a type of selective replacement of one mineral by the other. Aside from the compositional changes in plagioclase and amphibole, the major mineralogic change is between amphibole and quartz, but there is no evidence of selective replacement of amphibole or any other mineral by quartz. There is also no evidence of albitization of plagioclase. Only microcline has a habit suggestive of replacement in as much as it is commonly intersitial and has distinctly ragged borders at contacts with other minerals.

In the upper Illinois Creek-Park Run Creek area, the transition from hornblende gneiss to quartzo-feldspathic gneiss goes from a well layered hornblende gneiss to augen gneiss to massive quartzo-feldspathic gneiss. The transition is perpendicular to strike as in the Deerhorn Point area but the change takes place across a contact one mile wide. Despite these textural variations between this area and the Deerhorn Point area the mineralogical changes are the same. The percentage of quartz systematically increases from the hornblende gneiss to quartzo-feldspathic gneiss and the percentage of hornblende decreases. Plagioclase is richer in sodium in the more felsic rocks and amphibole shows the decrease in 2V noted in the samples at Deerhorn Point. Microcline forms large porphyroblasts that may have developed by replacement and some microcline also shows the interstitial habit noted at Deerhorn Point.

The mineralogic transitions noted in these border zones differs from those noted by Steven (1957, p. 372-373) and Swetnam (1962, p. 41-46) for similar rocks in the Northgate area of Colorado that borders the Medicine Bow Mountains on the south and in the Pelton Creek area of the southwestern Medicine Bow Mountains. Both Steven and Swetnam believe that the massive quartzo-feldspathic gneiss is metasomatic in origin. Field evidence cited by Steven and Swetnam is similar to that cited above, but textural evidence differs. Steven (1957, p. 343-345) cites evidence of alteration of hornblende to chlorite and biotite, in places through an intermediate actinolite stage, and of plagioclase altered to sodic albite with introduction of quartz and microcline. He regards the transformation of plagioclase to a more sodic variety as a replacement phenomena, and considers quartz and microcline as introduced by replacement. He proposes, in essence, that in early stages of metasomatism hornblende in the hornblende gneiss was converted to biotite, plagioclase was changed to a sodic variety, and considerable quartz was introduced. Continued metasomatism resulted in transformation of pre-existing minerals to microcline along grain boundaries, and progressive destruction of biotite along with additional introduction of quartz until the hornblende gneiss was converted to massive quartzo-feldspathic gneiss (quartz monzonite gneiss of Steven).

The major difference in the petrographic evidence cited by Steven and that cited by the writer is that for selective albitization of plagioclase and that for introduction of quartz by replacement. The writer does not deny that plagioclase is more sodic in the quartzo-feldspathic gneiss, but regular variations in plagioclase composition are expected in rocks of varying bulk composition. Textural evidence suggestive of replacement of calcic plagioclase has not been observed. The hornblende in the hornblende gneiss is locally replaced by biotite, chlorite, and epidote, and in some areas a secondary actinolite is present. This alteration of amphibolite is most commonly found in areas subjected to mild cataclasis, and may be largely a retrograde metamorphic effect, but there are rocks showing this conversion that exhibit no obvious deformation. In certain areas in the quartzo-feldspathic gneiss, quartz is unquestionably introduced. In pegmatized quartzo-feldspathic gneiss quartz is introduced or secreted along fractures and may replace primary minerals of the gneiss, but this type of texture was not noted in the transitional rocks. The evidence of introduction of microcline along grain boundaries is certainly supported by results of this study.

NOTES ON THE CHEMISTRY OF THE QUARTZO-FELDSPATHIC GNEISS

Chemical analyses have been made of rocks gradational from hornblende gneiss to massive quartzo-feldspathic gneiss in the Deerhorn Point area as well as of typical samples of quartzo-feldspathic gneiss from the southern Medicine Bow Mountains (Tables 43 and 44). Figure 19 is a Harker diagram showing the changes in the percentage of principal oxides plotted against silica at the contact zone. The Harker diagram shows an almost straight line decrease in iron oxide, calcium oxide, and magnesium oxide with corresponding increase in sodium oxide from the hornblende gneiss to the massive quartzo-feldspathic gneiss. The potassium content increases stepwise from the hornblende-rich rocks to the plagioclase-quartz rich rocks. Both water and titanium oxide also decrease from hornblende gneiss to massive quartzo-feldspathic gneiss. The aluminum oxide percent is remarkably consistent in the different rock types. These chemical changes are similar to those found where there is transformation by metasomatism to rocks nearer the composition of granite. Unfortunately they do not prove metasomatism because here we cannot be certain that this does not represent original compositional differences in the rocks or even local chemical transfer at the contact.

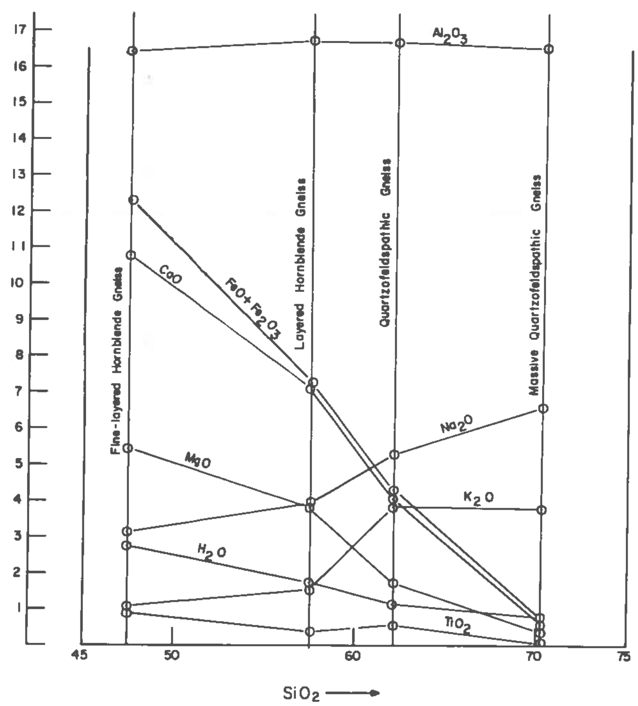


Figure 19—Harker diagram showing changes in the percentage of major oxides in contact zone at Deerhorn point.

Table 44 shows analyses of three typical samples of massive quartzo-feldspathic gneiss. The first comes from the locality at Deerhorn Point; second comes from outcrops on upper Illinois Creek-Park Run Creek; and sample three comes from the southwest corner of the Medicine Bow Mountains, sec. 16, T. 12 N., R. 80 W. The

Table 43—Chemical analyses of rocks gradational from hornblende gneiss to massive quartzo-feldspathic gneiss Deerhorn Point area, Medicine Bow Mountains

	1	2	3	4
Sample Number	62-68	62-67	62-66	62-65
SiO_2	47.35	57.34	62.05	70.26
Al_2O_3	16.44	16.73	16.73	16.51
Fe_2O_3	4.50	2.87	2.05	0.74
FeO	7.79	4.36	2.61	0.39
MgO	5.49	3.82	1.73	0.37
CaO	10.79	7.18	4.12	0.70
Na_2O	3.16	3.93	5.30	7.09
K_2O	1.04	1.55	3.83	3.78
TiO_2	0.91	0.43	0.56	0.04
MnO	0.20	0.13	0.08	0.02
CO_2	0.03	0.02	0.08	0.03
H_2O+	2.14	1.35	0.75	0.39
H_2O-	0.60	0.36	0.42	0.33

1 Very fine-layered hornblende gneiss
2 Layered hornblende gneiss with greater proportion of felsic layers.
3 Quartzo-feldspathic gneiss with rare mafic layers.
4 Massive quartoz-feldspathic gneiss.
Analyst Shirou Imai, Japan Analytical Chemistry Research Institute.

rocks show significant chemical differences suggesting that they are not part of a series of granitic rocks of relatively homogeneous composition emplaced at this level of the crust.

Table 44—Chemical analyses of massive quartzo-feldspathic gneiss, southern Medicine Bow Mountains

	1	2	3
SiO_2	70.26	72.41	75.53
Al_2O_3	16.51	13.74	11.75
Fe_2O_3	0.74	1.40	2.22
FeO	0.39	1.63	0.88
MgO	0.37	0.61	0.17
CaO	0.70	2.60	0.63
Na_2O	7.09	4.16	3.42
K_2O	3.78	2.32	4.36
TiO_2	0.04	0.23	0.21
MnO	0.02	0.03	0.01
CO_2	0.03	0.05	0.00
H_2O+	0.39	0.91	0.41
H_2O-	0.33	0.25	0.36

1 Massive quartzo-feldspathic gneiss, Deerhorn Point.
2 Massive quartzo-feldspathic gneiss, upper Illinois Creek-Park Run Creek area, sec. 25, T. 13 N., R. 79 W.
3 Massive quartzo-feldspathic gneiss, southwest corner of Medicine Bow Mountains, sec. 16, T. 12 N., R. 80 W.
Analyst Shirou Imai, Japan Analytical Chemistry Research Institute.

ORIGIN OF THE QUARTZO-FELDSPATHIC GNEISS

This group of rocks has had such a complex history, especially tectonic, that statements on origin are strictly in the realm of speculation. We probably deal with rocks of original igneous and sedimentary parentage that now show similar mineralogical and textural features. However, the entire body of feldspathic gneiss is probably not derived from original hornblende gneiss by metasomatism. In some areas where the quartzo-feldspathic gneiss is interlayered with hornblende gneiss it may have been a sedimentary rock.

The following facts support a sedimentary origin for the quartzo-feldspathic gneiss.

(1.) Contacts between layers of quartzo-feldspathic gneiss and hornblende gneiss are commonly sharp.

(2.) Individual layers of quartzo-feldspathic gneiss show continuity along strike.

There is evidence (which will be reviewed later) that the great part of the layering is tectonic, but even so, we may deal with a transposed original sedimentary layering. There is also local evidence of origin of felsic layers by replacement of hornblende gneiss, but the writer prefers to consider this a local phenomena enhanced by late potash metasomatism rather than the principal mode of origin of the rocks.

The massive quartzo-feldspathic gneiss including some bodies mapped as older granite may in part, be metasomatic in origin. Facts supporting a metasomatic origin for the massive quartzo-feldspathic gneiss are listed below.

(1.) The structure of the massive gneiss is identical with that of the host rock.

(2.) Contacts are gradational and remnants of host rock are present in the massive gneiss.

(3.) Chemical transitions at contacts are those expected for metasomatic transformation of hornblende gneiss to a rock approaching granite.

(4.) The massive gneiss shows considerable chemical and mineralogical variation more typical of rocks with inherited chemical differences than of metamorphosed intrusions of granitic composition. The writer's opinion is that the massive quartzo-feldspathic gneiss developed, in part, by metasomatism of a pre-existing series of hornblende gneisses and related rock.

Certainly our current information on these rocks allows no unique interpretation of origin. Detailed mapping may provide an answer, and such studies will be undertaken as part of the University of Wyoming's continuing studies of the Medicine Bow Mountains.

As a final comment on the quartzo-feldspathic gneiss, a large number of samples of the gneiss have been collected from different parts of the area for age determination. Most samples did not have rubidium to strontium ratios suitable for study. The samples finally selected for study were from the southwest corner of the Medicine Bow Mountains. The samples did not fit an isocron, but it can be stated that this gneiss does not retain evidence of the 2.4 b.y. event recorded for similar rocks in the north (Hills, et. al., 1968).

IGNEOUS ROCKS OF PRECAMBRIAN AGE SOUTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

QUARTZ DIORITE

A large body of quartz diorite covering about 34 square miles is located in the southcentral Medicine Bow Mountains south and southeast of Keystone (Pl. 1). The present shape of the intrusion is that of a wooden shoe because the central portion of it is cut by a gabbroic intrusion. A much smaller body is located in sec. 30 and 31, T. 13 N., R. 79 W. The quartz diorite is younger than hornblende gneiss and quartz andesine gneiss, but is older than other igneous rocks of the area with the possible exception of the amphibolite.

The large body of quartz diorite herein called the Keystone Quartz Diorite is a dark-gray medium-grained faintly to strongly foliated rock that is characterized by resistant elongate outcrops with widely spaced joints (Fig. 21). Some of the best exposures of this rock are along Douglas Creek south of the old mining town of Keystone and on Highway 230 west of Woods Landing. Although the Keystone Quartz Diorite is an irregularly-shaped intrusion, contacts with hornblende gneiss and quartz andesine gneiss are generally conformable. The quartz diorite is usually fine grained and more strongly foliated along contacts, and gradational contacts with hornblende gneiss are present along the southwestern border of the quartz diorite. The exception to the conformable relationship is along the eastern border of the quartz diorite where the unit locally crosscuts the hornblende gneiss at large angles and may contain numerous inclusions of the gneiss. Inclusions of gneiss are found throughout the quartz diorite as small lenticular pods and schlieren, but the most distinctive inclusions are in a north-trending belt east of Douglas Creek (Fig. 20) (Currey, 1959, p. 21-22). In this belt, inclusions make up as much as 30% of the rock and consist of hornblende gneiss identical to that along the southern border of the intrusion.

The mineralogy of the quartz diorite is shown in Table 45.

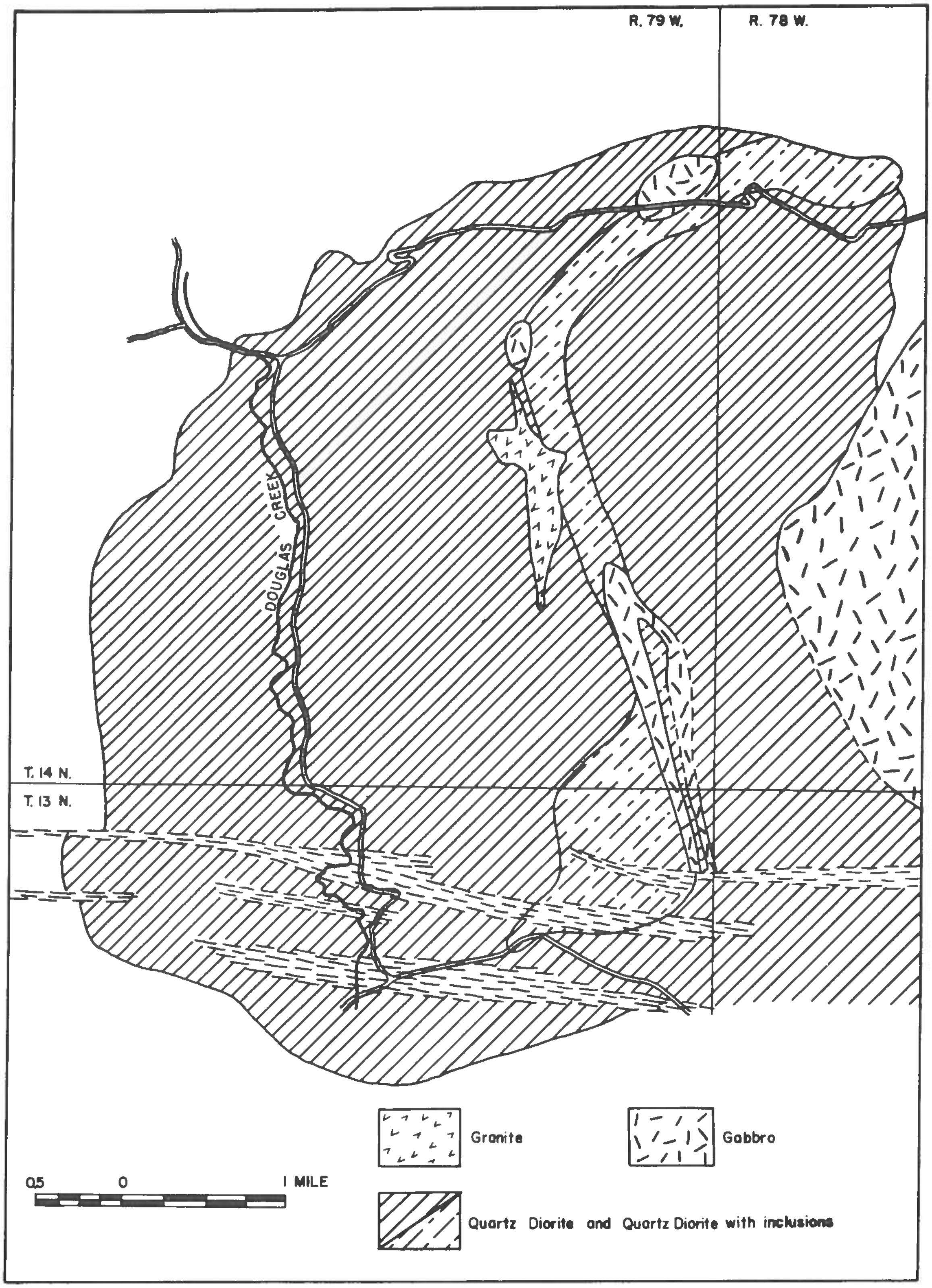


Figure 20—Sketch map of western part of Keystone Quartz Diorite showing inclusion-rich belt in central part of body. (After D. R. Currey.)

Table 45—Mineralogy of quartz diorite
(mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Plagioclase	10/10	58.9	49-72	Av. Comp., An[illegible], Range An[illegible]-An[illegible], as large platy crystals, and small interstitial grains
Quartz	10/10	17.5	8-32	Interstitial to plagioclase and amphibole
Amphibole	9/10	13.4	0-33	z∧c = 23, Nx = tan, Ny = very dark green, Nz = dark green
Biotite	10/10	5.5	1-11	Brown with very strong pleochroism
Epidote	10/10	2.5	1-4	
Chlorite	1/10	1.1	0-11	
Microcline	1/10	0.6	0-6	Interstitial
Sphene	6/10	0.4	0-2	
Opaque minerals, garnet, apatite, zircon, allanite		P	—	Opaque minerals include magnetite, ilmenite, and pyrite

Foliation is developed both by cataclasis and by parallel alignment of plagioclase and biotite in samples that do not show evidence of crushing The rocks having cataclastic textures are common in the west, and those simply showing alignment of minerals are more common in the east. There are zones in the southwest part of the intrusion of extreme crushing that are shown on Plate one. These zones can be recognized in the field by their unusually well developed foliation and distinct pink color caused by increase in potash feldspar. The increase in potassium in these crushed zones is shown below:

	Unsheared quartz diorite	Sheared quartz diorite
K_2O	0.37 percent	2.80 percent

The Keystone quartz diorite is regarded as an intrusive rock, introduced into hornblende gneiss and quartz andesine gneiss, that has been variously metamorphosed after or during its emplacement.

AMPHIBOLITE

Although there is every gradation between amphibolite and hornblende gneiss in the southern Medicine Bow Mountains, many small or large bodies are present that are mapped as amphibolite. These rocks are not as well-foliated as the hornblende gneiss, do not have well-developed

A

B

Figure 21—A. Outcrop of Keystone Quartz Diorite exposed east of Douglas Creek showing widely spaced joints typical of more massive variety of the unit. B. Inclusion-rich zone from exposures on Lake Creek, 6 inch rule. (Photographs by D. R. Currey.)

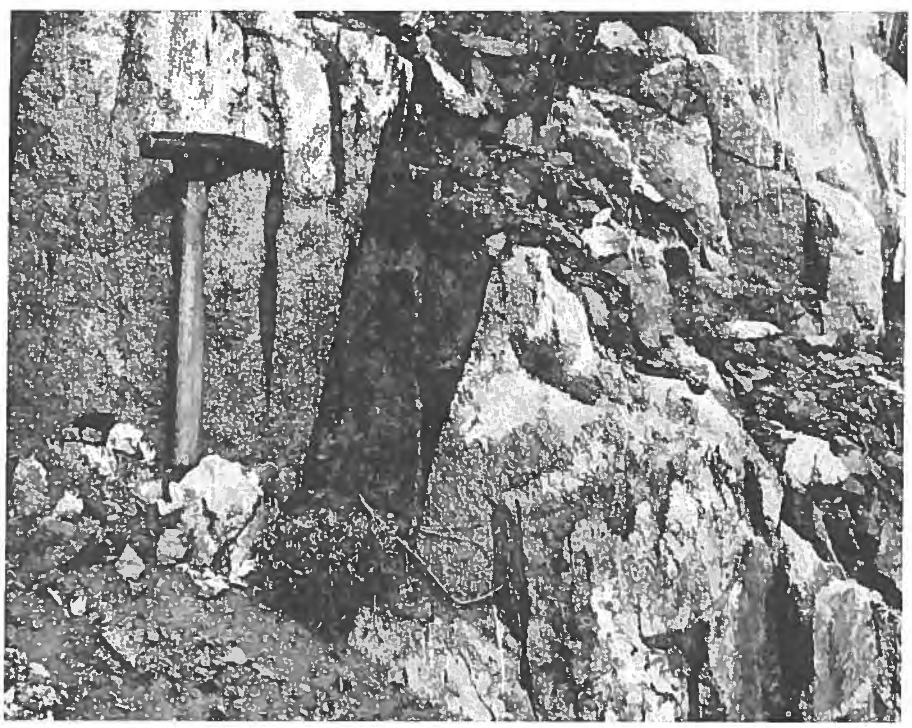


Plate 21—Amphibolite layers in quartzo-feldspathic gneiss from outcrop on State Highway 230, sec. 22, T. 13 N., R. 81 W. Note in both photographs that amphibolite bodies strike at an angle to foliation and trend of mafic inclusions in gneiss. Knife and hammer parallel trend of foliation.

layering and are richer in amphibole than the gneiss. Some of the large units of amphibolite are shown on Plate one, but there are many smaller bodies with thicknesses of a few inches to a few feet that cannot be shown at this scale.

The amphibolite is interlayered with hornblende gneiss and quartzo-feldspathic gneiss and most layers conform in strike and dip to the structure of associated gneiss. The amphibolite is completely recrystallized and retains no primary structure. Thus, in most cases, field evidence does not allow a classification as para- or ortho- types. In a few localities, however, field relationships indicate that the layers of amphibolite are cross-cutting with respect to the structure of the associated gneiss. On road cuts along Wyoming State Highway 230 east of the Big Creek ranch small layers of amphibolite cross-cut the structure of the enclosing quartzo-feldspathic gneiss (Pl. 21). In Plate 21 the layer of amphibolite is at an acute angle to the foliation of the gneiss which, in this case, is outlined by stretched mafic inclusions in the gneiss. In Figure 22, an amphibolite body cuts the gneiss nearly at right angles to the foliation but is in a V-shape perhaps as a result of later deformation. These amphibolite bodies may have been dikes of basaltic composition that have been completely recrystallized and brought to near conformity during a period of deformation and metamorphism that affected the original gneiss and contained dikes.

In other areas, the layers of amphibolite show remarkable continuity and parallel the structure of the associated gneiss almost perfectly. Examples of this kind of amphibolite are in the Elkhorn Point area near the southeast corner of the Medicine Bow Mountains. Plate 22 shows both thin and thick layers of amphibolite which have conformable contacts with interlayered quartzo-feldspathic gneiss. Even in these units, relationships suggest discordance. For example, in Plate 23. a deformed stringer of amphibolite cuts across quartzo-feldspathic gneiss and connects two thick bodies of amphibolite. This relationship suggests an igneous origin for the amphibolite.

Many of the amphibolite layers in the south are about the same width and have a very similar

Figure 22—Body of amphibolite at nearly right angles to the foliation of the enclosing quartzo-feldspathic gneiss. Foliation of gneiss parallels major joint surface in lower left. Knife for scale. Sec. 22, T. 13 N., R. 81 W.

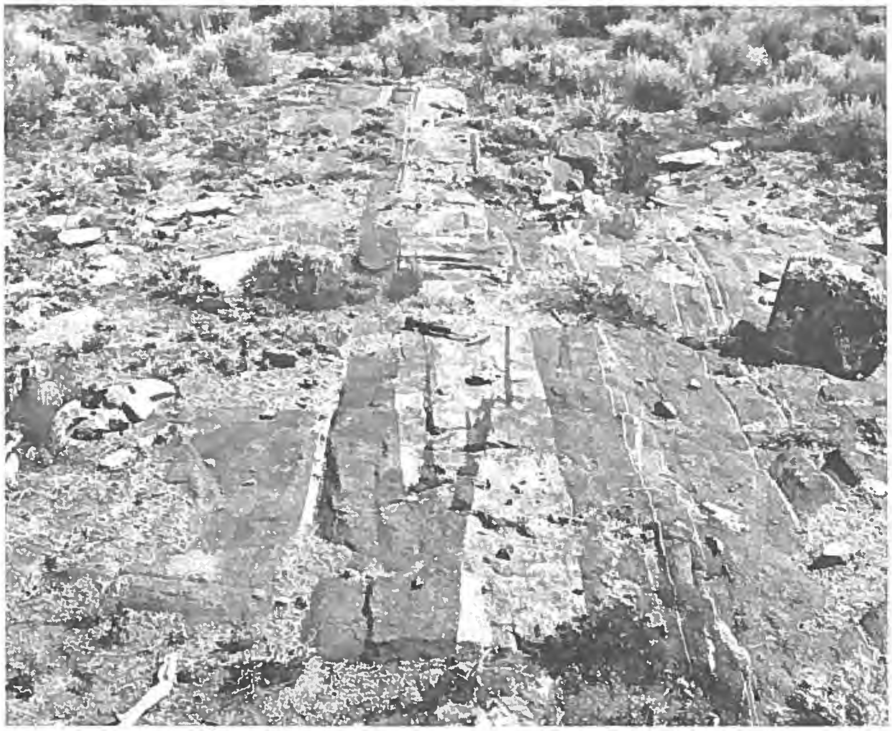
A

B

C

D

Plate 22—Examples of contacts between amphibolite and quartzo-feldspathic gneiss in the Elkhorn Point area. A. Most common field relationship where contact is parallel to foliation but actual contact is not exposed. B. Small amphibolite body with contacts parallel to gneissic foliation and trend of pegmatite layers. C. Sharp conformable contact between gneiss and amphibolite. D. Massive amphibolite that is rarely exposed as well as shown here. Hammer for scale in all photographs.



A

Two layers of amphibolite with contacts that conform in strike to foliation of gneiss but are connected by a deformed stringer.

B

Irregular-shaped, locally cross-cutting pegmatite in quartzo-feldspathic gneiss.

Plate 23—Amphibolite and pegmatite from Elkhorn Point area.

appearance to the more altered dikes and sills in the quartzo-feldspathic gneiss of the northwestern Medicine Bow Mountains. These field relationships suggest that most of the amphibolite is igneous in origin or ortho-amphibolite, but the field relationships can apply only to a specific body and it is very possible that some layers are sedimentary or that we have amphibolite of diverse origin in the same group of outcrops.

The mineralogy of the amphibolite appears in Table 46. It is typical of ortho-amphibolite as defined by Williams, Turner, and Gilbert (1954, p. 241-243) and Heinrich (1956, p. 255-256), but as noted earlier mineralogy cannot be considered diagnostic in determining origin of such rocks.

Table 46—Mineralogy of amphibolite (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
Amphibole	8/8	55.7	30-60	Nx = tan, Ny = green, Nz = blue-green, 2V = 78, N_x = 1.650-1.654, N_z = 1.666-1.671
Plagioclase	8/8	30.7	17-46	Av. comp. An[illegible], Range An[illegible]
Quartz	8/8	6.7	5-15	Some small rounded grains as inclusions in amphibole
Epidote	6/8	1.9	0-13	
Hypersthene	1/8	1.7	0-20	
Opaque minerals	6/8	1.2	0-3	Magnetite and pyrite
Apatite	8/8	0.7	0-3	
Sphene	5/8	0.5	0-3	
Biotite	3/8	0.3	0-2	
Zircon	2/8	—	—	

Inasmuch as these layers of amphibolite may be recrystallized equivalents of mafic dikes in the northwest, two chemical analyses were made of typical amphibolite from the Elkhorn Point area for comparison with igneous rocks north of the Mullen Creek-Nash Fork shear zone (Table 47). These are compared with gabbro, diabase, and ortho-amphibolite from the area north of the fault. They are close enough chemically in major elements to the known igneous rocks north of the fault so that an origin by metamorphism of the igneous rocks is possible. They are, however, high in Sr and low in the Ni as compared with the igneous rocks, making the correlation questionable. The average composition of these two amphibolites is plotted on Figure 18, no. 9, and these fall within the field of ortho-amphibolite compositionally, but within the area of overlap with para-amphibolite. Unfortunately the chemical analyses are not diagnostic.

In summary, some field evidence and mineralogy suggest an igneous origin for the amphibolite layers, but chemical analyses do not aid in classification. We may have individual layers in the same complex of similar mineralogy but different origin, but most units shown as amphibolite on Plate one are considered to be igneous by the writer.

Table 47.—Chemical analyses of amphibolite from the southwest compared with known igneous rocks from north of the Mullen Creek-Nash Fork shear zone.

	1	2	3	4	5
SiO_2	51.73	48.91	50.20	46.87	47.24
TiO_2	0.73	1.95	0.76	1.14	2.31
Al_2O_3	15.64	14.51	12.24	15.84	13.52
Fe_2O_3	2.55	4.39	2.61	2.98	4.89
FeO	7.98	8.81	7.84	9.61	9.92
MnO	0.15	0.14	0.16	0.21	0.23
CaO	11.26	7.76	12.04	10.52	9.58
MgO	6.51	5.12	9.82	6.48	5.40
Na_2O	2.07	2.74	1.39	3.08	3.63
K_2O	0.39	1.61	0.40	0.75	0.98
P_2O_5	0.06	0.16	0.06	0.23	0.36
H_2O+	0.77	2.73	1.56	1.75	2.19
H_2O-	0.53	0.55	0.38	0.47	0.44
CO_2	0.01	0.47	0.99	0.02	0.04
Cr	0.005	0.005	0.091	0.008	0.009
Ni	0.025	0.025	0.029	0.013	0.004
Sr	0.004	0.007	0.011	0.018	0.020

1. Gabbro from Gold Hill sill, sec. 17, T. 16 N., R. 80 W.
2. Diabase dike cutiing Medicine Peak Quartzite, sec. 17, T. 16 N., R. 80 W.
3. Orthoamphibolite (diabasic texture locally presreved), sec. 3, T. 13 N., R. 82 W.

4 and 5. Amphibolite, SW¼ sec. 16, T. 12 N., R. 80 W.

Japan Analytical Research Institute, Analyst, Tadashi Asari

SMALL MAFIC INTRUSIONS

Mafic intrusions ranging in width from 10 feet to one-half mile and extending along strike for a mile or more are common in the area south of the Mullen Creek-Nash Fork shear zone. These intrusive bodies are small as compared to the Lake Owens and Mullen Creek mafic complexes of the southern area, but many are as large as the Beaver Hills gabbro and Bennett Peak gabbro north of the shear zone. The great majority of these intrusions are metagabbro, but metaolivine gabbro, metadiorite, and metapyroxenite are also present. The composition of the intrusive bodies is shown on Plate one by symbol. The mafic intrusions are conformable in strike and dip to the enclosing metamorphic rocks although some may cross-cut the structure locally. All are metamorphosed to some degree especially along contacts, and as a general rule the centers of larger bodies are less metamorphosed than the smaller (Fig. 23). Where these bodies are completely altered to amphibolite or where unaltered samples were not found these bodies are designated as amphibolite on Plate one.

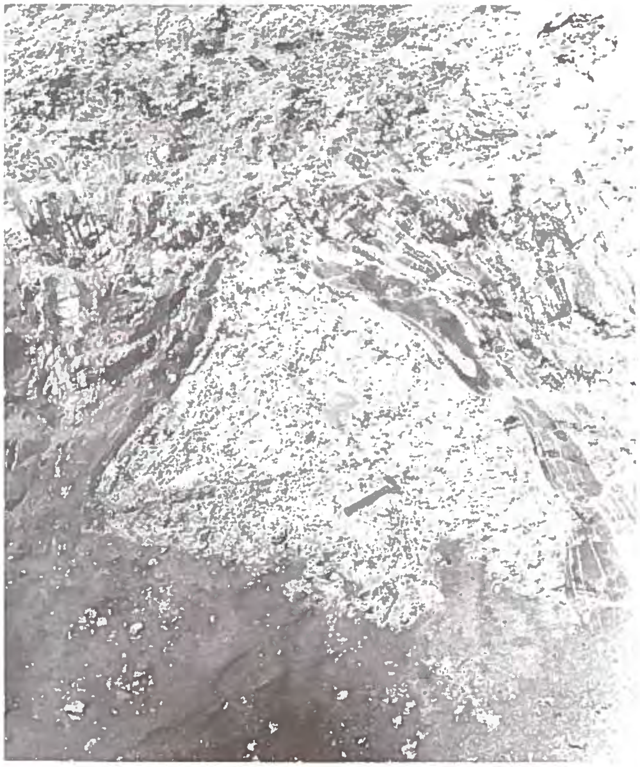


Figure 23—Small metagabbro body in hornblende gneiss showing lenticular nature of intrusion. Photograph taken on north side of Highway 230 road cut, sec. 19, T. 13 N., R. 77 W.

The petrography of typical examples of these intrusions will be reviewed below including the Parkview metaolivine gabbro, the Douglas Point metagabbro, the New Rambler metagabbro, the Queen metapyroxenite, and the Utopia metadiorite.[1] The Parkview metaolivine gabbro is a north trending bar-bell shaped body located in T. 12 N., R. 80 W. It conforms generally in strike and dip to the structure of the enclosing quartzo-feldspathic gneiss and amphibolite, and the foliation of gabbro at the contact dips the same as that of the gneiss. It is highly metamorphosed (gneissic in the south) and the northern part of the intrusion is the only area where remnants of olivine gabbro are present. Pod-like inclusions of quartzo-feldspathic gneiss are found in the olivine gabbro along its contact with gneiss. The mineralogy of less altered samples of the olivine gabbro is shown in Table 48. The olivine has reaction rims similar to those described for the Bennett Peak gabbro and the composition of the rock is similar to olivine-rich phases of this rock.

Table 48—Mineralogy of Parkview meta-olivine gabbro
(mode in volume percent)

Mineral	A	X	Range		Comments
Plagioclase	2 2	41.9	29.3	64.7	An [illegible]
Clinopyroxene	2 2	24.7	20.0	29.4	Augite, most grains partly altered to amphibole and rimmed by amphibole
Orthopyroxene	2 2	5.8	0.3	11.4	As part of reaction rim around olivine—most grains amphibolized
Olivine	2 2	3.3	0.3	6.0	Well developed reaction rims of orthopyroxene and amphibole and chlorite
Opaque minerals	2 2	0.1	0.1	0.1	
Amphibole	2 2	[illegible]	21.4	[illegible]	In various habits, all suggesting alteration of pyroxene. As rims around pyroxene and olivine and as [illegible] of small grains that are pseudomorphic after pyroxene. Actinolite
[illegible]	2 2	[illegible]	[illegible]	2.3	Secondary (?)

The New Rambler metagabbro is an interconnected series of mafic bodies located in secs. 28, 29, 32, and 33, T. 15 N., R. 79 W. and secs 4 and 5, T. 14 N., R. 79 W. (Pl. 1). The metagabbro is highly altered. Five samples have been studied and all show varying degrees of alteration towards amphibolite and two show intense cataclasis. Remnants of clinopyroxene are present in some samples that are largely altered to amphibole. Plagioclase is also extensively altered to saussurite, but an igneous texture that is typical of gabbro can be recognized in these altered rocks. One sample although largely altered to amphibole contains serpentinized olivine crystals as well as remnants of clinopyroxene. This may be a peridotitic variety of the gabbro.

The Douglas Point metagabbro is a sill of metagabbro that crops out in secs. 34 and 35, T. 13 N., R. 79 W. It is in structural conformity to the associated metasedimentary rocks and like many sills of this type is more resistant to erosion and stands above the surrounding rocks. The mineralogy of the sill is shown in Table 49.

Table 49—Mineralogy of Douglas Point gabbro
(mode in volume percent)

Mineral	One sample	%	Comments
Plagioclase		50.9	$An_{[illegible]}$ (±5%)
Clinopyroxene		35.0	Augite
Amphibole		9.2	As alteration of augite, viewed as secondary
Opaque minerals		4.9	Magnetite and ilmenite

The Utopia metadiorite is a lenticular conformable body of metadiorite located in sec. 4, T. 15 N., R. 78 W., northwest of the Utopia mine. According to McCallum (1964, p. 87) metadiorites of this type contain 40-65% hornblende usually in part chloritized), 25-60% plagioclase (An_{35-50}),

1. Gabbroic rocks given informal names for descriptive purposes.

5-15% chlorite, 2-8% magnetite and ilmenite, 2-5% secondary quartz, and 2-6% epidote, clinozoisite, sphene, biotite, apatite, and garnet. The plagioclase is saussuritized, but where determinations of composition could be made was andesine, and the textural relationships suggest it is primary. These metadiorites are thus probably originally diorites rather than gabbros with recrystallized plagioclase.

The Queen metapyroxenite is a conformable bifurcating body located in sec. 16, T. 15 N., R. 78 W., in the area of the Queen mine of Centennial Ridge. Like the New Rambler metagabbro the unit is not composed of a single rock type, but grades laterally and along strike into metagabbro and metadiorite. The amphibolized metapyroxenites contain from 70-90% amphibole, (hornblende, actinolitic hornblende, and/or actinolite), 0-15% clinopyroxene (augite-diopside), 2-25% chlorite, 1-10% magnetite, ilmenite, leucoxene, and limonite, 1-10% epidote, 1-3% sphene, 1-2% apatite, and 1-4% other accessory minerals (McCallum, 1964, p. 87). Textural relationships suggest that the amphibole is largely derived from pyroxene.

These mafic intrusives are regarded as having been emplaced after formation of the hornblende gneiss and related metamorphic rocks of the area including the quartzo-feldspathic gneiss, and quartz diorite, but as noted above all have been variously metamorphosed since their emplacement. In general, the mafic intrusives in the northeastern part of the southern area, i.e. from the Rambler mine area to Centennial Ridge, are more highly metamorphosed and more strongly foliated than the others. This may result from deformation related to the shear zones to the north. This universal metamorphism and the fact that these units do not cut each other makes it difficult to establish their sequence of intrusion.

LARGE MAFIC INTRUSIONS

There are two major bodies of mafic rock in the area south of the Mullen Creek-Nash Fork shear zone; the Mullen Creek mafic complex and the Lake Owens mafic complex (Pl. 1).[2] The Lake Owens mafic complex has been studied in more detail and will be reviewed first.

The Lake Owens mafic complex is located in the southeastern part of T. 14 N., R. 78 W., south of the village of Albany (Pl. 1). It is a large circular intrusive that covers an area of approximately 21 square miles. It has been studied by E. J. Catanzaro, (1957), and H. L. Stensrud (1963), and the following is a summary of their work. The rocks of the mafic complex include troctolite, olivine gabbro, olivine norite, gabbro and norite. Although these rock types can be recognized in the field the gradations from one rock to another are so common and local that two generalized rock zones have been recognized and mapped. These are: (1) olivine-bearing zones consisting of troctolite, olivine gabbro and olivine norite, and (2) gabbroic zones, consisting of gabbro and norite. The major olivine bearing zone is in the center of the complex and there are two smaller zones near the northern and southern borders of the complex (Pl. 1). There is also an east-west trending magnetite-rich zone about one mile long and one quarter mile wide near the southeastern border of the mafic complex. The magnetite is disseminated through the rocks and is concentrated in small stringers and pods.

The structure of the complex is defined by a steeply dipping primary foliation developed by alignment of the plagioclase crystals, and by the olivine bearing zones that form a semi-circular body concave to the north (Pl. 1). The dip of the foliation is vertical to steeply northeast. The complex cuts quartz diorite on the west and hornblende gneiss and quartz biotite andesine gneiss on the east and is in turn cut by coarse-grained granite on the north and felsic dikes on the west (Pl. 1). Its relationship to foliated granites that crop out to the west is not known.

The composition and modal analyses of the minerals and rocks are shown in Table 50. As can be noted in Table 50 some of these rocks can be regarded as leuco-varieties of gabbro. The samples examined were fresh and unmetamorphosed and generally showed a distinct alignment of plagioclase that is regarded as primary in the absence of any evidence of cataclasis. Olivine may or may not have reaction rims, but where present they are complete rims of orthopyroxene with or without amphibole exterior rims. The

2. These mafic complexes are given informal names for descriptive purposes.

amphibole exterior rims are developed where the olivine is in contact with plagioclase.

Table 50—Mineralogy of the Rocks of Lake Owens Mafic Complex (after Stensrud, 1963)

(mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
				Gabbro
Plagioclase	8/8	60	(52-70)	Av. comp., An[illegible]; Range An[illegible]
Clinopyroxene	8/8	25	(16-35)	Augite—locally altered to amphibole, Wo[illegible] En[illegible] Fs[illegible]
Orthopyroxene	8/8	11	(8-17)	En_{72}
Olivine	2/8	1	(0-3)	
Amphibole, spinel, magnetite, and pyrite	—	—	—	Amphibole (hornblende) after pyroxene, magnetite up to 9% in one sample studied. Spinel deep green
				Norite
Plagioclase	3/3	64	(57-73)	Av. comp., An[illegible], Range An[illegible]
Orthopyroxene	3/3	19	(15-25)	En[illegible]
Clinopyroxene	3/3	8	(2-15)	Augite, lamellae of orthopyroxene parallel to (100)
Olivine	1/3	—	—	Fo_{77}
Magnetite, hornblende and spinel	—	—	—	Amphibole as brown hornblende around augite
				Olivine gabbro
Plagioclase	6/6	63	(51-73)	Av comp., An[illegible], Range An[illegible]
Clinopyroxene	6/6	16	(4-22)	Augite, schiller structure
Olivine	6/6	13	(5-34)	Fo[illegible], Continuous reaction rims of orthopyroxene and amphibole
Orthopyroxene	5/6	8	(0-11)	May have inclusions of augite
Amphibole, spinel, magnetite	—	—	—	
				Olivine norite
Plagioclase	4/4	70	(52-78)	Av. comp., An[illegible], Range An[illegible]
Orthopyroxene	4/4	16	(7-34)	Interstitial to plagioclase and as rims around olivine, schiller structure common, En[illegible]
Clinopyroxene	3/4	3	(0-8)	Augite
Olivine	4/4	10	(8-12)	Fo[illegible], continuous reaction rims of orthopyroxene and amphibole
Magnetite and amphibole	—	—	—	
				Troctolite
Plagioclase	2/2	83	(same)	An[illegible]
Olivine	2/2	15	(13-17)	Fo[illegible], has rims of amphibole at plagioclase contact
Orthopyroxene, clinopyroxene, and magnetite	—	—	—	

Although the rocks of the Lake Owens mafic complex are the least altered of any mafic rocks in the Medicine Bow Mountains, the unit has not escaped metamorphism. Especially along the southern border of the complex, the rocks have a metamorphic foliation and are altered to amphibolite. The generally fresh nature of the rock types suggests that this unit may be younger than other mafic rocks of the area, but in as much as it is metamorphosed along the southern border, it is also possible that the massive body of mafic rock has simply resisted metamorphism except along borders.

The Mullen Creek mafic complex is located in T. 14 N., R. 80 W., and adjacent areas (Pl. 1). It is a large irregularly shaped body of mafic rocks of diverse composition covering an area of approximately 60 square miles. It is truncated on the northwest by the Mullen Creek-Nash Fork shear zone and is cut by numerous bodies of granite and quartz monzonite. It cuts hornblende gneiss along its eastern border but is so metamorphosed along its southern border as to appear to grade into the gneiss. In general, the rocks of the complex are much more highly deformed than those of the Lake Owens mafic complex, and except for local areas in the central part of the unit it is marked by secondary foliation.

This extensive body of mafic rocks is not well exposed through much of the outcrop area, and is located in one of the least accessible and most heavily timbered areas of the Medicine Bow Mountains. Mapping through most of the northern part of the complex is reconnaissance. We do not know if this is a series of multiple intrusions or one large body, but the evidence to date (i.e. gabbro cutting basalt) suggests that basalt was introduced first and then gabbro.

The rocks of the complex are metabasalt, olivine gabbro, gabbro, diorite and quartz diorite. Olivine gabbro is the most abundant rock type of samples that are so little metamorphosed that their original composition could be determined, but this comprises a relatively small part of the complex so the actual major rock type is not known. Most of the outcrops studied were coarse to medium grained so that the major part of the complex is a gabbroic rock of some type, but a number of localities were found where a fine-grained altered basalt was present that was cut by the gabbroic phase. Along the borders of the complex where the units are foliated and amphibolized much of the rock has the composition of diorite or quartz diorite. It is possible that this body is rimmed by quartz diorite on the west and north somewhat like the Lake Owens mafic complex. Especially along the northern border late granite and quartz diorite is nearly as common as more gabbroic rocks. Gabbro appears rich in potash feldspar and cataclastic textures and structures are common in all rocks. One area was studied in sec. 35, T. 14 N., R. 81 W. where the rocks of the complex appeared unmetamorphosed. Here the trend of the foliation was northwesterly and isolated outcrops suggested an interlayering of olivine gabbro, and anorthositic olivine gabbro. In all other areas studied, the foliation was secondary developed through cataclasis and recrystallization of the original gabbro.

The mineralogy of the various rock types of the mafic complex is shown in Table 51. The textures of the unaltered rocks are very similar to those of the Lake Owens mafic complex and the altered rocks like altered gabbro previously described.

Table 51—Mineralogy of Rocks of the Mullen Creek Mafic Complex (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
		Gabbro		
Plagioclase	2/2	55.0	53-57	An[illegible]
Clinopyroxene	2/2	22.8	22-24	Augite, Ext. 42, 2V = 65 (Est.)
Orthopyroxene	2/2	10.0	6-14	As discrete crystals and rims around olivine (Hyp.)
Olivine	1/2	1.8	—	Reaction rims at plagioclase contacts
Amphibole	2/2	7.2	3-11	After pyroxene as outer part of olivine reaction rim
Opaque minerals	2/2	1.8	—	Ilmenite
Apatite and chlorite	—	—	—	Chlorite as alteration of pyroxene and in veinlets cutting rock
		Olivine gabbro		
Plagioclase	3/3	56.4	50-60	An[illegible]
Clinopyroxene	3/3	15.7	10-20	Augite, $z \wedge c = 45°$
Orthopyroxene	3/3	7.7	6-10	Interstitial and as rims around olivine
Olivine	3/3	8.3	4-11	Reaction rims with plagioclase
Amphibole	3/3	7.4	5-11	
Opaque minerals	3/3	1.3	—	Magnetite and ilmenite
Biotite	1/3	1.3	—	Red-brown
Quartz	1/3	1.9	—	Secondary (?)
		Amphibolized gabbro (texture retained)		
Amphibole	5/5	46.1	37-54	Largely after pyroxene, outlines of pyroxene can be recognized in some slides. Nx = tan, Ny = green, Nz = blue-green, $z \wedge c = 15$, 2V = 60 (Est.)
Plagioclase	5/5	37.2	28-52	An[illegible] partly or wholly altered to zoisite, carbonate, clay mineral mixture.
Quartz	5/5	7.8	4-12	
Opaque minerals	5/5	4.7	3-7	Magnetite and ilmenite—altered to leucoxene and sphene
Epidote	3/5	1.7	0-3	Epidote after mafic minerals—clinozoisite after plagioclase
Chlorite	1/5	1.29	0-6	After pyroxene
Biotite	2/5	0.6	0-3	Associated with chlorite—after pyroxene
Sphene	3/5	0.4	0-1	
Talc	1/5	0.2	0-1	

GRANITE AND QUARTZ MONZONITE

There is a large number of small to medium-sized bodies of granite and quartz monzonite in the general area of the Mullen Creek mafic complex and east of this body where felsic units cut quartz diorite and the gneissic rocks (Pl. 1). These bodies of granite and quartz monzonite are medium-grained and foliated to varying degrees, and in sec. 17, T. 14 N., R. 78 W. northwest of Muddy Mountain and on the north end of Sheep Mountain they are cut by coarse-grained granite of the Sherman type. They are later than the gneisses and mafic igneous rocks, but earlier than the Sherman Granite. In addition to the granitic bodies noted above, it has been possible to separate other similar units that are located in the southern part of the mountains and are entirely within the gneissic complex. These units in the gneiss include some previously described as massive quartzo-feldspathic gneiss at Deerhorn Point, east of Fox Park and on Jelm Mountain, as well as smaller bodies such as those on Illinois Creek, secs. 26 and 27, T. 13 N., R. 79 W. All of these "granitic" bodies are mapped as older granite on Plate one. Whether they are all of the same age or are similar genetically is a question that will be considered below.

The character of these bodies of granite and quartz monzonite will be discussed by describing certain typical occurrences including the Big Creek granite, the Rambler granite, the Douglas Creek granite-quartz monzonite complex, the Prospect Mountain granite, the Park Run quartz monzonite sills, and by reviewing the character of the older granite previously described under massive-quartzo-feldspathic gneiss.[3] The bodies of older granite previously described are bodies that can be distinguished in the field by more massive texture and somewhat different coloration than the gneiss. These bodies of granite are more felsic in mineralogy, and richer in Na, K, and silica than the gneiss host rock. They have gradational contacts, are structurally conformable to the host rock, and contain remnants of host rock. The weight of the evidence suggests that they developed by replacement of the host rock.

The Big Creek granite is located in the north-central part of T. 13 N., R. 81 W., northeast and west of Big Creek (Pls. 1 and 3). It is in contact with hornblende gneiss on the south and cuts metagabbro of the Mullen Creek mafic complex on the north. Small dikes and sills of the same composition as the granite cut the metagabbro to the northwest and metagabbro and hornblende gneiss to the east. Contacts with hornblende gneiss are not well exposed, but on the western border of the granite contacts are marked by a transitional augen hornblende gneiss with large crystals of potash feldspar. Outcrops on the eastern border of the granite suggest that this is one granite body with interfingering contacts with gneiss (Pls. 1 and 3). It is possible, however, that this is a series of very closely spaced sills or dikes with thin screens of gneiss between them. Outcrops along Big Creek could be interpreted in this manner or as a zone of inter-fingering at the gneiss contact. Contacts with metagabbro are sharp and cross-cutting. The mineralogy of the granite is shown in Table 52, and in as much as the plagioclase is calcic oligoclase to andesine the granite actually ranges in composition from gran-

3. These granitic rocks are given informal names for descriptive purposes.

ite to quartz monzonite depending on the percentage of potash feldspar. All samples of the granite studied in thin section have a well defined foliation developed by cataclasis. The essential minerals are bent, fractured, and rotated and drawn into boudins, and biotite is commonly aligned parallel to the planes of fracturing.

Samples from the Big Creek granite have been dated by the Rb/Sr whole rock method (Hills, et. al., 1968). The six samples of granite do not define a single isochron, but lie on two isochrons indicating ages of 1470 ± 160 and 1715 ± 50 m.y.

Table 52—Mineralogy of granite and quartz monzonite (mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
			Big Creek "granite"	
Potash-Feldspar	3 3	46.0	43-48	Perthitic
Quartz	3 3	31.0	28-33	Large and small anhedral grains scattered through rock, also in veinlets cutting K-Feldspar. Strong undulatory extinction
Plagioclase	3 3	15.5	12-21	An[illegible]
Muscovite	3 3	2.9	1-6	In masses associated with biotite and epidote
Biotite	3 3	2.5	1-4	Greenish-brown partly altered to chlorite
Epidote	3 3	1.1	0.2-2	Inclusions of unknown radio-active mineral
Opaque minerals	2 3	0.8	—	Ilmenite and hematite
Sphene	2 3	0.2	—	
Zircon, garnet, chlorite	—	—	—	
			Douglas Creek granite — quartz monzonite complex	
Quartz	5 5	37.7	29-47	Micrographic texture in some samples
K-Feldspar	5 5	26.5	12-48	As large discrete crystals and as interstitial masses along grain boundaries
Plagioclase	5 5	27.0	16-47	An[illegible], may show reverse zoning
Biotite	4 5	4.0	0-7	May be altered to chlorite
Muscovite	5 5	2.2	0.2-5	
Epidote	5 5	2.4	0.3-5	
Opaque minerals	2 5	0.1	—	Magnetite
Apatite, chlorite, zircon, allanite, sphene	—	—	—	
			Prospect Mountain "granite"	
K-Feldspar	3 3	45.5	35-55	As large porphyroblasts and as interstitial aggregates
Quartz	3 3	23.2	16-29	Two types: rounded small grains and large anhedral grains
Plagioclase	3 3	19.6	19-20	An[illegible], some grains show reverse zoning, two generations of plagioclase in some rocks, secondary plagioclase fresh in appearance. Secondary plagioclase more calcic
Biotite	3 3	8.9	3-15	
Amphibole	2 3	0.9	0.2.5	Nx = tan, Ny = very dark green, Nz = dark blue green, $z \wedge c$ = 19, 2V = low
Epidote	2 3	1.3	0.2.7	In large crystals associated with biotite
Muscovite	2 3	0.3	—	
Sphene	2 3	0.1	—	
Opaque minerals	2 3	0.1	—	Magnetite
Zircon, apatite	—	—	—	Zircon as euhedral crystals

The Prospect Mountain granite is located east of the Big Creek granite in secs. 1 and 12, T. 13 N., R. 81 W., secs. 6 and 7, T. 13 N., R. 80 W. Like the Big Creek granite the Prospect Mountain granite is in contact with gabbro on the north and gneiss on the south. Contacts with gneiss are gradational and some contacts show a gradation from granite to porphyritic gneiss with augen of potash feldspar. There are many lenses of gneiss within the area mapped as granite.

Locally the granite is highly deformed with the development of cataclastic structure. In one area where contacts with gabbro could be seen (Fig. 24), the granite is clearly later than the gabbro, since stringers from the granite penetrate the gabbro. The most interesting aspect of the contact zone is the nature of foliation in the granite and gabbro. The granite is strongly foliated with the development of augen where crystals of potash feldspar have been deformed, but the gabbro is only faintly foliated, and even the stringers of granite in the body of gabbro are less well-foliated than the granite. There may be two explanations for this relationship:

1. The granite stringers cutting the gabbro may simply be rheomorphic stringers developed by melting at the gabbro contact.

2. The granite is later than the gabbro and both rocks have been deformed with the granite more easily deformed than the gabbro.

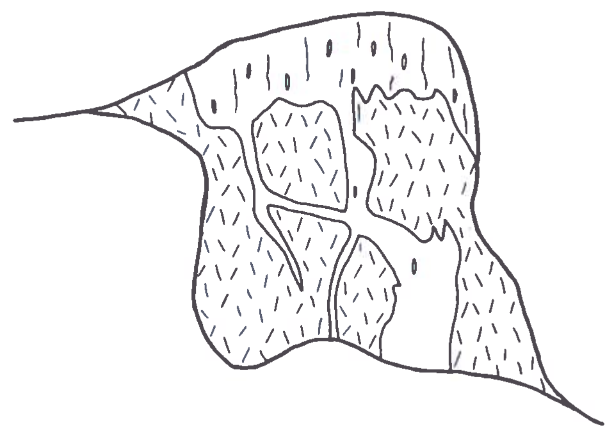


Figure 24—Field sketch showing relationship of gabbro of Mullen Creek mafic complex to granite at Prospect Mountain. Note that granite is porphyritic and sends stringers into faintly foliated gabbro (short dashes), but that granite is foliated with development of blastoporphyritic texture. Width of outcrop about two feet.

The latter explanation is more likely since the granite is usually more highly deformed than the gabbro even where large bodies of granite that are too large to have been formed by rheomorphism cut the gabbro. The mineralogy of the Prospect Mountain granite appears in Table 52.

The Douglas Creek granite-quartz monzonite complex is a series of interconnecting bodies of granite and quartz monzonite that cut gabbro of the Mullen Creek mafic complex and hornblende gneiss. The granite is in sharp contact with the gabbro and inclusions of gabbro in granite are common. As in the Big Creek granite and Prospect Mountain granite, contacts with hornblende gneiss are conformable and may be gradational. The granite is generally foliated and most thin sections show cataclasis, but there are some samples of granite in which biotite flakes are aligned and essential minerals show a crude preferred orientation that do not show evidence of cataclasis. Along upper Devils Gate Creek pegmatized varieties of granite are common. The granite is cut by stringers of pegmatite varying in width from one to ten inches that both parallel and crosscut the foliation. The amount of pegmatitic material may be as much as 50% of the rock. The mineralogy of the granite is in Table 53.

The Park Run quartz monzonite sills are located in secs. 26 and 27, T. 13 N., R. 79 W. near the junction of Park Run and Illinois Creeks (Pl. 1). These are a series of sills having the general form of phacoliths that are surrounded by hornblende gneiss. The quartz monzonite has sharp contacts with the hornblende gneiss and crosscutting dikelets of fine-grained quartz monzonite are common in the hornblende gneiss at the contact. The quartz monzonite is porphyritic having large porphyroblasts of microcline, and all samples studied are foliated.

The Rambler granite is in secs. 33 and 34, T. 15 N., R. 79 W., along Elk Creek (Pl. 1). It is shaped somewhat like the top of a mushroom with a dome-shaped top and two major appendages on either side. The body covers about one and one-half square miles and has a large gneissic zone in the central part. It is in sharp cross-cutting contact with the gabbroic intrusions and has gradational and conformable contacts with the quartzofeldspathic gneiss on its eastern side. Foliation present in the granite is the result of cataclasis. The mineralogy of the granite is in Table 53.

