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VOLUME 1
A SUMMARY OF THE GEOLOGY AND URANIUM POTENTIAL OF PRECAMBRIAN CONGLOMERATES IN SOUTHEASTERN WYOMING

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Laramie, Wyoming
February, 1981

PREPARED FOR U.S. DEPARTMENT OF ENERGY
GRAND JUNCTION OFFICE, COLORADO

This report is a result of work performed by the University of Wyoming through a Bendix Field Engineering Corporation Subcontract, as part of the National Uranium Resource Evaluation. NURE is a program of the U.S. Department of Energy's Grand Junction, Colorado, Office to acquire and compile geologic and other information with which to assess the magnitude and distribution of uranium resources and to determine areas favorable for the occurrence of uranium in the United States.

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GEOLOGICAL SURVEY AND THE GEOLOGICAL SURVEY OF WYOMING

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ABSTRACT

Since 1975 a cooperative effort between the United States Department of Energy, United States Geological Survey, Geological Survey of Wyoming, the Geology Department, University of Wyoming, and various industry groups has resulted in the discovery of a series of uranium-thorium- and gold-bearing conglomerates in Late Archean and Early Proterozoic metasedimentary rocks of southern Wyoming. The mineral deposits were found by applying the time and strata bound model for the origin of uranium-bearing quartz-pebble conglomerates to favorable rock types within a geologic terrane known from prior regional mapping.

No mineral deposits have been discovered that are of current (1981) economic interest, but preliminary resource estimates determined by the Kriging method of ore reserve estimation and an ore-grade cut-off of 100 ppm indicate that over 3418 tons of uranium and over 1996 tons of thorium are present in the Medicine Bow Mountains and that over 440 tons of uranium and 6350 tons of thorium are present in the Sierra Madre. Sampling has been inadequate to determine gold resources. High grade uranium deposits (deposits over 2 meters thick and averaging over 1000 ppm uranium) have not been detected by work to date but local beds of uranium-bearing conglomerate contain as much as 1380 ppm uranium over a thickness of 0.65 meters. This project has involved geologic mapping at scales from 1/6000 to 1/50,000 detailed sampling, and the evaluation of 48 diamond drill holes, but the area is too large to fully establish the economic potential with our present information.

Uranium-thorium-gold bearing quartz-pebble conglomerate is present in the Jack Creek Quartzite of the Late Archean Phantom Lake Metamorphic

Suite and the Magnolia Formation of the Early Proterozoic Deep Lake Group in the Sierra Madre and Medicine Bow Mountains respectively. The best developed and most continuous beds of quartz-pebble conglomerate are in the Jack Creek Quartzite of the northwest Sierra Madre, where radioactive conglomerate crops out in the overturned limb of a major fold for a distance of 13 kilometers. Unfortunately, the well-developed quartz-pebble conglomerates of the Late Archean Jack Creek Quartzite are a thorium rather than a uranium resource with thorium to uranium ratios averaging 6.99. The geochemistry of primary thorium-uranium minerals in the Jack Creek Quartzite indicate that these minerals are mixtures of a thorium-bearing monazite and the thorium silicate huttonite. These minerals along with monazite, zircon, and pyrite are the principal heavy minerals in the Jack Creek Quartzite. The thorium-uranium minerals of the Jack Creek Quartzite are believed to have been derived from 2800 m.y. or older Archean gneiss and granite of the Wyoming Province. Paleocurrent determinations in the Jack Creek Quartzite suggest a northerly source.

In contrast to the Jack Creek Quartzite, the Early Proterozoic quartz-pebble conglomerate of the Magnolia Formation of the Deep Lake Group of the Medicine Bow Mountains contains minerals richer in uranium and has an average thorium to uranium ratio of 2.49. The most promising deposits in the Magnolia Formation are in a three square mile area near the village of Arlington in the northeastern Medicine Bow Mountains. Here uranium-bearing quartz-pebble conglomerate is in the southwest plunging nose and in the complexly folded and overturned northwest limb of a major syncline. The principal uranium minerals in the Arlington locality are uranothorite and coffinite and uranium is also present in lesser amounts in monazite, zircon, and monazite-huttonite mixtures similar to the principal uranium-thorium-bearing minerals of the Sierra Madre. The heavy mineral suite is much more complex in the Magnolia Formation than in the Jack Creek Quartzite and contains minerals such as ilmenorutile and columbite that are almost certainly from a granitic source. Pyrite is the most abundant heavy mineral in both the Medicine Bow and Sierra Madre occurrences. No reliable paleocurrent measurements have been obtained in the quartz-pebble conglomerate beds of the Magnolia Formation but paleocurrent determinations in overlying quartzites suggest a northeast to north source. Inasmuch as the Magnolia Formation is believed to have been deposited after intrusion of uranium-rich Late Archean granites (2500-2600 b.y.) of the Wyoming Province, these granites of central and eastern Wyoming are considered to be the source of uranium-bearing minerals of the quartz-pebble conglomerate. The different source rocks of the Late Archean quartz-pebble conglomerate and Early Proterozoic quartz-pebble conglomerate are critical in that they determine the economic potential of the conglomerate.

In both the Sierra Madre and Medicine Bow Mountains radioactive quartz-pebble conglomerate is either at, or near, the base of a sedimentary succession that lies unconformably on older rocks. The quartz-pebble conglomerate is believed to have been deposited in braided streams and rivers that developed in the vicinity of tectonic (fault-controlled) highlands. Radioactive quartz-pebble conglomerate layers are individual beds or compound beds in coarse-grained quartzite and they occur at different levels in quartzite units that are up to 800 meters thick.

The Jack Creek Quartzite of the Sierra Madre is the basal unit of a Late Archean volcano-sedimentary succession, the Phantom Lake Metamorphic Suite, that has a maximum thickness of 4780 meters. The lower part (Jack Creek Quartzite) of the Phantom Lake Metamorphic Suite is a transgressive succession, which is fluviatile at its base and grades up-section into marine quartzite, schist, and marble; the middle part of the Phantom Lake Metamorphic Suite consists of volcanically-derived graywacke, paraconglomerate, and flows and tuffs of basaltic and andesitic composition; and the upper part of the Phantom Lake Metamorphic Suite is largely quartzite of marine origin. The Phantom Lake Metamorphic Suite is present in both the Sierra Madre and Medicine Bow Mountains, but radioactive quartz-pebble conglomerate is well developed only in the Sierra Madre.

The Phantom Lake Metamorphic Suite rocks are overlain unconformably by the Magnolia Formation which is the basal formation of the Early Proterozoic Deep Lake Group. The Deep Lake Group consists of five formations which are mostly of continental origin in the Medicine Bow Mountains and are mostly of marine origin in the Sierra Madre. The uppermost formation of the Deep Lake Group in the Medicine Bow Mountains and Sierra Madre is the glaciomarine Vagner Formation. Radioactive quartz-pebble conglomerate of the basal Magnolia Formation is best developed in the eastern Medicine Bow Mountains; there are local lenses of radioactive quartz-pebble conglomerate in the basal Magnolia Formation of the Sierra Madre.

A younger Proterozoic rock succession, the Libby Creek Group, is in fault contact with older rocks ranging from Archean Basement to rocks of the Deep Lake Group in both the Medicine Bow Mountains and Sierra Madre. The Libby Creek Group is divided into lower and upper parts which are separated from each other by major faults. The basal formations of the Lower Libby Creek Group, the Rock Knoll Formation and Headquarters Formation are glaciomarine deposits; these beds are overlain by a quartzite-dominated rock succession that is interpreted as various facies of a major delta. The entire Lower Libby Creek Group is well-developed in the Medicine Bow Mountains, but in the Sierra Madre the only beds present are quartzites of the delta facies.

The Upper Libby Creek Group consists of three formations, the Nash Fork Formation (stromatolitic dolomite and graphitic phyllite), Towner Greenstone (marine metavolcanic(?) rocks) and the French Slate (normal marine slate and phyllite). These formations, which are believed to represent a marine carbonate bank and marine deposits laid down well off-shore, are in fault contact with underlying rocks of the lower Libby Creek Group and are thought to have been transported tectonically some distance from their original site of deposition. None of the formations of the Libby Creek Group contain significant uranium, and inasmuch as local beds of hematitic quartzite are in the French Slate and other formations of the Libby Creek Group, it seems probable that uraninite could not have been transported as a detrital mineral when beds of the Libby Creek Group were deposited.

Archean granite cuts the rocks of the Phantom Lake Metamorphic Suite locally, but the most important igneous rocks in the Phantom Lake Metamorphic Suite and the Early Proterozoic Deep Lake and Libby Creek

Groups are sills and dikes that are tholeiitic in composition. These sills and dikes range in composition from ultramafic to mafic and make up 10-15 percent of the area underlain by metasedimentary and metavolcanic rocks.

There is a rough relationship between deformation, metamorphism, and age of the various rock successions. The rocks of the Phantom Lake Metamorphic Suite are isoclinally folded and metamorphosed to amphibolite facies. In both ranges beds are overturned and are generally recumbent in the Sierra Madre. Rocks of the Phantom Lake Metamorphic Suite are also affected by later deformational events that affected younger rocks so that they show evidence of multideformation in most areas. Rocks of the Deep Lake Group are generally less deformed than Phantom Lake Metamorphic Suite rocks, for example, in the central Medicine Bow Mountains they are in broad, open folds. Deformation of the Deep Lake Group rocks is not uniform; in some areas these rocks are more severely deformed and faulted, perhaps as a result of events related to island arc collision with the craton during Early Proterozoic time. Libby Creek Group rocks are in a steeply dipping homocline which may be the northern limb of a major syncline or simply a homoclinal succession, faulted at its southern border. In general, Deep Lake Group rocks and Libby Creek Group rocks are less metamorphosed than rocks of the Phantom Lake Metamorphic Suite, in fact, these rocks are greenschist facies in the central Medicine Bow Mountains, but are higher rank in other areas.

In order to understand the geology and structure of the Proterozoic metasedimentary rocks of the Sierra Madre and Medicine Bow Mountains it is necessary to point out that these rocks terminate abruptly against a major east to northeast striking shear zone located in southeastern Wyoming and that south of this shear zone metamorphic rocks are eugeoclinal facies. We believe that this shear zone is an area where island arcs located in what is now Colorado collided with a rifted margin of the Wyoming Province about 1700 m.y. ago. This collision is the major cause of deformation of the miogeoclinal Proterozoic metasedimentary rocks described above. The Proterozoic metasedimentary rocks were thrust over the older Phantom Lake Metamorphic Suite and Archean basement and finally were rotated to steeply dipping attitudes when the collision occurred. The rock succession is thus telescoped in this area and this factor along with syndepositional rifting accounts for the thick (maximum of 17,000 meters) metasedimentary and metavolcanic succession preserved in southern Wyoming.

The Late Archean-Early Proterozoic rocks of southern Wyoming are strikingly similar to the Thessalon-Livingstone Creek Formations and rocks of the Huronian Supergroup of southern Ontario, Canada. The stratigraphic position of uranium-bearing quartz-pebble conglomerate is about the same in both areas, and we suspect the geologic history is somewhat similar. The Huronian Supergroup contains the economically significant Blind River-Elliot Lake uranium-bearing quartz-pebble conglomerate, and the possibility that a similar deposit might be present in the United States, is of course, exciting. Such an occurrence in Wyoming cannot be ruled out; it was not possible for this investigation to evaluate large areas covered by glacial drift where favorable targets might be present, and drilling programs completed for this report were not adequate to test all possible targets.

However, a reasonable appraisal based on current information is that although the radioactive quartz-pebble conglomerate of the Phantom Lake Metamorphic Suite represents the right depositional environment, it is of the wrong age and source for economic uranium deposits. The radioactive quartz-pebble conglomerate of the Magnolia Formation is the right source and age, but deposits found to date are not as proximal a facies or perhaps not of a high enough energy environment to have economic concentrations of uranium minerals.

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by: A. J. Flurkey, R. S. Houston, K. E. Karlstrom, and A. L. Kratochvil

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By: R.S. Houston and
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INTRODUCTION

This project was undertaken to evaluate the uranium potential of quartz-pebble conglomerates in Early Proterozoic metasedimentary rocks of the Medicine Bow Mountains and Sierra Madre of southeastern Wyoming. This area was chosen for study because regional geologic mapping and geochronologic studies (Houston and others, 1968; Hills and others, 1968) demonstrated that these metasedimentary rocks had many similarities to uranium-bearing metasedimentary rocks in the Blind River-Elliot Lake area of Canada and that they were deposited at a time in the earth's history, and in tectonic and sedimentary environments, favorable for the concentration and preservation of fossil placer uranium deposits (Roscoe, 1969; Pretorius, 1976a, 1976b; Houston and Karlstrom, 1980).

This study includes the following:

(1) detailed geologic mapping of metasedimentary rocks of both the Medicine Bow Mountains and Sierra Madre, an area of about 2000 km^2;

(2) stratigraphic and sedimentological studies with emphasis on units containing radioactive quartz-pebble conglomerates;

(3) regional observation of mesoscopic structures throughout the area and structural analysis of the geometry of major fold systems;

(4) paleocurrent measurements in key stratigraphic units to aid in stratigraphic study and to determine source areas of minerals in radioactive quartz-pebble conglomerates;

(5) regional rock sampling on an approximate one mile or smaller grid and analysis of assays for uranium, thorium, and other selected elements for 2989 surface and subsurface samples;

(6) petrographic and mineralogical studies of selected samples to aid in genetic studies;

(7) geologic support of drilling programs done on favorable horizons. This included: selection and monitoring of Bendix drilling sites; logging and sampling of drill core; and detailed geologic mapping and radiometric surveying of drill site areas. Bendix drilling programs drilled approximately 5932 m (19,448 feet) of rock; and private sector groups provided 8088 m (26517 feet) of core and cuttings.

(8) statistical analysis of surface and subsurface geologic and geochemical data from areas of potential economic interest and determination of uranium and thorium resources.

The above program is a substantial geologic success in that all the objectives were met and the southeastern Wyoming conglomerates now rank as an important new occurrence of Precambrian uranium-bearing conglomerate, among relatively few such occurrences on our planet. However, the program has proven to be only a questionable economic success because grades of uranium, thorium, and gold mineralization are moderate and radioactive zones are relatively thin.

Major sets of radioactive quartz-pebble conglomerate beds were discovered and mapped in both the Sierra Madre and Medicine Bow Mountains. In the Sierra Madre, radioactive conglomerate beds (the Deep Gulch Conglomerate) were discovered at the base of the Jack Creek Quartzite of the Phantom Lake Metamorphic Suite and were traced about 13 km in surface outcrop. Geostatistical work suggests that these beds contain 442 tons of U_3O_8 at a cut-off of 100 ppm and 6353 tons ThO_2 at a cut-off of 100 ppm. The deposits therefore represent a thorium resource but are not a uranium resource for the near future. In the Medicine Bow Mountains, beds of radioactive conglomerate were discovered in both the Rock Mountain Conglomerate of the Phantom Lake Metamorphic Suite and the Magnolia Formation of the Deep Lake Group. Examination of surface outcrops of quartz-pebble conglomerate and information from drilling suggests that radioactive zones in the Rock Mountain Conglomerate are discontinuous, limited in extent, and are of no economic interest. On the other hand, radioactive quartz-pebble conglomerate beds in the Magnolia Formation of the Deep Lake Group have been located in isolated outcrops from Onemile Creek in the northeastern Medicine Bow Mountains to South Brush Creek in the west-central Medicine Bow Mountains. These radioactive conglomerate beds of the Magnolia Formation probably underly an area 35 km long and 5-10 km wide. Within this area the most radioactive conglomerate beds are present in the Onemile Creek area of the northeastern Medicine Bow Mountains, in an area approximately 2.7 km long and one km wide. This area contains conglomerates with up to 1620 ppm U and 1143 ppm Th. Two other areas contain radioactive conglomerates. The Threemile Creek area (EMB-11) has conglomerates with up to 365 ppm U and 344 ppm Th and the South Brush Creek area (MB-9R) has conglomerates with up to 77.8 ppm U and 52 ppm Th. However, in both of these areas the radioactive zones are thin and discontinuous with a few isolated exceptions. Examination of outcrops and limited drilling in other parts of the Medicine Bow Mountains did not reveal uranium- or thorium-bearing beds with more than 100 ppm uranium or thorium. Based on this information, the only area in the Medicine Bow Mountains of potential economic interest is the Onemile Creek area, but other areas underlain by radioactive conglomerates of the Magnolia Formation cannot be rejected entirely because outcrop and drilling data are too limited for economic appraisal.

The Onemile Creek area of the Medicine Bow Mountains is estimated to contain 1801 tons U_3O_8 and 1106 tons ThO_2. The mineralized beds have been mapped in detail on the surface and 10 holes have been drilled to support surface study. Mineralized beds of quartz-pebble conglomerate are believed to be relatively continuous throughout the study area and average about 310 ppm U_3O_8 and 290 ppm ThO_2, but beds or composite beds containing more than 1000 ppm uranium with a thickness of 6 to 10 feet are not present. Although this area has substantial uranium resources that may be exploited in future years, the current low price for uranium (1980) and high costs of mining in a hard rock area of this structural complexity suggests that these reserves will not be exploited in the near future.

Because of the large scope of this project, and the realization that different readers will be interested in different levels of detail, we have chosen to present a somewhat segmented report utilizing as many summarizations as was practical. In fact, the entire first section is entitled: "A summary of the geology and uranium potential of Precambrian conglomerates in southeastern Wyoming." This section is designed to stand alone and to point out the highlights of the southeastern Wyoming conglomerates -- both geological and geochemical. This section combined with the introductory material, should be sufficient for the more casual reader. Sections II and III are more detailed discussions of the geology of the Medicine Bow Mountains and Sierra Madre respectively, which emphasize the radioactive units. Presentation and discussion of results from the drilling program, plus geochemical data are presented in greater detail in a separate volume entitled Volume 2: Drill-hole data, drill-site geology and geochemical data from the study of Precambrian uraniferous conglomerates of the Medicine Bow Mountains and Sierra Madre of southeastern Wyoming. Details of our geostatistical analysis of uranium and thorium resources in southeastern Wyoming conglomerates are presented in Volume 3: Uranium assessment for Precambrian pebble conglomerates in southeastern Wyoming.

HISTORY

Geologic study of the Medicine Bow Mountains and Sierra Madre began prior to the turn of the century with reconnaissance work done by the Territorial Surveys (Hague, 1877). The first mapping and stratigraphic studies were by A. C. Spencer in the Sierra Madre (Spencer, 1904) and by Elliot Blackwelder (1926) in the Medicine Bow Mountains. Both of these were reconnaissance studies, but Spencer's work was a remarkable accomplishment in that he made a regional map of almost the entire Sierra Madre and his report included geologic maps of all important mines of the district as well as a perceptive classification of and review of the genesis of the copper-bearing mineral deposits. Spencer did his work in one field season July-October, 1902.

Blackwelder made several visits to the Medicine Bow Mountains from 1916 to 1925, but his primary work was done in brief (6 weeks) visits during these two seasons. Blackwelder mapped the distribution of metasedimentary rocks in the central Medicine Bow Mountains and made a careful study of the stratigraphy of the major metasedimentary rock sequence, now referred to as the Libby Creek Group. Blackwelder's description of the metasedimentary rocks was detailed and he was the first to recognize the glacial origin of rocks of the Headquarters Formation and the similarity of rocks of the Libby Creek Group to the Huronian Supergroup of the Lake Superior Region.

Significant mining operations in both the Sierra Madre (copper) and Medicine Bow Mountains (gold) ceased in the 1930s and, with the decline in interest in mines and prospects, geologic studies were discontinued. In the early 1950s H.D. Thomas, State Geologist of Wyoming, recognizing that the primary gap in geologic mapping in the State of Wyoming was in the rocks of Precambrian age, proposed to the Wyoming Legislature that funds be appropriated to map the rocks of Precambrian age, in the belief that new mapping might stimulate prospecting for mineral deposits or that knowledge of Precambrian geology might help in understanding the structure of oil-bearing strata of the Wyoming basins. This proposal was approved and geologic mapping began in 1957 under the direction of R. S. Houston. This mapping program was undertaken by Houston and graduate students at the University of Wyoming and some 8000 square kilometers were mapped between 1957-1968. The primary effort was in the Medicine Bow Mountains and Sierra Madre and a geologic report on the Medicine Bow Mountains, which included a map at a scale of 1:63,360, was published in 1968 (Houston and others, 1968). In addition, University of Wyoming theses were available on other Precambrian areas and most thesis maps could be obtained as open-file maps of the Geological Survey of Wyoming. The most detailed maps of the Sierra Madre were those of Ebbett (1970) which were eventually summarized along with other Sierra Madre studies in two reports (Houston and Ebbett, 1977, and Houston and others, 1975).

A companion geochronologic study of the Medicine Bow Mountains was undertaken by F. Allan Hills of the University of Minnesota and Yale University. The results of this study were summarized in two reports (Hills and others, 1968) and Hills and Houston (1979). Further geochronological and geochemical studies of rocks of the Sierra Madre were made by Allan F. Divis of the University of California, San Diego (Divis, 1976, 1977).

The above work established the framework of the Precambrian geology and geochronology of southeastern Wyoming and bracketed the age of thick metasedimentary and metavolcanic successions in the northern part of both the Sierra Madre and Medicine Bow Mountains as younger than about 2700 m.y. and older than about 1700 m.y. As Thomas had anticipated, the geologic studies did lead to renewed interest in the mineral resource potential of the area and exploration programs were undertaken by several mining companies in both ranges.

In the 1970s more detailed geologic and geochemical studies were initiated in the Medicine Bow Mountains under the auspices of the United States Geological Survey. Geologic mapping of four 7½ minute quadrangles was begun as part of a program to evaluate the platinum potential of the southern Medicine Bow Mountains. This new mapping was under the direction of R. S. Houston, University of Wyoming, and M. E. McCallum, Colorado State University -- McCallum had been responsible for a significant part of the geologic mapping of the Medicine Bow Mountains in the 1960s. Of the four geologic quadrangles, Lake Owen is published (Houston and Orback, 1976), Albany is in editorial stages, and mapping is complete for Keystone and Overlook Hill. The United States Geological Survey has also sponsored studies of the Sheep Mountain (Houston, in press) and Savage Run (McCallum, in preparation) areas of the Medicine Bow Mountains, as part of proposed Wilderness Surveys.

At the time of completion of the 1968 report (Houston and others, 1968) on the Medicine Bow Mountains it was known from a number of prior studies (Joubin, 1954; Roscoe, 1957) that Precambrian conglomerates in rocks of about the same age as those in the Medicine Bow Mountains contained economic concentrations of gold and uranium. Furthermore, Houston and others (1968), like Blackwelder before them (Blackwelder, 1935), were impressed by the lithologic resemblance between the rocks of the Deep Lake Group and Libby Creek Group of the Medicine Bow Mountains and metasedimentary rocks of the Huronian Supergroup in Canada -- metasedimentary rocks that contained uranium-bearing quartz-pebble conglomerate. For this reason, Houston and others (1968, p. 159) suggested that the better sorted conglomerates of the Deep Lake Group (the rocks that most closely resembled ore-bearing horizons in Canada) should be examined as a possible source of gold and other heavy minerals. This suggestion received little attention by American geologists (an exploration program of Gulf Oil Company in 1969-1970 is reported to have involved examination of quartz-pebble conglomerate of the northern Medicine Bow Mountains, but geologists were sent to more promising Canadian prospects before the work was completed), but Canadian geologists familiar with the Huronian (Young, 1970) also suggested the correlation. In the early 1970s Stewart Roscoe, of the Geological Survey of Canada, who had done much of the detailed geologic study of the Blind River area in Canada, examined some conglomerates brought to his attention by one of the authors (Houston) and determined that they were slightly radioactive. To our knowledge, Roscoe was the first person to recognize the presence of radioactive conglomerates in southeastern Wyoming.

In the Sierra Madre report of 1975 (Houston and others, 1975) a point was made of the fact that miogeoclinal metasedimentary rocks of the Sierra Madre are lithologically similar to rocks of the Deep Lake Group of the Medicine Bow Mountains. At the time of the Sierra Madre report it was widely accepted that uranium-bearing minerals in Precambrian pyritic, quartz-pebble conglomerates were preserved because of the lower oxygen pressure in the Precambrian atmosphere that existed prior to about 2,000 m.y. (Cloud, 1968; Roscoe, 1973). Houston and others (1975) again emphasized the lithologic resemblance of the Sierra Madre and Medicine Bow miogeoclinal rocks to the Canadian Huronian, and suggested that a more definitive correlation might come from additional isotope studies (at that time the metasedimentary rocks were known to be younger than 2500 m.y. and older than 1700 m.y. so a more closely bracketed age was needed to demonstrate that the metasedimentary rocks were older than the "magic" 2000-2200 m.y. oxygen pressure limit) or from proof that pyritic, quartz-pebble conglomerate that seemed to characterize the Early Proterozoic (2500 m.y.-2000 m.y.) are present in the Wyoming succession.

By 1975 proposals had been made to the United States Geological Survey to undertake geologic mapping programs in both the Medicine Bow Mountains and Sierra Madre specifically designed to search for heavy metal-bearing (uranium, gold, and thorium) conglomerates. These proposals were approved and work was begun in the Sierra Madre as a cooperative program with the Geological Survey of Wyoming in which Forrest Root of the Geological Survey of Wyoming was principal investigator and Paul Graff, graduate student at the University of Wyoming, was responsible for field studies. In the Medicine Bow Mountains funding was obtained by the United States Geological Survey from the Department of Energy to expand a wilderness study in the Snowy Range region of the Medicine Bow Mountains and to emphasize the search for heavy metal-bearing conglomerate. The Medicine Bow study was under the direction of R. S. Houston and the major portions of the field work were done by Karl Karlstrom and Raymond Lanthier, then graduate students at the University of Wyoming. Houston's geologic work has had continued support from the U. S. Geological Survey from 1975 to the present.

These programs were successful in that radioactive quartz-pebble conglomerate was discovered by industry groups in the Onemile Creek area of the Medicine Bow Mountains, in several areas of the central Medicine Bow Mountains by the Wyoming group (Miller and others, 1977; Houston and others, 1977; Karlstrom, 1977; Karlstrom and Houston, 1979a) and in the northwestern Sierra Madre by the Wyoming group (Graff and Houston, 1977; Houston and others, 1979).

In June and July of 1978 the Bendix Field Engineering Corporation, prime contractor for the Grand Junction office of the U. S. Department of Energy, initiated a subcontract (78-222-E) to the University of Wyoming, under the direction of R. S. Houston, to study uranium in quartz-pebble conglomerates. This subcontract was part of the National Uranium Resource Evaluation Program (NURE) and was part of a program (called "Topical Studies," "World Class," and "Special Projects" over the years) to examine the world's most important uranium ore deposits and to assess the favorability of finding such "world class" deposits in the United States.

Specifically, the subcontract called for a summarization, mainly from the literature, of a model for the genesis of Precambrian uranium-bearing conglomerates and a discussion of the southeastern Wyoming occurrences with respect to this genetic model. The results were presented by Houston and Karlstrom (1980).

Discoveries of surface radioactivity in conglomerates were very promising but were not proof of the presence of uranium-bearing conglomerates of the Blind River-Elliot Lake or Witwatersrand type because all outcrops are leached of uranium and pyrite and the best surface samples collected by the Wyoming group contained no more than 150 ppm uranium and 1000 ppm thorium. Numerous cubic cavities could be detected in the conglomerate matrix of the more radioactive samples and the conglomerate was iron-stained, suggesting that pyrite had been leached from the surface and that both pyrite and uranium minerals might be abundant at depth.

It was therefore essential that drilling programs be undertaken to determine if these were indeed pyritic quartz-pebble conglomerates and if they contained enough uranium to be of economic interest. In the summer of 1977, a private company began a drilling program in the Onemile Creek area of the Medicine Bow Mountains, drilling six holes (EMB-1, 2, 3, 4, 5, 6) and intersecting strongly radioactive, pyritic conglomerates in three of them (EMB-1, 2, 6). At about the same time, drilling was undertaken (four holes) by a private sector company (BOWS-1, 2, 3, and GH-1) on their properties in the central Medicine Bow Mountains. Their holes intersected only mildly radioactive conglomerates. At least one hole was also drilled in the central Medicine Bow Mountains by Rioamex, a division of Rio Algom Corporation (who own extensive interests in the Blind River-Elliot Lake deposits) during either the 1977 or 1978 season but the details of this hole are not available to us. In 1978, an additional five holes (EMB-7, 8, 9, 10, 11) were drilled in the Onemile Creek area, intersecting strongly radioactive conglomerate in all of them.

In 1979, Bendix initiated a second subcontract with the University of Wyoming, with principal investigators R. S. Houston, K. E. Karlstrom and L. E. Borgman, to continue geologic work in southeastern Wyoming, to provide geologic support for a helicopter supported, drilling program and to use geostatistical ore-reserve calculation techniques to come up with an estimate of the uranium resources present in Precambrian conglomerates of southeastern Wyoming. The Bendix drilling program had a slightly different emphasis than industry drilling programs. As part of the National Uranium Resource evaluation program, Bendix was charged by the U. S. Department of Energy to assess the potential uranium resources in Precambrian rocks of southeastern Wyoming at grades over 100 ppm U_3O_8 and to formulate and test genetic and exploration models for this type of uranium deposit. This necessitated drilling even mildly radioactive surface rocks to determine if they might contain over 100 ppm U_3O_8 in subsurface. It also required that drilling test out regional unconformities, especially the basal Proterozoic unconformity, which had been identified by geologic mapping. In 1979, Bendix drilled fifteen holes in the Sierra Madre where no subsurface information was yet available, and six holes (MB-4, 5, 9R, 11, 13, 14) in the Medicine Bow Mountains. Of these, 7 holes in the Sierra Madre intersected pyritic and anomalously radioactive conglomerate (SM-1, 1A, 2, 3, 4A, 4B, 9) and one hole in the Medicine Bow Mountains (MB-9R) intersected strongly radioactive conglomerates. The Bendix drilling program went uncompleted in 1979 due to bad weather and several proposed holes were left either undrilled or incompleted.

Private mineral companies were also active in the Sierra Madre during the summer and fall of 1979. They drilled four holes (JP-1, 2, 3, 4) in the Deep Gulch area of the northwest Sierra Madre on land which they had acquired in a winter staking effort shortly after the discoveries of radioactive conglomerates near Dexter Peak and Deep Gulch were announced by Graff and Houston (1977) through the U. S. Geological Survey. All of these holes intersected pyritic conglomerate but uranium contents were generally low.

In the summer of 1980, Bendix completed their drilling program, providing additional funding to the Wyoming group for geologic support of the drilling program and additional time so that geological and geostatistical studies could include results from the 1980 program. One additional hole was drilled in the Sierra Madre (SM-2D), four new holes (MB-10, 15, 16, 17) were drilled in the Medicine Bow Mountains and two uncompleted holes (MB-4, 11) were deepened in the Medicine Bow Mountains. Of these, the hole in the Sierra Madre (SM-2D) and one hole in the Medicine Bow Mountains (MB-16) intersected pyritic and radioactive conglomerates.

Additional data on Precambrian rocks in both mountain ranges were obtained from two deep holes drilled to test for oil and gas. In the winter of 1979, a 4573 foot deep hole was drilled in the northeastern Medicine Bow Mountains spudding in Quaternary deposits which conceal Precambrian rocks of the Vagner Formation in the hanging wall of the Arlington thrust fault and hoping to penetrate the thrust and test Mesozoic reservoirs in the footwall. Drilling was discontinued at 4573 feet in Precambrian quartzites.

Another deep oil and gas test was drilled by C and K Petroleum on the west flank of the Sierra Madre during the summer of 1980. Apparently from seismic interpretations they also hoped to pass through an otherwise undocumented Precambrian thrust sheet and into sedimentary rocks. This hole was spudded in Tertiary deposits, drilled 580 feet of Tertiary sandstone and conglomerate, 1445 feet of Paleozoic and Mesozoic sedimentary rocks (Flat-head through Chugwater) then 3831 feet of Precambrian rock including schists and quartzites of the Singer Peak Formation of the Deep Lake Group and intrusive metagabbro. Cuttings were logged from this hole and down-hole logs were provided.

ACKNOWLEDGEMENTS

This study could not have been completed without the aid of private companies who made available drill core from their properties in the Sierra Madre and Medicine Bow Mountains. We thank in particular, Doug Charlton, of Resource Associates of Alaska, who was especially helpful in discussions of Sierra Madre geology. We also had helpful discussions with Tom Geisck of Rioamex, a division of Rio Algom Corporation. Discussions and field trips with Stuart Roscoe in the early stages of the project and later with Larry Minter

of the Anglo American Corporation of South Africa were also extremely helpful. Communications and field trips with other Bendix subcontractors working on quartz-pebble conglomerate projects helped keep us in touch with the regional ramifications of the project. These included Uinta Arch project -- Paul Graff, Jim Sears, Greg Holden - Black Hills study -- Jack Redden; Raft River project -- Allan Black; southwestern Montana project -- Bob Cohenour, Dick Kopp; Needle Mountain project -- Larry Burns, Noel Tyler, Frank Ethridge-Kingston Peak study -- Donald Carlisle.

Mr. Kenneth Carrico of Rawlins, Wyoming allowed free access to his ranch properties on the northwest margins of the Sierra Madre and permission for Bendix to drill several holes. In addition to trespass privileges, Mr. Carrico was kind enough to give both camping and fishing privileges to the Wyoming group -- a maximum courtesy on Wyoming ranchland. He also arranged for exchange of information between the Wyoming group and C and K Petroleum Inc., who graciously provided us with cuttings and down-hole geophysical logs from an oil and gas test they drilled in Precambrian rocks on the northwestern flank of the Sierra Madre.

In the Onemile Creek area Mr. C. E. Pitcher of Arlington was most hospitable by allowing access through his property to our group as well as many visiting geologists who asked to examine the Onemile Creek occurrences. With few exceptions, property owners in both the Sierra Madre and Medicine Bow Mountains allowed us to study on private property and gave us access to public lands and we acknowledge their courtesy most gratefully.

The United States Forest Service assisted the Wyoming group in many ways such as giving us information on forest road conditions and allowing us access to new Forest Service air photograph coverage of key areas. We are particularly in debt to Ray Urbom of the Forest Service for his help during the course of the study.

This undertaking was a cooperative effort between the World Class Studies Group of the Bendix Field Engineering Corporation (now called Special Projects Group) and the Wyoming group. All non-geological aspects of the drilling program were under the direction of Bendix geologists and engineers. In addition, surface radiometric surveys, down-hole logging, and other specialized services were done by the Bendix support group. Bill Bird, of Bendix, worked with us in the earliest stages of the project; Rex Cole, Hal Gardner, and their supervisor, Bob Young, provided much in the way of cooperation and enthusiasm from 1978-1980 and during the 1979 drilling effort; Nick Theis provided helpful discussions and new insights into the mineralogical characteristics of fossil-placer uranium deposits during the time when he was our project monitor in 1980; Fritz Loomis and John Ludlum were also helpful during this period. Our most recent contacts, Arch Gridley, Jule Anderson, John Ludlam and Ken Karp have also been helpful, both in expediting the 1980 drilling effort and in helping on various aspects of the final report preparation. We also acknowledge Steve Mitchell of Bendix who was helpful in early stages of this project.

The analytical services for this project were handled under D.O.E. funding by the Union Carbide uranium resources evaluation group at Oak Ridge National Laboratory. John W. Arendt and Ron Helgerson of Union Carbide were especially cooperative and helpful in monitoring analytical services and in processing results. Ron Helgerson was particularly helpful in trying to straighten out the numerous cataloging and data processing problems which are the inevitable companion to a sampling project of this magnitude.

Various stages of the Bendix and Wyoming efforts were monitored by members of the Department of Energy. Dave Dahlem and Bill Chenoweth have given useful advice and counsel during the course of the study.

In contrast to some other scientific disciplines where much prior work is abandoned and fully superceded by more modern studies, geologic research, especially geologic mapping, is commonly based upon and certainly greatly benefits from prior work. The foundations for the field work for this study were laid by A. C. Spencer in the Sierra Madre and Eliot Blackwelder in the Medicine Bow Mountains. In general, geologic mapping in both ranges has progressed from reconnaissance to regional to detailed, but we have constantly referred to earlier studies to check and augment the more detailed mapping of the past five years. We have been especially fortunate at Wyoming because manuscript maps, rock samples, and thin sections from most previous mapping in southeast Wyoming are available for study here at the University of Wyoming. Individuals whose work was of great value in this study include McCallum (1964), King (1963), Childers (1957), Lanthier (1978) and Blackstone (1973, 1976) in the Medicine Bow Mountains and Ebbett (1970), Graff (1978), Lackey (1965), Gwinner (1979), Short (1958), Hughes (1973), Weid Ferris (1964) and Miller (1971) in the Sierra Madre. Detailed mapping by Lanthier (1978) in the Snowy Range area of the Medicine Bow Mountains and detailed outcrop mapping by Graff (1978) of quadrangles in the Sierra Madre was the basis for our study of the area mapped by these individuals. We found Graff's and Lanthier's mapping to be of exceptional quality and to require relatively little revision on our part.

Efforts here at the University of Wyoming have involved a large and diverse group. Core logging and sampling activities mainly done by undergraduate students: Tom Schmidt, David Inlow, Peter Swift, John Turner, Ramsey Bentley, Paul Kapp, Robert Spatz, Brooke Cholvin, Brad Pomeroy, Patricia Gallagher, and Baxter Pharr. Geostistical studies involved a number of graduate students: Cassandra Sever, Bill Quimby, Marta Buniak, Moj. Taheri, Mike Andrew, and Randy Hagen. Geochemical work was performed here at the University by Steve Boese. Clerical work was mainly done by Elaine Hertzfeldt.

OVERVIEW: A SUMMARY OF THE GEOLOGY AND URANIUM POTENTIAL OF PRECAMBRIAN CONGLOMERATES IN SOUTHEASTERN WYOMING

PRECAMBRIAN GEOLOGIC SETTING

Precambrian rocks of Wyoming are exposed in the cores of uplifts formed during the Late Cretaceous-Early Tertiary Laramide Orogeny (Figure 1.1). These uplifted blocks were eroded during the Laramide Orogeny and all major uplifts were stripped of their Paleozoic and Mesozoic sedimentary cover exposing a core of Precambrian rocks which were also eroded. The orogeny produced an Early Tertiary topography of north to northwest striking anticlinal uplifts cored by rocks of Precambrian age and synclinal basins floored with Paleozoic-Mesozoic and Early Tertiary sedimentary rocks. The anticlines were topographically high, with relief from 2 to 3 km between basin and uplift.

These anticlinal uplifts are commonly bounded on one or both sides by major thrust faults some of which have been shown to dip at low angles ($30^{\circ}\pm$) and to extend nearly to the base of the crust (Smithson and others, 1978). The displacement of some major thrust faults is therefore substantial and Precambrian rocks exposed on one or both sides of uplifts may be displaced vertical distances of more than 10 kilometers from their basin counterparts in the larger Laramide uplifts. This can be of critical economic importance because it may limit mining in rocks of Precambrian age entirely to thrust hanging walls.

The Early Tertiary topography of north to northwest striking anticlinal uplifts and basins was subsequently buried by continued erosion of the topographically high areas and by aggradation of vast amounts of volcanic debris in the basins. The magnificant mountains and basins of the Early Tertiary were almost entirely buried by the end of the Miocene. Fortunately

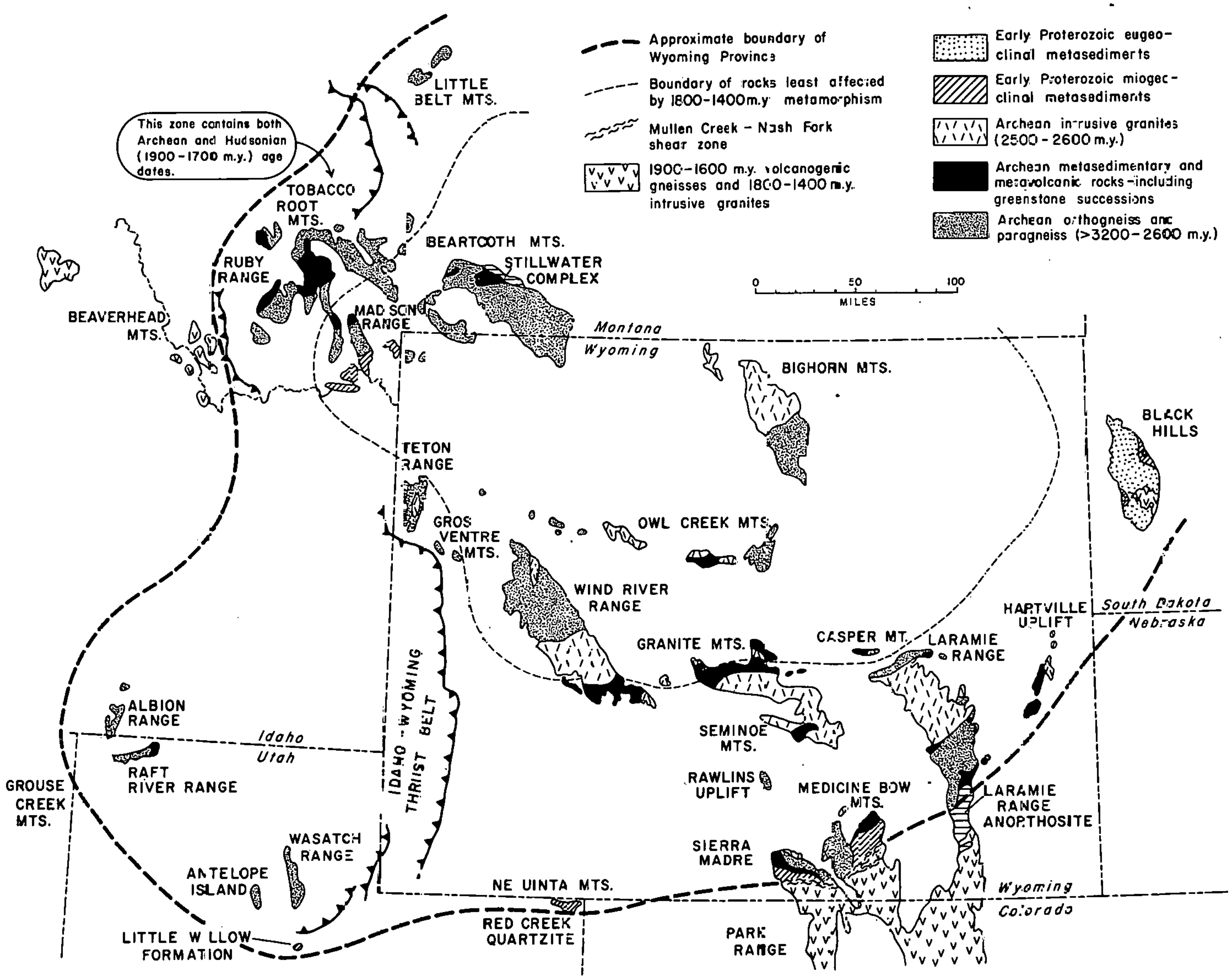


Figure 1.1 Geologic map of the Archean Wyoming Province

for Precambrian geologists, uplift in the Pliocene and Pleistocene, which was accompanied by a decrease in the supply of volcanic debris to this area, resulted in erosion of much of the cover of Tertiary sedimentary rocks, re-exposing the Early Tertiary topography. Today Wyoming is once again an area of topographically high anticlinal uplifts and topographically low synclinal basins, but the relief from valley floor to mountain top (Figure 1.2) is not as great as it must have been in the Early Tertiary.

The Precambrian rocks exposed in the cores of uplifts only constitute about 10-15 percent of the total area in Wyoming because the basins are much larger than the uplifts (Figure 1.1). Therefore Precambrian geologists must make regional interpretations based on limited outcrop areas for which, until recently, there has been only limited geologic mapping. The last 15 years have brought major strides in mapping of rocks of Precambrian age in Wyoming. About one-third of the Precambrian exposures have been mapped at scales of 1:48,000 and reconnaissance mapping has been done in other areas. Much of the work in the Wyoming Precambrian is unpublished but because over 60% has been done by University of Wyoming personnel and because geologists of the United States Geological Survey have made information available for this study we are in a reasonably good position to interpret the regional Precambrian geology.

THE ARCHEAN WYOMING PROVINCE: A DISCUSSION OF POTENTIAL SOURCE ROCKS FOR URANIUM AND THORIUM MINERALS

Since Engel (1963), it has been widely recognized that the majority of the Precambrian rocks in Wyoming are Archean and that the Wyoming

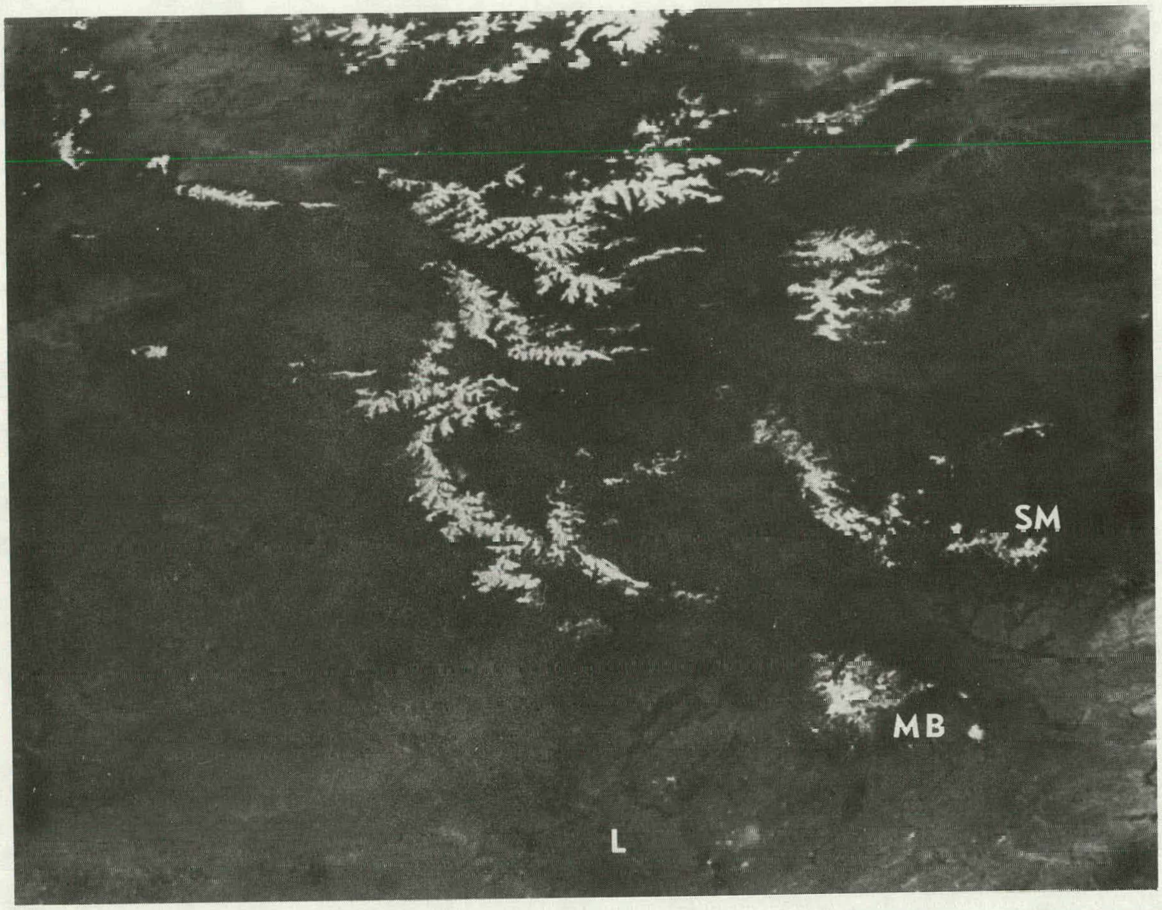


Figure 1.2. Air photograph taken with hand held camera by Skylab astronaut showing uplift and basin topography of central Rocky Mountains. View looking southwest. Lower right one-quarter of photograph is southeastern Wyoming; Laramie Mountains (L) in lower center (not snow covered); Medicine Bow Mountains (MB) lower right (snow covered with two small mountains (Pennock Mountain and Elk Mountain to right); and Sierra Madre (SM) in right center (snow covered and striking at an angle to main mass).

area includes some of the oldest rocks in North America. Engel proposed the name Superior-Wyoming Province for what he considered to be a contiguous Archean shield area containing both the Wyoming rocks and the Superior Province of the Canadian Shield. However, as shown in Figure 1.3 this continuity has not been substantiated by more recent subsurface data in North and South Dakota and southern Canada. Instead, it now appears that Archean rocks in Wyoming are separated from the main mass of Archean rock in the Superior Province by rocks of the Churchill Province which were strongly deformed during the Hudsonian Orogeny, about 1700 m.y. ago (Goldich and others, 1966; Lidiak; 1971, King, 1976, Figure 17). More recent workers (Houston and others, 1968; Houston, 1971; Condie, 1976; Houston and Karlstrom, 1980) have used the name Wyoming Province for the geochronological province in Wyoming and parts of Montana, Idaho, Utah and South Dakota (Figure 1.1) which contains Archean rocks ranging in age from 3300 to 2500 m.y. The Wyoming Province is therefore much like the Slave Province and North Atlantic craton: a continental nucleus within the Precambrian platform of North America which has remained stable for the last 2500 m.y. -- i.e., which has not been affected by tectono-thermal events of strong enough intensity to obscure earlier geologic history by rejuvenating Rb-Sr and U-Pb isotopic systems (Houston and others,1968; Houston and Karlstrom, 1980).

Geochronological studies are not yet detailed enough in the Archean to establish time stratigraphic subdivisions and such subdivisions may never be possible in much of the Archean terrane because of masking of earlier events by later ones. Most of the 3000 m.y. or older dates have been obtained in the central part of the Province -- Beartooth Mountains,

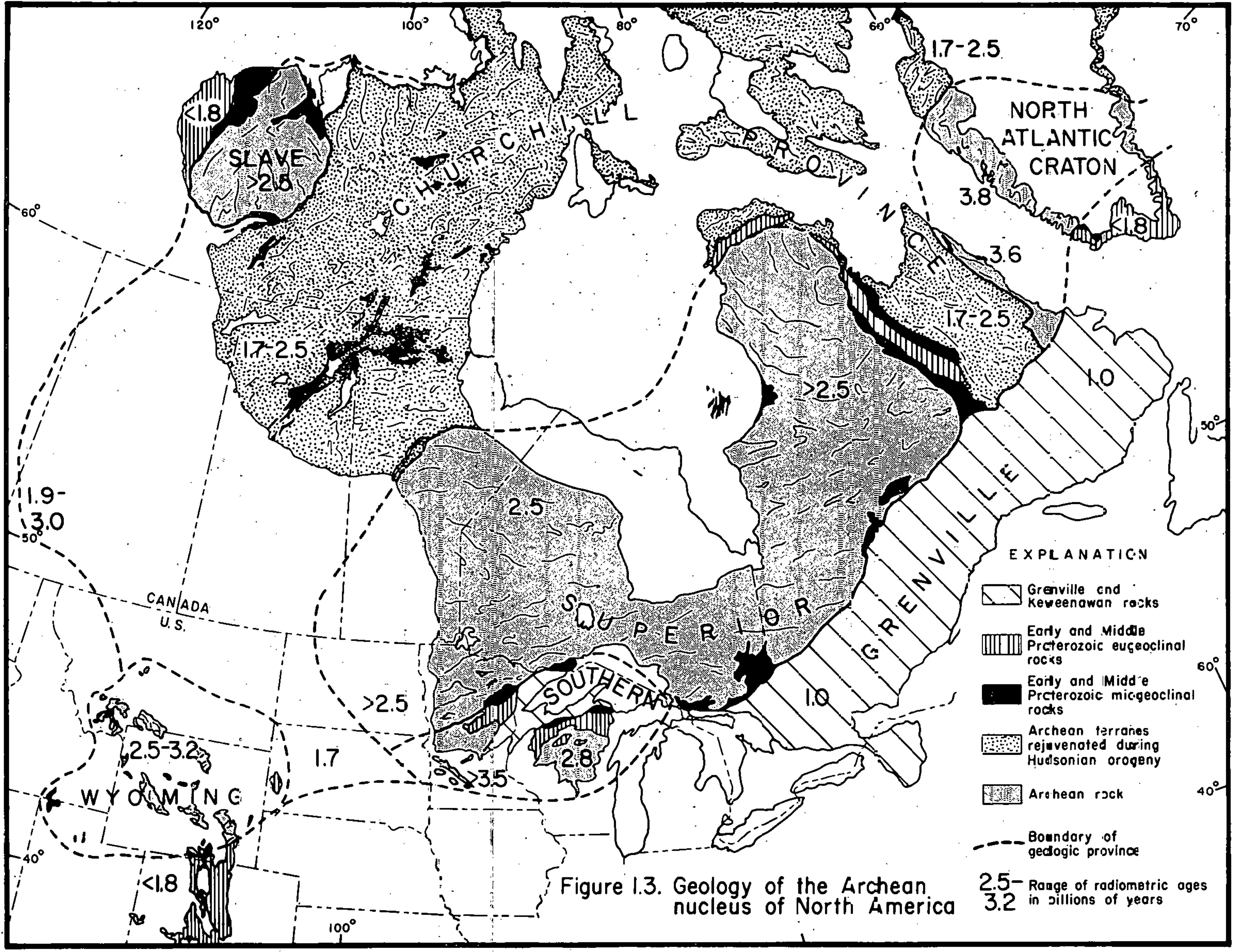


Figure 1.3. Geology of the Archean nucleus of North America

Wind River Range, Bighorn Mountains, and Owl Creek Mountains (Figure 1.1). There are several explanations for this observed distribution: it could be an artifact of still limited geochronological data; it may represent a central part of the province which was least affected by Late Archean and Early Proterozoic orogenies; or it may indeed represent an ancient cratonic nucleus or core zone in the Wyoming Province. The latter two interpretations are not mutually exclusive and we believe both may be true. The second is discussed in later sections; the third interpretation is discussed below.

Ancient Core

The three billion(+) m.y. old rocks in the central part of the Wyoming Province have some characteristics that lend support to the concept that this area is indeed an ancient cratonic nucleus. Like other Archean areas worldwide it is composed of a granitic gneiss terrane with associated greenstone belts. As is the case in virtually all other Archean areas, the age relationships between the gneissic terrane and the dominantly volcanic rocks of the greenstone belts is uncertain. Undoubtedly there is a range of ages for both gneissic rocks and greenstone belts. On geologic evidence alone it appears that the gneissic terrane is older than two of the major greenstone belts, one located in the Atlantic City area of the Wind River Mountains (Bayley and others, 1973) and one located in the Owl Creek Mountains (Thaden, 1980a, 1980b, 1980c, Granath, 1975). However, there are also remnants of very old greenstone belt rocks scattered through the gneissic terrane that may be older than most of the rocks of the gneissic terrane as evidenced by the fact that belt remnants are invaded by migmatite and mobilized gneiss.

Gneissic Terrane

The gneissic terrane deserves the term "basement complex" in that it consists of interlayered paragneisses and orthogneisses of varying composition, agmatite, and migmatite, all of which are quartzo-feldspathic rocks with varying grain size and proportion of mafic minerals such as biotite and amphibole. The quartzo-feldspathic rocks may be complexly folded on mesocopic scales or the strike of foliation may be consistent over large areas. These rocks are predominently white, gray, or grayish pink in color. Rare layers of amphibolite, hornblende gneiss, quartzite, iron formation, marble, and calc-schist may be present in the basement complex either as individual bodies of one rock type or in irregular patches which consist of several rock types. These rocks of the basement complex are cut by diabasic dikes and sills of varying ages. Mafic dikes and sills may be deformed and converted to amphibolite or they may be relatively fresh, exhibit chilled borders, and have diabasic textures.

The basement complex also contains large masses of "granite" which is essentially a more homogeneous phase of the gneissic complex which can be mapped as a separate unit. The "granite" may range in composition from quartz diorite to true granite and exhibits a variety of structural and textural relationships to the gneiss; most commonly the type referred to as catazonal by Buddington (1959).

Clearly this is a complex rock sequence with a long history of igneous activity, sedimentation, volcanism, and deformation. All rocks are intensely deformed and metamorphosed and most are upper amphibolite facies. There are, however, pockets of granulite facies rocks in a number of areas (Worl, 1968, and Worl, R., personal communication, 1980), and we suspect that much of this core area may have been subject to granulite facies metamorphism at some stage in its development.

So far as the writers are aware the Archean core area does not contain rocks with greater than normal primary uranium. We have not observed gneiss or granite to be unusually radioactive in reconnaissance studies done by Wyoming University geologists and no abnormally radioactive rocks are reported by geologists of the United States Geological Survey (Worl, R., personal communication, 1980, on Wind River Mountains). These rocks may well be depleted in K, Rb, Th, and uranium as is typical of basement gneisses older than 3000 m.y. (Tarney, 1976).

Greenstone Belts

Major greenstone belts of the Archean core area are in the southern Wind River Mountains and in the Owl Creek Mountains (Figure 1.1). The greenstone belts include metavolcanic rocks and metasedimentary rocks. In the Wind River Mountains (Bayley and others, 1973) metagraywacke is the most common metasedimentary rock type, but pelitic schists, conglomerate, quartzite and iron formation are also present. Metavolcanic rocks of the Wind River greenstone belt are metabasalt (both pillow basalt and amygdaloidal basalt), metatuff, meta-andesite, as well as hornblende-schist and amphibolite of probable volcanic origin. The base of the greenstone succession in the Wind River Mountains has not been recognized, because the characteristic younger granite intrusion that typically invades greenstone belt margins is present in the form of the Louis Lake Batholith. Bayley and others (1973, p. 6) suggested that the amphibole schist, metagabbro, and serpentinite inclusions in the border phases of the granite where it is in contact with rocks of the greenstone belt may represent mafic volcanics that were the original base of the greenstone belt succession. There are no reliable dates to establish the age of the Wind River greenstone belt succession. The Louis Lake Batholith has been dated as 2680 m.y. by Naylor and others, (1970) so the rocks of the Wind River greenstone belt are older than about 2700 m.y. but we cannot be certain how much older nor do we have evidence to indicate whether the greenstone belt rocks are older or younger than the gneissic terrane of the Wind River Mountains. The great majority of the rocks in the Wind River greenstone belt succession have been metamorphosed to amphibolite grade; only a small area of greenschist facies rocks is preserved in a fault-bounded block in the northern area of outcrop (Bayley and others, 1973, p. 26-27). There is no report of uranium or thorium occurrences in the rocks of the Wind River greenstone belt, but gold-bearing quartz veins and gold-bearing quartz-calcite veins are present in a zone of gabbroic intrusions in the northwestern part of the greenstone belt and these veins contain accessory pyrite, arsenopyrite, and chalcopyrite as well as native gold. Galena and pyrrhotite have also been reported in the veins (Spencer, 1916) but these minerals are apparently uncommon (Bayley and others, 1973, p. 29).

The Owl Creek greenstone belt is less well known than the Wind River greenstone belt. It consists of a northeast-trending body in the north-central Owl Creek Mountains (Houston, 1973, p. 50-51; Thaden, 1980a, 1980b, 1980c) and a similar and probably related body of metasedimentary and metavolcanic rocks in the Wind River Canyon to the west (Granath, 1975). The rocks of the Owl Creek greenstone belt are isoclinally folded, are of amphibolite grade and are engulfed by a younger granite. This deformation, metamorphism, and granite invasion has destroyed the great majority of primary structures in the metasedimentary and metavolcanic rocks making it difficult to establish a stratigraphic section and interpret the structure. Both Gliozzi (1967, p. 14) and Bayley and others (1973, p. 28) have suggested that the greenstone belt rocks may lie on an older gneissic basement. The gneissic basement consists of interlayered quartzo-feldspathic gneiss, amphibolite, amphibole schists, and amphibole gneiss. If Gliozzi's (1967) tentative stratigraphic succession is correct, the lower unit of the greenstone belt is a quartzite and quartz-rich gneiss succession which is overlain in turn by pelitic schists, and by an iron-rich group of rocks that include iron formation, ferro-amphibole schists and amphibolite, quartz-amphibole schists, biotite-garnet schists and a few beds of chert, marble, and pelitic schists. The overall rock succession of the Owl Creek greenstone belt is similar to the lower part of the greenstone belt succession in the Wind River Mountains. An interesting feature of the Owl Creek greenstone belt rock is an absence of rocks that can be positively identified as volcanic. Granath (1975, p. 89) has suggested that massive amphibolites exposed in the Wind River Canyon are probably mafic volcanic rocks but, even here, no primary structures are present. There are numerous mineralized fracture systems in the greenstone belt of the Owl Creek Mountains. Native gold, scheelite, chalcopyrite, galena, bismuthinite, and pyrite have been reported in these mineralized fractures. The area has been extensively prospected and a number of mines were developed that operated over short time spans.

Younger Granites

The Louis Lake batholith that intrudes the northern and western margin of the Wind River greenstone belt cuts both gneissic rocks of the Wind River Mountains and the rocks of the greenstone belt. The batholith contains mainly a gray, foliated, rather uniform granodiorite (Bayley and others, 1973, p. 18) that has several granitic satellite plutons and numerous pegmatite dikes. As reported above, the age of the Louis Lake batholith has been established as about 2700 m.y. (Naylor and others, 1967). No mineralized fractures have been reported in the Louis Lake batholith and the pegmatites are not mineralized although small amounts of beryl, lepidotite, columbite-tantalite and spodumene are present in some. Bayley and others, (1973, p. 31) have suggested that the Louis Lake batholith may be the source of gold and other metallic elements in the vein systems of the greenstone belt, but the lack of mineralization in the satellite stocks and pegmatites of the batholith suggests to us that this is unlikely.

The late granite that cuts greenstone belt rocks of the Owl Creek Mountains and Wind River Canyon is distinctly different from the Louis Lake granodiorite. It is a pink, medium- to coarse-grained alkali-rich granite which is not foliated except where sheared or near contacts with country rocks. It appears to have been a highly mobile body that literally engulfed the greenstone belt (see maps by Thaden, 1980a, 1980b) sending hundreds of veins, stringers, dikes, and satellite pegmatites into the metamorphic rocks of the greenstone belt. From the uranium geology viewpoint this granite is extremely interesting because fracture systems in the granite contain uranophane, uraninite, and coffinite (Yellich and others, 1978). Yellich and others, 1978, p. 35) state that the preferred theory for formation of the uraninite is supergene, but they also state that a potential does exist for hydrothermal deposits. Inasmuch as the uraninite in these veins has been dated as 40 m.y. (personal communication, J. F. Davis, 1980), it appears that the uranium is not Precambrian hydrothermal mineralization related to the granite, but is instead a mineral deposited from ground-water solutions during the Eocene. A key question that has yet to be answered is the source of the uranium -- was it leached from Eocene sedimentary and volcanic rocks or was it derived from the Precambrian granite by leaching of lower grade veins and veinlets? The Owl Creek granite is a legitimate source of uranium despite the fact that no uraninite of Precambrian age has been identified. Stuckless (1979) has determined the thorium, uranium, and lead isotope content of a number of granites in central Wyoming including the Owl Creek Granite. Stuckless (1979) finds that the Owl Creek granite has a uranium content of about 3 to 5 ppm which is about normal for granite but his Pb/U concordia diagrams (Stuckless, 1979, p. 174, Figure 2) show, that if uranium loss during the Cenozoic is considered, the granite was originally anomalously high in uranium. Stuckless (personal communication, 1980) has recently pointed out that the Owl Creek granite is not as uranium-rich as granites of the Granite Mountains in Wyoming but is still anomalous. The Owl Creek granite, therefore, could be the source of uranium deposited in fractures during the Eocene, and thus may also have been a source of uranium minerals in any placer deposit younger than the granite. The age of the Owl Creek granites has not been precisely determined. Minerals from two pegmatites of the Wind River Canyon that <u>may</u> be differentiates of the Owl Creek granite have been dated by Gilletti and Gast (1961) as 2640 m.y. and 2720 m.y. by the Rb/Sr method. These age determinations suggest that the Owl Creek granite is about the same age as the Louis Lake batholith. On geologic grounds we suspect that the Owl Creek granite is intermediate in age between granite of the Louis Lake type and younger granites of central Wyoming that will be discussed below.

A Geochronologic Boundary within the Wyoming Province

Figure 1.1 shows a line, roughly concentric to the Wyoming Province boundary, outside of which feldspar and mica mineral dates, and some whole-rock dates, were reset by the Hudsonian orogeny and subsequent cooling 1800-1400 m.y. ago, and inside of which mineral dates were generally not reset and retain their 2500 m.y. values. In the southern Wyoming Province, where this line is best defined (Peterman and Hildreth, 1978; Hills and Armstrong, 1974), the line trends east-west through the greenstone terranes of the Wind River and Granite Mountains and between Casper Mountain and the Laramie Mountains (Figure 1.1). A somewhat similar geochronologic boundary, separating partially rejuvenated (1700-1800 m.y.) rocks from Archean rocks exists in southwestern Montana (Giletti, 1966).

The major significance of both of these boundaries appears to be related to Proterozoic tectonic events. These aspects are discussed by Giletti (1966), Peterman and Hildreth (1978), Hills and Houston (1979) and Houston and Karlstrom (1980) and are summarized in a later section of this report on Proterozoic tectonics. However, this boundary also approximately concides with the boundary between an older Archean core of the Wyoming Province and probable younger Archean rocks near the margins of the province (Figure 1.1).

Archean Rocks South of the Geochronologic Boundary

South of the geochronologic boundary (and presumably south of the ancient core of the Wyoming Province) but north of the Wyoming Province boundary between Archean and Proterozoic rocks in southern Wyoming (heavy line in Figure 1.1) is a zone of Archean rocks about 200 km wide and over 500 km long which may be part of a somewhat younger Archean margin to the Wyoming Province. The Archean rocks are similar to rocks of the core area in that both areas contain a gneissic terrane with associated greenstone belts and both are cut by younger granites (Figure 1.1). The principal differences between this marginal zone and the Archean core area are:

(1) Late Archean granites of this marginal zone are distinctive radiogenic igneous rocks, anomously high in uranium.

(2) Some Archean metasedimentary and metavolcanic successions within the marginal zone that are in close proximity to the southeastern and eastern margin of the Wyoming Province contain abundant quartzites and other shelf-type sedimentary rocks that are not typical of greenstone belts.

(3) K/Ar ages in the marginal zone record evidence of a severe thermal disturbance approximately 1400-1700 m.y. ago. The disturbed area extends to the southeast limit of the Wyoming Province.

Is this marginal zone of Archean rocks in southern Wyoming one in which Archean rocks are actually younger than those of the core area or do we simply deal with a zone of Archean rocks not significantly different from the core in age, but which have been disturbed during later tectonothermal events?

Gneissic Terrane and Greenstone Belts

First we will consider the character of the gneissic terrane and greenstone belts south of the geochronologic boundary. As stated above, the gneissic terrane does not differ in lithology from that of the core area, but the gneisses of this area may be somewhat younger. Rocks of the gneissic terrane have been dated as 2860± 80 m.y. by Peterman and Hildreth in the Granite Mountains (a metamorphic date -- the gneiss may be 200-300 m.y. older) and 2960± 220 m.y. by Johnson and Hills (1976) for the northern Laramie Mountains. Both of these geochronological studies of the gneissic terrane have given dates of metamorphism so the gneiss protoliths may be older, but, in comparison with the Archean core gneisses which give metamorphic dates of 3100 m.y. (Reid and others, 1975 for Beartooth Mountains) and older than 3000 m.y. (Barker, 1976 for Bighorn Mountains), these gneisses of the Granite Mountains and northern Laramie Range still may be several hundred million years younger than their counterparts to the northwest. We admit that the distinction is not very compelling at this point, especially given the relatively large uncertainties in dating rocks this old. A much better defined age difference can be seen between the gneissic terranes of the central Laramie Range, northern Medicine Bow Mountains, and northeastern Sierra Madre and all of the areas mentioned above. Metamorphic dates of 2535± 45 m.y. for the central Laramie Range (Hills and Armstrong, 1974), 2500± 50 m.y. for the northern Medicine Bow Mountains (Hills and others, 1968) and around 2600 m.y. (Hills and Houston, 1979 and Divis, 1977) for the northeastern Sierra Madre, are about 500-600 m.y. younger than gneisses of the core area, Granite Mountains or northern Laramie Mountains and these southern gneisses are probably younger, but we must again emphasize that these metamorphic dates do not necessarily define the age of the protoliths.

Aside from geochronology, the only other significant difference between gneiss of the Archean core area and gneiss south of the geochronological boundary is metamorphic grade. The Archean gneissic rocks south of the geochronological front are all amphibolite grade and, so far as is known, no granulite facies rocks are present. It is thus possible that these Archean gneisses south of the geochronological boundary may have a higher uranium and thorium content than Archean core area gneisses. However, reconnaissance studies of background radioactivity of these rocks by University of Wyoming geologists has not been adequate to demonstrate such a difference between the Archean core area gneiss and that south of the geochronological front.