Table 53—Mineralogy of Rambler granite (after McCallum, 1964)

Mineral	Range in Percent	Comments
K-Feldspar	30-40	Perthitic, largely microcline
Quartz	30-40	As large equant grains and as fine-grained interstitial aggregate
Plagioclase	20-30	An_{13}, most grains partly altered to sericite
Biotite	1-3	Dark brown with layers of chlorite
Epidote	1-2	Associated with biotite and amphibole may have cores of orange-brown allanite.
Amphibole	1	
Opaque Minerals, Muscovite, and allanite	—	Opaque minerals largely magnetite

CRITIQUE ON GRANITIC ROCKS

It is possible that under the heading of older granite we have considered several entirely different type of rocks. A logical subdivision is to consider most of the granitic rocks in gneiss; typified by the Deerhorn Point body, as metasomatic, and to consider those bodies of granite that clearly cross-cut gabbro as magmatic, i. e. emplaced in a liquid or semi-liquid state. These "granites" may have formed at different times and at different levels in the crust. The writer has deliberately chosen not to do this so that the reader may note a general relationship between these rocks. We find, for example, that granite that is in contact with both gneiss and gabbro may show cross-cutting relationships with gabbro and conformable and even gradational contacts with gneiss. The nature of the contacts with the mafic igneous rock suggests the granite was emplaced in a liquid or semi-liquid state perhaps by forceful intrusion or stoping, but the relationship to the gneiss may suggest an origin by replacement or metasomatism.

Contrasting field relationships of the same granitic bodies are not unique to this area. Mercy (1963, p. 209-211) has reviewed some examples of granitic bodies that in one locale show field characteristics suggestive of replacement, and are cross-cutting elsewhere. He quotes Read (1958) as proposing that one can develop this relationship by partial mobilization of granitized (metasomatized) country rock, but offers an alternative (Mercy, 1963, p. 210) of magma emplacement in one area by forceful intrusion, in another by admission along planes of structural weakness (foliation and/or bedding) perhaps in a succession of episodes without disturbing the stratigraphic succession or structure of the host rocks to a great extent.

The choice between these two views of the origin of the older granite would depend on one's prejudice on the origin of granite. No matter how the granite formed it has been deformed subsequently with the development of cataclastic structures. The relatively young age for the Big Creek granite may also support later deformation (Hills, et. al., 1968) at least, the granite does not retain evidence of the 2.4 b.y. event for similar rocks at Baggot Rocks.

SMALL INTRUSIVE BODIES OF FELSIC COMPOSITION

Small dikes and sills of granite and aplite are common in the quartzo-feldspathic gneiss and hornblende gneiss of the southern area. The granitic bodies are similar in composition and general character to the granite just described. In fact, these bodies are commonly conformable in strike and dip to the adjacent gneiss and have gradational contacts. They are also affected by cataclasis as is the case for the larger intrusions. Many of these small granitic bodies appear to be emplaced in fault zones such as those in T. 13 N., R. 81 W., and the sills on Centennial Ridge (Pl. 1). The mineralogy of typical small granitic bodies is in Table 54.

Table 54—Mineralogy of granitic dikes
(visual estimates, volume percent)

Mineral	A	$\bar{X}$	Comments
K-Feldspar	4/4	36	Perthitic
Quartz	4/4	35	
Plagioclase	4/4	24	Av. comp. An[illegible], Range An[illegible]
Biotite	4/4	2	Green, partly altered to chlorite
Epidote, chlorite, opaque minerals, muscovite, amphibole, and zircon			

An exception to the general rule of composition conformity is the lenticular to pod-like bodies of luxullianite that are in the area north of the Rambler granite. These bodies are notably rich in black tourmaline and have a general composition as listed in Table 55. They are, however, similar to the other felsic intrusives because they are concordant with the gneiss and have a cataclastic foliation.

Table 55—Mineralogy of luxullianite dikes
(visual estimates, volume percent)
(after McCallum, 1964)

Mineral	Range in Percent	Comments
Quartz	20-30	Extreme undulatory extinction
Plagioclase	30-40	An[illegible], some antiperthitic
K-Feldspar	10-20	Largely perthitic microcline
Tourmaline	5-15	Pale to deep blue
Muscovite	3-8	May fill fractures in both quartz and feldspar

PEGMATITE

Pegmatites are common in the area south of the Mullen Creek-Nash Fork shear zone especially in the hornblende gneiss and related rocks (Pl. 1). The great majority of the pegmatites are simple conformable types composed of microcline, quartz, and plagioclase with muscovite, ilmenite, magnetite, garnet, and biotite as accessory minerals. These pegmatites are rarely over 10 feet wide and probably average less, but they may extend for distances 5 to 10 times their width along strike, and some of them are over one-half mile long. They have sharp contacts with the country rock but do not show a decrease in grain size at the contact. Pegmatites of this type are particularly abundant in an east-trending belt of hornblende gneiss that extends from one side of the Medicine Bow Mountains to the other and is confined to T. 13 N. (Pl. 1). The pegmatites are scattered throughout this belt but seem to be particularly concentrated in the axes of folds in the gneiss. Conformable pegmatites are also common on Centennial Ridge, and McCallum (1964, p. 121-124) has recognized a second type of conformable pegmatite in this area which is tourmaline rich and generally contains a higher percentage of accessory minerals, especially muscovite, garnet, and magnetite. A third type of conformable pegmatite is the sulphide-rich type that is uncommon, but a particularly good example is located on Big Creek in sec. 9, T. 13 N., R. 81 W. This pegmatite (Houston, 1961) is unusual because it contains gash veins and brecciated areas filled with coarsely crystalline quartz and chalcopyrite.

A second subdivision of pegmatite has cross-cutting relationships with the country rock and this pegmatite is generally more elliptical in plan. The length of these pegmatites may average three times their width. These pegmatites have sharp contacts with the country rock, and do not decrease in grain size at the contact, but they may have poorly developed zoning such as concentration of quartz pods in the center of the pegmatite or a poorly developed quartz core and a graphic granite outer border. Many are more complicated mineralogically; and, in addition to feldspar and quartz, contain such minerals as garnet, tourmaline, biotite, muscovite, fluorite, and rarely monazite, allanite, euxenite, and columbite. Pegmatites of this type are relatively common in the Big Creek area, T. 13 N., R. 81 W., and sec. 32, T. 13 N., R. 78 W., southwest of Fox Park (Pl. 1).

A third division of pegmatites is a replacement type found along the southwestern border of the Medicine Bow Mountains near the Colorado State line and in the North Park fluorspar district of Colorado (Steven, 1957). The replacement type of pegmatite has gradational contacts with the country rock, is not exceptionally coarse grained, is uneven texturally, is unzoned, and is characterized by simple mineralogy. According to Steven (1957, p. 350-351) most of these pegmatites are simple aggregates of quartz and feldspar.

A fourth and fifth division might be composed of the pegmatized areas in granite previously described and pegmatized areas in quartzo-feldspathic gneiss. The pegmatized areas in the quartzo-feldspathic gneiss have been described by Swetnam (1962) as zones of coarse grain size characterized by replacement potash feldspar and networks of intersecting quartz veinlets.

It seems probable that the pegmatites are not of the same age and that some pegmatites with similar field characteristics may have been formed during events greatly separated in time. It is the writer's view that the majority of the conformable and cross-cutting pegmatites formed or were introduced at approximately the time of formation of granite and quartz monzonite previously described. At least, one can be certain that they are not related to the coarse-grained granite of the Sherman type (this was originally proposed by the writer, Houston, 1961, p. 11, for the cross-cutting type) in as much as several of these pegmatites (notably the Platt pegmatite on Big Creek) have been dated as 1.6 b.y. years old and the granite of the Sherman type has been dated as 1.35 b.y. (Hills, et. al., 1968).

SHERMAN GRANITE

Perhaps the most distinctive and widespread igneous rock of the Medicine Bow Mountains is a pink, coarse-grained granite that closely resembles the Sherman Granite originally described by Blackwelder (Darton, Blackwelder, and Siebenthal, 1910) in the southern part of the Laramie Mountains of Wyoming. In his map of the Laramie-Sherman area of Wyoming, Blackwelder noted that the granite on the eastern edge of Sheep Mountain was of the Sherman type, and he concluded that there was one great batholith underlying much of southeastern Wyoming. The boundaries of the Sherman granite have been better defined since Blackwelder's mapping by Hagner and Newhouse (1957), Harrison (1951), and by our current study of the Medicine Bow Mountains (Fig. 25). Although little is known of the southwestern extent of the granite there is no doubt that it extends beneath the cover of the Laramie Basin since it is found in wells of the basin between the Boulder Ridge occurrence of the western Laramie Mountains and the occurrences of the eastern Medicine Bow Mountains (Fig. 25). Further support for the one granite concept can be had from age determinations made in both the southern Laramie Range (Aldrich, et. al., 1958; Ferris and Krueger, 1964) and the Medicine Bow Mountains (Hills, et. al., 1968) that fall in the same range of 1.35 b.y.

The character of the Sherman Granite is best understood by reviewing studies of the granite in both the Laramie Mountains and Medicine Bow Mountains, and the following will be a composite picture of the batholith as we know it today. The Sherman Granite is the last major igneous rock emplaced in the area. It cuts all metamorphic and igneous rocks previously described in this report and is later than anorthosite of the Laramie Mountains. It is cut by dikes of felsic and mafic composition that are best exposed in the Boulder Ridge area of the southwestern Laramie Mountains. Contacts between the granite and country rock are commonly sharp and irregular, and the granite is generally coarse-grained to the contact. There are areas where inclusions of country rocks are common along contacts and small rounded inclusions are scattered throughout the intrusion. Blackwelder (1910, p. 5) notes areas where the country rock is brecciated at the contact and the granite sends stringers and complex networks of veinlets into the country rock and has numerous angular misoriented blocks of country rock in it. There are areas, however, especially in finer-grained plagioclase-rich varieties of the granite where contacts are gradational, and Harrison (1951) has noted gradational contacts both parallel to and along strike.

There are three major types of granite in the areas mapped as Sherman Granite. The most abundant type is a pink, coarse-grained granite that is eguigranular and has crystals that average 5 to 6 m.m. in diameter. The second type is a medium-grained grey granite that is rich in plagioclase (calcic oligoclase) and is classed as quartz monzonite by Harrison (1951). A third type is a porphyritic granite that has large phenocrysts of microcline in a medium-grained to coarse-grained groundmass. Students of the Sherman Granite have described contacts between these various rock types as gradational and the distribution as random, but in the Medicine Bow Mountains the porphyritic phase is more common along borders and in apophyses.

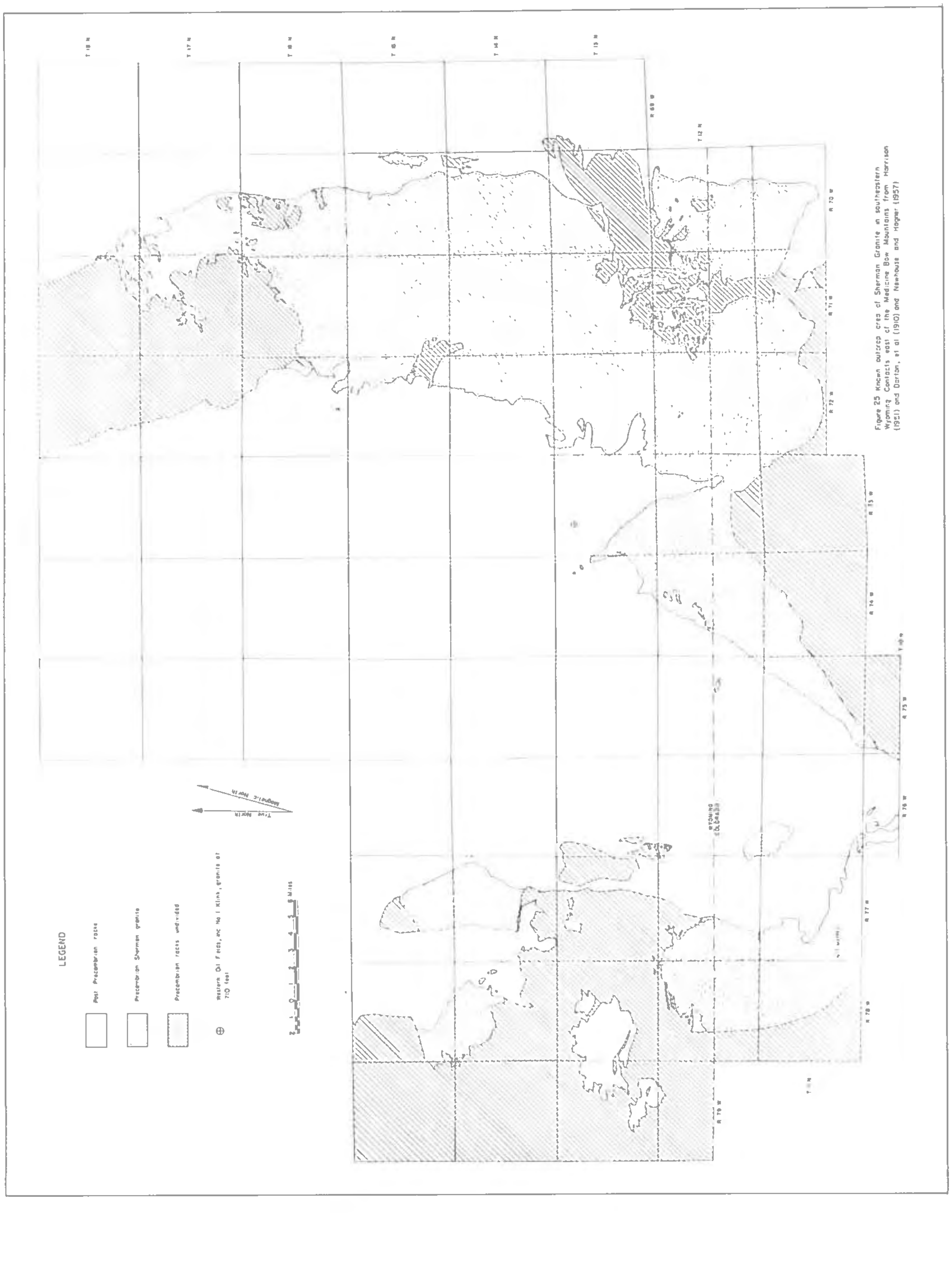


Figure 25—Outcrop area of Sherman Granite.

Harrison (1951) notes a well-developed foliation in the granite of the Laramie Mountains especially in the quartz monzonite variety. This is developed by a subparallelism of microcline crystals. Harrison (1951) states, however, that this foliation is poorly developed and difficult to recognize in the coarse-grained and porphyritic varieties of the granite. In the Medicine Bow Mountains the granite appears to be nonfoliated except where it has been subjected to a mild cataclasis. Comments on the field characteristics of the granite would not be complete without mention of the distinctive mushroom-shaped outcrops so common where the coarser-grained facies of the granite are exposed (Fig. 26).

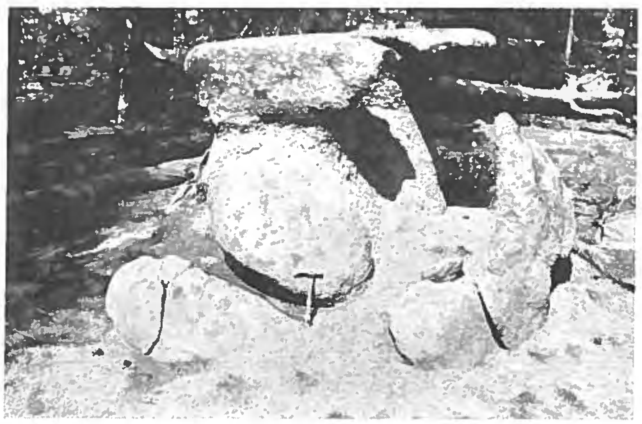


Figure 26—Mushroom-shaped outcrop typical of more massive phases of the Sherman Granite. Photograph by D. R. Currey.

The petrography of the granite has been studied in most detail by Harrison (1951, p. 55, table 2) and McCallum (1964c, p. 119) and Table 56 lists the mineralogy of typical samples of pink, coarse-grained granite from the southern Medicine Bow Mountains. Certain textural features of the granite noted by both Harrison and McCallum are significant. Quartz, microcline, plagioclase, and biotite are present in two generations, plagioclase commonly shows reverse zoning with a clouded interior of composition An_{26} and a clear exterior of composition An_{33}. Many crystals of microcline have rims of plagioclase around them resembling the rapakivi texture of Sederholm (1923), perthite is common, and myrmekite is present at contacts of plagioclase and microcline.

Table 56—Mineralogy of Sherman Granite, Medicine Bow Mountains
(pink, coarse-grained type)
(mode in volume percent)

Mineral	A	$\bar{X}$	Range	Comments
K-Feldspar	6/6	38.1	(25-53)	Mostly perthitic microcline, some slides show micrographic texture. Early altered K-feldspar may be replaced by late perthitic microcline
Quartz	6/6	34.0	(29-47)	May be two generations, early clouded grains with abundant inclusions, and late clear grains often interstitial
Plagioclase	6/6	20.5	(6-25)	Range An[illegible], two generations in some samples, early clouded and younger clear. Early plagioclase more calcic. Some crystals show ocillatory zoning, others have clear borders of more calcic plagioclase
Amphibole	6/6	2.5	(1-3)	Nz deep blue-green
Biotite	6/6	3.4	(2-6)	Greenish-brown
Opaque minerals	4/6	0.6	—	Magnetite and ilmenite
Epidote, apatite, allanite, sphene, chlorite, and zircon				

Another significant feature of the Sherman Granite is its relation to the felsic and mafic dikes common in the Boulder Ridge area and well-exposed in the Union Pacific Railroad cuts in the southern Laramie Mountains. The felsic dikes include aplite and fine-grained porphyritic granite, and the mafic dikes are metadiabase. These dikes are in north to northwest trending swarms and cut all rocks of the granite. Harrison (1951) has noted that the aplite dikes are typical cross-cutting dikes with fine-grained borders, but that in some areas on Boulder Ridge where these dikes are exposed in cross section the granite corrodes and actually sends stringers completely across the dikes. This same type of relationship has been noted for the metadiabase dikes by the writer where coarse-grained granite cuts the diabase.

Perhaps the last chapter in the history of the Sherman Granite is local fracturing where the granite is cut by small parallel shear zones. These zones may be the site of alteration of the granite to unakite. In some cases, hematite is also introduced and rarely chalcopyrite and pyrite. The granite is also cut by a few pegmatites; one notably rich in allanite crops out near Albany in the Medicine Bow Mountains.

We are in early stages of the study of the Sherman Granite and the above review gives some indication of the overall nature of the Sherman Granite and its great complexity.[1] One cannot help but be impressed by the similarity of this granite to Precambrian granites of Finland that

1. At the time this paper was in press Eggler (Eggler, D. H., 1968, Virginia Dale Precambrian ring-dike complex, Coloradio-Wyoming: Geol. Soc. American Bull., v. 79, p. 1545-1564) published results of a detailed study of a root zone of a ring-dike complex in the southeastern part of the Sherman Granite. He presents evidence to suggest that early gabbroic igneous activity was followed by formation of a ring-dike complex and introduction of successive magmas ranging from diorite to granite. This complex sequence of emplacement of magma may explain the seemingly anomolous structural relationships between mafic and granitic rocks described above.

have been described in detail by Sederholm (1923, 1926).

LATE FELSIC DIKES

North to northwest trending felsic dikes cut the central and southeastern part of the Medicine Bow Mountains (Pl. 1). These dikes cut all rocks, including pegmatites, with the exception of the Sherman Granite. They are dense, light-colored, fine-grained rocks that appear fresh in hand specimen. The dikes are resistant to erosion and stand above country rocks as narrow ridges. One northwest-trending dike located between Albany and Keystone is exposed over a distance of six miles along strike (Pl. 1).

The groundmass of the dikes is a feathery graphic intergrowth of feldspar and quartz. The dikes have medium-sized crystals of magnetite, chlorite, muscovite, amphibole, and biotite and larger crystals of K-feldspar, oligoclase and quartz (Table 57). The groundmass is too fine-grained and altered for satisfactory identification of feldspars, but the rocks are classed as porphyritic quartz latite on the basis of composition of the larger crystals.

REVIEW OF IGNEOUS ROCKS

The various "igneous" rocks both north and south of the Mullen Creek-Nash Fork shear zone present an almost bewildering variety. It seems worthwhile to review and compare these units so they can be better related to tectonic events that will be discussed later.

The most ancient "igneous" rocks north of the shear zone are augen gneiss and quartz diorite. These rocks may have formed during some metamorphic event and could be classed as synkinematic. "Granite" such as that at Baggot Rocks, although more felsic in composition (at least, in part) shows cross-cutting relationships to the gneiss country rock and may be later than the conformable augen gneiss and quartz diorite, but may still be a synkinematic intrusion. It is conceivable that these "igneous" bodies owe their generally conformable characteristics to later metamorphism and recrystallization that makes them appear to have characteristics of a "synkinematic" intrusion. The age determinations on the "granite" at Baggot Rocks, however, do appear to support of synkinematic origin for this unit.

These units are certainly old "igneous rocks" and are cut by dikes of mafic composition. They therefore predate one period of intrusion of basaltic magma. Because there are probably several episodes of intrusion of basaltic magma, it is difficult to relate individual igneous bodies of "gabbroic" composition to those of "granitic" composition.

Igneous rocks south of the shear zone are in some respects similar to those of the north. The most ancient "igneous" rocks may be quartz diorite and perhaps some "older granite" and certain massive varieties of quartzo-feldspathic gneiss. These rocks could certainly be classed as synkinematic or formed during a metamorphic event. In this area, however, certain bodies of "gabbroic" composition such as rocks of the Mullen Creek mafic complex and many of the sills of amphibolite are probably older than these synkinematic intrusions.

Some of the older "granite" masses south of the shear zone are clearly later than quartz diorite and gabbro. The "granite" on Big Creek that cuts gabbro is strongly foliated and may either be synkinematic or a deformed and recrystallized granite. If this granite is synkinematic and if pegmatites nearby were formed in the same metamorphic event we have an illustration of a remarkably abrupt change in the ages of granite on either side of the shear zone—2.4 b.y. to 1.6-1.7 b.y.

Certainly the most intriguing aspect of the igneous bodies of gabbroic composition is the possibility that many of the amphibolite sills south of the shear zone equate to dikes of basaltic composition north of the fault. This also implies an abrupt transition from one side of the shear zone to the other. Other bodies of gabbro and basalt may be younger than the older granite especially dikes and sills in the shear zones, but generally where older granite and rock of mafic composition is in contact the granite is younger.

The Sherman Granite is post kinematic and is the most recent igneous rock of Precambrian age with the exception of certain felsic dikes best shown in the Medicine Bow Mountains and mafic dike swarms perhaps best exposed on Boulder Ridge east of the Medicine Bow Mountains.

So far, the information we have gathered from geologic study and from age determinations suggests that the felsic igneous rocks south of the shear zone belong to an entirely different geologic series than those of the north. The majority of the "older granite" and pegmatites may have formed in a metamorphic event or events between 1.5 and 1.7 b.y., some 700 million years after the major granite making event north of the shear zone. It is also very likely that some of the highly metamorphosed mafic igneous rocks south of the shear zone correlate with much less deformed and metamorphosed units north of the fault. Metamorphosed felsic units equivalent to those north of the fault may also be present in the south but either we have selected the wrong "igneous" bodies to date or the events south of the fault were severe enough to wipe out evidence of an earlier age.

Table 57—Mineralogy of typical late felsic dikes (mode in volume percent)

Mineral	Percent	Comments
Plagioclase	39.0	An[illegible], strongly zoned large crystals, very fine-grained crystals in groundmass
Quartz	27.2	As large crystals and in groundmass. Micrographic feldspar quartz intergrowth in groundmass
K-Feldspar	10.6	As large crystals and in groundmass
Biotite	7.3	As large crystals but smaller in size than feldspar and quartz
Chlorite	6.2	As large crystals but smaller in size than quartz and feldspar
Muscovite	5.2	As large crystals but smaller in size than quartz and feldspar
Opaque Minerals	4.4	Large crystals of magnetite euhedral smaller in size than quartz and feldspar

CATACLASTIC ROCKS

Rocks deformed by crushing, and recrystallization accompanying crushing are classed as cataclastic. It is not always possible to determine what portion of a cataclastic rock is formed by crushing or grinding and what portion is formed by recrystallization during the grinding process, but Williams, Turner, and Gilbert, (1954, p. 199-208) have proposed that the term cataclasite be used for a rock deformed by crushing and grinding, the term mylonite be used for a rock subjected to crushing and grinding in the extreme so that the resulting product is very fine grained, and the term phyllonite be used where the reduction in grain size is accompanied by recrystallization of minerals. In the Medicine Bow Mountains cataclastic rocks are most common in the great shear zones of the central part of the mountains and cataclasites, mylonites, and phyllonites are present in these zones.

With the possible exception of the late felsic dikes all rock types of Precambrian age in the Medicine Bow Mountains have been subjected to cataclastic deformation in some degree. In general, this has been noted in the descriptions of the rocks. Commonly this cataclastic deformation is not confined to zones in the rocks nor is it the major textural feature of the rock. It is difficult to recognize in the field and is commonly noted only in thin section. There are, however, many areas usually associated with shear zones and faults where the rocks have been subjected to more concentrated cataclastic deformation and it is these areas that are the subject of this review. The areas of cataclastic deformation are shown by short dashes on plate one. It should be noted that while the major areas of cataclastic rocks are in or associated with faults and shear zones, there are belts of cataclastic rocks where crushing and grinding have taken place in the rock with little or no offset.

For a general review of the characteristics of the cataclastic rocks emphasis will be placed on rocks associated with the Mullen Creek-Nash Fork shear zone since these rocks are typical of those observed elsewhere in the Medicine Bow Mountains, and have been studied in most detail. The fault system which includes the Mullen Creek-Nash Fork shear zone is a series of northeast-trending shear zones that coalesce in the central part of the mountains into one great fault, the Mullen Creek-Nash Fork shear zone that has been traced from the northwest quarter of T. 13 N., R. 81 W. to the northwest quarter of T. 16 N., R. 78 W., a distance of 27 miles. There are four northeast trending shear zones in addition to the major one (Pl. 1). The three northernmost shear zones, including the Mullen Creek-Nash Fork shear zone are in a general zone of cataclastic rocks. This zone is three and one-half miles wide in the northeast and wedges out to the southwest where the three shear zones join (Pl. 1). The arrangement of the shear zones is such that the cataclastic zone is divided into two wedge-shaped bodies. The northern wedge adjacent to the Mullen Creek-Nash Fork shear zone is a cataclastic migmatite (Pl. 1). The rock unit is a biotite-plagioclase-quartz gneiss with numerous layers of pink granite, averaging 3 to 6 inches in width (Fig. 27). The biotite-plagioclase-quartz gneiss is cataclastic with textures as shown in Plate 24. The granite layers are only slightly effected by cataclasis and are regarded by McCallum (1964, p. 78-88) as introduced after the main period of cataclasis. The southern wedge is largely composed of biotite augen gneiss with large composite eye-shaped

A

B

C

Plate 24—A. Well foliated augen gneiss showing numerous lenticular eyes and some eye merging. Note the presence of a few thin layers of granitic material (light layers). Located near Bear Lake in sec. 26, T. 16 N., R. 79 W. B. Strongly sheared augen gneiss with pronounced schistocity. Augen are markedly granulated and stretched, and small lenses of merged eyes are quite prominent (crude layers developed). Located on north canyon wall of Libby Creek canyon in sec. 23, T. 16 N., R. 79 W. C. Strong layering developed by passive flow. Remnants of early fold systems preserved in some layers. Located north of Libby Creek in sec. 24, T. 16 N., R. 79 W. (Photographs by M. E. McCallum.)

A B

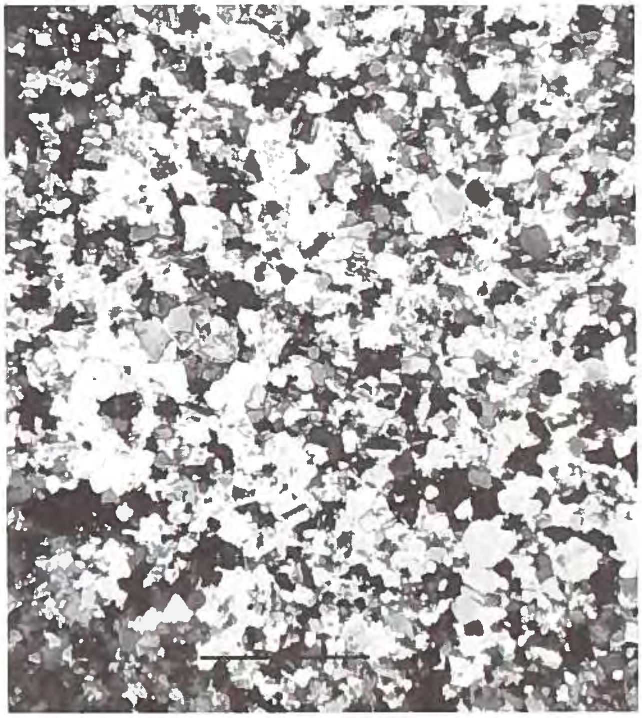
Figure 27—A. Coarsely layered gneiss or veinite of the cataclastic migmatite unit. Darker areas between light granitic layers are chiefly well-foliated biotite augen gneiss. Located along north shore of Silver Run Lake in sec. 26, T. 16 N., R. 79 W. B strongly layered and silicified mylonite in the central prong of the Mullen Creek-Nash Fork shear system, sec. 33, T. 16 N., R. 79 W. Photographs by M. E. McCallum.

masses of potash feldspar and plagioclase in a matrix of biotite, muscovite, and granulated feldspar and quartz. South of this wedge there are local cataclastic areas in the rocks, but with the exception of several prominent shear zones, rock units are not marked by extensive crushing or granulation.

The shear zones proper are planar zones of more intense granulation with bodies of mylonite that have the general appearance of very fine-grained quartzite in mylonized gneiss and slate or phyllite in mylonized mafic rock. A typical cross-section through the major shear zone is exposed in T. 14 N., R. 81 W., where this fault crosses the North Platte River. The rocks north of the shear zone are quartzo-feldspathic gneisses, and those south of the shear zone are igneous rocks of gabbroic and dioritic composition. From the north side of the shear zone across a zone approximately 1000 feet wide, quartzo-feldspathic gneiss grades into augen cataclasite, augen cataclasite to layered cataclasite, and layered cataclasite to exceedingly fine-grained mylonite having the appearance of fine-grained quartzite (Pl. 25D; Fig. 27). On the south side, the mylonite grades into mylonized mafic rock having a slaty appearance, slaty mylonite to a rock having the appearance of phyllite (phyllonite), then to foliated potash-rich quartz diorite, and finally into undeformed igneous rock. This latter zone is approximately 1200 feet wide. The area affected by shearing is nearly one-half mile wide at this point.

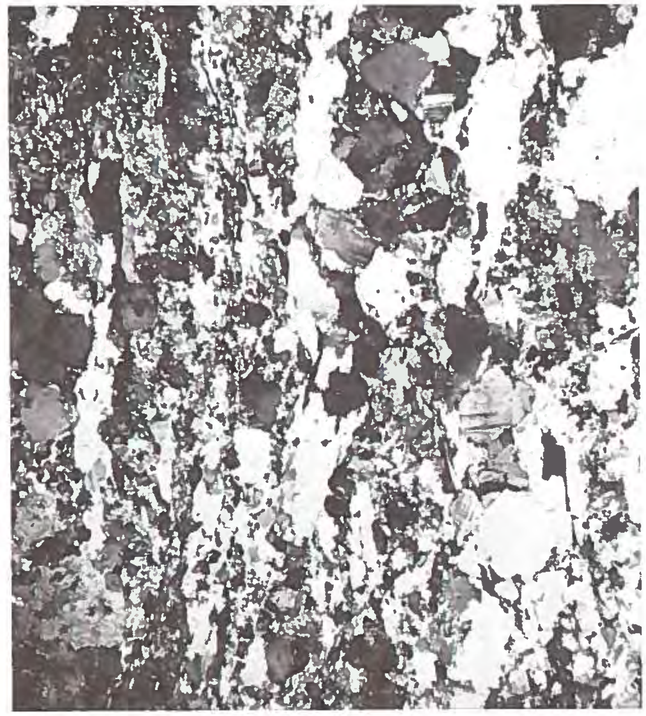
Sills and dikes of both mafic and felsic composition are in general areas of cataclastic rocks. The majority of these are basaltic in composition and are sills that may or may not be affected by cataclasis. Obviously these units were introduced after deformation, but many were affected by later shearing, some drawn into boudins.

In the Medicine Bow Mountains the cataclastic rocks are most commonly derived from quartzo-feldspathic gneiss and mafic igneous rocks and are rarely from metasedimentary rocks. Cataclasites derived from mafic igneous rocks and metasedimentary rocks are easily recognized in the field because of striking textural differences from the original rock. In isolated outcrops, cataclasite and mylonite derived from mafic igneous rocks might look like slate or phyllite, but the units can usually be traced to undeformed bodies of mafic rocks. Cataclasite derived from quartzo-feldspathic gneiss presents a particular problem in identification, because of its close similarity in texture and general field appearance to normal



A

Typical quartzo-feldspathic gneiss from north of shear zone; essentially undeformed. Crossed-nicols. Line 2 mm. long, all photos same scale.



B

Cataclastic quartzo-feldspathic gneiss with parallel crushed zones, lenticular masses of quartz, but with areas not completely reduced. Crossed nicols.

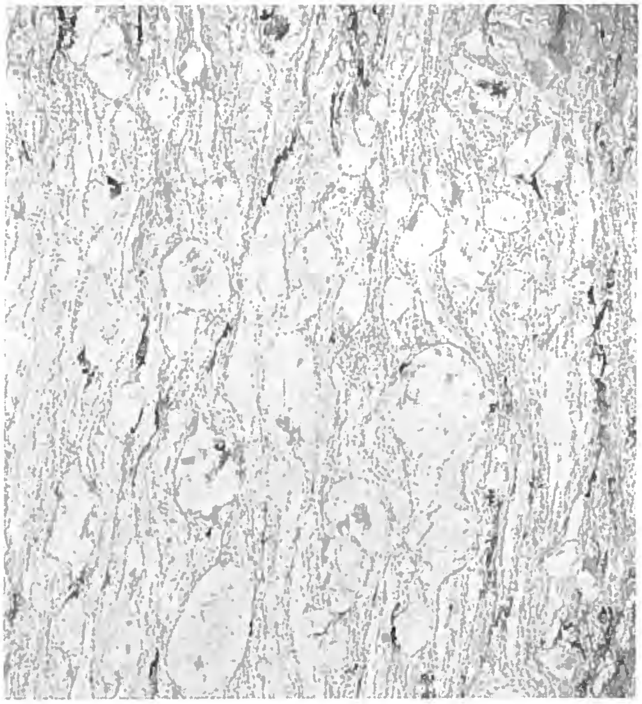
C

Cataclastic gneiss with typical lenticular masses; shattered but not yet completely reduced to mylonite. Plain light.

D

Gneiss near true mylonite with well-developed layering and exceedingly fine-grained layers. Few relics of original gneiss left.

Plate 25—Photomicrographs of cataclastic rocks from the Mullen Creek-Nash Fork shear zone on west side of Platte River showing stages in reduction of gneiss to layered mylonite.



A

Cataclastic quartzo-feldspathic gneiss from outcrop northeast of University of Wyoming Science Camp, sec. 13, T. 16 N., R. 79 W. Note rolled grains of feldspar, plain light.

B

Same as A with crossed-nicols.

C

Cataclastic gneiss from Mullen Creek-Nash Fork shear zone showing late fractures filled with epidote. Plain light.

D

Cataclastic gneiss from Mullen Creek-Nash Fork shear zone showing brecciated cataclasite with breccia zone filled with hematite. Plain light. Line 2 mm. long, all photos same scale.

Plate 26—Photomicrographs of cataclastic rocks.

gneiss. It is particularly difficult to recognize when original gneissic foliation parallels foliation resulting from cataclasis. However, most cataclasites developed from quartzo-feldspathic gneiss are distinguished from normal gneiss by the habit of quartz. In normal quartzo-feldspathic gneiss, quartz is usually as discrete grains which may or may not be elongate. In the cataclasite quartz is commonly in parallel stringers which are discontinuous, but are 10 to 20 times longer than they are wide. Such cataclasite may also contain spherical grains and lens-like bodies developed during the grinding (Pl. 26 A & B).

Major changes seen under the microscope are textural, but mineralogical changes are found especially in the cataclasites developed from mafic igneous rocks. The least deformed cataclasite derived from quartz diorite is marked by lenticular masses of partly crushed quartz diorite separated by crushed areas of fine-grained minerals (Pl. 27A). The fine-grained areas are marked by an abundance of chlorite, quartz, and biotite. In some samples, muscovite and potash feldspar is present. These rocks grade into true phyllonites in which there has been further reduction in grain size accompanied by recrystallization. The phyllonites are composed primarily of fine-grained quartz, chlorite, plagioclase, and muscovite but may have small lenticular masses of feldspar that has not been completely reduced during the grinding process (Pl. 27B). These phyllonites may have quartz veinlets parallel to the foliation and layers exceptionally rich in chlorite. Some are marked by complex folding of the folation surfaces in which the limbs of the folds are sheared and quartz is introduced along the new surfaces. Mineralogical and textural changes like those described above have been noted in cataclastic hornblende gneiss, amphibolite, and metabasalt (Pl. 27C). These phyllonites derived from mafic rocks approximate the mineralogy of rocks of the green-schist facies indicating retrograde metamorphism of rocks from the almandine amphibolite facies, or development of rocks of greenschist facies from previously unmetamorphosed igneous rocks. The mineralogical changes also suggest introduction of water, silica and potassium since quartz, potash-bearing minerals and hydrous minerals are much more abundant in the cataclasite.

Cataclasites developed from quartzo-feldspathic gneisses show somewhat similar textural changes but less pronounced mineralogical changes than those of the mafic rocks. Less deformed cataclastic quartzo-feldspathic gneiss has lenticular remnants of gneiss in a matrix of fine-grained quartz, feldspar, biotite, and minor chlorite. The lenticular masses may be composite bodies of quartz and feldspar or single crystals of potash feldspar or plagioclase and usually consist of partly broken mineral grains (Pl. 25B). Parallel, discontinuous veinlets of quartz are distributed throughout the rock. More deformed gneiss is marked by a strong parallelism and greater elongation of the lenticular bodies as well as long stringers of quartz parallel to the foliation. Some grains are circular and surrounded by fine-grained crushed material (Pl. 25B). Further deformation results in the development of mylonites with pronounced layering (Pl. 25D). These rocks are marked by layers of quartz alternating with layers of quartz, feldspar, sericite, epidote, and chlorite. Some quartz layers are medium-grained, and other layers are fine-grained. These rocks contain rare subrounded to rounded grains of plagioclase and quartz (Pl. 25D). The final products of crushing are dense, hard rocks composed of submicroscopic material. There is evidence in the rocks of introduction of quartz and increase in the proportion of hydrous minerals in the more highly deformed units, but the mineralogical changes do not indicate an increase in potash. Plagioclase of the undeformed gneiss is calcic oligoclase, but in the cataclasites oligoclase and albite have both been identified in the fine-grained portion of the rocks. Apparently the quartzo-feldspathic gneiss has locally been reduced to the greenschist facies of regional metamorphism during the process of cataclasis.

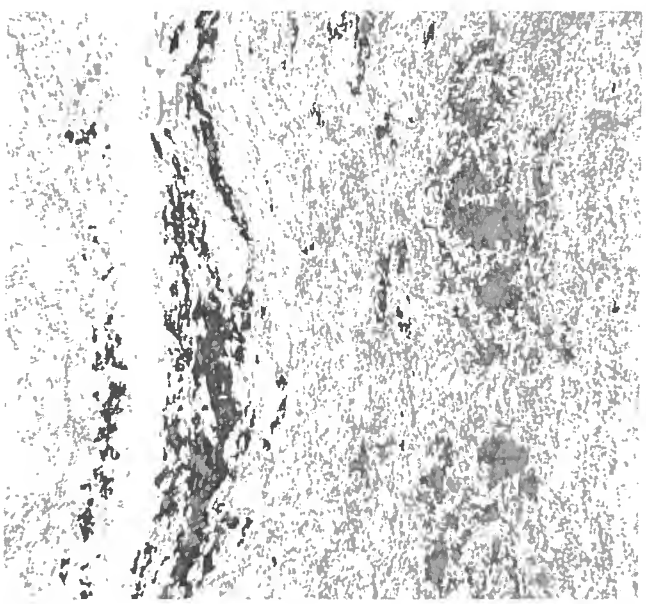
Some concept of the chemical changes in the gneiss can be had by reference to Table 58. Sample A is slightly crushed quartzo-feldspathic gneiss; sample B is a strongly deformed gneiss with well-developed lenticles in a crushed matrix, and sample C is mylonite. The chemical changes are those predicted from the mineralogy except that the mylonite is silicified to a high degree. Sample D is slaty mylonite developed from an igneous rock. Here again the chemical changes are as predicted from the mineralogy if one ac-

A

Cataclastic quartz diorite from margin of Mullen Creek mafic complex on west side of Platte River. Lenticular masses of plagioclase are relics of original rock. Plain light. Line 2 mm. long, all photos same scale.

B

Same as A but rock reduced to "phyllite". Relics of original diorite still recognizable but most mafic minerals transformed to chlorite epidote mixture. Rock classed as phyllonite; looks like phyllite in field. Plain light.



C

Mylonized basalt sill from Mullen Creek-Nash Fork shear zone. Strongly foliated, grains reduced in size and mineralogy of lower almandine amphibolite facies. White layer is lenticular body of quartz. Fine-grained matrix consists of epidote, quartz chlorite and amphibole. Plain light.

D

Cataclastic texture of late granite dikes north of the Mullen Creek-Nash Fork shear zone. Sample from sec. 27, T. 14 N., R. 82 W. Crossed-nicols.

Plate 27—Photomicrographs of cataclastic rocks.

cepts the average tonalite as typical of the rock that has been deformed.

Table 58—Chemical composition of cataclastic rocks from the Mullen Creek-Nash Fork shear zone exposed in T. 14 N., R. 81 W. where the fault crosses the North Platte River

	A	B	C	D	E
SiO_2	72.22	74.32	89.56	70.12	66.15
Al_2O_3	14.54	14.33	5.60	15.31	15.56
Fe_2O_3	0.85	0.47	0.16	1.83	1.36
FeO	0.45	0.24	0.47	1.75	3.42
MgO	0.46	0.30	0.27	0.97	1.94
CaO	1.31	0.86	0.36	0.55	4.65
Na_2O	3.68	4.20	2.28	2.78	3.90
K_2O	5.29	4.55	0.99	3.54	1.42
TiO_2	0.10	0.04	0.10	0.29	0.62
MnO_2	0.02	0.01	0.01	0.04	0.08
CO_2	0.06	0.03	0.02	0.16	?
H_2O+	0.84	0.75	0.53	2.65	0.69
H_2O-	0.39	0.47	0.22	0.48	

A Slightly crushed quartzo-feldspathic gneiss.
B Strongly deformed gneiss with lentiels in a crushed matrix.
C Mylonite developed from gneiss.
D Mylonite developed from tonalite.
E Average tonalite after Nockolds.
Analyst: Shirou Imai, Japan Chemistry Research Institute.

Metasedimentary rocks deformed by cataclasis are quartzite, slate, and metadolomite. The quartzite is simply more finely ground and may be marked by hematite along fractures. The slate is reconstituted structurally and silicified. The most interesting changes are in the metadolomite. McCallum (1964, p. 129-130) has noted a number of lenticular bodies of tremolite-talc marble and serpentine marble in the Mullen Creek-Nash Fork shear zone. These en echelon masses of marble may have developed from previously larger masses of siliceous metadolomite of the Nash Fork Formation caught up in the fault zone. The tremolite-talc marble contains approximately 75% calcite, 10% dolomite, 10% tremolite, and 5% talc with traces of opaque minerals, rutile, and antigorite, and the serpentine marble contains approximately 70% calcite, 29% antigorite, and 1% olivine (Fo_{82}) and opaque minerals. McCallum (1964, p. 130) views these rocks as developed by dedolomitization of siliceous dolomites of the Nash Fork Formation. The tremolite of the tremolite-talc marble is thought to have developed by reaction between dolomite, silica, and water to form tremolite, calcite, and carbon dioxide under conditions approximating the greenschist facies of regional metamorphism set up during the deformation. Under higher temperature conditions (almandine amphibolite facies of regional metamorphism) tremolite may have reacted with excess dolomite to form forsterite. Because both tremolite and forsterite are partly converted to antigorite, a final period of alteration under lower temperature conditions probably followed the major period of cataclasis.

From the above discussion, one can surmise that the process of crushing and grinding that results in the development of the cataclasite and mylonites is accompanied by introduction of constituents and may take place under lower temperature and perhaps higher water vapor pressure than those of earlier periods of regional metamorphism. In some examples the temperature may have approximated that of the almandine amphibolite facies but the majority of the rocks have been reduced to greenchist facies. These rocks of the shear zones have probably been deformed a number of times under varying conditions of temperature and pressure, but the major imprint is that of a late period of crushing and grinding under temperature-pressure conditions approximating the green schist facies, i.e., no less than 300 degrees Centigrade and 3000 bars load pressure (Fyfe, Turner, and Verhoogen, 1958, p. 166-173).

As a final note in the discussion of cataclasites one must point out certain areas especially in the quartzo-feldspathic gneiss of the northwest part of the Mountains and in hornblende gneiss in the south where zones and belts of quartzo-feldspathic gneiss or hornblende gneiss are commonly at large angles to the general structural trend. The gneiss in these belts may have textures somewhat like those described above, especially the lenticular bodies of quartz, but show no evidence of retrograde metamorphism and are completely recrystallized. These zones are formed in much the same manner as those described above, but under temperature-pressure conditions that allow complete recrystallization of the rocks and that approximate those of prior periods of metamorphism.

ROCKS OF PALEOZOIC AND MESOZOIC AGE

Geologic mapping of rocks of Paleozoic and Mesozoic age in the Medicine Bow Mountains was done primarily by Beckwith (1938, 1941, 1942). Beckwith's work was along the flanks of the mountains in the southeastern and northwestern part. Additional mapping includes work of Thomas (1928) in the Centennial area, Bauer (1951) near Arlington, Ashley (1947) in the northwest and more recently Gries (1964) in the Pass Creek Basin. Although the major emphasis of the present program of mapping has been the study of rocks of Precambrian age, Paleozoic and Mesozoic rocks have been mapped in order to complete the geology and to better understand relations between structures of Precambrian age and those of Laramide age. D. L. Blackstone, Jr., of the University of Wyoming has mapped sedimentary rocks along the northeast flank of the Mountains and H. D. Thomas, late State Geologist of Wyoming, assisted greatly in interpreting the stratigraphy of units of Paleozoic and Mesozoic age. Additional stratigraphic studies have been made of these units by Thomas (1936, 1951, 1953, 1957), Maughan (1963, 1964), Mallory (1960), Knight (1929), Pipiringos (1957), and Shaw (1953). Reviews of the structure and stratigraphy are presented in the Guidebook for the eighth annual field conference of the Wyoming Geological Association and the University of Wyoming and in Wyoming Stratigraphy (Burke, 1956).

The above reports treat the stratigraphy of the rocks of Paleozoic and Mesozoic age in such detail that it need not be repeated here, but the general lithology is given in Table 59. Notes on the stratigraphy pertinent to discussions of structure will be given below.

DISCUSSION OF PALEOZOIC-MESOZOIC MAP UNITS

Most of the Paleozoic-Mesozoic formations are defined according to standard nomenclature of Burke (1956), but some formations have been redefined in recent years so it is necessary to discuss certain formations briefly to clarify the map units.

MISSISSIPPIAN, PENNSYLVANIAN, AND EARLY PERMIAN

The stratigraphy of rocks of Mississippian, Pennsylvanian, and Early Permian age is complicated by facies changes and erosional unconformities from south to north in the Medicine Bow Mountains (Fig. 28). In the southern Medicine Bow Mountains arkosic sandstone and conglomerate of the Fountain Formation lie on rocks of Precambrian age. These rocks are probably middle and late Pennsylvanian in age (Maughan and Wilson, 1960). The Fountain Formation is overlain by a distinctive cross-bedded sandstone unit called the Casper Formation. On the southeast margin of the Medicine Bow Mountains the Casper Formation consists of a lower cross-bedded sandstone, a middle shale, and an upper cross-bedded sandstone. The Casper Formation of this area is equivalent to that described by Knight (1929, p. 49-78) and is early Permian in age (Maughan and Wilson, 1960, p. 41) (Fig. 28).

The Fountain Formation is approximately 400 feet thick along the east-central and southeastern margin of the Mountains, but in the general area north of the Mullen Creek-Nash Fork shear zone these rocks are absent or only present as arkosic lenses less than 10 feet thick in the Amsden Formation. The Casper Formation is variable in thickness ranging from 483 feet on the west side of Sheep Mountain to 208 feet near Centennial (Knight, 1929, p. 52).

In the northwestern part of the Medicine Bow Mountains the Mississippian Madison Limestone lies on rocks of Precambrian age. The Madison Limestone includes a basal arkosic conglomeratic sandstone, a middle unit of limestone, and an upper unit of yellow-gray siltstone (Maughan, 1963, p. 623). This unit thins from about 55 feet at Elk Mountain to patchy outcrops of limestone a few feet thick along the east flank of the Mountains near Cooper Hill and along the west flank of the Mountains in outcrops at Elk Hollow Creek.

In the Elk Mountain-Coad Mountain area the Madison Limestone is overlain by dark red shales and siltstones of the Amsden Formation. These distinctive red beds can be found in the western Medicine Bow Mountains as far south as Elk Hollow Creek, but the Amsden Formation is thinner in the south and especially in the Mullison Park area, sec. 3, T. 16 N., R. 81 W. it contains beds of arkose and arkosic sandstone. In the northeastern Medicine Bow Mountains near Arlington the red shales and siltstones of the Amsden Formation are not well developed, and to the southeast the Amsden Formation is missing (Fig. 28). On the geologic map (Pl. 1) the Madison Limestone and Amsden Formation are shown as a single map unit.

The Casper Formation of the northern Medicine Bow Mountains can best be defined in the Elk Mountain-Coad Mountain area where the distinctive cross-bedded sandstone overlies red beds of the Amsden Formation and is in turn overlain by red beds of the Goose Egg Formation. In this

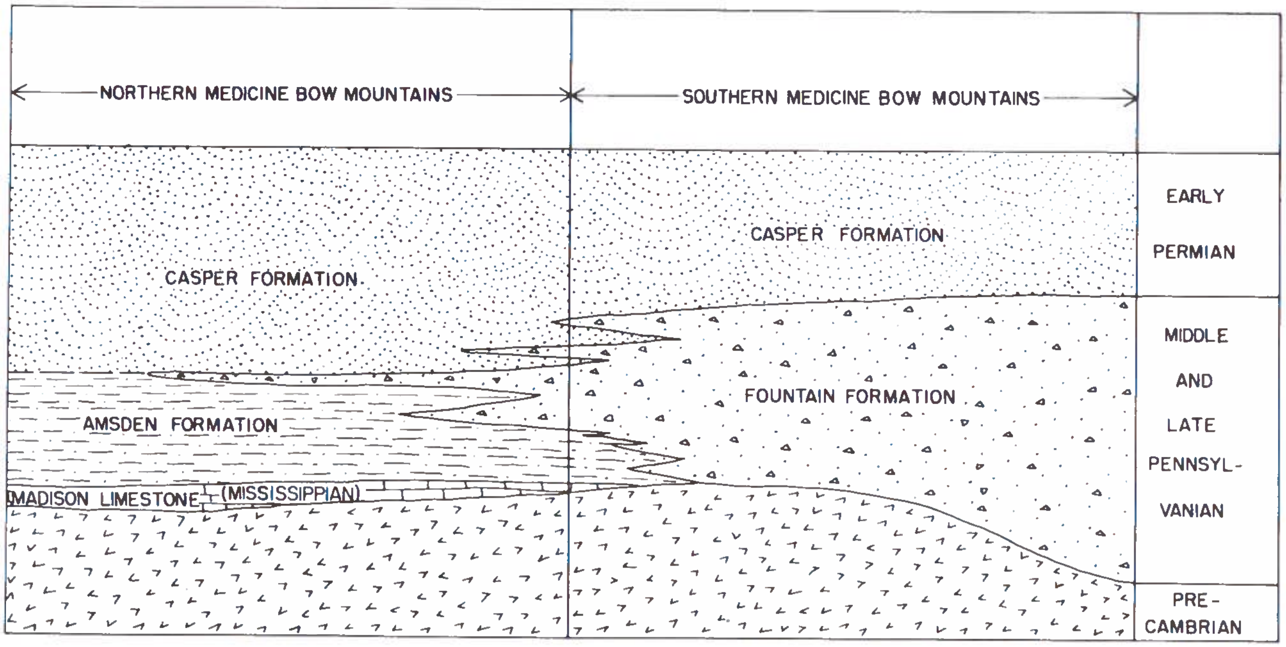


Figure 28—Stratigraphic relationships of Mississippian, Pennsylvanian and Early Permian formations of the Medicine Bow Mountains. Modified from Thomas (1957, p. 6, Fig. 7).

area the Casper Formation may include thin beds of limestone especially near the base. In this area the Casper Formation probably ranges in age from late Pennsylvanian to early Permian (Fig. 28). It has been mapped as Tensleep Formation by Beckwith (1941). The problem in terminology involving the Tensleep and Casper Formations is discussed by Maughan (1967, p. 138-139).

In the northeastern Medicine Bow Mountains near Arlington the Casper Formation lies on beds that equate to the Mississippian Madison Limestone and Pennsylvanian Amsden Formation. These beds are much thinner than to the northwest and include light gray limestone of the Madison Limestone and red siltstones and limestones that may equate to the Amsden Formation. In some areas (southeast of Arlington) the Madison-Amsden interval is a conglomerate a few feet thick consisting of cobbles and pebbles of Madison Limestone in a mudstone matrix. The Casper Formation of this area is Pennsylvanian in the lower part and early Permian in the upper part. The lower part is more calcareous and finer grained and the upper part is a distinctive cross-bedded sandstone typical of the formation. According to Hyden and others (1967) the lower and upper Casper Formation is separated by a unconformity that marks the boundary between the Pennsylvanian and Permian part of the Formation. The Casper Formation of this area is as much as 550 feet thick (Hyden and others, 1967).

PERMIAN-TRIASSIC

Permian rocks of the Medicine Bow Mountains also exhibit changes in lithology from north to south. In the north red beds that overlie the Casper Formation in the Elk Mountain-Coad Mountain area are more like the Goose Egg Formation of the type locality (Fig. 29) but to the south and east the upper limestone and evaporite members of the Goose Egg Formation are missing. The Forelle Limestone Member of the Goose Egg Formation is well developed throughout the Medicine Bow Mountains and has been used to subdivide map units both in the north and south. In the north we have mapped a restricted Goose Egg Formation that includes units above the Casper Formation and below the Difficulty Red Bed Member (Fig. 29). In the south the Forelle Limestone and Satanka Shale have been mapped as Satanka-Forelle undivided.

The Chugwater Formation as mapped here includes red siltstones, shales, limestone, and evaporites in the lower part that are of Permian age (upper Goose Egg) and an upper sequence of reddish-orange siltstones, shales, and sandstone that are Triassic in age (Fig. 29). In many areas

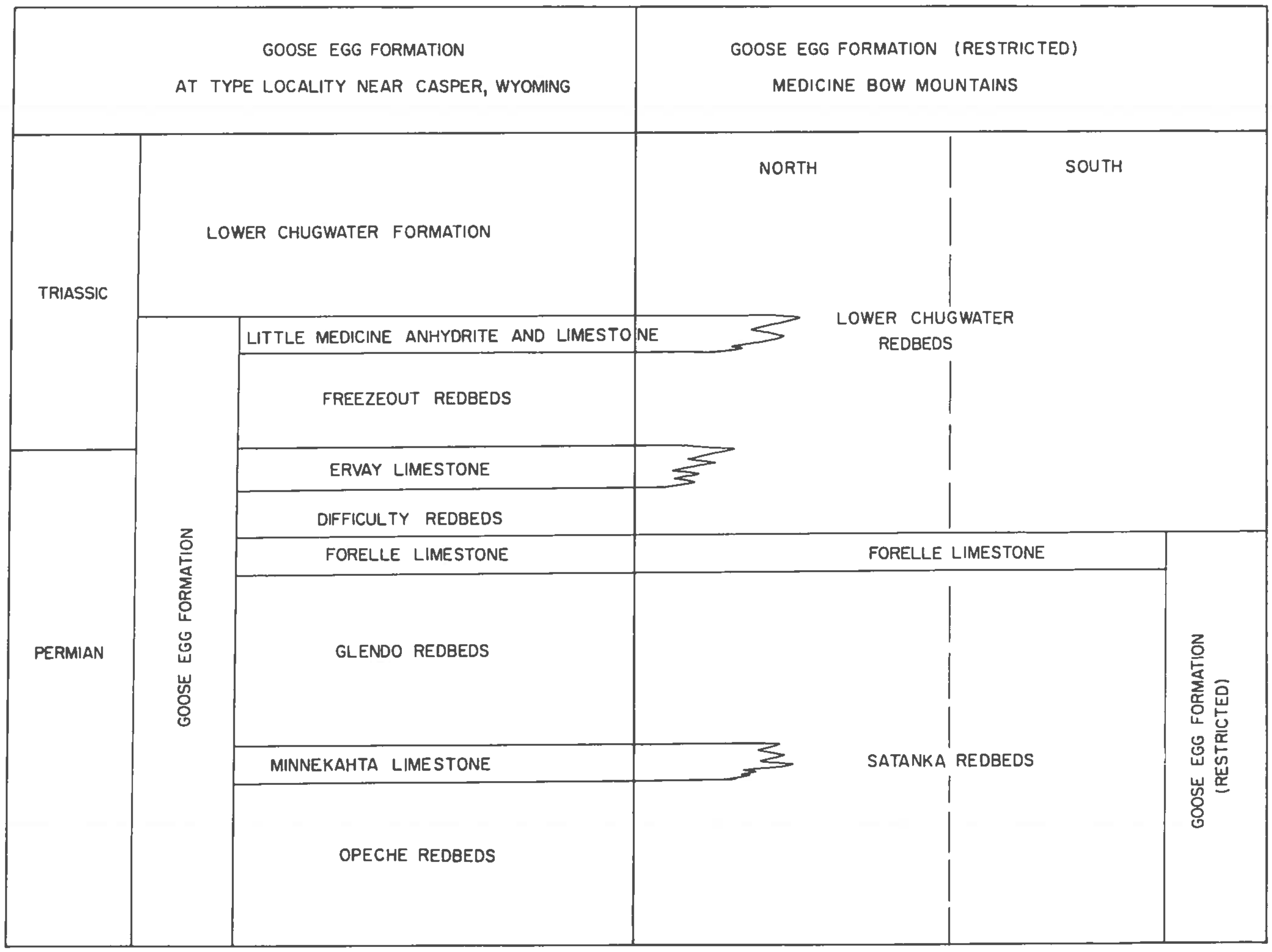


Figure 29—Stratigraphic relationships of Late Permian-Early Triassic formations of southern Medicine Bow Mountains. After Burk and Thomas (1956) and Maughan (1964).

these red sandstones, siltstones, and shales could not be assigned to a specific formation especially in isolated outcrops. These beds were mapped as Permo-Triassic red beds undivided (Pl. 1).