As shown in Figure 1.1, greenstone belt rocks are present in the Granite Mountains (Houston, 1973), Casper Mountain (Beckwith, 1939; Burford and others, 1979), Seminoe Mountains (Bishop, 1963; Bayley, 1968), and in the central and northern Laramie Range (Graff and others, personal communication, 1980). The greenstone belt rocks of the Granite Mountains and Casper Mountain are poorly known but contain amphibolitic and ultramafic igneous rocks of probably volcanic parentage, quartzite, iron formation and layered schists that may have been graywacke. No uranium or thorium have been reported in these rocks. The greenstone belt of the Seminoe Mountains has been mapped by Lovering (1929), Bishop (1963) and Bayley (1968) and is thus much better known than the Granite or Casper Mountain greenstone belts. Even so, the stratigraphic succession of the Seminoe greenstone belt remains uncertain. The base of the stratigraphic succession in the Seminoe Mountains is not exposed; the lower most exposed rocks are probably a series of slates and metagraywackes, with interbedded chert, graphite schist, quartzite and thin iron formation, this succession is overlain by a well-developed quartz-magnetite-grunerite iron-formation that may be up to 100 feet thick. The top of the greenstone belt consists of amphibolite, hornblende gneiss and schist of probable volcanic origin. The Seminoe greenstone belt rocks have numerous quartz veins and mineralized fractures that contain chalcopyrite, pyrite, and native gold. Bishop (1963, p. 40) reported that his X-ray emission spectrograph analysis of dump samples detected copper, bismuth, lead, gold, silver, uranium and iron.

About eight miles north of the Seminoe greenstone belt, at a locality on Heath's Peak, there is a large inclusion of metasedimentary and metavolcanic rocks that appears to be a fragment of the Seminoe greenstone belt engulfed by younger granite. The inclusion consists of amphibolite, quartz-muscovite-graphite schist, chlorite schist, calc-silicate rock, and quartzite. This inclusion is of particular interest to us because it is mineralized and contains uraninite that is Precambrian in age (Love, 1970, p. C126). Uraninite occurs in the border phase of the granite and in chlorite-biotite schist of the inclusion which is in contact with granite Harshman and Bell, 1970). The mineralized border also has pyrite, pyrrhotite, molybdenite, marcasite, chalcopyrite, sphalerite, and galena -- pyrite is the most abundant mineral in the area. The quartz-muscovite-graphite schist is also mineralized and contains pyrite, pyrrhotite, marcasite, chalcopyrite, and sphalerite; but it does not contain uraninite. Sulphide minerals are also present in shear zones in the inclusion.

The Heath's Peak occurrence proves that uranium was present in younger granites of this area and was transported and deposited in rocks of associated greenstone belts in Precambrian time. Bishop's analysis of the mineralized fractures of the Seminoe greenstone belt suggests that uranium may also have been deposited in veins and fracture systems. The presence of abundant pyrite in association with the Heath's Peak uraninite may also be significant in terms of reconstructing a source of heavy minerals for later, Proterozoic-type fossil placer deposits.

In the Laramie Range, greenstone belt-type rocks are present both within the gneissic terrane and as large xenoliths within late granite batholiths. In the Esterbrook area of the northern Laramie Range (Spencer, 1916; Segerstrom and others, 1977), greenstone belt-type rocks occur as large xenoliths in the Laramie batholith, a late granitic intrusion of the northern Laramie Range (Condie, 1969). The greenstone belt rocks include amphibolite, hornblende gneiss and schist, fine-grained quartzite, muscovite biotite quartz schist, graphite schist and iron formation. Spencer's (1916, p. 5877) descriptions of mines and prospects of the Esterbrook area demonstrate that there are numerous mineralized zones within the greenstone belt rocks. Essentially we deal with widespread low-grade mineralization consisting primarily of pyrrhotite and pyrite, but with local enrichment of galena, chalcopyrite, and sphalerite -- no significant gold or silver has been reported either by Spencer (1916) or by Segerstrom and others (1977).

Abnormal radioactivity was detected by geologists of the United States Atomic Energy Commission in the Esterbrook-Laramie Peak area during the 1950s

(Guilinger, 1956; Smith, 1954).

According to Guilinger (1956) primary and secondary uranium minerals are in Precambrian igneous and metamorphic rocks of the North Laramie Peak district. The principle uranium occurrences are in shear zones that cut schists near the margins of granite masses. Pitch blevel, uraninite, coffinite, and uranophane were listed by Guilinger (1956, p. 4) as present in the mineralized area. A key occurrence of uranium minerals was at a locality referred to as the Trail Creek Group of claims by Spencer (1916, p. 70-71).

South of the Laramie batholith and north of the Laramie anorthosite (Figure 1.1) greenstone belt-type rocks have been recognized in a number of places and recent reconnaissance studies by Paul J. Graff, James W. Sears, and Greg Holden of Research Associates of Wyoming and by Cassandra K. Sever, graduate student at the University of Wyoming, have given us a better understanding of the geology and uranium mineralization of this area.

Recent mapping by Graff and others (1981), based in part on earlier studies by graduate students of the University of Wyoming (Fields, 1963; Hodge, 1966; Bothner, 1967; Toogood, 1967, and Smith, 1967) demonstrates that a classic greenstone belt succession is present in the area north of the Laramie anorthosite (see Figure 1.1 where eastern one-half of greenstone belt is shown in black). The greenstone belt, referred to as Elmer's Rock greenstone belt, has a basal amphibolite group which consists of mafic and ultramafic amphibolite that grades up-section into mafic amphibolite with local pillow structure. This is overlain by a felsic amphibolite that perhaps was a pile of andesitic flows originally. The basal amphibolite group of Elmer's Rock greenstone belt contains local layers of gneiss (probably metagraywacke) banded iron formation continuous and thin beds of ultramafic rocks that were probably flows and matrix supported conglomerate. Overlying the amphibolite group is a succession of clastic metasedimentary rocks that includes graywacke, pelitic schist, graphite schist, and minor quartzite, marble, and calc-silicate rocks.

The Elmer's Rock greenstone belt is underlain by a granitic gneiss terrane that is believed to have been partly mobilized so that gneissic domes up-welled through the greenstone belt succession and so that mobilized phases of the gneissic terrane invaded the rocks of the greenstone belt. This complex gneissic "basement" in the cores of domes contains early biotite gneisses and gray granite which is invaded by younger pink and red granite. According to Graff (personal communication, 1980) the pink granites are the most radioactive of these various "basement" units and where large masses of pink granite have developed (as on Squaw Mountain) the units give a distinct radioactive anomaly for the area.

Compared with the Esterbrook area there are relatively few mines and prospects in the Elmer's Rock greenstone belt or in underlying granitic gneiss terrane. Most prospects have been for non-metallic mineral deposits such as marble, graphite, and vermiculite, but veins and shear zones containing pyrite and chalcopyrite are present in a number of areas and Wilson (1966, p. 217) reported anomalous radioactivity in graphite schist at one locality in the Cooney Hills. Reconnaissance surveys of the Elmer's Rock greenstone belt by the writers and Alan Hills of the United States Geological Survey (1977-1979) and more detailed examination of the rocks of this area by Graff and others (1981) have demonstrated that there are several units that display above background radioactivity. These include paraconglomerate, coarse-grained quartzite, and the previously mentioned pink granite. None of these units appear to contain radioactive elements in economic quantity.

Greenstone belt-type rocks are also exposed in the vicinity of Garret (black area in southwest corner of Laramie batholith, Figure 1.1), where the rock units are referred to informally as the Potato Creek suite. The Potato Creek suite consists of amphibolite, hornblende gneiss, biotite gneiss and schist, iron formation, calc-schist and quartzite. The greenstone belt rocks are enclosed withn a "basement" of migmatite, agmatite, and granite. The "basement" granitic gneiss is probably older than the greenstone belt-type rocks of the Potato Creek suite, but contacts with the "basement" gneisses were not observed. A preliminary study of uranium and thorium content of the "basement" gneisses, greenstone belt rocks of the Potato Creek suite, and red granite sills, dikes and large intrusives that cut both of the above units was made by Cassandra Sever. Table 1.1 shows that uranium and/or thorium may be locally concentrated in some metasedimentary rocks of the greenstone belt (i.e. metagraywacke and quartzite), but these elements are not present in significant amounts in either the greenstone belt rocks or the gneissic rocks and gray granite that are considered basement. Uranium and thorium are, however, enriched in the young pink granite and are notably enriched in sheared and altered pink granite (Table 1.1). It is also apparent from this study that fractured rocks rich in epidote and biotite are more radioactive than other rocks. This is apparently true of all rocks but fractured and altered pink granite has areas up to 100,000 counts per minute or about 50 times local background.

Greenstone belt-type rocks are also present in southern Wyoming in the northeastern Medicine Bow Mountains, the central Sierra Madre, and in a

TABLE 1.1 GEOCHEMISTRY OF ARCHEAN ROCKS FROM THE POTATO CREEK AREA, LARAMIE MOUNTAINS, WYOMING; ANALYSES BY STEVE BOESE, UNIVERSITY OF WYOMING

Sample No.	Description	% SiO_2	% Al_2O_3	% CaO	% MgO	% TiO_2	% Na_2O	% K_2O	% Fe_2O_3	% MnO	% P_2O_3	Sum	ppm U_3O_8	ppm Th**
GRANITIC ROCKS														
PC-79-2	Gray gneissic granite	75.8	12.8	0.74	0.10	0.1	3.51	4.99	1.27	0.03	0.04	99.3	7.1	11
PC-79-7	Sheared pink granite sill, epidote-enriched	68.5	14.1	5.31	0.16	0.1	3.05	3.29	3.35	0.06	0.12	98.1	34.1	40
PC-79-9	Fine-grained pink granite sill	75.9	12.2	0.87	0.19	0.2	2.71	5.83	1.34	0.04	0.03	99.3	12.3	11
PC-79-10A	Red (batholithic) granite, very radioactive	68.6	8.7	2.41	0.32	0.4	2.45	3.19	11.5	0.05	1.52	99.1	75	535**
PC-79-10B	Red granite	71.7	8.8	4.87	0.31	0.2	2.72	2.61	3.52	0.04	3.55	98.3	96	250**
PC-79-10C	Red granite (altered)	46.4	10.2	0.42	0.13	1.0	2.54	4.23	35.0	0.03	0.10	100.0	18.4	41
PC-79-12	Gray gneissic granite	73.8	13.5	1.81	0.31	0.3	4.79	2.83	1.66	0.04	0.06	99.1	1.6	5
METASEDIMENTARY AND METAVOLCANIC ROCKS														
PC-79-1	Metabasalt	50.9	7.4	14.5	14.8	0.6	0.53	0.19	11.1	0.20	0.04	100.3	0.2	15
PC-79-3	Layered amphibolite	52.3	13.3	10.3	5.70	1.1	1.83	0.83	14.9	0.23	0.10	100.6	0.9	16
PC-79-4	Epidote quartzite	91.5	3.1	2.07	0.04	0.1	0.48	0.07	1.38	0.02	0.02	98.8	3.5	5
PC-79-5	Quartzite	66.8	13.7	11.9	0.13	0.1	0.33	0.09	4.29	0.03	0.02	97.3	13.4	7
PC-79-6	Iron formation	52.7	1.9	2.18	1.33	0.1	0.18	0.18	43.7	0.06	0.07	102.5	4.2	4
PC-79-8	Sheared biotite gneiss	63.1	15.9	3.00	1.81	0.8	4.08	2.36	7.6	0.07	0.11	98.8	113	43
PC-79-11	Biotite gneiss (metagraywacke?)	70.5	12.5	2.50	0.89	0.8	4.24	1.46	5.01	0.08	0.17	98.2	4.2	15

*Thorium values are accurate to ±5 ppm except those distinguished by ** which are within ±25 ppm.

number of areas in the western Medicine Bow Mountains and eastern Sierra Madre. One of the largest masses of greenstone belt-type rocks in southern Wyoming is in the Arlington area of the eastern Medicine Bow Mountains where eight square km are underlain by amphibolite, hornblende gneiss and schist, lesser amounts of felsic gneiss, quartzite, and paraconglomerate. These rocks are referred to as the Stage Crossing Gneiss. They are in fault contact with younger(?) metasedimentary rocks and are cut by granitic rocks ranging from tonalite to pink granite.

A similar, and probably correlative, assemblage is found in a 10-15 square km area in cirques north of the continental divide in the central Sierra Madre. These rocks, called the Continental Divide Metavolcanic Rocks, include amphibolite, garnet amphibole schist, hornblende gneiss, quartzite, marble, and two outcrops of beautifully preserved pillow basalt (Figure 1.4). The Continental Divide assemblage is intruded by a gray tonalite to (granodiorite), shown in Figure 1.5, which has yielded a date of 2700 m.y. (Carl Hedge, personal communication, 1979) and appears to be conformable with underlying granitic paragneisses and overlying quartzites of the Phantom Lake Metamorphic Suite.

Remnants of metasedimentary and metavolcanic rocks are present in small masses within a dominantly granitic gneiss terrane in a number of areas of the western Medicine Bow Mountains and eastern Sierra Madre. These rocks range from individual layers of quartzite or marble in felsic gneisses to areas with enough amphibolite, hornblende gneiss, paragneiss, quartzite, marble, and various mafic schists to be separated as a metasedimentary-metavolcanic suite in geologic mapping (Plate 5).

The relationship between these metasedimentary and metavolcanic rocks and the much more widespread Archean granitic gneiss terrane is generally

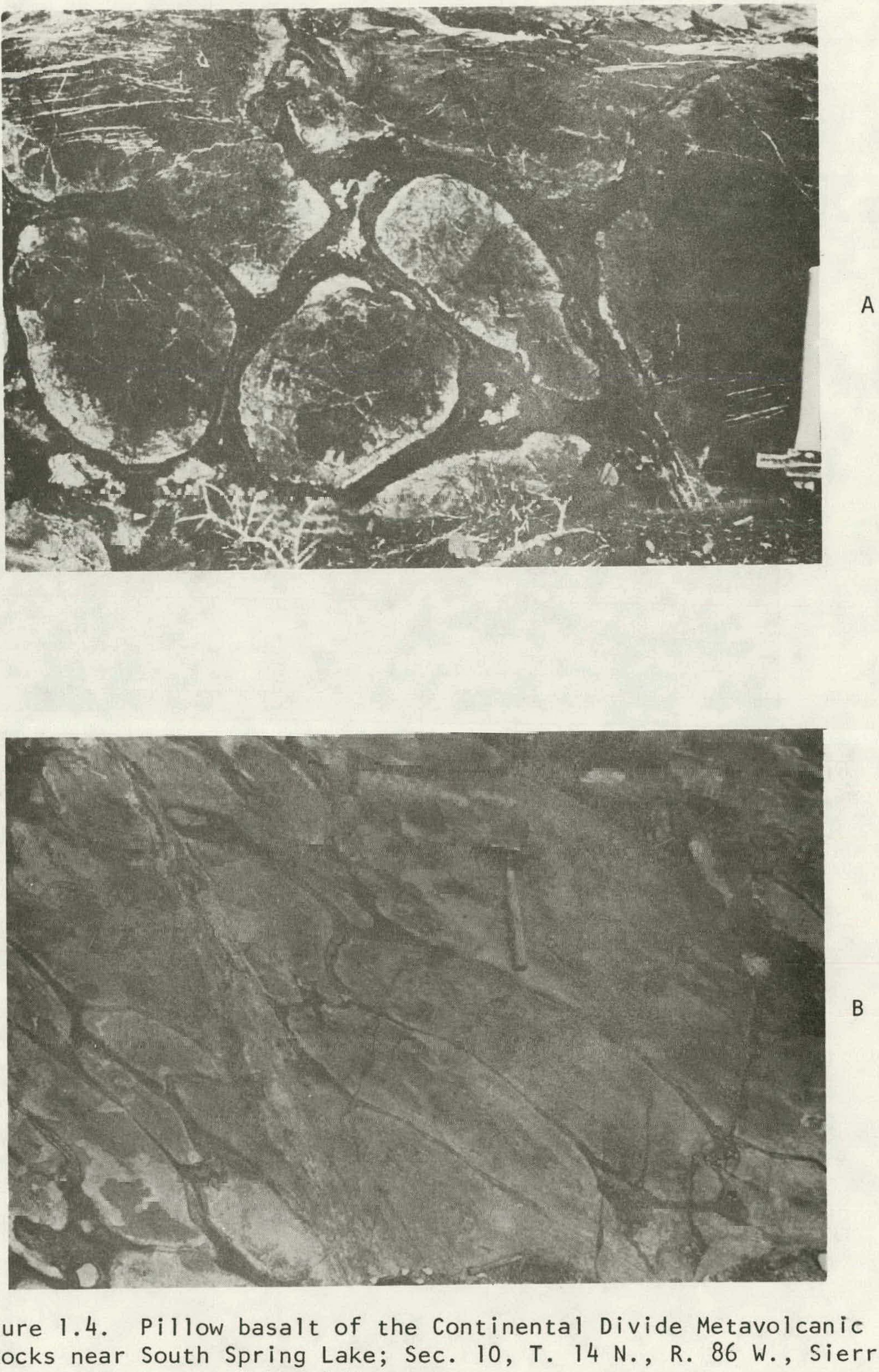


Figure 1.4. Pillow basalt of the Continental Divide Metavolcanic Rocks near South Spring Lake; Sec. 10, T. 14 N., R. 86 W., Sierra Madre. Outcrops right side up.

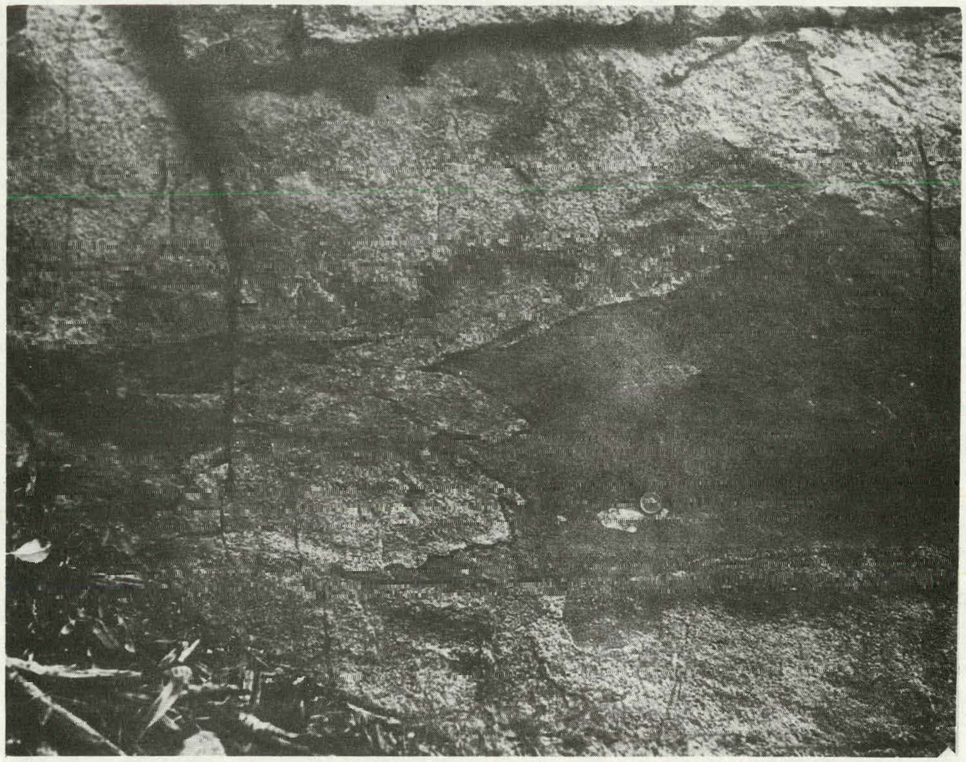


Figure 1.5. Archean granite cutting amphibolitic paraconglomerate layer in lower beds of Phantom Lake Metamorphic Suite. Note clast in paraconglomerate, near nickel right center of photograph. Photograph looking south; located in center of Sec. 5, T. 14 N., R. 86 W., in area north of Jack Creek Mine, central Sierra Madre.

structurally conformable; that is, adjacent bodies of felsic gneiss or foliated granite are in conformable contact with these units or appear to grade into them. These metasedimentary-metavolcanic rocks may have been part of a larger body that once constituted a greenstone belt, but this remains conjectural.

Proterozoic-type Metasedimentary and Metavolcanic Rocks of Archean Age

In the northern Sierra Madre, northern Medicine Bow Mountains, Hartville Uplift, and Black Hills (Figure 1.1), there are rock sequences that contain either a preponderance of, or significant amounts of quartzite of continental or marine origin, normal marine slate, metadolomite, and other units that suggest shallow marine or continental environments of deposition. These rock types are more typical of metasedimentary successions of Early Proterozoic age than of typical volcano-sedimentary Archean greenstone belt successions. They may represent a transition from the greenstone belt environment to the more modern Proterozoic-type sedimentary successions (Houston and Karlstrom, 1980). This transitional nature is further supported by the fact that these Early Proterozoic-type Archean successions do contain graywackes and metavolcanic rocks in greater proportion than a typical Early Proterozoic metasedimentary succession.

In the northern Sierra Madre and northern Medicine Bow Mountains the Archean, Early Proterozoic-type succession is referred to as the Phantom Lake Metamorphic Suite. The stratigraphy of the Phantom Lake Metamorphic Suite is shown in Figure 1.6 and will be discussed in more detail later, but rock types include quartzite, marble, paraconglomerate, quartz-pebble conglomerate, slate, graywacke (probably in good part volcaniclastic) tuffs, basalt, and andesite. These rock types are partly of continental

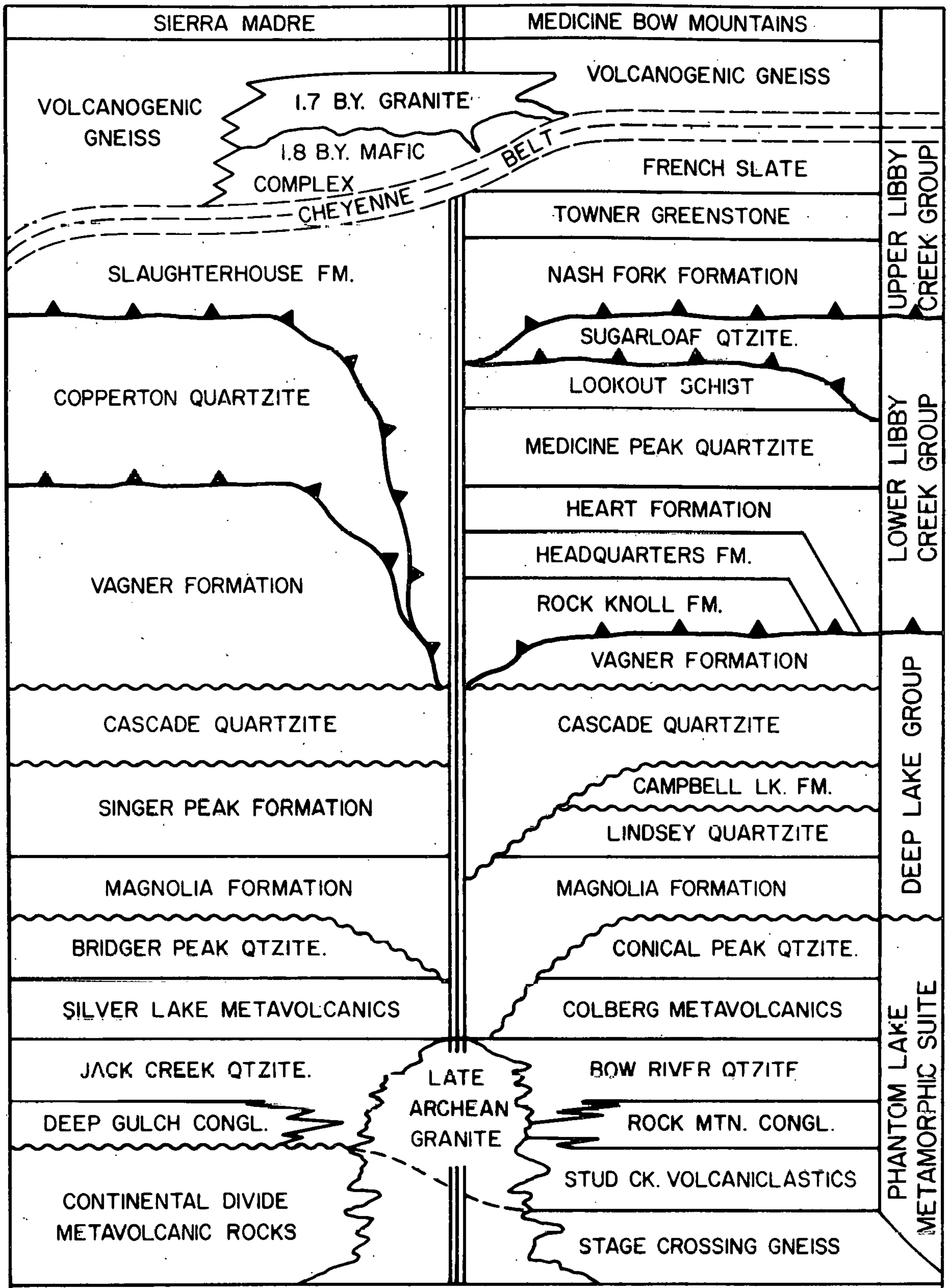


FIGURE 1.6. COMPARATIVE STRATIGRAPHIES OF METASEDIMENTARY ROCKS IN THE SIERRA MADRE AND MEDICINE BOW MOUNTAINS, WYOMING.

and partly marine origin. We estimate that 60-70% of these rocks are continental or near shore marine. The total rock succession includes 40 percent (or less) rocks of volcanic origin, and a significant portion of these volcanic rocks is considered continental. The rocks of the Phantom Lake Metamorphic Suite were deposited on a continental platform or continental margin. Clearly there was a well-developed continental land mass prior to deposition of these rock types which contributed detrital quartz and granitic rock fragments and they are more like Proterozoic and Phanerozoic sedimentary rocks than rocks of the greenstone belts.

We consider rocks of the Phantom Lake Metamorphic Suite to be Archean in age because the older strata of the suite are cut by tonalite (Figure 1.5) dated as about 2700 m.y. (Carl E. Hedge, personal communication, 1979) and younger strata of the suite are cut by similar, but undated granitic rocks.

The continental sedimentary rocks of the Phantom Lake Metamorphic Suite, in particular, were a source of heavy minerals that could be reworked and deposited in conglomerates of Early Proterozoic age. In fact, the Phantom Lake Metamorphic Suite contains significant deposits of uranium and thorium minerals in quartz-pebble conglomerates that were undoubtedly deposited in braided streams and rivers. These Archean quartz-pebble conglomerates differ significantly from their Early Proterozoic counterparts of this area because they contain a much higher proportion of thorium than uranium. It is thus clear that the Early Proterozoic conglomerates were not simply reworked from Late Archean deposits.

Recent geologic mapping by George L. Snyder of the United States Geological Survey (Snyder, 1980) in the Hartville uplift of southeastern Wyoming (Figure 1.1) has clarified the Precambrian stratigraphy of this area. Snyder's (1980) stratigraphic succession of the Whalen Group includes a lower (base unknown) unit consisting of quartzite that grades upward and laterally into a tremolite dolomite with local stromatolites. This unit is overlain by amphibolite (some derived from pillow basalt) and chlorite schist units that were probably largely metavolcanic rocks. These rocks are, in turn, overlain by a schist which was partly derived from normal shale and partly derived from graywacke. The uppermost unit of this succession is largely dolomite with local stromatolites. Essentially, the Whalen Group is a lower shallow marine unit, overlain by a deeper water unit of submarine volcanic rocks, shale, and graywacke, and an upper shallow marine unit. This sequence is quite similar to the tripartite sequence of the Phantom Lake Metamorphic Suite which has a basal continental and near-shore marine quartzite succession, overlain by a continental and near-shore marine volcanic succession, with a near-shore marine quartzite succession at the top. New geochronological studies by Z. E. Peterman of the United States Geological Survey (Snyder, 1980, p. 4) support this correlation by demonstrating that granite cutting the Hartville section is older than about 2600 m.y. To our knowledge, the Whalen Group does not contain uranium- or thorium-bearing rocks.

The final Proterozoic-type metasedimentary-metavolcanic succession which may be of Archean age is in the Black Hills of South Dakota where a lower platform-type succession consisting of quartzite-iron-formation, ortho- and paraconglomerate, arkose and slate is succeeded by a thick eugeosynclinal graywacke-metavolcanic rock group (Redden, 1980). Part of the lower platform-type rocks are probably Archean in age (Redden, 1980, p. 19) but all that is known about the bulk of this rock group is that it is older than the 1700 m.y. Harney Peak granite that cuts it. Part of the lower platform-type succession may correlate with Phantom Lakte Metamorphic Suite rocks but dating is inadequate to verify the correlation.

Late Archean Granites

It is clear from the above discussion that pink granite is widespread south of the geochronological boundary and that this late intrusive rock is distinctively radioactive as compared with most older rocks. We have noted above background radioactivity in pink granites of the Granite Mountains, Seminoe Mountains, Freezeout Hills, and north central Laramie Range. In addition, Stuckless (1979) has demonstrated, through detailed isotope studies, that pink granites of the Granite Mountains and Laramie Range have anomalously high thorium content and that, were it not for uranium loss during the Cenozoic, these same granites would have anomalously high uranium contents as well.

The pink granites are part of a Late Archean granite-forming event that took place in the Wyoming Province between 2,450-2,600 m.y. (Peterman and Hildreth, 1978, p. 14). Peterman and Hildreth (1978, p. 14) list granites of Lankin Dome and Tincup Mountain in the Granite Mountains, the Bears Ears pluton in the Wind River Mountains, Mount Owen quartz monzonite in the Teton Range, Baggot Rocks granite of the Medicine Bow Mountains, Baggot Rocks granite of the Sierra Madre, and Laramie granite of the Laramie Mountains as belonging to this group, and we suggest that pink granites of the Seminoe Mountains and Freezeout Hills also belong to this group.

We are not certain if all granites belonging to the 2,450-2,600 m.y. age group contain thorium and uranium values in greater than normal amounts. But, most of the ones in southern Wyoming do and these granites are certainly a logical source for the uranium and thorium minerals in Early Proterozoic quartz-pebble conglomerates. If so, minerals and granite clasts derived from such a source should not appear in metasedimentary rocks older than about 2400 m.y.

Both Late Archean quartz-pebble conglomerates of the Phantom Lake Metamorphic Suite and Early Proterozoic quartz-pebble conglomerates of the Deep Lake Group of the Medicine Bow Mountains and Sierra Madre have associated paraconglomerate that contains granite clasts. We have not made a statistically valid study of granite clasts in paraconglomerate, but our field observations in the Medicine Bow Mountains and Sierra Madre suggests that the percentage of pink and red granite and granite gneiss clasts is inversely related to age of the paraconglomerate. Therefore, there may be a relationship between the formation of a Late Archean granite source rock and the deposition of uranium minerals in quartz-pebble conglomerates. The quartz-pebble conglomerates of the Late Archean Phantom Lake Metamorphic Suite probably originated prior to formation of the uranium-rich pink and red granites, and this may account for the paucity of uranium in these conglomerates, whereas the Early Proterozoic quartz-pebble conglomerate of the Deep Lake Group in the Medicine Bow Mountains probably originated after formation of the uranium-rich pink and red granites and this may account for the relative abundance of uranium in these conglomerates as compared with the Late Archean type. Obviously, local source will also play a role in determining the mineralogy of quartz-pebble conglomerate and, in a given area, quartz-pebble conglomerate or the right age may have been formed in a drainage basin with no granite source and thus show a different heavy mineral suite.

Model for Evolution of the Archean Wyoming Province

The granitic gneiss terrane of the Archean core area of the Wyoming Province probably was formed prior to 3000 m.y. ago. The formation of this crustal block may have involved granulite facies metamorphism over much of the area and must have involved re-working of older granite or felsic igneous rocks and formation of new granite bodies. Pre-existing metasedimentary and metavolcanic rocks were largely reconstituted leaving rare remnants in a larger body of granitic gneiss. Uranium, thorium, rubidium and potassium were probably depleted during the various metamorphic and granite forming events.

Southeast of the core area, a granitic gneiss terrane very similar to that of the core area was subjected to metamorphism and about 2800-2900 m.y. ago in an area including the Granite Mountains and nothern Laramie Mountains and later, about 2500-2600 m.y. ago in the central Laramie Mountains, Medicine Bow Mountains and Sierra Madre. These metamorphic events may simply be evidence of later re-working of the southern parts of the Archean core; if so, the re-working may have resulted in a further depletion in U, Th, Rb, and K in the granitic gneiss terranes. Another possiblity is that the two "younger" strips of granitic gneiss may have been added to the older core by a process resembling modern arc-arc collision. This latter possiblity would probably mean that younger granitic gneiss terrane would have a higher uranium content than the core area.

Archean greenstone belts have been interpreted to have formed in a variety of ways: as rifts within granitic gneiss terranes that are limited in extent (Anhausser and others, 1969), as foundered shield volcanoes developed in a granitic gneiss terrane (Condie and Hunter, 1976), as portions of early mafic crust (Glickson, 1976) and as basins formed in a similar manner to modern interarc basins (back-arc basins) of the western Pacific (Goodwin, 1973; Windley, 1977).

We suspect that there is a relationship between the development of the greenstone belts and the granitic gneiss terrane south of the Archean core of the Wyoming Province. The age of the rocks of the greenstone belts is unknown although all are older than 2500-2600 m.y. The greenstone belt rocks may have developed in a back-arc basin or basins and in intra-arc basins that developed between the Archean core area and a block of Archean core that moved southeast (Figure 1.7A). The Archean core area may have extended to the northwest margin of the southern Wind River greenstone belt and to the southern Owl Creek Mountains (Figure 1.1); the block of Archean core that separated may be repesented by rocks of the northern and central Laramie Range, northwestern Medicine Bow Mountains, eastern Sierra Madre and perhaps the Hartville Uplift (Figure 1.1). Closure of these back-arc basins or basin could have resulted in deformation of the rocks of the back-arc basins, island arcs and any remnants of Archean crust remaining in the back-arc basins.

If some subducted plates dipped southeast a substantial mass of K-rich granite might have formed under the southeastern Archean block. This might account for the formation of the Late Archean potassium- and uranium-rich granites discussed above.

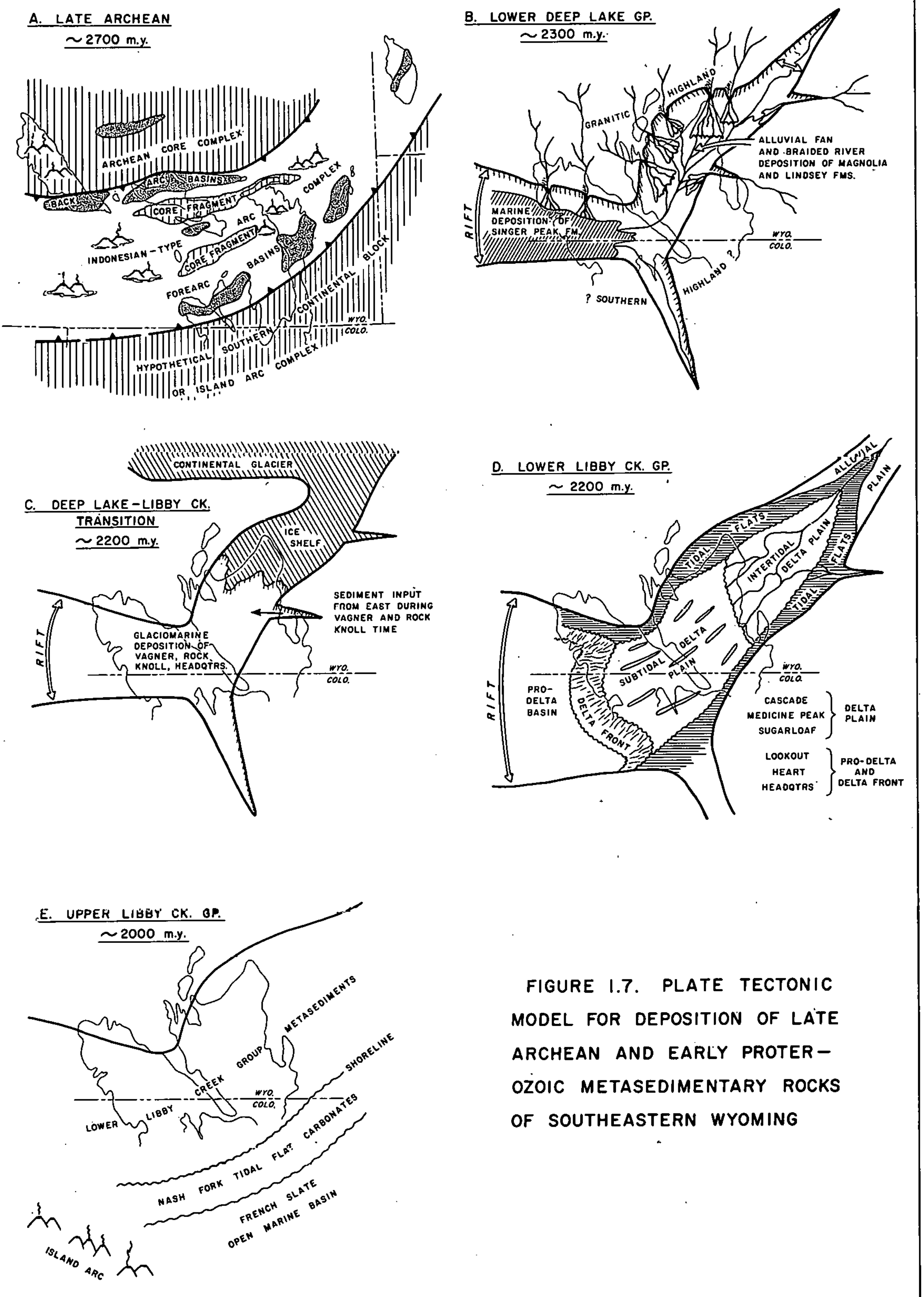


FIGURE I.7. PLATE TECTONIC MODEL FOR DEPOSITION OF LATE ARCHEAN AND EARLY PROTEROZOIC METASEDIMENTARY ROCKS OF SOUTHEASTERN WYOMING

Inasmuch as the platform-type metasedimentary and metavolcanic rocks (Phantom Lake Metamorphic Suite of the Sierra Madre and Medicine Bow Mountains, metasedimentary and metavolcanic rocks of the Hartville area, and possibly some older metasedimentary rocks of the Black Hills) are older than the Late Precambrian K- and U-rich granites, they were undoubtedly deposited prior to closure and may be broadly contemporaneous with development of the back-arc basins. These metasedimentary and metavolcanic rocks are regarded as platform or continental margin deposits of the fragment of original Archean core which moved southeast.

The above model is over-simplified and highly speculative, but it is consistent with some of the previous work done in this area. For example, geochemical studies of both the Laramie batholith (Condie, 1969) and Late Archean granites of the Granite Mountains (Peterman and Hildreth, 1978) suggest origin by partial melting of lower crust or upper mantle with lower crust contamination. A plate tectonic concept somewhat like the one suggested here has been proposed previously by Condie (1972, p. 104-112) for the greenstone belt of the southern Wind River Mountains and Condie cites some parallels between the chemistry of rocks of this greenstone belt and modern island arc systems of the South Pacific.

Summary: Probable Source for Uranium and Other Heavy Minerals in Early Proterozoic Conglomerates

It is quite clear from the review of the Archean source area that the most probable source of uranium and thorium minerals is the Late Archean granites, and their associated contact zones, and shear zones either in the granite or in metasedimentary and metavolcanic rocks near the granite. It is also obvious that other minerals such as gold, pyrite, and other sulphide minerals could have come from a variety of sources in the Archean greenstone belts, and that these minerals are not in any way tied to a Late Archean granite source rock.

Therefore, in the Wyoming Province, quartz-pebble conglomerates older than about 2700 m.y. probably will not contain significant uranium minerals whereas quartz-pebble conglomerates younger than 2600-2700 m.y. have a high probability of containing uranium minerals. The presence of uranium minerals in quartz-pebble conglomerates is also limited by the development of a more strongly oxidizing atmosphere by about 2000 m.y., so that the search for uranium bearing quartz-pebble conglomerate can best be confined to the 2600-2000 m.y. old deposits in this area.

PROTEROZOIC

There are two Proterozoic terranes in southeastern Wyoming, northeastern Utah and northwestern South Dakota (Figure 1.1). One contains a combination of platform and continental margin metasedimentary and metavolcanic rocks and is essentially miogeoclinal; another contains marine volcano-sedimentary rocks and is essentially eugeoclinal. As shown in Figure 1.1, the miogeoclinal rocks crop out in two small areas of the northeastern and southwestern Black Hills, the northern Medicine Bow Mountains, northern Sierra Madre, and northeastern Uinta Mountains of Utah. These miogeoclinal metasedimentary rocks lie on Archean basement and are located north of and within 30-50 km of an irregular east-northeast to northeast striking zone of cataclastic rocks referred to as the Cheyenne Belt (Houston and others, 1980) which marks the southern boundary of the Wyoming Province (Figure 1.1). There are no known rocks of Archean age south of this boundary.

Stratigraphy and Paleogeography of Early Proterozoic Miogeoclinal Metasedimentary Rocks

The most complete sections of miogeoclinal metasedimentary and metavolcanic rocks are in the Medicine Bow Mountains and Sierra Madre. In both of these ranges the basal Proterozoic rock succession is referred to as the Deep Lake Group, but formations within the Deep Lake Group of the Sierra Madre and Medicine Bow Mountains are not identical (Figure 1.6). In fact, it has been necessary to utilize different formation names in the two ranges because of facies changes between sections less than twenty miles apart (Figure 1.6).

In both the Sierra Madre and Medicine Bow Mountains the basal formation of the Deep Lake Group is the Magnolia Formation which lies unconformably on older metasedimentary and metavolcanic rocks of the Archean Phantom Lake Metamorphic Suite and, in the Medicine Bow Mountains, on greenstone belt-type rocks of the Stage Crossing Gneiss and associated pink granitic gneiss of the Baggot Rocks-type granite.

The Magnolia Formation is divided into a basal Conglomerate Member and an upper Quartzite Member (Karlstrom and Houston, 1979a, 1979b). The Conglomerate Member is a discontinuous basal conglomerate which contains pebbles and cobbles of quartz, quartzite, granite, and metavolcanic rocks in varying proportions in a subarkosic to arkosic matrix. In the Medicine Bow Mountains granite clasts are common in the northeast whereas metavolcanic rock clasts are more common in the southwest probably reflecting differences in local source areas. Lithologies in the Conglomerate Member include arkosic polymictic paraconglomerates which are only slightly radioactive, strongly radioactive quartz-pebble conglomerates, and pebbly and granular quartzites. These rocks are complexly interfingered but, in places, are systematically ordered in fining-upward successions of probable fluvial (alluvial fans?) origin. Strongly radioactive quartz-pebble conglomerates have been noted within arkosic paraconglomerates in many areas of the Medicine Bow Mountains but the Onemile Creek area of the northeastern Medicine Bow Mountains is the only known area where quartz-pebble conglomerates are thick enough, extensive enough, and contain enough uranium and thorium to be of potential economic interest. The Conglomerate Member has not been mapped as a separate unit in the Sierra Madre but slightly radioactive quartz-pebble conglomerates are present at the base of the Magnolia Formation in the western Sierra Madre.

The Conglomerate Member of the Magnolia Formation grades up-section into very coarse-grained quartzites and quartz-granule conglomerates referred to as the Quartzite Member in the Medicine Bow Mountains (Karlstrom and Houston, 1979a, 1979b). These rocks are also interpreted by us to be fluvial in origin but they appear to have been deposited at a greater distance from the source than rocks of the Conglomerate Member.

Paleocurrent measurements of the Magnolia Formation in the Medicine Bow Mountains, primarily from the Quartzite Member, suggest deposition in a southwesterly-directed braided river system (Plate 2). Paleocurrents from the Magnolia of the Sierra Madre are directed south and southeast.

The Magnolia Formation in the Medicine Bow Mountains, is overlain conformably by the Lindsey Quartzite which is a fluvial deposit similar in most respects to the Magnolia Formation. The Lindsey Quartzite is finer-grained than the Magnolia Formation, but is composed of fining-upward sets which have conglomeratic bases. The Lindsey Quartzite, however, has numerous shale partings in the quartzite suggesting a more distal source and lower-energy fluvial deposition. Paleocurrent measurements suggest continuing southwest-directed rivers for deposition of the Lindsey Quartzite.

In the Sierra Madre the Magnolia Formation is overlain by the Singer Peak Formation (Graff, 1979) which is a muscovite, chlorite garnet phyllite with thin beds of quartzite in the upper part. Locally, poorly sorted granite-clast paraconglomerate is present in the upper part of the unit. The Singer Peak Formation is thought to be a marine facies of the Lindsey Quartzite of the Medicine Bow Mountains (Figure 1.7B).

In the Medicine Bow Mountains the Lindsey Quartzite is overlain disconformably by the Campbell Lake Formation which is a two-fold unit with a basal paraconglomerate and an upper phyllite. The paraconglomerate beds in the upper part of the Singer Peak Formation of the Sierra Madre may correlate with the Campbell Lake Formation (Figure 1.6). The Campbell Lake Formation and upper Singer Peak paraconglomerate beds may represent glacial diamictites or alluvial fan debris flow deposits.

The Cascade Quartzite overlies the Campbell Lake Formation unconformably and is a distinctive quartzite succession which is widespread in both the Medicine Bow Mountains and Sierra Madre. In the Medicine Bow Mountains the basal beds of the Cascade lie unconformably on beds as old as Archean. The Cascade Quartzite is a much more mature quartzite than any of the quartzites that underlie it, it contains abundant stable clasts such as quartz and chert and has very small percentages of phyllosilicates compared with older quartzite successions. The sedimentary structures of the Cascade include both planar and trough crossbedding and can be interpreted as fluvial or near-shore marine; the great lateral extent of the unit and its consistent lithology suggest a marine origin.

The Vagner Formation lies unconformably on the Cascade Quartzite and is present in both the Medicine Bow Mountains and Sierra Madre. The Vagner Formation is a paraconglomerate-marble-phyllite-quartzite succession interpreted to be glaciomarine in origin (Figure 1.7C).

Overall, the Deep Lake Group is a transgressive succession. The lower part records fluvial sedimentation on a continental platform; the upper part records shallow marine and glaciomarine sedimentation on a continental shelf. Unfortunately, it has not been possible to directly date rocks of the Deep Lake Group. However, the basal beds of the Deep Lake Group in the Medicine Bow Mountains unconformably overlie granitic rocks similar to the 2500 m.y. old Baggot Rocks Granite (Hills and others, 1968) so the Deep Lake Group is Proterozoic. An upper limit is given by a 2000 m.y. date on the Gaps Trondhjemite (Carl Hedge, personal communication, 1980) which crosscuts the Lower Libby Creek Group in the Medicine Bow Mountains.

As shown schematically in Figure 1.6, the upper part of the Deep Lake Group in both the Sierra Madre and Medicine Bow Mountains is truncated by major thrust faults which bring younger rocks of the Libby Creek Group against older rocks of the Deep Lake Group and Archean basement. In the Medicine Bow Mountains a complete section of the Libby Creek Group (over 6 km) is exposed whereas only a partial section (2 km) is exposed in the Sierra Madre. The Libby Creek Group is divided in this report into an upper and lower part. The lower part consists of continental margin-type siliciclastic rocks (primarily marine quartzite) whereas the upper part, which is thrust over the lower, is a carbonate bank, shale, submarine volcanic succession that was probably deposited well off-shore.

The basal formation of the Lower Libby Creek Group in the Medicine Bow Mountains is the Rock Knoll Formation which is a quartzite-phyllite succession with local layers of conglomerate. The Rock Knoll quartzites are plagioclase-rich arkoses and subarkoses which are similar to quartzites interlayered with paraconglomerates of the underlying Vagner Formation of the Deep Lake Group and the overlying Headquarters Formation of the Lower Libby Creek Group, both interpreted to be glaciomarine. We believe, therefore, that the Rock Knoll Formation is also glaciomarine in origin. There is no Rock Knoll equivalent in the Sierra Madre.

The Rock Knoll Formation of the Medicine Bow Mountains is overlain by the Headquarters Formation which is comprised of paraconglomerate, quartzite, and laminated schists and phyllite. These rocks are interpreted to be glaciomarine in origin on the basis of lithologic characteristics such as poor sorting and only faint stratification in the paraconglomerates, the presence of angular dropstone clasts in the paraconglomerate, the association of paraconglomerates and laminated (probably marine) phyllites, and chemical similarities between rocks of the Headquarters Formation and other Early Proterozoic glacial-related units. The Headquarters Formation has not been identified in the Sierra Madre.

Inasmuch as the Vagner Formation, Rock Knoll Formation and Headquarters Formation are all regarded as glaciomarine, they may be related to the same depositional episode and may not be separated by a major time break. However, our tectonic interpretation indicates that the Vagner Formation was deformed into broad folds prior to the thrusting which brought the Rock Knoll and Headquarters Formations against rocks of the Vagner Formation.

In the Medicine Bow Mountains the Headquarters Formation is overlain conformably by the Heart Formation which is composed chiefly of argillaceous quartzite. The Heart Formation is interpreted to have been deposited in a macrotidal deltaic environment ranging from prodelta sedimentation at the base to delta front at the top. No beds equivalent to the Heart Formation have been identified in the Sierra Madre.

The Heart Formation is conformably overlain in the Medicine Bow Mountains by the Medicine Peak Quartzite which is a 1700 m thick succession of aluminous (kyanite and pyrophyllite) quartzarenite. Similar quartzites (1400 m thick) in the Sierra Madre are called the Copperton Quartzite (Graff, 1978). Sedimentological studies of the Medicine Peak Quartzite indicate that, like the Heart Formation, it was deposited on a macrotidal delta (Figure 1.7D). The base of the Medicine Peak Quartzite is thought to represent subtidal sand shoals on a macrotidal delta plain, the middle Medicine Peak Quartzite is believed to be deposits laid down in subtidal to intertidal delta plain channels and bars and the upper Medicine Peak Quartzite, like the lower, is considered to be subtidal sand shoals on the delta plain. The Copperton Quartzite of the Sierra Madre is much more highly deformed than the Medicine Peak Quartzite and has not been studied sedimentologically. However, it too, probably represents deltaic sedimentation.

The Medicine Peak Quartzite is overlain by the Lookout Schist in the Medicine Bow Mountains (Figure 1.6). The Lookout Schist, a complex succession of interlayered quartzite, phyllite, and schist, is interpreted as a delta front (upper and lower Lookout Schist) and prodelta deposit (chemical sediments of the middle Lookout Schist). Similar rocks in the Sierra Madre are lumped with the Copperton Quartzite.

The uppermost formation of the Lower Libby Creek Group in the Medicine Bow Mountains is the Sugarloaf Quartzite which lies conformably on the Lookout Schist. The Sugarloaf Quartzite is believed to consist of deposits laid down in subtidal to intertidal delta plain environments.

Considered together, the deposits of the Deep Lake and Lower Libby Creek Group are related to a prolonged period of Early Proterozoic siliciclastic sedimentation in which detritus was derived from the Wyoming Province and carried by major river systems. Early deposition in the Deep Lake Group took place in high-energy alluvial fan and braided river environments on the continental platform (Figure 1.7B); upper Deep Lake and Lower Libby Creek Group sedimentation was deltaic and took place on the continental margin where marine tides and currents could influence sedimentation (Figure 1.7D). Processes of deltaic sedimentation was interrupted at least once and perhaps several times by glacial episodes, which deposited glaciomarine sediments in the uppermost Deep Lake and lowermost Libby Creek Groups (Figure 1.7C).

In the Medicine Bow Mountains, where we have sufficient paleocurrent measurements to aid in interpreting depositional environments, it is clear that the major current direction throughout the depositional episode was southwest (Plate 2). If the overall shoreline trend of the Early Proterozoic of this area was northeast, as implied by the present day orientation of the Cheyenne Belt, both non-marine and marine deposits of the Medicine Bow Mountains must have been laid down in major rivers and river-delta systems in which clastic transport was sub-parallel to the Wyoming Province shoreline. We feel that the best explanation for this situation is that the clastic deposits of the Medicine Bow Mountains were laid down in grabens or a graben oriented sub-parallel to an Early Proterozoic rifted margin (Figure 1.7), perhaps like grabens described by Burke (1976, p. 92-112) that developed during the initial stages of rifting of the Atlantic Ocean in the Mesozoic. However, another valid interpretation might postulate a large embayment in the shoreline in this area of the Wyoming Province in the Early Proterozoic.

There are a number of depositional environments that might be favorable for deposition of clastic uranium or uranium-thorium minerals in the Deep Lake and Lower Libby Creek Groups. These include the braided stream environments of the Magnolia Formation and Lindsey Quartzite, and the deltaic channel deposits of the Cascade Quartzite and middle part of the Medicine Peak Quartzite. In fact, radioactive conglomerate beds have been found in all of these units although the only extensive uranium concentrations found so far are in the Magnolia Formation.

The Upper Libby Creek Group represents a major change in paleogeography in southern Wyoming. In both the Medicine Bow Mountains and Sierra Madre, the Upper Libby Creek Group is separated from the Lower Libby Group by a major thrust fault that brings rock types of an entirely different facies over the beds of the Lower Libby Creek Group (Figure 1.6). The Upper Libby Creek Group, from oldest to youngest, consists of the Nash Fork Formation, Towner Greenstone, and French Slate. These deposits are marine and are believed to represent a carbonate bank (Nash Fork Formation), submarine volcanics (Towner Greenstone), and normal shales (French Slate) deposited well off-shore from the chiefly deltaic deposits of the Deep

Lake and Lower Libby Creek Groups. As shown in Figure 1.7E, these deposits may have developed along a rifted margin of the Atlantic-type (Dewey and Bird, 1970). We believe the deposits of the Upper Libby Creek Group must have extended for many miles along the Early Proterozoic continental margin perhaps into the Minnesota-Michigan area where the stromatolitic dolomites similar to those of the Nash Fork Formation are found in the Trout Lake Formation of the Cuyuna Range, the Bad River Dolomite of the Gogebic Range, and the Kona Dolomite and Randville Dolomite of the Marquette Range Supergroup of Michigan (Houston and Karlstrom, 1980).

We must emphasize that, in southeastern Wyoming, the complete Upper Libby Creek Group is preserved only in the Medicine Bow Mountains. In the Sierra Madre a marine carbonate succession named the Slaughterhouse Gulch Formation is correlated with the Nash Fork Formation (Figure 1.6), but no beds equivalent to the Towner Greenstone or French Slate have been identified. The Slaughterhouse Formation of the Sierra Madre is more highly deformed than the Nash Fork Formation of the Medicine Bow Mountains, stromatolites have not been recognized and the Slaughterhouse Formation contains a higher proportion of interbedded slates and phyllites than the Nash Fork Formation. We suggest that the Slaughterhouse Formation was deposited in deeper water than the Nash Fork Formation.

The age of rocks of the Upper Libby Creek Group is uncertain. We suspect that they correlate with the Marquette Range Supergroup of the Lake Superior Region and, if so, are between about 2100 m.y. and about 1900 m.y. old. As shown in Figure 1.1, there are no migeosynclinal rocks equivalent to those of the Sierra Madre and Medicine Bow Mountains in the Laramie Range or Hartville Uplift where metasedimentary and metavolcanic rocks are dated as Archean. However, in the northeastern Uinta Mountains of Utah and in the Black Hills of South Dakota there are metasedimentary rocks that we believe correlate with some parts of the miogeosynclinal succession of the two southern Wyoming mountain ranges.

Redden (1980) describes a succession of metasedimentary rocks in the Nemo area of the northeastern Black Hills of South Dakota that he classes as platform-type in contrast to the eugeosynclinal metasedimentary and metavolcanic rocks of most of the central Black Hills. These rocks include paraconglomerate, quartz-pebble conglomerate, quartzite, phyllite, dolomite, and iron formation. Redden (1980, p. 34-35) demonstrates that these deposits were formed in an area undergoing extensional tectonics and suggests deposition in small shallow basins (grabens?). The environments of deposition range from fanglomerates near faults to fluvial deposits in streams or rivers, to deeper water (marine?) deposits of iron formation and dolomite. Radioactive uranium-thorium-gold bearing quartz-pebble conglomerates are present in the fluvial deposits of the Tomahawk Tongue of the Boxelder Formation of this area. The geochemistry and mineralogy of the fluvial deposits of the Tomahawk Tongue resemble those of the Onemile Creek area of the Medicine Bow Mountains, and it is possible that these rocks are the same age. Redden (1980, p. 78-79) cites evidence that the minerals in the fluviatile conglomerate are derived from the 2500 m.y. old Little Elk granite of this area and he shows that the deposits are older than the Blue Draw metagabbro dated as about 2100 m.y. If our concept of a rifted margin can be extended to the Black Hills, Redden's platform-type rocks may have been formed in grabens developed during early stages of an aulacogen. Platform-type metasedimentary rocks may also be present in the western Black Hills in the Bear Mountain Dome (Kleinkopf and Redden, 1975).

In the northeastern corner of Utah a group of rocks referred to as the Red Creek Quartzite crop out below the Late Proterozoic Uinta Mountain Group (Graff and others, 1980) and lie on a basement of granitic gneiss of Archean age. The Red Creek Quartzite is primarily massive quartzite but it contains at least three subdivisions that consist of alternating schist, quartzite and graywacke, with an occasional bed of marble. The Red Creek Quartzite is probably a marine facies of some part of the Lower Libby Creek Group of the Medicine Bow Mountains and Sierra Madre. It is more like rocks of the Sierra Madre than rocks of the Medicine Bow Mountains suggesting that marine incursions moved form west to east along the Early Proterozoic continental margin and that there is less probability of extensive fluvial deposits west of the Sierra Madre.

Eugeoclinal Metasedimentary and Metavolcanic Rocks of Proterozoic Age

Eugeoclinal metasedimentary and metavolcanic rocks that may have been deposited at about the same time as the miogeoclinal rocks discussed above are present in the southern Sierra Madre, southern Medicine Bow Mountains, southern Laramie Mountains, the Black Hills of South Dakota, and in the Front Frange and Park Range of northern Colorado.

These rocks are so highly deformed and metamorphosed in southeastern Wyoming and northern Colorado that their origin is hard to decipher. In local areas where deformation is not extreme, marine graywacke-turbidite successions and both marine and continental volcanics can be recognized. Inasmuch as these eugeoclinal rocks are located south of the Cheyenne Belt, it has been suggested that they developed in island arcs and accreted to the Wyoming Province about 1700 m.y. ago (Hills and Houston, 1979).

The Black Hills aulacogen has better preserved metasedimentary and metavolcanic successions than those of southeastern Wyoming and northern Colorado and metavolcanic rocks are believed to be much less common than in the other two areas. The most common rock types in the Black Hills are graywacke, phyllite, slate, basalt, chert, iron formation, graphitic slate, and various iron and aluminum-rich slates. These rocks are not regarded as island arc deposits. Instead, we interpret them to be marine sediments deposited in a narrow ocean basin during Early Proterozoic rifting.

Proterozoic Igneous Rocks

The most common igneous rocks of Proterozoic age north of the Cheyenne Belt are gabbroic sills and dikes. These intrusives crosscut Archean basement and all three metasedimentary successions and undoubtedly range in age from Archean to late Early Proterozoic (1600 m.y.). Unfortunately, we have been unable to separate Archean from Proterozoic intrusives by mapping. However, limited geochemical data does suggest at least three different compositions which may reflect different age intrusives. Figure 1.8 shows three chemical fields on both the alkali-silica plot (Figure 1.8A) and the AFM plot (Figure 1.8B): 1) two ultramafic bodies which crosscut Archean "basement" in the Sierra Madre form one field; 2) large gabbroic sills in the Phantom Lake Suite and Deep Lake Group from both the Sierra Madre and Medicine Bow Mountains from another, closely grouped, field; and 3) the dikes and sills from the Libby Creek Group in both ranges form a third. The first two are distinctly tholeiitic; the last is higher in alkalies and lower in silica and is transitional between tholeiite and calc-alkalic rocks.

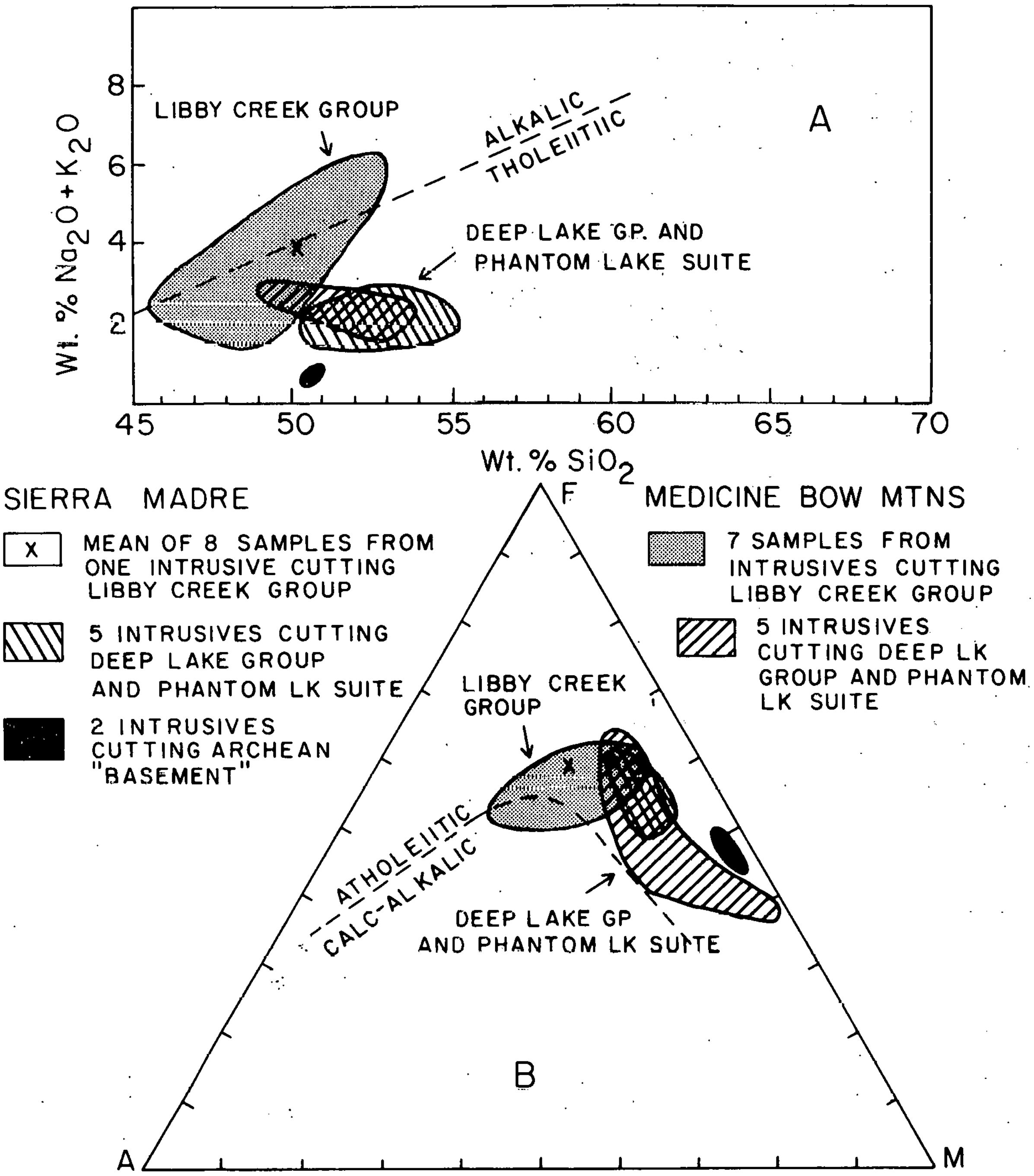


Figure 1.8. Geochemistry of mafic intrusive rocks from the Sierra Madre and Medicine Bow Mountains. A. Alkali-silca plot; the alkalic rock-tholeiite boundary is from MacDonald (1968). B. APM plot; A = $Na_2O + K_2O$; F = Fe_2O_3 + FeO; M = MgO; tholeiite-calc-alkalic rock boundary from Irvine and Baragar (1971).

Geochronologic studies done elsewhere in the Wyoming Province (Stueber and others, 1976; Peterman and Hildreth, 1978, Reed and Zartman, 1973) have defined episodes of intrusion of gabbroic magma around 2500 m.y., 2100-2000 m.y., and 1700 m.y. and we tentatively correlate these ages with the fields discussed above: rocks of field 1 are probably Archean; rocks of field 2 are believed to be 2100-2000 m.y.; and rocks of field 3 may be either 2000 m.y. or 1700 m.y. or some of each. The 2000-2100 m.y. age for many of the large sills in the Deep Lake Group and Phantom Lake Suite is reasonable on geologic grounds. These intrusives are located within about 30 km north of the Cheyenne Belt and are believed to have been emplaced during a rifting episode. The most logical time for rifting is about 2100 m.y. -- the age of tensional tectonics (rifting?) and emplacement of the Nipissing Diabase in the Canadian Shield (Fairbairn and others, 1969) and the age of the emplacement of the Blue Draw Metagabbro in the Black Hills (Redden, 1980).

Many uranium-thorium bearing quartz-pebble conglomerate beds are in contact with major sills of gabbroic composition. The uranium has clearly been redistributed near dike contacts and there is some enrichment locally. We have found no evidence, however, that these mafic intrusive bodies were a source of uranium in the quartz-pebble conglomerates.

The only felsic igneous rock of Proterozoic age north of the Cheyenne Belt is the Gaps Trondhjemite. This unit crops out in several small intrusive bodies which mainly crosscut rocks of the Lower Libby Creek Group but which also appear to crosscut the upper Deep Lake Group in the Sierra Madre. The trondhjemite is always associated in outcrop with gabbroic dikes and sills and it plots on the AFM diagram in about the same field as the gabbroic intrusives which crosscut the Libby Creek Group (Figure 1.8).

Therefore, we believe that the Gaps Trondhjemite is co-genetic with (but a felsic differentiate) of these gabbroic intrusives. The Gaps Trondhjemite has recently been dated as 2000 m.y. old (Carl Hedge, personal communication, 1980) which implies that it and at least some of the gabbroic intrusives in the Libby Creek Group were related to the proposed 2100-2000 m.y. rifting event in southeastern Wyoming.

The Gaps Trondhjemite is also of interest with respect to uranium. Outcrops near Lewis Lake contain fracture systems locally filled with uraninite. So far, the maximum thickness of mineralized fractures found is on the scale of millimeters and maximum rock assays yield 1000 ppm U. However, we have not extensively prospected all outcrops of Gaps Trondhjemite and we have no subsurface information so that more extensive and richer uraninite veinlets may conceivably be found in the future. The source of the uranium is not clear but we assume it was concentrated from low-grade sedimentary rocks during ascent of the felsic magmas.

Proterozoic igneous rocks of a much greater variety are present south of the Cheyenne Belt. The oldest intrusives are a series of large layered mafic complexes of the southern Sierra Madre and Medicine Bow Mountains which are dated as about 1800 m.y. (Synder and Hedge, 1978). Granitic granodioritic, adamellite and tonalite intrusives are present that range in age from about 1750 to 1600 m.y. and late anorthosite (1500 m.y.) and granite (1400 m.y.) are also common south of the Cheyenne Belt (Hills and Houston, 1979). The layered complexes and various felsic intrusives which range in age from about 1800 m.y. to 1700 m.y. may be comagmatic with volcanic rocks (now hornblende gneisses) believed to have been formed in Proterozoic island arcs. The younger intrusives of granite and anorthosite are post-tectonic and their origin is not known.

SUMMARY OF REGIONAL GEOLOGIC AND TECTONIC HISTORY AND THEIR SIGNIFICANCE IN URANIUM PROSPECTING

Figure 1.9 is a summary of our interpretation of the Precambrian geologic history of southeastern Wyoming. As shown, an Archean basement of gneissic rocks older than about 2700 m.y. existed in southeastern Wyoming, or in a locality near southeastern Wyoming, prior to deposition of rocks of the Phantom Lake Metamorphic Suite. The basement consisted of gneissic rocks and greenstone belts with the former predominating in the Sierra Madre and Medicine Bow Mountains. These gneissic rocks of Archean age had a well developed fabric which probably formed during an episode of isoclinal folding and metamorphism of protoliths of the gneisses. The earliest folds recognized in the Archean gneisses are folds of this early foliation that trend north (Houston and others, 1968).

Sedimentary and volcanic rocks of the Phantom Lake Metamorphic Suite were deposited on this basement probably between 2900 m.y. and 2700 m.y. Initial clastic sedimentation took place in braided rivers and streams on a well-developed landmass contemporaneous with subaerial volcanism in other parts of the basin.

The minerals and rock fragments in quartz-pebble conglomerates of the lower Phantom Lake Metamorphic Suite were derived from a gneissic terrane containing foliated gray gneisses and granodiorites. The thorium to uranium ratio of constituents of the granitic source was high, as is reflected in high Th/U in the Archean conglomerates. The shape and extent of the landmass and the original depositional basin are unknown.