LATE TRIASSIC

Throughout the Medicine Bow Mountains the Chugwater Formation is overlain by a sequence of orange siltstones and sandstones that superficially resemble the rocks of the Chugwater. At the base these beds have a pebble conglomerate composed of pebbles of limestone, shale, wood fragments, and fragmentary remains of vertebrate fossils. These rocks have been named the Jelm Formation (Knight, 1917, p. 168) and as originally defined by Knight and mapped by Beckwith, the pebble conglomerate is placed at the base of the Jelm Formation, and the unit includes approximately 250 feet of siltstone and sandstones that underlie the rocks of the Sundance Formation. The Jelm Formation can be established as Triassic in age by phytosaur fragments that have been found in the Jelm of the Laramie Basin (personal communication P. O. McGrew, 1964). It can be further designated as Late Triassic because rocks equivalent to the Jelm of the Laramie Basin overlie the Alcova Limestone (Burke, 1953, p. 29-33) that has been dated as Late Triassic (Zangerl, 1963, p. 117-125).

LATEST CRETACEOUS-EARLY TERTIARY

Rocks of Latest Cretaceous and Early Tertiary age include the Medicine Bow Formation, Ferris Formation, Hanna Formation, and Wind River Formation. Especially along the borders of the Medicine Bow Mountains these formations

Table 60—Lithology of the Medicine Bow Formation at Citizens Coal Mine
(NE¼, Sec. 29, T. 16 N., R. 77 W.)
After S. H. Knight

Ferris-Hanna Formations (undivided)
(angular unconformity)

Lithology	Thickness in feet
Covered (probably green shale)	50+
Conglomerate, light brown to dark reddish brown, massive, matrix, askosic, gravelly, sandstone; gravel consists of fragments of feldspar, quartz, and small fragments of silicified sandstone (Cloverly Formation), silicified wood (Morrison Formation), siliceous shale (Mowry Formation), silicified pebble conglomerate (Cloverly Formation), silicified fossil bone (Morrison Formation), chert (Morrison and Cloverly Formations), quartzo-feldspathic gneiss and granite (Precambrian), some cobbles are present that are largely fragments of silicified sandstone and conglomerate from the Cloverly Formation and Precambrian quartzo-feldspathic gneiss, gravel and cobble size clasts are subrounded	15
Sandstone, tan to gray, medium to coarse-grained, forms prominent ridge	55
Sandstone, tan to dark brown, medium-grained, forms prominent ridge	6
Shale, brown, gray, tan, carbonaceous locally, interbedded sandstone	63
Sandstone, brown to reddish-brown, medium to coarse-grained, ferruginous, forms ridge	6
Sandstone, tan to gray, medium-grained, carbonaceous, leaf imprints locally	148
Limestone, tan, composed of oysters	5
Sandstone, brown and gray, medium- to fine-grained, interbedded carbonaceous shale, clams abundant, *Corbula* ? sp., probably brackish water	40
Sandstone, brown and gray, medium- to fine grained, interbedded carbonaceous shale and thin (6" to one foot) coal seams, leaf imprints in sandstone	140
Coal (poorly exposed)	3
Total	530+
Conformable Contact with Lewis Shale	

contain abundant conglomerate, conglomeratic sandstone, and arkosic sandstones and are marked by disconformities, angular unconformities, and channeling. With the exception of the Medicine Bow Formation, the lithology of the flank facies (i.e. lithologic facies of the formations near the Mountain uplift) is very similar and the writer has not been able to subdivide the Ferris, Hanna, and Wind River Formations. The conglomeratic flank facies of the Ferris, Hanna, and Wind River Formation is shown on the geologic map (Pl. 1) as Ferris-Hanna undivided. These rocks probably range in age from Paleocene to early Eocene. Their complex lithology reflects the rapid erosion of rocks of the uplifted Medicine Bow Mountains during the Laramide orogeny. The coarse-grained

Table 61—Lithology of the Upper Lewis Formation, Medicine Bow Formation, and Ferris-Hanna Formation (undivided) at Mill Creek locality
(Sec. 7, T. 16 N., R. 77 W.)

Ferris-Hanna Formations, undivided (top covered, dip 25 degrees)

Lithology	Thickness in feet
Sandstone and conglomerate, sandstone, white to gray, medium to coarse-grained, locally arkosic, cross-bedded, thin lenses of carbonaceous shale and lignite, conglomerate, gray to light brown, very poorly sorted, matrix, arkosic sandstone, clasts, variable in size up to 1.5 feet in diameter, average cobble size, largely rounded to subrounded quartzite (Precambrian) but includes petrified wood (Morrison Formation ?), pink quartzo-feldspathic gneiss (Precambrian), vein quartz (Precambrian), silicified shale (Mowry Formation), and silicified sandstone and conglomerate (Cloverly Formation). Both conglomerate and sandstone poorly consolidated but removal of cobbles from conglomerate leaves distinct molds. Conglomerate is channeled into sandstone and discontinuous, channels up to four feet deep	20
Shale, white, non-fissile, grades upward to dark-gray lignitic shale (Pollen sample no. Sp-5A and Sp-5B)	24
Conglomerate, gray, very poor sorting, matrix arkosic sandstone, clasts; largely gravel size	5
Siltstone, white to purple, mottled, massive, shows jointing perpendicular to bedding	43
Conglomerate, white to gray, poorly sorted, matrix, white sandstone, clasts, mostly gravel size	3
Siltstone, white, massive, non-bedded, jointing perpendicular to bedding	42
Total	137+
Angular unconformity (contact between Ferris-Hanna and Medicine Bow Formation not exposed, top of Medicine Bow Formation overlapped by beds of the Ferris-Hanna Formation)	
Medicine Bow Formation (top overlapped, dip 60-70 degrees)	
Shale, gray, locally carbonaceous and lignitic (covered)	45
Sandstone, white to light gray, very fine-grained, thin shale and lignite layers	5
Shale, white to gray (largely covered)	43
Sandstone, brown, fine-grained, massive, locally iron-stained layers, small iron-stained concretions	5
Shale, gray to black, lignitic (largely covered)	60
Sandstone, gray, fine grained, locally iron stained layers small ironstone concretions	7
Shale, gray, lignitic	6
Sandstone, white, fine grained to medium grained, ironstone concretions	14
Shale, light gray, thin lignite layers	10
Sandstone, gray, fine-grained	2
Shale, light gray, thin lignite layers	27
Sandstone, gray, fine-grained, iron stained	6
Shale, white to gray, carbonaceous, locally thin lignite layers	7
Sandstone, white to gray, medium-grained	4
Shale, gray, silty	5
Sandstone, brown, fine-grained, slabby parting, calcareous, iron-stained	2
Shale, black, highly carbonaceous	6
Sandstone, white to gray, fine-grained	1
Shale, gray to black, highly carbonaceous	6
Sandstone, gray, fine-grained with beds of siltstone and gray shale	10
Shale, black, carbonaceous	2
Sandstone, white, medium to fine grained, cross-bedded	5
Shale, black, lignitic	4
Sandstone, white, medium to fine grained	5
Shale, black, lignitic	7
Sandstone, white, cross-bedded	6
Shale, gray	3
Sandstone, gray, fine grained	2
Shale, steel blue, four-inch layers of lignite	4
Sandstone, white to gray, fine grained cross-bedded	6
Shale, steel gray, carbonaceous (Pollen Sample No. SP-4B D3506-B)	1
Coal, (Pollen Sample No. D-3506-A SP-4A)	2
Shale, gray blue, carbonaceous	1
Sandstone, light gray, fine-grained, friable, cross-bedded	4
Shale, black to steel gray, locally lignitic	11
Shale, dark gray	6
Sandstone, light tan, medium to fine-grained, beds of siltstone	5
Shale, dark gray, highly carbonaceous	15
Shale, light tan, sandy	10
Shale, dark gray, highly carbonaceous	25
Sandstone, tan, medium-grained, ironstone concretions	3
Siltstone, buff to tan, three feet of dark gray shale at base	14
Sandstone, tan, very fine-grained, one foot rusty-brown, calcareous layer at top	5
Shale, dark gray, grading upward for nine feet to siltstone, at nine feet above base two feet thick gray calcareous siltstone, siltstone becomes sandy upward with upper three foot interval, light gray, thin bedded silty sandstone	14
Shale, dark gray to dark brown with ironstone concretions at base, 13 feet above base brown papery shale with jarosite coating	27
Siltstone, light gray, thin-bedded	20
Shale, dark gray, interlayered siltstone and fine-grained sandstone	45
Sandstone, buff, very fine grained, thin bedded, glauconitic	9
Shale, dark gray, interlayered with silty shale and brown thin bedded siltstone, ironstone concretions in zones 17 to 22 feet above base	35
Shale, gray to brown, becomes more silty towards the top	9
Coal	1
Total	576+
Conformable Contact with Lewis Shale	
Lewis Shale	
Siltstone, dark brown, sandy, very thin-bedded, carbonaceous	1
Siltstone, light gray, thin bedded, brown carbonaceous shale laminae	7
Shale, dark gray, silty, with interbedded siltstone and fine-grained calcareous sandstone (very poor exposures)	936
Shale, dark gray, locally silty and sandy, scattered light brown, thin-bedded siltstone layers (Pollen Sample No. SP-3)	55
Thickness to base of Lewis Formation	1,624
Total	2,621

conglomeratic facies was probably formed in alluvial fans and the disconformities and angular unconformities developed as a result of deformation accompanying the uplift.

Medicine Bow Formation

The marine shales of the Lewis Formation are overlain conformably by a succession of brackish water and non-marine shales, sandstones, and coal seams correlated by Knight (1953, p. 67) with the Medicine Bow Formation because these rocks are lithologically similar and are in the same stratigraphic position as the rocks of the Medicine Bow Formation of Late Cretaceous age named by Bowen (1918, p. 229-230) in the Hanna Basin of Wyoming. The Medicine Bow Formation crops out along the margin of the Mill Creek syncline northeast of Centennial and northwest of Rex Lake along the east central flank of the Mountains (Pl. 1). These units are overlain unconformably by rocks of Paleocene and Eocene age. The top of the Medicine Bow Formation is not exposed in the area studied because of onlap by rocks of Paleocene and Eocene age, but the lithology of the lower part of the formation is given in Tables 60 and 61.

Following Bowen (1918, p. 230) the contact between the Lewis Shale and Medicine Bow Formation has been placed below a series of well-developed coal seams in the Medicine Bow Formation. At the type locality, the Medicine Bow Formation consists of alternating shale and massive to thin-bedded, ripple-marked, and cross-bedded sandstone with several beds of coal in the lower third of the unit. Lithologically, it is very similar to units on the margin of the Mill Creek syncline described in Tables 60 and 61. In addition, Bowen (1918, p. 230) states that the Medicine Bow Formation in the Hanna Basin contains remains of fresh and brackish water invertebrates, land plants, and bones of vertebrates. No vertebrate fossil remains have been found in the Mill Creek area, but fresh water clams, **Corbula** and well preserved leaf imprints are common in the formation especially on the south side of the Mill Creek syncline. Two samples from a coal and black shale 259 feet above the base of the Medicine Bow Formation (Table 61) have been examined by Robert H. Tschudy of the United States Geological Survey for spores and pollen. According to Tschudy:

> Both samples yielded Late Cretaceous pollen and spore assemblages. The dominant taxon was a monosulcate pollen with characteristic reticulate sculpture. This taxon is common to abundant in the Cretaceous Medicine Bow formation near its type locality. In both samples this taxon exceeded 25% of the total pollen and spore flora. Other pollen species characteristically limited to Upper Cretaceous were found. These include species of **Proteacidites** and **Aquilapollenites** and a new genus we have coded as C3-rt30. More than 25 distinct taxa were recognized in each of the two samples. No fossils characteristically limited to the post-Cretaceous were seen.

Thus the Medicine Bow Formation of the Mill Creek syncline seems to correlate on both lithologic and paleontologic grounds with that of the type locality of the Hanna Basin, and is Late Cretaceous in age. The thickness of the unit in the Mill Creek syncline may differ considerably from that of the type area where the Medicine Bow Formation is 6,200 feet thick. In the Mill Creek syncline all that can be said is that the formation is greater than 576 feet thick.

Recently Hyden, McAndrews, and Tschudy (1965, p. K3-K8) have described a unit called the Foote Creek Formation which crops out north and east of the area included in this report, in sec. 13, T. 20 N., R. 79 W., and sec. 18, T. 20 N., R. 78 W., Carbon County, Wyoming and along the north part of the Laramie Basin. At its type locality NW 1/4 sec. 18, T. 20 N., R. 78 W., the Foote Creek Formation measures only 228 feet thick, but lithologically it is quite similar to the Medicine Bow Formation of the Mill Creek syncline. In fact, an oyster bed is present about 86 feet above the base of the formation that strongly resembles the bed found in the Medicine Bow Formation along the south margin of the Mill Creek syncline. Furthermore, **Corbicula fracta** is reported in the macrofossil suite of the formation. In addition, a coal near the base of the formation about seven feet above the contact with the Lewis shale contains an assemblage of pollen and spores of Late Cretaceous age like that of the lowermost coal of the Medicine Bow Formation near its type locality. Except for thickness, the Foote Creek Formation at its type locality is very similar to the Medicine Bow Formation at its type locality and in the Mill Creek syncline. Hayden, McAndrews, and Tschudy (1965, p. K3) prefer to use a different name for the unit because of the great difference in thickness and a slight difference in age. The age difference is indicated because samples from the Foote Creek Formation in a drill hole in sec. 32, T. 19 N., R. 77 W., yielded excellent palynologic specimens of Paleocene age.

Whether we continue to use the name Medicine Bow Formation or in some areas the Foote Creek Formation, it is clear that following deposition of the Lewis Formation the sea began to retreat from this area, and in place of the marine shales of the Lewis Formation, brackish water muds and sand with associated peat were deposited in paralic swamps. In the Medicine Bow Mountain area, the uplift was sufficient to expose the Precambrian to erosion. This is shown by the presence of fragments of Precambrian quartzo-feldspathic gneiss and granite in the conglomerate of the Medicine Bow Formation at Citizens Coal Mine (Table 60). As noted by Knight (1953, p. 67), the presence of this conglomerate in beds only 515 feet above marine Cretaceous rocks is the first positive evidence of local uplift during the Laramide Orogeny. The conglomerate is dominated by fragments of the more resistant rocks of the underlying Paleozoic and Mesozoic formations, and fragments of Precambrian rocks make up less than 10% of the total (Table 60, Fig. 30). Clasts of quartzite that are the dominant constituent of conglomerates that overly the Medicine Bow Formation are not present. This indicates that the source of the Precambrian rocks was either in the southern Medicine Bow Mountains or local, perhaps in the Corner Mountain area, but the Precambrian of the northern Medicine Bow Mountains where quartzite is presently the major rock type could not have been exposed.

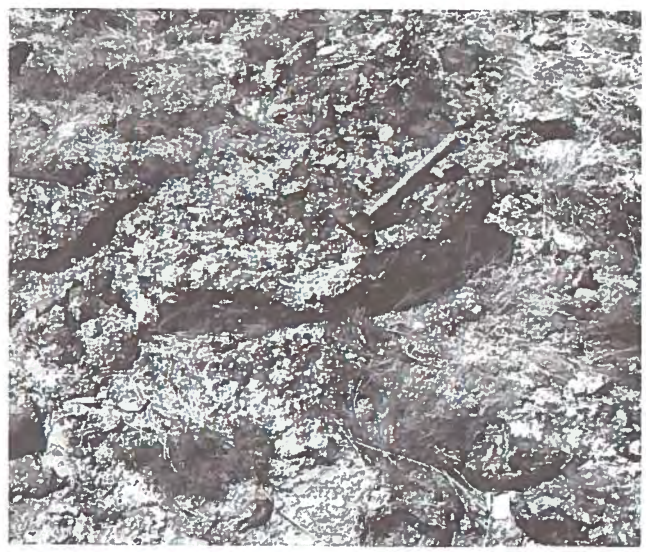
We must emphasize that the deposition of the Medicine Bow conglomerate is part of a continuing uplift that probably began in Lewis time so the initial rise of the Mountains was in the Cretaceous. Deposition was continuous (without an angular unconformity) since the conglomerate lies conformably on older rocks of the Medicine Bow Formation, but since the conglomerate is high in the section and dated samples come from the lower one-half of the section, we cannot be certain that the conglomerate was deposited in Cretaceous time.

Ferris-Hanna Formations, Undivided

The term Ferris-Hanna Formations, undivided is used to designate the conglomeratic flank facies of rocks of Paleocene-Eocene age that lies with angular unconformity on the Medicine Bow Formation (T. 16 N., R. 77 W.). This unit is thought to include coarse-grained facies of the early Eocene Wind River Formation in the upper part, but in T. 17 N., R. 77 W., typical finer-grained feldspathic sandstone of the Wind River Formation lies with angular unconformity on the lower beds of the Ferris-Hanna Formation. It is thus possible to demonstrate two periods of deformation, one prior to deposition of the beds of the Ferris-Hanna and one during deposition (Fig. 31).

Beds of the Ferris-Hanna Formations, undivided, are widespread in the northern Medicine Bow Mountains (Pl. 1). In the northeast they are in the Mill Creek syncline of T. 16 N., R. 77 W. where the thickness may exceed 3000 feet and in a major syncline bordering the Arlington fault that extends northwest from the south fork of Mill Creek to sec. 13, T. 19 W., R. 79 W., north-

A



B

Figure 30—A. Outcrop of Medicine Bow Formation showing almost right angle change in strike of unit in sec. 29, T. 16 N., R. 77 W. (See Plate 1). Conglomerate located on far side of resistant bed in upper left. B. Conglomerate located 515 feet above base of Medicine Bow Formation and composed largely of fragments of resistant rock types of underlying Mesozoic sedimentary rocks. White fragments are silicified shale of Mowry Formation.

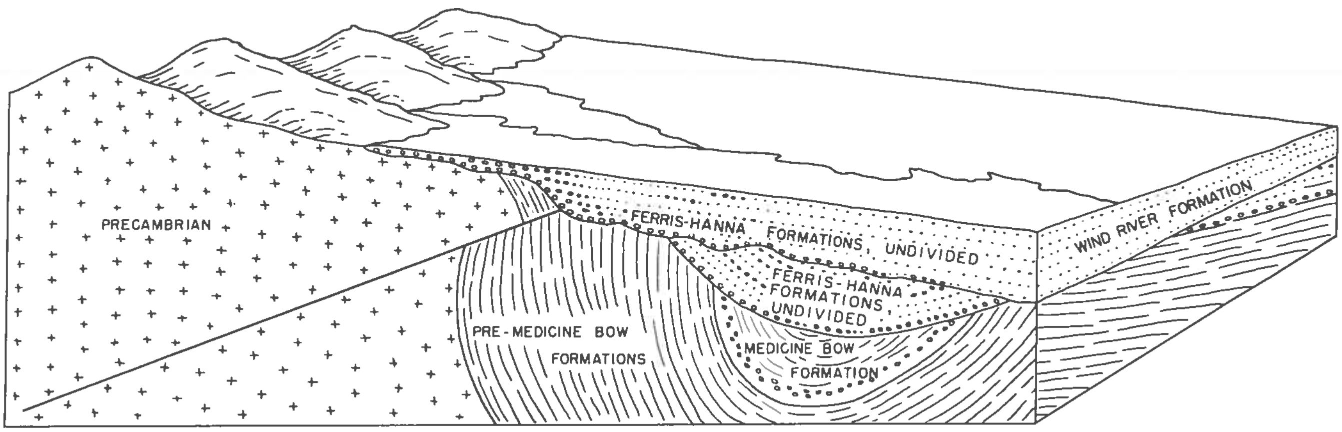


Figure 31—Block diagram showing probable relationship between the Medicine Bow and Ferris-Hanna Formations, and Wind River Formation of the Mill Creek syncline northeast of Centennial. Note that the upper part of the Ferris-Hanna Formations may grade basinward into the Wind River Formation and that both lie with angular uncomformity on the lower part of the Ferris-Hanna Formations in the Mill Creek syncline. Modified from S. H. Knight (1953).

west of Arlington. They are also present in a syncline in the Kennaday Peak area where thickness may exceed 2000 feet and in a great plate of gently dipping sediments from Phantom Lake north to T. 19 N. (Pl. 1).

Rocks of the Ferris-Hanna Formations, undivided, include feldspathic sandstone, arkose, carbonaceous shale, conglomeratic sandstone, and thick beds of conglomerate (Fig. 32). The lithology of partial sections is in Tables 61 and 62 and a composite stratigraphic section with typical lithology is in Table 63.

Table 62—Ferris-Hanna Lithology, NW ¼ SW ¼, Sec. 10 T. 18N., R. 80 W., above the East Fork of the Medicine Bow River, Carbon Co., Wyoming (after Gries, 1964, p. 63)

Lithology	Feet
Shale and sandstone, largely covered	50+
Shale, yellow to rusty-brown, limonitic, micaceous, very fissile, local stringers of medium to coarse-grained sandstone	15
Sandstone, buff to brown, cross-bedded with stringers of pebbles up to one inch in diameter in coarse-grained sandstone	7
Sandstone, white to buff, poorly consildated, one-inch carbonaceous shale layer at bottom	3
Shale, dark-gray, fissile, with limonitic stains on bedding planes. Contains many partial prints of leaves and plant remains (Pollen Sample No. SP-2)	3
Sandstone, yellow, limonitic, friable, coarse-grained with inclusions of shale	3
Shale, rusty-brown, platy, locally sandy, carbonaceous films on bedding planes	6
Shale, rusty-brown to medium gray, sandy	22
Conglomerate, greenish-gray to rusty brown, matrix; rusty-brown, arkosic sandstone, clasts, pebbles, cobbles, and boulders, averaging from 4 inches in diameter up to two feet. Cobbles largely quartzite with some weathered metasedimentary rocks, well rounded	71
Total	180+
Angular unconformity with Cloverly Formation	

This conglomeratic flank facies of the Medicine Bow Mountain is correlated with the Ferris Formation and Hanna Formation of the Hanna Basin (Bowen, 1918). These two formations are also conglomeratic and are similar lithologically to the units described above. Like the Ferris-

Table 63—Composite Stratigraphic Sequence of the Ferris-Hanna Formations (undivided) from Isolated Outcrops between Cooper Hill and Arlington (after King, 1963, p. 73)

Unit	Top
8.	Dark-brown sandstone
7.	Brown sandstone with ferruginous concretions
6.	Carbonaceous shale
5.	Yellowish-green sandy shale
4.	Conglomerate—boulder-sized clasts common
3.	Reddish unconsolidated sandy soil with boulders of Precambrian rocks
2.	Light-to dark- sandstone, some with quartz pebbles
1.	Dark-brown conglomerate with clasts largely cobble sized. Clasts chiefly locally derived Precambrian rocks, inter bedded white to gray, medium to coarse grained Arkosic sandstone base

Hanna Formations, undivided, of this area they are formed from material eroded from uplifts that developed during the Laramide Orogeny. The Ferris Formation of the type locality has a basal conglomerate that lies on the Medicine Bow Formation. This conglomerate may not be derived from local uplifts bordering the Hanna Basin (Bowen, 1918, p. 230) and is generally finer-grained than conglomerates of the Hanna Formation of the type locality or in the Medicine Bow area. Beds of the Ferris and Hanna Formations of the type locality are also much thicker (13,600') than similar beds of the Medicine Bow area, but aside from differences in thickness, probably caused by greater rates of subsidence in the Hanna Basin, and some differences in lithology, because of variations in source area, these units are very similar. As will be shown below the Ferris and Hanna Formations of the type locality and the Ferris-Hanna Formations, undivided, of this area are also roughly equivalent in age.

We must admit, however, that these units are not precise equivalents and one solution to the problem is to propose new names for the units in the Medicine Bow area. This approach has been used by Hyden, McAndrews, and Tschudy (1965, p. K2-K3) who state that units above the Lewis Shale; the Medicine Bow Formation, Ferris Formation, and Hanna Formation can not be correlated from the Hanna to the Laramie Basin. These writers suggested that new names be introduced because of the great difference in thickness (20,000 feet in the Hanna Basin versus 800 feet in the north part of the Laramie Basin), probable difference in source areas, and slight age difference in specific units. We have already noted that the Foote Creek Formation of Hyden and others is generally equivalent to the Medicine Bow Formation as described from the Mill Creek syncline area of the Medicine Bow Mountains, but does differ considerably in thickness from the Medicine Bow Formation of the type locality. In the Laramie Basin the Foote Creek Formation is overlain by a unit consisting of conglomeratic sandstone, interlayered with shale, conglomerate, coal, and fine-grained sandstone that has been designated the Dutton Creek Formation by Hyden and others (1965, p. K-11). This formation must equate, in part, to the rocks referred to as Ferris-Hanna Formations, undivided, in this report. It is considered late Paleocene in age, however, and is overlain by the Wind River Formation so it may be older than some rocks included in the Ferris-Hanna, undivided, of this report.

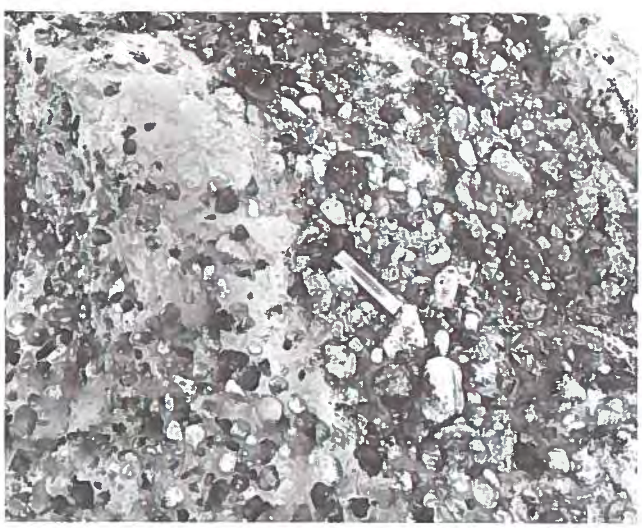


Figure 32—Conglomerate of the Ferris-Hanna Formations (undivided) from outcrop along Mill Creek. Note large boulder in upper right. Most light colored cobbles are quartzite from the Precambrian of the northern Medicine Bow Mountains.

Age of the Ferris-Hanna Formations (undivided)

In the Medicine Bow Mountains beds mapped as Ferris-Hanna Formations, undivided range in age from Paleocene to Eocene. In the Mill Creek syncline ten samples were collected for spores and pollen study from the base of the formations to the top near the axis of the syncline. Four of these samples had adequate material for study and all were from basal beds; i.e., at the base of the formations on Mill Creek and along the southeast margin of the syncline near the contact with the Medicine Bow Formation. All samples yielded spores and pollen that were assigned to the Paleocene by Woodside and Tschudy. Montagne (1955, p. 21) collected leaf remains from a locality near the base of the section at Kennaday Peak at an elevation of 9000 feet in sec. 28, T. 17 N., R. 81 W. These were regarded as Paleocene by R. W. Brown (personal communications J. Montagne, 1960). Other samples were collected in the northern part of the Medicine Bow Mountains and only one of these yielded material satisfactory for spores and pollen analyses. This sample was from dark-gray shale 102 feet above the base of the Ferris- Hanna Formations, undivided at a locality in sec. 10, T. 18 N., R. 80 W., along the east fork of the Medicine Bow River. According to Woodside the pollen and spore assemblage from this sample was Eocene in age, possibly equivalent to some part of the Wasatch flora.

In the type locality, the Ferris Formation was considered by Bowen to be equivalent to the Fort Union in age (Paleocene), but noted that fragments of bones of **Triceratops** were present in the basal part of the formation. Palynological study of samples from the Ferris Formation in the type locality confirmed a Late Cretaceous age for the basal conglomerate of the Ferris and a Paleocene age for the upper part (Hyden, McAndrews, and Tschudy, 1965, p. K3). The age assignment of the Hanna Formation was conjectural and perhaps it ranges in age from Paleocene to early Eocene.

Summary of Stratigraphy of Rocks of Latest Cretaceous and Early Tertiary Age

Rocks of the latest Cretaceous age are present along the margin of the Mill Creek syncline and on both lithologic and paleontologic evidence are considered to be equivalent to the Medicine Bow Formation of the Hanna Basin. In sec. 29, T. 16 N., R. 77 W. a conglomerate containing largely clasts of locally derived sedimentary rocks is present 515 feet above the base of the formation that is the first direct evidence of uplift during the Laramide Orogeny.

Conglomeratic sandstones interbedded with conglomerate, carbonaceous shale, feldspathic sandstone, and coal overlay the Medicine Bow Formation unconformably. These rocks of Paleocene-Eocene age are called Ferris-Hanna Formations, undivided because they are similar lithologically and are approximately the same age as these two formations as they were described in the Hanna Basin (Bowen, 1918). The Ferris-Hanna Formations, undivided, of this area probably include rocks equivalent in age to the Dutton Creek Formation of Hyden and others and the Wind River Formation as this unit is mapped in the Laramie Basin.

Problems in terminology for these units, especially the rocks called Ferris-Hanna, undivided, in this report are due largely to facies changes that take place from the source area (mountain uplift) to the adjacent basins, the unconformities that develop near the uplift where deformation is greatest, and the nature of the local source which will contribute material of different composition in different areas. The thickness of the formation in any given locality is probably related to the rate of subsidence of the basin of deposition.

Whether the Ferris Hanna Formations of this report are distinctive enough to be given a new name is debatable. Certainly the unit cannot be correlated with the Dutton Creek Formation which is overlain unconformably by the Wind River Formation, but it is not clearly correlatible with the Ferris Formation of the type locality on lithologic grounds. This problem will have to await further stratigraphic and paleontologic study.

It is clear, however, from this study that from Latest Cretaceous to Early Eocene time the Medicine Bow Mountains were uplifted and a conglomeratic flank facies developed along the margin of the uplift. We can demonstrate from lithologic and structural relationships at least two episodes of deformation (Fig. 31). Exposures are not adequate to allow further documentation, but this is certainly a minimum. Deformation probably began as early as Lewis time (75 m.y.$\pm$) and continued spasmodically through the Late Cretaceous and Early Eocene (52 m.y.$\pm$), a time span of approximately 23 m.y. It must be emphasized that the deformation and resultant angular unconformities shown in Figure 31 are largely confined to the mountain front and that basinward (often less than 10 miles from the present mountain) there is not only a striking change in lithology and grain size in these units but beds of Cretaceous, Paleocene, and early Tertiary age may be conformable.

STRUCTURE

The most interesting aspect of the structure of the Medicine Bow Mountains is the evidence that a record of very early events of the Precambrian is retained in an area subject to stress or earth movements perhaps several times during the Precambrian and having one major tectonic event after the Precambrian. Tectonic or thermal events since 2.4 b.y. ago were sufficient only to mix or homogenize strontium isotopes in the basement gneiss on a local scale. Complete remobilization, melting, or otherwise mixing on the scale necessary to eradicate completely older structures has, therefore, probably not happened since 2.4 b.y. ago. It is therefore possible that structure developed 2.4 b.y. ago or earlier is preserved locally. The early structural record is preserved in the basement gneisses of the northwestern part of the area. The metasedimentary rocks that lie on this basement complex are also deformed, probably twice. Thus we have north of the shear zone a very complex tectonic record. South of the shear zone we deal with a more highly deformed series of rocks than in the north with clear evidence of refolding, but there is no record of early events reflected in the ages determined for the rocks of this area. Some event, probably tectonic, that has not been transmitted to the north or that was less strongly developed in the north, has affected the rocks south of the fault. This striking contrast in the structure and the age of rocks north and south of the Mullen Creek-Nash Fork shear zone suggests that the Medicine Bow Mountains are part of a transitional zone between the very ancient rocks of the Precambrian of central Wyoming and rocks of the Front Range of Colorado that from age determinations are either younger or have suffered a thermal or tectonic event that has wiped out evidence of the early ages (Hills, et al. 1968. The Mullen Creek-Nash Fork shear zone may therefore be part of a boundary between two geologic provinces that existed in Precambrian time. If this is a real boundary, one can see the importance of contrasting the structure of rocks north and south of the shear zone and of trying to establish a chronology of structural events. Therefore, in the discussion of the structure of the Precambrian rocks emphasis will be placed on structural evolution.

A complex sequence of igneous rocks of both mafic and felsic composition is present in the gneisses and metasedimentary rocks. These igneous rocks range from older than or equal to 2.4 b.y. to 1.35 b.y., and all igneous rocks have been metamorphosed to some degree. The Sherman Granite (1.35 b.y.) is least affected by deformation, but even this relatively young rock is locally sheared. These igneous rocks, especially the basaltic and gabbroic intrusions, may be the key to the structural evolution of the area, and an attempt will be made to evaluate the significance of mafic intrusive bodies in this respect.

The last intrusive event prior to deposition of Cambrian rocks may have been the introduction of the Sherman Granite (1.35 b.y.); so there is a gap of at least 750 million years (if one uses 600 m.y. as a time of initial deposition of Cambrian rocks) in the geologic and structural record for this area. There may not have been penetrative deformation in this 750 million year interval, but there were certainly intervals of erosion and faulting. The major deformation during the Laramide has been partly discussed in the chapter on Paleozoic and Mesozoic rocks. During Laramide deformation the Medicine Bow Mountains were uplifted and especially in the northern part of the mountains individual blocks of Precambrian rock have been disrupted, displaced, and rotated during the orogeny. Particularly in the northwest, blocks of Precambrian rocks were separated from each other by faults (many partially or wholly covered by sedimentary rocks of Tertiary and Quaternary age) so that the Precambrian rocks are not in their pre-Laramide position. Tilting and faulting during late Tertiary may have resulted in reorientation of Precambrian structures. The effect of the post-Precambrian deformation is such, that in many cases, the present day orientation of structures in the rocks of Precambrian age may be quite different from that of 1.35 b.y. ago.

The long and complex structural history of the Medicine Bow Mountains means that one cannot simply map the structures of the Precambrian rocks and ignore the Laramide deformation, nor can one consider Laramide structure independently of the Precambrian. In fact, the probability that a given structural trend or fold orientation in rocks of Precambrian age is original must be small. Ideally the structure of such an area should be reviewed in order of age, but this requires that the record be complete and all late structures be removed by reconstruction of surfaces prior to a given structural event. This cannot be done in the Medicine Bow Mountains for later structures have in many cases masked earlier ones and despite the remarkably complete record there are large gaps such as the 750 million year interval between introduction of the Sherman Granite and deposition of Cambrian rocks. We will discuss the structures in rocks as they are present today, and review the structures in the older (Precambrian rocks) rocks first. Where information is available the sequence of structural events will be considered in each area, but particularly in the discussion of Precambrian structure the reader is forewarned that although in a given area or domain a sequence of structural events may be outlined, and orientation of these structures may not be that of the time of formation. Furthermore, we cannot demonstrate that all structural features of the Precambrian rocks are Precambrian in age.

STRUCTURE OF ROCKS OF PRECAMBRIAN AGE

The structure of the rocks of Precambrian age will be reviewed by considering two major divisions: (1) rocks north and (2) rocks south of the Mullen Creek-Nash Fork shear zone. The northern area is divided into three provinces: (1) quartzo-feldspathic gneiss of the northwest; (2) older metasedimentary rocks of the northeastern part of the mountains; and (3) younger metasedimentary rocks of the central part of the mountains. The quartzo-feldspathic gneiss of the northwest slope will be reviewed first.

STRUCTURE OF THE QUARTZO-FELDSPATHIC GNEISS OF THE NORTHWEST

The area covered in the discussion of the quartzo-feldspathic gneiss of the northwest is shown in Plate 4. This area includes those basement gneisses and igneous rocks north of the

Mullen Creek-Nash Fork shear zone and west of the metasedimentary rocks. The rock types of the area have been described previously. Although age relationships can be established among these various units,—i.e. older hornblende gneiss, biotite gneiss, and quartzite-quartzo-feldspathic gneiss formed, in part, by transformation of older rocks; younger quartz diorite, quartz monzonite, and granite; all of these rocks have been deformed and the structure of the rocks shown in Plate 4 is partly superimposed on earlier structures. The rocks of mafic composition, basalt and gabbro found in the other units, have in part been deformed with the gneissic rocks and, in part, deformed independently of the gneissic rocks. Some mafic bodies may post-date deformation of the gneiss. Therefore, if information is to be gained on timing of events in an area of this sort lacking in marker beds, clues must come from a comparison of the structure of the mafic rocks and the gneissic units as well as from "absolute" age determinations.

In a review of regional structure in which adequate information for structural analysis is not available various compromises in the presentation of data must be made between a purely descriptive approach and a more quantitative approach. We feel that there is enough information for a preliminary discussion of the structure of this unit, but the reader should recognize at the outset that this review is not meant to be a structural analysis in the strict sense.

The structure of the quartzo-feldspathic gneiss area will be reviewed by subdividing it into domains (Pl. 4). Ideally a structural domain should be an area with uniform geometric properties or be statistically homogeneous in the scale of the domain. An attempt was made to subdivide the area of quartzo-feldspathic gneiss into domains having similar geometric properties, but, for the most part, these domains proved too large and too complex geometrically to be considered statistically homogeneous. The major value of these domains is to provide a subdivision that can be used in discussions. In addition, information on planar structures is limited in most domains so that the β diagrams for each domain must be interpreted with caution.

The major structural features of each domain are summarized in Table 64. This summary includes trend and plunge of β, lineation, attitude of foliation, zones of attenuation and shear zones, and the relationship of the structure of the mafic bodies (amphibolite and metadiabase) and the attitude of foliation of the gneiss. These various elements of the structure will be discussed separately in the following paragraphs, and this will be followed by a more detailed analysis of the structure of certain key domains.

Beta Diagrams

Plate 4 is a tectonic map of the Medicine Bow Mountains with β diagrams superimposed. Not all domains are contoured because structural information is too limited in some areas. These β diagrams are contours of intersections of foliation and commonly are the equivalent of B or the direction of plunge of the fold axis but in this area such equivalency can only rarely be shown. The β diagram is not considered as useful for structural analysis as the π diagram (plot of poles to planes of foliation) because β diagrams magnify small field errors and many intersections not related to B occur in diagrams depicting gently or isoclinally-folded foliation (Ramsey, 1963). The β diagram is used here because it shows the fold axis directly and is more easily visualized by the general reader than the π diagram.

In most domains of the quartzo-feldspathic gneiss the β diagrams do not show a strong concentration of foliation intersection or a maximum that might be equated to a fold axis, B. In some domains, VIIA, VIIC, XI there is a suggestion of a small circle distribution of β. This small circle distribution of β could be interpreted as evidence of refolding by flexure of some pre-existing fold axes. It may also be an apparent small circle distribution caused by interference between north-plunging and northwest-plunging folds. The possibility of refolding will be considered further in the discussion of key domains.

In other domains, XIV, XVII, XIX, XXI, XXII, and XXV, there is a suggestion of a great circle distribution of β. Such distribution could result from isoclinal folds in these areas and in some areas minor isoclinal folds have been observed (Table 64), but this type of distribution of β may simply result from minor errors in foliation measurement in the field and be more apparent than real. A great circle distribution of β could also result from passive folding of some early fold

A

Well-lineated quartzo-feldspathic gneiss from the Cedar Ridge area in which lineation seen in plane of foliation plunges parallel to the axes of minor folds.

B

Lineation in hornblende gneiss shown by elongation of amphibole crystals and by parallel aggregates of elongate amphibole crystals. Sec. 10, T. 12 N., R. 79 W.

Plate 28—Examples of lineation.

axes, but this cannot be verified in these domains.

A strong β concentration has been noted in only two domains, XXIV and XXVI. These two domains are near the Mullen Creek-Nash Fork shear zone. These steeply plunging folds may result from late folding of foliation associated with movement in the fault.

Lineation

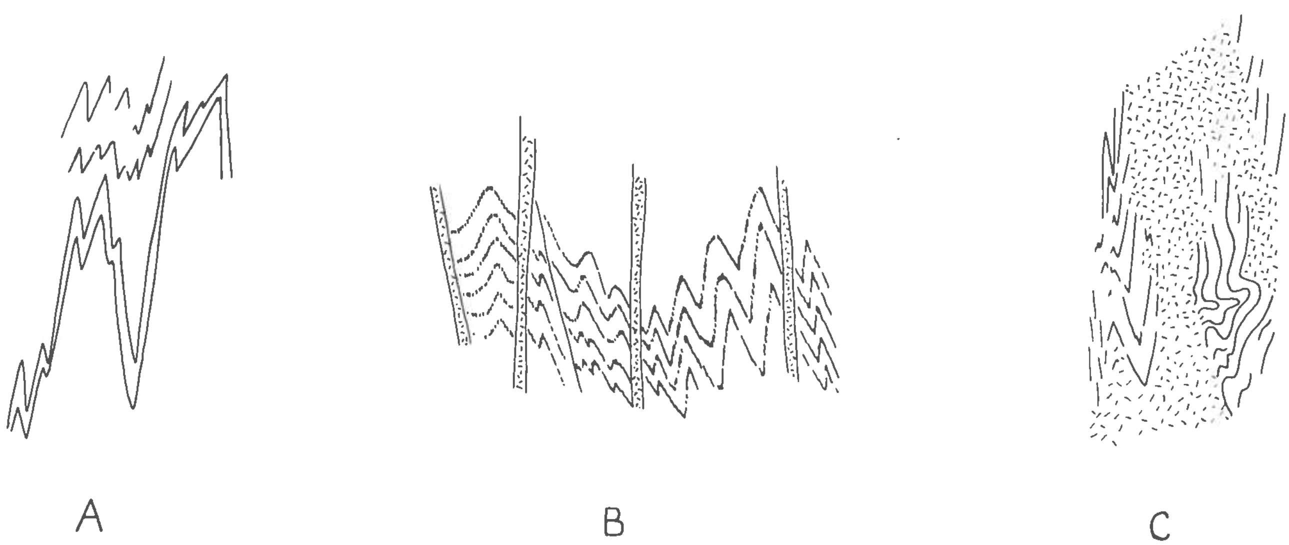
On the geologic and tectonic maps (Pls. 1 and 4) two types of lineation are distinguished; mineral lineation and minor fold axes. The mineral lineation may include biotite streaks measured on the plane of foliation and a preferred orientation of elongate minerals such as amphibole. In most cases the "mineral" lineations are biotite streaks measured in the plane of foliation the genesis of which is unknown. This "mineral" lineation rarely parallels B (the fold axis), however, and in most domains it is variable in attitude suggesting refolding.

Minor folds are common in some domains. They are of both flexural and passive types (Donath and Parker, 1964). The plunge of minor fold axes may parallel β , but in some domains there is evidence of folding of minor folds about new axes. Table 64 summarizes the attitude of the linear features in domains where information is available, and details of lineation of specific domains will be discussed below.

Foliation

For the most part the foliation measured in the quartzo-feldspathic gneiss is compositional layering which is parallel to the orientation of platy minerals. In the great majority of outcrops there is only one foliation or S-surface developed, but in some domains isolated outcrops show more than one S-surface. Two S-surfaces are noted where there is evidence of passive folding of the gneiss and the second surface or S_2 may be subparallel to the axial plane of the folds (Fig. 33). This S_2 foliation is rarely distinguished from S_1 in plan view, so it is not possible to distinguish the two S-surfaces in the great majority of outcrops.

In the area of quartzo-feldspathic gneiss the foliation plotted on the tectonic and geologic map is the foliation parallel to compositional layering in a given outcrop. This is probably an S_1 foliation for the most part, but it could be S_2 in outcrops where S_1 and S_2 are parallel or where S_1 has largely been obliterated. Because it was usually impossible to distinguish between S-surfaces of different ages, an additional limitation is placed on the interpretations of the β diagrams. In the discussion of individual domains to follow some distinction will be made between S-surfaces



A B C

Figure 33—Field sketches of minor folds from area of quartzo-feldspathic gneiss. A—Sketch of felsic layer in biotite gneiss at Baggot Rocks showing passive folds with thick hinges and attenuated limbs (felsic layer approximately one inch wide). B—Sketch of folded biotite gneiss in Beaver Hills showing passive folding of biotite rich layers and orientation of biotite parallel to axial plane of fold. Pegmatite is emplaced in axial plane of folds. Distance between layers of pegmatite about one foot. C—Sketch of minor folds in biotite gneiss in eastern part of Beaver Hills (sec. 2, T. 14 N., R. 82 W.) showing folded gneiss locally replaced by pegmatite. Width of outcrop one foot.

of different ages and how they may be combined in the β diagrams giving more than one β maximum.

Zones of Attenuation and Shear Zones

In some areas of the quartzo-feldspathic gneiss, folded gneiss passes into zones of nearly constant strike and dip of foliation that are referred to here as zones of attenuation. In both small scale (hand specimen) and large scale (map scale) some folds have greatly attenuated limbs and these may grade into zones where folds are so drawn out and disrupted that there is little evidence of the original fold remaining. This type of attenuation is best seen in hand specimen (Pl. 31), but map scale examples have been recognized especially on Cedar Ridge (domain VII). These attenuated zones are probably the end product of passive folding. There is no way to determine whether appreciable offset occurs on opposite sides of the zones of attenuation, but such zones could be akin to shear zones and do have some similarities in the field.

Shear zones are also common in the quartzo-feldspathic gneiss. Most shear zones have been recognized in areas where there is an abrupt change in strike of foliation and where cataclastic structure can be recognized in the gneiss. Most shear zones are probably faults with a development of cataclastic structure and mylonite resulting from movement in the fault zone. This is especially true where the shear zone makes a large angle with the general trend of gneissic foliation. Other shear zones that nearly parallel the general trend of gneissic foliation may be genetically related to the zones of attenuation.

The general trend of these zones of attenuation and shear zones is presented in Table 64. Most of these zones have nearly vertical foliation. Typical examples of these structures will be discussed below.

Relationship of Structure of Mafic Dikes (Metadiabase and Amphibolite) to Attitude of Foliation of the Gneiss

In most domains mafic dikes or sills are concordant in strike and dip to the foliation of the gneiss (Table 64), and where complex folds are present the dikes appear to be folded with the gneiss. Examples of this type of relationship are shown in Figure 34, B, C, and D. In other domains some dikes cut the gneissic foliation at an acute angle (Pl. 17), but appear to be concordant in plan to the general trend of foliation. One interpretation of this structural relationship is that the mafic bodies were emplaced in the gneiss in fractures oriented at various angles to the foliation but were brought to near concordance in a later period of folding.

The dikes in zones of attenuation are also concordant to the foliation. These dikes may show button-hook and V-shaped forms that may be remnants of greatly attenuated folds (Fig 34 A, E).

Other mafic bodies are clearly cross-cutting with respect to the foliation of the gneiss and commonly cut the structure at large angles. These dikes are post-tectonic, but they are largely altered to amphibolite.

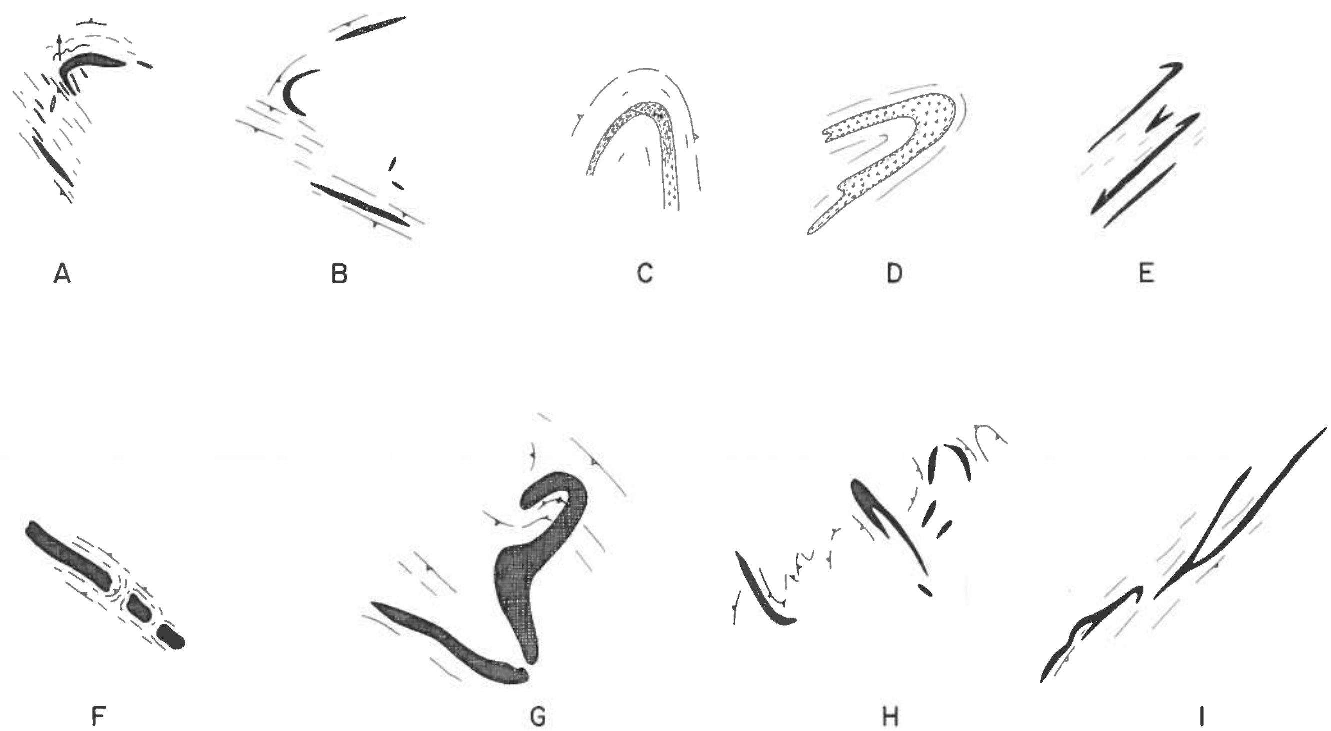
The folded dikes support a concept of multideformation for the quartzo-feldspathic gneiss. Specific examples will be discussed below.

Domain Seven (Cedar Ridge-Baggot Rocks)

Domain seven is the westernmost segment of the Medicine Bow Mountains and includes rocks underlying Cedar Ridge and adjacent hills on the north and Baggot Rocks and adjacent areas on the south (Pl. 4). The rocks within this domain are better exposed than those elsewhere in the quartzo-feldspathic gneiss, and as a result the details of the structure are better known. Rocks of this domain are biotite gneiss, quartzo-feldspathic gneiss, and granite. Biotite gneiss is most abundant on Cedar Ridge and in that body of rock located east of Baggot Rocks (Pl. 1). Quartzo-feldspathic gneiss is the most common rock elsewhere except at Baggot Rocks which is, in part, granite. Dikes and sills of basalt and other mafic rocks are common, and all are at least partly converted to amphibolite, and some are so deformed as to be gneissic in texture. For convenience in discussion, domain seven is subdivided into three subdomains (Pl. 4).

Subdomains A and B

Subdomain A is Cedar Ridge proper which is an area of complexly folded biotite gneiss in which contained sills and dikes of basalt and diabase are folded with the gneiss. The northern part of this subdomain roughly from sec. 2, T. 16 N., R. 83 W. to the north end of Cedar Ridge is an area where folds plunge generally N. 45° W. at



A.—Dike (cross-hatch) conformable with gneiss in northeast deflected in area of attenuation to southwest. Pegmatites (black) introduced in attenuated zone. Maximum width of the dike about 50 feet. Cedar Ridge area.

B.—Dike (cross-hatch) showing deflection in attenuated area as in A. Note small dike at angle to foliation. Maximum width of dike about 40 feet. Cedar Ridge area.

C.—Field sketch of small dike in Cedar Ridge area showing strong development of foliation in axes of flexural fold and on left limb of fold. Dike about 3 feet wide.

D.—Field sketch of small dike in Beaver Hills area showing development of marginal foliation and more massive core. Dike about 6 feet wide.

E.—Diagrammatic sketch of habit of dikes in areas of attenuation.

F.—Boudinage in dike located in northern part of Baggot Rocks area. Maximum width of dike about 50 feet.

G.—Dike showing distinct cross-cutting relationship with foliation of biotite gneiss. Cedar Ridge area. Maximum width of dike about 280 feet.

H.—Dikes in complexly folded biotite gneiss at north end of Cedar Ridge. Dike appear disrupted but conform to foliation of gneiss. Width of dike about 40 feet.

I.—Dikes in Domain 23 showing bifurcating form and discordant structure to foliation of quartzo-felspathic gneiss. Width of dike about 20 feet.

Figure 34—Structural relationship of dikes and sills of Orthoamphibolite in Quartzo-feldspathic Gneiss.

30° as indicated by β maxima and small fold axes (Fig. 35). These folds are both on a small scale, and large scale, and are flexural folds. They probably developed by folding of the foliation of the biotite gneiss.

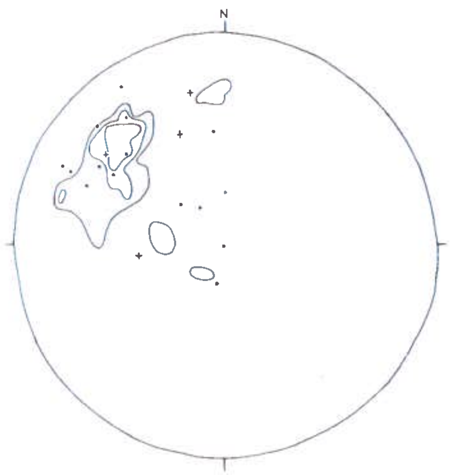


Figure 35—Lower hemisphere equal area plot of minor fold axes (dot) and mineral lineation (x) for the Cedar Ridge area subdomain VIIA. β contours of 78 intersection of S surfaces; 6, 9, and 15 percent of one percent of area.

Through much of this area the antiforms and synforms in the gneiss have axial planes that dip steeply southwest, and for this reason the trend of the axes of the folds is at an angle to the axial trace of the folds (Fig. 36). The flexural folds grade into passive folds in zones of attenuation where steeply-dipping foliation trends northwest. The overall structure of the area is one of complexly folded areas separated by zones of northwest-striking foliation where one structural trend dominates. These zones of northwest-striking foliation are areas where folding movement was passive, and if folds are present they are so disrupted that nothing is left but small fold remnants. In the zones of attenuation the gneiss is generally less-well foliated, and locally pegmatites are present. The β diagram and plot of the plunge of small fold axes for subdomain A suggest refolding about a small circle axis (Fig. 35). This may be more apparent than real because the domain may not be homogeneous and northwest trending concentration of intersection of S surfaces (a possible S_2 surface) probably reflects zones of attenuation where planar structures strike northwest. The rather wide scatter of fold axes may reflect measurements on folds of more than one generation.

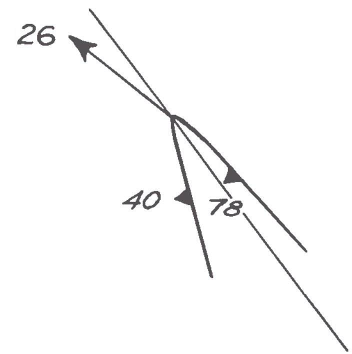


Figure 36—Sketch of minor fold in subdomain A showing trend of fold axis at angle to strike of axial trace of fold.

In secs. 26 and 23, T. 17 N., R. 83 W., a gneissic zone cuts across the structural grain at nearly right angles (Pls. 1 and 4). This zone of quartzo-feldspathic gneiss approximately 300 feet wide strikes northeast and appears unrelated to other northwest-striking structures common in the area. This is one of several such zones in the quartzo-feldspathic gneiss and may be a late structure superimposed on earlier northwest patterns.

In the southern part of subdomain one, an area in sec. 11, T. 16 N., R. 83 W. has been mapped in detail by the writer and S. B. Smithson of the Department of Geology, University of Wyoming to study the minor structure of the gneiss and the relationship of the sills of orthoamphibolite to the gneissic foliation. In this area the major fold in biotite gneiss is a flexural fold trending southeast (Fig. 37). The fold is an antiform outlined by foliation of biotite gneiss and two sills of orthoamphibolite that conform to the structure of the gneiss. The antiform is flanked on both the northeast and southwest by zones of attenuation where the biotite gneiss is less-well foliated and appears somewhat cataclastic and orthoamphibolite bodies are more disrupted and sheared. The biotite gneiss has a well-developed lineation as well as foliation. This lineation is the trace of the axial plane of passive folds on the plane of foliation (Pl. 29). The orthoamphibolite also contains a mineral lineation developed by alignment of the long axes of amphibole crystals. Both lineations are folded with the gneiss and orthoamphibolite (Fig. 37). Although various interpretations can be made from the plots of distribution of lineation in the orthoamphibolite, and lineation in the gneiss (Fig. 37), there is a suggestion of a distribution about a small circle with axis of rotation oriented trending south with a shallow plunge. This is consistent with the view that folding of pre-existing foliation of the biotite gneiss and contained sills and dikes took place by flexure. The interpretation is also supported by the constant thickness of the dike in the hinge of the fold.