Paleocurrent data and lithfacies variations in the fluvial rocks of the lower Phantom Lake Suite suggest a northern source for the sediment. The paleocurrents in the Deep Gulch Conglomerate of the northwestern Sierra

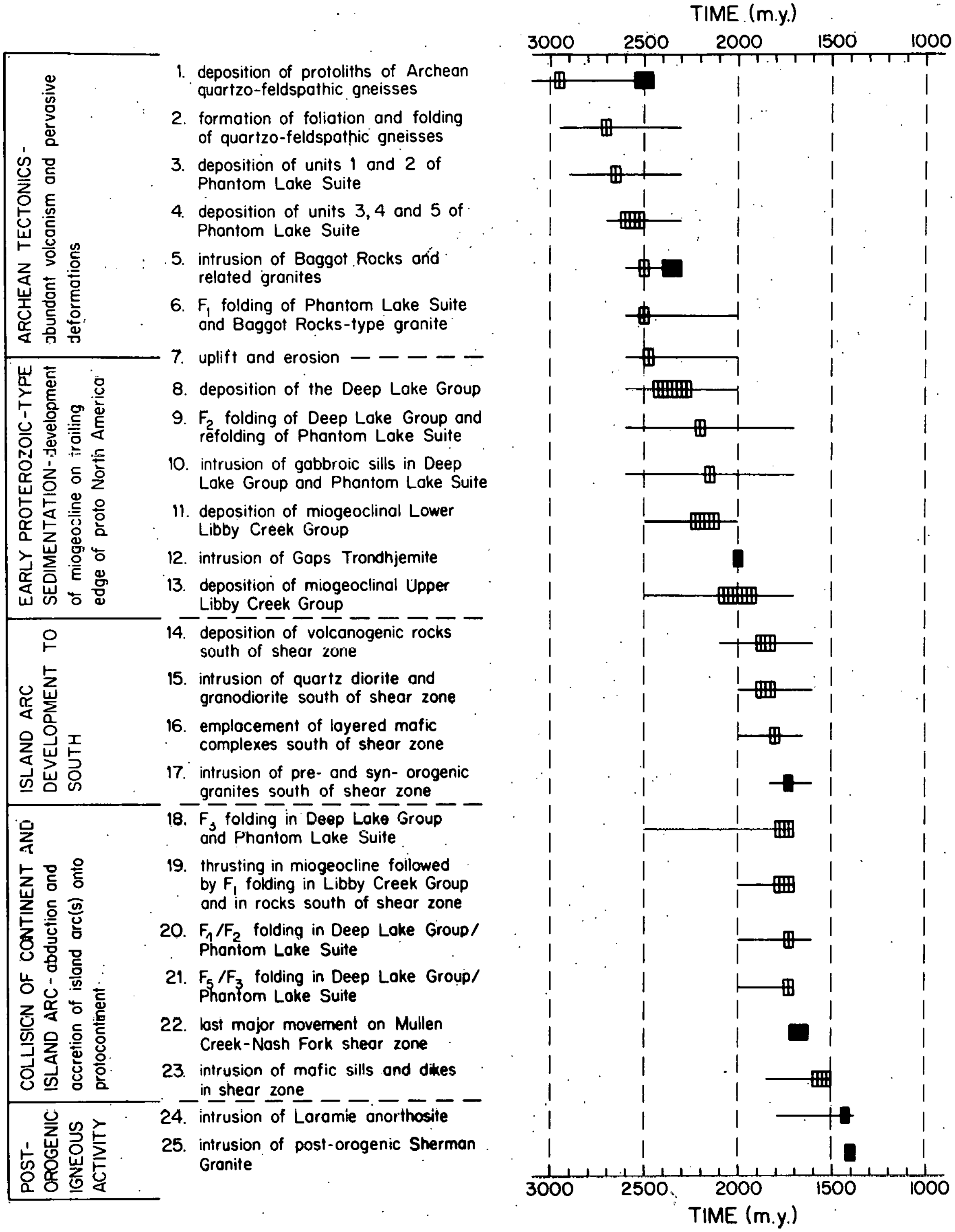


Figure 1.9. Precambrian geologic history of the Medicine Bow Mountains. solid box=geochronological data, showing analytical error; vertical lines=preferred interpretation; solid line=possible range of timing of event. See Table 2.27 for references and justification of dates shown.

Madre were mainly directed south (Plate 7), although the complexity of folding in this area makes these directions somewhat questionable; and the Rock Mountain Conglomerate of the northeastern Medicine Bow Mountains passes south into volcaniclastic rocks suggesting a southward prograding alluvial fan system. Both of these areas were probably close to the margins of the Phantom Lake Suite depositional basin. Fluvial rocks in both ranges are of limited lateral extent and grade quickly (over several hundred meters) upward into shallow marine quartzites (and marble in the Sierra Madre) with bimodal-bipolar paleocurrents directed northeast and southwest. These paleocurrents presumably represent ebb and flood currents in intertidal depositional settings and suggest either a northwest-trending shoreline in this area or deposition in a northeast-trending embayment or trough (similar to the later, Proterozoic, rift basin) in which seas could transgress and regress. Deposition of marine quartzites was interrupted during middle Phantom Lake Suite deposition by widespread volcanism and deposition of thick, mainly subaerial, volcanoclastic sediments. Uppermost Phantom Lake Suite deposition was again of marine quartzites which exhibit northeast-southwest bimodal-bipolar paleocurrent distributions.

The maximum thickness of the sedimentary and volcanic rocks of the Phantom Lake Metamorphic Suite is unknown because of erosion and cover of younger rocks, but it must have exceeded 4780 meters. It is conceivable that the sedimentary succession may have been thick enough to cause local melting of granitic material at the base of the sedimentary wedge at a time of higher heat flow during the Archean because granitic intrusives dated as about 2700 m.y. intrudes the lower part of the succession in the

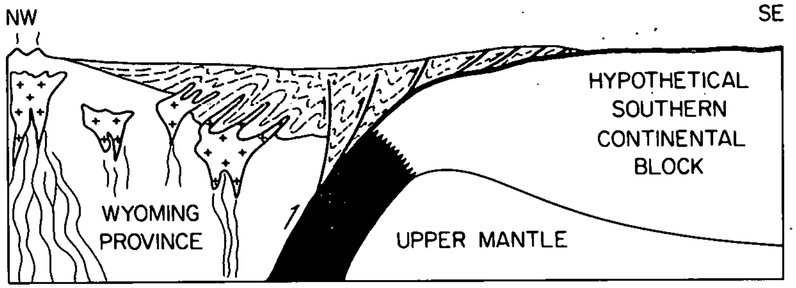
Sierra Madre. However, it seems more probable to us that the granite may have formed during the deformation and metamorphism of the Phantom Lake Metamorphic Suite, which are the most highly deformed and metamorphosed of any miogeoclinal rocks in southeastern Wyoming. In the Sierra Madre and Medicine Bow Mountains Phantom Lake Metamorphic Suite rocks have been deformed into tight to isoclinal folds and metamorphosed to amphibolite facies.

In the Sierra Madre, isoclinal folds and axial plane schistosity of the earliest recognizable fold system (F_1 of Figure 1.9 and Plate 6) have been complexly refolded and now dip at relatively shallow angles south in the central Sierra Madre, west in the western Sierra Madre, and northwest in the northwestern Sierra Madre (Plate 6). F_1 appears to be a nappe system but we are unsure of the initial direction of vergence. As shown in Figure 1.10A,F_1 folds in the Medicine Bow Mountains now strike generally northeast and are overturned, with axial planes dipping northwest (southeast vergence; Plate 3). Deformation of rocks of the Phantom Lake Metamorphic Suite in both ranges was severe enough in this first episode to destroy many primary sedimentary structures so that our interpretations of the stratigraphy and structure of these rocks are less well-constrained than those for the Early Proterozoic metasedimentary sequences.

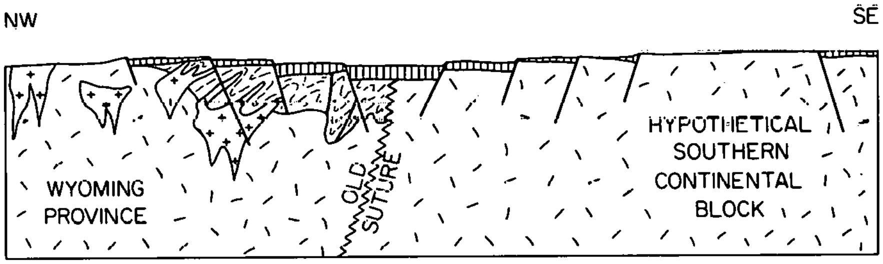
Deformation and metamorphism of rocks of the Phantom Lake Metamorphic Suite was probably part of a general episode of Late Archean orogeny in the Wyoming Province or it may have been part of several episodes of orogeny spanning the time from about 2700 m.y. to 2500 m.y. Granite intrusives dated as about 2700 m.y. (Sierra Madre, Hedge, personal communication, 1980), 2600 m.y. (Hartville Uplift, Snyder, 1980), 2500 m.y. (Laramie Mountains, Hills and Armstrong, 1974) and 2500 m.y. (Medicine Bow

Mountains, Hills and others, 1968) crosscut the various sedimentary and metavolcanic successions in southeastern Wyoming. The 2600 to 2700 m.y. granites appear to be more mafic (granodiorites and tonalites predominate) and the younger, 2500 m.y. old granites appear to be richer in K-feldspar (quartz monzonites and granodiorites predominate). The granitic rocks may have formed by different processes or from different types of source rocks (I-type older granites versus S-type younger granites?). Or, Archean granitic rocks in southeastern Wyoming may reflect an increasing component of upper crustal contamination of mantle or lower crust derived magma, with time. These problems would be a fruitful area for further research. Certainly by about 2400 m.y. the general outline of the Archean Wyoming Province with granitic gneiss terrane, infolded greenstone belts, infolded miogeoclinal successions, and Late Archean granitic intrusions were established and a period of uplift and erosion began.

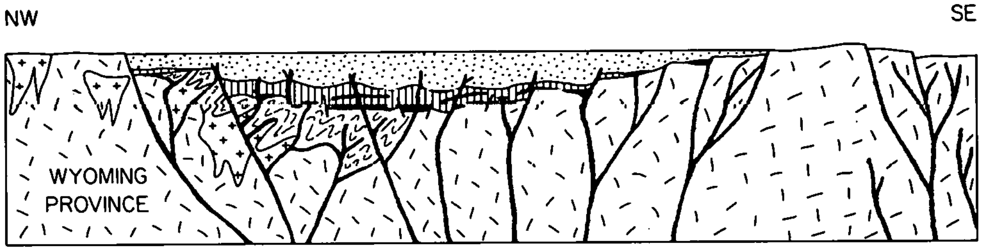
By about 2400 m.y., a once more extensive Wyoming Province in southeastern Wyoming began to rupture and intracontintal rifts and aulocogens developed. Prior to rifting and in the initial stages of rifting, alluvial fans and braided river systems developed on the eroded Archean landmass in the Sierra Madre, Medicine Bow Mountains, and probably in the Black Hills of South Dakota. The river systems must have extended over a wide area because the present outcrop areas extended over a lateral distance of over 420 kilometers. It is these Early Proterozoic fluvial sediments that contain uranium and thorium minerals in high enough concentrations to be of possible economic interest. The lower thorium to uranium ratio of these deposits compared to the Phantom Lake Suite fluvial rocks



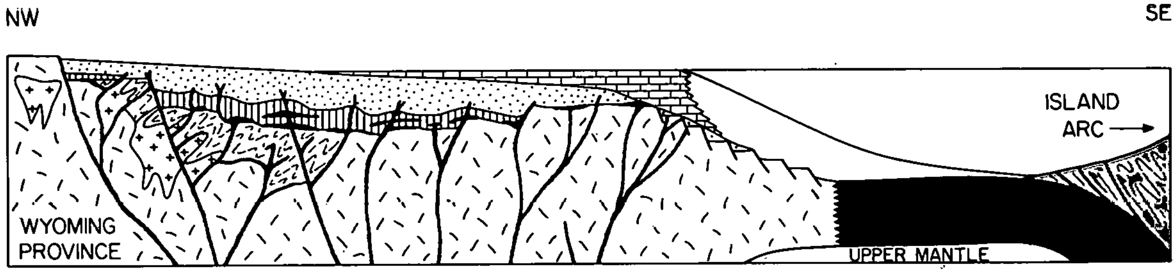
A. LATE ARCHEAN OROGENESIS ~2500 m.y.: SOUTHEAST VERGING DEFORMATION IN PHANTOM LAKE SUITE (F_1); EMPLACEMENT OF SYNOROGENIC GRANITE SILLS AND PHACOLITHS.



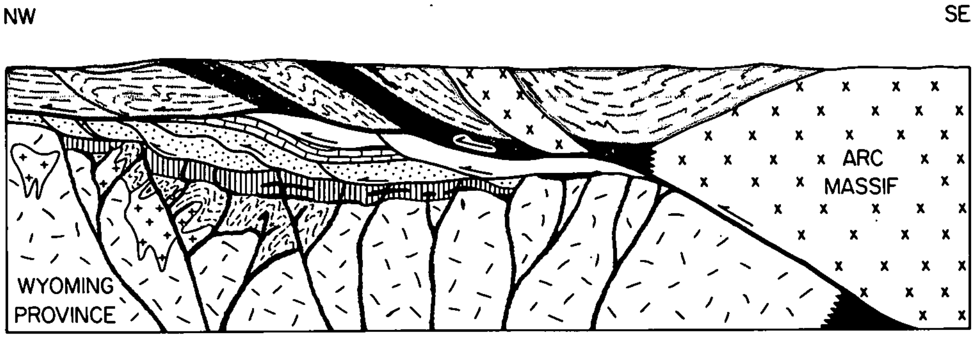
B. RIFT VALLEY SYSTEM ~2300 m.y.: UPLIFT AND RIFTING OF STABILIZED ARCHEAN CRATON; FLUVIAL DEPOSITION OF LOWER DEEP LAKE GROUP, INCLUDING RADIOACTIVE CONGLOMERATE.



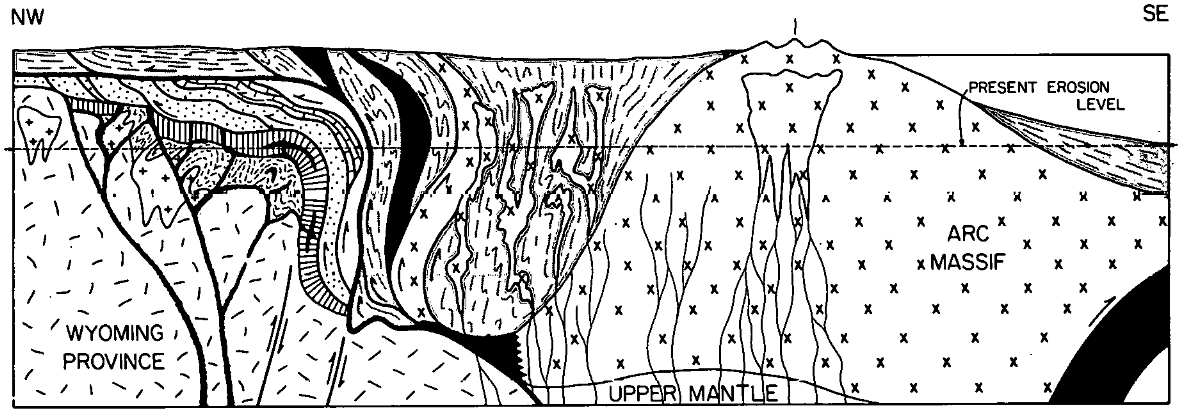
C. PROTO-OCEANIC GULF ~2200 m.y.: DELTAIC DEPOSITION OF UPPER DEEP LAKE AND LOWER LIBBY CREEK GROUPS; FOLDING OF DEEP LAKE GROUP (F_2); INTRUSION OF THOLEIITIC SILLS.



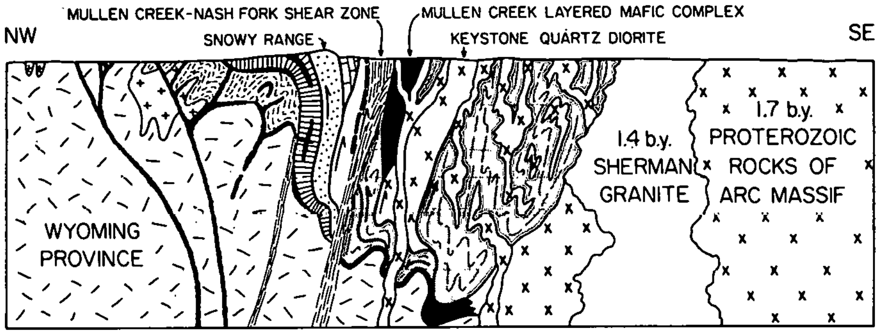
D. OPEN OCEAN ~2000 m.y.: CARBONATE/SHALE DEPOSITION IN UPPER LIBBY CREEK GROUP; APPROACH OF ISLAND ARC FROM SOUTH.



E. CONTINENT-ISLAND ARC COLLISION ~1800 m.y.: OBDUCTION OF ISLAND ARC; THRUSTING IN LIBBY CREEK GROUP AND FOREARC VOLCANOGENIC ROCKS.

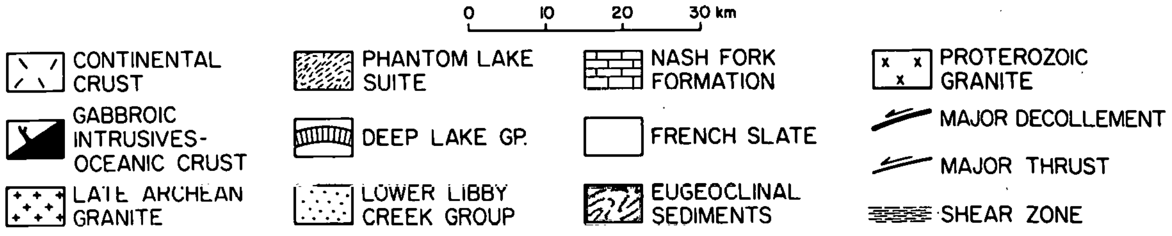


F. FLIP IN DIRECTION OF SUBDUCTION ~1700 m.y.: ROTATION OF BEDS AND THRUSTS; CONTINUED THRUSTING; F_1 AND F_2 FOLDING IN LIBBY CREEK GROUP; F_3 AND F_4 FOLDING IN DEEP LAKE GROUP AND PHANTOM LAKE SUITE; MAXIMUM METAMORPHISM.



G. STRIKE SLIP MOVEMENT ON SHEAR ZONE ~1600 m.y.; INTRUSION OF POST-OROGENIC SHERMAN GRANITE ~1400 m.y.; UPLIFT DURING LARAMIDE OROGENY ~70 m.y.; EROSION TO PRESENT TOPOGRAPHY.

FIGURE 1.10. PLATE TECTONIC MODEL FOR THE PRECAMBRIAN TECTONIC HISTORY OF THE MEDICINE BOW MOUNTAINS, WYOMING.



is thought to reflect the presence of uranium-rich, Late Archean granites (~2500 m.y.) in the source area that was being eroded at this time.

As rifting and break-up of the Archean continental mass continued, narrow ocean basins formed in southeastern Wyoming and the Black Hills. Marine incursions took place from the southwest in Wyoming (Figure 1.7B) where the rifted margin trended northeast to east, and from the southeast in the northwest-striking Black Hills aulacogen. Upper Deep Lake Group deposition included both fluvial and marine sediments and probably represents predominantly deltaic sedimentary environments. Superimposed on this deltaic deposition was an episode of glaciation which deposited glaciomarine sediments in the Sierra Madre and Medicine Bow Mountains.

The beds of the Deep Lake Group were folded probably during and certainly by the end of deposition of the Vagner Formation, the uppermost formation of the Deep Lake Group. Deformation was not severe and broad open folds (F_2 of Figure 1.9 and Plate 3) with axes striking northeast were formed. During and/or after this deformation basaltic magma of tholeiitic affinity was introduced in the rifted areas forming large gabbroic sills and dikes in the Sierra Madre, Medicine Bow Mountains (Figure 1.10C) and the Black Hills. A sill in the northeastern Black Hills referred to as the Blue Draw Metagabbroic and dated as about 2100 m.y. (Redden, 1980) may have formed during this rifting episode.

Continued or renewed glaciation took place in the initial stages of deposition of rocks of the Lower Libby Creek Group. This glacial period probably correlates with a glacial episode recorded in southern Canada and represented by the Gowganda Formation of the Huronian Supergroup. This glacial episode may have been part of a continental glaciation over much of the Canadian Shield at that time (Young, 1970). As the glaciers withdrew, a major delta was re-established in the Medicine Bow-Sierra Madre area and marine sandstone and shale of the Red Creek Quartzite were deposited in northeastern Utah. In the Black Hills, the preserved rocks which may correlate with the Lower Libby Creek Group are a deeper water facies marked by iron formation.

Basaltic magma continued to be introduced to form sills and dikes in formations of the Lower Libby Creek Group of the Sierra Madre and Medicine Bow Mountains. A differentiate of one of the mafic intrusions, the Gaps Trondhjemite, is dated as about 2000 m.y. (Carl Hedge, personal communication, 1980) and thus places an upper limit on the age of the rocks of the Lower Libby Creek Group.

During the latter part of Lower Libby Creek time, the southern continental block had separated from the Wyoming Province entirely (Figure 1.10D and the aulacogen of the Black Hills had developed into a substantial northwest striking seaway, by partial separation of the Wyoming Province from the Canadian Shield. Extensive carbonate banks developed on the continental margin of southeastern Wyoming and along similar rifted-margins in Minnesota and Michigan as well. In the Black Hills, deep water facies marked by turbidite structures in graywacke, subgraywacke, and shale were deposited.

By about 1900 m.y., island arcs were forming south of the Cheyenne Belt while marine deposition continued on the continental margin. Ultimately the island arcs began to move or be subducted towards (southeast dipping subduction zones) the Atlantic-type margin of southeastern Wyoming (Figure 1.10D). As the island arcs are moved toward the continent, marine successions of the Upper Libby Creek Group were thrust over rocks of the Lower Libby Creek Group and a series of parallel thrusts were formed in the Libby Creek Group of the Sierra Madre and Medicine Bow Mountains (Figure 1.10E). A similar event probably was taking place in northeastern Utah (Graff and others, 1980). Ultimately the island arcs were partially thrust over the continent and welded to the continent resulting in severe deformation and metamorphism and juxtaposition of two very different Proterozoic metasedimentary terranes (Figure 1.10E, F). We believe that the Black Hills aulacogen began to close at about this same time so that there was complex interference of structures developed during closure of the northwest striking Black Hills trough and the collision of island arcs with the east- to northeast-striking continental margin of southeastern Wyoming.

Deformation during collision of the island arc(s) and the Wyoming Province was undoubtedly extremely complex and not confined to the rift area where we see Early Proterozoic metasedimentary rocks preserved today. Available evidence permits our interpretation that initial sub-horizontal thrusting in the miogeocline formed an extensive foreland thrust belt which may have extended hundreds of kilometers onto the Wyoming craton. Then, continued compressional stresses caused rotation of thrusts to steep attitudes in the marginal area and accompanying northeast-trending F_1 folds in the Libby Creek Group (Figure 1.9). This rotation may have been due to a temporary shift in direction of subduction during collision (Figure 1.10E). Rocks of the underlying Deep Lake Group and Phantom Lake Metamorphic Suite were refolded and further compressed at this time (F_3 folding in Deep Lake and Phantom Lake, Figure 1.9). The metasedimentary rocks of the Deep Lake Group and Libby Creek Group were metamorphosed to greenschist facies during this episode, about 1700 m.y. ago, as indicated by 1700 m.y. metamorphic dates for the Lookout Schist of the Lower Libby Creek Group of the Medicine Bow Mountains (Hills and others, 1968).

In the Sierra Madre and Medicine Bow Mountains, as well as elsewhere in the Wyoming Province (Wind River Mountains, Worl, 1968; Owl Creek Mountains, Gliozzi, 1967; Black Hills, Kleinkopf and Redden, 1975), late folds with northwest trends and late penetrative S-surfaces that strike northwest are superimposed on earlier structures. This northwest-trending deformation (F_4/F_2 folding in Deep Lake Group/Libby Creek Group of Figure 1.9) must have taken place prior to about 1740 m.y., if it is synchronous with Black Hills folding, because the metasedimentary and metavolcanic rocks of the Black Hills that have northwest fold axes are cut by Harney Peak Granite dated as about 1740 m.y. (Riley, 1970). The northwest oriented F_4/F_2 folds are best developed in zones which are separated from each other by large blocks relatively unaffected by the event. In the Medicine Bow Mountains northwest-striking penetrative surfaces are best developed in the northeastern margin of the mountains and are coincidental with and probably controlled the location of the northwest-striking Arlington fault of Laramide age. The development of these late northwest folds and S-surfaces is believed to be related to closing of the Black Hills aulacogen which may have been triggered by collision of island arcs against the Wyoming Province. This correlation might depend on the angle of approach of island arcs to the continent or the configuration of the continent at time of collision -- both unknown.

Following the development of the northwest structures north of the Cheyenne Belt, there appears to have been a change from compressive stresses to shear stresses across the Cheyenne Belt. F_5/F_3 folds in the Deep Lake

Group/Libby Creek Group have vertical fold axes and are presumed to represent rotational forces and these appear to have formed during intense cataclasis and shearing in the Cheyenne Belt shear zones. These deformations appear to reflect initiation of transform (left-lateral) motion along the plate suture. Granitic and gabbroic intrusives invade the shear zones and local melting occurred in the shear zone to develop migmatite. In the Sierra Madre, the Sierra Madre granite of Divis (1977) cuts across the sheared rocks of the Cheyenne Belt and is itself sheared. This granite is dated as about 1635 m.y. (Hills and Houston, 1979) and about 1645 m.y. (Divis, 1977) and is thought to have been formed during the late stages of the collision event. If so, it suggests that most of the collisional orogenesis took place before about 1640 m.y.

The Mullen Creek-Nash Fork shear zone of the Medicine Bow Mountains, which is the major manifestation of the Cheyenne Belt, can be traced northeast to the Laramie Range where it crops out in the Richeau Hills then to an area south of the Hartville Uplift where sheared Precambrian rocks are brought up along Tertiary normal faults. The Mullen Creek-Nash Fork shear zone can also be traced southwest into Sierra Madre where it changes strike to a more westerly trend. This shear zone was probably the site of strike-slip movement in the late stages of collision, as discussed above, but it also may have continued to move later in the Proterozoic as a transform fault (Warner, 1978), after the collision. In any event, the Proterozoic movement on this fault is probably the last event of a fascinating Late Archean-Early Proterozoic history of southeastern Wyoming.

The above sequence of events suggests that at one time much of the southeastern part of the Wyoming Province was covered by sedimentary rocks of Early Proterozoic age. It suggests that an Early Proterozoic landmass of about 2400 m.y. had a series of braided streams and rivers that flowed across a deeply weathered Archean terrane of relatively low relief and occupied graben-like rift valley basins. It implies that uranium-thorium bearing quartz-pebble conglomerate might have been deposited over a far greater area than the three localities in which they have been found. From an explorationists viewpoint this seems promising, but there are two lines of evidence that argue against it.

(1) The best chance of preservation of sedimentary and volcanic rocks of Early Proterozoic age that have been exposed to erosion periodically for the last 1600 m.y. would be in rifts, grabens, or infolded synclines. These structures are probably uncommon, and their most probable occurrence is near a continental margin where they have already been located.

(2) The geochronological boundary in the Archean terrane of southeast Wyoming, as defined by Hills and Armstrong (1974) and by Peterman and Hildreth (1978), is a line extending roughly eastward from the southern Wind River Mountains to the northern Laramie Range (Peterman and Hildreth, 1978, p. 18, fig. 8) south of which mineral ages decrease to between 1400 m.y. and 1600 m.y. Peterman and Hildreth (1978, p. 20) suggest that this age pattern could have been generated by vertical uplift of this southern block between 1400-1600 m.y. The area affected by this "uplift" may have been the area covered by a foreland thrust belt containing Early Proterozoic metasedimentary rocks. The total thickness of metasedimentary rocks of Early Proterozoic age in the Medicine Bow Mountains is on the order of 10 kilometers. The metasedimentary rocks of the craton were probably less than one-half this thickness but would have been tectonically thickened by thrust faulting. Peterman and Hildreth (1978) suggested that an uplift of

ten kilometers or so was required to cause a 300° C. temperature change of minerals and thus alter mineral ages downward. This is about the right amount of erosion to remove the entire Early Proterozoic cover of the southern Wyoming Province craton and is indirect support for total removal of Early Proterozoic sediments of southern Wyoming, except in cratonic margin areas.

If Early Proterozoic sedimentary rocks were deposited north of the geochronological boundary in the central Wyoming Province their potential of preservation might be greater, but unfortunately this area has relatively few Late Archean granites that are the best source of uranium and thorium minerals. Thus it appears that the Wyoming Province is not a promising area to prospect for additional uranium-thorium bearing quartz-pebble conglomerate of Early Proterozoic age.

DISTRIBUTION, STRUCTURE AND U, Th ASSAYS OF RADIOACTIVE CONGLOMERATES

Plate 5 shows the distribution of radioactive conglomerates in the Sierra Madre (Deep Gulch Conglomerate of the Jack Creek Quartzite and Magnolia Formation) and Plate 1 shows the distribution of radioactive conglomerates in the Medicine Bow Mountains. In both ranges, radioactive conglomerates occur in fluvial metasedimentary rocks of Late Archean and Early Proterozoic age.

Uranium- and Thorium-Bearing Conglomerates of Archean Age

Sierra Madre

In the Sierra Madre, uranium- and thorium-bearing conglomerates of Late Archean age occur in the Jack Creek Quartzite of the Phantom Lake Metamorphic Suite, which crops out in the central and northwestern Sierra

Madre (Plate 5). The Jack Creek Quartzite is the basal subdivision of the Phantom Lake Metamorphic Suite (Figure 1.6) and is in the hinge of an overturned and faulted synclinorium that has been traced throughout the northern and western Sierra Madre. The structural interpretation of Plates 5 and 6 shows a complex synclinorium (the Rudefeha Synclinorium of Plate 6) with an east-striking axial plane overturned and dipping south at a low angle in the central Sierra Madre. The Jack Creek Quartzite is exposed on the north limb of the synclinorium in the central Sierra Madre and is covered on the south, where the Early Proterozoic Magnolia Formation lies unconformably on rocks of the Phantom Lake Metamorphic Suite and covers much of the south limb of the synclinorium. In Section 27, T. 15 N., R. 87 W., the axial plane of the synclinorium is rotated so that it strikes north and here both limbs of the overturned syncline are exposed and the Jack Creek Quartzite is exposed on both the western and eastern limb of the fold (Plate 5). The eastern limb of the fold in Section 26, T. 15 N., R. 87 W. contains several intralimb folds in the Jack Creek Quartzite; the western limb is partly engulfed by a large gabbroic sill (Plate 5).

In Section 22, T. 15 N., R. 87 W. the trace of the axial plane of the synclinorium is offset by an east-west fault, the Vulcan Mountain fault of Plate 6, which is interpreted to be a north-verging thrust fault which disrupts the continuity of the major synclinorium (Plate 5). The synclinorium reappears north of the fault (now called the Divide Peak Syncinlorium in Plate 6) in Section 18, T. 15 N., R. 87 W. where the axial plane strikes northeast and is overturned toward the southeast (Plate 5).

The Jack Creek Quartzite is exposed in the northwest and southeast limbs of the synclinorium and the southeast limb of the synclinorium has four intralimb folds that cause repetition of the Jack Creek Quartzite (Plate 5).

Farther northwest, the trace of the axial plane of the synclinorium is again offset by a left lateral fault of major proportions (the Savery Creek fault) that strikes northwest and extends some 30 km from Section 36, T. 16 N., R. 88 W. in the northwest to Section 22, T. 14 N., R. 85 W. in the southeast, where it runs into the Quartzite Peak fault. The axial plane of the synclinorium reappears on the northeast side of the fault in NE½ Section 6, T. 15 N., R. 87 W. where it strikes northeast and is overturned toward the southeast. In this northernmost exposure of the Divide Peak synclinorium, the overturned fold gradually changes strike to east-northeast in Section 26, T. 16 N., R. 87 W. (Plate 5).

The complexity of the synclinorium is best shown in this northern outcrop area where the northwest limb is faulted (northwest side up) and the southwest limb is thrown into a series of folds that result in repetition of the Jack Creek Quartzite (Plate 5). This compound folding and faulting shown in the northern outcrop area probably is present in other parts of the synclinorium, but indications of stratigraphic tops are lacking and it has not been documented.

Radioactive, uranium- and thorium-bearing quartz-pebble conglomerates have been recognized in several places in the northwestern Sierra Madre. Radioactive beds of the Deep Gulch Conglomerate can be traced for about 2 km at the Carrico Ranch (Sec. 12, T. 15 N., R. 83 W. and Secs. 6, 7, T. 15 N., R. 87 W), about 1.5 km at Deep Gulch (Sec. 36, T. 16 N., R. 88 W.)

and about 1.5 km at the Manning Ranch (Secs. 29, 30, T. 16 N., R. 87 W). This represents a 7.5 km long, semi-continuous trend of the Deep Gulch Conglomerate. Surface assays for U and Th range up to 72 ppm U and 1100 ppm Th at the Carrico Ranch; 205 ppm U and 840 ppm Th at Deep Gulch; and 21 ppm U and 410 ppm Th at the Manning Ranch. Drilling yielded up to 720 ppm U and 2600 ppm Th at Carrico Ranch (Sm-2); up to 490 ppm U and 483 ppm Th at Deep Gulch (JP-1); and 48 ppm U and 653 ppm Th at Manning Ranch (JP-4). Weakly radioactive outcrops of the Deep Gulch Conglomerate occur on the north side of a fault in Sec. 19, T. 16 N., R. 87 W.; in isolated outcrops in the N½, Sec. 4, T. 15 N., R. 87 W.; and at the extreme eastern limit of outcrop of the Jack Creek Quartzite in the Divide Peak synclinorium, Sec. 35, T. 16 N., R. 87 W. (up to 14 ppm U and 130 ppm Th).

Strongly radioactive conglomerates also occur in the upper Jack Creek Quartzite near Dexter Peak. One outcrop of quartz-pebble conglomerate near the top of Dexter Peak, Sec. 21, T. 15 N., R. 87 W., yielded up to 131 ppm U. However, two drill holes spotted south, along strike, yielded maximum values of only 8.2 ppm U and 66 ppm Th. Mildly radioactive conglomerates (11 ppm U and 23 ppm Th) of the Jack Creek Quartzite also crop out in Sec. 25, T. 15 N., R. 87 W., but maximum values in the subsurface (SM-6) were only 31 ppm U and 24 ppm Th.

Local beds of radioactive conglomerate are also present in units that overlie the Jack Creek Quartzite. Paraconglomerates of the Silver Lake Metavolcanic Rocks contain thin zones of quartz-pebble conglomerate which appear to be a better-sorted facies of the paraconglomerates. These occur in the SE¼, Sec. 27, T. 15 N., R. 87 W.; in the S½, Sec. 31, T. 15 N., R. 87 W.; in the SW¼, Sec. 11, T. 14 N., R. 86 W.; and in drill hole SM-9

(0.8 ppm U, 9 ppm Th). Pyritic quartz-pebble conglomerate was also intersected at the base of the Bridger Peak Quartzite in SM-9 and this conglomerate yielded 35 ppm U and 40 ppm Th.

Medicine Bow Mountains

In the Medicine Bow Mountains radioactive quartz-pebble conglomerate beds of Late Archean age occur in the complexly folded units of the Stud Creek Volcaniclastic Rocks and Rock Mountain Conglomerate of the Phantom Lake Metamorphic Suite. Mildly radioactive conglomerates are also locally present in the Colberg Metavolcanic Rocks and in the Conical Peak Quartzite which are the uppermost subdivisions of the Phantom Lake Metamorphic Suite.

The Stud Creek Volcaniclastic Rocks contains thin beds of arkosic paraconglomerate and quartz-granule conglomerate interbedded with the volcaniclastic rocks. The most extensive bed (about 1 km long and 100 m wide) is located on the north side of Stud Creek, in Secs. 10, 15, T. 18 N., R. 79 W. These rocks contain 20 ppm U and 37 ppm Th. Another radioactive conglomerate found in the Stud Creek Volcaniclastic Rocks was a one-meter thick paraconglomerate in SE¼ Sec. 11, T. 18 N., R. 79 W. which contained 6.3 ppm U and 58 ppm Th.

More strongly radioactive conglomerates are found in the overlying Rock Mountain Conglomerate, which crops out in two main areas in the northern Medicine Bow Mountains. Slightly radioactive (3 times background) quartz-granule conglomerates of the Rock Mountain Conglomerate crop out in a northeast-striking layer from the East Fork of Wagonhound Creek (NE¼, Sec. 5, T. 18 N., R. 79 W.) to the west ½ Sec. 27, T. 19 N., R. 79 W., a distance of about 3 km (Plate 1). Maximum grades from this zone were less than 5 ppm U and Th. The second outcrop area, near Rock Mountain (SW¼, sec. 1, T. 18 N., R. 79 W.), is more promising. These radioactive beds occupy an area of about one km^2 in the S½, Sec. 2, T. 18 N., R. 79 W. Most of the conglomerate outcrops are only mildly radioactive (10 ppm U and 10 ppm Th) but some layers contain up to 270 ppm U and 95 ppm Th in outcrop. However, these zones are quite lenticular and thin (centimeters thick) and two drill holes, MB-10 and MB-15, failed to intersect strongly radioactive rocks. Maximum subsurface values were 190 ppm U and 86 ppm Th in MB-15.

Radioactive conglomerates in the Colberg Metavolcanic Rocks are like those found in the Silver Lake Metavolcanic Rocks of the Sierra Madre. They are the cleaner and better-sorted facies of the paraconglomerate. The best example is a quartz-pebble conglomerate in a prospect pit in SW¼, Sec. 10, T. 18 N., R. 79 W. which contain 3 ppm U and 17ppm Th. A more weakly radioactive rock is the Colberg Metavolcanic Rocks in a quartz-granule conglomerate in SW¼, Sec. 26, T. 19 N., R. 79 W. which contains 2.6 ppm U and 14 ppm Th.

Slightly radioactive quartzites in the Conical Peak Quartzite crop out in NE¼, Sec. 14, T. 17 N.,R. 80 W. These coarse-grained quartzites occupy small troughs and scours and contain up to 1 ppm U and 4 ppm Th in outcrop. These rocks are not the upper part of a fluvial succession because MB-13 drilled a thick section of non-radioactive fine-grained quartzites in this area. They may be coarser-grained beach sands, within the shallow marine Conical Peak Quartzite. Radioactive beds of uncertain origin are also present in the Conical Peak Quartzite in the SE¼, Sec. 1, T. 17 N., R. 80 W.

Uranium - and Thorium-Bearing Conglomerates of Early Proterozoic Age

Sierra Madre

Radioactive quartz-pebble conglomerate beds are present locally in basal beds of the Early Proterozoic Magnolia Formation of the Deep Lake Group in the Sierra Madre. The Magnolia Formation is not as well defined in the Sierra Madre as in the Medicine Bow Mountains. It crops out mainly in the western part of the Sierra Madre where Magnolia beds are interpreted to lie unconformably on the Phantom Lake Metamorphic Suite. The presence of an unconformity cannot be verified at the west margin of the Sierra Madre but along the section line between Sec. 34, T. 15 N., R. 87 W. and Sec. 27, T. 15 N., R. 87 W. (Plate 5) beds of the Magnolia Formation strike east almost at right angles to the trace of the axial plane of the overturned Rudefeha synclinorium of the Phantom Lake Metamorphic Suite (see Graff, 1978 for an alternative interpretation of this structure). The Magnolia Formation and younger formations of the Deep Lake Group are folded into broad, northeast-trending folds in the SE parts of T. 15 N., R. 87 W. and the NW parts of T. 14 N., R. 87 W. (Plate 5), and are severely disrupted by a large gabbroic intrusion. The Magnolia Formation can be mapped to Sec. 18, T. 14 N., R. 86 W. (Plate 5) where the strike of the beds begins to change from northwest to east-west and the Magnolia Formation then is mapped on a consistent east strike for a distance of about 20 km, almost to the east limit of outcrop of metasedimentary rocks in the Sierra Madre.

Radioactive quartz-pebble conglomerate beds in the Magnolia Formation are best developed in the western outcrop area, in NE¼ Sec. 33, T. 15 N., R. 87 W.; SW¼ Sec. 27, T. 15 N., R. 87 W.; N½ Sec. 34, T. 15 N.,

R. 87 W.; and in SE¼ Sec. 35, T. 15 N., R. 87 W., but even in these localities conglomerate beds are lenticular and only weakly radioactive and two drill holes (SM-5 and SM-11) failed to detect significant mineralization in the conglomerate layers intersected at depth (maximum values are 3 ppm U, 7 ppm Th in outcrop and 27 ppm U, 110 ppm Th in drill core). East of these outcrops, the Magnolia Formation is poorly exposed but in the NE¼ Sec. 16, T. 14 N., R. 85 W. beds of slightly radioactive (3 ppm U, 33 ppm Th in outcrop) quartz-pebble conglomerate are exposed and east of this outcrop, conglomerate layers have been mapped for about 6 km to Sec. 18, T. 14 N., R. 84 W. These conglomerate beds are finer grained and more feldspathic than beds in the west and two drill holes (SM-12 and SM-8) that intersected these conglomerate layers at depth did not reveal significant mineralization (up to 3 ppm U, 33 ppm Th).

The above interpretation, that a basal Proterozoic Magnolia Formation unconformably overlies the Phantom Lake Suite, is one that attempts to bring the better defined stratigraphy of the Medicine Bow Mountains into the Sierra Madre. We believe that there is good justification for the correlation between ranges because beds of the Deep Lake Group and Libby Creek Group that overlie the Magnolia Formation of the Sierra Madre can be correlated with equivalent units in the Medicine Bow Mountains with a high degree of confidence. It is possible, however, that the Magnolia Formation of the Medicine Bow Mountains is missing in the Sierra Madre and that beds we have called Magnolia in the Sierra Madre are actually part of the Jack Creek Quartzite and Silver Lake Metavolcanic Rocks of the Phantom Lake Metamorphic Suite (this is a preferred interpretation of Douglas Charlton of Resource Associates of Alaska who had done detailed mapping in this area). Additional detailed mapping with particular attention to topping criteria might help solve this problem. However, from an economic viewpoint, this does not appear to be a critical problem inasmuch as no conglomerates of economic interest have been detected in this unit.

Medicine Bow Mountains

In the Medicine Bow Mountains radioactive quartz-pebble conglomerates of Early Proterozoic age have been found in the Magnolia Formation, Lindsey Quartzite, and Cascade Quartzite of the Deep Lake Group and in the Medicine Peak Quartzite of the Libby Creek Group.

The distribution of the Magnolia Formation is shown in Plate 1. In the central Medicine Bow Mountains, T. 16 N., R. 80 and 81 W., the Magnolia Formation crops out in a northeast-trending anticline along Arrastre Creek, where it lies unconformably on the Phantom Lake Metamorphic Suite, and in a faulted syncline near the confluence of Little Brush Creek and Brush Creek. Outcrops of the Magnolia Formation are quite poor in these areas, but the best and most continuous outcrops are north of Arrastre Lake (Sec. 10, T. 16 N., R. 80 W). Here lenticular beds of radioactive quartz-pebble conglomerate occur within a thick section of arkosic paraconglomerates near the base of the Magnolia Formation. This zone of paraconglomerate can be traced laterally about 1000 m on the east limb of the anticline but does not crop out on the west limb. Maximum surface values are 8 ppm U and 38 ppm Th. A private company drilled one hole to try to intersect these conglomerate beds below the zone of weathering (PL-1) and failed to encounter radioactive quartz-pebble conglomerate at depth (maximum values were 11 ppm U and 30 ppm Th). Another drill hole (GH-1) was drilled by a company in the private sector, on the northwest flank of the Arrastre anticline (W½ Sec. 3, T. 16 N., R. 80 W.) to a depth of 1212 feet, but this hole failed to penetrate the basal beds of the Magnolia Formation and did not encounter radioactive quartz-pebble conglomerate (maximum values were 32 ppm U and 48 ppm Th). Two holes were also drilled in the syncline near Brush Creek (Secs. 14, 23, T. 16N., R. 81 W). One, MB-9R, encountered thick sections of mildly radioactive arkosic paraconglomerate (maximum values of 78 ppm U, 52 ppm Th). The other hole, MB-17, had caving problems due to the thick glacial overburden and did not reach conglomerates of the Magnolia Formation.

The Magnolia Formation also crops out over large areas in the central Medicine Bow Mountains, in T. 17 N., R. 79 W. (Plate 1). Outcrops are poor throughout the central Medicine Bow Mountains but there are several areas where basal Magnolia beds are exposed. In Sec. 22, T. 17 N., R. 79 W., quartz-pebble conglomerates overlie metabasalts of the Phantom Lake Metamorphic Suite. These basal beds contain up to 3.2 ppm U and 13 ppm Th in outcrop and 14 ppm U and 36 ppm Th in drill hole MB-11. In the NE¼, Sec. 14, T. 17N., R. 80 W., similar mildly radioactive quartz-granule conglomerates contain 1.8 ppm U and 16 ppm Th in outcrop. In the Crater Lake area (Sec. 35, T. 18 N., R. 79 W.), quartz-granule conglomerates containing 1 ppm U and 26 ppm Th unconformably overlie the Colberg Metavolcanic Rocks. In general, the basal Magnolia Formation in the Central Medicine Bow Mountains is a quartz-granule conglomerate of the Quartzite Member and the Conglomerate Member does not appear. Thus, this area is of limited economic interest for uranium-bearing conglomerates.

A complex synclinorium is exposed in the northern part of T. 17 N., R. 78 and 79 W. and in T. 18 N., R. 78 and 79 W. (Plate 1). This synclinorium is disrupted by faulting and its east limb is mostly removed by faults. However, a small piece of the Magnolia Formation is preserved on the east limb in Secs. 28 and 33, T. 18 N., R. 78 W. In this area, the Magnolia Formation contains a few quartz-pebble conglomerates with up to 13 ppm U and 490 ppm Th in outcrop and up to 14 ppm U and 30 ppm Th in drill hole MB-4.

A more complete section of the Magnolia Formation is exposed on the northwest limb of the fold where Magnolia conglomerates uncontormably overlie rocks of the Phantom Lake Metamorphic Suite. Radioactive quartz-pebble conglomerates have been detected on the northwest limb of this syncline from the SE¼ Sec. 5, T. 17 N., R. 79 W., to Sec. 6, T. 18 N., R. 78 W., a distance of 10.5 km. In the southwest, the Magnolia Formation consists of a polymictic arkosic paraconglomerate overlain by quartz-granule conglomerate. Paraconglomerates contain up to 20 ppm uranium and 190 ppm thorium in outcrops. Several holes have been drilled to intersect the conglomerates below the zone of weathering in the general area south of Rock Creek (Plate 1). These are BOWS-1, BOWS-2, and BOWS-3 and MB-5 (Platel). Rocks encountered in these drill holes were, in general, only mildly radioactive (up to 10 ppm U, Th) but values up to 64 ppm U and 27 ppm Th were obtained in BOWS-1 and values up to 110 ppm U and 41 ppm Th were found in BOWS-3. These data suggest that radioactive conglomerates are present in this area but grades are too low and the conglomerates too limited in thickness and extent to warrant further exploration.

North of Rock Creek, the Magnolia Formation is composed of rocks similar to those found south of Rock Creek: a basal paraconglomerate unit and an upper granular quartzite unit. Most of these rocks are only weakly radioactive. However, in the NW¼, Sec. 19, T. 18 N., R. 78 W., strongly radioactive quartz-pebble conglomerates occur in the upper part of the paraconglomerate unit, near the gradational contact with the overlying quartzites. These conglomerates contain up to 20X background radioactivity and surface samples collected in this area contain up to 51 ppm U and 130 ppm Th. EMB-5 and EMB-11 were designed to intersect these radioactive beds below the zone of weathering. EMB-5 encountered a major fault, then was terminated before it reached the target horizons. EMB-11 intersected a number of layers of radioactive quartz-pebble conglomerate that were interbedded with an arkosic and chloritic paraconglomerate containing large clasts (up to 10 cm) of vein quartz, quartzite, pink granite, granite-gneiss, and chlorite schists. Several of these zones contained in excess of 100 ppm U and Th. The subsurface extension of the most radioactive conglomerate found in outcrop contained up to 365 ppm U and 344 ppm Th. This suggests that this radioactive zone is continuous in a down-dip direction for at least 300 m.

North of the above locality (NW¼ Sec. 19, T. 18 N., R. 78 W) the Magnolia Formation is folded and cut by a major sill of gabbro (Plate 7). However the Magnolia Formation is exposed on the east flank of a minor fold between SW¼ Sec. 18, T. 18 N., R. 78 W., and NE¼ Sec. 7, T. 18 N., R. 78 W. (Plate 1). Exposures are not continuous but beds of radioactive conglomerate are exposed on the surface in a number of areas. The average

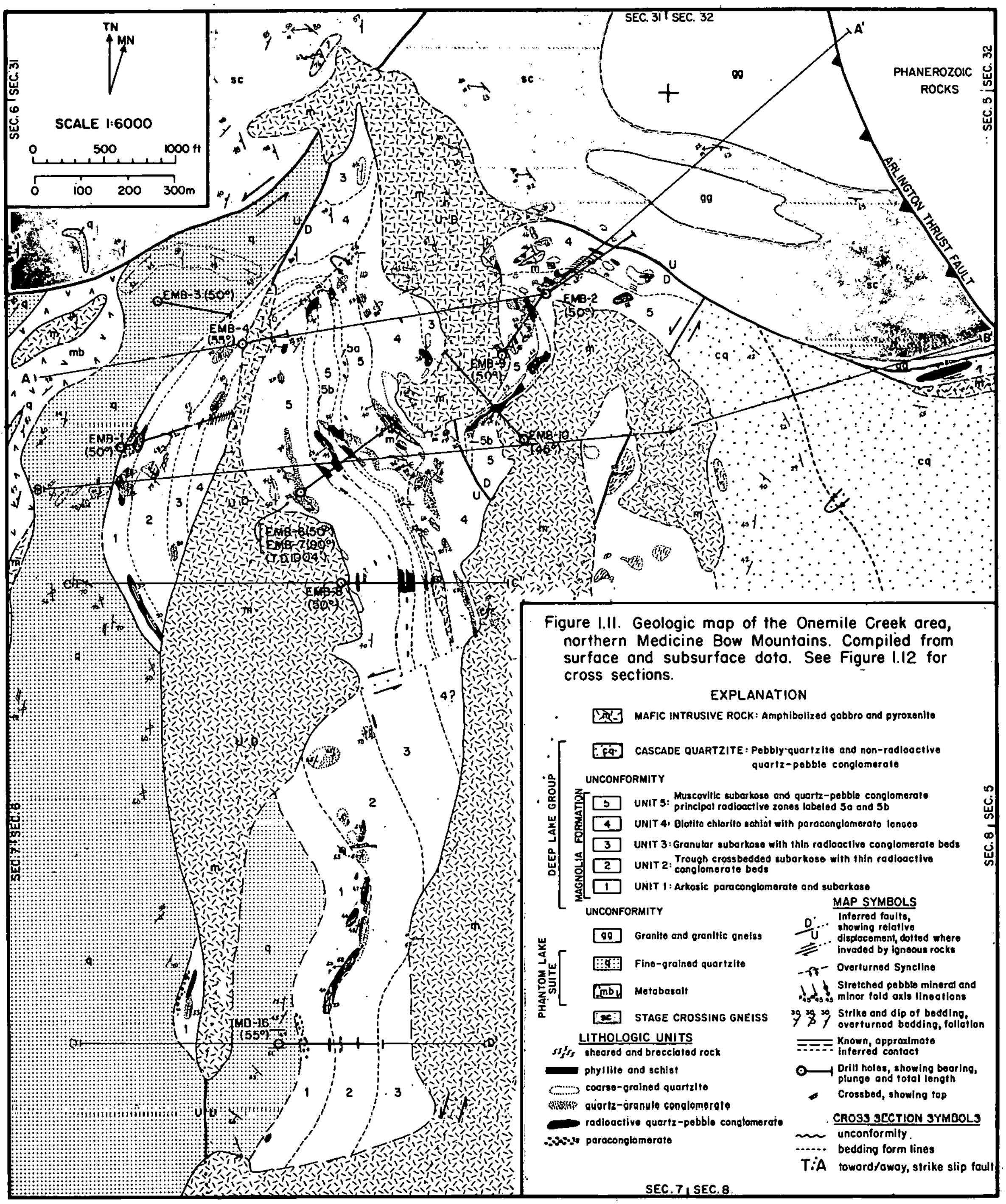


Figure I.II. Geologic map of the Onemile Creek area, northern Medicine Bow Mountains. Compiled from surface and subsurface data. See Figure I.12 for cross sections.

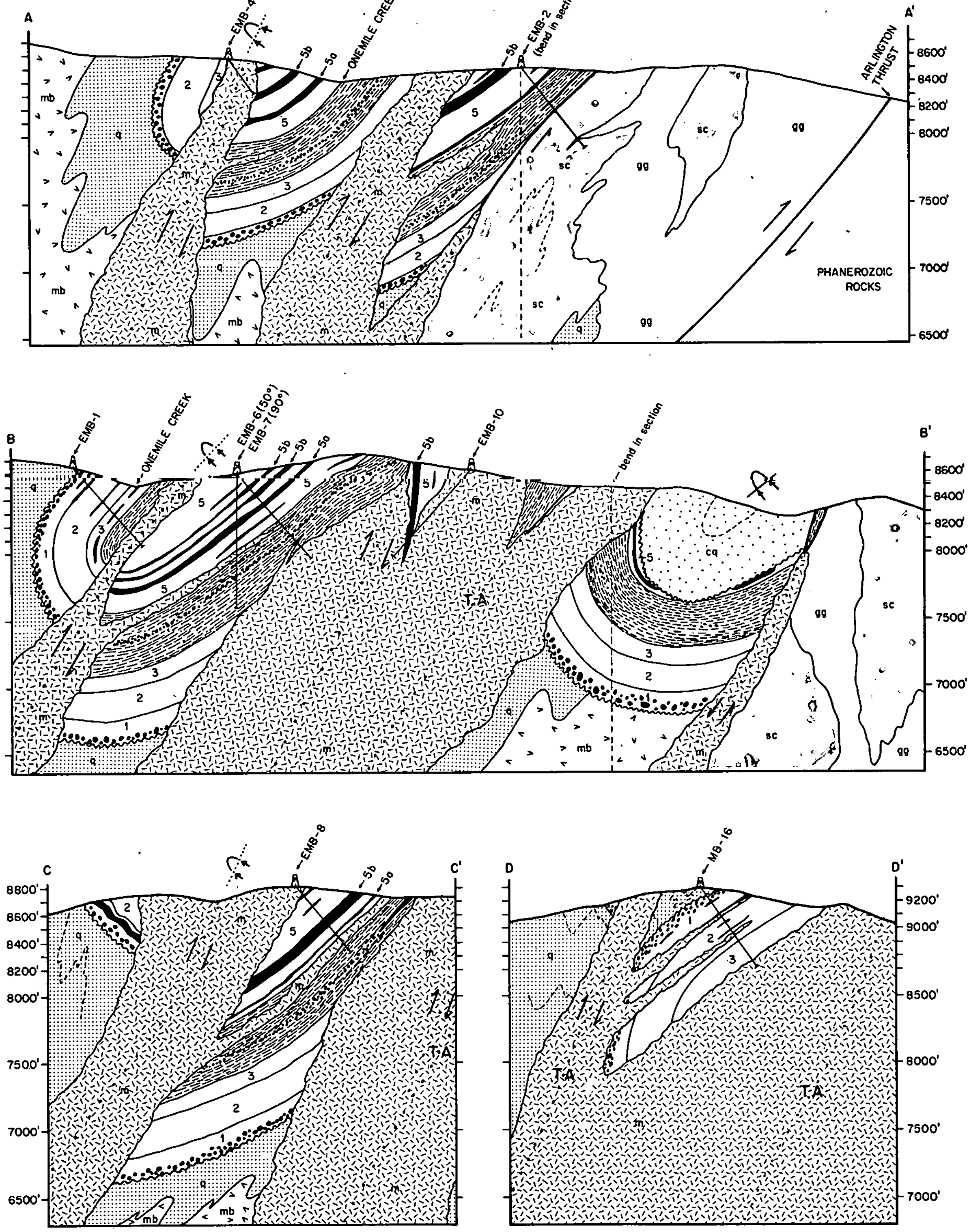


Figure 1.12. Cross sections and drill sections from the Onemile Creek area.

tenor of surface samples of the radioactive quartz-pebble conglomerate of this area is 5 ppm uranium and 10-30 ppm thorium and maximum assays are 21 ppm uranium and 67 ppm thorium. These radioactive quartz-pebble conglomerate beds are less promising than those to the south (and north) and are believed to be too lenticular and too low grade to be of economic interest. No drill holes have been located in this area.

The most promising economically of any radioactive conglomerates in the Medicine Bow Mountains, and in southern Wyoming in general, crop out in the Onemile Creek area, Secs. 6 and 7, T. 18 N., R. 78 W. Figures 1.11 and 1.12 illustrate the distribution of radioactive conglomerate layers of this area. The radioactive conglomerate layers are in the faulted and overturned nose of a tightly folded syncline. The Magnolia Formation has been deformed, at least twice, and metamorphosed to amphibolite facies, but tops of beds can still be recognized and ten holes, nine drilled by a private company and one drilled by Bendix as part of this study (Figure 1.11) allow us to make a more comprehensive analysis of this area than any other in southeast Wyoming.

The radioactive quartz-pebble conglomerate units of this area are in the lower conglomeratic part of the Magnolia Formation. The basal unit is a partly open- and partly closed-framework conglomerate with clasts of pink granite, (up to 35 cm in float), vein quartz, quartzite, and rare chlorite schist in an arkosic matrix. This is overlain by coarse-grained sericitic quartzite with local thin layers of quartz-pebble conglomerate. Next is a phyllite with layers of paraconglomerate, and at the top, is a coarse-grained sericitic quartzite with well-developed beds of radioactive quartz-pebble conglomerate. The radioactive quartz-pebble conglomerate beds in the upper unit of the Magnolia Formation in this area are interpreted to be continuous layers through most of this area (Figure 1.11). There are two principal beds of radioactive conglomerate (5a and 5b) that are variable in thickness but may be up to 20 meters thick. 5a and 5b are compound units composed mostly of radioactive quartz-pebble conglomerates but with interlayered quartz-granule conglomerate and coarse-grained quartzite. Typical uranium assays for the radioactive quartz-pebble conglomerates in the upper unit are about 100 ppm and thorium assays are about 80 ppm. Uranium values up to 400 ppm over a thickness of three meters are present in EMB-6. The maximum uranium value in this area is 1620 ppm from a 5 cm zone in MB-16 and the maximum thorium value is 1380 ppm over 20 cm in EMB-6. According to Borgman and others (1980), estimated uranium resources are about 1801 tons U_3O_8 at an average grade of 310 ppm U_3O_8 and estimated thorium resources are about 1106 tons ThO_2 at an average grade of 290 ppm ThO_2. This resource estimate is from geostatistical analysis of core and surface samples and is discussed in detail in Volume 3.

The Lindsey Quartzite and Cascade Quartzite of the Deep Lake Group and the Medicine Peak Quartzite of the Libby Creek Group all contain well-developed quartz-pebble conglomerate beds, but very few of these quartz-pebble conglomerate beds are radioactive. The Lindsey Quartzite has weakly radioactive quartz-pebble conglomerate beds in the northeastern Medicine Bow Mountains. Some of the best developed and most radioactive quartz-pebble conglomerate beds are at the base of fining-upwards sets in the Lindsey Quartzite in exposures in NE¼ Sec. 30, T. 18 N., R. 78 W., on the north side of Rock Creek. This is in the area where the Lindsey Quartzite thins to zero thickness, presumably due to unconformable overlap of the Cascade Quartzite (Plate 1). Maximum values are 24 ppm U and 630 ppm Th in the Lindsey Quartzite in this area. Radioactive conglomerates of the Lindsey also crop out on a cliff face on the west side of the North Fork of Rock Creek (NE¼, Sec. 1, T. 17 N., R. 79 W.). Maximum values obtained from these beds was 6 ppm U and 7 ppm Th.

Radioactive conglomerate beds have been reported in the Cascade Quartzite near Cascade Lake (Sec. 35, T. 17 N., Rn. 80 W) and Rio Amex Company is reported to have drilled in this area, but the results of this drilling program are not available to us, and we have not detected strongly radioactive beds at any locality in the Cascade Quartzite. Rio Amex reported several assays of 3-5 ppm Au from the Cascade.

Quartz-pebble conglomerate beds are common in the middle and upper Medicine Peak Quartzite, but surface outcrops with above normal radioactivity have not been detected. On the other hand, one glacial boulder of quartz-pebble conglomerate from the Medicine Peak Quartzite, collected from an area south of the Gap in the central Medicine Bow Mountains, contained 7.4 ppm uranium and 330 ppm thorium.

MINERALOGY AND PETROLOGY OF URANIUM- AND THORIUM-BEARING UNITS

Uranium- and thorium-bearing units inthe Sierra Madre and Medicine Bow Mountains include quartz-pebble conglomerates, arkosic paraconglomerates, and subarkosic to arkosic granular quartzites. The most radioactive rocks in the Sierra Madre are subarkosic moscovitic quartz-pebble conglomerates of Unit 3 of the Deep Gulch Conglomerate (Plate 7) and the most radioactive rocks in the Medicine Bow Mountains are subarkosic, muscovitic small-pebble (quartz, granite, and quartzite) conglomerates of Unit 5 of the Magnolia Formation (Plate 4). Although these beds differ in age, the major constituents are basically the same (except granite fragments are rare in the Sierra Madre). These constituents include quartz, rock fragments of granite and quartzite, K-feldspar, plagioclase, muscovite, chlorite, biotite, pyrite and a heavy mineral suite (which is discussed in detail later). Arkosic and subarkosic quartzites were originally bimodal, argillaceous sandstones and quartz-pebble conglomerates were originally trimodal, argillaceous conglomerates. Tables 1.2 and 1.3 show modal analyses of the sand and granule size fractions from the main radioactive units in the Medicine Bow Mountains and Sierra Madre.

Pebbles in the conglomerates are well rounded, generally moderately sorted, and tightly packed. The most radioactive conglomerates appear to be pebble-supported in both ranges, although stretching of pebbles in the Onemile Creek area of the Medicine Bow Mountains makes it difficult to decipher original packing densities. Clasts in the Deep Gulch Conglomerate are entirely quartz and quartzite, with an average size range of 0.7 to 3.7 and maximum size of 5 cm. Clasts in Unit 5 of the Magnolia Formation are quartz, quartzite, and granite. They range in size

TABLE I.2 SUMMARY OF THE PETROGRAPHY OF THE MAGNOLIA FORMATION ARRANGED APPROXIMATELY NORTHEAST (TOP) TO SOUTHWEST (BOTTOM).

MAGNOLIA FORMATION CONGLOMERATE MEMBER

	Qtz.	Qtzte.	K-spar	Plag.	Granite	Musc.	Chlor.	Biot.	Pyrite	Hem	Heavies
ONEMILE CREEK AREA											
Unit 1, mean and range	52	1.5	1.5	1.5	3.8	24	.7	10	–	5	
of 9 samples	14-70	0-10	0-14	0-10	0-25	5-47	0-7	0-70	–	0-6.4	
Unit 2, mean and range	51	2	8	7	5	25	.1	.7	.6	.7	
of 7 samples	30-62	0-10	3-25	0-20	0-20	5-45	0-.5	0-3	0-4	0-3	
Unit 3, mean and range	57	included	11	6	–	21	5	–	–	–	
of 2 samples	46-67	w/qtz.	6-16	5-6	0-.75	11-31	1-8		–	–	
Unit 4, mean and range	71	4	2.7	.3	–	17	–	–	5	–	
of 4 samples	63-82	0-8	2-4	0-1	–	10-25	–	–	0-20	–	
Unit 5, mean and range	58	1.4	1.3	1	2.8	26	1.5	.4	7.6	–	Apatite
of 43 samples	27-80	0-16	0-15	0-6	0-32	4-50	0-15	0-14	0-30	–	Zircon
Grand Mean of 65 samples	57.9	1.6	1.7	1.8	2.9	24.9	1.2	.26	5.4	.1	
THREEMILE CREEK AREA (EMB-5 & 11)											
Mean and range of	48	10	3.3	2.3	3.3	23.3	2.4	4.2	.6	1.4	1.2 Garnet
17 samples	10-68	0-23	0-13	0-10	0-24	0-75	0-10	0-20	0-5	0-20	Zircon Tourmaline
ARRASTRE ANTICLINE AREA (PL-1)											
Mean and range of	65	6.5	–	.3	.1	21.4	4.1	1	1	–	.5 Zircon
10 samples	45-82	0-25	–	0-3	0-1	1-30	0-15	0-.5	0-5	–	Amphibole
BRUSH CREEK AREA (MB-9)											
Mean and range of	58	4	1.3	2.6	–	16	5.6	2.2	.6	–	Garnet
3 samples	45-62	0-20	0-5	0-5	–	0-20	0-10	0-5	0-2	–	

MAGNOLIA FORMATION QUARTZITE MEMBER

	Qtz.	Qtzte.	K-spar	Plag.	Granite	Musc.	Chlor.	Biot.	Pyrite	Hem	Heavies
THREEMILE CREEK AREA (EMB-11)											
Mean and range of	51	4	4	8	3	20.5	2.1	2	5.4	–	Zircon
5 samples	20-76	0-20	0-10	0-20	0-10	5-50	0-5	0-5	0-15	–	Apatite
MB-4 AREA											
Mean and range of	70	4	9.31	4.3	.6	8.	.6	1.2	–	.4	Zircon Amphibole
8 samples	37-90	0-20	0-20	0-13	0-5	2-15	0-7	0-5	–	0-2	Apatite
NORTH FORK ROCK CREEK CRATER LAKE AREA (ME-11)											
Mean and range of	59	4.3	1.6	4.5	1	12.	1.8	.1.5	–	.9	Amphibole
10 samples	44-93	0-15	0-10	0-10	0-5	3-25	0-5	0-7	–	0-2	Calcite Apatite Zircon
ARRASTRE ANTICLINE AREA (PL-1 & GH-1)											
Mean and range of	56	5	3	8	.3	11.3	3.3	2	.3	.8	Tourmaline
8 samples	54-82	0-20	0-7	0-30	0-2	0-25	0-10	0-5	0-2	0-2	Zircon Calcite

Modal percentages represent entirety of poorly sorted rocks, granule and subgranule matrix for Bimodal conglomerates.

TABLE I.3. PETROGRAPHY OF UNIT 3, DEEP GULCH CONGLOMERATE. PERCENTAGES ARE FROM POINT-COUNTED THIN SECTIONS OF QUARTZ-PEBBLE CONGLOMERATES.

Drill Hole Depth	Qtz.	Qtzite	Plag.	Perthite	Micro-cline	Ortho-clase	Musc.	Chlor.	Biot.	Opaq.	Zircon
SM-1 175.4	61	6	–	1	7	3	23	–	–	1	–
SM-1 175.7	45	24	–	–	10	3	10	–	–	7	1
SM-1 192.8	67	1	–	9	3	–	17	–	–	2	–
SM-1 207	66	6	Tr	8	–	11	7	–	–	Tr	Tr
SM-1A 541.6	60	13	Tr	9	3	8	6	–	–	1	–
SM-2 149	53	13	2	7	8	12	6	–	–	Tr	Tr
SM-2 150.1	57	10	Tr	Tr	6	6	17	–	–	4	Tr
SM-2 187	63	14	1	10	Tr	3	Tr	9	Tr	Tr	Tr
JP-1 408.7	70	2	Tr	2	21	–	2	–	–	2	–
JP-1 410	56	25	–	–	13	–	3	–	–	5	Tr
JP-1 412	61	15	–	Tr	19	–	5	–	–	1	–
JP-1 416.5	62	6	1	–	23	–	4	–	–	2	Tr
JP-1 421	71	1	1	1	19	–	4	–	–	3	Tr
JP-2 311	61	4	1	Tr	11	9	9	–	–	3	1
JP-4 284	62	9	1	Tr	9	9	11	–	Tr	2	Tr

from granules to boulders 7.5 cm in diameter but are most commonly 1-3 cm in diameter. Many of the conglomerates contain 100 percent quartz pebbles; other contain up to 20 percent quartzite and granite pebbles.

The matrix of the conglomerates is composed of quartz, feldspar, and phyllosilicates. The phyllosilicates, muscovite, chlorite, and biotite, are considered to be metamorphic minerals formed by recrystallization of an argillaceous matrix. However, we recognize that a part of the phyllosilicate fraction could be primary. Micas make up about 25 percent of the matrix of most conglomerates and some samples contain as much as 50 percent. The most radiocative conglomerates tend to be rich in muscovite and sericite, but poor in biotite and chlorite. Individual grains of quartz and feldspar are too deformed to determine their original shape; most are strained and exhibit sutured contacts (Figure 1.13B) and extreme flattening is a feature of the more highly deformed rocks (Figure 1.13C). We believe, however, that most of the sand-sized grains are original and were not formed during metamorphism. Some small clear grains of plagioclase and microcline in the muscovite "matrix" may be metamorphic minerals. Quartz generally exhibits undulatory extinction and some grains have trains of fluid inclusions. K-feldspar occurs as microcline or perthite, in about equal amounts, and individual grains are usually clear although cloudy and altered K-feldspar grains are present in some samples. Plagioclase grains are variable in composition and are commonly more deformed than K-feldspar (Figure 1.13). Pyrite is present in the matrix in highly radioactive conglomerates. In arkosic and subarkosic quartzites, pyrite is present as euhedral grains scattered through the rock and it makes up less than one percent.