In this same area, there are a number of very sharp folds in both gneiss and orthoamphibolite along the margin of the major antiform.

Figure 34 shows the variety of relationships between foliation of the gneiss and form of the amphibolite bodies. In sketches C and D, the form of the amphibolite body is similar to that of the enclosing gneiss. In these less-deformed bodies one can recognize diabasic textures in some of the amphibolite bodies, but as shown in D the mafic body is always strongly foliated at the contacts with gneiss. In many cases, the folded amphibolite is more highly deformed on one limb than the other as in A and C, and the more deformed part of the body is well foliated and shows no original texture. In the more deformed zones (attenuated areas), amphibolite bodies conform in strike and dip to the foliation of the gneiss, appear misoriented, and show but-

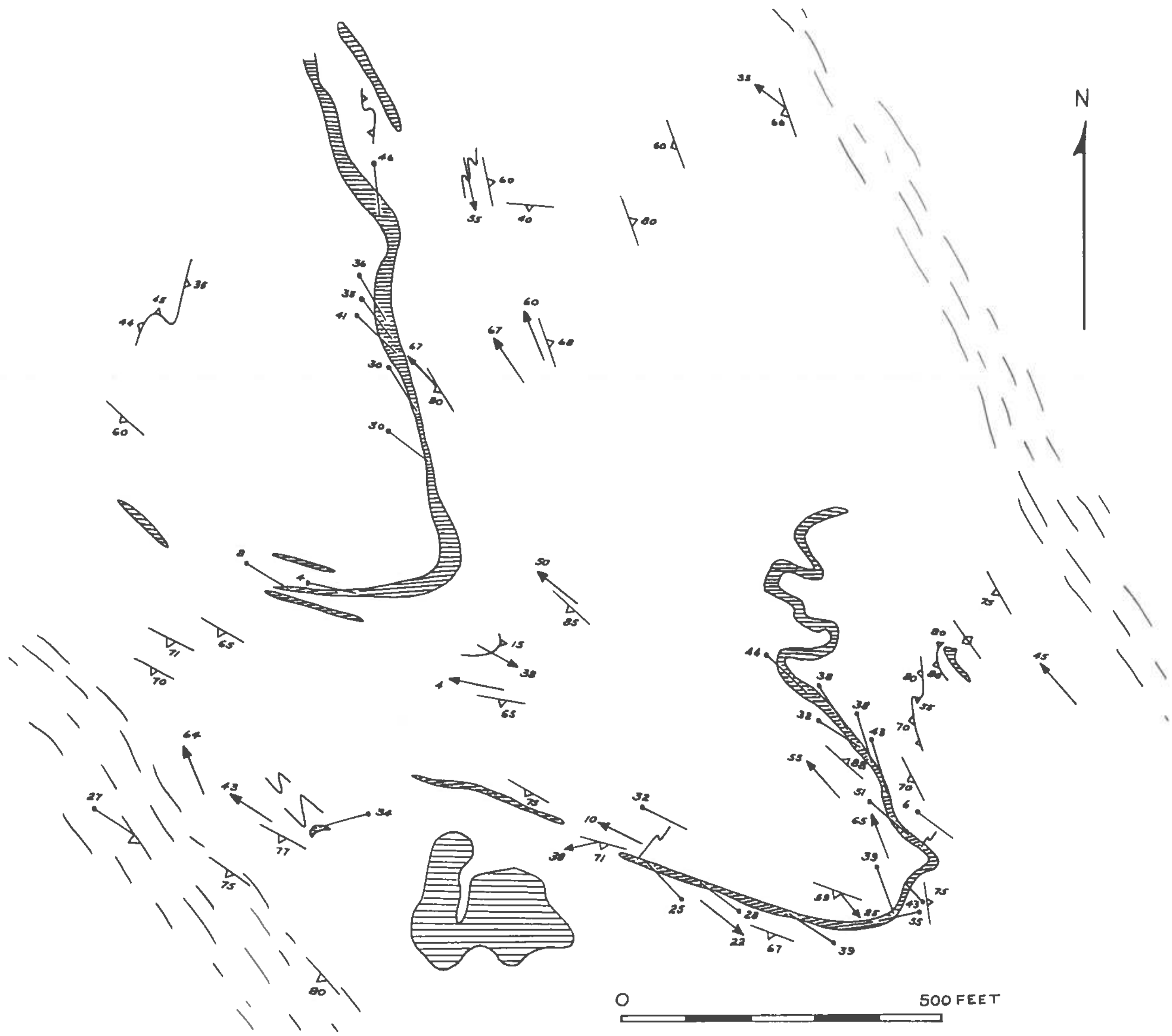


Figure 37—Geologic sketch map of area in northern part of domain seven (Cedar Ridge) showing dikes of orthoamphibolite (dark) folded with biotite gneiss. Mineral lineation shown with dot in direction of plunge; plunge of minor fold axes with arrow in direction of plunge. Dashed lines show trend of steeply dipping foliation. Sketch map by R. S. Houston and S. B. Smithson. Sec. 11, T. 16 N., R. 83 W.

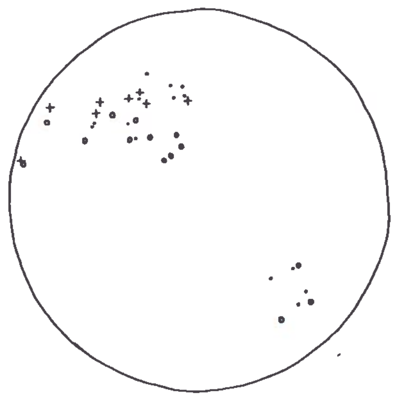


Figure 38—Plot of plunge of lineations shown on Figure 37. Dots are mineral lineations from southeast dike, x are mineral lineations of northwest dike, and circles are plunges of minor folds measured in gneiss. Lineation on equal area projection lower hemisphere.

tonhook-shaped and V-shaped forms (Fig. 34 B, E and H). In these areas, relationship between structure of the gneiss and amphibolite is often difficult to determine, but where outcrops of gneiss are present near amphibolite bodies, the structure of the gneiss is usually conformable to the strike of the amphibolite body as illustrated in A. The amphibolite bodies may simply reflect the deformation of the gneiss, and in the zones of attenuation the disrupted and misoriented bodies, the V-shaped and buttonhook-shaped bodies, and even the seemingly elongate sills are simply remnants of passive folds that have been so disrupted and drawn out that the only evidence of them is retained in the shape of the amphibolite bodies.

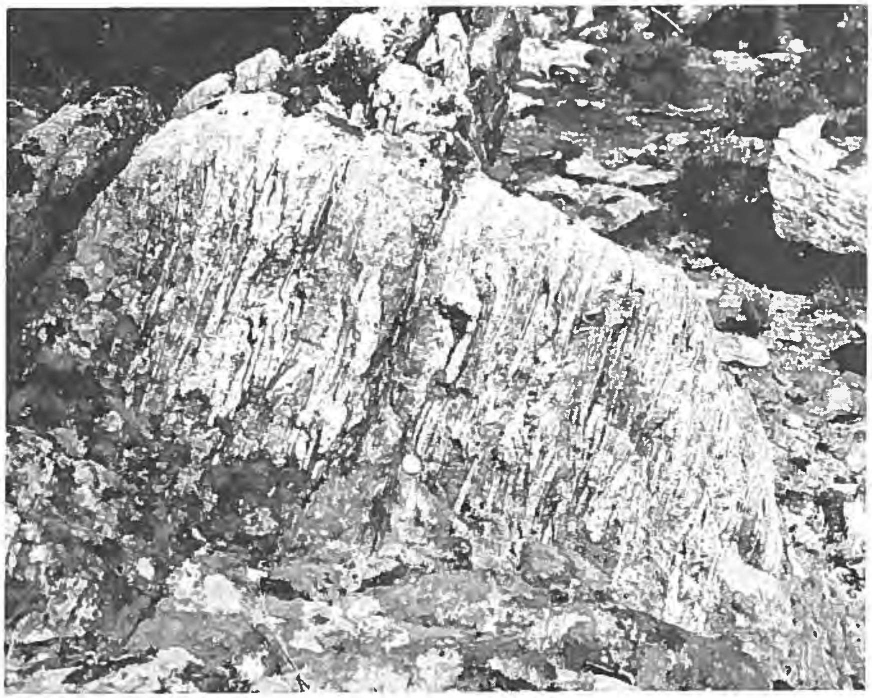
There are many exceptions to the general rule of structural relations discussed above. Many amphibolite bodies do not conform structurally to the gneiss. For example, the large body in the southeast quarter of sec. 11, T. 16 N., R. 83 W., is partly concordant and partly discordant (Fig. 34, G). This may be a preservation of original discordant relationships or it may be a dike emplaced after or during folding.

Although most folds (flexural and passive) trend either northwest or southeast in conformity with the general structure some are aberrant and trend north or south. These folds may be part of an earlier system. In the Cedar Ridge area, the northwest-trending structural pattern has been superimposed on pre-existing structures. The units that were present in the area prior to the development of this structure are preserved, in part, in the less severely deformed blocks such as that illustrated in Figure 37. These units were biotite gneisses cut by generally conformable mafic dikes and sills. The gneiss and contained dikes have been variously deformed by flexure. The areas with flexural folds grade into areas having elements of passive folding and to planar zones or belts of attenuation. Dikes have been disrupted and rotated in the more complexly folded areas and in the zones of attenuation. The gneiss has been largely recrystallized to a more coarsely-foliated quartzo-feldspathic type in the zones of attenuation, and pegmatite that developed during or after this last period of deformation cuts all units. It is possible that a later set of dikes was introduced during or after folding.

Subdomain B covers a north-trending ridge east of Cedar Ridge (Pl. 4). The rocks of this area are largely quartzo-feldspathic gneiss, exhibiting a considerable variation in structure. Outcrops are not good so relatively few structural observations have been made, but a beta diagram shows two maxima, one much less well developed, similar to that of the Cedar Ridge domain and another steeply-plunging east-southeast maximum (Pl. 4). In this domain two structural trends of foliation dominate; northerly and northwest. Unfortunately exposures are such that it is difficult to see how the two trends are related. If one assumes the northwest trend is later (S_2) as at Cedar Ridge, the northerly trend may be early (S_1), and the β maxima plunging steeply east-southeast may be the result of intersection of two S surfaces of different ages.

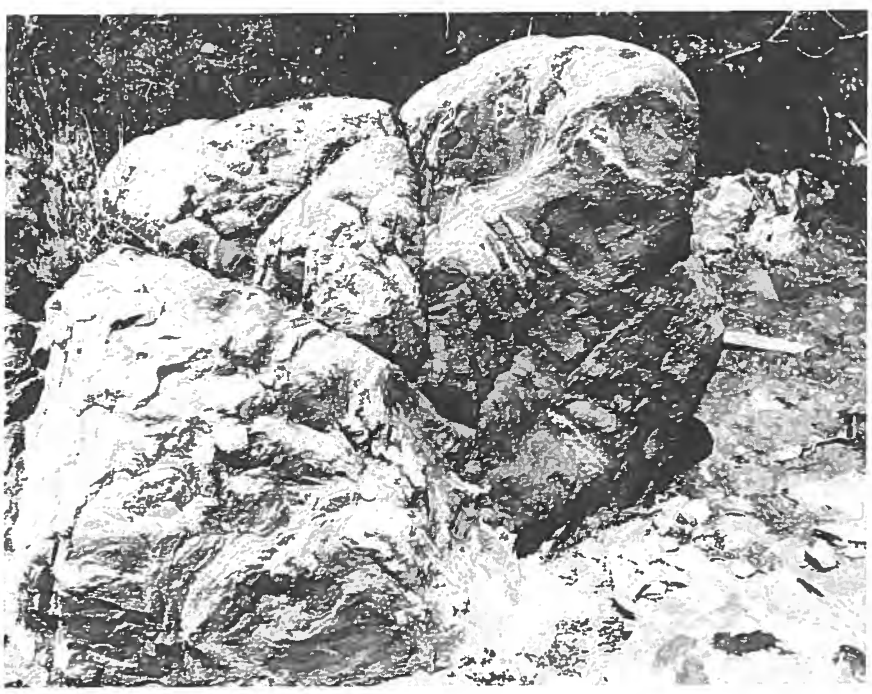
Subdomain C

Subdomain C is Baggot Rocks which is underlain by quartzo-feldspathic gneiss, biotite gneiss and foliated granite, which rocks are cut by mafic dikes and pegmatite. With the exception of the pegmatites and some of the mafic dikes, all units are foliated and there is a general conformity of structure in the various rock types. The main structural feature is a north-trending, antiform located in the central part of Baggot Rocks where the Encampment River occupies a gorge in the Precambrian rocks (Pls. 1 and 4). The western limb of the antiform is cut by a north- to north-northwest-striking shear zone located west of the Encampment River. In the southern and western segment of Baggot Rocks a synform trends north and is best exposed in the NW¼ sec. 16, T. 15 N.,



A

Minor fold in biotite gneiss showing passive folds viewed roughly perpendicular to fold axes. Quarter for scale.



B

Lineation parallel to the plunge of axes of minor folds seen on plane of foliation of biotite gneiss. Quarter for scale.

Plate 29—Lineation in biotite gneiss of Cedar Ridge. Outcrops in sec. 2, T. 16 N., R. 83 W.

R. 83 W. The folds in the Baggot Rocks area are probably flexural folds in the quartzo-feldspathic gneiss and are essentially folded compositional layering of the gneiss. The foliation of the granite conforms to that of the gneiss. The fold pattern of the area is shown by a β diagram of all foliation in gneiss and granite (Pl. 4). This β diagram shows a maximum of N 10 W at 45° plunge, but there is a considerable spread with another maximum at N 45 ° W plunging 45°. Minor folds in the quartzo-feldspathic gneiss trend northwest, with a plunge between 40° and 70°. Mineral lineations in the quartzo-feldspathic gneiss are mineral streaks on the plane of foliation. The trend and plunge of the mineral lineations is not parallel to the trend and plunge of minor folds, in fact, mineral lineations are at an acute angle to the fold axes so the area is not homogeneous (Pl. 4).

Along the northeastern margin the granite grades into gneiss through an extensive zone of migmatite, but along the southern and western borders the granite may either grade into gneiss in a zone a few feet wide or have sharp crosscutting contacts with the gneiss. In areas where the granite cuts (across) the gneiss it is commonly more massive in texture. Actually interpretation of the structural relationship of granite and gneiss would depend to a large extent on which granite outcrops were examined. The relationship between mafic dikes and the major rock types is quite variable, and the dikes change in character from north to south. In the northern and eastern part of Baggot Rocks, mafic dikes are characteristically conformable to the structure of the gneiss and granite. The dikes are more highly deformed and are amphibolite. In some areas in foliated granite, dikes are disrupted and boudinage structure is developed with the foliation of the granite wrapped around the ends of the segments of the boudins (Fig. 34 F). These dikes generally strike northwest. In the southern and western part of Baggot Rocks, the dikes are less deformed and many show original igneous textures. These dikes strike north and although they are largely conformable some, especially in the southeastern part, cut the structure of the gneiss and granite (Pl. 1).

In the south, where the dikes are partly crosscutting, fold axes generally trend northward, and the dikes parallel the axes and cut through the hinges of these folds. In the northeast, the general structural pattern is northwesterly and the dikes are conformable, more deformed and disrupted. The development of structures oriented northwest may be a later event in this history of this area and the development of later northwest oriented structures here may be related to development of similar later northwest oriented structures at Cedar Ridge. Some "mineral" lineations observed in the northwestern area may be caused, in part, by an intersection of planes striking north and planes striking northwest, and this relationship has been noted in the field in one outcrop along the west side of the Encampment River. Other evidence in favor of late deformation is cataclasis of the granite in the northern and eastern part of Baggot Rocks. Areas with strong northwest elements as well as areas with a northern structural element are present in the northeast, and some north-striking dikes are deflected along northwest-trending zones of attenuation much as they are at Cedar Ridge. The β diagrams may also reflect the two periods of deformation. As was the case at Cedar Ridge, there is a suggestion of refolding but one may note two major concentrations of "S" surface intersections, one trending north and one trending northwest (Pl. 4). The accumulated evidence supports a development of a late passive structural element that strikes northwest. The small circle distribution of β in this domain may therefore result from interference of two structural trends rather than flexural folding of early folds about a new axis. The geometry could be even more complicated however with superposition of the northwest system on refolded structures. Furthermore, since we may deal with dikes and even "granite" of more than one age, it is possible that dikes in the north are actually older than those in the south thus reversing assumptions concerning timing of events. These problems may be solved by future detailed mapping of the Baggot Rocks area.

The presence of two structural trends at Baggot Rocks is not as well documented as it is to the southwest in an area on the east margin of the Sierra Madre. An area there is very similar to Baggot Rocks and it has been mapped in detail by Ferris (1964, 1966). The major structure is a northerly-trending synform. The structure is considered to be pre-intrusion of mafic magma, but structure is largely obliterated to the south as a result of penetrative shearing that is later than intrusion of mafic magma and has developed steeply dipping northwest foliations.

Pegmatites in the Baggot Rocks area are late relative to the mafic dikes which they cut. Pegmatites are largely concentrated in the hinge of the major fold located in the southern part of the area (Pl. 1). Some of the pegmatites are deformed cataclastically in the zones of attenuation.

Another area of interest in Domain seven is located northeast of Baggot Rocks (Pl. 4) in an area that is underlain mainly by biotite gneiss with accompanying sills of amphibolite. The sketch map (Fig. 39) of a part of this area shows a north-trending synform with biotite gneiss and amphibolite folded conformably. The fold is disrupted on the northeast flank by northwest-striking zones of attenuation that cut the pre-existing structure. Fragments of the east flank of the fold are separated from each other by the steeply-dipping northwest-striking gneiss, and the sills are buckled and disrupted with some showing buttonhook structure. One sill is located in the belt of attenuation and is conformable in structure to the steeply-dipping structure of the belt (Fig. 39). This may be another case of disruption of early gently-plunging folds by the northwest-striking structures.

Figure 39—Geologic sketch map showing orthoamphibolite (cross-hatch) folded with biotite gneiss. Scale approximately one inch equals 1000 feet. Sec. 35, T. 16 N., R. 83 W.

Age Determinations in Domain Seven

In the Baggot Rocks and Cedar Ridge area both biotite gneiss and granite have been dated by the whole-rock Rb/Sr method, and minerals from the granite have been dated separately. The location of the samples dated is shown in Plate 2. The samples of biotite gneiss were selected from areas where this rock was not cataclastically deformed, i.e., outside of the attenuated areas, and the samples of granite are from the southwestern part of Baggot Rocks where the granite is not cataclastically deformed. We (Hills, et. al., 1968) hoped, in other words, to get evidence of early events in the rocks if such a record was preserved. Five samples of biotite gneiss give concordant ages of 2.41 b.y. $\pm$ .50, and seven samples of granite give concordant ages of 2.34 b.y. $\pm$ .50 Zircon concentrates from the granite gave a primary age of 2.4 b.y. $\pm$ .50. The minerals of the granite which are plagioclase, epidote, apatite, potash feldspar, and biotite are dated, as 1.5 b.y. in contrast to the ancient event recorded in the whole rock determinations.

The old ages for the Baggot Rocks granite and biotite gneiss relate these units to the Superior province of the Canadian shield. Geologically the old ages are determined in the rocks that should be a part of the old basement, and we thus have a general correlation between age determinations and geology. The problem that remains is to correlate the ages with the geology or tectonic history of the gneiss and granite. Hills et. al., (1968) consider the 2.41 b.y. ($\pm$ 0.50) age of the gneiss to be (1) a time of initial gneiss formation (i.e. formation of the gneiss during a period of metamorphism and/or metasomatism), or (2) a time of homogenization of an older terrain through intense metamorphism. They favor the first hypothesis because the gneiss has an initial Sr^{87}/Sr^{86} ratio that is very low. Zircon and whole-rock Rb/Sr ages for the granite agree indicating that the time of primary origin of the granite is approximately 2.4 b.y. ago.

Domain Ten

Domain ten is occupied by quartzo-feldspathic gneiss with steeply-dipping foliation that strikes west and dips northeast. Mafic dikes in this domain are concordant to the structure of the gneiss, are highly deformed, and largely converted to amphibolite. In the southern part of this domain north-striking foliation is present and the dikes are concordant (Fig. 40), but there is an area between the north-striking and the east-striking dikes where the dikes are Z-shaped and are discordant structurally to the east-trending gneissic foliation. This unusual shape of the dike may result from later movement along east-striking foliation planes of the gneiss, and may be another example of disruption of early north-striking structure by later S-surfaces.

Figure 40—Geologic sketch map showing relationship of foliation in quartzo-feldspathic gneiss to trend in plan of orthoamphibolite (cross-hatch). Scale one inch equals approximately 1600 feet. Southern part of domain ten. Secs. 25 and 36, T. 16 N., R. 83 W.

Domain Twenty-Five

Domain twenty-five is located along the drainage of Beaver Creek west of State Highway 130 (Pl. 4). The major rock type in the area is quartzo-feldspathic gneiss, but small patches of hornblende gneiss are common. The foliation in the gneiss strikes generally N. 20° W. and the dip of the foliation ranges between 30 and 70 degrees to the northeast (Pl. 4). The southern part of the domain is an antiform with its axial plane dipping east (Pls. 1 & 4). Minor folds of both passive and flexural types have been noted in the gneiss. The passive folds in the northern part of the area have axes that trend northwest and plunge 25 degrees. In Domain twenty-five the relationship between mafic dikes and the structure of the gneiss is well shown. This is also an area having an unusually heavy concentration of pegmatites, and the relationship between pegmatites, granite dikes, gneiss and mafic dikes is also clear (Fig. 41). In contrast to the Cedar Ridge area (domain seven) mafic dikes are not folded with the gneiss. Two large dikes show the buttonhook shape in plan (Fig. 41). The long dimension of the dikes parallels the foliation of the gneiss, but the buttonhook portion of the dike cuts across the strike of gneissic structure. A third large dike cuts the foliation at a high angle except at its extremities where the trend of dikes and gneissic foliation is parallel. If the northeast-striking dike was introduced in a fracture that cut the gneissic structure, the extremities of the dike may have been deflected by later movement. Through most of this area the dike strikes north-northwest conformable to the foliation of the gneiss, but in domain twenty-five there are five zones (Fig. 41) where the foliation strikes northwest. In these zones, pegmatites are more abundant and dikes are deflected and disrupted. These northwest-striking structures may have developed after dike emplacement and during a stage of pegmatite formation.

Granite dikes are both conformable to the foliation and cross-cutting. The northeast-striking dikes in the northern part of domain twenty-five were probably introduced in fractures developed during movement on the Mullen Creek-Nash Fork shear zone. Similarly to dikes in domain twenty-two (Pl. 4), these dikes are marked by cataclastic structure. A possible sequence of

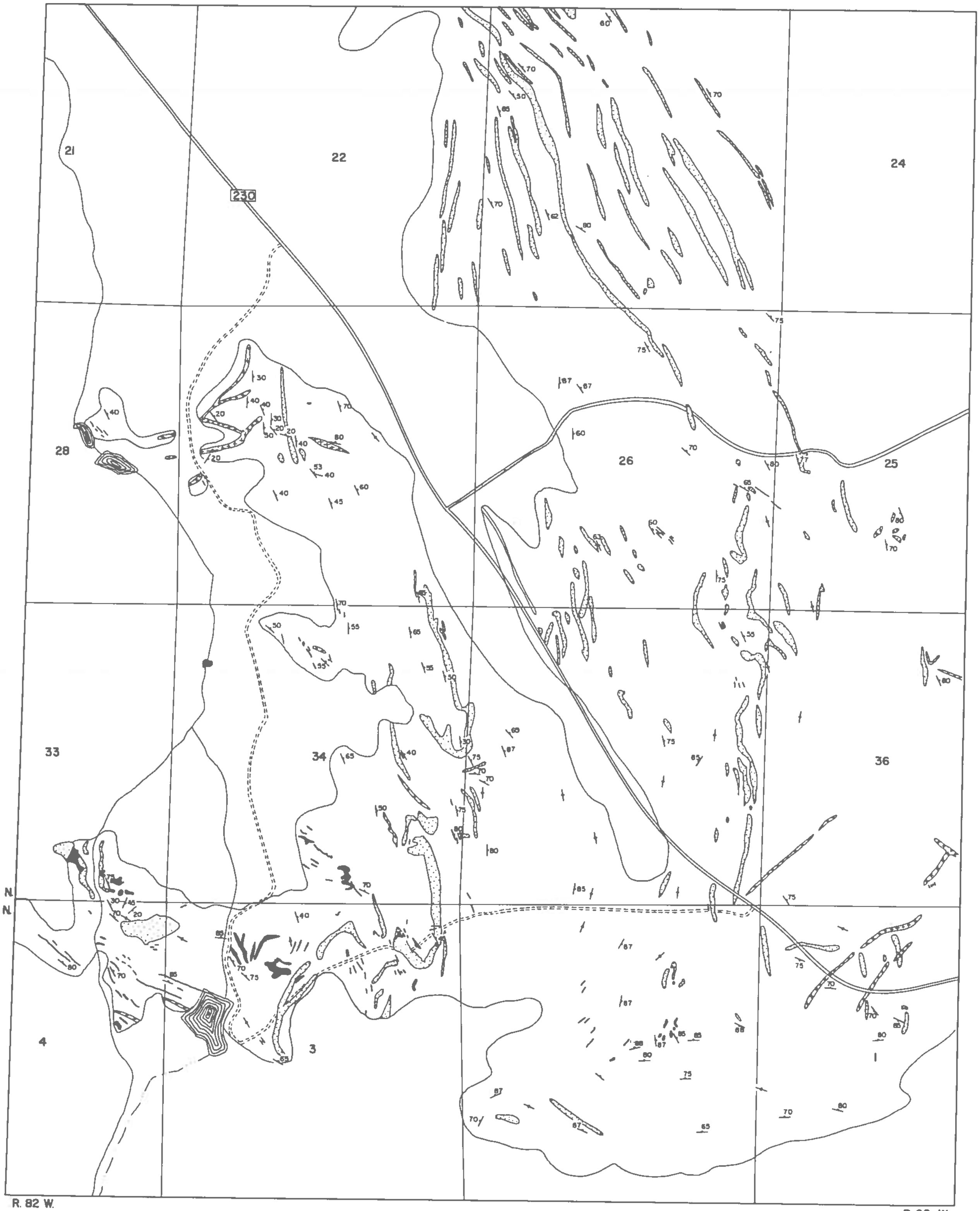


Figure 41. Geologic map of domain 25 and adjacent areas. Stippled pattern *amphibolite,* cross pattern *granite dikes,* solid black *pegmatite,* and enclosed colorless area is *quartzo-feldspathic gneiss.*

events in this area would be: (1) deformation to develop a structural pattern in the gneiss, (2) introduction of mafic dikes, (3) further movement along northwest foliation planes of the gneiss in zones, causing folding of the dikes, (4) introduction of pegmatites after or during phase three, (5) fracturing during the development of the Mullen Creek-Nash Fork shear zone and emplacement of granite dikes in the fractures, (6) movement along these same fractures to deform the granite dikes.

SUMMARY OF STRUCTURE OF THE QUARTZO-FELDSPATHIC GNEISS

In the above discussion certain facts have been established about the structure of the gneiss that allow us to speculate on a structural evolution. The facts are as follows.

1. In some domains north-striking foliation has been disrupted by northwest-striking foliation.
2. Some antiforms and synforms have axes that plunge north or north-northwest. These structures are cut by north-striking dikes, locally.
3. North-striking dikes are deformed along later northwest- to west-striking foliation.
4. In some areas, both dikes and gneiss have been complexly folded. The later structures strike northwest; movement has been, in part, passive and pegmatites were introduced during or after development of the northwest structure.
5. A number of northeast-striking gneissic zones and shear zones cut all structures. The area is bordered on the south by the major northeast-trending Mullen Creek-Nash Fork shear zone.
6. In some areas, the general northwest-striking structure is folded with the development of secondary antiforms and synforms that trend west or east.
7. A later generation of mafic dikes and larger gabbroic intrusives cuts the regional structure at large angles and either strikes east or northeast. Some of these mafic bodies are cut by pegmatite.
8. In and adjacent to the Mullen Creek-Nash Fork shear zone, dikes of both mafic and felsic composition parallel the shear zone. These dikes may also be sheared.
9. Two events are suggested by age determinations:
   (1) A period of gneiss and granite formation at 2.4 b.y.
   (2) A period of redistribution and equilibration of Sr isotopes between minerals in granite at 1.5 - 1.6 b.y.

The best documented event in the area was a period of gneiss and granite formation at 2.4 b.y. as indicated by whole rock Rb/Sr age determinations. The granite and gneiss are cut by mafic dikes, that were introduced either in a late stage of gneiss formation or as a late event, after uplift. Assumptions made concerning the timing of dike introduction with respect to orogeny are of primary importance to interpretation of structural evolution. If the dikes were introduced before deformation ceased, structures that are definitely later than the development of the early folds may simply be late-stage phenomena of the same orogenic event. On the other hand, if the dikes were introduced after uplift (at a higher level in the crust) deformation of dikes would constitute a separate orogenic event. This point has been made by Poldervaart, (Eckelmann and Poldervaart, 1957, p. 1258) in studies of the Beartooth Mountains of Wyoming and Montana. Poldervaart believes that certain Precambrian dikes and small mafic bodies were introduced after a main period of granitization, but before a pegmatite stage of the same event. The evidence for this is that these mafic bodies cut granitized rock and pegmatite but are in turn cut by pegmatite. O'Hara (1961, p. 859-860) notes that the extent of recrystallization of chilled margins of Scourie dike of the Scottish Highland varies with the width of the dike. He believes that this is best explained by recrystallization immediately after intrusion of the dike while rocks were still deeply buried and capable of forming an amphibolite facies assemblage at recrystallized borders. O'Hara (1942, p. 206) notes that a dike in the Lewisian complex of the Scottish Highlands was emplaced while the enclosing rocks were deeply buried because the dike has a stress-free texture that has been metamorphosed to the amphibolite facies whereas the enclosing gneiss is of the granulite facies.

In the Medicine Bow area, exposures are not good enough to test O'Hara's criteria and the gneiss and dikes are of the same metamorphic grade. However, dikes cut gneiss, granite, and pegmatite and are themselves cut by pegmatite. Therefore, we might use Poldervaart's criteria and state that the dikes were introduced after a main stage of granite formation but prior to pegmatite introduction; this timing implies introduction while the rocks were still deeply buried.

An alternate to the above concept is that the dikes were introduced at a higher level in the crust and were affected by a later separate period of deformation. In the rare areas where dike-gneiss contacts were observed, the dike is finer grained at the margin. It (the fine-grained border) is generally sheared to the point where one could not be entirely certain that grain size reduction was original but some dikes had porphyritic cores and one showed a recrystallization of gneiss at the contact. This does not prove, but does suggest introduction at shallow depth. Many of the dikes are severely metamorphosed and as was noted at Cedar Ridge lineation in the dikes and host gneiss is refolded. This suggests a second, separate, metamorphic event of considerable intensity.

Late east-trending folds and shear zones as well as northeast-striking shear zones must have developed after the second metamorphic event since these structures cut all earlier ones and dikes of both mafic and felsic composition either occupy these zones or parallel them. The dikes paralleling these zones and in these zones are deformed; this observation suggests continuing movement or periodic movement in these fracture systems. The movement on these structures may be as late as Laramide.

The Bennett Peak gabbro is clearly cross-cutting with regard to the gneissic structure and is not deformed. It is locally amphibolized, however, and cut by both late mafic dikes and pegmatite. It may well have been intruded at a deeper level in the crust under enough cover so that the border zone might attain amphibolite facies mineralogy. Dates (Hills, et. al., 1968) by the Rb/Sr and K/A method on minerals of these pegmatites indicate a minimum age of 1.6 b.y., but an anomalous initial Sr isotope ratio suggests a greater age.

This review suggests a very complex history for the quartzo-feldspathic gneiss. It is the writer's opinion that the structural history suggested is minimal. Because of the structural complexity it is difficult to relate the age determinations to a specific geologic event. The relatively young age (1.5 - 1.6 b.y.) on minerals of the granite and pegmatite cutting the Bennett Peak gabbro probably relates to a period of folding of the overlying metasediments where ages for metamorphic events are around 1.6 b.y. If this is true most of the tectonic history of the gneiss and granite may have preceeded the event of ~ 1.6 b.y. Perhaps the 2.4 b.y. event is related to the last severe deformation of the gneiss and granite when the northwest structures developed. Another alternative is to consider the dikes as introduced late in an orogenic event and view the evolution of north-trending and northwest-trending structures as part of one major orogeny that terminated around 2.4 b.y.

PALINSPASTIC MAP OF QUARTZO-FELDSPATHIC GNEISS

In order to show how much the Precambrian distribution of these rocks in the quartzo-feldspathic area may have differed from the present, a palinspastic map has been constructed (Pl. 30). In an area lacking in marker beds and devoid of stratigraphic control, unlimited possibilities exist for construction of such a map. It is possible, however, to place certain blocks in their pre-Laramide position by using present position of Paleozoic and Mesozoic formations and, at least, the assumed major horizontal displacement can be removed from faults of Precambrian age. We may also make an assumption that an early structural trend of Precambrian antiforms and synforms was northerly.

In domains eleven, fourteen, fifteen, sixteen, seventeen, eighteen, twenty, and twenty-three, an antiform of regional extent may be present (Pl. 30). Elements of this structure have been brought back to a general northerly trend in the construction of the palinspastic map. The original structure may have been a broad fold cut by numerous nearly conformable dikes. Later deformation compressed the fold especially on the west, and the dikes were deformed and disrupted. The reader will recognize the inadequacies of such a reconstruction, but one may note that certain faults and shear zones fit together, and the northwest-trending structures of Cedar Ridge are better related to structures in domains nine and ten. Structures that presently trend east may have had northwest orientations originally, as in domains nine and sixteen. The antiform on Pennock Mountain may possibly even be related to that of domain eleven, suggesting several miles of right-lateral movement on the Laramide fault bounding the south margin of Pennock Mountain.

This map should show how the orientation of early Precambrian structures has been affected by movement on both later faults of Precambrian age and major faults of Laramide age. Again we must emphasize that the distribution of present structures may differ considerably from past structures, and comments on structural evolution are of greater value as a suggestion of sequences of events than as a statement of disposition of structural elements through time.

STRUCTURE OF THE OLDER METASEDIMENTARY ROCKS (DEEP LAKE FORMATION)

The stratigraphy of the older metasedimentary rocks or the rocks of the Deep Lake Formation is incompletely known so that structural studies must be based on top and bottom criteria. In the central and western exposures of this unit, the quartzite has well-preserved cross-bedding so that criteria are available to determine general structure. Plate 4 shows the structure of these rocks based on the attitude of bedding in the isolated outcrops of quartzite.

In the central and western area underlain by the older metasedimentary rocks, broad flexural folds with relatively gently dipping limbs have axes that trend northeast (Pls. 1 & 4). Quartzite of this area is not highly metamorphosed and primary structures are well developed. Interlayered schist, metavolcanic rocks (?), and schistose conglomerate contain minerals typical of the greenschist facies and a foliation generally conformable to the strike and dip of the quartzite bedding. In contrast to the general northeast trend of structure, isolated outcrops of schist and conglomerate have a northwest trending foliation dipping steeply northeast, nearly at right angles to the strike of bedding in the quartzite. Northeast of Arrastre Lake, sec. 10, T. 16 N., R 80 W., quartzite with well developed cross-bedding contains thin layers of conglomerate that parallel the bedding in quartzite. These grade along strike into deformed zones where the bedding of the quartzite is disrupted and the schistose conglomerate layers have a northwest-trending foliation. These later foliation trends suggest a recrystallization of less-competent units in response to movement from a different direction than that resulting in the northeast fold pattern. In support of this, a case can be made for a set of secondary northwest trending fold axes in the Deep Lake Formation (Pl. 4).

In the eastern part of the area underlain by older metasedimentary rocks, fold axes change direction from northeast to northwesterly in a broad flexure. The rank of metamorphism increases to the northeast where most of the same units described above have minerals typical of the almandine-amphibolite facies (of regional metamorphism). The units in the northeast are more highly deformed than those to the south and west. Limbs of folds are more tightly appressed, and most of the primary structure, even in the quartzite, is destroyed. Two synclines are recognized in this area, one east of Rock Creek in T. 18 N., R. 78 W., and another southwest of Arlington in Tps. 18 and 19 N., R. 79 W., (Pl. 4), on the basis of cross-bedding in the quartzite. The structure of rocks located between these two synclines is not known because of disruption by large mafic intrusive bodies and lack of top and bottom criteria in the metasedimentary rocks.

In the southeastern area underlain by older metasedimentary rocks (east of the Rock Creek fault), the general trend of bedding and/or foliation in the quartzite is northwest, but this mass of Precambrian rocks is on the hanging wall of a Laramide thrust plate and is probably not in its original orientation.

A number of dip faults that strike northwest and dip steeply northeast are present in the Gold Hill area (Pls. 1 & 4). Similar faults which are found in the Deep Lake area (Pls. 1 & 4), located due south of Deep Lake, dip approximately 86 degrees west. None of these faults offsets mafic igneous rocks and some of the fault planes have dikes of mafic composition emplaced in them; so they are regarded as pre-emplacement of mafic magma. Several northeast-striking faults that strike approximately parallel to the general structural trend are found in sec. 15, T. 16 N., R. 80 W., one is west of Rock Creek Knoll, sec. 35, T. 17 N., R. 79 W. and another is southwest of Arlington along Rock Creek (Pls. 1 & 4). A number of areas where there is an abrupt steepening of bedding and local cataclasis in the quartzite are found such as along the North Fork of Rock Creek, and in sec. 15, T. 17 N., R. 79 W. These changes in structure may be related to northeast trending faults in the metasedimentary rocks. The number of faults recognized in the metasedimentary rocks is minimal. Many of the sills and dikes of mafic composition were probably emplaced along fault traces. Some of the dikes are bordered by brecciated or sheared quartzite, and many contain inclusions of brecciated and sheared quartzite which suggest that the dikes were emplaced in disrupted and faulted zones.

The great number of sills and dikes of gabbro, norite, and diabase in the metasedimentary rocks must have been emplaced after folding and faulting, but some of these units show the same increase in degree of metamorphism to the north and northeast that is present in the metasedimentary rocks. North of T. 17 N., these units are almost entirely converted to amphibolite.

A case can be made for two periods of deformation in the older metasedimentary rocks. The early deformation resulted in the formation of a northeast fold system and was probably followed by faulting and introduction of mafic magma. The metasedimentary rocks were recrystallized to greenschist facies. This was followed by later deformation that resulted in the development of broad northwest-trending folds, steeply dipping northwest foliation in some less competent rocks, and an increase in rank to amphibolite facies to the northeast where deformation was most severe.

STRUCTURE OF YOUNGER METASEDIMENTARY ROCKS (LIBBY CREEK GROUP)

The stratigraphy and structure of the younger metasedimentary rocks or the rocks of the Libby Creek Group is better known than that of the Deep Lake Formation. These rocks are well exposed near Medicine Bow Peak and the early work of Blackwelder (1926) along with information obtained in this study indicates a normal stratigraphic succession (Table 5) is exposed in the central range with top to the southeast. The rocks form the northwest flank of a major syncline that plunges northeast (Pl. 1). The plunge of minor fold axes in sericitic layers in the quartzite averages 60° northeast, but inasmuch as the beds may have been tilted by movement on the Mullen Creek-Nash Fork shear zone this may not be the original plunge of the syncline. To the southwest the trough of the syncline is deformed so that these minor fold axes as well as the previously established microfabric of the quartzite (Houston and Parker, 1963, p. 199-200) are rotated through an arc of a small circle (Figs. 42 and 43). This later deformation like the early fold is the flexural type (Houston and Parker, 1963, p. 200-220). Since the publication of Figure 42 by Houston and Parker, further studies by Parker and Kelley (1966) have confirmed the small circle distribution of the old minor fold axes using the analytical method of Kelley (Parker and Kelley, 1966). The southeast limb of the syncline is missing, and has either been removed by erosion on the uplifted (south) side of the Mullen Creek-Nash Fork shear zone or displaced, by horizontal movement on the fault, beyond the area of outcropping rocks of Precambrian age in southeastern Wyoming or to some area now covered by post-Precambrian rocks.

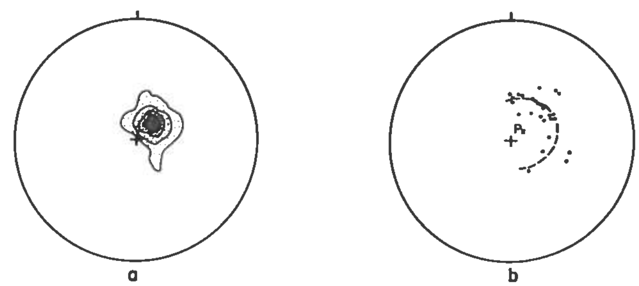


Figure 42—Bedding and lineation diagrams: (a) β diagram of 30 bedding surfaces around fold with contours of 435 intersections at 3-10-20-30-40 percent per 1 percent area; (b) crenulations from the fold with small circle (dashed) that gives best fit to present orientation distribution. Axis of small circle cone (P) shown. Diagrams are lower hemisphere, equal-area (Schmidt) projections. (Figure after Houston and Parker, 1963, Figure 2.)

Several large faults displace the metasedimentary rocks. Typical of these is the Lewis Lake fault (Pl. 4) that follows the drainage of French Creek where there is apparent north movement on the east side. This fault crosses into the Lewis Lake area where the Sugarloaf Quartzite is omitted. In the area east of Lewis Lake Metadolomite of the Nash Fork Formation is in contact with Lookout Schist. As noted by Blackwelder (1926, p. 636) and verified by the writer slivers of Sugarloaf Quartzite are present locally along the fault trace between Brooklyn Lake and Lewis Lake. East of Brooklyn Lake the Sugarloaf Quartzite is present in its full thickness and a normal stratigraphic section is exposed from the Lookout Schist to the French Slate.

Northeast of Brooklyn Lake the Medicine Peak Quartzite thins greatly and part of the section may be removed by faulting. This possible fault in the Medicine Peak Quartzite may be related to the Lewis Lake fault. There may be a single major fault that cuts across the section at an acute angle and slices out various members of the sedimentary succession. In effect, younger rocks would be thrust over older rocks with increasing displacement to the northeast. This fault might resemble the Laramide thrust fault at the southwest end of Centennial Valley, T. 14 N., R. 78 W. except that older rocks are thrust over younger in the Laramide example. Whether this fault cuts out the upper part of the Medicine Peak Quartzite, the lower part, or cuts through the center of the unit is questionable. Blackwelder (1926, p. 651) thought the fault may cut out the upper part of the quartzite and be at the contact between the Lookout Schist and Medicine Peak Quartzite as it is projected northeast of Brooklyn Lake. An alternate possibility is that the fault might follow a dike located between sec. 2 and 3, T. 16 N., R. 79 W., and cut out the lower part of the Medicine Peak Quartzite at the contact with the Heart Formation. This possibility is strengthened by the presence of a distinct topographic depression at the contact between the Heart Formation and Medicine Peak Quartzite of this area. It is true, however, that some beds of the lower part of the Medicine Peak Quartzite are present because some deep bluish-green quartzite is in the North Twin Lakes area at the contact with the Heart Formation.

An alternate interpretation of these faults is shown on Plates 1 and 4. Here the Lewis Lake Fault is interpreted as a thrust with a hinge east of Brooklyn Lake and with major displacement to the west near Lewis Lake. The southward extension of the Lewis Lake fault along French Creek is interpreted as an associated tear fault with left lateral movement. The thinning of the Medicine Peak Quartzite is interpreted as a similar fault with a hinge east of North Gap Lake and increasing displacement to the east. By either interpretation this fault or faults is considered to be some type thrust moving younger rocks over older.

In two other areas the Sugarloaf Quartzite is missing; in sec. 33, T. 17 N., R. 78 W. on Rock Creek Ridge and T. 15 N., R. 80 W. in the hinge of the major syncline. This may be depositional in both areas although on Plate 4 it is interpreted as a folded fault in the hinge of the syncline.

Another major fault that cuts the metasedimentary rocks is the French Creek fault located in T. 15 N., and Ranges 80 and 81 W. (Pl. 4). Unlike the above faults the actual zone of brecciation and shearing can be seen in the field. Where the fault cuts gneiss in T. 15 N. the gneiss is cataclastically deformed and mylonite is present locally, and where it cuts through quartzite this rock is brecciated and cemented by hematite. There is over a mile of apparent right-lateral displacement on this fault in T. 15 N., R. 81 W. The fault could not be traced beyond sec. 24, T. 15 N., R. 81 W., but it may be related to a northwest trending fault located in the northwest quarter of T. 15 N., R. 80 W. This fault may be a thrust fault with hanging wall on the southwest. The French Creek fault would therefore be a tear or transverse fault related to this thrust somewhat like the Laramide fault at Corner Mountain, located about 2 miles north of Centennial.

Faults of various ages offsetting the metasedimentary units at nearly right angles to the strike of the beds are common. Some were formed before intrusion of mafic magma because dikes occupy the fault planes, or faults are cross cut by bodies of igneous rocks, but others are later than the magma. The fault that cuts the metaquartzite at an acute angle in the area northwest of Brooklyn Ridge offsets the Lewis Lake fault and may be late Precambrian or even Laramide in age. The north trending Rock Creek fault in the eastern part of the area is Tertiary in age (Knight, 1953, p. 71).

Two of these dip faults are exposed well enough so that the fault can be observed in the field. The fault located in secs. 24 and 25, T. 16 N., R. 80 W., southwest of Lake Marie is shown in the frontispiece. The fault breccia is made up of angular fragments of quartzite cemented by hematite. The Gap fault located in secs. 5, 8, and 17, T. 16 N., R. 79 W. extends, at least, as far

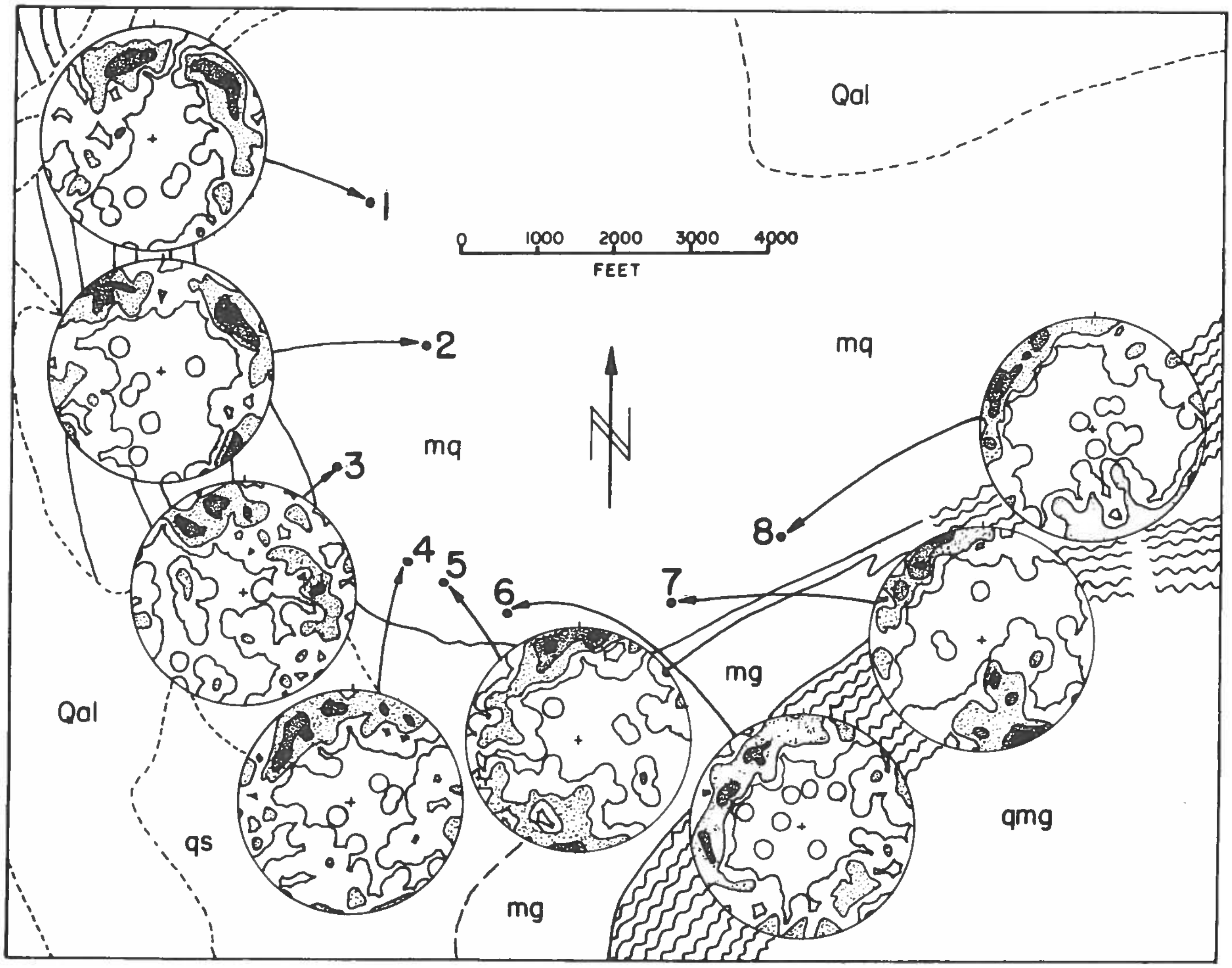


Figure 43—Quartz fabrics in the axis of the fold in rocks of the Libby Group at the southwest margin of outcrop. All diagrams are lower hemisphere, equal area projections of quartz C-axes in normal geographic orientation with contours at 0.5-2-4-6 percent of 1 percent of area. Each diagram represents 200 C-axes. Note the relative rotation of principal maxima from samples 1-8. Mq is Medicine Peak Quartzite; qs, Heart formation, mg amphibolite, and qmg, quartzo-feldspathic gneiss south of Mullen Creek-Nash Fork shear zone. See Plate one, southwest corner of T. 15 N., R. 81 W. and northeast corner of T. 14 N., R. 81 W. for geologic orientation. Fabric by R. B. Parker. (Figure after Houston and Parker, 1963, Figure 3).

south as Libby Lake and may offset the Lewis Lake fault. Where this fault cuts quartzite the quartzite is brecciated and in places the breccia zone is filled by igneous rocks of different composition. North of Lewis Lake, granite invades the fault and is itself sheared suggesting repeated movement. East of North Gap and South Gap Lake amphibolite, probably altered mafic igneous rock, fills the fault. North of North Gap Lake quartzite is brecciated and sheared and the brecciated quartzite is cemented by hematite. This fault appears to have a greater displacement to the north than in the south, but it could not be traced beyond the southern half of sec. 5, T. 16 N., R. 79 W. The exceptional thinning of the Headquarters Schist near Reservoir Lake may somehow relate to this fault.

CRITIQUE OF STRUCTURE OF THE METASEDIMENTARY ROCKS

The two series of metasedimentary rocks, (Deep Lake Formation and Libby Creek Group) are structurally conformable where they are in contact and have probably had a similar structural history. In the central area of outcrop, these units are very little deformed. Primary structures are beautifully preserved and the rocks are of the greenschist facies. Folds in the Deep Lake Formation are broad open folds that plunge at low angles. However, both to the northeast near Arlington and southwest in T. 15 N. the rocks are more highly deformed and the grade of metamorphism increases. The fact that some sills and irregular intrusive bodies of gabbroic composition are also more highly deformed and metamorphosed both to the northeast and southwest is significant evidence for two periods of folding and metamorphism separated by an invasion of basaltic magma.

The early period of folding resulted in folds with axes that trend northeast, the later period of folding developed ill-defined but northwest-trending folds. The faults in the metasedimentary rocks are probably of different ages. If we assume that there was only one main period of invasion of basaltic magma, most faults are pre-intrusion and may have developed during the early folding. However, since there is more than one invasion of basaltic magma in the quartzo-feldspathic gneiss and since dikes and sills of basalt are very common in the Mullen Creek-Nash Fork shear zone basalt of more than one generation may be present in the metasedimentary rocks. We have no satisfactory method of dating the faulting.

The refolded hinge of the syncline (Houston and Parker, 1963), in the southwest is a significant structural feature of the younger metasedimentary rocks. The lineation of the quartzite was folded by flexure. This refolding may be related to movement on the Mullen Creek-Nash Fork shear zone.

AGE DETERMINATIONS ON THE METASEDIMENTARY ROCKS

Two rocks from the Deep Lake Formation and three different units of the Libby Creek group have been dated. A biotite-garnet gneiss from higher rank units of the Deep Lake Formation near Arlington (Pl. 2) yields a tentative date of 1840 ± (?) m.y. by whole rock method. In the Libby Creek Group, a phyllite of the Headquarters Schist from a location near North Twin Lake (Pl. 2) yields a whole-rock-biotite date of 1550 ±40 m.y. A laminated schist from the Lookout Schist near Lake Marie (Pl. 2) gives an isochron of 1650 ±60 m.y. based on five whole-rock samples. The French Slate from a locality near Mill Pond (Pl. 2) gives an isochron of 1550 ±425 m.y. based on six whole-rock samples. The two dates of 1550 m.y. have been interpreted as metamorphic dates whereas the 1650 m.y. date could be a time of sedimentation but may also be a metamorphic date (Hills et. al., 1968). The significance of the 1840 ±(?) date in the Deep Lake rocks is unknown.

The older units (stratigraphically) do yield older ages, but since these units have probably suffered more than one metamorphic event, there need be no correlation between stratigraphic ages and dates for metamorphism. In fact, our structural interpretation would suggest that the higher rank gneiss from the Deep Lake Formation should appear younger than the phyllite of the Headquarters Schist if both dates are times of metamorphism.

At present, all we can say is that all these rocks are older than the youngest date (biotite; 1550 m.y.) and the Lookout Schist is at least 1650 ±60 m.y. old.

STRUCTURE OF PRECAMBRIAN ROCKS SOUTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

Domain One (S)

The structure of rocks south of the shear zone is best reviewed by subdividing this area into six domains. Domain one (S) is to the northeast (Pl. 4) and is bounded by the northeast trending Mullen Creek-Nash Fork shear zone on the north and a related east-trending shear zone on the south. The Precambrian rocks form a wedge-shaped block that has its greatest width on the east and tapers to a thin point on the west where the two bounding shear zones join. The greater part of this domain was studied by McCallum and the reader is referred to papers by McCallum (1964a, 1964b) for details of the structure of the rocks. Throughout this entire complex the most impressive feature of the rocks is cataclasis. Very few samples have been examined from this area that do not show some evidence of crushing and grinding. In addition to the general cataclasis, four shear zones that are zones of more intense granulation and mylonization (Pl. 4) cut this domain. A further subdivision of the rocks of the domain can be made by considering three subdomains that are bounded by these shear zones (Pl. 4). Subdomain A to the north is largely migmatite. The host rock for the granitic layers and veinlets is a cataclastic quartzo-feldspathic gneiss. Conformable layers of granite are located in the gneiss but some layers of granite locally cross-cut the cataclastic layering of the gneiss. The granitic layers are less deformed than the enclosing gneiss, but most layers of granite are crushed to some degree. Even cross-cutting granite veinlets are deformed in some areas by further movement along foliation planes of the gneiss. This migmatite complex is cut by numerous conformable mafic dikes or sills. These dikes are largely amphibolite and most have been subjected to cataclasis. The foliation of the rocks of subdomain A is developed by alignment of platy minerals during cataclasis and by segregation of minerals into layers during cataclasis. The strike of the foliation is northeast and is consistent. The dip is steep but variable ranging from 30 degrees southeast to 45 degrees northwest (Pls. 1 & 4), but averaging 70 to 80 degrees. Mineral lineations shown as biotite streaks on the plane of foliation and by preferred alignment of amphibole crystals plunge steeply south or southwest. Southwest-plunging minor fold axes have also been noted, and according to McCallum the minor folds may show folded mineral lineations.

Subdomain B is largely biotite-rich gneiss and augen gneiss. These rocks also show cataclastic structure. They grade into more granitoid rocks to the south and on Corner Mountain that are shown as quartzo-feldspathic gneiss on Plate 1. The quartzo-feldspathic gneiss is not so well foliated as the biotite gneiss and does not show strong granulation megascopically. It is marked by minor zones of cataclasis that can be seen under the microscope. Locally the quartzo-feldspathic gneiss is intensely sheared and is texturally similar to the cataclastic quartzo-feldspathic gneiss of subdomain one. The foliation of this subdomain strikes northeast and dips steeply. Most foliation surfaces dip 80 degrees southeast. No linear structures were observed in these rocks.