Non-Radioactive Heavy Minerals

A number of heavy minerals are concentrated in the matrix of the radioactive conglomerates. Where we have convincing top criteria, the heavy minerals can be shown to be concentrated near the base of conglomerate beds and this is further verified by radioactivity scans of layers which contain high uranium and thorium values. We will divide the heavy mineral suite into non-radioactive and radioactive heavy minerals and discuss the non-radioactive minerals first. We have relied, in part, on the microprobe studies of selected samples of conglomerate from the Onemile Creek area of the Medicine Bow Mountains by Desborough and Sharp (1979, p. 38-39) but Desborough and Sharp are not to be held responsible for our conclusions regarding origin of these heavy minerals.

Pyrite is the most abundant heavy mineral in radioactive conglomerates of both ranges. It is typically found as euhedral grains and in masses of aggregates inlayers with other heavy minerals (Figure 1.14). The general characteristics of the pyrite are like those of other minerals of the conglomerate. In less deformed samples, pyrite is in equant or even subhedral grains whereas in deformed and flattened samples pyrite is stretched and flattened in much the same manner as quartz or feldspar (Figure 1.13). Although uncommon, some pyrite fills fractures and individual grains or aggregates of pyrite may be connected by veinlets (Figure 1.15A). In some pyrite aggregates, and in a few individual grains, the pyrite is subhedral and has a partly rounded shape . These grains appear to have a secondary pyrite overgrowth (Figure 1.15B). This same feature has been noted in some quartz grains in less deformed Sierra Madre specimens. This may be a metamorphic texture or it may be local preservation of an

A

B

C

Figure. 1.13. Photomicrographs in transmitted polarized light, crossed nicols, showing degree of flattening of sulphide (black) and silicate grains (shades of gray) in quartz pebble conglomerate. Black grains are pyrite, gray grains mostly quartz, fiberous grains muscovite and biotite. A. shows pyrite-muscovite layer with subrounded and sutured quartz grains in upper left (quartz grain, upper left is 0.40 mm in diameter). B. shows flattened and sutured quartz and pyrite (pyrite grain upper left is 0.22 mm wide). C. shows extreme flattening; elongate black mass in center is pyrite (0.08 mm wide), irregular black mass in right center is recrystallized mixture of urarothorite and monazite. Samples from Onemile Creek locality, Medicine Bow Mountains.

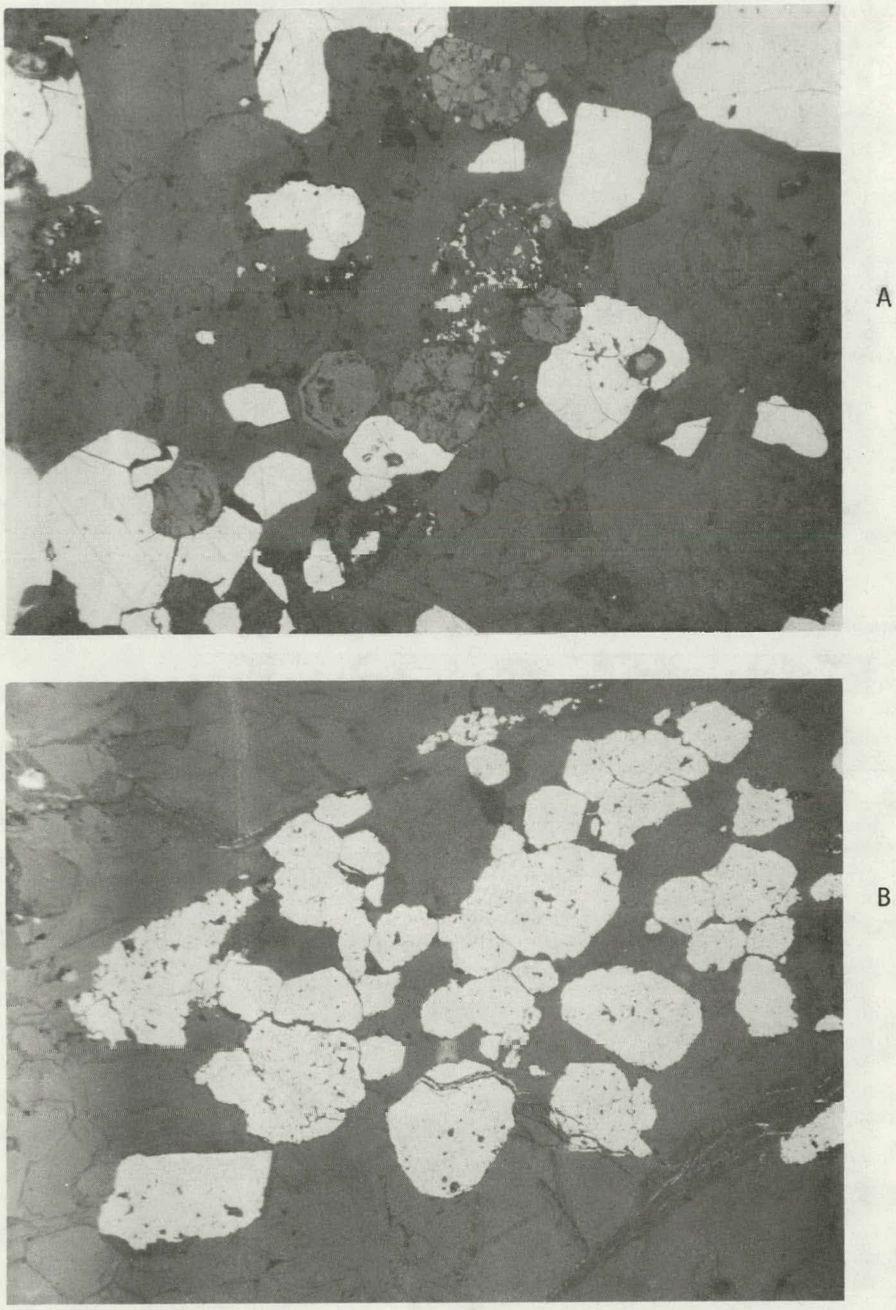


Figure 1.14. A. Photomicrograph showing heavy minerals in quartz pebble conglomerate from Carrico Ranch locality, Sierra Madre. White grains are pyrite, euhedral gray grains showing zoning are zircon, dark gray grains with specks of white pyrite are monazite-huttonite (?) mixtures. Euhedral zircon grain in lower left center of photograph is 0.15 mm in diameter. B. Photomicrograph showing subhedral grains of pyrite (white) in pyrite-rich layer of quartz pebble conglomerate from Onemile Creek locality, Medicine Bow Mountains. Gray grains are silicate minerals--mostly quartz and feldspar. Grain diameters of pyrite are about 0.22 mm.

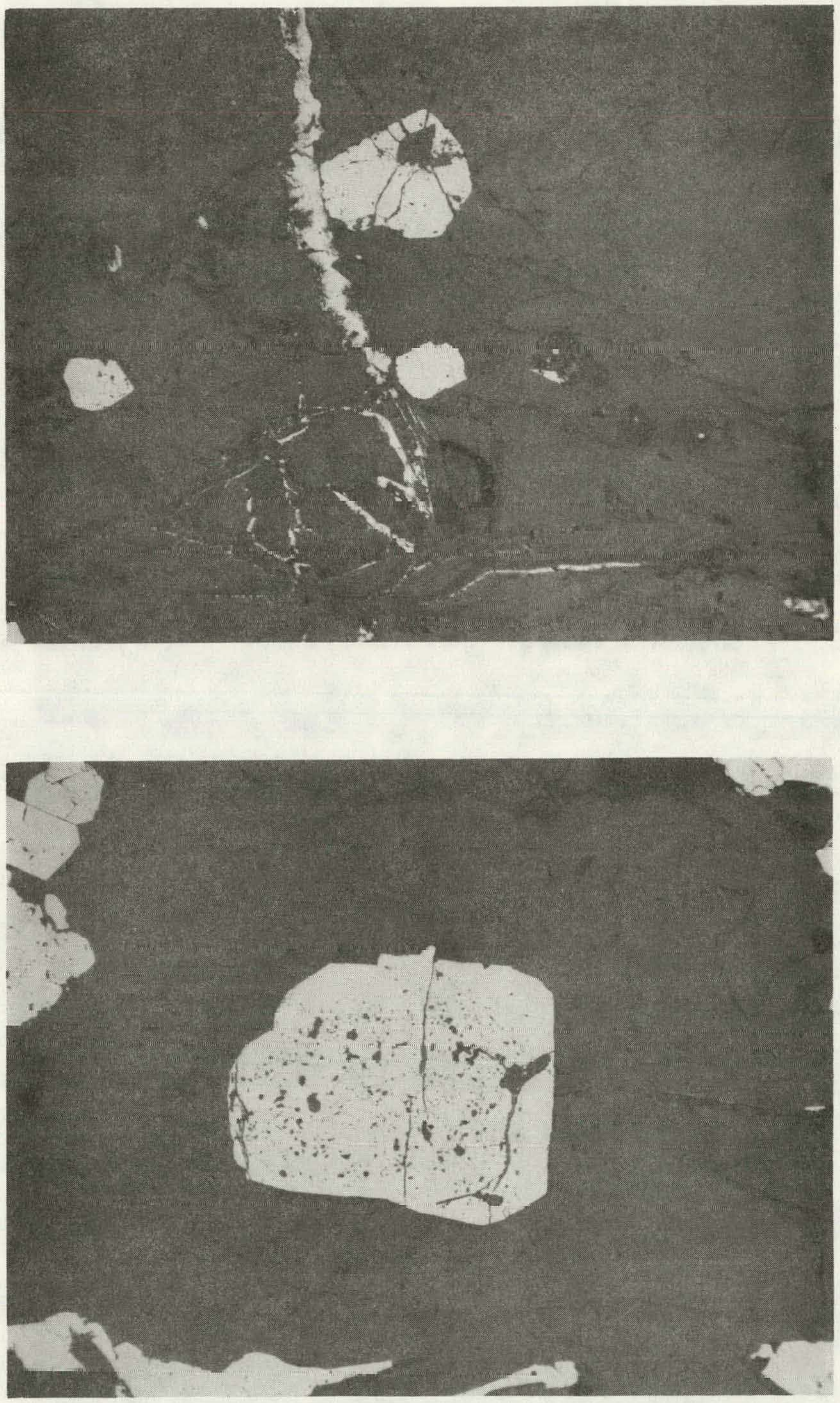


Figure 1.15. A. Photomicrograph showing individual grains and veinlets of pyrite (white) in quartz pebble conglomerate. Length of large pyrite grain is 0.25 mm. B. Pyrite grain showing clear overgrowth of pyrite on subrounded grain with numerous inclusions. Width of pyrite grain is 0.33 mm. Photomicrographs of samples from Onemile Creek locality, Medicine Bow Mountains.

original form. There is, however, no entirely convincing evidence that the pyrite in these samples was originally rounded.

Other non-radioactive heavy minerals are ilmenorutile, apatite, galena, chalcopyrite, bornite, marcasite, sphene, ilmenite, columbite, magnetite, anatase, rutile, and spessartine garnet. There is also a nickel-cobalt-iron sulphide identified by Desborough and Sharp by microprobe study that may be a member of the linnaeite group.

Chalcopyrite and bornite are uncommon and are present as small irregular masses in the muscovite "matrix" or as inclusions in pyrite aggregates and as individual grains attached to pyrite. Marcasite is also sparse and is present as an alteration product of pyrite. Galena is present in two habits: as small masses in pyrite and as crystals attached to pyrite aggregates and as minute crystals in uranium-thorium minerals.

Ilmenorutile is a relatively abundant oxide in the heavy mineral suite and is present in round spherical or elliptical grains with 20 to 30 percent interstitial quartz. As suggested by Desborough and Sharp (1979, p. 38) the ilmenorutile may be a recrystallized Nb-rich detrital mineral, but we believe it has retained its original form. Other oxides that are considered detrital by the authors are magnetite, ilmenite, and columbite. These minerals are very uncommon constituents of the heavy mineral suite, however.

Zircon is common in the conglomerate but volumetrically minor because of the small size of the grains. Rounded and subrounded zircon grains are clearly detrital. Some zircon grains appear metamict and others are, in part, extremely fine aggregates perhaps developed during metamorphic recrystallization. Garnet and apatite are typically well rounded and are detrital in origin.

Particular mention is made of graphite which is present in round grains but is sparse. The graphite may be of organic origin inasmuch as sulphur was identified by microprobe studies of Desborough and Sharp in one specimen.

Rutile is present as deep red, rounded grains and as very fine aggregates with anatase that have developed through alteration of titanium-bearing minerals such as sphene, ilmenorutile, and ilmenite.

Radioactive Heavy Minerals

Radioactive heavy minerals are coffinite, thorite, thorogummite, monazite, huttonite(?), and zircon. According to Desborough and Sharp (1979, p. 39) the uranium-thorium silicates, coffinite and thorite vary widely in composition. The coffinite contains 45-61 weight percent uranium and up to several weight percent thorium and lead whereas the thorite contains 30-45 weight percent thorium and up to several percent uranium, lead, and yttrium. These uranium and thorium minerals are in masses that may have been rounded originally but now appear recrystallized and to have developed irregular borders or overgrowths (Figure 1.16). We are uncertain if the uranium-thorium silicates are recrystallized detrital grains or metamorphic minerals.

Some of the uranium and thorium in the quartz-pebble conglomerate is present in zircon and monazite. As noted above, some zircon appears metamict or recrystallized and most of the monazite is believed to be recrystallized and therefore is a metamorphic mineral (Figure 1.17). These two minerals contain up to six weight percent thorium (Desborough and Sharp, 1979, p. 38).

Minute crystals of uranium and thorium silicate (coffinite) are in graphite and in nickel-cobalt-iron sulphide grains.

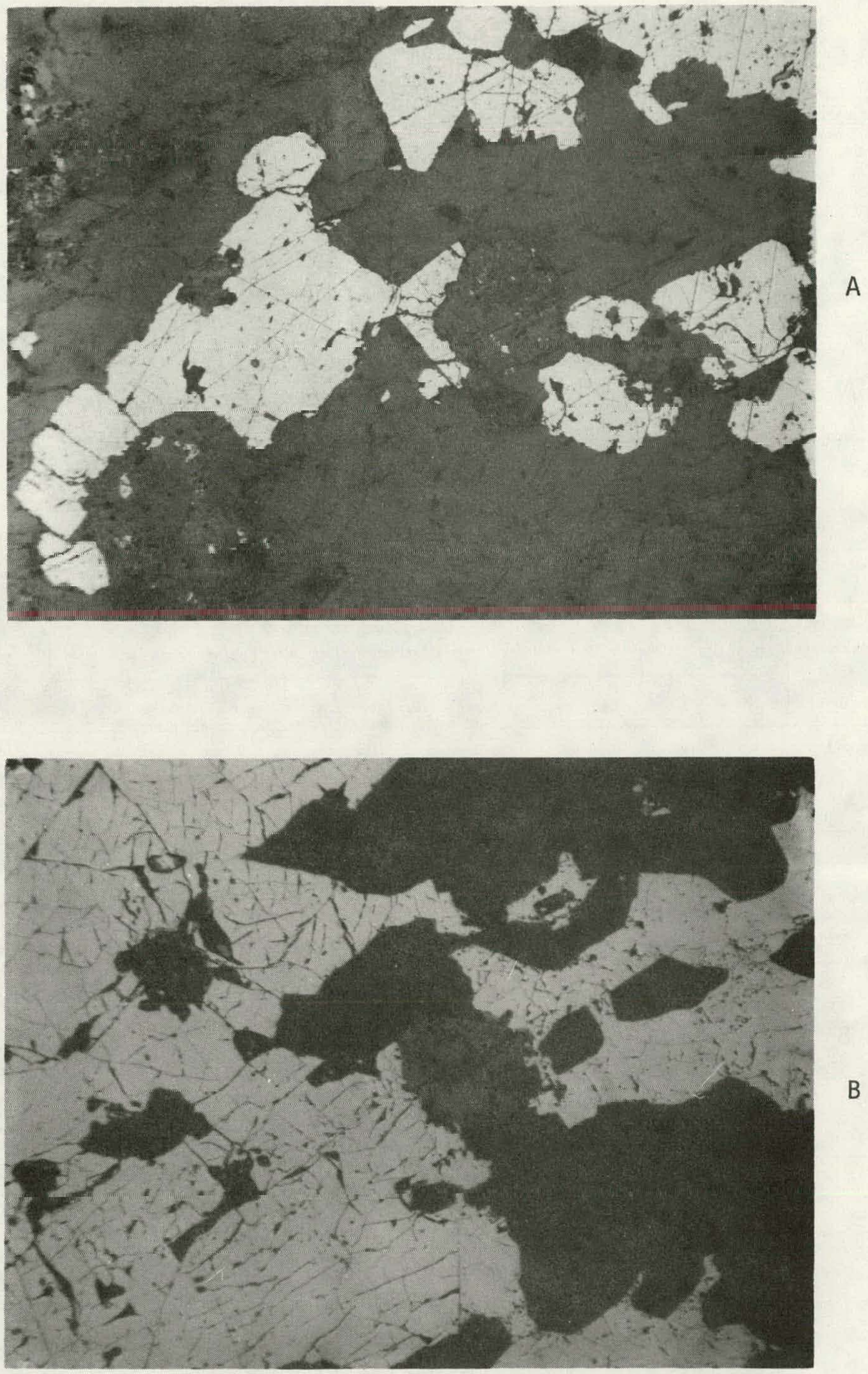


Figure 1.16. A. Photomicrograph showing recrystallized masses of pyrite (white) and subrounded diffuse masses of metamict uranothorite (two dark gray masses center and lower left) with patches of pyrite as minute inclusions. Uranothorite mass in center is 0.25 mm diameter. B. Photomicrograph of coffinite (gray grain in center of photograph). White is pyrite, dark gray is quartz and feldspar. Coffinite is veined with a dark material with orange internal reflections which is tentatively identified as uranothorite. Length of coffinite grain is 0.15 mm. Samples from Onemile Creek locality, Medicine Bow Mountains.

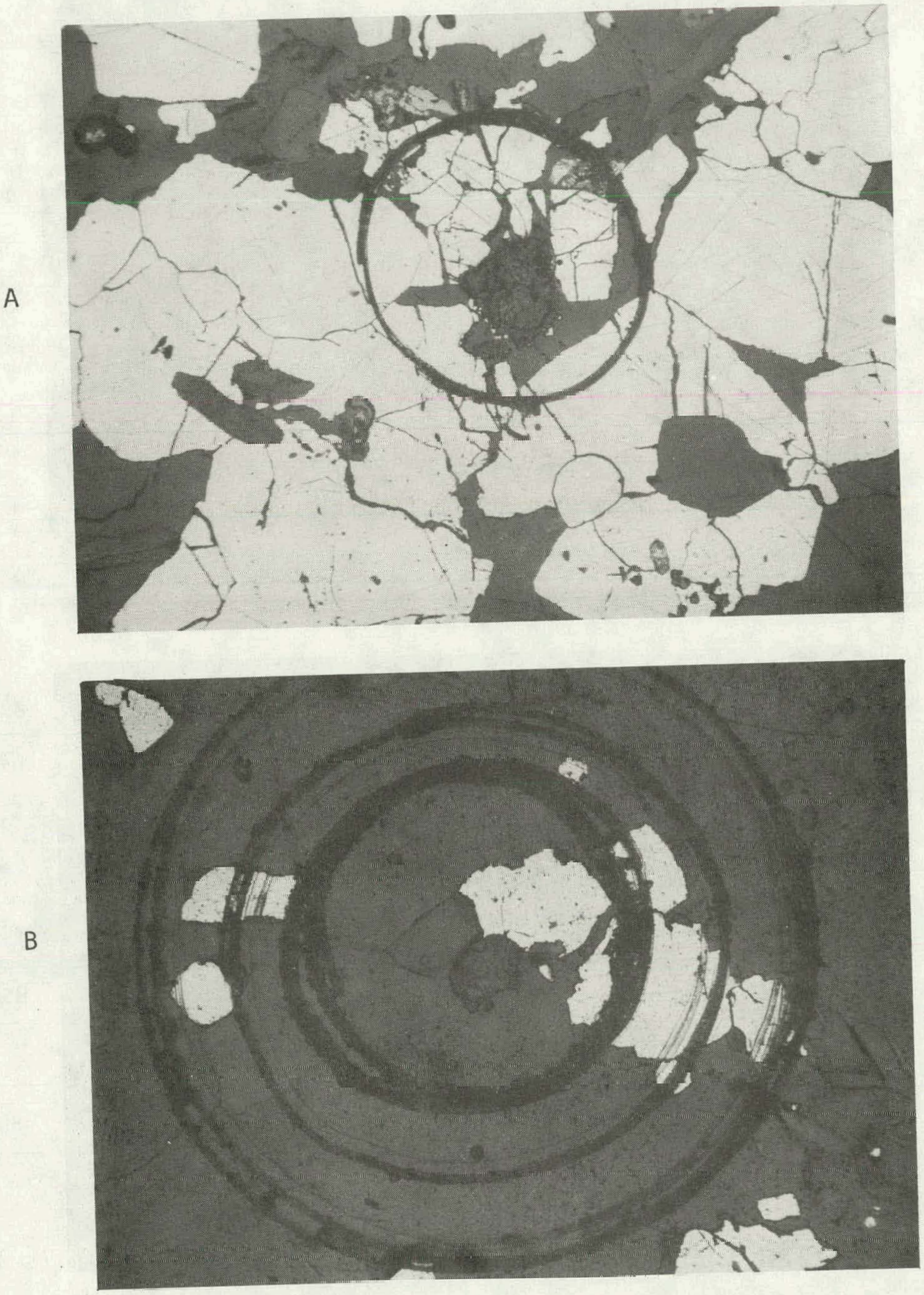


Figure 1.17. A. Grain believed to be metamict and a mixture of monazite and huttonite (?). Element ratios as determined by X-ray spectrograph are Al=2, Si=20, P=89, Th=81, Ca=50, S=16, Fe=48, Ti=2, U=52, Ce=2, and La=3. Euhedral white grains are pyrite, other lighter gray grains are quartz and feldspar. Metamict grain is 0.13 mm in diameter. B. Dark grain in center of photomicrograph is monazite-huttonite (?) mixture with element ratios as determined by X-ray spectrograph of Al=2, Si=50, P=83, Th=53, U=28, Ca=20, S=24, Fe=19, Ce=4, and La=2. White grains are pyrite. Width of dark grain is 0.10 mm. Samples from Carrico Ranch locality of Sierra Madre.

Comparison of Heavy Mineral Suites of the Medicine Bow Mountains and Sierra Madre

In polished section, the principal uranium-thorium minerals are similar in the Carrico Ranch area of the Sierra Madre and the Onemile Creek area of the Medicine Bow Mountains. They are dark gray masses that are subrounded in the less deformed rocks and are flattened and clearly recrystallized in more deformed rocks. However, in spite of apparent similarities, geochemical data and geostatistical analysis of uranium and thorium mineralization in the two areas suggested that the Sierra Madre conglomerates were richer in thorium and the Onemile Creek conglomerates were richer in uranium. The geochemical difference suggests a difference in the heavy mineral content of the two rocks and we initially assumed that the Sierra Madre conglomerates contained substantially larger quantities of zircon and monazite and that these minerals accounted for the preponderance of thorium over uranium in the Sierra Madre. This view proved to be essentially correct.

A comparative study was made of radioactive heavy minerals in the two areas by selecting and marking individual grains in polished thin section and making a semi-quantitative chemical analysis of these grains using an X-ray spectrometer attached to a scanning electron microscope. This particular analytical procedure does not make determinations of weight percentages of elements in minerals and cannot detect elements of an atomic weight less than sodium. Instead, the ratios of elements in minerals are determined and this is useful in mineral identifications especially if minerals of known composition are determined along with the unknowns. Twenty-seven mineral grains were analyzed from the Carrico Ranch locality of the Sierra Madre and sixty-one samples were analyzed from the Onemile Creek locality of the Medicine Bow Mountains. As many as ten to fifteen analyses were made on selected grains from each area to test methods and the homogeneity of the grains.

These analytical results are in Tables 1.4, 1.5, 1.6, and 1.7. In the Sierra Madre (Table 1.4) the uranium- thorium-bearing minerals are monazite, a mineral tentatively identified as huttonite, and mixtures of these two minerals. In polished section these mineral phases and mixed phases have a rather diffused appearance and are considered metamict (Figure 1.17). They are dark gray and may have an orange internal reflection. The uranium-thorium minerals are arranged in Table 1.4 to show general increase in silica from left to right. Note that monazite from Blind River, Canada and from Onemile Creek, Medicine Bow Mountains are included for comparative purposes. Inasmuch as monazite and huttonite are considered isostructural (Deer, Howie, and Zussman, 1962, p. 340-341) we suspect that the table shows a systematic change in composition from a thorium monazite to the grains with a higher proportion of the thorium-cerium silicate, huttonite. The huttonite phase has not been verified by X-ray however. From an economic viewpoint it is clear that these minerals and mixed mineral phases of the Sierra Madre are high thorium and low uranium. The thorium/uranium ratio of the 10 grains analyzed from the Sierra Madre is about three. Note also in Table 1.5 that zircon which is far more abundant in the Sierra Madre than in the Medicine Bow Mountains contains small amounts of uranium and thorium.

The major uranium-thorium bearing mineral phases of the Onemile Creek locality of the Medicine Bow Mountains are believed to be uranothorite, thorogummite, and coffinite (Table 1.6). There are probably

TABLE 1.4. RATIOS OF ELEMENTS, DETERMINED BY A X-RAY SPECTROMETER ATTACHED TO A SCANNING ELECTRON MICROSCOPE, IN MINERAL PHASES FROM THE CARRICO RANCH, SIERRA MADRE, WYOMING (FROM KARLSTROM AND OTHERS, 1981).

Element	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Al	—	—	2	2	2	4	10	7	1	6	2	—	—	—	—
Si	5	4	20	33	50	58	80	90	88	90	84	88	90	91	3
P	86	87	89	90	83	88	77	54	33	61	—	—	—	—	—
S	2	1	16	14	24	16	21	9	5	32	3	11	10	9	—
K	2	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Th	6	1	81	72	53	51	37	31	20	38	—	0.4	0.2	—	—
U	5	2	52	3	28	32	2	0.6	12	5	0.1	—	—	—	—
Ca	—	—	50	35	20	18	8	8	10	18	—	3	4	1	—
Fe	10	14	48	41	19	15	26	11	8	16	5	8	6	5	5
Ce	21	20	2	1	4	3	—	—	—	2	—	—	—	—	1
La	14	12	3	0.5	2	2	—	—	—	1	—	—	—	—	—
Mg	*	*	*	*	*	*	*	*	*	*	—	—	—	—	—
Ti	*	*	*	*	*	*	*	*	*	*	—	—	—	—	97
Zr	*	*	*	*	*	*	*	*	*	*	91	71	63	96	—

*not reported

—not detected

1. Monazite, Carrico Ranch, Sierra Madre
2. Monazite, Carrico Ranch, Sierra Madre
3. Monazite-Huttonite(?), Carrico Ranch, Sierra Madre
4. Monazite-Huttonite(?), Carrico Ranch, Sierra Madre
5. Monazite-Huttonite(?), Carrico Ranch, Sierra Madre
6. Monazite-Huttontie(?), Carrico Ranch, Sierra Madre
7. Huttonite(?), Carrico Ranch, Sierra Madre
8. Huttonite(?), Carrico Ranch, Sierra Madre
9. Huttonite(?), Carrico Ranch, Sierra Madre
10. Huttonite(?), Carrico Ranch, Sierra Madre
11. Zircon, Carrico Ranch, Sierra Madre
12. Zircon, Carrico Ranch, Sierra Madre
13. Zircon, Carrico Ranch, Sierra Madre
14. Zircon, Carrico Ranch, Sierra Madre
15. Rutile, Carrico Ranch, Sierra Madre

OTHER HEAVY MINERALS

Element	1	2	3	4	5	6	7	8	9	10	11
Al	2	39	—	—	—	3	2	—	—	—	—
Si	13	88	—	—	—	95	84	88	90	91	3
S	—	9	96	97	97	—	3	11	10	9	—
K	15	19	—	—	—	—	—	—	—	—	—
Th	4	3	—	—	—	0.1	—	0.4	0.2	—	—
U	34	22	—	—	—	0.1	0.1	—	—	—	—
Ca	2	3	—	—	—	—	—	3	4	1	—
Fe	3	6	24	21	28	1	5	8	6	5	5
P	—	7	—	—	—	—	—	—	—	—	—
Ce	—	—	—	—	—	—	—	—	—	—	1
La	—	—	—	—	—	—	—	—	—	—	—
Mg	—	1	—	—	—	—	—	—	—	—	—
Ti	95	60	—	—	—	—	—	—	—	—	97
Zr	—	—	—	—	—	66	91	71	63	96	—

1. Brannerite, Blind River, Canada
2. Brannerite (?), Onemile Creek, Medicine Bow Mountains
3. Pyrite, Onemile Creek, Medicine Bow Mountains
4. Pyrite, Onemile Creek, Medicine Bow Mountains
5. Pyrite, Onemile Creek, Medicine Bow Mountains
6. Zircon, Onemile Creek, Medicine Bow Mountains
7. Zircon, Carrico Ranch, Sierra Madre
8. Zircon, Carrico Ranch, Sierra Madre
9. Zircon, Carrico Ranch, Sierra Madre
10. Zircon, Carrico Ranch, Sierra Madre
11. Rutile, Carrico Ranch, Sierra Madre

TABLE 1.5. RATIOS OF ELEMENTS DETERMINED BY X-RAY SPECTROMETER ATTACHED TO A SCANNING ELECTRON MICROSCOPE IN MINERAL PHASES FROM THE SIERRA MADRE AND MEDICINE BOW MOUNTAINS, WYOMING.

URANOTHORITE—THOROGUMMITE—COFFINITE

Element	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Al	3	15	2	5	13	—	16	4	24	5	11	3	—	1	1	3	6	3
Si	61	81	88	73	73	42	82	70	78	45	52	55	32	19	21	78	25	41
S	7	9	11	14	15	11	22	22	27	12	17	16	—	14	7	32	40	10
K	13	—	8	13	25	10	8	10	31	10	5	11	34	6	10	41	44	58
Th	87	42	41	77	64	87	87	80	82	85	85	89	50	92	87	—	—	1
U	72	36	43	71	68	63	63	61	53	68	57	70	82	80	77	74	64	94
Ca	3	—	—	1	2	—	3	2	2	2	2	6	—	4	—	—	—	—
Fe	3	6	4	4	5	3	6	5	9	2	4	—	6	5	2	—	3	9
P	6	—	12	6	20	21	24	25	32	26	16	16	—	15	19	11	—	—
Ce	—	1	—	—	—	—	—	—	2	1	1	1	—	5	0.5	—	—	—
La	—	—	—	—	—	—	—	—	1.5	0.5	0.5	0.5	—	2	1	—	—	—
Mg	—	9	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Ti	—	—	—	—	!	—	—	—	—	—	—	—	—	—	—	—	—	17

1. Uranothorite, Blind River, Canada
2. Uranothorite, Onemile Creek, Medicine Bow Mountains
3. Uranothorite, Onemile Creek, Medicine Bow Mountians
4. Uranothorite, Onemile Creek, Medicine Bow Mountains
5. Thorogummite, Onemile Creek, Medicine Bow Mountains
6. Thorogummite, Onemile Creek, Medicine Bow Mountains
7. Thorogummite, Onemile Creek, Medicine Bow Mountains
8. Thorogummite, Onemile Creek, Medicine Bow Mountains
9. Thorogummite, Onemile Creek, Medicine Bow Mountains
10. Thorogummite, Onemile Creek, Medicine Bow Mountains
11. Thorogummite, Onemile Creek, Medicine Bow Mountains
12. Thorogummite, Onemile Creek, Medicine Bow Mountains
13. Mixture, Onemile Creek, Medicine Bow Mountains
14. Mixture, Onemile Creek, Medicine Bow Mountains
15. Mixture, Onemile Creek, Medicine Bow Mountains
16. Coffinite, Onemile Creek, Medicine Bow Mountains
17. Coffinite, Onemile Creek, Medicine Bow Mountains
18. Coffinite + brannerite (?), Onemile Creek, Medicine Bow Mountains

TABLE 1.6. RATIOS OF ELEMENTS DETERMINED BY X-RAY SPECTROMETER ATTACHED TO A SCANNING ELECTRON MICROSCOPE IN MINERAL PHASES FROM THE MEDICINE BOW MOUNTAINS, WYOMING.

some admixtures of monazite and huttonite inasmuch as these uranium-thorium silicates contain a higher proportion of phosphate than might be expected. These three minerals are also believed to be largely metamict and are typically dark gray (Figure 1.16) in polished section, and rarely show the orange internal reflection noted in the Sierra Madre. Note, in particular, that the thorium/uranium ratio of 17 grains analyzed in the Onemile Creek samples is about one. In addition to the thorium and uranium, in silicates, Table 1.5 shows that zircons contain small amounts of uranium. Also, a titanium-rich phase that may be brannerite is present in the Onemile Creek locality.

Multiple analyses were made of eleven grains from the Sierra Madre and Medicine Bow Mountains to determine if the grains were homogeneous. We wished to consider the possibility that the uranium-thorium silicates might have developed by replacement of another mineral such as uraninite and we believed that a remnant of the original mineral might be detected or that some systematic chemical change might be detected that would indicate replacement. Of the eleven grains analyzed five were surprisingly homogeneous (Table 1.7) when we consider the fact that they are metamict and probably have adsorbed cations such as potassium and iron. Six grains proved to be mixtures of various minerals such as monazite, apatite, rutile, iron-spinels as well as the uranium-thorium silicates. These mixed grains, however, were largely uranium-thorium silicates.

We believe, therefore, that the primary uranium-thorium minerals are uranothorite and coffinite in the Onemile Creek locality of the Medicine Bow Mountains and thorium monazite and huttonite(?) in the Sierra Madre. All of these minerals are now largely metamict, the uranothorite is altered to thorogummite, and other grains are changed compositionally by adsorption and hydration. We are not certain if coffinite is an original mineral in the Onemile Creek locality because we do not know of any detrital mineral suites with coffinite.