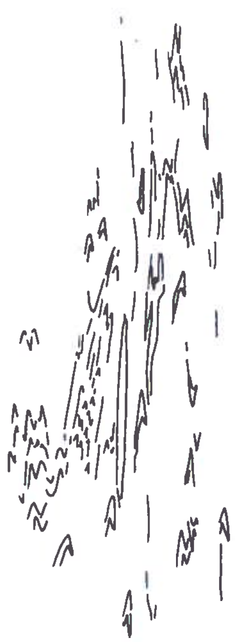
Subdomain C is composed of quartzo-feldspathic gneiss on the northwest and a layered sequence of hornblende gneiss, quartzo-feldspathic gneiss, biotite gneiss, garnet gneiss, and marble on the south (Pls. 1 and 4). The character of the quartzo-feldspathic gneiss is similar to that described in subdomain two, but southeast-plunging mineral lineations were noted in secs. 32 and 33, T. 16 N., R. 78 W. (Pls. 1 and 4). The layered sequence is a striking group of rocks characterized by interlayering on various scales of hornblende gneiss and quartzo-feldspathic gneiss. The layers are parallel and conform in strike to the foliation of the gneiss. Most outcrops show layering of some type with thickness of a few inches to 100 feet or more that could not be subdivided even on a map scale of 600 feet to the inch. On Plate one the units are shown subdivided by the most abundant rock type in any one area. This complex intermingling of rock types may have been produced by extreme attenuation and stretching of fold limbs during passive folding.

If these layered rocks were deformed by cataclasis, they were largely recrystallized during the process, because there is less evidence of crushing and grinding in these rocks than in those of the other two subdomains. The foliation of the rocks of subdomain three strikes northeast and is steeply dipping for the most part. Within the layered sequence most foliations dip south in the northwest and most dip north in the southeast (Pls. 1 & 4). The rocks of the layered sequence are cut by many small and large mafic igneous bodies that conform structurally to the foliation of the layered sequence. The mafic igneous rocks are all deformed to various degrees and may show textural variations ranging from slightly altered relatively massive gabbroic rocks to amphibolite with distinct layering. In general, the smaller bodies are most deformed and the larger bodies are more deformed along their borders. Felsic dikes or sills are common in this sequence, and these units also are structurally conformable. Like the mafic intrusives these bodies are variously deformed, but most show some evidence of cataclasis under the microscope and some are so deformed that the cataclastic structure can be recognized in hand specimen.

Pegmatites are also abundant in the layered sequence and are for the most part structurally conformable to the foliation of the gneisses, but some are cross-cutting and others show both relationships. The pegmatites are apparently post-shearing since few show evidence of deformation.

Domain Two (S)

Domain two (S) is an east-west zone in the central part of the area that is dominated by large igneous intrusives (Pl. 4). The domain is bounded on the north by the southernmost shear zone of domain one and on the south by an east-trending belt of small shear zones (Pl. 4). The western limit of this domain is Douglas Creek and the North Platte River. The Sherman Granite and related rocks are post-kinematic, but other igneous rocks were emplaced prior to or during the last period of deformation and metamorphism. Numerous small intrusives of both mafic and felsic composition are also earlier than the last period of deformation since all are foliated to some degree. Hornblende gneiss and related rocks including calc-biotite gneiss, sillimanite-biotite gneiss and hornblende diopside gneiss are the most abundant metamorphic rocks of this domain. The gneiss crops out in the southern part of the domain where it is in conformable contact with the early intrusives, the Keystone Quartz Diorite and the Mullen Creek mafic complex. Lenticular bodies of hornblende gneiss are enclosed within the quartz diorite and these bodies conform in strike to the trend of foliation in the quartz diorite. At the eastern limit of the quartz diorite, the contact between quartz diorite and hornblende gneiss is at an angle to the structure of the gneiss, and in certain zones in the quartz diorite there are swarms of inclusions of hornblende gneiss. The gneiss is obviously earlier than the quartz diorite but the general conformity of metamorphic foliation in both units suggests that both were metamorphosed after or during emplacement of the igneous rock. Through most of it's outcrop area the foliation of the gneiss is developed by fine layering of hornblende-rich and hornblende-poor layers. The orientation of platy minerals is parallel to this compositional layering. In some areas, coarser layers are present with thicknesses on the order of feet. In some better exposures of the gneiss it is evident that the layers are the attenuated limbs of passive folds (Fig. 44A), and this layering perhaps partly developed by passive folding is itself folded (Fig. 44B). The general strike of the foliation is east except for a small body of gneiss between the Keystone Quartz Diorite and the Mullen Creek mafic complex where the foliation trends generally north and in the area south of Sheep Mountain where foliation trends are northeast. Dip of the foliation is north mostly between 75 and 80 degrees (Pls. 1 & 4). In one area northeast of Fox Park, sec. 23, T. 13 N., R.



A

B

Figure 44—Field sketches of passive folds in hornblende gneiss of domain II S. A—Folded felsic layers in hornblende gneiss with preservation of hinges of folds and limbs greatly attenuated. Outcrop about six inches wide. B—Folded felsic layers in hornblende gneiss. Width of outcrop about three feet.

78 W., synforms and antiforms with axes that plunge 30° west-northwest are in the gneiss. This is one of the few areas where the gneiss could be studied in cross-section, and other folds of this type may be present, but even here both north and south of the folded gneiss the dip of foliation was north. Two types of lineation were noted in both gneiss and quartz diorite; a mineral lineation developed by biotite streaks on the plan of foliation and by preferred orientation of amphibole crystals, and a lineation developed by the plunge of axes of minor folds. The majority of both types of lineations plunge northwest (Pls. 1 & 4).

The southern part of domain two (S) is underlain by quartzo-feldspathic gneiss whose foliation strikes similarly to that of the hornblende gneiss and quartz diorite. Another unit with structure conformable to that of the hornblende gneiss is quartz-biotite-andesine gneiss which is in T. 13 N., as large and small tabular inclusions in the quartz diorite and in a thin tabular body along the southeastern border of the quartz diorite (Pl. 1). In the area between the Mullen Creek mafic complex and the Keystone Quartz Diorite, hornblende gneiss grades into quartz-biotite andesine gneiss along strike. The quartz-biotite andesine gneiss rims the northwestern border of the Keystone Quartz Diorite and is in a southwest-plunging synform on the northwest side of the intrusion (Pl. 1). This gneiss grades into quartzo-feldspathic gneiss and quartz-biotite andesine gneiss is deflected to the northeast at the northern border of domain two (S) (Pl. 4). This change in strike defines a fold about a steeply plunging axis and is probably related to movement along northeast trending structures in domain one.

Perhaps the most interesting structures of domain two (S) are those of the older intrusives. The Mullen Creek mafic complex in the western part of the area is a complex of mafic rocks that is variously metamorphosed. Local areas in the complex are massive gabbroic rocks that are layered and have a primary igneous foliation parallel to the layering. Where noted, this foliation trends northwest and dips close to 90 degrees (Pl. 1). Metamorphic foliation developed by shearing and recrystallization of the gabbroic rocks trends northeast along the northwestern border of the complex parallel to the strike of the Mullen Creek-Nash Fork shear zone, and along the southern border of the complex, the metamorphic foliation parallels the strike of foliation of the hornblende gneiss. Borders are gradational in this southern area suggesting that a part of the gneiss was derived from the mafic rocks by shearing and recrystallization. The bodies of granite within this complex are also foliated, and although contact relationships show the granite is later than the gabbroic rocks, some outcrops of the granite are better foliated than the earlier gabbro. Some of the foliation in the granite is cataclastic, but many thin sections examined show a preferred orientation of platy minerals in ungranulated rocks.

The structure of the Keystone Quartz Diorite has been partly reviewed in the discussion of hornblende gneiss and in the description of the quartz diorite. The northeast-striking foliations in the western part of the body are cataclastic, but the east-striking foliation in the south and east does not show cataclastic textures. Several east-trending shear zones transect the southwestern part of the quartz diorite. Foliation in and related to these zones is cataclastic. As stated above, both of these large intrusive bodies have been metamorphosed, and the only good clue to the original structure of these intrusives is the northwest striking primary foliation of the Mullen Creek mafic complex.

Late intrusive bodies of domain two are the Lake Owens mafic complex, Sherman Granite, pegmatites, and felsic dikes. The rocks of the Lake Owens mafic complex cut the hornblende gneiss and Keystone Quartz Diorite. Their relationship to the foliated granite is not known because of poor exposures at contacts. The unit is regarded as late because of its cross-cutting relationships to other rocks. The steeply dipping foliation of the unit (Pl. 1) is regarded as primary except along the extreme southern border of the unit where an east-trending metamorphic foliation is present.

Pegmatites are abundant in the domain especially in the hornblende gneiss (Pl. 1). The pegmatites are later than all units with the exception of Sherman Granite and the felsic dikes. Their relationship to the rocks of the Lake Owens mafic complex is not known since they do not cut this unit. They are regarded as earlier than the Sherman Granite because age determinations on similar pegmatites in domains three (S) and four (S) range between 1.5 and 1.6 b.y. whereas the granite is between 1.35 and 1.4 b.y. (Hills et. al., 1968). The pegmatites are cut by the felsic dikes. A few of the pegmatites may possibly be late since one of two were noted in the Sherman Granite, but these could not be distinguished from others in the hornblende gneiss. Most pegmatites are conformable, but many cut the foliation of the gneiss at right angles, and some show both conformable and cross-cutting relationships with the gneissic structure. The cross-cutting pegmatites are more common in metaigneous rocks and most trend northwest (Pl. 1).

The Sherman Granite of domain two (S) is a large disconformable wedge-shaped body in plan that underlies most of Sheep Mountain and extends into the Medicine Bow Mountains proper northwest of Albany (Pl. 1). The granite is massive to faintly foliated. One northeast-trending fault cuts the granite in an area southeast of Cinnabar Park (Pl. 1). The granite may be sheared at the border between domains one and two (at the northern border of the granite where an east trending shear zone is present) but as noted by McCallum the intensely sheared rocks were all part of the layered sequence of domain one and no sheared granite could be found. The granite does border the shear zone however, and there was no evidence of granite cutting sheared rocks.

The youngest igneous rocks are felsic dikes emplaced in north to northwest-striking fractures. The most prominent of these dikes cuts the western border of the Lake Owens mafic complex. The dike has been regarded as Tertiary (?) by Currey (1965), but there is no conclusive evidence of this. Another north-trending dike is east of Fox Park and several northwest-trending dikes are along the east margin of the mountains near Woods Landing.

Northeast-trending shear zones are common along the northwest border of domain two (S). Many of these intensely sheared areas must have been developed in the metamorphic rocks without important horizontal movement (Pls. 1 & 4). East-trending shear zones are common in the southern part of the Keystone Quartz Diorite and at the southern border of the domain. Again map views of these structures (Pls. 1 & 4) do not indicate major horizontal movement. Most of these sheared zones are probably not major faults, but simply areas that yielded to stress by crushing and granulation rather than folding.

The Mullen Creek-Nash Fork shear zone in the northwest corner of domain two (S) is an extremely complex zone of mixed rocks. The major rock type in the shear zone is sheared quartz monzonite and quartz diorite, but sills and dikes of granitic and gabbroic composition are common. This portion of the shear zone may have been invaded by felsic magma that also cuts the Mullen Creek mafic complex. If so, the felsic rock was also sheared, but not as severely as elsewhere in the fault zone, because mylonite is not as common here as in the areas to the east and west.

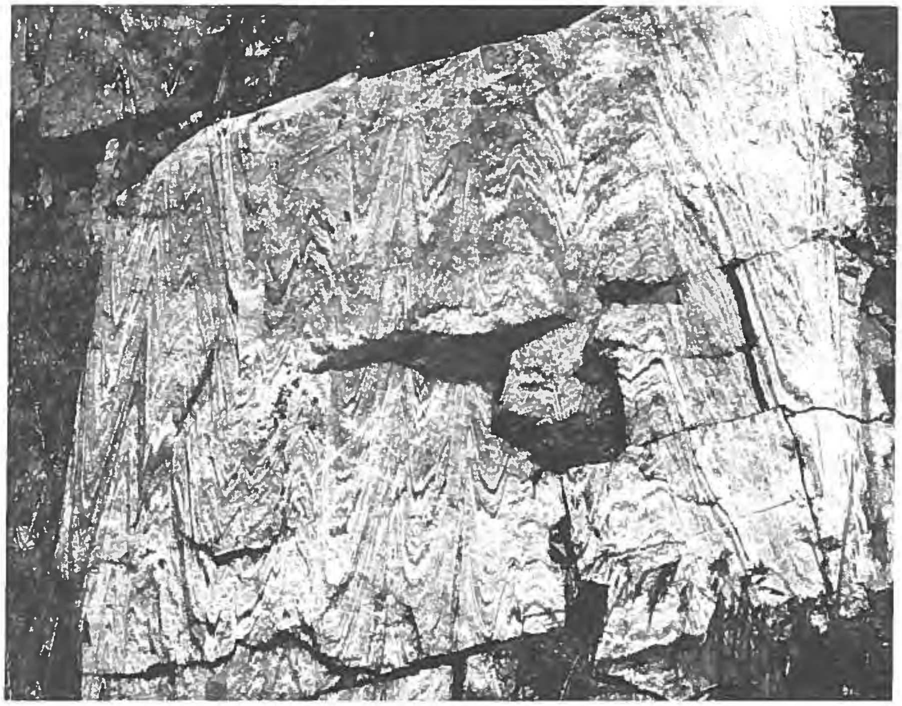
A number of small northwest-trending shear zones were noted along the western margin of the Keystone Quartz Diorite by Currey (1965). Many of these shear zones are mineralized with copper sulphides and native gold.

Northwest trending faults are common in the eastern part of the domain, south of the Sherman Granite (Pls. 1 & 4). Many of these faults are breccia zones partly filled with calcite. Some may be Laramide rather than Late Precambrian. A half-moon shaped fault is present in the southeastern part of domain two (Pls. 1 & 4). This fault zone is filled with epidote and quartz and the fault is probably Precambrian.

Domain three (S)

Domain three (S) is bordered on the northwest by the Mullen Creek-Nash Fork shear zone, the North Platte River and Douglas Creek on the east and Elkhorn Point on the south (Pl. 4). The structure is better studied in this area than in other areas south of the fault because of superior outcrops.

Beta diagrams (Pl. 4) show a major difference in structure between rocks of this domain and the quartzo-feldspathic gneiss north of the fault. In contrast to the general northwest strike of foliation and the prevalence of north- or northwest-plunging antiforms in the area north of the fault, the foliation of domain three strikes northeast to east-northeast, and antiforms plunge steeply east-northeast (Pl. 4). The steeply plunging antiforms are refolded as shown by folded mineral lineation and by the general geometry of the folds. Minor folds developed during the last period of folding contain folded mineral lineation (Fig. 45). The regional plunge of these minor folds in domain three is northeast or east-northeast (Pl. 4) corre-



A

Passive folds showing greatly attenuated limbs of folds and thickening in fold hinge

B

Close-up of passive folds in upper left hand part of A showing detail of structure. Note extreme attenuation of dark amphibole-rich layer in upper right.

Plate 31—Passive folds in hornblende gneiss in Big Creek area.

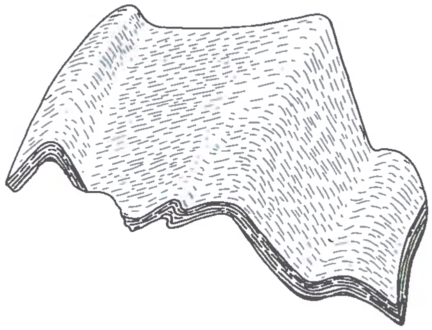


Figure 45—Mineral lineation (preferred orientation of amphibole) folded by flexure of earlier foliation. Specimen about 8 inches from left to right. Collected from near axis of fold in subdomain III S B.

sponding roughly to the plunge of major antiforms and synforms.

Domain three (S) has been divided into subdomains having similar geometric properties (Pl. 4). The most important of these subdomains will be reviewed below. A detailed map has been prepared of subdomains A, B, C, and D (Pl. 3). In subdomains A - D antiforms and synforms plunge east-northeast (Pl. 3). The major rock type of this area is hornblende gneiss that is interlayered with quartzo-feldspathic gneiss. Near the cores of antiforms the gneiss is more massive and is largely quartzo-feldspathic gneiss. The gneiss is cut by olivine gabbro of the Mullen Creek mafic complex in the north and both gabbro and gneiss are cut by medium-size bodies of granite and granite sills. Pegmatite is especially common in the hornblende gneiss and is later than the gneiss. With the exception of pegmatite and granite dikes, which are locally crushed, all of these rocks are affected by the late folding.

Subdomain three (S)B is an antiform bordered on the north by an east-trending shear zone. The structure of this domain is best shown in Plate 3. This antiform is a compound structure having elements of east-trending and northeast-trending fabric. This can be noted in the beta diagram (Pl. 4) where an east-trending set of S surface intersections and a northeast-trending set of S surface intersections is shown. The fold has two types of lineation, mineral lineation on the plane of foliation and minor fold axis. There are probably two sets of "mineral" lineations, one developed by the intersection of east trending and northeast trending foliation planes and a second set in which the mineral lineation is sub-parallel to the axes of passive folds (Fig. 46) in the gneiss. Mineral lineations in which the orientation of minerals nearly parallels the axes of passive folds have been observed in the field and are designated as such in Plate 3. Other mineral lineations were not identified as to type in the field. One may recognize folded mineral lineations in the minor flexural folds (Fig. 45). There are also two sets of these minor folds, one plunging east and another northeast.

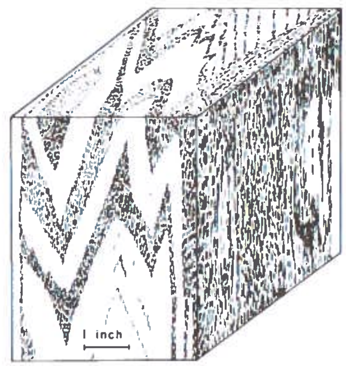


Figure 46—Sketch of passive fold in hornblende gneiss showing amphibole-rich layers (dark) with orientation of amphibole crystals in a direction subparallel to the axis of the fold. Sample from outcrop along Big Creek; (approximate width of front face of block 4 inches.)

The early formed passive folds have been refolded about a new axis and are exposed as an antiform. East of Big Creek the attitudes of mineral lineation change around the nose of the antiform, and minor folds that probably developed during the late folding show visibly folded mineral lineation (Pl. 3). Figure 47 is a plot of mineral lineations measured east of Big Creek along with late minor fold axes. Both types of lineation show a considerable spread, but the minor fold axes roughly coincide with a secondary β concentration in the northeast (Pl. 4). The distribution of the "mineral" lineations could be interpreted as a small circle distribution about the β concentration in the northeast and the late minor fold axes; or if one observation (the easternmost in Figure 47) is ignored a great circle distribution about β. Thus we do not know if the late folding was passive or flexural. Additional study of linear features would seem to be the obvious answer to this but measurement by Professor R. B. Parker of the University of Wyoming of more than 100 lineations on this structure also failed to show a distribution of linear features that could be classed as refolding by flexural or passive mechanism.

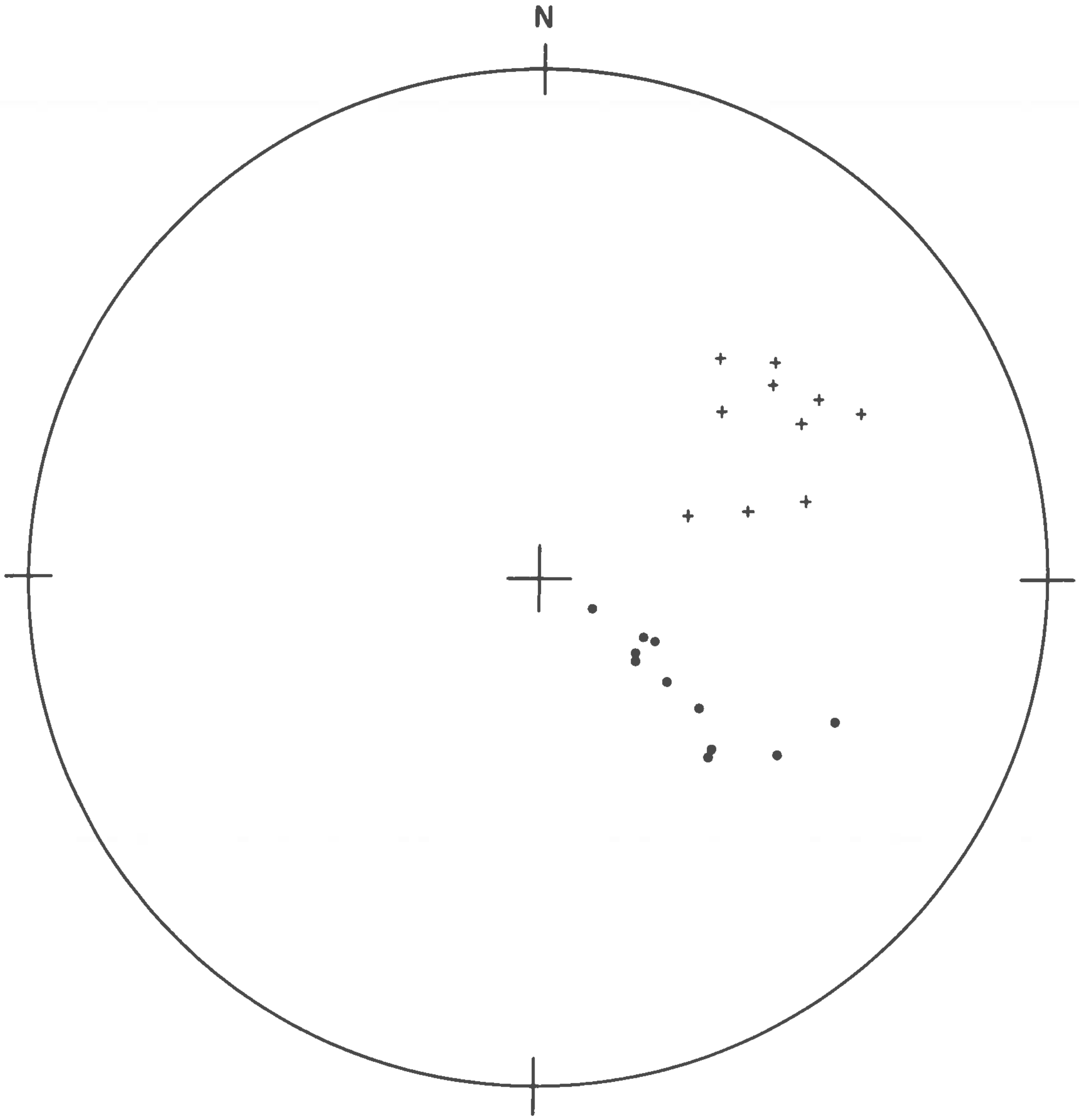


Figure 47—Mineral lineations from subdomain III S B. Circles are mineral lineations measured east of Big Creek. Crosses are minor flexural folds that show folded mineral lineations. Plot on equal area (Schmidt) net, lower hemisphere.

This structure may have developed by deformation of a pre-existing structure by a combination of flexural and passive movement. We have probably measured S surfaces and "mineral" lineations of more than one type. The east-trending concentration may be caused by passive flow whereas the ill-defined northeast concentration may result from flexural folding about a new northeast axis. The initial folding may have been by flexure which gradually changed to passive movement. Certainly the plan view of the structure in the west (west of Highway 230) and in the north suggest a type of passive fold (Pl. 3).

Subdomain three (S)A is a synform plunging east-northeast bordered on the south by an east-trending fault and on the north by gabbro of the Mullen Creek mafic complex. The synform is in hornblende gneiss, but granite occupies the axis of the synform in the north. On the surface (in plan view) the granite grades into hornblende gneiss through augen gneiss and quartzo-feldspathic gneiss. In cross-section, however, as noted in exposures along Big Creek, the granite can be seen to cut both hornblende gneiss and quartzo-feldspathic gneiss locally. The entire rock sequence is foliated, however, with the texture in

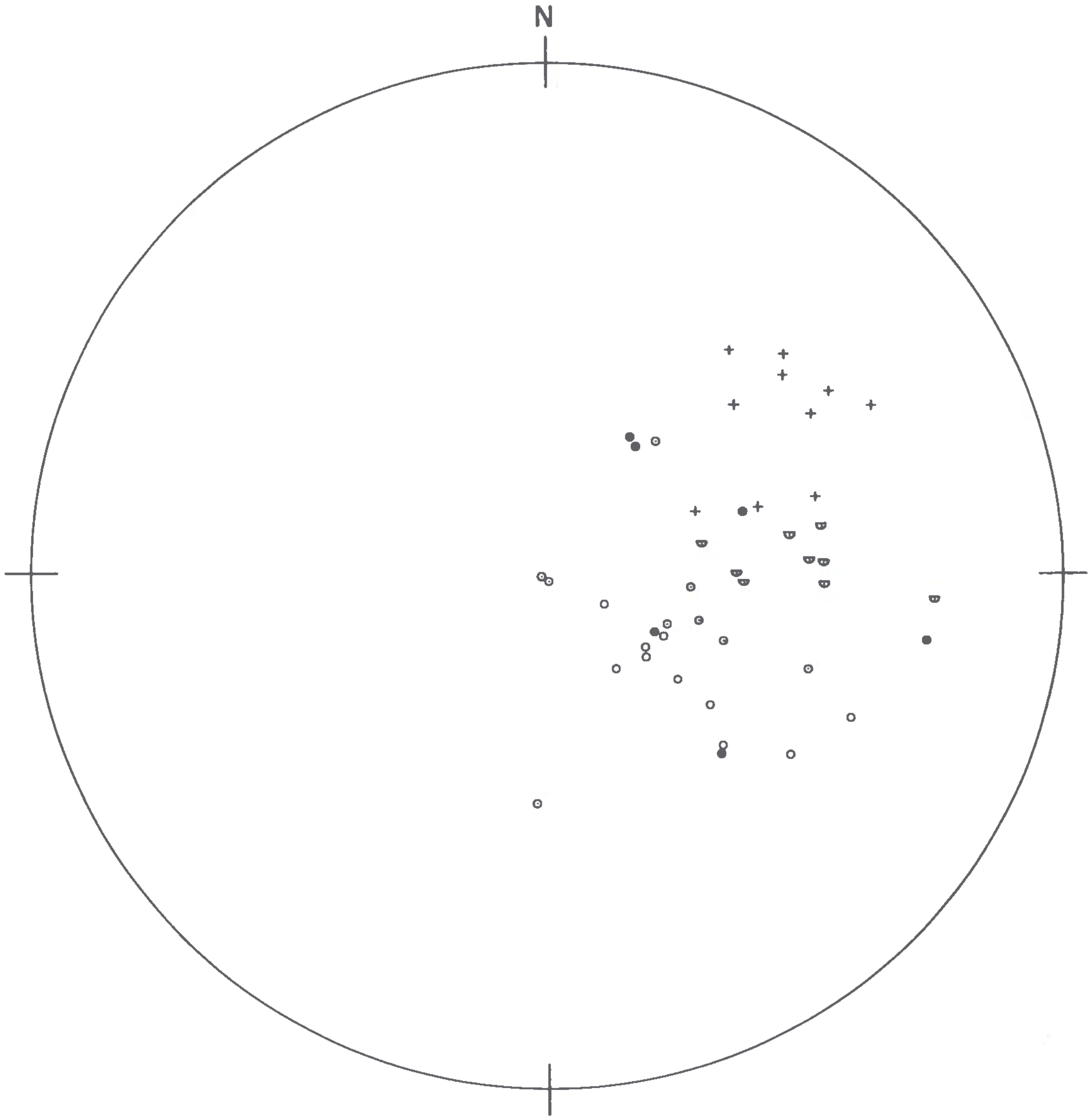


Figure 48—Plot of all linear features in subdomains III S A and III S B. Crosses are minor folds (late folds of the flexural type with folded mineral lineation), half moon are lineations that may be caused by intersections of planes, open circles are mineral lineations measured in eastern part of III S B, circles with dots are mineral lineations measured in III S A, and solid dots are mineral lineations where axis of passive fold nearly parallels mineral lineations. Plot on equal area (Schmidt) net, lower hemisphere.

the granite strongly cataclastic. The rock sequence, including granite, has been deformed.

This synform has probably been refolded in the same manner as the antiform of domain three (S)B. The mineral lineations shown in Plate 3 shows the same type of distribution as in three (S)B. Figure 48 is a plot of the linear features in both domains three (S)A and three (S)B. These linear features show a considerable variation suggesting refolding, but we have no satisfactory interpretation for their distribution.

Subdomain three (S)C is an antiform with a distinctive geometry (Pl. 4). Figure 49 is a sketch of the strike of foliation in this structure as observed on the ground and as seen on air photos. The structure is a type of interference pattern produced by superposition of the one fold set on another. According to Ramsay (1963, p. 151-153) there are only three fundamental types of interference patterns and a comparison with Ramsay's illustration suggests that the pattern in subdomain C is a type where the movement direction of the second fold makes a high angle with the axial planes of the first folds. An interference pattern of this type could be produced by interference of first folds with axial planes that trend northwest and second folds with axial planes trending northeast. This is significant inasmuch as folds in older units north of the shear zone have northwest-trending axial planes. Mineral lineations and minor folds were not studied in detail in subdomain C.

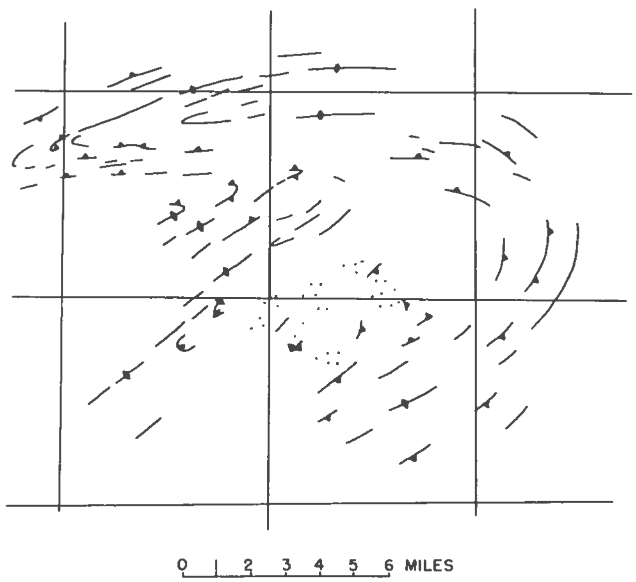


Figure 49—Sketch of trend of foliation in subdomain III S. C., showing interference pattern produced by superposition of one fold set on another. Dotted area more massive quartzo-feldspathic gneiss.

The east-striking shear zone separating domains three (S)A and three (S)B is one of cataclasis and silicification that has been traced about 8 miles along strike. Cataclastic foliation in this zone has a vertical dip. Another lesser shear zone that strikes northeast and has a cataclastic foliation that dips 70° northwest is located north of Elkhorn Point at the southern margin of domain three (S).

Domain Four (S)

Domain four (S) is south of the general east-striking belt of shearing that marks the southern limits of domains two (S) and three (S), and extends from the western border of the Medicine Bow Mountains to the east limit of R. 79 W. (Pl. 4). Rocks of this domain are largely quartzo-feldspathic gneiss on the west and hornblende gneiss on the east. The regional structural trend is northeast except for an area in the northwestern part of the domain and a belt of northwest-striking structures in the south-central part of the domain (Pls. 1 & 4).

In the western part of domain four (S) the quartzo-feldspathic gneiss contains many conformable bodies of amphibolite. In Plates 32 and 33 which shows the structure in greater detail, both amphibolite and quartzo-feldspathic gneiss are complexly folded. The northern structure in subdomain four (S)A (Pl. 4) is a circular area outlined by steeply dipping foliation of the gneiss and amphibolite. The general dip of foliation is to the center of the structure except along the western margin where dips are eastward. The general attitude of the amphibolite bodies resembles that of a series of cone sheets, but it is more likely a compound structure developed by repeated folding.

Several northeast trending shear zones are present that may be related to the larger shear zones that bounds domains two, three, and four. Mineral lineations which are biotite streaks on the plane of foliation, have been recorded in this structure (Pls. 1 & 4). Most of the lineations trend and plunge steeply northeast.

The circular structural pattern is cut by an east-striking fault at its southern margin. South and southeast of this fault, a spectacular series of folds, that resemble finger prints in plan view are developed (Pls. 32 and 33). This area was mapped by W. G. Myers in 1958 (Myers, 1958, p. 1738) as one of the first of the mapping projects in the Medicine Bow Mountains. Although many interesting and complex structures have been mapped in the area since this time none have been as enigmatic as this. It is the writer's opinion that the continuity of individual units shown in Plate 32 is questionable, because outcrops are such that individual amphibolite bodies cannot be traced along their entire length. However, the structural plan can be observed on air photographs (Pl. 33). Both northwest and northeast-trending structures are in this area. The complex surrounding the barbell-shaped Parkview gabbro trends northwest whereas the structure south of the gabbro and northeast of the gabbro trends northeast. A belt of northwest-trending structures which also includes a body of gabbro is present to the southeast (Pl. 4). One of the structures in the belt of northwest-trending structures is

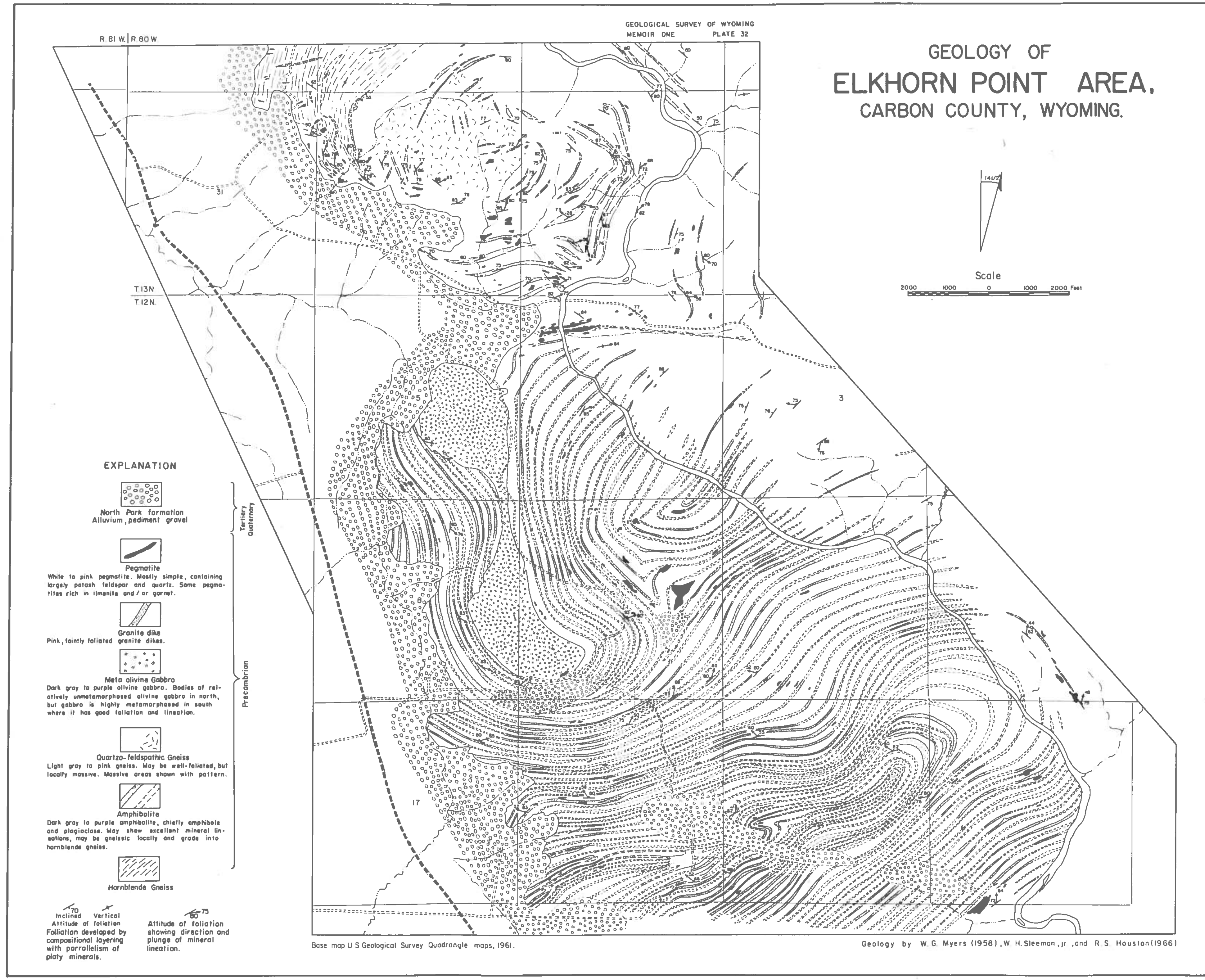

GEOLOGICAL SURVEY OF WYOMING
MEMOIR ONE PLATE 32
GEOLOGY OF
ELKHORN POINT AREA,
CARBON COUNTY, WYOMING.
Scale
2000 1000 0 1000 2000 Feet
R 81 W. R 80 W.
T13N
T12N
EXPLANATION
North Park formation
Alluvium, pediment gravel
Tertiary
Quaternary
Pegmatite
White to pink pegmatite. Mostly simple, containing largely potash feldspar and quartz. Some pegmatites rich in ilmenite and / or garnet.
Granite dike
Pink, faintly foliated granite dikes.
Meta olivine Gabbro
Dark gray to purple olivine gabbro. Bodies of relatively unmetamorphosed olivine gabbro in north, but gabbro is highly metamorphosed in south where it has good foliation and lineation.
Quartzo-feldspathic Gneiss
Light gray to pink gneiss. May be well-foliated, but locally massive. Massive areas shown with pattern.
Amphibolite
Dark gray to purple amphibolite, chiefly amphibole and plagioclase. May show excellent mineral lineations, may be gneissic locally and grade into hornblende gneiss.
Hornblende Gneiss
Precambrian
Inclined Vertical
Attitude of foliation
Foliation developed by compositional layering with parrallelism of platy minerals.
Attitude of foliation showing direction and plunge of mineral lineation.
Base map U S Geological Survey Quadrangle maps, 1961.
Geology by W. G. Myers (1958), W. H. Sleeman, jr., and R. S. Houston (1966)

Plate 33. Aerial Photograph of Elkhorn Point area.

closed. This belt of northwest structures is bounded on the northwest and southeast by rocks with northeast-trending foliation. Throughout this area foliation in both the gneiss and amphibolite layers is steeply dipping and the foliation conforms to the strike of individual layers of amphibolite. Steeply plunging mineral lineations of the planes of foliation of gneiss and amphibolite are consistent in plunge in individual structures but show no regional consistency (Pl. 4).

If the amphibolite layers were originally igneous these structures may be folded dikes or sills, but individual dikes would have had to be introduced in a fracture system with remarkably regular spacing. One must also find some mechanism to explain the abutment of northwest and northeast trending folds shown in the area east of the Parkview gabbro (Pls. 1 & 4).

Stevens (1957, p. 357-364) offers an alternative to the idea that these are folded dikes or sills. He considers the closed structure in the south-central part of domain four as a diapiric structure developed during upward movement of mobilized gneiss. The general structural pattern of these units does resemble diapiric structures developed in salt domes (Balk, 1949, 1953; Hoy, 1962; Muehlberger, 1959, and Kupfer, 1963), but on a much larger scale. The complex fold patterns seen in plan in salt mines develop by upward movement and attenuation of original horizontal bedding of evaporite sequences. If we propose a similar origin for these units, the original layers of amphibolite in the gneiss may have been sub-horizontal sills or even mafic tuffs in the gneiss, now greatly attenuated. Such a concept would also have to explain the divergent orientation of structures, but interference folds as well as closed folds are found in salt structures (Kupfer, 1963).

The problem of origin of these structures is not solved and will require a much more detailed study than was possible in this investigation; but regardless of genesis, the area has two structural elements that trend northwest and northeast.

A series of poorly developed, northwest-trending shear zones are found along Pelton Creek (Pl. 4) in domain four (S). East of these shear zones the major rock type is hornblende gneiss, whereas west of the shear zone the major rock type is quartzo-feldspathic gneiss. There is no way to determine the displacement on these faults, but the apparent offset does not appear large in the southern part of the area. East of Pelton Creek exposures are not good in the Precambrian, and mapping has been largely reconnaisance, but this appears to be an area of interference between northwest- and northeast-trending structures.

Domain Five (S)

Domain five (S) is located in the southeast sector of the Medicine Bow Mountain (Pl. 4). The major rock types are quartzo-feldspathic gneiss and hornblende gneiss with a general east-to east-northeast-striking foliation. At the extreme southeast corner of the domain, a body of quartz-biotite andesine gneiss has a northeast foliation. These units are cut by Sherman Granite in the southern part of the domain which is, for the most part massive, coarse-grained granite. Late north-striking and northwest-striking felsic dikes cut the gneiss, and the Sherman Granite is cut by two northwest-striking faults in its southern part. The fault on the west is mineralized.

Domain Six (S)

Domain six (S) is at the north end of Sheep Mountain. The rocks of this area are similar to those on Centennial Ridge but the structural trend of the units is northwest instead of northeast. The major rock type is hornblende gneiss which is locally calcareous and contains lenses of marble. There is no marker bed in the gneiss but the foliation of the gneiss which is defined by compositional layering and parallel oriented platy minerals is complexly folded. Both large and small folds, which have northwest-striking axial planes that dip to the southwest have been recognized in the gneiss. Most of the minor-fold axes noted plunge approximately 40° southeast (Pls. 1 & 4). The hornblende gneiss is cut by granite that has a cataclastic foliation that roughly parallels the foliation of the hornblende gneiss in some areas. This gneissic granite clearly cuts the hornblende gneiss which is locally engulfed by numerous stringers and irregular veinlets of granite. Massive, coarse-grained Sherman Granite cuts this group of rocks at nearly right angles to the structure of the gneiss (Pl. 1). The older granite is probably equivalent to granites in domain two where their age relationship to the Sherman Granite was not known, but here there is no question but that the Sherman Granite is later. The northwest structural grain of this area suggests that this is part of the zone of northwest-trending structures located west of the Sherman Granite contact in the northern part of domain two (Pl. 4). The rocks of this area may have been protected from severe deformation during the late orogeny because they were located between the large mafic bodies on the south and were bounded by a shear zone on the north.

METAMORPHISM

The older rocks of the Medicine Bow Mountains, the quartzo-feldspathic gneiss and related rocks of the northwest and the gneissic units south of the shear zone, belong in the almandine amphibolite facies (Fyfe, Turner, and Verhoogen, 1958, p. 228) of regional metamorphism. In both the northwest and south, typical mineral assemblages of quartzo-feldspathic gneiss are quartz, microcline, plagioclase, An_{15-31}, biotite, epidote, and muscovite. In both areas typical mineral assemblages in hornblende gneiss are quartz, plagioclase, An_{21-59}, amphibole, epidote, and sphene. South of the Mullen Creek-Nash Fork shear zone calcareous rocks have such minerals as calcite, diopside, epidote, garnet, green-brown hornblende, plagioclase, An_{40-55}, and quartz ($\pm$ microcline).

Mafic igneous rocks in both the northwest and south show a wide range in intensity of metamorphism. This has been reviewed in the descriptions of these units, but particular note should be made of the fact that a single body of mafic rock may range from diabase to gneissic amphibolite. In the northwest the conformable dikes and sills are more highly metamorphosed than the cross-cutting bodies. However, in individual conformable bodies we find portions of the unit with original igneous textures and mineralogy in which there is labradorite, hypersthene, and augite, only slightly altered. The plagioclase is commonly fresh, but the pyroxene generally partly altered to green amphibole and quartz. These relatively unaltered areas grade to completely recrystallized orthoamphibolite composed of andesine, dark-green to brown amphibole, quartz, sphene, garnet, and epidote. We deal with rocks with texture and mineralogy partly retained or orthoamphibolite of the almandine amphibolite facies. There are no mafic rocks of intermediate grade. The same characteristics hold for the cross-cutting bodies except reaction rims are more common in these units.

South of the fault most of the small mafic bodies, dikes and sills, do not retain original textures. Most of these are of uncertain origin, but those considered orthoamphibolite contain bluish-green to green-brown amphiboles, andesine to sodic labradorite, and quartz, ($\pm$ epidote, and sphene). Larger gabbroic intrusions show the same characteristic as the cross-cutting bodies in the northwest.

The igneous rocks of basaltic composition have followed what has been called the **abnormal** trend of metamorphism (Wiseman, 1934). That is, these units essentially change from igneous rocks with preserved texture to amphibolite of the almandine amphibolite facies. This type of response in mafic igneous rocks has been noted by Wiseman (1934) and been described in detail by Sutton and Watson (1951) in the Northwest Highlands of Scotland. The geologic setting in the Northwest Highlands, is very similar to that of the northwestern part of the Medicine Bow Mountains because the dikes that have responded to metamorphism in the **abnormal** manner are in gneissic units (biotite and hornblende gneiss) that have been refolded. If dikes cut gneiss and are then refolded with the gneiss during a second episode of metamorphism, one might assume that the second metamorphism was accomplished under dry condition. This **abnormal** response to metamorphism shown by the mafic dikes may be a type of metamorphism developed in a water deficient environment (Poldervaart, 1953, p. 265-266).

METASEDIMENTARY ROCKS

The metasedimentary rocks of the Deep Lake Formation and Libby Creek Group are of lower rank than metamorphic rocks of the older units. These rocks are generally of the greenschist facies of regional metamorphism, but there is a regional increase in rank both to the northeast and southwest. The key mineralogical changes are shown in Table 65 and the locations of the various zones in Plate 2. The changes in mineralogy are found in areas that were recognized in the field as zones of increasing deformation and probably increasing rank of metamorphism. No attempt was made to map isograds in the field, but enough samples were studied petrographically to make a rough approximation of regional mineral variations.

Table 65—Regional mineral zonation of metasedimentary rocks

Rock Type	Mineral	Central Zone	Zone 2	Zone 3	Zone 4
Pelitic Rocks (some calcareous)	Quartz				
	Muscovite				
	Chlorite				
	Biotite				
	Garnet				
	Chloritoid				
	Albite				
	Oligoclase				
	Andesine				
	Amphibole (b-g)				
	Epidote				
	Staurolite				
	Kyanite				
Mafic Igneous Rocks	Quartz				
	Pyroxene (P)				
	Plagioclase (P)				
	Chlorite				
	Clinozoisite				
	Amphibole (BG)				
	Epidote				
	Garnet				
	Biotite				
	Albite				
	Andesine				
Greenstone	Quartz				
	Calcite				
	Actinolite				
	Chlorite				
	Albite				
Metadolomite	Dolomite				
	Quartz				
	Tremolite				
	Zoisite				
	Oligoclase				
	Microcline				
	Calcite				
	Biotite				

b-g = brown-green, BG = blue-green, P = primary

Zones correlate with Plate 2

The pelitic rocks (Table 65) include rocks of varying composition, from normal slate to calcareous phyllites and schists. For this reason, the appearance of chloritoid, garnet, and possibly amphibole may reflect compositional changes of the rocks as much as facies changes. In the northeast near Arlington, some of the higher rank rocks may possibly be part of the older basement infolded with the metasediments, but, even so, the increase in grade is apparent in metasedimentary rocks that probably correlate with those of the central area. South of the central area, there is inadequate sampling of pelitic rocks to determine facies changes (Pl. 2), but in zone four the pelitic units are of the almandine amphibolite facies with such assemblages as kyanite, biotite, quartz, and muscovite, calcite, blue-green amphibole, biotite, calcic oligoclase, and quartz, and staurolite, quartz, biotite, muscovite, garnet, and kyanite.

The metadolomite has been studied only from the central area and zone two. The mineral assemblages shown (Table 65) are of the greenschist facies, if oligoclase is considered to be a clastic mineral in this suite.

Greenstone (Towner Greenstone) was studied only from the central area where the mineral assemblages belong to the greenschist facies.

The mafic igneous rocks show interesting mineralogic changes both in dikes of basaltic composition and gabbro. With the exception of the appearance of albite in some samples, the response to metamorphism is not unlike the **abnormal** response shown by these units in the gneissic terrain. Our information to date shows that mafic rocks containing albite are the dikes and sills (including the Towner Greenstone) in the pelitic rocks and metadolomite of the Libby Creek Group. These rock types show what has been termed the **normal** response of mafic rocks to metamorphism, i.e. development of minerals typical of the greenschist facies. However, the large gabbroic bodies do not have albite, and show the **abnormal** trend. The gabbroic rocks were introduced after the early period of folding of these sediments, and they may have been metamorphosed in a late event affecting all units. This late event may have been one of "dry" metamorphism. Also the gabbroic units studied most carefully are primarily in quartzite that does not have water as a major component, and therefore water may not have been available when they were metamorphosed.

A major problem in this interpretation is the presence of kyanite in the Medicine Peak Quartzite which is overlain and underlain by rocks considered to represent a greenschist facies. The kyanite is considerably more abundant in the southwest where pelitic rocks are higher in metamorphic rank, but it is nevertheless common in the central area of assumed low rank pelitic rocks.

The complex structural history proposed for the Precambrian of the Medicine Bow Mountains carries a connotation of a similar complex history of metamorphism. Inasmuch as all rocks are of the almandine amphibolite facies except the metasedimentary rocks in the northern axial remnant of the mountain mass, it is apparent that successive metamorphic events accompanying structural events have either been from low grade to high grade, or that successive metamorphic events have been of nearly the same metamorphic grade. If we assume, as is suggested by study of gneiss and dikes in the western part of the mountain, that gneiss was present prior to dike introduction then folded dikes may indicate that once established the almandine-amphibolite facies minerals of the gneiss either persisted or were recrystallized under the same conditions in which they formed. The evidence for an ancient gneiss- and granite-forming event of 2.4 b.y. preserved at Baggot Rocks and Cedar Ridge suggests that, locally, early formed gneiss and granite has not been completely recrystallized during a second metamorphic event.

Another factor involved in this seeming persistence of the same metamorphic grade may be the fluid phase. Once almandine-amphibolite grade is established in a series of rocks the water available to promote mineralogical transformation in the rocks may be decreased 10 fold (Balk, 1936) over the amount available in original slate and even more over the amount available in shale or graywacke. Even if temperature-pressure conditions are established in which a low grade rock could form, the lack of water and other volatiles in the rocks undergoing metamorphism may prevent the establishment of the new lower grade except in shear zones or fractures where volatiles may be locally concentrated. The fact that dikes and gabbroic bodies simply show a direct transformation to almandine-amphibolite grade suggests that the rocks were either metamorphosed a second time under temperature-pressure condition of this facies or water was unavailable for transformation to the greenschist facies.

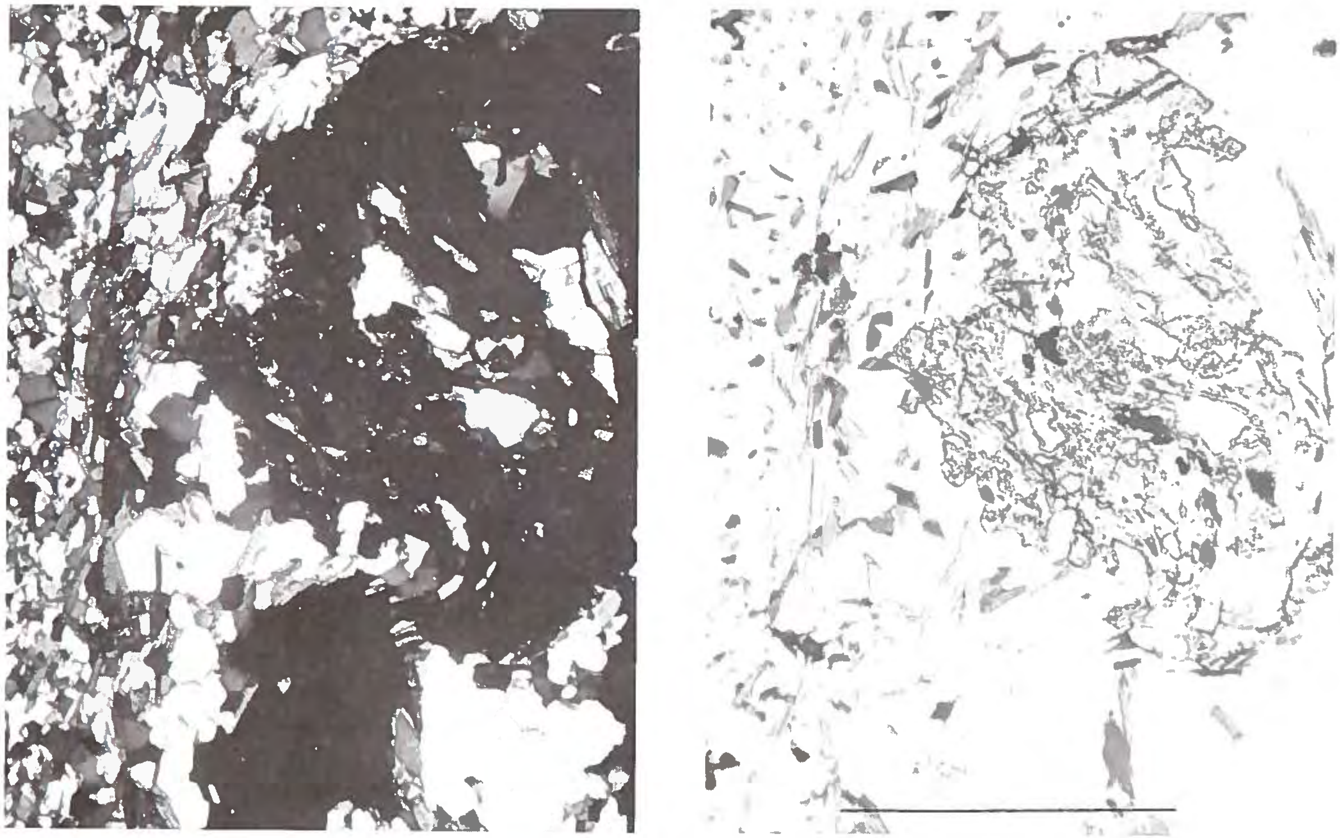
This problem of persistence of high-temperature metamorphic assemblages has been discussed in terms of reaction kinetics by Fyfe (1958, p. 96-100) who concludes that persistence of high-temperature forms are likely for polymetamorphic transitions in a "dry" system but less likely if the system is hydrous. Perhaps the real problem is the availability of adequate fluid phase to promote any significant recrystallization in a second metamorphic event. The dikes and other mafic igneous rocks obviously do recrystallize, however, and in the process some water must become available either from external sources or from the mafic rock itself. In this respect, the dikes and sills in the little disturbed central area of metasedimentary rocks where rocks of greenschist facies are present are particularly interesting. These dikes were emplaced after the first period of folding, and since they are in the center of the complex they are undeformed by the second episode of folding. They are altered, however, and where present in the carbonate and slates some of the dikes have minerals especially along borders (i.e. albite, chlorite) characteristic of greenschist facies. Undeformed dikes in quartzite are also altered but are amphibolized and do not contain albite. Perhaps during a second episode of folding and metamorphism or a late thermal event, these dikes have responded to temperature changes dependent on available water and volatiles. Where water is present and deformation is not extreme as in the slates and carbonates the dikes are of greenschist facies, but where water is absent as in quartzite they are of the almandine amphibolite facies. Many of these dikes are only slightly altered, and even altered dikes retain original ignous textures. We must emphasize that many or perhaps most of the mafic igneous rocks were emplaced at such a late date as to never have been exposed to severe metamorphism of any type or they were undeformed and therefore did not recrystallize.

POLYMETAMORPHISM

There is no unequivocal evidence of polymetamorphism in the minerals of the metasedimentary rocks. The pelitic rocks in the northeast and southwest extremity of outcrop show some paragenetic relationships that are suggestive, but these textures could be interpreted as evidence for progressive metamorphism.

Garnet in garnet-biotite schists located in sec. 2, T. 18 N., R. 79 W., show helicitic fabric (Fig. 50). In some layers of the schist small folds in the schist continue through the garnet porphyroblasts suggesting garnet formation after folding of the schistosity. In the other layers elongated quartz inclusions show an S-shape and the trend of inclusions does not parallel the schistosity (Fig. 50). The inclusions in the garnet are approximately the same size as the minerals of the matrix. This suggests that the garnets grew and were rotated during folding of the schistosity and continued to grow after folding or deformation ceased. This type of fabric is found in both schists of the northeast and southwest, but has not been recognized in the central area of metasedimentary rocks. This fabric may have developed during the proposed late stage of folding of these rocks or it may simply be evidence of deformation during progressive metamorphism.

In the southwest where pelitic schists contain porphyroblasts of biotite, garnet, staurolite, and kyanite, the staurolite and kyanite may be late.



A B

Figure 50—Photomicrograph showing folded trails of elongated quartz and opaque minerals in garnet-biotite schist southwest of Arlington, sec. 2, T. 18 N., R. 79 W. Quartz inclusions are about the same size as matrix. Trend of inclusions shows helicitic fabric at an angle to trend of schistosity shown in left. Garnet probably rotated during growth. Line 1 mm. long. A—Crossed nicols. B—Plain light.

For example, in biotite-staurolite schists, large porphyroblasts of biotite contain trains of quartz smaller in size than that of the matrix, whereas staurolite crystals contain inclusions about the same size as the matrix. The biotite porphyroblasts are crushed and rotated, but the staurolite crystals are oriented parallel to the schistosity. We suggest deformation of early-formed biotite schists with development of staurolite in a later metamorphic event at higher temperature. We cannot say, however, if these two events are greatly separated in time or if they result from continued deformation during progressive metamorphism.

METASOMATISM

Field evidence of metasomatism is common in the gneissic units. In the northwest (north of the Mullen Creek-Nash Fork shear zone and west of the metasedimentary rock), there is evidence of transformation of both biotite gneiss and hornblende gneiss to more massive quartzo-feldspathic gneiss. In addition, many small granitic bodies and pegmatites show replacement relationships with the gneissic host. Examples of field relationship are in Figures 16 and 33. The timing of the metasomatism and granitization in the northwest area is unsolved. In the late stage of deformation of the gneiss and the contained igneous bodies marked by the development of northwest zones of passive flow, there was some metasomatism and granitization since in these zones the gneiss is more granitic, and both pegmatite and migmatite are more common. Locally also mafic dikes are migmatized. However, metasomatism and granitization probably accompanied earlier periods of deformation and distinctions have yet to be made between the early and late events.

South of the Mullen Creek-Nash Fork shear zone field evidence of metasomatism and granitization is also common. This has been cited in the discussion of the quartzo-feldspathic gneiss and older granite. The last stage of deformation in the south, which caused the refolding of the gneissic unit and deformation of virtually all rock types except the Sherman Granite and perhaps some mafic bodies, was accompanied by granitization. This is demonstrated by the granitization of cores of antiforms in the gneiss and the common replacement locally of hornblende gneiss by quartzo-feldspathic gneiss. Most of the pegmatite and pegmatized areas probably developed in late stages of this event.

If one considers the problem in a regional perspective, there is a general increase of evidence for metasomatism toward the south. Immediately south of the shear zone there is evidence of deformation by cataclasis and shearing certainly accompanied by metamorphism and metasomatism, but the rocks do not seem completely recrystallized. Pegmatites are conformable and cross-cutting, but pegmatites interpreted as replacement in origin are rare. Near the Colorado border the rocks appear more thoroughly recrystallized and shear zones are completely healed. Most granitic bodies have vaguely defined contacts and field characteristics are suggestive of granitization. Along the State line and south into Colorado (Steven, 1957) replacement pegmatites become the common variety and all rocks appear completely recrystallized. The amphibolite dikes (?) in the southwest are good examples of complete recrystallization. These units show no trace of original texture and structure.

Perhaps the reason Steven (1957) and Swetnam (1962) have suggested a metasomatic origin for quartzo-feldspathic gneiss in the south is that the area is one where there was greater mobility at contacts between granitic and mafic rocks.

RETROGRADE METAMORPHISM

The return to low temperature-pressure conditions that must take place after periods of deformation and uplift is not accompanied by establishment of minerals stable at these T-P conditions, i.e. retrograde metamorphism. This is the general case in metamorphic rocks, and it is fortunate otherwise we would have no record of metamorphic events. Nevertheless, as is the case in most metamorphic terrains, there is here some evidence of retrograde metamorphism. For example, garnet, amphibole, and epidote of the almandine-amphibolite facies rocks may be partly chloritized, and kyanite and sillimanite may be altered to muscovite, in part. These changes are most common in fractured rocks and a control of alteration by microfractures is typical. However, many rocks in both older gneisses and the metasedimentary rocks show evidence of retrograde metamorphism without obvious fracture control. Perhaps there was retrograde metamorphism of the almandine-amphibolite facies rocks during one of the latter events of Precambrian, or we may confuse the effect of wet metasomatism with retrograde metamorphism, but most retrograde effects developed in areas subject to cataclasis.

The best examples of retrograde metamorphism are in the shear zones, and these have been reviewed earlier in the discussion of cataclastic rocks.

PRECAMBRIAN TECTONIC HISTORY

Two approaches can be used in interpreting the Precambrian tectonic history of this area. Both approaches represent the results of work in progress, and give only tentative answers. As a first approach, we can simply review details of the structure of the Medicine Bow Mountains and try to establish a chronology of events using dated rocks, where possible. As a second approach, we can place the mountain area in a regional perspective and interpret the geologic history as it might relate to the structural evolution of the Precambrian rocks of the central and northern Rocky Mountains.

We will first consider a sequence of structural events utilizing only that information derived from this study of the Medicine Bow Mountains. Field evidence indicates that the gneiss of the northwest slope and possibly the gneissic granite near Arlington is the Precambrian basement. This is substantiated by age determination that indicate a 2.4 b.y. age for the gneiss and a 2.2 b.y. minimum age for the gneissic granite near Arlington. The metasedimentary rocks that lie on the basement are older than 1.6 b.y. and younger than the basement, and are thus bracketed between 2.4 b.y. and 1.6 b.y. These metasedimentary rocks are deformed into well defined northeast trending folds and ill-defined northwest trending folds. Our supposition has been: (1) that these rocks were deformed; (2) gabbroic magma was introduced; and (3) that the rocks were deformed again, but we do not know if these events were greatly separated in time or if they were part of one period of deformation. We also know that the hinge of the major syncline developed in the Libby Creek Group has been refolded by flexure in the vicinity of the Mullen Creek-Nash Fork shear zone. It is clear that the metasedimentary rocks have had a complex history, but were never subject to deformation comparable to that affecting the basement rocks. They are generally of low metamorphic rank and not highly deformed except to the northeast and southwest where both degree of deformation and metamorphic rank increase to some degree.

This complicated structural history of the cover sedimentary rocks during Precambrian time implies an even more complex history for the basement for we must assume that whatever structural history the basement may have had prior to 2.4 b.y. has been further compounded during folding of the sedimentary cover. There is no agreement among geologists on basement behavior during folding of sedimentary cover. This point will be discussed further in the review of Laramide deformation, but the basement could deform in a passive manner (i.e. movement on closely spaced fractures) while the cover deformed by flexure; or faults in the basement might pass into folds in the sedimentary cover. If we examine the palinspastic map of the basement (Pl. 30) we see northeast trending flexures and shear zones that may have developed during the initial stages of folding of the cover, but unfortunately these structures appear to be later than northwest trending structural elements of the basement. There seems to be a reverse order of structural evolution of basement and cover. In other words the northwest trends are earlier than the northeast trends in the basement, but are later than the northeast trends of the cover. Another difficulty with this attempt to correlate structure is the evidence of more intense deformation of the basement including pegmatite formation during the development of northwest structure while the cover sedimentary rocks are broadly folded and metamorphosed but not severely deformed. In brief, we do not have a straight forward correlation between structure in the basement and in the sedimentary cover.

Undoubtedly the basement was deformed during folding of the cover, but earlier basement structures may have been oriented in such a manner that they show little evidence of later deformation. In essence, the orientation of the basement structure may have controlled, to some degree, the pattern of folding in the cover. If this is true the north trending folds of the basement and later northwest basement structure may be older than the fold systems of the cover. This implies an extraordinarily complex structural history in the area north of the shear zone.

South of the shear zone the regional structural trend is more like that of the northern Colorado Front Range (Tweto, 1966, p. 304) than that of the northern Medicine Bow Mountains, and lithologies also differ with a greater abundance of hornblende gneiss and associated paragneiss. There is nothing to tie the rocks on either side of the fault together with the exception of recurring scraps of evidence that an early north to north-northwest structural trend may have existed in the area south of the fault. The evidence for this has been cited before and includes the geometry of folds in domain three (S), local northwest-trending structures sandwiched between belts of northeast trending structure, and early northwest-trending foliation in mafic intrusive bodies. Virtually every rock type in the southern area is metamorphosed after or during emplacement. Certainly these rocks are more highly deformed than those to the north, and the structure prior to the last major episode of deformation may have been similar to that of the north. This orogeny (?) that affected the rocks south of the fault may have caused the development of northeast-to-east-trending folds and faults in the north as well but the severe deformation was mainly south of the fault. Refolding of the hinge of the French Creek fold may have taken place at this time and the Mullen Creek-Nash Fork shear zone was probably reactivated. Indeed, this may have been the time in which the major movement in the fault occurred.

Additional support for a major episode of folding and metamorphism south of the fault comes from age determination that indicate that gneiss, granite, and pegmatite (Hills, et. al., 1968) retain no evidence of an early age. This suggests that all rocks studied are either younger or a late orogeny has wiped out evidence of the early event of 2.4 b.y.

Certain igneous intrusives and pegmatite both north and south of the fault were either formed in late stages of this event or are post-kinematic. These include the Bennett Peak gabbro north of the fault and the definitely post-kinematic Sherman Granite south of the fault, as well as many mafic and felsic dikes that either occupy or parallel the major northeast shear zone.

If we consider the structure of the Medicine Bow Mountains from a regional perspective we can relate the gneiss northwest of the fault to the rocks of the Superior Province of Wyoming. The rocks of the Superior Province of Wyoming have been referred to as the Wyoming Province by Engel (1963) and include units that have not been severely enough deformed or thermally affected since about 2.5 b.y. to eliminate evidence of very old geologic events as determined by dating minerals. The age of these rocks approximates that of the Superior Province of the Canadian shield hence the tentative name correlation. This area of ancient rocks has been most recently outlined by Goldich, et. al. (1966, Fig. 1) and it extends through most of the Wyoming Precambrian. It also includes rocks to the northwest in the Little Belt Mountains of Montana and east to the Black Hills of South Dakota. The westernmost outcrops of these rocks are in the Teton Range of Wyoming and the southernmost in the Medicine Bow Mountains. Unfortunately, within this area exposures of Precambrian rock are only in the cores of Laramide uplifts and these exposures make up about one tenth of the total area. Because of lack of outcrop we may never be able to adequately interpret the geology of the Province, but in the last 10 years enough of the geology of this 10 percent has been mapped so that speculation on regional geology is possible. Since there is no general review of the geology of this area a brief summation has been prepared.

Perhaps the most detailed work in the Wyoming Province has been done by Arie Poldervaart and his students in the Beartooth Mountains of Wyoming and Montana (Eckelmann and Poldervaart, 1957; Spencer, 1958, Harris, 1959, Casella, 1964, Prinz, 1964, Butler, 1966, and Larsen et. al. 1966). The rocks of the area are largely granitized and metamorphosed quartzo-feldspathic gneisses with rare interlayers of metasediment (quartzite), para-amphibolite, and iron-silicate rock) that have been invaded at least four times by basaltic magma. According to Prinz (1964, p. 1217-1248) there were three periods of basaltic intrusion in the Precambrian and one period in the Tertiary. The early mafic intrusions are pre-granitization and are deformed and altered to amphibolite. This is followed by emplacement of metadolerites near the end of granitization, and finally emplacement of a "late" Precambrian dolerite that is not highly metamorphosed. The major structural pattern in these rocks is a series of south-plunging folds, some broad open folds, and others more tightly appressed and marked by passive flow. Larsen (1966, p. 1289) believes that this fold system developed during granitization and later than metamorphism. This pattern of folds and complex sequence of dike emplacement resembles in some degree history of the early gneisses of the Medicine Bow Mountains.

In the northern Bighorn Mountains Osterwald (1955, 1959) has mapped a series of granitic rocks that he believes formed by replacement of older gneiss. The older gneiss contains interlayers of amphibolite, mafic schist, and rock types that may be tuff or agglomerate. The gneiss has a northeast striking fabric and the granitic rocks are in a northwest plunging synform. The granitic rocks and gneisses are cut by diabase dikes some of which are deformed and invaded by the granite suggesting remobilization of the granite in some late event.

In the east central Bighorn Mountains Hoppin (1961) has mapped an area largely underlain by quartzo-feldspathic gneiss where early foliation and layering in the gneiss is generally masked by "shear" fractures that are high angle and strike dominantly N 10° E. Mafic dikes cut the gneiss and also strike N 10° E. Palmquist (1967, p. 283-298) states that the Horn area of the southern Bighorn Mountains is comprised largely of granitic gneisses containing conformable bodies of amphibolite, biotite schist, calc-silicate rock, marble, quartzite, banded quartz-hematite schist and garnetiferous rock. These metamorphic rocks form a complex antiform plunging to the north-northwest. The gneisses are cut by at least, one dike of orthoamphibolite.

Southwest of the Horn area in the eastern Owl Creek Mountains[1] of Wyoming Millgate (1966, Personal communication) and Gliozzi (1967) have mapped a sequence of metasedimentary rocks containing iron formation, quartzite, and various pelitic schists and gneisses. The metasedimentary rocks are folded into a northeast trending syncline, and it is possible that a fold set with steep south plunging axes has been superimposed. The gneiss and metasedimentary rocks are cut by diabase dikes that are, in part, altered to amphibolite, but not severely deformed. The entire sequence is cut by a relatively unmetamorphosed granite and associated pegmatite. West of the eastern Owl Creek Mountains Precambrian rocks are exposed in Wind River Canyon where the major rocks type is interlayered felsic schists and amphibolite similar to that of the eastern Owl Creek Mountains (Engel, 1947; Condie, 1967). These layered rocks are cut by a granitic intrusion and associated pegmatite that is probably related to that of the eastern Owl Creek Mountains.

Little is known about the isolated bodies of Precambrian rock exposed in the area northwest of the Wind River Canyon. Long (1959) states that the major rock types of the Phlox Mountain area are granite gneiss and biotite schist. The foliation of these rocks strikes northwest and dips steeply east. The gneiss is cut by numerous diabase dikes that strike roughly east southeast.

The westernmost large exposure of Precambrian rocks in Wyoming is in the core of the Teton Range. Reed (1963, p. C1-C6) reports that the oldest Precambrian rocks are completely folded interlayered metasedimentary rocks including amphibole gneiss, biotite gneiss, amphibolite, sillimanite schist, quartzite, and marble. These metasedimentary rocks have associated granite gneisses that Reed believes were formed by granitization of the metasedimentary sequence during deformation. Following a late episode of folding, quartz monzonite and peg-

1. The eastern Owl Creek Mountains include a large block of Precambrian rocks east of Wind River Canyon.

matite were introduced and finally diabase dikes that have chilled borders and are unmetamorphosed. The major structural trend of foliation in the area is northeast, but Reed emphasizes that the actual fold patterns have not been worked out.

Turning to the Wind River Range we find that the greater part of this major uplift is unmapped. The majority of the rocks in the central and northern part of the mountain are probably gneisses of granitic composition. Richmond (1945) and Baker (1946) have studied a part of the northwest margin of the Wind River Range which they describe as primarily gneiss and gray, porphyritic granite. Baker (1946, p. 567) states that the gneiss contains rare layers of amphibole schist and green quartzite and that it is cut by pegmatite. These units are cut by basalt dikes that strike northwest and east. In the north Wind River Range near Downs Mountain Worl (personal communication, 1967) has mapped an area consisting largely of quartzo-feldspathic gneiss and that contains some partly granitized remnants of iron formation. Oftedahl (1953) made a traverse across the central Wind River Range and reports that the rock types are a gneiss-granite complex with areas of migmatite and agmatite. This gneiss- granite complex is cut by possibly two generations of older mafic intrusions and irregular dikes that have been partly granitized. There are scattered "late" Precambrian dikes that are not metamorphosed. Oftedahl states that there is no regionally consistent structure in these rocks. Parker (1962) states that an area in the central Wind River Range near Grave Lake is composed of various types of felsic gneisses and agmatite. These units are intruded by equigranular granodiorite and porphyritic adamellite. The entire complex is cut by northeast trending Late (?) Precambrian diabase dikes. Perry (1967) mapped an area in the east central Wind River Range in the Paradise Basin area. The major rock types are quartzofeldspathic gneisses of probable sedimentary and volcanic origin. Granite is in the gneissic units but is folded with the gneiss. In the southern part of the area the gneisses are recumbently folded about a gently plunging north-northeast axis, but in the north these rocks are in turn folded about steeply plunging axes that trend generally north.

The geology of the southern Wind River Range is better known than that of the central and northern part because of recent mapping by Bayley (1963, 1965), Hodge and Worl (1965). In the Atlantic City area Bayley describes an older sequence of metasedimentary rocks including pelitic schists, iron formation, and various types of metavolcanic rocks which he designated the Goldman Meadows Formation. This is overlain by the Miners Delight Formation which consists of mafic lava flows, graphite schist, ellipsoidal meta-andesite and metagraywacke. This metasedimentary sequence is cut by serpentinite metadacite, and meta gabbro dikes. These igneous rocks are deformed and schistose. The entire sequence is intruded by the Louis Lake Batholith, a large body of quartz diorite and granodorite which is massive in the core but strongly foliated along its border. The last igneous rocks emplaced are diabasic gabbro dikes. The metasedimentary rocks are exposed in a major northeast trending syncline. The Louis Lake Batholith also shows a strong northeast foliation. West of Bayley's mapping Worl (1963) and Hodge (1963) have recognized a series of pelitic and quartzo-feldspathic metasedimentary rocks that have been invaded by granite and associated pegmatite. This entire complex is cut by unmetamorphosed diabase dikes. These units are probably a western extension of the metasedimentary sequence mapped by Bayley, but iron formation has not been recognized. Worl (1963) suggests a complex tectonic history for the metasedimentary rocks with formation of early folds with a general northeast trend and late folds that are in part passive that trend northwest.

East of the Wind River Range isolated masses of Precambrian rock make up the Granite Mountains. These rocks are little known, but the westermost Precambrian outcrops are reported to be composed of hornblende gneiss and metasedimentary rocks. These rocks have a dominant northeast structural trend as observed on air photographs. A large area of Precambrian rock exposures in the northern part of this area has been mapped in reconnaissance by Carey (1959) who reports the Precambrian units are mainly hornblende gneiss and quartzofeldspathic gneiss. The dominant structural trend is northwest in these units. Other masses of Precambrian rock that have been seen by the writer are largely granite gneiss with rare interlayers of metasediment. These units are cut by diabase dikes most of which trend northeast.

Mapping in the Seminoe Mountains located southeast of the Granite Mountains, by Bishop (1963) and Bayley (1965) indicate that this area contains a sequence of metasedimentary rocks including iron formation, various types of hornblende gneiss and schist, and quartzite. These units are cut by mafic igneous rocks that are also metamorphosed and all rocks are invaded by granite. The granite is well foliated but only the marginal portion has been studied. There may be a late mafic dike series in this area because dikes cut the granite body. However, the granite-dike relationship is like that described by Osterwald in the Bighorn Mountains where the dikes are also invaded by granite. Perhaps there are mafic dikes of two ages in the granite. The general structural trend of the metasedimentary rocks is northeast, but here again there is a suggestion of a late steeply plunging northwest axis.

The northern Laramie Mountains is less well known geologically than any area of comparable extent in Wyoming. Beckwith (1939) mapped a small area in the northernmost segment near Casper, Wyoming, where the major rock types are granite gneisses containing remnants of hornblende schist, amphibolite and quartzite. The gneiss also contains schistose bodies of serpentinite that were originally ulta-mafic intrusions. These units are cut by amphibolized and somewhat schistose dikes of metadiabase which are, in turn, cut by granite pegmatite. The age relationship of the granite gneiss is uncertain; Beckwith thought it was the youngest Precambrian rock and that the gneissic structure was inherited from the schists. Possibly this "granite gneiss" has also had a history of remobilization. The general structural pattern of these rocks is northeast. In the northeastern Laramie Mountains small areas were mapped by Spencer (1916), Osterwald (1946), Albanese (1949) and Patterson (1950). According to these students of Laramie Mountain geology the older rocks are various types of hornblende schists and gneisses, with lesser biotite schist and rare layers of magnetite schist and graphite schist. The most abundant rocks are granite gneisses that are believed to have been derived from hornblende gneiss by metasomatism. These gneisses grade into more massive granite locally. One small body of pyroxenite was mapped by Albanese (1949). Granite pegmatite is common, and Osterwald (1946) reports a late body of soda granite in the Owen Creek area. Spencer (1916) who studied an area near Esterbrook is the only one of these observers who mentions mafic dikes. He states that the dikes are younger than the granite or granite gneiss, but he obviously had some difficulty with dike-granite relationships because he considers the possibility of two ages of dikes. The general structural trend in three of these areas is northeast, but the area mapped by Patterson on the eastern margin of the Range has a northwest structural pattern.

In the central Laramie Mountains north of Wyoming State Highway 87, recent mapping by Fields (1963), Hodge (1966), Toogood (1967), Bothner (1967), and Smith (1967) of small but connected areas give some information about a belt of rocks extending east-west across the Laramie Mountains. These rocks are generally north of the anorthosite complex of the south central Laramie Mountains and are north of the Mullen Creek-Nash Fork shear zone if it is extended to the Laramie Mountains proper. They may be part of the Wyoming Province, but this is uncertain because none of the rocks have been dated. The older rocks are probably mostly metasedimentary and perhaps metavolcanic rocks. These include hornblende gneiss and schist, calc-schist, quartzite, marble, conglomerate, pelitic schists (with sillimanite, kyanite, and andalusite), and biotite schist. Various types of quartzo-feldspathic gneisses are interlayered with these units, grade into the units and are commonly structurally conformable. Some of these more felsic units do show local cross-cutting relationships to the metasedimentary rocks and are thus later. The origin of these felsic gneisses is not known but as in other areas they may be cut by metadiabase dikes which are also deformed. In the southern part of this area the metasedimentary units and gneisses are cut by anorthosite and related rocks and these rock types are intruded by Sherman Granite. The structure of these units is extremely complex suggesting several episodes of deformation and metamorphism. Early fold systems may have had north to northwest trends and may have been, in part, recumbent. Later fold patterns trend northeast. Along the eastern border of the Laramie Mountains the isolated mass of Precambrian rocks of the Cooney Hills has a pronounced northwest structural pattern.

Rocks of Precambrian age are also exposed in the Hartville area of eastern Wyoming that may be within the Wyoming Province. The older gneisses of this area are largely hornblende gneiss and granite gneiss. Millgate's (1965) map of the area shows a broad north plunging antiform which appears to be a product of refolding since mineral lineations show a systematic change in attitude around the axis of the antiform.

The easternmost exposures of Precambrian rock of the Wyoming Province are in the Black Hills of South Dakota (Darton and Page, 1925). This area may have an older basement overlain by metasediments including graywacke, quartzite, quartz schist, slate and various higher rank pelitic schists and gneisses, calc-schist, metadolomite, and iron formation. The metasedimentary rocks are cut by amphibolite, foliated granite, and pegmatite. The major granite bodies are in the south and are pegmatitic cross-cutting granites superficially resembling the Sherman Granite of southern Wyoming. The general structural trend of the metasedimentary rocks is north-northwest. Detailed mapping of metasedimentary rocks in the Lead area by (Noble, et. al., 1949) shows that the metasedimentary rocks of the area have been isoclinally folded to produce north-northwest fold axes and later deformation primarily passive in nature has developed cross-folds that have axes that trend northwest. Gabbroic magma was introduced after the early phase of folding, but prior to the late stage of folding.

In summary several generalizations can be made about the geology of the Precambrian rocks of the Wyoming Province.

(1.) In most areas the oldest rocks are metasediments that have been severely deformed, metamorphosed, and probably partly granitized. Because of granitization and metamorphism only remnants of the original sedimentary sequence remain. The most abundant sedimentary rocks may have been graywacke, metavolcanic rocks, and shales or graywacke that are now primarily hornblende gneiss and schist. Quartzite, iron formation, and various calcareous metasedimentary rocks are also present. Foliated granite, quartz diorite and different granite gneisses associated with the metasedimentary rocks may be deformed igneous rocks, metamorphosed volcanic rocks, or products of granitization.

(2.) There may be a younger series of metasedimentary rocks that includes iron formation, metavolcanic rocks, and graywacke in three areas; South Wind River Range, Seminoe Mountains and Owl Creek Mountains, all in Wyoming. These metasedimentary rocks are in the northeast trending synclines and are not as extensively granitized as other metasedimentary units.

(3.) The metasedimentary rocks and gneisses contain numerous bodies of mafic igneous rocks that have been metamorphosed and largely converted to amphibolite. These units are of various ages, but usually post-date some major episode of folding.

(4.) Mafic dikes and sills are common. These dikes and sills are of more than one age—some are highly deformed and metamorphosed—others are relatively fresh and have chilled borders. Some dikes are veined by enclosing granite, suggesting remobilization of the granite mass. In general, one set of less metamorphosed (Late (?) Precambrian) dikes is present in the northern and central part of the Wyoming Province but to the south and east there are no examples of completely unmetamorphosed dikes.

(5.) There is no consistent structural trend. Fold axes trend north, north-northwest, northwest, northeast, and east. Evidence for refolding is good in some areas and strongly suggested in virtually every area mapped. In several localities early structural patterns are northeast and late structural patterns are northwest, but this is not universally true. There is no well-defined regional structural trend that could be related to a major orogenic event. Such a pattern could be present and not outlined by mapping to date or it may simply be impossible to establish with less than 10 percent outcrop.

Age determinations indicate that the rocks of the Wyoming Province have not been completely reconstituted since about 2.5 b.y., but dates determined so far are difficult to relate to any regional deformation and they are quite variable ranging from 3.1 b.y. to around 1.6 b.y. The 3.1 b.y. date is for the formation of detrital zircon in the Beartooth Mountains (Catanzaro and Kulp, 1964, Catanzaro, 1966), and suggests some metasedimentary rocks are very old indeed. If those areas where metasedimentary rocks are most highly deformed and granitized are considered older terrain we find that Beartooth pegmatites are dated as 2.7 b.y. (Gast el. al., 1958), Bighorn augen gneiss from Osterwald's area as 2.5 b.y. (Gast et. al., 1958). Teton Range pegmatite 2.6 b.y. (Giletti and Gast, 1961), and northern Laramie Mountains biotite gneiss 2.3 b.y. (Giletti and Gast, 1961). If the seemingly less metamorphosed iron formation-graywacke suite is considered younger than these units we find that granite and pegmatite that cut these metasedimentary rocks in the Owl Creek Mountains is dated as 2.3 b.y. (Aldrich et. al., 1958) or 2.6 b.y. (Giletti and Gast, 1961) but that biotite from this granite is dated as 1.3 b.y. (K/Ar) and 1.9 b.y. (Rb/Sr), (Aldrich et. al., 1958). Granite that cuts these metasedimentary rocks in the southern Wind River Range is dated as 2.2 b.y. (Giletti and Gast, 1961) and 2.2, 3.0, and 3.3 by lead alpha dates on zircon from the granite (Cannon et. al., 1966) and 2.6 and 2.2 b.y. for hornblende and biotite respectively from the granite by K/Ar (Cannon et. al., 1966). It is obvious from this review that there is no clear separation of the assumed older and younger terrain by age determination. Perhaps all of these units have been affected by some geologic event in the general 2.5 b.y. time period that caused reconstitution of the rocks but failed to eliminate geologic evidence of a complex earlier history. In any event, I see no simple relationship between the enormously complex geologic history suggested for this area and age determinations. The age determinations are of great value, however, in telling us that these rocks are among the oldest in North America.

Another interesting aspect of the geochronology is the occasional young date recorded in minerals from this older terrain. For example, 1.3 b.y. (K/Ar) and 1.9 (Rb/Sr) on Owl Creek granite (Aldrich et. al., 1958), 1.57 (K/Ar) on K-feldspar and 1.8 (K/Ar) on biotite from pegmatite of the northern Wind River Range (Basset and Giletti, 1963), 1.7 and 1.8 b.y. (K/Ar) on biotite from metamorphic rocks of the Northern Wind River Range (Bassett and Giletti, 1963) and 1.6 b.y. (K/Ar) on muscovite and 1.6 b.y. (Rb/Sr) on muscovite from pegmatite of the Black Hills (Aldrich et. al., 1958). These dates must reflect some aspect of the later thermal history of this Province, but there is no obvious relationship to an orogenic event.

Certainly the Wyoming Province has had an extraordinarily complex history. Unfortunately this history cannot be reduced to an ordered sequence of events that can be related to the history of the older gneisses of the Medicine Bow Mountains. The sequence of events suggested for the gneiss of the Medicine Bow Mountains is similar to that of the Beartooth and perhaps northern Bighorn Mountains, but elsewhere there is no correlation or the suggested history may be the reverse of that of the Medicine Bow Mountains. The thermal history is like that of other parts of the Wyoming Province with a major event of around 2.5 b.y. and a second event of around 1.6 b.y., but here again this cannot be correlated with specific geologic events within the Wyoming Province.

Part of the puzzle of the older rocks may be understood by considering the marginal, less highly deformed and less metamorphosed metasedimentary rocks along the southeast margin of the Wyoming Province. In three areas, all lying immediately north of the age and structural disconformity marked, in part, by the Mullen Creek-Nash Fork shear zone, metasedimentary rocks are present. These areas are the Medicine Bow Mountains, Sierra Madre, and Hartville Uplift. The rocks are shelf-type sedimentary rocks for the most part, including thick quartzite, "algal" limestone, tillite (?), and iron formation. Lithologically they are more like some part of the Animikie series (Huronian) of the Lake Superior Province of the United States than any of the other units within the Wyoming Province. The age determinations on these units in the Medicine Bow Mountains also support this generalization. In the Hartville area Ebbett (1956) and Millgate (1965) describe an older pelitic schist overlain by metadolomite, iron formation, and dark phyllite. In the Sierra Madre Spencer (1904) and Ebbett (1967, personal communication) describe from oldest, conglomerate, metabasalt and graywacke, "slaty" phyllite, quartzite, phyllite and limestone, conglomerate, quartzose limestone and phyllite, quartzite, pebbly quartzite, and phyllite. The metasedimentary rocks of the Sierra Madre are more like the older metasedimentary series of the Medicine Bow Mountains (Deep Lake Formation) and these of the Hartville are more like the younger metasediments (Libby Creek Group), but we cannot correlate these units by use of any definitive marker or other precise criteria.

If these rocks are Middle Precambrian why are they preserved only along the margin of the Wyoming Province? One possibility is that they were more highly deformed and infolded along the margin of a fold belt located to the south, and have been stripped off the stable, less deformed Wyoming Province to the northwest. Ideally this would suggest a common regional structural trend for these units and the rocks of the Colorado Front Range, but this is not entirely true. The northeast structural trend of the Medicine Bow sedimentary sequence fits fairly well with this hypothesis, but the structural trend of the metasedimentary rocks of the Sierra Madre is west-northwest and the Hartville uplift probably changes from east to north.

Before we consider these structural problems it is necessary to review briefly the general geology of the Precambrian of Minnesota and Michigan and the Colorado Front Range. Goldich et. al. (1961) have reviewed the Precambrian geology and geochronology of Minnesota. In Minnesota the general geology is quite similar to that of Wyoming. In fact, it is reasonable to extrapolate southwest from Minnesota to Wyoming and relate the structural evolution of the two areas. In northern Minnesota the older Precambrian is a complex of metasediments, gneisses, and intrusive rocks that broadly resemble the rocks of the Wyoming Province. It is quite possible, for example, that one of the Early Precambrian iron formations (Soudan) of this region correlates with the iron formation of the Wyoming Province. According to Goldich these older rocks were deformed during the Algoman orogeny, that terminated around 2.5 b.y. ago. Sedimentary rocks were deposited on the basement between 2.5 b.y. and 1.7 b.y. ago. The sedimentary rocks, Huronian or Animikie, were deposited as a miogeosynclinal facies along the border of the ancient shield, but to the southeast the sediments are eugeosynclinal. These metasedimentary rocks were deformed during the Penokean orogeny with very mild deformation in the northwest, but with increasing metamorphic rank and intensity of folding to the southeast. The Penokean orogeny terminated at about 1.7 b.y. This general history is similar to that of the Medicine Bow Mountains, at least, through an orogeny terminating at around 1.7 b.y. The general northeast trend of the Penokean fold belt is also compatible with the northeast trend of the metasedimentary rocks of the Medicine Bow Mountains.

If we wish to extrapolate the Penokean fold belt to southeast Wyoming it must also extend to Colorado, and the Front Range should be within the heart of the deformed region. Recent reviews of Front Range geology and geochronology suggest that this could be true. Hedge et. al. (1967) postulate a period of regional metamorphism in the northern Front Range at 1.7 b.y. The rocks of the northern Front Range are like those of the southern Medicine Bow Mountains except that sillimanite-biotite gneisses are more common. The sillimanite-biotite gneiss and various quartzo-feldspathic gneisses are interlayered with hornblende gneiss and amphibolite. The amphibolite is conformable and Lovering and Goddard (1950) have considered them to be flows and sills. Braddock (1966) has recognized graywacke with well-developed graded bedding in the Big Thompson Canyon area where these rocks are somewhat lower rank than in most parts of the Front Range. These rocks could be a more highly metamorphosed eugeosynclinal facies of the metasedimentary rocks of the Medicine Bow Mountains or more likely they may be in part older metasediments and in part equivalent in age to these units. They have been deformed several times Moench et. al. (1962), Harrison and Wells (1959) and Braddock (1966) but the regional structural trend is east-northeast.

This broad concept of an older basement with a complex history in Wyoming bordering a fold belt mainly developed in Colorado is an appealing one, and fits the regional geologic framework, but certain problems in detail suggest that this is an over simplification. The main problem is with the "miogeosynclinal" sedimentary rocks in the Sierra Madre, Medicine Bow, and Hartville areas. These units are not continuous along the margin of the proposed fold belt nor do they have the same general structural trend. There is evidence that they have been deformed more than once in the Medicine Bow Mountains and possibly the Hartville area. There is a possibility that these units have been disrupted and folded about new axes after the Penokean orogeny.

Goldich's (1966, p. 5401-5402) concept of a Black Hills orogeny at approximately 1.7 b.y. must have some relation to this problem. In the Black Hills there may be an older basement overlain by deformed Middle Precambrian sedimentary rocks (Goldich et. al., 1966, p. 5400-5402). The Middle Precambrian metasediments may include the rocks mapped by Noble et. al. (1949) in the Lead area. The structural trend of these rocks is north-northwest however in contrast to the east-northeast to northeast structural trend along the Wyoming-Colorado border. Goldich et. al. (1966) have proposed a general northwest fold belt in the area perhaps extending into Canada to the north (Wilson and Brisbin, 1962) where it would become part of the Churchill province of the Canadian shield. If this is a true orogenic belt the Wyoming Province is bounded on the east by a northwest to north-northwest trending fold belt and on the south and east by an east-northeast to northeast trending belt of about the same age. This would isolate the Wyoming Province from the main Superior Province of the Canadian shield and would suggest a coalescence of north-northwest and northeast structure about in the Hartville area of Wyoming. Interestingly the trend of fold axes in Middle (?) Precambrian metasedimentary rocks of this area changes from east in the west to north as the Black Hills are approached.

This possible coalescence of two major fold belts in southeast Wyoming may explain the complexity of structure, the superposition of northwest trends on northeast in some areas, and the disrupted nature of the units along the margin of the fold belt. It is also possible that the fold belt was disrupted in some post-Penokean event as late as the Laramide orogeny.

PROBLEMS

In the above review we have actually done more to outline problems that must be considered before an interpretation of the Precambrian tectonic history of the Medicine Bow can be made than to evolve a satisfactory interpretation. The major problems are: (1) How are northwest structures in gneiss of the northwest slope related to late northwest folds of the metasedimentary rocks. If these structures developed simultaneously the tectonic history of the metasedimentary rocks had to be very complex before an east-northeast fold belt developed in Colorado. This would seem to rule out the interpretation that considers the metasedimentary series as a miogeosynclinal facies of rocks in a Colorado fold belt. (2) If we interpret the northwest structures of the gneiss as early we must find some means to disrupt the miogeosynclinal rocks north of the Mullen Creek-Nash Fork shear zone after development of the fold system in Colorado. There is ample time for this to happen about 1.7 b.y. The major question is whether it could have happened in the Laramide. Perhaps some of the late northwest structures of the Wyoming Province are actually Laramide, but since the metasedimentary rocks of the Medicine Bows seem to have been raised to almandine amphibolite rank when the northwest structures developed a Laramide origin is unlikely. We must then appeal to an event in the Precambrian at some time between 1.7 b.y. and 0.6 b.y. The only event that the writer can relate to this is folding of the Late Precambrian Belt series that took place in Idaho prior to 1.2 b.y. (Reid, et. al., 1967, Hobbs et. al., 1965). This period of folding developed northwest folds, but whether the effects of this event could have been felt in the Wyoming Province is questionable. (3) What is the relationship between Goldich's Black Hills orogeny and the deformation of rocks in the Colorado Front Range where structural trends are almost 90 degrees from those of the Black Hills?

Perhaps the answers to these questions will be found with detailed mapping of the Medicine Bow Mountains or elsewhere in the Precambrian of the Rocky Mountains, or as stated by Gilluly (1963, p. 139-140) in his review of the Rocky Mountain Precambrian "it may be fruitless to try to reconstruct regional chains when none can be confidently traced across a single state. Later structures obscure our reading of the older."

RELATIONSHIP OF PRECAMBRIAN AND LARAMIDE STRUCTURE

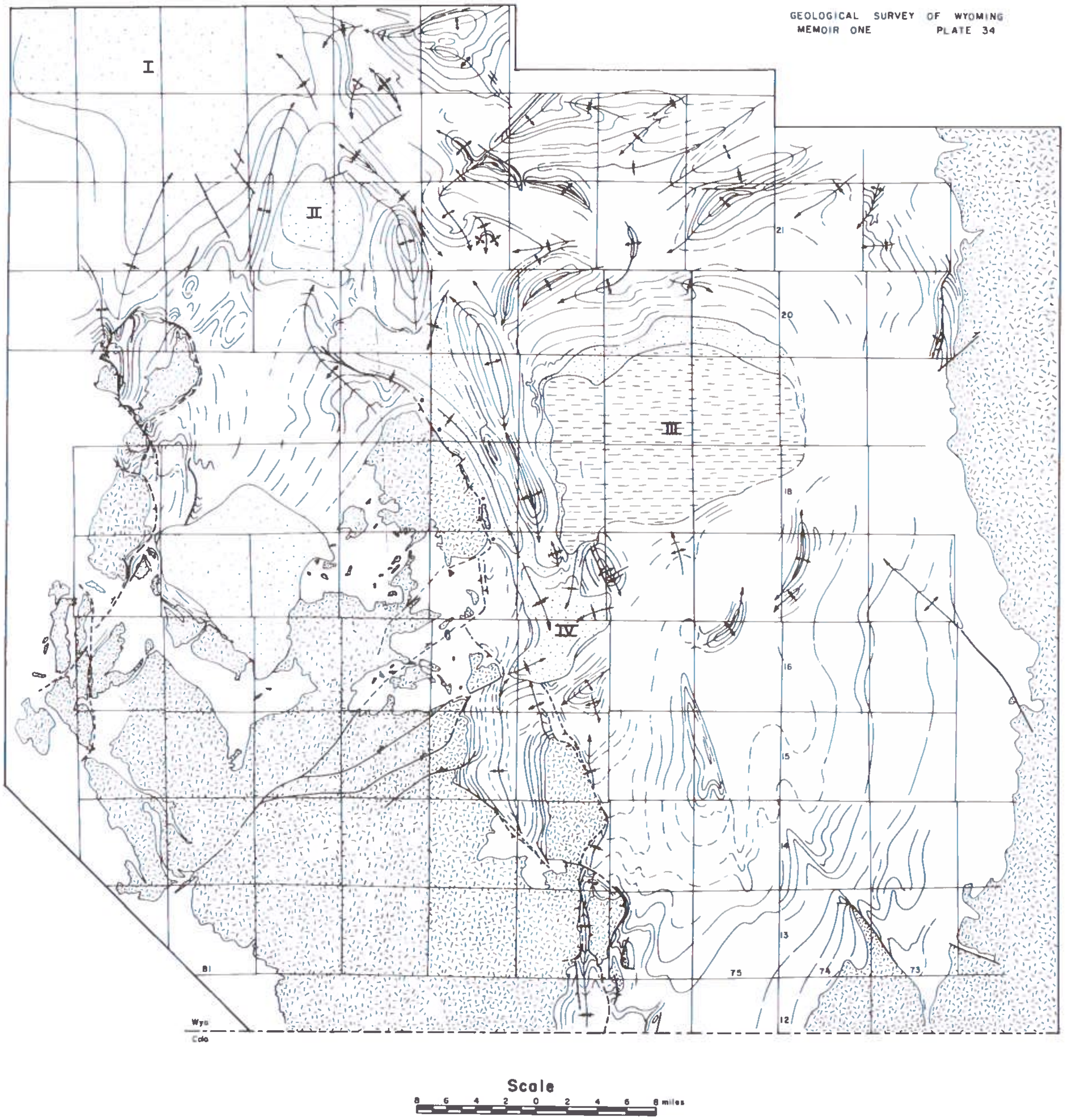
The Laramide structure has been covered in considerable detail by previous workers (Beckwith, 1938, 1942, 1941; Knight, 1953; Blackstone, 1953, 1965). The role of the basement (the Precambrian) has never been properly evaluated in consideration of Laramide structure because of lack of data therefore this aspect of Laramide structure will be reviewed here.

To understand the role of basement rocks in the development of complex structures in the Paleozoic and Mesozoic rocks along the flanks of the mountains; one must first review basic stratigraphic and structural data from field studies spanning the last 30 years. Laramide deformation began in this area in Latest Cretaceous time, and uplift was sufficient to strip Paleozoic and Mesozoic rocks from the mountains and expose rocks of Precambrian age. The maximum thickness of sedimentary rocks overlying the Precambrian surface at the beginning of the uplift could not have exceeded 14,000 feet even with the most optimistic estimates of thickness. If one uses high thermal gradients of 30 to 50 degrees C./km. (Tuttle and Bowen, 1958, p. 123), the temperature of the rocks near the base of the sedimentary section would be no more than 150° C. The load pressure on the basement from this thickness of sedimentary rocks would probably be less than 1,000 bars (Fyfe, Turner, and Verhoogan, 1958, p. 34). The effect of this sedimentary cover on the Precambrian basement would be relatively small, in fact we are probably at a depth where directed stress would exceed load stress on the basement rocks. We might also anticipate that rocks at this level in the crust subject to stress would yield by fracture, and that relief from stress would be gained by upward movement. It is also reasonable to assume that movement might be on preexisting surfaces oriented in such a manner that they could respond to directed stress. In the next paragraphs we will review the field evidence to see if there is support for such a view.

Initial uplift during the Laramide Orogeny may have begun in Late Cretaceous since rocks of Late Cretaceous age are thinner in the Medicine Bow Mountain area than elsewhere in southeastern Wyoming (Haun and Kent, 1965, fig. 21, p. 1795). Uplift sufficient to expose the basement did not take place until latest Cretaceous during the time of deposition of the Medicine Bow Formation. Conglomerate has been found in this formation in only one area in the immediate vicinity of the mountain along the south border of the Mill Creek Syncline (Pl. 1).

This formation is also exposed along the northern and northwestern margins of the Cooper Basin, the northeast margin of the Carbon Basin, and around the entire margin of the Hanna Basin except in the northeast (Pl. 34, Love and Weitz, 1955). The Medicine Bow Formation does not contain conglomerate in any of these areas except in the Mill Creek syncline. As suggested by Knight (1953) this conglomerate may be the first direct evidence of the Laramide orogeny of this area. We must emphasize, however, that the conglomerate is in the upper part of the Medicine Bow Formation of the Mill Creek syncline, and we are not certain of the age of this part of the formation. None-the-less this is the earliest known conglomerate of the Medicine Bow area and it is confined to the Mill Creek syncline. Since the axis of the Mill Creek syncline strikes northeast, it is possible that the early basin had a northeastern configuration paralleling strong northeast trending Precambrian structures of the Medicine Bow Mountains. The present form of the Medicine Bow Mountains, however, is that of a north-trending anticline, and most of the flanking structures in sedimentary rocks also trend north or northwest (Pl. 34). There are many folds and faults with northeast orientations in the basin east and northeast of the Medicine Bow Mountains (Pl. 34) and some of these structures appear to be deflected by north trending structures and some are clearly disrupted by thrust faults with north to northwest trends. We cannot eliminate the possibility that some or many of the northeast-trending structures developed simultaneously with north or northwest structures, but some may have been early and may correlate with northeast structures of the basement.

In Paleocene and early Eocene during deposition of the Ferris-Hanna (?) Formations, the Medicine Bow Mountain area was uplifted and structures developed that were like those present



TECTONIC MAP SHOWING LARAMIDE STRUCTURE
NORTH AND EAST OF THE MEDICINE BOW MOUNTAINS

Lines show trend of strike of sedimentary bedding and contacts between major units as determined from surface outcrop and air photographs.
Precambrian rocks shown with short dashes, Paleozoic-Mesozoic unpatterned, Ferris-Hanna Formations with dots, Wind River Formation, horizontal rule. I = Hanna Basin, II = Carbon Basin, III = Cooper Basin, IV = Mill Creek syncline.

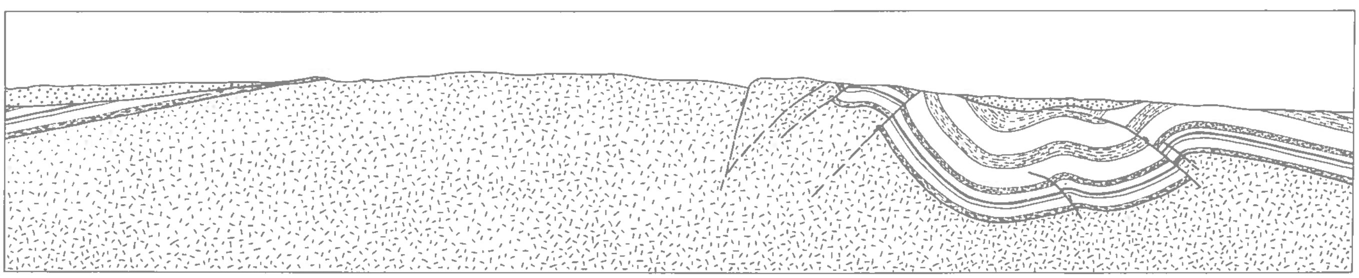


Fig. 51—Idealized Laramide Uplift

today. Uplift was accompanied by thrusting and the great majority of structures especially in the area north of the Mullen Creek-Nash Fork shear zone show uplift and thrusting on the east side (Pl. 34). The structures show certain basic characteristics, as cited below.

(1.) Blocks of Precambrian basement are uplifted, and displaced to the northeast on west dipping fault planes. Beds dip steeply on the thrust (northeast) side of the block and gently on back side of the block (southwest) (Fig. 51, Pl. 5).
(2.) Some blocks are also thrust to the southeast, but most thrust blocks terminate on the southeast in a northeast-trending transverse fault. Some of these transverse faults break out of Precambian shear zones (Pl. 1).
(3.) In the Coad Mountain-Pennock Mountain area as well as the Arlington area, blocks of Precambrian basement have been rotated during faulting.
(4.) Most thrust faults have sedimentary rocks on the hanging wall of the thrust that are steeply dipping, and locally overturned (Pl. 5).
(5.) Red gypsiferous shale and sandstone beds of the Chugwater and Goose Egg Formations are commonly the youngest units exposed on the hanging wall of the thrust. Apparently when the fault propagates into these beds of low competence the fault plane follows bedding planes in the red beds rather than cutting across more competent younger units.
(6.) The dip of the thrust planes has not been determined, but stratigraphic relationships show that some of the faults are low angle approaching 15 degrees in dip, at least, at the surface.
(7). Most of the thrust blocks have a syncline in sedimentary rocks in the foot wall below the thrust sheet, and anticlines and synclines with less amplitude may be present basinward (Fig. 51, Pl. 5).
(8.) Opposing thrusts with dips opposite to the main thrust may be present on the basinward side of the syncline.
(9.) Folding of sedimentary rocks was accomplished by flexure, and synclines near the surface that have steeply dipping limbs may pass downward to relatively moderately warped basement. Anticlines in sedimentary rocks may pass downward to faulted basement, rather than into tight folds with steeply dipping limbs.

LARAMIDE FOLDING OF ROCKS OF PRECAMBRIAN AGE AND BASEMENT CONTROL

To evaluate the role of the basement in the development of these Laramide faults and folds we must first consider the basic question of basement folding. Does the field evidence indicate that the basement has folded with the sedimentary rocks, i.e. does a flexural fold in the sedimentary rocks lie on a basement that has the exact configuration of the fold in sedimentary rocks. This was the view of Beckwith (1941, 1942) as shown by key cross-sections prepared by him (Fig. 52). Basement behavior of this type is difficult to conceive because it requires that a rigid relatively homogeneous group of rocks yield to stress in almost the same manner as a group of layered sedimentary rocks. The sedimentary rocks can yield by slippage on bedding planes, but the basement must either behave plastically under light load pressure or yield by movement on fractures, joints, or microfractures to take the exact configuration of folds in the overlying sedimentary rocks. Despite these difficulties many geologists who have studied folds of the type shown here have interpreted them in the same manner as Beckwith (Knight, 1953, Pl. 3, 5; Hudson, 1955, p. 2038-2052; Berg, 1962, p. 2019-2032; Love and Keefer, 1965; Blackstone, 1956, p. 3-20; Warner, 1956, p. 129-144; Richards, 1955, Pl. 4, Wilson, 1934, p. 498-522). The reason the basement is shown folded is that some well studied folds actually show a basement configuration comparable to that of overlying sedimentary beds (Wilson, 1934, p. 498-522; Hudson, 1955, p. 2039-2052), and there are many areas where map views of a base-

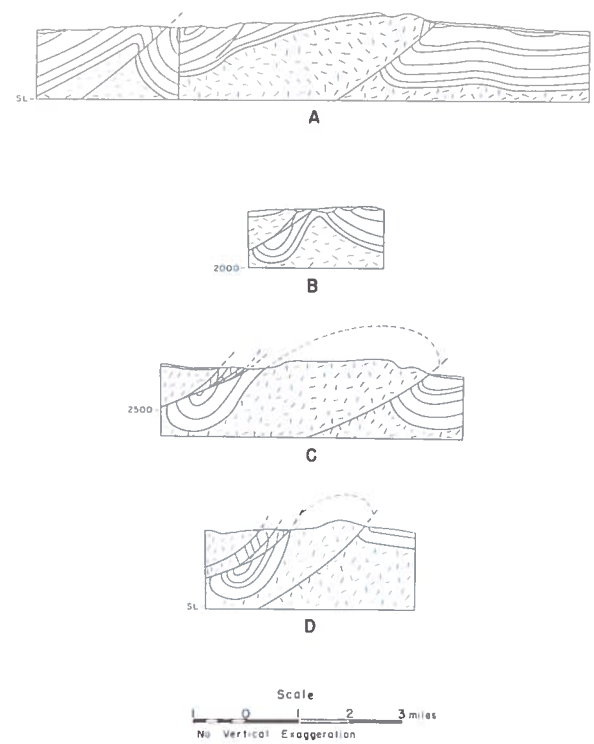


Figure 52.—Geologic cross sections by Beckwith illustrate his concept of tight folding of the Precambrian basement and fault planes with upward concavity. A—near northern end of Elk Mountain and Sheephead Mountain. B—South of Coad Mountain. C—South of Sheephead Mountain. D—Central Jelm Mountain. (Bearing of cross sections generally east)

Plate 35—Air phtograph of folded basement on Sheephead Mountain, secs. 32 and 33, T. 20 N., R. 82 W. (Pl. 1). Looking north Sheephead Mountain in left center of photograph. Dark core of overturned anticline is Precambrian gneiss. First white layer on Precambrian basement is Madison Limestone. Note that Precambrian-Madison contact conforms to geometry of overlying beds. (Photo by Pownall, University of Wyoming Photo Service.)

ment-sedimentary rock contact show an apparent perfect agreement in the structure of the folded sedimentary rocks and the shape of the basement. Furthermore, it is difficult to explain overturned sedimentary rocks on the hanging wall of a thrust without folding of the basement (Fig. 51, Pl. 5).

In the northern Medicine Bow Mountains, many structures show an apparent sedimentary rock-basement contact that is folded. Such structures are present at the north end of Elk Mountain, on Sheephead Mountain, on Coad Mountain, at the south end and south of Pennock Mountain, and west of Arlington (Pls. 1 and 5).

The fold on Sheephead Mountain T. 19 N., R. 82 W., is an anticline with Madison Limestone lying on quartzo-feldspathic gneiss of Precambrian age (Pls. 1, 5 and 35). The fold appears to be a very sharp asymmetric anticline with steep dips on the east side and the Precambrian rocks appear to be folded with the limestone. Beckwith's interpretation of this structure is shown in Figure 52A. The quartzo-feldspathic gneiss is so shattered in this area that reliable structural attitudes could not be found; so it is possible that the basement in this case has simply been deformed by closely spaced faults. At the south end of this structure the Tensleep Sandstone is overturned, but this may have been accomplished by a later thrust exposed to the south on Bear Mountain. In any event, the basement takes the configuration of the Laramide fold (Pl. 35).

At the north end of Elk Mountain, T. 20 N., R. 81 W., structural relationships are very similar to those at Sheephead Mountain, except that the Precambrian rocks are foliated granite. This may be a case of basement folding as indicated by Beckwith's cross-section (Fig. 52A). There is not enough information concerning basement to prove or disprove the thesis. Observed foliation in the granite strikes at right angles to the fold axes, and there is no evident relationship between the Laramide structure and the foliation of the Precambrian basement rock.

The south end of Coad Mountain, T. 18 N., R. 82 W., is folded and faulted with Madison Limestone in contact with well-foliated augen gneiss. This fold is thrust to the northeast and has an associated tear fault on the southeast side (Fig. 53, Pl. 1). The block involved is rotated counter clockwise. On the nose of the anticline the Madison Limestone is sliced into segments along northeast trending faults rather than folded around the nose of the anticline (Fig. 53). The Casper Formation however, is folded without obvious offset. By looking down the plunging nose of the anticline (Fig. 53) one can see that the fold has developed, in part, by fracturing of Madison Limestone on a series of faults and that the faults die out in the overlying Paleozoic formations that are simply folded by flexure. However, if these faults are removed the basement shows a configuration like that of the overlying northeast trending Laramide fold.

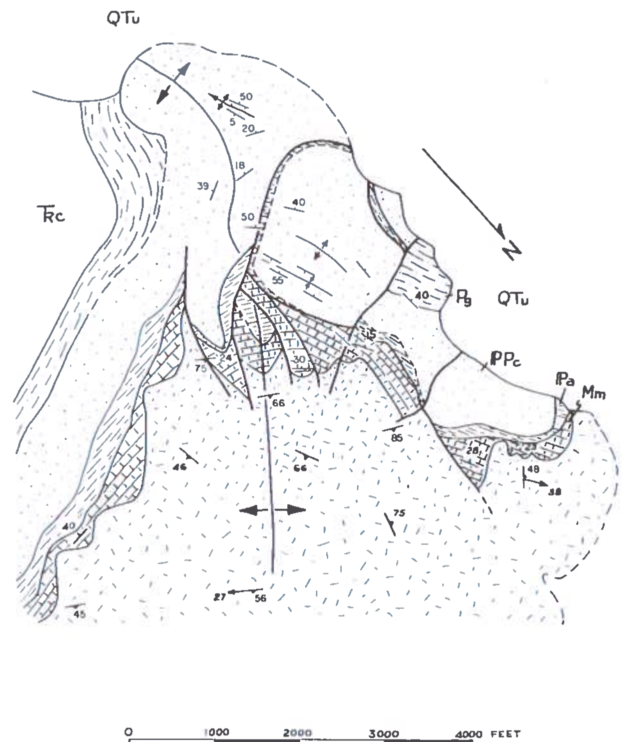


Figure 53—Laramide fold at south end of Coad Mountain. Precambrian rock is Coad Mountain augen gneiss, Mm = Missiissippian Madison Limestone, a = Pennsylvanian Amsden Formation, Pc = Casper Formation of Pennsylvanian and Permian age, Pg = Goose Egg Formation of Permian age, and c = Triassic Chugwater Formation. See text for explanation.

Pennock Mountain, T. 17-18 N., R. 82 W., is a large block that is thrust both to the northeast and southeast (Pl. 1). Offset to the northeast was greater than to the southeast since beds of the Niobrara Formation are in contact with Precambrian rocks on the hanging wall of the thrust in the northeast corner of the structure (Pl. 1). Three large blocks of Precambrian rock southeast of Pennock Mountain have been faulted and rotated during the movement of the Pennock Mountain block (Pl. 1, T. 17 N., R. 82 W.). It is probable that a large segment of Precambrian rocks moved at the time of formation of the Pennock Mountain thrust. The northeast trending fault at the southern end of the mountain may extend southwest for a distance of 8 to 10 miles (Pl. 1).

At the south end of Pennock Mountain, Madison Limestone lies on a basement of quartzo-feldspathic gneiss and related rocks that have a northwest-trending foliation. The Laramide structure is a northeast - trending syncline with the northwest limb overturned and thrust over the faulted southeast limb of the fold. South of this thrust the basement is part of the west limb of a major syncline. In the vicinity of the fault (secs. 11, 12, 13, 14 and 24, Tps. 17 and 18 N., R. 82 W.) the west limb of the syncline is faulted and overturned and the basement may be folded.

The role of the basement in the development of these structures can be partially evaluated. The overturned northwest limb of the fold at the south end of Pennock Mountain lies on gneiss with foliation that strikes perpendicular to the strike of the bedding in the Madison Limestone. The slope north of the Madison outcrops is covered with rubble so that there is no way to determine whether or not a fault in the Precambrian rock north of the Madison contact may have caused the reversal of dip of the Madison. In the southwest exposure of the fold (sec. 15) the sedimentary rocks are offset along a north-striking fault that parallels the foliation of the gneiss, and southwest of this fault the bedding in the sedimentary rocks is arched with the concave side to the west. In this area, the foliation of the gneiss roughly parallels the bedding of the sedimentary rocks but does not have the same dip. The gneiss near the contact with sedimentary rocks is complexly fractured, but no major fault on the west side of the overturned beds was noted. The basement seems to have the configuration of the Laramide fold, but this could not have developed by movement on the planes of foliation of the basement gneiss, unless movement was horizontal.

The two blocks of Precambrian rock immediately south of Pennock Mountain (secs. 12 and 14) both have Madison Limestone in contact with quartzo-feldspathic gneiss on the east flank. This repetition may be caused by a normal fault. In the western block, the strike and dip of bedding in the Madison Limestone conforms to the strike and dip of foliation in quartzo-feldspathic gneiss that underlies it. This is one of the rare instances where there is structural conformity between the foliation of Precambrian rocks and the strike and dip of overlying sedimentary beds.

South of this area (sec. 24, T. 17 N., R. 82 W.) a small north-trending body of Precambrian rocks has Madison Limestone on the east side. This body of Precambrian rocks is part of a larger block that is present to the southeast, and the larger block has relatively gently dipping Madison Limestone in its northeast flank. The dip of the Madison Limestone increases from 25° on the northeast flank of the larger block until it is overturned and dips 65° toward the Precambrian where it is in contact with the smaller block (Pl. 1). There are two structural trends in the Precambrian. The foliation in quartz monzonite as well as the bedding in a small body of quartzite strikes northwest, but a series of northeast trending bodies of amphibolite are in the quartz monzonite, and locally foliation in the quartz monzonite trends northeast. In the small north-trending body of Precambrian rock, amphibolite is present that may have this northeast trend. In the quartzite, northeast-trending foliation developed by closely spaced fractures, and a small inclusion of quartzo-feldspathic gneiss in the amphibolite also has northeast-trending foliation. The interesting aspect of this structure is that the structural change in the Precambrian units is related to the increase in dip of the Madison Limestone and overturning of the beds of this unit. The Madison Limestone and overlying units are offset by northeast-trending faults that parallel the foliation of the Precambrian rocks. If the foliation in the quartzite and the trend of the amphibolite body is Precambrian in age, then the Laramide deformation is controlled in part by slippage on planes of foliation. However, if the foliation of the quartzite is Laramide, this may be a case of Laramide microfractures developing in rocks of Precambrian age.

Another possible case of folding of Precambrian rocks is in the northeast corner of the Medicine Bow Mountains near Arlington, T. 19 N., R. 78-79 W. Here Precambrian rocks are displaced over beds of the Ferris-Hanna Formation along the northwest-trending Arlington thrust fault. Beds on the hanging wall of the thrust in T. 19 N., R. 79 and 80 W., form a northwest-plunging anticline. The relationship between underlying Precambrian rocks and the sedimentary units is especially clear here because metasedimentary rocks of Precambrian age are in contact with Madison Limestone (Pl. 1). To understand the development of this fold, one must relate it to the development of the Arlington fault. The Arlington thrust fault is a hinge-type thrust that rotated counterclockwise. It is bounded on the south by a northeast-trending transverse fault. The mass of Precambrian rocks was uplifted along the north side of the transverse fault and thrust to the northeast with the maximum displacement at the south end of the fault. Displacement decreases along the fault to the northwest where the fault passes into a northwest plunging anticline in T. 20 N., R. 80 W. (Pl. 34). The northwest plunging anticline at the north end of this thrust developed as a result of the rotation of the fault block, and its relationship to underlying Precambrian rocks is particularly revealing as regards the behavior of the different rock types during folding. On the northeast hanging wall of the thrust, the Casper Formation lies on Precambrian. There are a few thin remnants of Madison Limestone and limestone-conglomerate beneath the Casper. In the southernmost exposures (secs. 8, 9, 16, T. 18 N., R. 78 W.) the Casper is overturned and dips beneath the Precambrian rocks at 60 degrees southwest, but further north (sec. 5) the Casper dips east at 89 degrees where it lies on the Precambrian. In the vicinity of the fold (T. 19 N., R. 79 W.) the Madison Limestone is present and rests unconformably on the Precambrian where it strikes northeast and dips northwest between 10° and 18° (Pl. 1). The strike of the Madison Limestone changes from N. 80° E. to N. 45° E. along the broad gently plunging fold developed at the Madison-Precambrian contact. It is only slightly flexed and the northwest dip of the Madison is probably a result of tilting of the block of basement rock to the northwest. The anticline outlined by the Cloverly Formation in T. 19 N., R. 79 and 80 W. is a much sharper fold with a greater amplitude than the slight flexure at the Madison-Precambrian contact. This is an unusually good illustration of strikingly different behavior of units above and below the incompetent, easily deformed red beds. This also illustrates how complex folds may develop during uplift that are, in part, independent of basement structure. The problem of basement folding is thus less severe, but we are still faced with explaining the steeply dipping and overturned beds of the Casper Formation on the hanging wall of the thrust. In most areas where overturned beds are on the hanging wall of the thrusts the actual contact with the rocks of Precambrian age is not exposed, but in secs. 8 and 9, T. 18 N., R. 78 W. the Casper Formation is overturned and dips at 60° towards the Precambrian. Exposures in an old mine shaft north of Three-mile Creek and along the ridge south of the creek show a well developed fault containing brecciated fragments of Precambrian rocks and abundant fault gouge between the Precambrian and the overturned Casper Formation. Thus the structure here is similar to that south of Sheephead Mountain and the overturning of the Casper Formation can be explained by faulting. The strike of foliation in all of the Precambrian units including hornblende gneiss, quartzo-feldspathic gneiss and quartzite is parallel to the strike of bedding in the Casper, but the dip along the contact may or may not conform to the dip of the beds of the Casper. The thrust is located on the east limb of a north-northwest trending syncline in the metasedimentary units, however, so that most beds in the quartzite and most of the foliation in the gneissic units dips southwest parallel to the dip of the overturned beds of the Casper. The thrust sheet may conform to the structure of the metasedimentary rocks and the overturning of the beds of the Casper Formation may have been accomplished by movement on fault planes parallel to the foliation and bedding in the metasedimentary units.

The major north-trending syncline located in the Centennial Valley, T. 15 N., Rs. 77-78 W. south of Centennial is another possible example of basement folding (Pls. 1 and 5). On the east flank of Centennial Ridge and on the west flank of Sheep Mountain, Paleozoic rocks in contact with the Precambrian dip 45° E. and 45° W. respectively (Pls. 1 and 5). As illustrated in the cross-section for this area, this may be an example of basement folding producing a syncline with a wave length on the order of six miles (Pl. 5).

In our review of basement folding we find that in many examples, folding can be explained by major and minor faults in the basement that may or may not be transmitted to the surface, but in other examples, especially folds with large wavelengths it is clear that the basement was involved during Laramide deformation perhaps by movement on microfractures not noted in this study. The folds do not show a systematic relationship to the major structures of the Precambrian basement.

RELATIONSHIP OF LARAMIDE FAULTS TO BASEMENT STRUCTURE

It is not possible to discuss Laramide folding and faulting independently as so many of the major faults have been partly reviewed above, but key structures have not been discussed, and in this review of faulting an attempt will be made to specifically relate the faults to Precambrian structure. Many of the major thrust faults in the Medicine Bow Mountains start at some point we will refer to as the hinge and show increasing displacement in a direction away from this hinge. An excellent example is the thrust fault located at the southwest border of Centennial Valley in Tps. 14 and 15 N., R. 78 W. The hinge of this fault is in sec. 28, T. 15 N., R. 78 W. and displacement on the fault increases to the southeast where Precambrian rocks are thrust over successively younger rocks. The fault thus appears to show a counterclockwise rotation in map view (Pl. 1). Most of the major thrust faults are covered along much of their trace, but the thrusts that bound the mountain front from Centennial to the northern limit of the map as well as some of the thrust faults in the Coad Mountain-Pennock Mountain area are believed to show greater displacement to the southeast.

Thrust faults that are present on the east flank of the Medicine Bow Mountains from sec. 21, T. 15 N., R. 78 W., north of the Middle Fork of the Little Laramie River, north to the limit of the map (Pl. 1) pass into shear zones of Precambrian age at their southern limit. Starting with the Arlington thrust fault in sec. 14, T. 19 N., R. 79 W. we will consider these faults from north to south. The Arlington thrust fault hinges outside the map area but as shown in Plate 34, the hinge of this fault is probably located in the southeast corner of T. 20 N., R. 80 W. Displacement on the Arlington fault increases to the south with major offset at Corner Mountain (T. 16 N., R. 78 W.) where the hanging wall of the thrust is probably displaced several miles over the footwall (Pls. 1 and 5). The fault is segmented with two major divisions between Corner Mountain and Coyote Hill (sec. 2, T. 17 N., R. 78 W.). At Coyote Hill the fault turns almost at right angles and strikes northeast. This may be a tear fault on the hanging wall of the thrust, and may be related to a fault in Precambrian metasedimentary rocks at Rock Creek Knoll (sec. 35, T. 17 N., R. 79 W.). A segment of the Arlington thrust fault continues southeast to sec. 33, T. 17 N., R. 78 W. where it again turns sharply and seems to follow the strike of the Mullen Creek-Nash Fork shear zone. There may be right lateral movement along the Mullen Creek-Nash Fork shear zone and the Arlington thrust fault may be offset by a transverse fault at this point. The final segment of the fault continues southeast and again turns abruptly at Corner Mountain and passes into a northeast striking shear zone of the Precambrian age. The apparent right lateral displacement on the northeast striking fault at Corner Mountain is over two miles. The hanging wall may have moved about two miles northeast and the thrust may simply terminate in the northeast striking tear fault. The right lateral movement may not be entirely confined to the thrust sheet, however, since it is possible that the footwall may also be offset and the northeast striking fault may be both a transverse and tear fault.

The major point to be made is that the thrust sheet does not simply die out to the southeast, but turns and joins tear faults and/or transverse faults that are clearly reactivated Precambrian shear zones. These faults thus show a counterclockwise rotation in plan and probably also considerable greater uplift on the north side of the tear or transverse fault. The transverse fault has a steep dip if the dip of foliation in the shear zone is an indication of the dip of the fault so the movement on these faults is probably both vertical and right lateral.

A most interesting example of a transverse fault is exposed in secs. 20-22, T. 15 N., R. 78 W., north of the Middle Fork of the Little Laramie River. This Laramide fault strikes northeast and passes into a Precambrian shear zone. In map view there appears to be right lateral displacement on this fault, but the fault changes strike from northeast to north in sec. 15, T. 15 N., R. 78 W. where it is in red beds of the Chugwater Formation. At its northern limit beds of the Casper Formation appear thrust over the beds of the

Chugwater Formation. It is impossible to determine how much of the apparent right lateral displacement on this fault is due to vertical movement, but it is possible that this fault is an incipient fault of the Corner Mountain type. It may show the initial stage of development of a tear fault in the thrust sheet that could allow development of a low angle thrust like that at Corner Mountain.

There is little evidence of structural control by the foliation of Precambrian rocks of the basement in the southern Medicine Bow Mountains. Most of the folds and faults of this area trend north whereas the major trend of foliation in the rocks of Precambrian age is either northeast or east (Pl. 1). The Laramide folds and faults are generally at right angles to the trend of foliation in the Precambrian rocks, but one interesting departure from this rule is the structure along the southeast flank of the Medicine Bow Mountains proper (Pl. 1). Here in Tps. 12 and 13 N., R. 77 W., west of the highway, the foliation in quartz-andesine gneiss trends northeast roughly parallel to the strike of bedding on the west limb of a Laramide fold, but in sec. 34, T. 13 N., R. 77 W., the strike of foliation in the quartzo-feldspathic gneiss of this area is east, and an east trending shear zone is located approximately at this change in strike of the rocks of Precambrian age, (Pl. 1). North of this shear zone, the Precambrian rocks are thrust over rocks of Cretaceous age. The trace of this north-trending thrust is along the east margin of the mountains and can be followed to the south end of Sheep Mountain where it is offset by a northwest-trending fault located approximately at the contact between the Sherman Granite and older gneissic rocks of the Precambrian (Pl. 1). The thrust then continues to the north along the east margin of Sheep Mountain. The southern segment of the thrust is bounded by transverse faults, one controlled by a Precambrian shear zone and the other by a contact between granite and gneiss, and this is the only evidence of control of Laramide structure by basement in the southern Medicine Bow Mountains.

Another significant Laramide structure is the Jelm Mountain uplift (Pls. 1 and 5). This is an uplifted and rotated block of Precambrian rocks that is thrust to the east on the northeast flank of the block and thrust to the west on the southwest flank of the block (Pls. 1 and 5). The rotation has been clockwise. If this block is rotated back to its former position the foliation of the Precambrian rocks is in closer structural conformity with that of the Medicine Bow Mountains proper.

Review of Evidence for Basement Control

The interpretation of the wedge shape for the Jelm Mountain block requires that two thrust faults oppose each other opposite the syncline that separates Jelm Mountain from the east margin of the Medicine Bow Mountains (Pls. 1 and 5). Beckwith (1938, p. 1531-1533) interpreted both the fault on the west side of Jelm Mountain and on the east margin of the Medicine Bow Mountain as west dipping thrust faults (Fig. 52, D). Apparently Beckwith later changed his view because in his paper on the structure of the upper Laramie River Valley (the area due south of the southeast margin of this map area) he shows opposing thrust faults in almost the same situation described above (Beckwith, 1942, pl. 1). Apparently as these opposing thrust faults developed erosion kept pace with movement of the fault blocks otherwise one might expect the basement to be intensely fractured on opposite sides of the intervening syncline. In secs. 27 and 34, T. 14 N., R. 77 W., the hanging wall of the thrust fault on the east side of the Medicine Bow Mountains is complexly fractured where these two faults oppose each other (Pls. 1 and 5).

Most Laramide folds have axes that trend north or northeast. The north-trending folds are not related to major structures in the Precambrian rocks (Table 66). The northeast trending folds may be related to movement on Precambrian shear zones, but most northeast trending folds are east of the Mountains proper and cannot be directly related to the basement.

Faults of Laramide age show positive relationships to Precambrian structure (Table 67). Of 15 faults evaluated, 9 and possibly 10 (Pennock Mountain) show a positive correlation with a structure of Precambrian age. This is distinct contrast to folds. We may therefore consider the possibility that the process of basement folding or flexing takes place independently of basement structure whereas faulting of the basement is largely controlled by pre-existing structure. Possibly the early north-trending folds developed when load was sufficient to allow local basement folding under stress, but as load was reduced by erosion, relief from stress was achieved by uplift and faulting. As movement continued along fault planes, original folds were rotated from their former orientation and the basement may simply have moved as large blocks of relatively rigid material.

Table 66—Relationship of Precambrian structure to Laramide folds

Orientation of Laramide Fold	Orientation of Precambrian Structure
Elk Mountain	
N—S axial plane dips west	General trend of foliation N 70 E, Dip steep (60° ±)
Bear Mountain	
N—S axial plane dips west	General trend of foliation N 70 W, Dip steep (70° ±)
Coad Mountain	
Axial plane strikes N 60 E	Antiform plunging N 30 W
Pennock Mountain	
Uncertain	Antiform and synform plunging northwest
Centennial Valley	
Axial plane of syncline strikes N—S	Developed largely in granite (non-foliated)
Jelm Mountain	
N—S axial plane of anticline	E trending synform in Precambrian rocks
Sheep Mountain	
N—S axial plane dips west	Developed in Sherman Granite largely. At north end isoclinal folds in Precambrian gneiss—axial planes trend WNW

Table 67—Relationship of Precambrian structure to Laramide faults

Orientation of Laramide Fault	Orientation of Precambrian Structure
Elk Mountain	
Major thrust strikes N 25° W, dips west	Trend foliation N 70° E, dip steep (60° ±)
Bear Butte	
Major thrust strikes N 10 W, dips west	General trend of foliation N 70 W, dip steep (70° ±)
Coad Mountain	
Major thrust strikes N 30 W, dips west	Antiform—plunging N 30 W
Pennock Mountain	
Major thrust strikes from NW to N, dips west	Antiform and synform plunging NW
Thrust and or tear fault at south end of mountain strikes N 30° E, dips NW	
Arlington Area	
Main northern segment of thrust strikes N 30 W, dips southwest	Precambrian syncline strikes from N to N 30 W on hanging wall of thrust and dips 60° ± southwest
Tear fault near south end of major segment strikes N 60 E	Trend of bedding in quartzite variable from N 30 E to N 60 E
Tear fault at south end of segment probable strike NE	Fault Reactivated trace of Mullen Creek-Nash Fork shear zone
Corner Mountain	
Probable strike of major thrust N 55 W, dip southwest	General strike of foliation N 60 E, dip steep
Tear fault at south end of Corner Mountain thrust, strike N 60 E	Fault reactivated Precambrian shear zone
Middle Fork	
Tear fault strikes N 75 E	Fault reactivated Precambrian shear zone
Sheep Mountain	
Thrust strikes N—S generally and extends to within 4 miles north of Colorado border, dips west	Variable attitude of Precambrian structure—not related to trace of fault
Tear fault at south end of Sheep Mountain strikes N 80° W	Fault parallel to contact between granite and gneiss
Tear fault 4 miles north Colorado border strikes E	Fault reactivated Precambrian shear zone
Jelm Mountain	
Fault strikes generally north—may dip east or west	E striking synform in Precambrian rocks—steep dip of foliation

UPLIFT VERSUS COMPRESSION

The Laramide structures of the Rocky Mountain foreland have been attributed to vertical and horizontally applied stress and the relative importance of each has been debated for many years. Deformation by horizontally applied stress has been favored through most of this century, but a trend towards emphasizing vertical stress[1] for foreland folding and faulting of the Rocky Mountain Foreland (Osterwald, 1961; Prucha et. al., 1965; Kerr and Christie, 1965), is evident in more recent studies. It is sometimes difficult to determine whether a writer is simply impressed with the great amount of vertical movement as compared to horizontal movement or whether he is thinking of vertical stress vs. horizontal stress. Major vertical movement can, of course, result from horizontally applied stress, but Prucha (1965, p. 991) specifically states that the Laramide structural pattern in the Wyoming Province of the Rocky Mountain Foreland reflects dominantly vertical movement in the basement rather than regional horizontal **compression** and Osterwald (1961, p. 234) states that most uplifts and basins in the foreland result from deformation, with large components of vertical movement, more nearly related to epeirogeny than to orogeny.

Examination of geologic cross-sections of the Medicine Bow Mountains (Pl. 5) shows that the vertical component of movement is clearly greater than the horizontal component.

The major thrust faults of the Medicine Bow Mountains probably are concave downward or dip more steeply as the basement is approached (Pl. 5.) This has not been documented by this study and it is possible that some low angle thrusts may extend into the basement. Prucha, et. al. (1965, p. 984) have pointed out that thrusts should be concave upward if produced by regional compression and since they find no faults with such an attitude they discount this process. Beckwith obviously had this in mind when he drew cross-sections of the flanking folds of the Medicine Bow Mountains since he shows most thrusts as concave upward (Fig. 53) and he believed that the thrusts developed by regional compression. These two major points, the predominance of vertical movement and shape of the thrust proper suggest vertical movement. With a major vertical component of movement one would think that virtually any fault with a dip of 45° ± would have to be concave downward ultimately under the influence of gravity. In any event, in the Medicine Bow Mountains we have good evidence to suggest a strong vertical

1. About the same results might be expected if the basement were subject to tension with the basins actually representing down dropped blocks.

component of movement, but the shape of the fault plane is conjectural.

Proponents of vertical stress suggest that folds pass downward into faults and that the initial structures developed were faults; and folds were a secondary phenomena confined to the layered sedimentary rocks. It is exceedingly difficult to prove or disprove this thesis in the Medicine Bow Mountains. A number of examples have been given (i.e. Coad Mountain) where folds in the sedimentary rocks do indeed pass downward into faults in the basement. On the other hand when we examine some of the major and minor folds (north end of Sheephead Mountain, Coad Mountain, northwest of Arlington, north end of Elk Mountain), we find that the sedimentary rocks are warped around the seemingly folded basement. Perhaps more significant, cross-sections through the Centennial Valley suggest a simple syncline. In this connection, the general lack of correlation between basement structure and folds and the positive correlation between basement structure and faults suggest that the basement may have been folded independently of pre-existing structure, but as stress was continuously applied and cover removed it began to break-up along some pre-existing zones of weakness.

Particularly in the northern Medicine Bow Mountains, strike-slip movement along pre-existing fractures may be characteristic of the Laramide structure. This is easily explained as a result of horizontal stress, but such movement seems less likely in rocks subject to vertical stress. We must emphasize, however, that the examples of strike-slip movement have not been proven.

Rotation of blocks of the basement can be noted in both plan (Pl. 1) and cross-section (Pl. 5). The series of blocks along the northeast side of the Medicine Bow Mountains have moved to the northeast on the hanging wall of thrusts with right lateral movement on bounding tear transverse faults (counterclockwise rotation in plan). This type of rotation can be explained as a result of strike-slip movement on the northeast-trending shear zones causing counterclockwise rotation of basement blocks. The north-striking thrust faults on the east side of the mountains would be expected to decrease in displacement to the north in such a situation.

If we consider the fault north of the Middle Fork of the Little Laramie River (secs 15, 20, and 21, T. 15 N., R. 78 W.) from the viewpoint of purely vertical movement without horizontally applied stress it seems to the writer that the basement must first be folded. If movement on this fault is purely vertical the north side would be elevated, but the syncline with east dipping bedding in sedimentary rocks would have to be present prior to the development of the fault. If this fault developed prior to folding (the concept of vertical faults preceeding or passing into folds at higher levels) the relationship of bedded sedimentary rocks on the south side of the fault in section 21 should be like that on the east side of the fault in section 15 (that is beds should dip away from the fault in all directions perhaps like the structures at the south end of Pennock Mountain). Thus even with purely vertical movement on this fault folding must preceed faulting and to the writer the folds must develop under horizontal compression.

Additional support for horizontal stress is the development of wedge-shaped blocks with opposing thrusts on opposite sides of synclines. This is especially well-shown in the Jelm Mountain area of the southeastern Medicine Bow Mountains (Pls. 1 and 5).

In summary, the Laramide structure is best explained as a result of tangential stress. The basement appears to have somehow developed folds perhaps concentrated in zones near the present mountain front during early stages of deformation that were completely independent of gross structure within the Precambrian rocks. As stress continued and overburden was removed a strong vertical component of movement began to develop and the basement began to break up into blocks, controlled in part by pre-existing faults, shear zones and major linements of the basement structure. Deformation was concentrated in fault zones and many complex structures developed locally as a result of crowding and rotation of fault blocks. Folds in the sedimentary rocks may be detached from basement folds or greatly exaggerated as a result of flowage in incompetent units.

This interpretation is quite similar to that of Foose and others (1961) who state that the rise of the Beartooth block of northern Wyoming and south central Montana resulted from horizontal compression. These writers envision early folding under horizontal compression followed by vertical uplift along major faults, and where vertical displacement was great enough and there was no buttressing against horizontal compression, horizontal displacement with associated imbrication and tear faulting resulted (Foose and others, 1961, p. 1167).

Although evidence in the Medicine Bow area seems to support a tangential stress, it should not be interpreted as ruling out vertical uplift on a regional basis. Certainly during regional deformation one area may be subject to horizontal compression while another subject to vertical stress or tensions. It is also possible that basement folding and faulting along the mountain flank is a function of crowding during vertical uplift. It must be emphasized that the problem of vertical versus horizontal stress is as important to the student of Precambrian rocks in the Rocky Mountains as it is to the student of Laramide structure. If the basement can be folded under light load as is suggested by this study, are folds in the central areas of Precambrian rocks entirely immune or are some of the "Precambrian" structures actually Laramide?

ECONOMIC GEOLOGY

Although the description of mineral deposits examined during this study will be reported elsewhere (see chapter on plan and purpose) a brief review of the nature of mineralization will be included here.

The Medicine Bow Mountain area includes numerous local concentrations of minerals of possible economic interest. No metallic deposits are being mined today, but many small mines were developed at the turn of the century. Petroleum and natural gas are produced from structures located east and north of the mountain area, and local deposits of non-metallic minerals have been mined in recent years. The mineral deposits will be discussed as they relate to rock types of different ages starting with the Precambrian.

Quartzo-feldspathic Gneiss north of the Mullen Creek-Nash Fork Shear Zone

The major mineral deposits in this unit are vermiculite deposits that are in two antiforms; one south of Baggot Rocks and another east of Beaver Creek and west of State Highway 230, Tps. 13 and 14 N., R. 82 W. (Pl. 1). These deposits are in areas where amphibolized and deformed mafic dikes and sills are cut by pegmatite. The amphibolite is transformed to vermiculite through a biotite schist facies. These deposits were probably formed during a late stage of deformation when pegmatite was formed or emplaced in the antiforms. The deposits are small but numerous and have been mined in recent years as a source of vermiculite for plant fertilizer.

Deep Lake Formation

The major mineral deposits in rocks of the Deep Lake Formation are in quartz veins, that for the most part, are in or associated with mafic intrusive bodies. These veins are usually along the margins of mafic intrusive bodies but may be in quartzite and other metasedimentary rocks near contacts with mafic intrusive bodies.

The major concentration of veins is in the Gold Hill area (center of T. 16 N., R. 80 W.) where quartz veins containing native gold, chalcopyrite, and pyrite are present. Most of these veins strike northwest and those noted on the surface are small (around 3' in thickness) and discontinuous. The quartz veins show ribbon structure and sericite, pink feldspar, epidote, and amphibole are present as gangue minerals.

A number of very poorly exposed prospects with limonite gossans apparently derived from sulphide minerals are present in the southern part of T. 18 N., R. 79 W. These gossans are in areas where scattered outcrops of metagabbro, very fine-grained amphibolite and rocks of questionable volcanic origin are present.

The Deep Lake Formation has many beds of conglomerate some clean, well-sorted and consisting mostly of clasts of granule and pebble size and others with open framework, chlorite matrix, and larger clasts. The better sorted conglomerates should be examined as a possible source of gold and other heavy minerals of economic interest. The stratigraphy of this unit is not well enough known to specify a particular part of the unit for exploration but this type of conglomerate is perhaps more often found near the top of the unit where it is in contact with Headquarters Schist.

Libby Creek Group

Mineralized quartz veins associated with mafic intrusions are also present in the Libby Creek Group. These veins resemble those in the Deep Lake Formation and are common in the south where a large body of metagabbro (?) is exposed.

The Nash Fork Formation contains a number of mineral occurrences of possible economic interest. Local traces of copper are in metadolomite of this unit, sec. 19, T. 15 N., R. 79 W., and the phyllite may be exceptionally rich in iron sulphide and graphite in some areas, most notably that body of phyllite in the center of the unit where it follows the course of French Creek (Pl. 1).

Shear Zones

Many of the shear zones shown on Plate one are mineralized locally. For example, copper and gold prospects are common along the Mullen Creek-Nash Fork shear zone especially in that area west of the Platte River. Preliminary sampling of sulphide rich zones in these shear zones have shown disappointing gold assays, but only a very small percentage of outcrops have been studied.

Rocks South of the Mullen Creek-Nash Fork Shear Zone

Pegmatites that contain rare earth minerals are found in the Big Creek area (Houston, 1961) and are present locally in an east trending belt of amphibolite and hornblende gneiss in the southern Medicine Bow Mountains. Some pegmatites also contain copper sulphide minerals in unusual amounts for pegmatites (Houston, 1961, p. 10). Most of these pegmatites are in the axes of antiforms somewhat like the concentrations of pegmatite associated with vermiculite north of the Mullen Creek-Nash Fork shear zone. These pegmatites are probably too small to be considered a source of feldspar for commercial use.

Reports on the major mineral occurrences south of the Mullen Creek-Nash Fork shear zone have been published (Houston, 1961; Currey, 1965; McCallum, 1968; McCallum and Orback, 1968). The most important are northwest striking gold-copper quartz veins peripheral to the Keystone Quartz Diorite (Currey, 1965), and copper-gold-platinum prospects in the New Rambler and Centennial areas (McCallum and Orback, 1968; McCallum, 1968, Theobald and Thompson, 1968). The reader interested in mineral prospects of this area should consult the above reports and examine theses (listed on Plate one) that show geology in greater detail than on Plate one and describe mines and prospects of this area.

Petroleum and Natural Gas

The major oil fields that developed as a result of the uplift of the Medicine Bow Mountains are east and north of the map area (Pl. 1). Two areas within the map area may be of interest to petroleum geologists. East of Rock Creek Ridge in T. 17 N., R. 78 W., outcrops of the Cloverly Formation may show a reversal of dip. Outcrops are very poor in this area, but it is, at least, possible that an anticline is present. East of Pennock and Coad Mountains isolated outcrops suggest that faulted anticlines are present (Pl. 1). Two dry holes have been drilled on a structure at the south end of these prospects (Gries, 1964).

Miscellaneous Mineral Prospects

As is the case for most large areas, there are many occurrences of rocks and minerals that may have ultimate economic value. For example, metadolomite of the Nash Fork Formation makes good road material, deep green quartzite found in the upper part of the Medicine Peak Quartzite is valuable as a decorative stone and in the manufacture of granules, and peat, common on upland surfaces of the central Medicine Bow Mountains and has been used as a plant fertilizer and soil builder in the Laramie area (McCallum, 1964, p. 134 and 135).

A number of areas both north and south of the Mullen Creek-Nash Fork shear zone have been mined for placer gold. Many of the streams in the vicinity of Keystone were placered around 1900 with the most extensive mining on Douglas Creek (Currey, 1965). Placer operations have also been carried out on the northeast margin of the Medicine Bow uplift with major activity in the upper reaches of Mill Creek. A unique placer operation that consisted of washing of conglomerate of the Ferris-Hanna Formations, undivided, was carried out in sec. 19, T. 16 N., R. 77 W. Water for this operation came from springs and was stored in large barrels and released for washing the conglomerate.

Suggestions for Prospecting

We would like to first re-emphasize that additional information on mines and prospects is available in original works cited on Plate one, and that details of certain mineralized areas are presented in Preliminary Reports of the Geological Survey of Wyoming.

Although no mineral deposits of major economic importance have been found in the Medicine Bow Mountains certain factors indicate that prospecting would be warranted. Much of the northern, central, and south central part of the Medicine Bow area is characterized by very poor exposure. The northern and central area is covered with glacial deposits and the high level erosion surface of the south-central Medicine Bow Mountains is either deeply weathered creating poor exposure or covered with a thin veneer of gravel and claystone. Prospecting, which was done chiefly from 1880-1910 was by surface methods only, and to date modern methods, geochemical and geophysical techniques, have not been tried.

Certainly this area deserves much more detailed geologic mapping than has been done in this study, and it is the writer's hope that detailed geologic study plus use of modern prospecting tools will result in discoveries of mineral deposits of economic value.

Conclusion

The geologist is normally and legitimately concerned with materials of economic value; those rock and mineral products than can be used for the benefit of mankind. In many areas studied by geologists, the country is such that equal or greater value may come simply from the natural beauty of the area. Few areas in the United States offer such a wide variety of scenery as the Medicine Bow Mountains. The crest of the mountain near Lake Marie is one of the most beautiful areas in the State of Wyoming. We hope that this geologic report will be of interest to those concerned simply with scenery and that the photographs will illustrate more than simply geologic phenomena.

REFERENCES CITED

Albanese, J. P., 1949, A petrographic and structural study of a periodotite body in the Laramie Mountains, Wyoming: unpublished M.S. thesis, University of Wyoming, 64 p.

Aldrich, L. T., Wetherill, G. W, Davis, G. L., and Tilton, G. R., 1958, Radioactive ages of micas from granitic rocks by Rb-Sr and K-Ar methods: Am. Geophys. Union Trans., v. 39, p. 1124-1134.

Ashley, William, 1947, Geology of the Kennaday Peak-Pennock Mountain area, Carbon County, Wyoming: unpublished M.A. thesis, Univ. of Wyoming, 71 p.

Atwood, W. W., Jr., 1937, Records of Pleistocene glaciers in the Medicine Bow and Park Ranges: Jour. Geology, v. 45, p. 113-140.

Baker, C. L., 1946, Geology of the northwestern Wind River Mountains, Wyoming: Geol. Soc. America Bull., v. 57, p. 565-596.

Balk, Robert, 1949, Structure of Grand Saline salt dome, VanZandt County, Texas: Am. Assoc. Petrol. Geol. Bull., v. 33, p. 1791-1829.

———, 1953, Salt structure of Jefferson Island salt dome, Iberia and Vermilion Parishes, Louisiana: Am. Assoc. Petrol. Geologists Bull., v. 37, p. 2455-2474.

Balk, R., and Barth, T. F. W., 1936, Structural and petrological studies in Dutchess County: Geol. Soc. America Bull., v. 47, p. 685-774, 775-850.

Barth, T. F. W., 1962, Theoretical petrology: John Wiley Pub. Co., New York City, 416 p.

Bassett, W. A., and Giletti, B. J., 1963, Precambrian ages in the Wind River Mountains, Wyoming: Geol. Soc. America Bull., v. 74, p. 209-212.

Bauer, E. J., 1951, Geology of the Wagonhound Creek area, Carbon County, Wyoming: unpublished M.A. thesis, Univ. of Wyoming, 58 p.

Bayley, R. W., 1963. A preliminary report on the Precambian iron deposits near Atlantic City, Wyoming: U.S. Geol. Survey Bull., 1142-C, 23 p.

———, 1965, Geologic map of the South Pass City quadrangle, Fremont County, Wyoming: U. S. Geol. Survey map GQ-458.

———, 1965, Geologic map of the Atlantic City quadrangle, Fremont County, Wyoming: U. S. Geol. Survey map GQ-459.

———, 1965, Geologic map of the Miners Delight quadrangle, Fremont County, Wyoming: U. S. Geol. Survey map GQ-460.

———, 1965, Geologic map of the Louis Lake quadrangle, Fremont County, Wyoming: U. S. Geol. Survey map GQ-461.

Beckwith, Radcliffe H., 1938, Structure of the southwest margin of the Laramie Basin, Wyoming: Geol. Soc. America Bull., v. 49, p. 1515-1544.

———, 1939, Asbestos and chromite deposits of Wyoming: Geol. Survey of Wyoming Bull. 29, 31 p.

———, 1941, Structure of the Elk Mountain district, Carbon County, Wyoming: Geol. Soc. America Bull., v. 52, p. 1445-1486.

———, 1942, Structure of the upper Laramie River Valley, Colorado-Wyoming: Geol. Soc. America Bull., v. 53, p. 1491-1532.

Beeler, Henry C., 1904, Wyoming mines and minerals in brief: Cheyenne, Wyo., 15 p.

———, 1905, Mining in the Grand Encampment copper district, Carbon and Albany Counties, Wyoming: Cheyenne, Wyo., 32 p.

———, 1906, Mineral and allied resources of Albany County, Wyoming: Laramie, Wyo., 80 p.

———, 1908, Wyoming mines, 1907; a summary of the conditions and progress in the copper, gold, placer, iron, asbestos, sulphur, stone, plastic minerals, gravel, oil, natural gas, and coal industries: Laramie, Wyo., 46 p.

Berg, Robert R., 1962, Mountain flank thrusting in Rocky Mountain foreland, Wyoming and Colorado: Am. Assoc. of Petrol. Geologists Bull., v. 46, p. 2019-2032.

Bishop, D. T., 1963, Retrogressive metamorphism in the Seminoe Mountains, Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 49 p.

Blackstone, D. L., Jr., 1953, Notes on the tectonic map of a portion of southern Wyoming and northern Colorado: in Wyo. Geol. Assoc., Guidebook, 8th Ann. Field Conf., p. 85-86.

———, 1956, Introduction to the tectonics of the Rocky Mountains, in Am. Assoc. Petrol. Geologists Rocky Mtn. Sec., Geological Record, p. 3-19.

———, 1965, Development of Laramide structure in Laramie Basin, Wyoming: (Abs.) Am. Assoc. Petrol. Geologists Bull., v. 49, p. 115.

Bothner, W. A., 1967, Structure, petrology, and economic geology of the Cooney Hills, Platte County, Wyoming: unpublished Ph.D. thesis, Univ. of Wyoming, 130 p.

Blackwelder, Eliot, 1926, Pre-cambrian geology of the Medicine Bow Mountains: Geol. Soc. America Bull., v. 37, p. 615-658.

Bouma, Arnold, H., 1962, Sedimentology of some flysch deposits; a graphic approach to facies interpretation: Elsevier Pub. Co., Amsterdam, 168 p.

Bowen, Charles F., 1918, Stratigraphy of the Hanna Basin, Wyoming: U. S. Geol. Survey Prof. Paper 108, p. 227-241.

Braddock, W. A., 1966, Precambrian geology on the east flank of the Front Range near Ft. Collins, Colorado: (Abs.) Geol. Soc. America Special Paper 87, p. 277.

Buddington, Arthur F., 1939, Adirondack igneous rocks and their metamorphism: Geol. Soc. America Mem. 7, 354 p.

Burke, Creighton A. (Editor), 1956, Subsurface stratigraphy of the pre-Niobrara formations in Wyoming, Pt. 1 of Wyoming Stratigraphy: Wyo. Geol. Assoc., Casper, 97 p.

———, 1953, Electric log correlation of the Triassic rocks of southeastern Wyoming: in Wyo. Geol. Assoc., Guidebook, 8th Ann. Field Conf., p. 29-33.

Burk, Creighton A., and Thomas, Horace D., 1956, The Goose Egg formation (Permo-Triassic) of eastern Wyoming: Geol. Survey of Wyo. Rept. Inv., no. 6, 11 p.

Butler, J. R., 1966, Geologic evolution of the Beartooth Mountains, Montana and Wyoming—Pt. 6 Cathedral Peak area, Montana: Geol. Soc. America Bull., v. 77, p. 45-64.

Campbell, John A., 1963, Permo-Triassic red beds, northern Denver Basin: Rocky Mtn. Assoc. of Geologists, Guidebook, Northern Denver Basin and adjacent uplifts, p. 105-110.

Cannon, R. S., Jr., Bayley, R W., Stern, T. W., and Pierce, A. P., 1966, Ancient rocks and ores in south-central Wyoming: (Abs.) Geol. Soc. America Spec. Paper No. 87, p. 279.

Carey, B. D., 1959, Geology of the Rattlesnake Hills Tertiary volcanic field, Natrona County, Wyoming: unpublished Ph D. Thesis, Univ. of Wyoming, 247 p.

Casella, C. J., 1964, Geologic evolution of the Beartooth Mountains, Montana and Wyoming—Pt. 4, Relationship between Precambrian and Laramide structures in the Line Creek area: Geol. Soc. America Bull., v. 75, p. 969-986.

Catanzaro, E. J., 1957, A preliminary petrographic study of the Lake Owens mafic complex, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 39 p.

————, 1966, Correlation of some Precambrian rocks and metamorphic events in parts of Wyoming and Montana: Mountain Geologist, v. 4, p. 9-21.

Catanzaro, E. J., and Kulp, J. L., 1964, Discordant zircons from the Little Belt (Montana), Beartooth (Montana), and Santa Catalina (Arizona) Mountains: Geochim. et. Cosmochim. Acta, v. 28, p. 87-124.

Chayes, Felix, 1956, Petrographic modal analysis: John Wiley and Sons, New York, 113 p.

Childers, Milton O., 1957, The geology of the French Creek area, Albany and Carbon Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 58 p.

Clarke, F. W., 1924, Data of geochemistry: U. S. Geol. Survey Bull. 770, 140 p.

Condie, Kent, 1967, Petrologic reconnaissance of the Precambrian rocks in the Wind River Canyon, central Owl Creek Mountains, Wyoming: Contrib. to Geology, Vol. 6, No. 1, p. 123-128.

Craig, Lawrence C., and Cadigan, Robert A., 1958, The Morrison and adjacent formations in the Four Corners area: in Intermountain Assoc. Petrol. Geologists, Guidebook, 9th Ann. Field Conf., p. 182-192.

Currey, Donald R., 1959, Geology of the Keystone area, Albany County, Wyoming, unpublished M.S. Thesis, Univ. of Wyoming, 64 p.

————, 1965, The Keystone gold-copper prospect area, Albany County, Wyoming: Geol. Survey of Wyo. Prelim. Rept. 3, 12 p.

Darton, Nelson H., Blackwelder, Eliot, and Siebenthal, C. E., 1910, Description of the Laramie and Sherman quadrangles, Wyoming: U. S. Geol. Survey Geol. Atlas, No. 173, 17 p.

Darton, N. H. and Page, Sidney, 1925, Description of the central Black Hills: U. S. Geol. Survey Geol. Atlas, No. 219, 34 p.

Davidson, P. S., 1966, The stratigraphy of the Upper Cretaceous Lewis Formation, Albany and Carbon Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 207 p.

Davis, John C., 1963, Origin of the Mowry shale: Contrib. to Geol., v. 2, p. 135-146.

Davis, J. R., 1966, Stratigraphy and depositional history of upper "Mesaverde" Formation of southeastern Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 122 p.

Dietrich, R. V., 1960, Banded gneisses: Jour. Petrology, v. 1, p. 99-120.

Dobbin, C. E., Bowen, C. F., and Hoots, H. W., 1929, Geology and coal and oil resources of the Hanna and Carbon basins, Carbon County, Wyo.: U.S. Geol. Survey Bull. 804, 88 p.

Donath, Fred A., and Parker, Ronald B., 1964, Folds and folding: Geol. Soc. America Bull., v. 75, p. 45-62.

Dorf, Erling, 1959, Climatic changes of the past and present: Mich. Univ. Mus. Paleontology Contr. v. 13, p. 181-210.

Dott, Robert H., Jr., 1961, Squantum "tillite", Massachusetts—Evidence of glaciation or subaqueous mass movements?: Geol. Soc. America Bull., v. 72, p. 1289-1305.

Dott, Robert H., Jr., and Howard, John K., 1962, Convolute lamination in non-graded sequences: Jour. Geology v. 70, p. 114-121.

Ebbett, Ballard, E., 1956, Structure and petrology of the metasedimentary rocks of the Good Fortune mining area, Platte County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 150 p.

Eckelmann, Frank D., and Poldervaart, Arie, 1957, Archean history of the Quad Creek area, Pt. 1 of Geologic evolution of the Beartooth Mountains, Montana and Wyoming: Geol. Soc. America Bull., v. 68, p. 1225-1261.

Eicher, Don L., 1960, Stratigraphy and micropaleontology of the Thermopolis shale: Yale Univ. Peabody Mus. Nat. History Bull. 15, 126 p.

Engel, A. E. J., 1947, Geology of the central Owl Creek Mountains, Wyoming: (Abs.) Geol. Soc. America Bull., v. 58, p. 1177-1178.

————, 1963, Geologic evolution of North America: Science, v. 140, p. 143-152.

Engel, A. E. J. and Engel, Celeste, 1951, Origin and evolution of hornblende-andesine amphibolite and kindred facies: (Abs.) Geol. Soc. America Bull., v. 62, p. 1435-1436.

Eskola, Pentti, 1932, Conditions during earliest geological times: Ann. Acad. Sci. Fennicae, Ser. A, v. 36, p. 16.

Evans, B. W., and Leake, B. E., 1960, The composition and origin of the striped amphibolites of Connemara, Ireland: Jour. Petrology, v. 1, p. 337-363.

Faessler, D. M., 1948, Simon Lake area, Papineau County: Quebec Bur. Mines, Geol. Survey, Geol. Rept. 33, 29 p.

Fenton, Carroll L., and Fenton, Mildred A., 1939, Precambrian and Paleozoic algae: Geol. Soc. America Bull., v. 50, p. 89-126.

Ferris, Clinton S., Jr., 1964, Petrology and structure of the Precambrian rocks southeast of Encampment, Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 74 p.

————, 1966, Structural analysis of the Precambrian rocks southeast of Encampment, Wyoming: Contrib. to Geology, v. 5, p. 13-19.

Ferris, Clinton S., Jr., and Krueger, Harold W., 1964, New radiogenic dates on igneous rocks from the southern Laramie Range, Wyoming: Geol. Soc. America Bull., v. 75, p. 1051-1054.

Fields, E. D., 1963, Precambrian rocks of the Halleck Canyon area, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 91 p.

Foose, R. M., Wise, D. U., and Garbarini, G. S., 1961, Structural geology of the Beartooth Mountains, Montana and Wyoming: Geol. Soc. America Bull., v. 72, p. 1143-1172.

Fyfe, W. S., Turner, F. J., and Verhoogen, J., 1958, Metamorphic reactions and metamorphic facies: Geol. Soc. America Mem., no 73, 259 p.

Gast, P. W., Kulp, J. L., and Long, L. E., 1958, Absolute age of early Precambrian rocks in the Bighorn Basin of Wyoming and Montana, and southeastern Manitoba: Am. Geophys. Union Trans., v. 39, p. 322-334.

Giletti, B. J., and Gast, P. W., 1961, Absolute age of Precambrian rocks in Wyoming and Montana: in Geochronology of rock systems: New York Acad. Sci. Annals, v. 91, p. 454-458.

Gilluly, James, 1963, The tectonic evolution of the western United States: Quart. Jour. Geol. Soc. London, v. 119, p. 139-140.

Gliozzi, James, 1967, Petrology and structure of the Precambrian rocks of the Copper Mountain district, Owl Creek Mountains, Fremont County, Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 141 p.

Goldich, S. S., Nier, A. O., Baadsgaard, Halfdan, Hoffman, John H., and Krueger, H. W., 1961, The Precambrian geology and geochronology of Minnesota: Minnesota Geol. Survey Bull. 41, 193 p.

Goldich, S. S., Lidiak, E. G., Hedge, C. E., and Walthall, F. G., 1966, Geochronology of the Midcontinent Region, United States—Part 2, northern area: Jour. Geophysical Research, v. 71, p. 5389-5408.

Green, Jack, and Poldervaart, Arie, 1955, Some basaltic provinces: Geochimica et Cosmochimica Acta, v. 7, p. 177-188.

Gries, John C., 1964, The structure and Cenozoic stratigraphy of the Pass Creek Basin area, Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 69 p.

Hagner, Arthur F., 1951, Anorthosite of the Laramie Range, Albany County, Wyoming, as a possible source of alumina: Geol. Survey of Wyo. Bull. 43, 15 p.

————, 1944, Wyoming vermiculite deposits: Geol. Survey of Wyo. Bull. 34, 47 p.

Hague, Arnold, and Emmons, S. F., 1877, Descriptive geology: U. S. Geological Expl. 40th Parallel (King), 2, 890 p.

Hamilton, Homer E., 1964, Stratigraphy and depositional history of the Muddy formation in the Freezeout Mountains, northeastern Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 125 p.

Hanley, John B., Heinrich, Eberhardt W., and Page, Lincoln R., 1950, Pegmatite investigations in Colorado, Wyoming, and Utah, 1942-44: U. S. Geol. Survey Prof. Paper 227, 125 p.

Harris, R. L., Jr., 1959, Gardner Lake area, Pt. 3 of Geologic evolution of the Beartooth Mountains, Montana and Wyoming: Geol. Soc. America Bull., v. 70, p. 1185-1216.

Harrison, J. E., 1951, Relationship between structure and mineralogy of the Sherman granite, southern part of the Laramie Range, Wyoming-Colorado: unpublished Ph.D. Thesis, Univ. of Illinois.

Harrison, J. E., and Wells, J. D., 1959, Geology and ore deposits of the Chicago Creek Area, Clear Creek County, Colorado: U. S. Geol. Survey Prof. Paper 319, 92 p.

Haun, John D., and Kent, Harry C., 1965, Geologic history of Rocky Mountain region: Am. Assoc. Petrol. Geologists Bull., v. 49, p. 1781-1800.

Hayden, Ferdinand V., 1863, On the geology and natural history of the upper Missouri: Trans. Am. Philos. Soc., New Ser., v. 12, p. 1-218.

————, 1868, Second annual report of the United States Geological Survey of the Territories, embracing Wyoming: U. S. Gen. Land Off. Rp. 1868, p. 111-120, 220-255.

————, 1872, Preliminary Report of the United States Geological Survey of Montana and Portions of adjacent Territories, being a fifth annual report of progress: Washington, D.C., 1872, 538 p.

Hedge, C. E., Peterman, Z. E., and Braddock, W. A., 1967, Age of the major Precambrian regional metamorphism in the northern Front Range, Colorado: Geol. Soc. America Bull., v. 78, p. 551-558.

Heinrich, Eberhardt W., 1956, Microscopic petrography: McGraw-Hill Co., New York City, 296 p.

Hensley, Frank S., Jr., 1955, Precambrian algae of the Medicine Bow Range, Wyoming: (Abs.) Geol. Soc. America Bull., v. 66, p. 1676.

Hills, F. Alan, Gast, Paul W., and Houston, Robert S., 1965, Chronology of some Precambrian igneous and metamorphic events of the Medicine Bow Mountains, Wyoming: (Abs.) Geol. Soc. America Special Paper, No. 82, p. 92.

Hills, F. Alan, Gast, Paul W., Houston, Robert S., and Swainbank, Ian, 1968, Precambrian geochronology of the Medicine Bow Mountains of southeastern Wyoming; Geol. Soc. America Bull., v. 79, p. 1757-1784.

Hobbs, S. W., Griggs, A. B., Wallace, R. E., and Campbell, A. B., 1965, Geology of the Coeur d'Alene District, Shoshone County, Idaho: U. S. Geol. Survey Prof. Paper 478, 139 p.

Hodge, D. S., 1963, Polymetamorphism of Precambrian rocks in the southwestern Wind River Mountains, Fremont County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 49 p.

————, 1966, Petrology and structural geometry of Precambrian rocks in the Bluegrass area, Albany County, Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 135 p.

Hodge, D. S., and Worl, R. G., 1965, Multiple metamorphic episodes in Precambrian rocks near South Pass, Wyoming: Contrib. to Geology, v. 4, p. 51-57.

Hoppin, R. A., 1961, Precambrian rocks and their relationship to Laramide structure along the east flank of the Bighorn Mountains near Buffalo, Wyoming: Geol. Soc. America Bull., v. 72, p. 351-367.

Houston, Robert S., 1961, The Big Creek pegmatite area, Carbon County, Wyoming: Geol. Survey of Wyo. Prelim. Rept. 1, 11 p.

Houston, Robert S., and Parker, Ronald B., 1963, Structural analyses of a folded quartzite, Medicine Bow Mountain, Wyoming: Geol. Soc. America Bull., v. 74, p. 197-202.

Hoy, Robert B., 1962, Structure of Winnfield salt dome, Winn Parish, Louisiana: Am. Assoc. Petrol. Geologists, Bull., v. 46, p. 1444-1459.

Hudson, Frank S., 1955, Folding of unmetamorphosed strata superjacent to massive basement rocks: Am. Assoc. Petrol. Geologists Bull., v. 39, p. 2038-2052.

Hyden, Harold J.; McAndrews, Harry; and Tschudy, Robert H., 1965, The Foote Creek and Dutton Creek formation, two new formations in the north part of the Laramie Basin, Wyoming: U. S. Geol. Survey Bull. 1194-K, p. K1-K12.

Hyden, H. J.; Houston, R. S.; and King, J. S., 1968, Geologic map of the White Rock Canyon quadrangle, Carbon County, Wyoming: U. S. Geol. Survey open file report.

Hyden, H. J.; King, J. S.; and Houston, R. S., 1967, Geologic map of the Arlington quadrangle, Carbon County, Wyoming: U. S. Geol. Survey map GQ-643.

James, Harold L., 1954, Sedimentary facies of iron formation: Econ. Geol., v. 49, p. 235-293.

James, Harold L., Clark, L. D., Lamey, C. A., and Pettijohn, F. J., 1961, Geology of central Dickinson County, Michigan: U. S. Geol. Survey Prof. Paper 310, 176 p.

Jamison, C. E., 1911, Mineral resources of Wyoming: Wyoming State Geologist's Bull. 1, p. 40.

Kasteler, John I., and Frey, Eugene, 1949, Diamond drilling at the Rambler copper mine, Albany County, Wyoming: U. S. Bur. of Mines, R. I. 4544, 6 p.

Kelley, J. C., 1966, A fortran IV program for the determination of rotation axes from fabric data: Contrib. to Geology, V. 5, No. 1, p. 3944.

Kerr, J. W., and Christie, R. L., 1965, Tectonic history of Boothia uplift and Cornwallis fold belt, Arctic Canada: Am. Assoc. of Petrol. Geologists Bull., v. 49, p. 905-926.

King, James R., 1961, Geology of the Boswell Creek area, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 83 p.

King, John S., 1963, Petrology and structure of the Precambrian and post Mississippian rocks of the northeastern Medicine Bow Mountains, Carbon County, Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 124 p.

Knight, Samuel H., 1917, Age and origin of the red beds of southeastern Wyoming: (Abs.) Geol. Soc. America Bull., v. 28, p. 168-169.

————, 1929, The Fountain and the Casper formations of the Laramie Basin; a study on genesis of sediments: Wyoming Univ. Pub. Sci. Geology, v. 1, 82 p.

————, 1953, Summary of the Cenozoic history of the Medicine Bow Mountains, Wyoming: in Wyo. Geol. Assoc., Guidebook, 8th Ann. Field Conf., 1953, p. 26-28.

Knight, S. H., and Keefer, D. K., 1966, Preliminary report on the Precambrian stromatolites in the Nash formation, Medicine Bow Mountains, Wyoming: Contrib. to Geology, v. 5, p. 1-11.

Koschmann, Albert J., 1960, Mineral paragenesis of Precambrian rocks in the Tenmile Range, Colorado: Geol. Soc. America Bull., v. 71, p. 1357-1370.

Kupfer, Donald H., 1963, Structure of Morton Salt Company mine, Weeks Island Salt dome, Louisiana: Am. Assoc. Petrol. Geologists Bull., v. 46, p. 1460-1467.

Lapadu-Hargues, P., 1953, Sur la composition chimique moyenne des amphibolites: Bull. Soc. Geol. France, Ser. 6, p. 153-173.

Larsen, L. H., Poldervaart, Arie, and Kirchmayer, Martin, 1966, Geologic evolution of the Beartooth Mountains, Montana and Wyoming—Pt. 7 Structural homogeneity of gneisses in the Lonesome Mountain area: Geol. Soc. America Bull. v. 77, p. 1277-1292.

Lees, G. M., 1952, Foreland folding: Quar. Jour. Geol. Soc. London, v. 108, p. 1-34.

Logan, B. W., Rezak, R., and Ginsburg, R. N., 1964, Classification and environmental significance of algal stomatolites: Jour. Geol., v. 72, p. 68-83.

Long, J. S., 1959, Geology of the Phlox Mountain area, Hot Springs and Fremont Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 59 p.

Love, J. L., Weitz, J. L., and Hose, R. K., 1955, Geologic map of Wyoming: U. S. Geol. Survey.

Love, John D., and Keefer, W. R., 1965, Contrasting tectonics of crustal blocks in central and northwestern Wyoming: (Abs.) Am. Assoc. Pet. Geologists Bull., v. 49, p. 114.

Lovering, T. S., and Goddard, E. N., 1950, Geology and ore deposits of the Front Range, Colorado: U. S. Geol. Survey Prof. Paper 223, 319 p.

Mallory, William W., 1960, Outline of Pennsylvanian stratigraphy of Colorado: in Guide to the Geology of Colorado, Rocky Mtn. Assoc. of Geologists, Denver, Colorado, p. 23-33.

Masters, John A., 1952, The Frontier formation of Wyoming: in Wyo. Geol. Assoc. Guidebook, 7th Ann. Field Conf., p. 58-62.

Matus, Irwin, 1958, The geology of the lower French Creek area, Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 38 p.

Maughan, Edwin K., 1963, Mississippian rocks in the Laramie Range, Wyoming, and adjacent areas: U. S. Geol. Survey Prof. Paper 475-C, p. C23-C27.

————, 1964, The Goose Egg formation in the Laramie Range and adjacent parts of southeastern Wyoming: U. S. Geol. Survey Prof. Paper, 501-B., p. B53-B60.

————, 1967, Eastern Wyoming eastern Montana and the Dakotas, in Paleotectonic investigations of the Permian system of the United States: U. S. Geol. Survey Prof. Paper 515, p. 129-156.

Maughan, E. K., and Wilson, R. F., 1960, Pennsylvanian and Permian strata in southern Wyoming and northern Colorado, in Rocky Mtn. Assoc. Geologists, Guide to the geology of Colorado: Denver, p. 34-42.

McCallum, Malcolm E., 1962, Glaciation of Libby Creek Canyon, east flank of Medicine Bow Mountains, southeastern Wyoming: Contrib. to Geology, v. 1, p. 21-30.

————, 1964a, Cataclastic migmatites of the Medicine Bow Mountains, Wyoming: Contrib. to Geol., v. 3, p. 78-89.

————, 1964b, Dedolomitized marble lenses in shear zone tectonites, Medicine Bow Mountains, Wyoming: (Abs.) Geol. Soc. America Spec. Paper 82, p. 129-130.

————, 1964c, Petrology and structure of the Precambian and post-Mississippian rocks of the east-central portions of the Medicine Bow Mountains, Albany and Carbon Counties, Wyoming: unpublished Ph.D. Thesis, Univ. of Wyoming, 164 p.

McCallum, M. E., 1968, The Centennial Ridge gold-platinum district, Albany County, Wyo.: Geol. Survey of Wyo. Prelim. Rept. 7, 13 p.

McCallum, M. E., and Orback, C. J., 1968, The New Rambler copper-gold-platinum district, Albany and Carbon Counties, Wyo.: Geol. Survey of Wyo. Prelim. Rept. 8, 12 p.

McKee, Edwin D., Reynolds, Max A., and Baker, Claud H., Jr., 1962, Experiments on informational recumbent folds in crossbedded sand: U. S. Geol. Survey Prof. Paper 450-D, p. D155-D160.

Mercy, Edward L. P., 1963, The geochemistry of some caledonian granitic and metasedimentary rocks: in the British Caledonides, Oliver and Boyd, Edinburgh, p. 189-216.

Michalek, D. D., 1952, Precambrian geology of Jelm Mountain, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 53 p.

Miller, W. G., 1905, The cobalt-nickel arsenides and silver deposits of Temiskaming: Ontario Dept. of Mines, Ann. Rep., v. 4, pt. 2.

Millgate, M. L., 1965, The Haystack Range, Goshen and Platte Counties: Geol. Survey of Wyoming Prelim. Rept. 5, 9 p.

Moberly, Ralph M., Jr., 1960, Morrison, Cloverly, and Sykes Mountain formations, northern Bighorn Basin, Wyoming and Montana: Geol. Soc. America Bull., v. 71, p. 1137-1176.

Moench, R. H., Harrison, J. E., and Sims, P. K., 1962, Precambrian folding in the Idaho Springs-Central City area, Front Range, Colorado: Geol. Soc. America Bull., v. 73, p. 35-58.

Montagne, John, 1955, Cenozoic history of the Saratoga Valley, Wyoming and Colorado: unpublished Ph.D. Thesis, Univ. of Wyoming, 148 p.

Mook, Charles C., 1916, A study of the Morrison formation: New York Ac. Sci., An. 27, p. 39-191.

Muehlberger, William R., 1959, Internal structure of the Grand Saline salt dome, VanZandt County, Texas: Texas Univ., Bur. Econ. Geology Rept. Inv., No. 38, 18 p.

Myers, William G., 1958, Geology of the Sixmile Gap area, Albany and Carbon Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 74 p.

————, 1958, Geology of the Six Mile Gap area, Carbon County, Wyoming: (Abs.) Geol. Soc. America Bull., v. 69, p. 1738.

Nace, Raymond L., 1936, Summary of the late Cretaceous and early Tertiary stratigraphy of Wyoming: Geol. Survey of Wyo. Bull. 26, 271 p.

Neely, Joseph, 1934, The geology of the north end of the Medicine Bow Mountains, Carbon County, Wyo.: Geol. Survey of Wyo. Bull 25, 15 p.

Newhouse, Walter H., and Hagner, Arthur F., 1949, Cordierite deposits of the Laramie Range, Albany County, Wyoming: Geol. Survey of Wyo. Bull 41, 18 p.

————, 1957, Geologic map of anorthosite areas, southern part of the Laramie Range, Wyo.: U. S. Geol. Survey Mineral Res. map. MF-119.

Noble, J. A., Harder, J. O., and Slaughter, A. L., 1949, Structure of a part of the northern Black Hills and Homestake mine, Lead, South Dakota: Geol. Soc. America Bull., v. 60, p.321-352.

Nockolds, S. R., 1954, Average chemical composition of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007-1032.

Oftedahl, Christofer, 1953, Petrologic reconnaissance in the Precambrian of the western part of the Wind River Mountains, Wyoming: Norsk. Geol. Tidsskrift, B. 32, p. 1-17.

O'Hara, M. J., 1961, Petrology of the Scourie dike, Sutherland: Min. Mag. v. 32, p. 848-865.

————, 1962, Some intrusions in the Lewisian complex near Badcall, Sutherland: Edinburgh Geol. Soc. Tr., v. 19, p. 201-207.

Orback, Clarence J., 1960, Geology of the Fox Creek area, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 100 p.

Osterwald, F. W., 1946, Relation of granitization and mineral facies to structure in northeastern Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 58 p.

————, 1955, Petrology of Precambrian granites in the northern Bighorn Mountains, Wyoming: Jour. Geology, v. 63, p. 310-327.

————, 1959, Structure and petrology of the northern Bighorn Mountains, Wyoming: Geol. Survey of Wyo. Bull. 48, 47 p.

————, 1961, Critical review of some tectonic problems in Cordilleran foreland: Am. Assoc. Petrol. Geologists Bull., v. 45, p. 219-237.

Palmquist, J. C., 1967, Structural analysis of the Horn area, Bighorn Mountains, Wyoming: Geol. Soc. America Bull., v. 78, p. 283-298.

Parker, R. B., 1962, Precambrian agmatites of the Wind River Range, Wyoming: Contrib. to Geology, v. 1, p. 13-19.

Parker, R. B. and Kelley, J. C., 1966, Least-Squares analysis of fabric data: II—Geological examples: Geol. Soc. America Sp. Paper No. 87, p. 298 (Abs.)

Patterson, K. D., 1950, Origin of graphite in western Platte County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 77 p.

Perry, K., 1967, High-grade regional metamorphism of Precambrian gneisses and associated rocks, Paradise Basin quadrangle, Wind River Mountains, Wyoming: (Abs.) Geol. Soc. America Rocky Mtn. Section Program, p. 52-53.

Pettijohn, Francis J., 1942, Basal Huronian conglomerates of Menominee and Calumet districts, Michigan: Jour. Geology, v. 51, p. 387-397.

————, 1957, Sedimentary Rocks, 2nd E.: Harper and Bros., New York City, 718 p.

————, 1963, Chemical composition of sandstones—excluding carbonate and volcanic sands: U. S. Geol. Survey Prof. Paper 440-S, 21 p.

Pipiringos, George N., 1957, Stratigraphy of the Sundance, Nugget and Jelm formations in the Laramie Basin, Wyoming: Geol. Survey of Wyo. Bull. 47, 63 p.

Poldervaart, Arie, 1953, Metamorphism of basaltic rocks: A review: Geol. Soc. America Bull., v. 64, p. 259-274.

Poldervaart, Arie, and Van Backstrom, J. W., 1949, A study of an area at Kakamas (Cape Province): Trans. and Proc. of Geol. of South Africa, Vol. LII.

Potter, P. E., and Pettijohn, F. J., 1963, Paleocurrents and basin analysis: Academic Press, New York City, 296 p.

Prinz, Martin, 1964, Geologic evolution of the Beartooth Mountains, Montana and Wyoming, Part 5—Mafic dike swarms of the southern Beartooth Mountains: Geol. Soc. America Bull., v. 75, p. 1217-1248.

Prucha, John A., Graham, John A., and Nickelsen, Richard P., 1965, Basement-controlled deformation in Wyoming Province of the Rocky Mountain foreland: Am. Assoc. of Petrol. Geologists Bull., v. 49, p. 966-992.

Ramsay, John G., 1963, Structure and metamorphism of the Moine and Lewisian rocks of the northwest Caledonides: in the British Caledonides, Oliver and Boyd, Edinburg, p. 143-170.

Read, H. H., 1958, Donegal granite: Sci. Progr., vol. 46, p. 225-340.

Reed, J. C., Jr., 1963, Structure of Precambrian crystalline rocks in the northern part of Grand Teton National Park, Wyoming: U. S. Geol. Survey Prof. Paper 475-C, C1-C6.

Reid, R. R., Greenwood, W. R., Nord, G. L., and Cook, J. D., 1967, Geology of the St. Joe area, Idaho: Part I, Structural Geology: Geol. Society America Rocky Mtn. Section Program, p. 56.

Richards, Paul W., 1955, Geology of the Bighorn Canyon—Hardin area, Montana and Wyoming: U. S. Geol. Survey Bull., 1026, 93 p.

Richmond, G. M., 1945, Geology of northwest end of the Wind River Mountains, Sublette County, Wyoming: U. S. Geol. Survey Oil and Gas Invs. Prelim. map 31.

Ricketts, Louis D., 1888, Annual report of the Territorial geologist to the Governor of Wyoming, January 1888: Cheyenne, Wyo., 87 p.

————, 1890, Annual report of the Territorial geologist to the Governor of Wyoming, January 1890: Cheyenne, Wyo., 80 p.

Rubey, William W., 1929, Origin of the siliceous Mowry shale of the Black Hills region: U. S. Geol. Survey Prof. Paper 154, p. 153-170.

Ruehr, Ben B., 1961, Geology of the Devils Gate area, Albany and Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 48 p.

Sanders, J. E., 1960, Origin of convoluted laminae: Geol. Mag. (London), v. 97, p. 409-421.

Schoen, Robert, 1953, Geological report on the Cooper Hill mineral district, Carbon County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 40 p.

Sederholm, J. J., 1916, On synantetic minerals: Bull. comm. geol. Finlande, No. 48.

————, 1923, On migmatites and associated Precambrian rocks of southwest Finland, Comm. geol. Finlande Bull. 58.

————, 1926, On migmatites and associated Precambrian rocks of southwest Finland: Comm. geol. Finlande Bull. 77.

Shaw, Alan B., 1953, Preliminary survey of the Foraminifera of the lower shale of the Niobrara formation in the Laramie Basin: in Wyo. Geol. Assoc., Guidebook, 8th Ann. Field Conf., p. 47-55.

Shrock, Robert R., 1948, Sequence in layered rocks: McGraw-Hill Co., New York City, 507 p.

Smith, B. D., 1967, Geology and geophysical investigations of an area of Precambrian rocks, central Laramie Range, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 104 p.

Spencer, A. C., 1904, The copper deposits of the Encampment district, Wyoming: U. S. Geol. Survey Prof. Paper 25, 107 p.

————, 1916, Economic geology of the north Laramie Mountains, Converse and Albany Counties, Wyoming: U. S. Geol. Survey Bull. 626, p. 47-81.

Spencer, E. W., 1958, Structural trends in the Beartooth Mountains, Montana and Wyoming: in Billings Geol. Soc. Guidebook, 9th Ann. Field Conf. p. 16-23.

Stensrud, Howard L., 1963, Geology of the Lake Owens mafic complex, Albany County, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 48 p.

Stevens, Rollin E., Niles, W. W., Chodos, A. A., Filby, R. H., Leininger, R. C., Ahrens, L. H., Fleischer, M., Flanagan, F. J., 1960, Second report on a cooperative investigation of the composition of two silicate rocks: U. S. Geol. Survey Bull. 1113, 126 p.

Steven, Thomas A., 1957, Metamorphism and the origin of granitic rocks, Northgate District, Colorado: U. S. Geol. Survey Prof. Paper 274-M, p. 335-377.

Sutton, Janet, and Watson, John, 1951, The pre-Torridonian metamorphic history of the Loch Torridon and Scourie areas in the northwest highlands, and its bearing on the chronological classification of the Lewisian: Geol. Soc. London, Quart. Jour., v. 106, p. 241-307.

————, 1951, Varying trends in the metamorphism of dolerites, Geol. Mag., v. 88, p. 25-35.

Swetnam, Monte N., 1961, Geology of the Pelton Creek area, Albany and Carbon Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 78 p.

———, 1962, The origin of Precambrian crystalline rocks in the Pelton Creek area, Medicine Bow Mountains, Wyoming: Contrib. to Geol., v. 1, p. 41-47.

Ten Haaf, 1956, Significance of convolute lamination: Geol. Mijnbouw, v. 18, p. 188-194.

Theobald, P. K., Jr., and Thompson, C. E., 1968, Platinum and associated elements at the New Rambler Mine and vicinity, Albany and Carbon Counties, Wyoming: U. S. Geol. Survey Circ. 607, 14 p.

Thomas, Horace D., 1928, Geology and deformation of the Centennial Valley, Wyoming: unpublished M.A. Thesis, Univ. of Wyoming, 32 p.

———, 1934, Phosphoria and Dinwoody tongues in lower Chugwater of central and southeastern Wyoming: Am. Assoc. Petrol. Geologists Bull., v. 18, p. 1655-1697.

———, 1936, Frontier-Niobrara contact in the Laramie Basin, Wyoming: Am. Assoc. Petrol. Geologists Bull., v. 20, p. 1189-1197.

———, 1951, Summary of Paleozoic stratigraphy of the region about Rawlins, Southcentral Wyoming: in Wyo. Geol. Assoc. Guidebook, 6th Ann. Field Conf., p. 32-36.

———, 1957, Geological history and petroleum geology of Wyoming: Oil and Gas Compact Bull., v. 16, p. 13-16.

Thomas, H. D., Thompson, M. L., and Harrison, J. W., 1953, Stratigraphy of the Casper Formation, Pt. 1 of Fusoulinids of the Casper Formation of Wyoming: Geol. Survey of Wyo. Bull. 46, p. 5-14.

Toogood, D. J., 1967, Repeated folding in the Precambrian basement of the Laramie Mountains, Albany and Platte Counties, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 65 p.

Tuttle, Orville F., and Bowen, Norma L., 1958, Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8KAlSi_3O_8$ -SiO_2 -H_2O: Geol. Soc. America Mem. 74, 153 p.

Tweto, Odgen, 1966, Regional features of Precambrian rocks in north-central Colorado: (Abs.) Geol. Soc. America Spec. Paper No. 87, p. 304.

Van Hise, Charles R., and Leith, C. K., 1909, Precambrian geology of North America: U. S. Geol. Survey Bull. 360, 939 p.

Walker, K. R., Joplin, Germaine A., Lovering, J. F., and Green, R., 1960, Metamorphic and metasomatic convergence of basic igneous rocks and lime-magnesia sediments of the Precambrian of northwestern Queensland: Jour. Geol. Soc. Australia, v. 6, p. 149-177.

Warner, Lawrence A., 1956, Tectonics of the Colorado Front Range: in Am. Assoc. Petrol. Geologists Rocky Mtn. Sec., Geological Record, p. 129-144.

Wilcox, Ray, and Poldervaart, Arie, 1958, Metadolerite dike swarms in Bakersville-Roan Mountain area, North Carolina: Geol. Soc. America Bull., v. 59, p. 1323-1368.

Williams, Howell, Turner, Francis J., and Gilbert, Charles M., 1954, Petrography—An introduction to the study of rocks in thin section: W. H. Freeman and Co., San Francisco, Calif. 406 p.

Wilson, Robert W., 1934, a study of the jointing in the Five Springs Creek area east of Kane, Wyoming: Jour. Geol., v. 42, p. 498-522.

Wilson, H. D. N., and Brisbin, W. C., 1962, Tectonics of the Canadian Shield: Royal Soc. Canada Spec. Pub. 4, p. 60-75.

Wiseman, J. D. H., 1934, The central and southwest Highland epidiorites: A study in progressive metamorphism: Geol. Soc. London, Quart. Jour., v. 90, p. 354-417.

Worl, R. G., 1963, Superimposed deformations in Precambrian rocks near South Pass, Wyoming: unpublished M.S. Thesis, Univ. of Wyoming, 53 p.

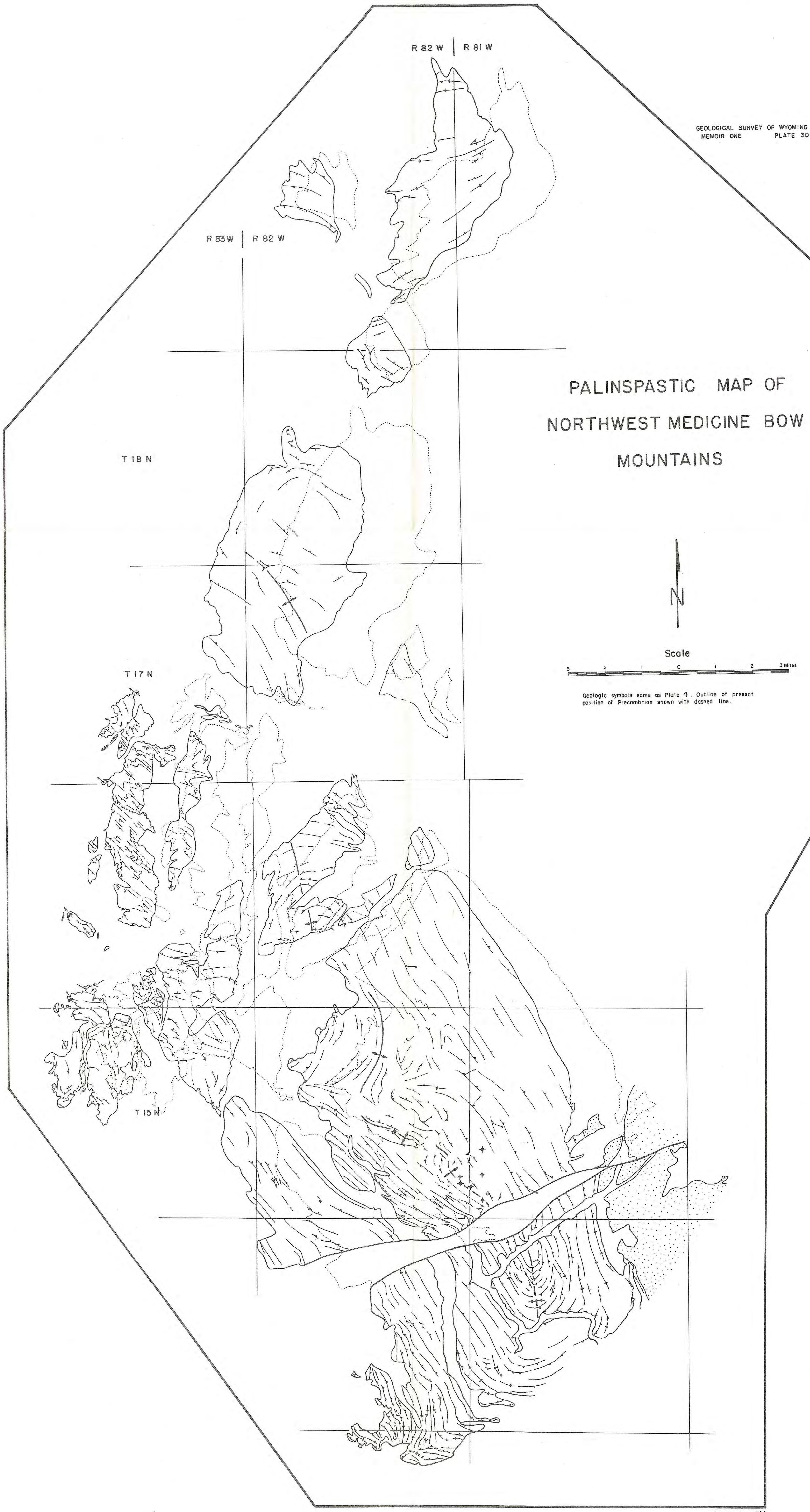
Zangerl, Rainer, 1963, Preliminary results of a restudy of *Corosaurus Alcovensis* Case, the only known New World Northosaur: Contrib. to Geol., v. 2, p. 117-124.

Table 64. Summation of major structural features of domains shown on plate 4.

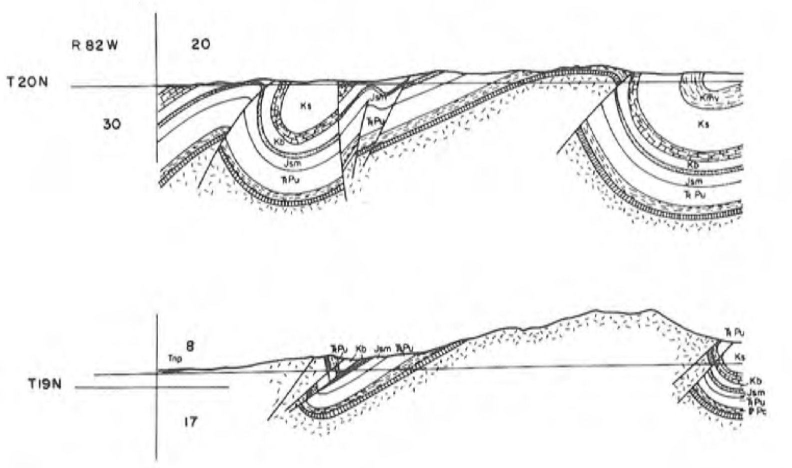
Domain	Trend and Plunge of β	Lineations	Attitude of foliation	Zones of Attenuation and shear zones	Relationship of structure of mafic dikes (amphibolite metadiabase) to attitude of foliation of gneiss
I	–	–	Northeast strike Steep dip.	–	Concordant with foliation of gneiss.
II	–	–	Northwest strike Dip south at high angle Areas of best developed foliation show evidence of cataclasis.	–	Concordant with foliation of gneiss.
III	–	Mineral lineation plunges 14-30 degrees northwest. May parallel axes of antiforms that plunge northwest. At south end of Coad Mtn. lineation plunges southeast at 10-20 degrees.	General trend of foliation northwest. On both Coad and Pennock Mtns. Antiforms plunge northwest at low angle.	Northwest trending zone of penetrative movement at southwest part of Pennock Mtn.	Majority concordant with foliation of gneiss. Some dikes cut foliation approximately at right angles.
IV	–	–	Foliation not well developed. In northeast strike approximately east-west, dip north. In southwest strike of foliation north-northwest, dip variable.	–	–
V	–	–	Foliation in gneiss strikes approximately east, dip 65 north. Cataclasis parallel to gneiss foliation. Bedding in quartzite strikes northwest, dip overturned 70 degrees southwest. Cataclastic foliation in quartzite strikes east, dip 65 north.	–	Mafic bodies concordant to strike of foliation.
VI	–	–	Steeply dipping foliation strikes northwest or northeast.	–	Mafic bodies trend northeast nearly perpendicular to northwest foliation.
VII	–	Both mineral lineation and fold axes present. Minor folds generally plunge northwest or southeast, but show considerable variation in attitude. Mineral lineation is not consistent.	General northwest trend of foliation but considerable variation because of complex folds present in area.	Northwest trending zones common.	–
VIIA	Suggestion of small circle distribution of β. Major concentration N45W plunges 30°. Secondary concentration defines plane which strikes northwest dips steeply south. Small circle distribution may be apparent.	Both mineral lineation and small fold axes. Small folds show variable attitude but most plunge northwest. Mineral lineations not consistent.	General northwest trend of foliation. Dip variable. Considerable variation in strike and dip of foliation because of complex folding.	Northwest trending zones common. Dip of foliation in these zones steeply southwest. One northeast trending zone with foliation dipping northwest.	Most bodies concordant with foliation of gneiss. Appear folded with gneiss.
VIIB	Beta maxima steeply plunging to east-southeast. Maxima may result from intersection of two S surfaces of different ages, one trending north, the other northwest.	–	Suggestion of interference between one set of foliation surfaces that trend north and another that trend northwest.	–	Concordant with foliation of gneiss.
VIIC	Suggestion of small circle distribution of β. Maximum at N45W plunge 45°. Maximum at N5W, plunge 45°.	Both mineral lineation and minor fold axes. Mineral lineations variable in attitude. Minor folds plunge northwest.	Foliation variable in attitude.	Northwest trending. One north trending and one northeast trending shear zone.	Most bodies concordant with foliation of gneiss, but cross-cutting dikes common in southwest.
VIII	–	Mineral lineation plunges northwest.	Northerly trend, steeply dipping.	East trending shear zone.	Most bodies concordant with foliation of gneiss. One dike cuts foliation at close to ninety degrees.
IX	–	–	Northwest strike, dip of foliation northeast averaging 70 degrees.	Northwest trending zones.	Most bodies concordant with foliation of gneiss.
X	–	–	General strike west to west-northwest. Dip variable but steep. Attitude of foliation variable, however, with north to northeast strike common.	–	Most bodies concordant with foliation of gneiss. Some north trending dikes cut foliation at nearly right angles.
XI	Suggestion of small circle distribution of β. May be refolding about new axes.	–	Foliation outlines an antiform that plunges gently north-northwest.	Northeast trending shear zone.	Mafic bodies rare, but those present are concordant with foliation.
XII	–	–	Not mapped because of trespass problems, but probably similar to XI	–	One large mafic body that is concordant with gneissic foliation.
XIII	–	Faint mineral lineations plunge southeast.	Consistent northwest strike of foliation. Dip of foliation northeast.	–	Mafic bodies not well exposed, but seem to be concordant with gneissic foliation.
XIV	Beta diagram reflects the broad flexure of foliation. Based on only 10 foliations.	–	Strike of foliation northeast and dip west. Foliation shows a broad flexure with an axis plunging north 80 west at 60 degrees.	East trending shear zone at northern border of domain.	Both large and small mafic intrusive bodies cut the foliation of the gneiss at large angles.
XV	Suggestion of girdle distribution, but based on only 10 foliations.	–	Foliation shows broad flexture as in XIV, but dips east on eastern border of domain and west on western border. Gently plunging antiform may be present.	–	Small mafic bodies concordant with strike of foliation. Large bodies clearly cross-cutting.
XVI	Beta diagram reflects east trending zone of attenuation in center of domain and slight flexure of northeast trending foliation in north.	Fold axes plunge northwest.	Strike northwest in south dip vertical. Strike northeast in north dip 30-45 degrees to northwest. Foliation in zone of attenuation strikes east and dips vertical.	East trending zone in center of domain.	Small mafic bodies concordant with gneissic foliation.
XVII	Beta diagram reflects flexure in foliation, thus folding of foliation.	Minor folds trend N65W. A mineral lineation plunges southeast at a low angle.	General strike of foliation northwest. In the north a sharp flexure forms a synform with axis plunging N35E.	Zone trending N70E with dip of foliation south between 25-45 degrees.	Mafic bodies concordant with gneissic foliation. Most mafic bodies are in N70E zone of attenuation.
XVIII	Beta diagram suggests gently plunging fold to northwest and southeast.	Mineral lineation plunges between 13 and 22 degrees southeast. Local variation with change in foliation.	General strike of foliation northwest. Dip averages 30-40 degrees east.	–	Mafic bodies appear concordant, but many cut foliation at small angle.
XIX	Beta diagram probably reflects small changes in attitude of northwest trending foliation, but great circle distribution is suggested.	Mineral lineations plunge about 30 degrees southeast.	General strike of foliation northwest. Dip 60 degrees to east.	–	Mafic bodies appear concordant, but some cut foliation of gneiss at a small angle.
XX	Beta diagram suggests gently plunging fold with northwest axis.	–	General strike of foliation northwest. Probable antiform outlined by foliation with steep limb on west.	–	Mafic bodies appear concordant, but cut foliation of gneiss at small angle.
XXI	Great circle distribution of β may reflect isoclinal folding of gneiss.	Minor fold axes plunge northwest or southeast.	General strike of foliation northwest. Dip steeply east.	–	Most mafic bodies are concordant with gneissic foliation. In northwest some bodies cut foliation at nearly right angles.
XXII	Great circle distribution of β may reflect isoclinal folding of gneiss.	Minor fold axes plunge northwest or southeast.	General strike of foliation N30W. Dip steeply to east.	Northeast trending shear zones.	Most mafic bodies are concordant with gneissic foliation. Some small dikes cut foliation at right angles.
XXIII	Beta diagrams suggest complex structure.	Mineral lineations plunge south.	General strike of foliation north-northwest. Antiform in western part of domain. Most dips of foliation are steeply east.	Northwest trending zones of attenuation.	Most mafic bodies are concordant, but some bodies cut foliation at small angle.
XXIV	Well-defined β maxima. Almost vertical plunge.	Mineral lineation plunges south.	Foliation outlines antiform with nearly vertical plunge.	Northwest trending zones of attenuation.	Mafic bodies are concordant with foliation of gneiss.
XXV	Domain is not structurally homogeneous. Major concentration of beta probably reflects southeast plunging fold.	Axes of minor folds plunge northwest at 25 degrees.	Foliation outlines northwest trending antiform with axial plane dipping east.	Northeast trending zones in south.	Some mafic bodies are concordant with foliation of gneiss; others cut the gneiss at an acute angle.
XXVI	Well-defined β maxima. Almost vertical plunge.	–	Foliation outlines antiform with nearly vertical plunge..	Northeast trending shear zones. Northwest trending shear zones in west.	Mafic bodies are both concordant and discordant with foliation of gneiss.

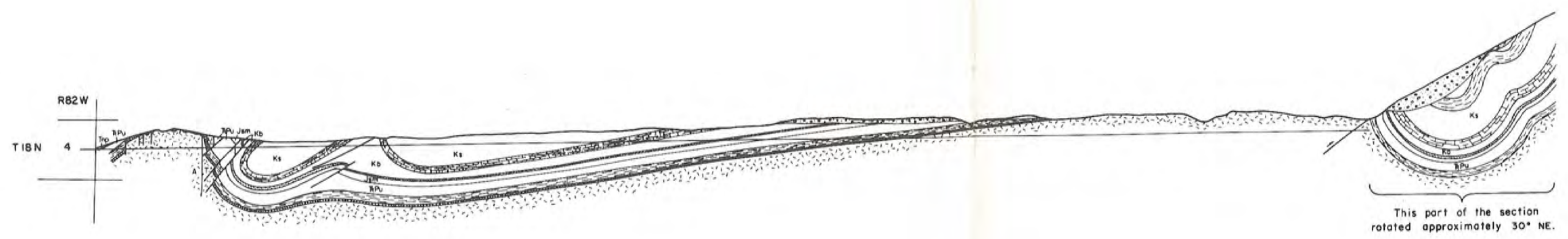
Table 59. Lithology of Post - Precambrian Rocks

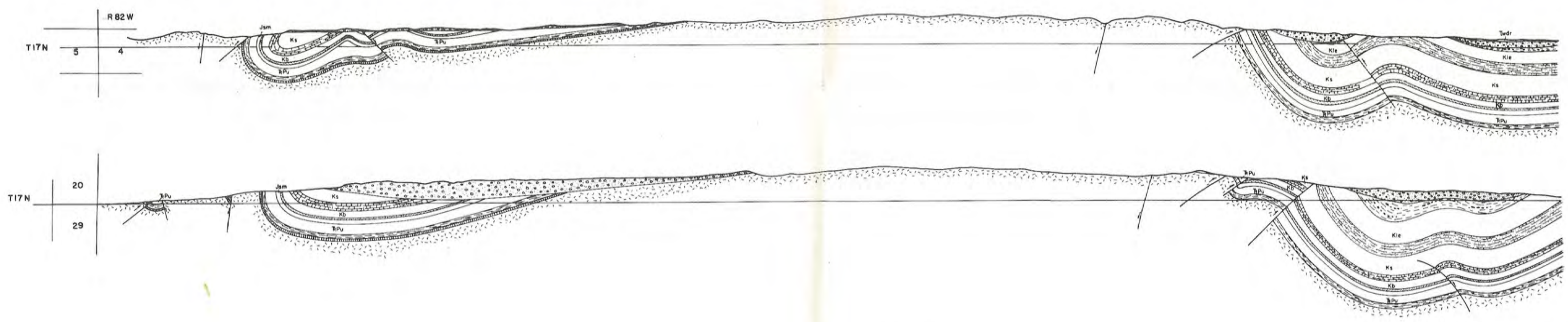
ERA	SYSTEM OR PERIOD	SERIES OR EPOCH	FORMATION	THICKNESS	ENVIRONMENT OF DEPOSITION	REFERENCE
CENOZOIC	Quaternary	Recent	No formal designations - Alluvium, colluvium, terrace gravel, pediment gravel, and lag gravel.	Variable		
		Pleistocene	No formal designations - Locally derived till, outwash, loess, and varved siltstone.	Variable	Glacial	Atwood (1937) McCallum (1962)
	Tertiary	Pliocene ?	No formal designations - Some conglomerates, arkoses, and arkosic sandstones designated as QTu on Plate one may be Pliocene in age.	Variable		
		Miocene	North Park Formation - Light gray fine-grained sandstone, algal limestone, and volcanic claystones with layers of ash. Locally derived calcareous conglomerate and exotic volcanic conglomerate.	Variable Local maximum $\approx$ 300'	Fluviatile and lacustrine	Montagne (1955)
		Oligocene	White River Formation - White to cream calcareous volcanic claystone, locally bentonitic.	Variable; Local maximum $\approx$ 400'	Fluviatile	Knight (1953)
			Major Unconformity			
		Eocene	Wind River Formation - Gray arkosic sandstone interbedded with variegated red, gray, and green claystone.		Fluviatile	Knight (1953) Nace (1936)
		Paleocene	Ferris - Hanna Formations, undivided - Variable sequence of poorly consolidated arkosic sandstone, shale, coal seams, and locally derived conglomerate.	Variable Local maximum $\approx$ 3000'	Coarse facies in alluvial fans and river channels. Fine facies in flood plains and swamps.	
			Major Unconformity			
MESOZOIC	Cretaceous	Late	Medicine Bow Formation - Gray shale, sandstone, brown siltstone, concretionary mudstone with pelecypods, and coal. Locally derived conglomerate 515 feet above base.	Local section incomplete, truncated by erosion. Greater than 600'	Brackish water marine to non-marine swamp and fluviatile.	Knight (1953) Dobbin and others (1929)
			Lewis Shale - Gray to brownish-gray shale, thick interbedded fine-grained gray sandstones, zones of glauconitic ironstone concretions near middle.	2400 - 3000'	Marine	Davidson (1966)
			Mesaverde Formation - White Pine Ridge Sandstone Member at top, interbedded sandstone, coal, and carbonaceous shale.	Ranges from 1400' in the south to 3500' near Elk Mtn.	Near shore marine grading to paralic swamp.	Davis (1966)
			Steele Shale - Gray shale with thin sandstone layers near top.	2700'	Marine	Darton and others (1910)
			Niobrara Formation - Dark gray shale with 3 or 4 beds of chalk in upper part, highly fossiliferous.	670 - 800'	Marine	Thomas (1936) Shaw (1953)
			Frontier Formation - Wall Creek Sandstone Member (1 to 3 thin beds of sandstone) at top, dark gray to black shale, bentonite beds, siderite concretions at bottom.	650'	Marine	Masters (1952)
		Early	Mowry Shale - Black siliceous shale, thin beds of bentonite, fish scales abundant, weathers to silver gray.	120 - 165'	Marine Probably deep water	Rubey (1929) Davis (1963)
			Muddy Sandstone - White fine to medium grained sandstone, some carbonaceous shale.	Discontinuous Local maximum $\approx$ 20'	Barrier bar and beach, near shore marine.	Eicher (1960) Hamilton (1964)
			Thermopolis Shale - Dark gray to black shale, interbedded tan siltstone.	40-100' thinner in south	Tidal flat	Eicher (1960)
			Cloverly Formation - Three fold unit; basal locally conglomeratic sandstone, middle pink siltstone and shale, upper iron stained thin bedded sandstone.	130'	Lower sandstone non-marine fluviatile, middle shale non-marine, upper sandstone marine (?).	Moberly (1960)
	Jurassic	Late	Morrison Formation - Bluish gray or variegated arenaceous claystones, interbedded fine-grained sandstone and limestone, reptile bone fragments common.	280 - 300'	Non-marine flood plain and fluviatile	Mook (1916) Craig and Cadigan (1958) Baker (1965)
			Sundance Formation - White cross-bedded sandstone, glauconitic green shales, red to white shaly sandstone, red to orange siltstones, highly fossiliferous.	280' at Elk Mtn., thins southward to zero near Ring Mtn.	Marine	Pipiringos (1957)
			Major Unconformity			
	Triassic	Late	Jelm Formation - Orange and red siltstone and sandstone, clay pebble conglomerate at base.	130 - 250'	Conglomerate probably continental fluviatile, sandstones marine (?).	Knight (1917) Pipiringos (1957)
		?	Chugwater Formation - Mostly red shale and red siltstone, lower part may contain beds of limestone and evaporite. Includes upper Goose Egg Formation of Burk and Thomas (1956).	Average 700'	Shallow water marine	Thomas (1934) Burk (1953) Campbell (1963)
PALEOZOIC	Permian	?				
		Middle	Goose Egg Formation - Restricted in north, Forelle Limestone and Satanka Shale in south, mostly red siltstone and red shale, thin beds of limestone, gypsum.	Variable Average 130'	Shallow water marine	Burk and Thomas (1956) Maughan (1964)
		Early	Casper Formation - Mostly white cross-bedded sandstone, may include limestone, shale, and calcareous sandstone, variable in age from north to south.	Variable - 208' near Centennial, 550 feet near Arlington.	Near shore marine	Knight (1929) Maughan (1967)
	Pennsylvanian	Late				
		Middle and Late	In the north - Amsden Formation - Deep red shale and sandstone interbedded with pink limestone, deposited on karst topography. In the south - Fountain Formation - Purple arkose, feldspathic sandstone, red siltstone, red shale, locally conglomeratic.	Amsden Formation 150' at Elk Mtn. but grades into Fountain Formation which reaches a maximum of 400' in the south.	Amsden Formation - shallow water marine. Fountain Formation - continental, probably deposited in alluvial fans.	Knight (1929) Maughan and Wilson (1960)
			Major Unconformity			
	Mississippian	Early	Madison Limestone - Light gray cherty limestone, dolomite, thin bed of conglomeratic sandstone at base.	55' at Elk Mtn., removed by pre-Pennsylvanian erosion in south.	Shallow water marine	Maughan (1963)
			Major Unconformity			
PRE-CAMBRIAN			—	—	—	—

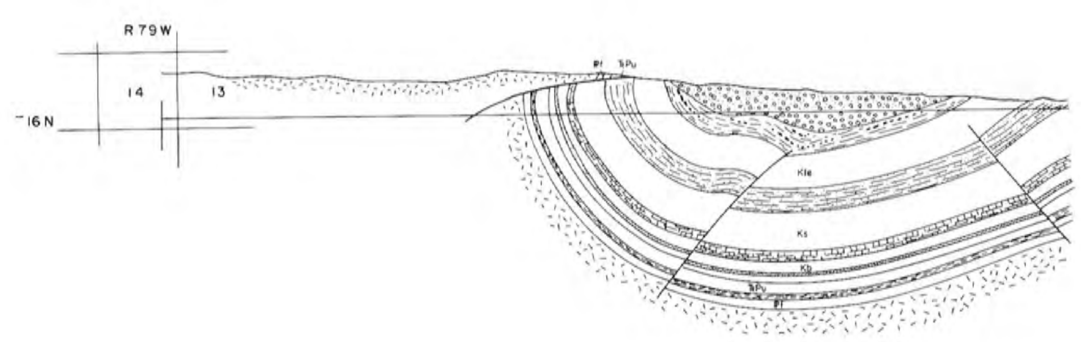


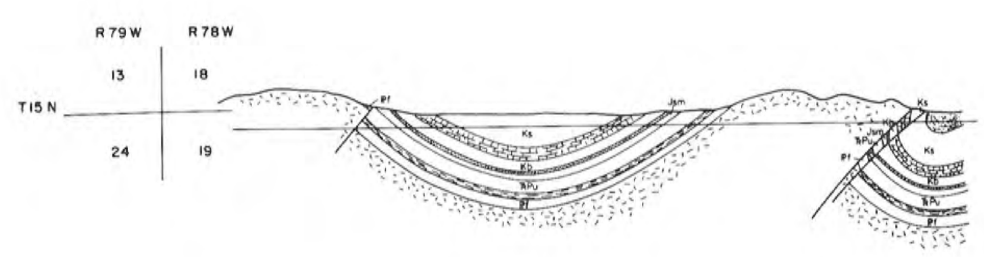
R.S. Houston, 1966

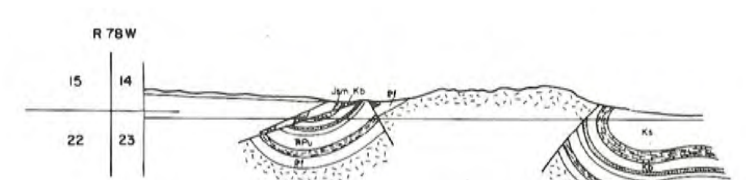


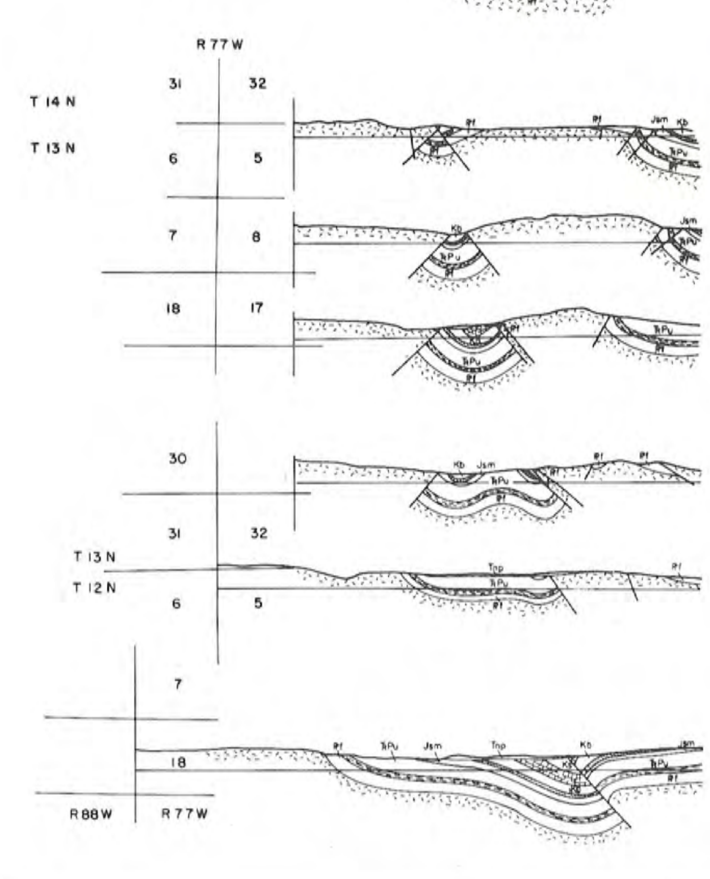












LEGEND

- Tnp — North Park formation
- Twdr — Wind River formation
- Ferris-Hanna formation
- Medicine Bow formation
- Kle — Lewis shale
- Mesaverde formation
- Ks — Steele shale
- Niobrara formation
- Kb — Benton Group
- Cloverly formation
- Jsm — Sundance-Morrison formation
- ₮Pu — Permo-Triassic red beds
- Casper formation
- Pf — Fountain formation
- Madison formation (includes Amsden formation of Pennsylvanian age)
- Precambrian

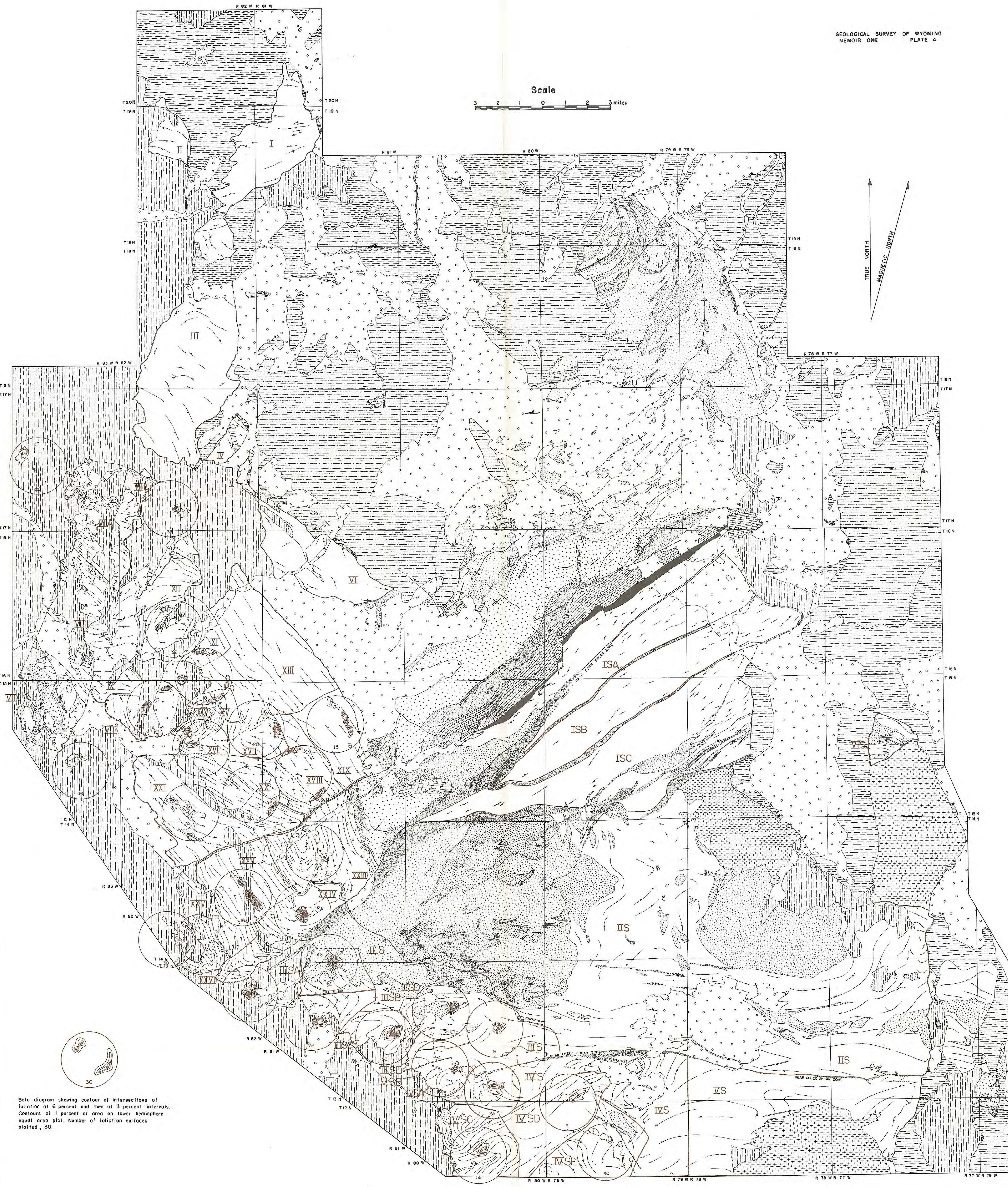
GENERALIZED CROSS-SECTIONS SHOWING LARAMIDE STRUCTURE OF THE MEDICINE BOW MOUNTAINS

Cross-sections rotated about east-west line drawn along line of section. East-west line at 7,000 foot elevations. Geology projected from nearest outcrop. No vertical exaggeration.

3 2 1 0 1 2 3 Miles

R S Houston 1966

GEOLOGICAL SURVEY OF WYOMING
MEMOIR ONE PLATE 4



Beta diagram showing contour of intersections of foliation at 6 percent and then at 3 percent intervals. Contours of 1 percent of area on lower hemisphere equal area plot. Number of foliation surfaces plotted, 30.

LEGEND

DIVISION OF ROCKS YOUNGER THAN PRECAMBRIAN

QUATERNARY-TERTIARY UNDIVIDED
Includes glacial deposit, lag gravel, alluvium, pediment gravel, and terrace deposits.

LATE TERTIARY
Includes North Park formation of Miocene age, White River formation of Oligocene age, and Wind River formation of Eocene age.

EARLY TERTIARY
Includes deposits of Paleocene and early Eocene age; the Ferris-Hanna formations.

PALEOZOIC-MESOZOIC
Includes rocks ranging in age from Mississippian to latest Cretaceous.

DIVISION OF ROCKS OF PRECAMBRIAN AGE

ROCKS NORTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

GABBROIC INTRUSIONS

YOUNGER METASEDIMENTARY ROCKS
Including in order of age; oldest quartzite mainly, Lookout schist, Sugarloaf quartzite, metadolomite, Towner greenstone. All units of the Libby group.

OLDER METASEDIMENTARY ROCKS
Including quartzite, conglomerate, and metavolcanic rocks of the Deep Lake formation.

GRANITE-QUARTZ MONZONITE

FELSIC GNEISS
Including biotite gneiss, quartzofeldspathic gneiss, and augen gneiss.

ROCKS SOUTH OF THE MULLEN CREEK-NASH FORK SHEAR ZONE

SHERMAN GRANITE

OLDER GRANITE

GABBROIC INTRUSIONS

GNEISS
Including hornblende gneiss, quartzofeldspathic gneiss, diopside gneiss, sillimanite gneiss, and marble.

Shear zone

Structural trend showing direction of plunge of lineation.

Structural trend showing vertical dip of foliation.

Generalized trend of foliation showing direction of plunge of antiform.

TECTONIC MAP OF THE MEDICINE BOW MOUNTAINS SHOWING STRUCTURAL SUBDIVISIONS AND BETA DIAGRAMS

EXPLANATION

Tertiary and Quaternary

Alluvium, pediment gravel, and North Park Formation, undivided

Precambrian

Granite Pegmatite
Pink to-white coarse-grained granite pegmatite composed essentially of potash feldspar and quartz; most pegmatites are simple unzoned types, but a few have poorly developed zoning and some contain such minerals as garnet, tourmaline, biotite, muscovite, fluorite, and a complex group of rare earth minerals.

Foliated Granite
Pink to gray medium to coarse-grained foliated granite composed principally of microcline, quartz, and biotite; some of the rocks are massive. This unit also includes some fine-grained dike-like bodies.

Metapyroxenite
Green to purple, highly altered.

Mafic Igneous Rocks
Dark colored mafic igneous rocks varying in composition from olivine gabbro to quartz diorite. Texture and degree of metamorphism vary widely, usually massive, but local facies are well foliated amphibolite.

Quartzo-feldspathic Gneiss
Gray to pink quartzo-feldspathic gneiss may be interlayered with hornblende gneiss, amphibolite, and biotite gneiss. Grades to massive gneiss, only faintly foliated. Massive gneiss shown by short dash.

Hornblende Gneiss and Amphibolite
Dark gray to purple hornblende gneiss and amphibolite. May be interlayered with biotite gneiss and quartzo-feldspathic gneiss.

Geologic Contact
dashed where gradational or inferred.

Shear Zone

30
Minor synform showing bearing and plunge of axis

80
Minor fold or folds in compositional layering or foliation, showing plan of fold and bearing and plunge of fold axis.

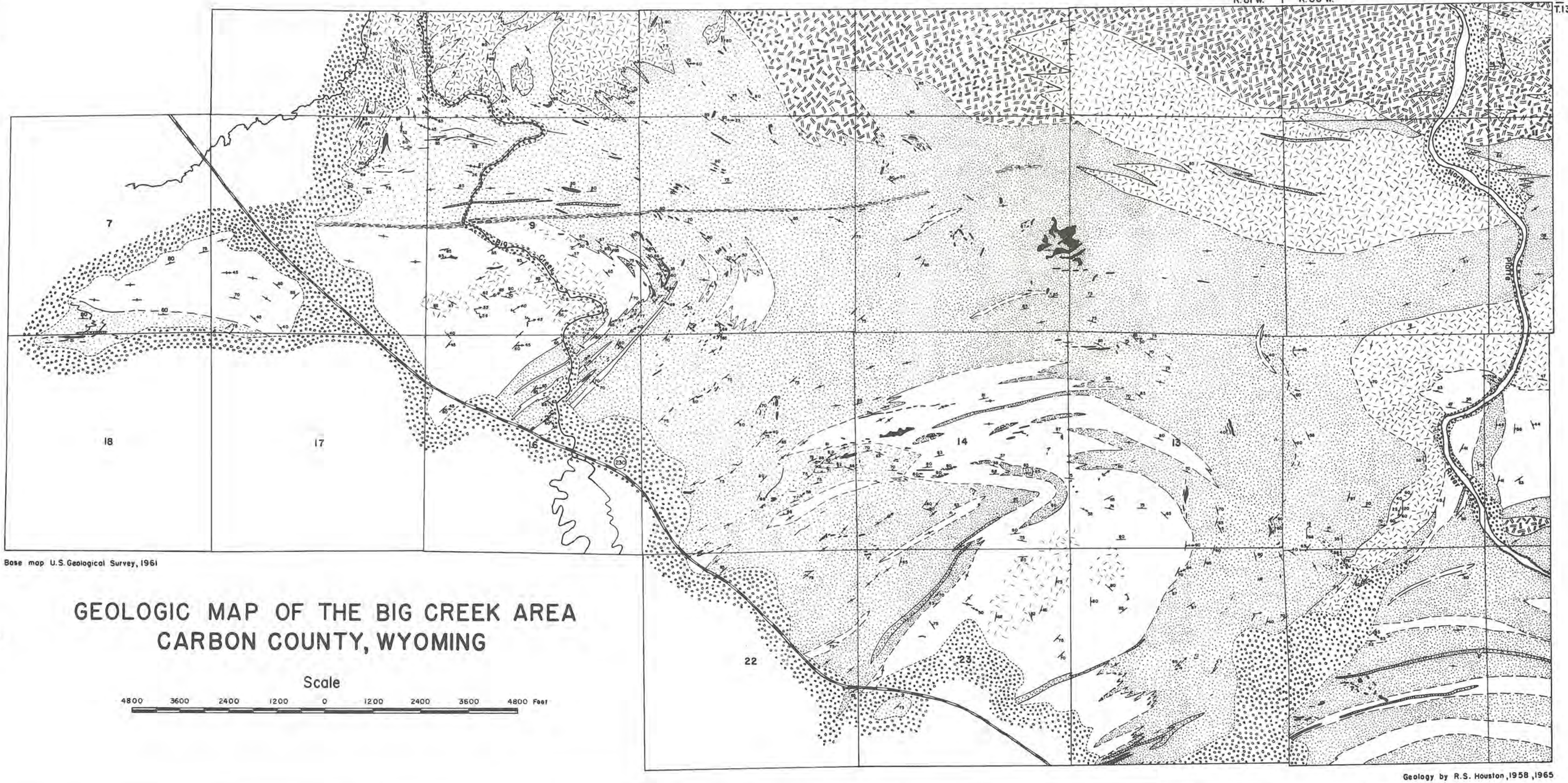
70
Inclined Vertical
Strike and Dip of Foliation
foliation developed by compositional layering in gneiss, usually parallel to orientation of platy minerals. Some foliation is layering caused by cataclasis.

60
Bearing and Plunge of Mineral Lineation
Orientation of amphibole crystals, biotite streaks on plane of foliation. May include lineation resulting from intersection of planes.

75
Bearing and plunge of Mineral Lineation
Where mineral lineation is known to parallel axis of passive fold.

80 67
Strike and dip of foliation and bearing and plunge of lineation

80 71 45
Coexisting minor structural features



Base map U.S. Geological Survey, 1961

GEOLOGIC MAP OF THE BIG CREEK AREA CARBON COUNTY, WYOMING

Scale
4800 3600 2400 1200 0 1200 2400 3600 4800 Feet

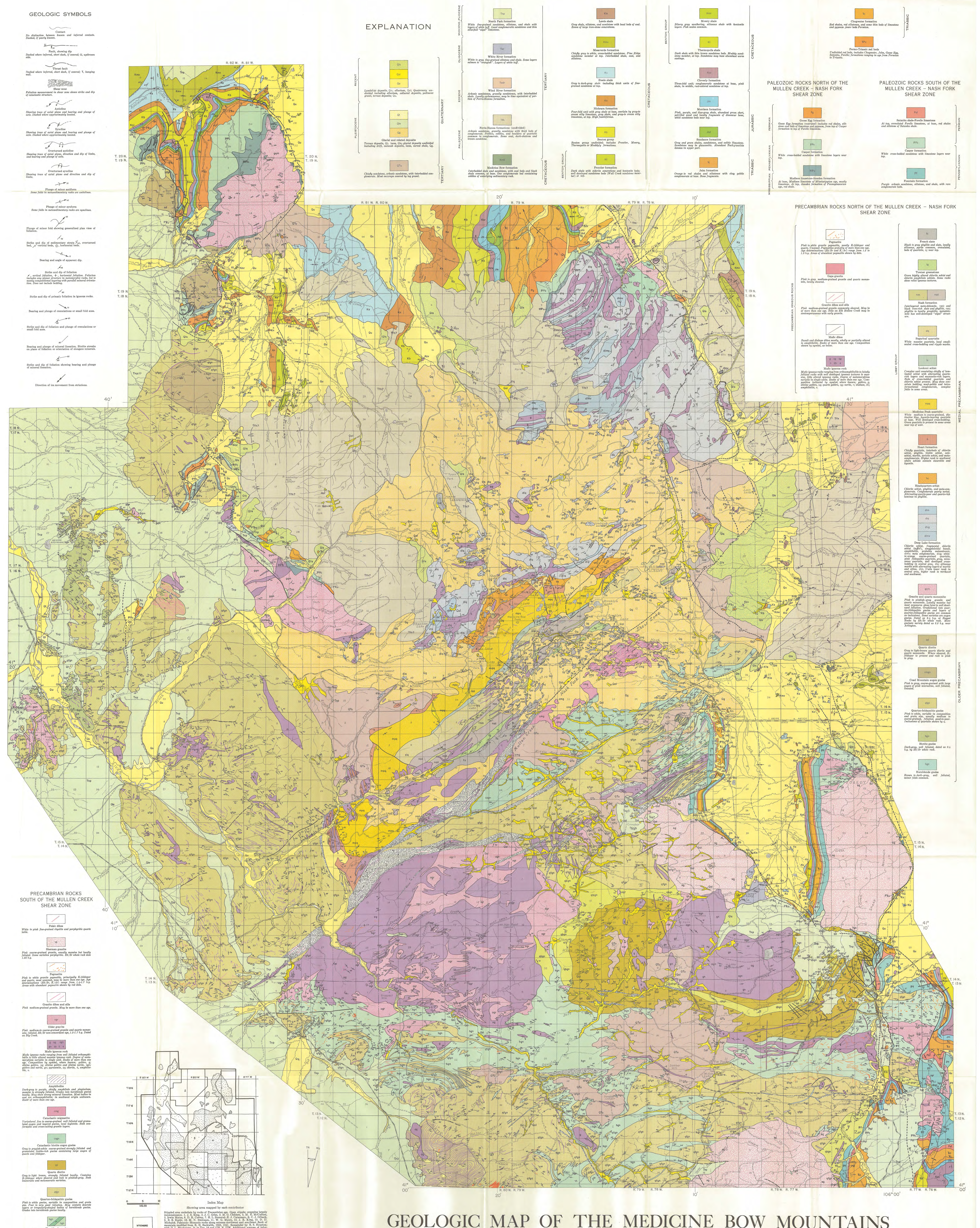
Geology by R.S. Houston, 1958, 1965

SAMPLE LOCATION MAP
ALSO SHOWING METAMORPHIC ZONES IN METASEDIMENTARY ROCKS

Scale

Dots are locations where samples were collected for petrographic study, M-70 sample localities for age determination, C-2 sample localities for chemical analyses.

Memoir One
PLATE ONE, GEOLOGICAL SURVEY OF WYOMING



GEOLOGIC MAP OF THE MEDICINE BOW MOUNTAINS

ALBANY AND CARBON COUNTIES, WYOMING

by

Robert S. Houston

Contributions by M. E. McCallum, J. S. King, B. B. Ruehr, W. G. Myers, C. J. Orback, J. R. King, M. O. Childers, Irwin Matus, D. R. Currey, J. C. Gries, H. L. Stensrud, E. J. Catanzaro, M. N. Swetnam, D. D. Michalek, R. H. Beckwith, D. L. Blackstone, Jr., Eliot Blackwekler, Joseph Neely, S. H. Knight, E. J. Bauer, W. H. Ashley, John Montagne, H. D. Thomas, and Harold Hyden. Age determinations by Allan Hills and Paul Gast. Work done 1957-1965 as cooperative project of Geological Survey of Wyoming and Department of Geology, University of Wyoming.

Williams & Heintz Map Corporation, Wash., D.C. 20027