Element	1	2	3	4	5	6	7	8	9	10	11	12	13	14	$\bar{X}$	Range
Al	4	4	3	2	2	1	2	3	3	12	18	14	2	2	5	1-18
Si	43	39	47	37	41	38	36	42	56	66	52	42	46	44	45	36-66
P	29	24	26	28	23	25	19	28	29	31	20	12	34	29	26	12-34
S	16	12	13	10	13	16	12	12	10	11	14	9	11	12	12	9-16
K	8	10	12	3	5	8	2	11	16	14	11	21	4	9	10	2-21
Th	77	86	81	86	82	82	87	87	87	84	88	86	86	85	85	77-88
U	58	71	71	62	67	63	73	75	78	69	68	70	64	66	68	58-78
Fe	3	2	3	2	3	1	0.5	1	3	3	1	5	2	4	2	0.5-5
Ce	0.5	0.5	1	3	1	0.5	0.5	1	2	—	1	1	2	0.5	1	nd-3
La	—	0.5	0.5	1	1	0.5	1	—	0.5	—	0.5	0.5	0.5	0.5	0.5	nd-1
Ca	2	1	2	2	2	0.5	1	2	1	3	3	1	2	3	2	0.5-3

TABLE 1.7. RATIOS OF ELEMENTS IN FOURTEEN ANALYSES OF A SINGLE URANIUM-THORIUM SILICATE GRAIN FROM THE MEDICINE BOW MOUNTAINS, WYOMING AS DETERMINED BY X-RAY SPECTROMETER ATTACHED TO A SCANNING ELECTRON MICROSCOPE.

We emphasize that the mineral identifications are from chemical analyses by microprobe and X-ray spectrograph combined with optical studies. Other mineral phases might be found by X-ray study and some tentative identifications might be changed. It is clear, however, that the mineral suites of the Onemile Creek locality of the Medicine Bow Mountains and the Carrico Ranch locality of the Sierra Madre are distinctly different and that the Onemile Creek area is a uranium prospect whereas the Carrico Ranch area is a thorium prospect.

Metamorphism and Source of Radioactive Minerals

Radioactive conglomerates from both the Onemile Creek area of the Medicine Bow Mountains and the Carrico Ranch area of the Sierra Madre have been affected by amphibolite facies regional metamorphism and by contact metamorphism adjacent to gabbroic intrusions. Inasmuch as we have found no weakly metamorphosed counterparts of these conglomerates in southern Wyoming, we are unable to demonstrate that any of the principal uranium-thorium minerals are definitely detrital. At present, there is not enough information on heavy minerals in potential source rocks such as Late Archean granites of the Wyoming Province to predict mineral suites that might be present in conglomerates derived from them. We do know that uraninite is present in some vein deposits and in contact metamorphic deposits associated with the Late Archean granites so that uraninite may have been an original mineral of the conglomerates. If so, it has been reconstituted during amphibolite facies metamorphism. It is also possible that coffinite and thorite were the primary detrital minerals, but these minerals have not been identified so far in the potential source rocks.

Where radioactive conglomerate is in contact with gabbroic intrusives, as in some localities of the Onemile Creek area, the uranium has been mobilized and precipitated in veinlets. We have noted veinlets of uranophane in samples from the Onemile Creek area, but only in close proximity to gabbroic intrusions.

SEDIMENTOLOGY OF URANIUM-BEARING UNITS

Uranium-bearing beds in the Sierra Madre and Medicine Bow Mountains are of two types: lenticular beds of quartz-pebble conglomerate interlayered with various types of paraconglomerates and basal successions containing interbedded quartz-pebble conglomerates, coarse-grained quartzites and arkosic paraconglomerates which unconformably overlie older "basement". The first type is of little economic significance because of lack of continuity of the radioactive zones. The second type shows the greatest continuity and may, in the long run, constitute mineable deposits.

Lenticular beds of quartz-pebble conglomerate interlayered with paraconglomerate occur in various units in the Medicine Bow Mountains and Sierra Madre. In the Sierra Madre, these rocks are found in the Silver Lake Metavolcanic Rocks of the Phantom Lake Metamorphic Suite. In the Medicine Bow Mountains, they are found in the Stud Creek Metavolcanic Rocks, Rock Mountain Conglomerate, and Colberg Metavolcanic Rocks of the Phantom Lake Suite.

The paraconglomerates in the Stud Creek Volcaniclastic Rocks and Rock Mountain Conglomerate crop out in the northern Medicine Bow Mountains (Plate 1). The units are highly deformed and few primary sedimentary features are preserved (bedding is often unrecognizable) so our interpretations of sedimentology and depositional environments are speculative. The Rock Mountain Conglomerate is a thick unit containing arkosic matrix paraconglomerate and granular arkose which crops out only in the northernmost Medicine Bow Mountains. To the south, the conglomerates are absent as a mappable unit but similar conglomerates occur as lenses in the underlying Stud Creek Volcaniclastic Rocks. This facies relationship, and the lithology of the paraconglomerates, suggests that the Rock Mountain Conglomerate was being deposited in alluvial fans in the north part of the basin at about the same time volcaniclastic sediments with minor alluvial deposition, took place to the south.

The thin zones of radioactive quartz-pebble conglomerate within arkosic paraconglomerate in these units are interpreted to be fossil placer accumulations of radioactive minerals which represent small areas on alluvial fans which experienced significant sediment reworking, i.e., where compound gravel bars, representing multiple periods of deposition and erosion, were relatively long-lived. However, the evidence for penecontemporaneous volcanism suggests unstable tectonic conditions which would favor rapid, debris flow, deposition (e.g. the paraconglomerates) instead of widespread braided stream-type deposition. In addition, the presence of the marine Bow River Quartzite gradationally(?) overlying the Rock Mountain Conglomerate suggests that fluvial conditions were superceded by marine conditions as seas transpgressed. Thus, environments favorable for fluvial, fossil placer accumulations were apparently of limited extent and duration and we don't expect to find thick or continuous radioactive quartz-pebble conglomerates within these units.

Thin quartz-pebble conglomerates also occur within paraconglomerates of the Colberg Metavolcanic Rocks of the Medicine Bow Mountains and the Silver Lake Metavolcanic Rocks of the Sierra Madre. These paraconglomerates are open framework conglomerates which have a matrix of chlorite and amphibole or arkose, and clasts, up to one meter in diameter, of granite, quartzite, and mafic rock fragments. Some of these paraconglomerates are of volcano-sedimentary origin and contain a predominance of clasts of mafic volcanic rocks and in some areas where paraconglomerates are interbedded with metabasalts, the conglomerates are interpreted as metamorphosed flow breccias. In other areas, the paraconglomerates have taken an arkosic matrix and the majority of the clasts are granite and quartzite. These paraconglomerates are interpreted to be alluvial or submarine debris flows.

In Ts. 18 and 19 N., R. 79 W. of the Medicine Bow Mountains, thick beds of paraconglomerates in the Colberg Metavolcanic Rocks are interbedded with basalt (amphibolite) and rocks considered to be metatuffs. The paraconglomerate beds thin to the north where they are interbedded with quartzite, mica schist, garnet-staurolite-mica schists and amphibolite (basalt?). The paraconglomerate is slightly radioactive (5-3X background) locally, and contains associated lenticular beds of radioactive quartz-pebble conglomerate in two areas: as thin lenses in paraconglomerate in the south, where the conglomerate is best developed, and as more continuous quartz-pebble conglomerate layers in the north, where the paraconglomerate is thin and quartzite and schist more common. The racioactive quartz-pebble conglomerate layers are not of economic interest in either of the above localities (maximum uranium is about 3 ppm and maximum thorium is about 14 ppm), but the occurrences may be significant as far as the depositional environment is concerned. The paraconglomerates are interpreted as alluvial fan deposits with a source in a volcanic highland because of the abundance of mafic volcanic rocks as clasts in the paraconglomerate and the fact that interlayered flows and tuffs are present. The paraconglomerate also contains large rounded granite clasts, granite gneiss, quartzite, and phyllite which could have been derived from underlying metasedimentary rocks of the Phantom Lake Metamorphic Suite and granite-gneiss basement, The source area, therefore, was probably an Archean highland with active volcanoes. The paraconglomerates may have been high viscosity mudflows deposited relatively close to source and the lenticular radioactive quartz-pebble conglomerate layers in the paraconglomerate are interpreted as fluvial channels in the upper reaches of the alluvial fan. The radioactive quartz-granule conglomerate layers north of of the thick well-developed paraconglomerates are interbedded with quartzite and schist and are interpreted as stream channel deposits at the distal end of the fan.

Sierra Madre occurrences in the Phantom Lake Metamorphic Suite are strikingly similar to those of the Medicine Bow Mountains but in some areas exposures are better and structures are preserved in the paraconglomerate that have not been recognized in the Medicine Bow Mountains. The best exposures and best developed paraconglomerate beds of the Sierra Madre are in the cirques north and east of Bridger Peak (Secs. 11, 12 and 13, T. 14 N., R. 86 W.). Here, beds of paraconglomerate exceed 200 meters in thickness. Clasts, up to one meter in diameter, are granite, quartzite, phyllite, and metavolcanic rocks and the matrix is either arkosic or amphibolitic graywacke (Graff, 1978, p. 24). This paraconglomerate succession is richer in volcanic clasts in the lower part and has a greater abundance of granite and other non-volcanic rocks up-section. According to Graff (1978, p. 25) boulder conglomerate layers grade upward into gritstone containing widely spaced, angular clasts of granite. Channels of cross-bedded arkose cut the conglomerate and, in general, the unit is better sorted up-section. In the well-exposed sections, the conglomerate unit displays both fining-upward and coarsening upward sequences (Figure 1.18). Lenticular beds of radioactive quartz-pebble conglomerate have been mapped in the paraconglomerate of the Sierra Madre and, as in the Medicine Bow Mountains, we interpret them to be part of major alluvial fan sequences.

Paraconglomerates of the Silver Lake Metavolcanic Rocks and the Colbert Metavolcanic Rocks, and associated volcanics, are interpreted

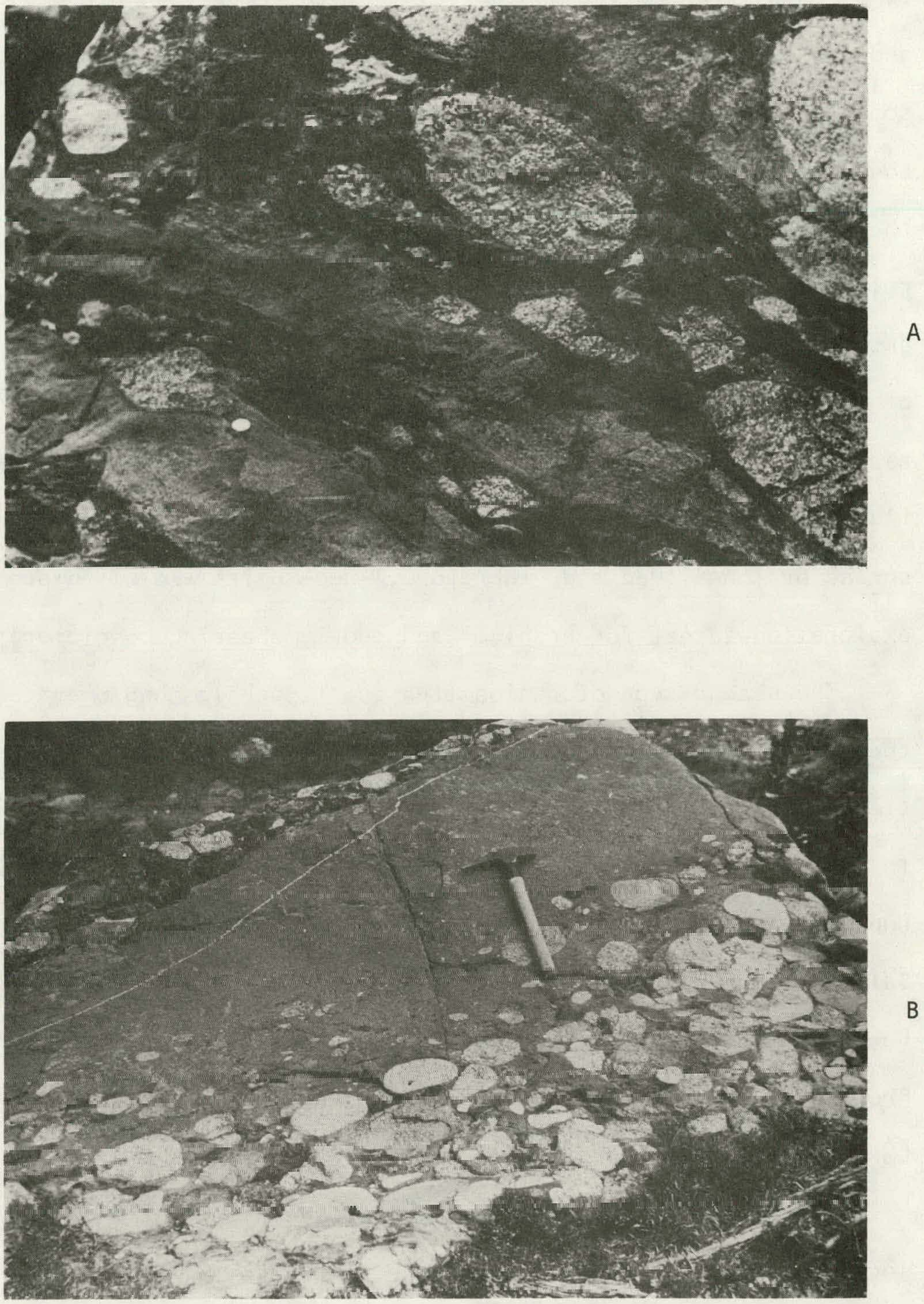


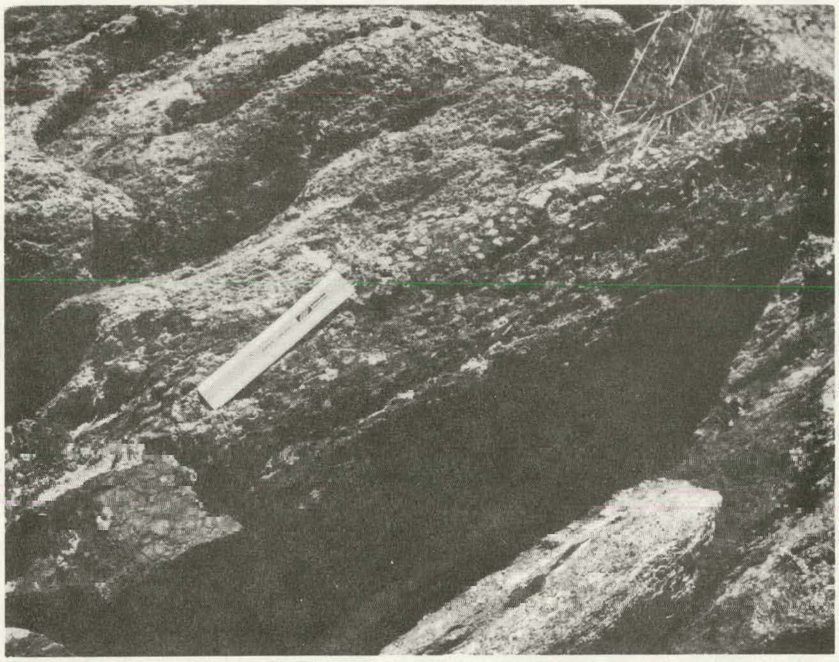
Figure 1.18. Photographs showing graded bedding in conglomerate of the Silver Lake Formation, central Sierra Madre. Note deformed laminae in graywacke layer of A.

to be mainly non-marine. However, in both areas, these rocks overlie and are overlain by marine quartzite successions. It is possible that some interbedded quartzites and phyllites at the distal ends of the paraconglomerate succession are marine and therefore the deposits may be of the fan-delta type (Wescott and Ethridge, 1980) where alluvial fans and related deposits are developed along active continental margins or island-arc systems and are partly non-marine and partly marine. The thin zones of radioactive conglomerates are postulated to be fluvial deposits but, as in the Rock Mountain Conglomerate and Stud Creek Volcaniclastic Rocks, fluvial environments favorable for placer accumulations were not widespread or long-lived and, therefore, these units are not a promising exploration target for uranium- and thorium-bearing conglomerates.

The second type of radioactive quartz-pebble conglomerates, those that occur in basal conglomerate-quartzite successions above unconformities, are present in both the Sierra Madre and Medicine Bow Mountains. In the Sierra Madre, strongly radioactive rocks of the Deep Gulch Conglomerate occur locally at the base of the Phantom Lake Metamorphic Suite and unconformably overlie older Archean gneisses. In the Medicine Bow Mountains, strongly radioactive rocks occur at the base of the Early Proterozoic Magnolia Formation which unconformably overlies the Phantom Lake Suite and Archean granitic rocks.

The Sierra Madre occurrence is best developed at the Carrico Ranch locality, Sec. 12, T. 15 N., R. 88 W. and Secs. 6 and 7, T. 15 N., R. 87 W., but extends northeast of that locality to Sec. 29, T. 16 N., R. 87 W., a distance of approximately 7 kilometers. At the Carrico Ranch the Deep Gulch Conglomerate lies on a gneissic basement which is both paragneiss (isoclinally folded locally) and orthogneiss. In most areas examined, metadiabase sills are intruded between the basement gneiss and the basal beds of the Deep Gulch Conglomerate, but in several areas the basal beds of the Deep Gulch Conglomerate lie on a quartz muscovite-rich rock which may be a regolith. If this is a regolith, it is deformed and metamorphosed and has developed a planar structure locally.

The basal 200 meters of the Jack Creek Quartzite has been divided into five units that can be recognized in most outcrops of the Carrico Ranch area (Plate 7). The lower unit (Unit 1 of Plate 7) is an arkose which is medium- to coarse-grained, poorly sorted, and is interbedded with muscovitic-rich arkose, thin quartz-pebble conglomerate layers, and thin arkosic conglomerate layers. The arkose conglomerate layers of Unit 1 contain abundant granite clasts and angular K-feldspar pebbles. Both the arkosic conglomerate and quartz-pebble conglomerate layers are slightly radioactive (2-3X background). Unit 1 is overlain by an arkose and subarkose (Unit 2 of Plate 7) which is coarse-grained and rich in muscovite. The arkose and subarkose contain well-developed small-scale trough crossbeds and lenticular beds of quartz-granule conglomerate, quartz-pebble conglomerate, and arkosic conglomerate. These conglomerates are also slightly radioactive (2-5X background). Unit 3 (Plate 7) contains pyritic and radioactive quartz-pebble conglomerates (up to 60X background) interbedded with granular to pebbly subarkosic quartzites. These conglomerates occur at the base of the fining-upward stratification sequences and pass up-section into coarse-grained subarkose with well developed trough and planar crossbeds. Beds of quartz-pebble conglomerate are from 17 to 75 cm thick (Figure 1.19) and individual beds


A


B

Figure 1.19. Layers of quartz pebble conglomerate in Deep Gulch Conglomerate member of the Jack Creek Quartzite, Carrico Ranch locality, northwest Sierra Madre. Note dark color of conglomerate layers due to oxidation of pyrite.

can be traced for a distance of two kilometers in the Carrico Ranch area and one bed may extend through the entire outcrop area of Unit 3, a distance of 7 kilometers (Plate 7). Units 1, 2, and 3, together, form the Deep Gulch Conglomerate.

Above the Deep Gulch Conglomerate in the northwestern Sierra Madre is a succession of fine- to medium-grained, planar crossbedded quartzites. Near Carrico Ranch, the lower parts of this quartzite succession were mapped as Unit 4, a medium-grained, planar crossbedded, well sorted subarkose, and Unit 5, a black phyllite with lenses of arkose (Plate 7). In other areas, the quartzite succession of the Jack Creek Quartzite was not subdivided, but includes lenses of highly deformed metalimestone, metagraywacke, phyllite, and paraconglomerate. Paraconglomerate is not present interbedded with units of the Deep Gulch Conglomerate at the Carrico Ranch locality, but lenses of paraconglomerate have been identified in Unit 4 at the Deep Gulch locality (Plate 7) and as discontinuous lenses higher in the Jack Creek Quartzite succession. Highly deformed paraconglomerates are also present in the "basement" rocks below the Deep Gulch Conglomerate at Deep Gulch.

The Deep Gulch Conglomerate (Unit 1 to Unit 3) is interpreted to be a fluvial succession deposited in a braided river system. Unit 1 is a fining-upward sequence which represents a reworked grus which was deposited in channels and braid bars unconformably on gneissic basement. Unit 2 generally coarsens-upwards and contains abundant trough crossbeds and fining-upward stratification sequences. Unit 2 is interpreted to represent aggrading channels in a braided river system. Unit 3 contains the major radioactive conglomerate zones in the Deep Gulch Conglomerate.

These are interpreted to represent deposition of gravels on longitudinal bars in braided rivers, which developed on prograding wet alluvial fans. The coarsening-upwards succession represented by Units 2 and 3 appears to represent faulting and uplift along the basin margins and the most radioactive conglomerate layers of Unit 3 are interpreted to represent compound longitudinal gravel bars which were relatively long-lived. Continued reworking of sediment in these bars, in response to progradation of the fan system, caused heavy minerals to be concentrated in these bars.

Alternate interpretations of the depositional environment of units of the Deep Gulch Conglomerate are feasible and include a glacial connection. It is possible that the braided river system developed because of the excess of clastic material present in a glaciated source area. We have no evidence of glaciation at the time of deposition of the Deep Gulch Conglomerate but it is quite possible that a glaciated area existed north of the present outcrop area.

Units of the Jack Creek Quartzite above the Deep Gulch Conglomerate are interpreted as a marine transgressional succession. In the Carrico Ranch locality, fluvial rocks pass up-section into marine quartzites, phyllites, and thin metalimestone. In other areas, the basal, fluvial rocks of the Deep Gulch Conglomerate may be entirely absent and fine-grained, marine quartzites appear to rest directly on basement rocks. The transgressional character of the Jack Creek Quartzite is unfortunate from an economic viewpoint. It suggests that fluvial deposition was not long-lived enough to produce really thick or extensive alluvial systems and it suggests that the basal

Phantom Lake Suite unconformity is not necessarily a favorable exploration target.

In the Medicine Bow Mountains, the most radioactive conglomerates occur at the base of the Magnolia Formation. This unit contains anomalously radioactive zones throughout the Medicine Bow Mountains but the thickest and most radioactive conglomerates found so far crop out in the Onemile Creek area of the northern Medicine Bow Mountains. The Onemile Creek area (Secs. 5, 6, and 7, T. 15 N., R. 78 W.) has many characteristics in common with the Deep Gulch area of the Sierra Madre. At Onemile Creek, however, the units of interest are in the lower conglomerate member of the Magnolia Formation of the Deep Lake Group and are Early Proterozoic instead of Late Archean. Paraconglomerate is also more prevalent than in the beds of the Lower Jack Creek Formation. Before discussing the depositional environment of radioactive beds at Onemile Creek it is desirable to review the general stratigraphy and regional distribution of beds of the Magnolia Formation. In the Medicine Bow Mountains it has been possible to subdivide the Magnolia Formation into two mappable units which are referred to as the Conglomerate Member and Quartzite Member of the Magnolia Formation (Karlstrom and Houston, 1979a, 1979b). The Conglomerate Member ranges in thickness from 0 to 330 meters and is composed of paraconglomerate, quartz-pebble conglomerate, and interlayered pebbly and granular quartzite. The quartzite member ranges from 400 to 600 meters in thickness and is coarse-grained quartzite composed of rounded granules of quartz (Karlstrom, 1977).

The Conglomerate Member of the Magnolia Formation is of particular interest in this study because it is the host of all significant uranium-thorium deposits in the Medicine Bow Mountains. In general, the Conglomerate Member has paraconglomerate at the base which grades upward into conglomeratic quartzite. The paraconglomerate at the base of the Conglomerate Member is variable in thickness and contains beds that range from true paraconglomerate (open-framework) to beds of orthoconglomerate (closed or largely closed framework). Clasts range in size up to tens of centimeters in diameter, and are granite, granite gneiss, phyllite, quartzite, amphibolite, hornblende gneiss, sericite schist and garnet biotite schist in a matrix of arkose. Thick beds of paraconglomerate are present at the base of the Conglomerate Member at localities west of the confluence of Brush Creek and Little Brush Creek (Sec. 22, T. 16 N., R. 81 W.) and in outcrops east of the Arrastre Creek (Sec. 10, T. 16 N., R. 80 W.) Paraconglomerate beds have not been recognized in the limited exposures of the Conglomerate Member at the North Fork Rock Creek (although an outcrop of paraconglomerate in Sec. 19, T. 18 N., R. 78 W. may be part of the basal Magnolia Formation) or on the Medicine Bow River. Thick beds of paraconglomerate are again present in the Conglomerate Member in outcrops on the north side of Rock Creek (Sec. 19, T. 18 N., R. 78 W). These beds thin to the northeast towards Onemile Creek and to the southwest, towards Deep Creek (Plate 1). In general, the paraconglomerates are best developed on the northwest limbs of synclines and are believed to thin to the southeast. We believe that the paraconglomerates represent alluvial fans that had a source (perhaps a fault scarp) some distance northwest of their principal outcrop areas; i.e. north of Rock Creek and west of the confluence of Little Brush and Brush Creeks. The paraconglomerates grade up-section and along strike into the Quartzite Member which is interpreted as braided stream or river deposits. The Conglomerate Member succession is therefore transgressive or it may represent a decrease in tectonic activity in the source area through time.

The Magnolia Formation paraconglomerate is radioactive and, in contrast to paraconglomerates of the Phantom Lake Metamorphic Suite, it has radioactivity above background in virtually every outcrop examined. The uranium and thorium content of the paraconglomerate is generally low, however, averaging 10-20 ppm uranium and 20-30 ppm thorium. In local areas, beds of quartz-pebble conglomerate are interlayered with paraconglomerate and some beds of paraconglomerate are quartz-rich and better sorted than typical paraconglomerate -- these beds normally contain a higher percentage of uranium and thorium, up to 545 ppm uranium and 1143 ppm thorium in the Brush Creek area (MB-9) and 110 ppm U and 190 ppm Th near Rock Creek (BOWS-1,-2,-3). Also, a zone of uranium- and thorium-rich paraconglomerate north of Rock Creek, in drill hole EMB-11, which has local layers of quartz-pebble conglomerate and zones of quartz-rich paraconglomerate, is nearly 200 feet thick and averages over 100 ppm uranium, and as such constitutes a low-grade uranium resource.

The paraconglomerate of the basal Magnolia Formation obviously had a different source than paraconglomerates of the Phantom Lake Metamorphic Suite. Magnolia paraconglomerates have fewer clasts of volcanic rocks, more quartz, and an arkosic rather than a phyllitic or amphibolitic matrix. The paraconglomerates were derived from a mixed source consisting of Archean granite, gneiss, metasedimentary and metavolcanic rocks.

The proportions of these various rocks in the local source area, within about 10-15 miles of the alluvial fan system, we assume, is reflected in clasts of the paraconglomerate. In the area west of the confluence of Brush Creek and Little Brush Creek and at Arrastre Lake, typical clasts in the paraconglomerate are phyllite, quartzites, mafic volcanics, and vein quartz. Granite clasts are uncommon in these localities (although they are present in two of eight localities studied). In contrast, granite is a common constituent of paraconglomerates of the basal Magnolia Formation in outcrops extending from the south side of Deep Creek to Onemile Creek (Plate 1), and it is in these localities where the most uranium-rich paraconglomerates (and quartz-pebble conglomerates) are found.

The Conglomerate Member of the Magnolia Formation gets progressively more radioactive as one goes north from Rock Creek and this change is accompanied by a change in lithology, from dominantly paraconglomerate in the south to dominantly quartz-pebble conglomerate in the north. The best developed quartz-pebble conglomerates are in the vicinity of Onemlle Creek (Secs. 5, 6, and 7, T. 18 N., R. 79 W.) where the Magnolia Formation is exposed in the nose of an overturned syncline (Figure 1.11). Only the Conglomerate Member of the Magnolia Formation is exposed in the Onemile Creek area and here, it has been subdivided into five units (Figures 1.11 and 1.12). Unit 1 is an arkosic paraconglomerate with abundant large granite clasts interlayered with subarkose (Plate 4). Unit 1 grades up-section into a trough cross-bedded subarkose with thin lenticular beds of radioactive quartz-pebble conglomerate referred to as Unit 2. Unit 2 grades upward into a granular subarkose (Unit 3) with thin lenticular beds of radioactive quartz-pebble conglomerate. Unit 3 is overlain by biotite chlorite schist with paraconglomerate lenses (Unit 4), and Unit 4 is overlain by muscovite-rich subarkose with thick and continuous beds of radioactive quartz-pebble conglomerate (Unit 5, Figures 1.11 and 1.12). The radioactive quartz-pebble conglomerate beds of Unit 5 are the most radioactive and most persistent of any in the Medicine Bow Mountains. As illustrated in Figure 1.11, there are two main zones (5a and 5b) in Unit 5 that contain radioactive quartz-pebble conglomerate. Individual zones of radioactive quartz-pebble conglomerate are up to 20 m thick. However, these radioactive zones are not single beds of conglomerate but are intervals of coarse-grained quartzite with numerous quartz-pebble conglomerate layers. The individual layers of quartz-pebble conglomerate range in thickness from that of a single pebble to composite zones tens of feet thick. Unfortunately these quartz-pebble conglomerate beds are not uniformly mineralized and may range from as little as 10 ppm to over 1000 ppm uranium. The variation in uranium values is believed to be related to the location within a channel, with generally higher values at the base of individual channels.

We interpret the basal Conglomerate Member of the Magnolia Formation as deposits of an alluvial fan system with the paraconglomerate representing more proximal, mudflow deposits; the mixed paraconglomerate-quartz-pebble conglomerate representing mid-fan deposits; and the coarse-grained quartzite with beds of quartz-pebble conglomerate representing parts of the distal fan. Overall, the Conglomerate Member of the Magnolia Formation is transgressional but there are obvious local episodes of progradation superimposed on the transgressional event. The presence of the most uranium- and thorium-rich quartz-pebble conglomerate at Onemile Creek is thought to be related to proximity to a source underlain by Late Archean granites as suggested above, but we must emphasize that this is also the only area of outcrop of the Conglomerate Member which contains thick quartz-pebble conglomerates so that the uranium and thorium mineralization might also be related to favorable depositional conditions. We cannot rule out the possibility of finding other thick and persistent beds of quartz-pebble conglomerate in unexplored subcrops of the Conglomerate Member elsewhere in the Medicine Bow Mountains, even in areas at some distance from known granitic source areas.

GEOCHEMISTRY OF URANIUM-BEARING ROCKS

A variety of geochemical techniques have been used to study uranium-bearing rocks of the Sierra Madre and Medicine Bow Mountains and to prospect for additional deposits. Uranium and/or thorium analyses have been made of rock chips, stream sands and silts and water samples. Radon analyses have also been made of water samples to test this method in exploration. The results of the geochemical analyses have been considered in two ways: as a prospecting tool and as an aid in understanding geologic features of the rocks such as surface alteration, mineralogy, and geochemical changes through time.

Geochemical Prospecting

Water

Uranium content of surface waters was studied as part of a hydrogeochemical reconnaissance study of the United States Department of

Energy (Weaver and others, 1978) and as part of a more detailed geochemical study of the United States Geological Survey in a key area of the Medicine Bow Mountains (Miller and others, 1977). Neither of these studies was successful in delineating uranium concentrations in quartz-pebble conglomerate. The reconnaissance water sampling program of the Rawlins NTMS quadrangle (Weaver and others, 1978) was undertaken by sampling about every 5 square mile interval, and water samples were collected from rocks ranging in age from Precambrian to Recent. The striking feature of the reconnaissance study was the low uranium values obtained in water from rocks of Precambrian age as compared with those obtained in younger rocks. The uranium concentrations in water of the Rawlins NTMS quadrangle ranged from less than the detection limit of 0.2 parts per billion (ppb) to 448 ppb uranium, with a mean value of 6 ppb (Weaver and others, 1978, p. 10-18). Anomaly thresholds were set at 50 ppb for the water samples (Weaver and others, 1978, p. 11). Not only were there no values of 50 ppb in the areas underlain by rocks of Precambrian age, these areas had the lowest uranium values of any water samples (Weaver and others, 1978, pl. 3); maximum between 2.01-5.00 ppb. Water samples from known uranium-bearing quartz-pebble conglomerate localities did not show an increase in uranium over other rocks of Precambrian age.

Chemical analyses of waters collected from streams, lakes, springs, and seeps of an area of about 30 km^2 near Arrastre Lake of the Medicine Bow Mountains (Miller and others, 1977) were made at intervals of two samples every 2.5 km^2. The waters of the Arrastre Lake area are extremely dilute sodium bicarbonate types and, according to Miller and others (1977, p. 17), the small amount of dissolved solids in these waters suggests an extremely short contact time between waters and rocks of the area. Dilute waters of this type would not be expected to contain high uranium concentrations and this appears true inasmuch as the mean uranium content of 24 samples is 0.25 ppb, and the maximum value is 0.50 ppb. There is no clear relationship found between the uranium values in waters draining areas of known low-grade radioactive quartz-pebble conglomerates of the Arrastre Lake area and waters draining other areas.

These two studies suggest to the writers that the determination of uranium in water is a poor method of prospecting for uranium in quartz-pebble conglomerates or in other crystalline rocks of Laramide uplifts. Two factors are involved: the short contact time between rocks and these dilute waters, discussed above, and the extreme leaching of uranium from surface outcrops, which will be discussed below. Obviously, this conclusion applies to typical occurrences of the Rocky Mountains and may not be valid in other geologic and topographic situations.

Stream Sediment Sampling

Stream sediment sampling should be a more useful exploration technique than water sampling in high-relief crystalline areas of the type investigated here, but the reconnaissance sampling of Weaver and others (1978) and detailed sampling of Miller and others (1977) in the Medicine Bow Mountains and Charleton (personal communication, 1980) in the Sierra Madre did not show any correlation between known uranium-bearing quartz-pebble conglomerate occurrences and abnormally high concentrations of uranium in stream sediment.

In contrast to water samples, where no uranium concentrations above normal were detected in the mountain areas underlain by rocks of Precambrian age, a number of uranium anomalies were obtained in this area in stream sediment studies of Weaver and others (1977). The samples containing high uranium values were primarily along the Mullen Creek-Nash Fork shear zone of the Medicine Bow Mountains and in the vicinity of the confluence of Billie Creek and the Encampment River of the Sierra Madre; none were near occurrences of uranium-bearing quartz-pebble conglomerate (Weaver and others,1977, pl. 4). Sediments of these two areas consist of reworked Pleistocene glacial deposits as well as sediments derived from bedrock. However it is also true that rock samples from outcrop and drill core in this area are not especially rich in uranium (up to 11 ppm U and 38 ppm Th).

Charlton's sampling in the west-central Sierra Madre, reported in detail in Volume 2 of this report, showed uranium values in stream sediment up to 12 ppm, with mean values about 3-5 ppm. There was no obvious correlation between uranium values and local rock type, although his sampling did not extend into the northwestern Sierra Madre where known radioactive conglomerates occur.

The apparent lack of a correlation between uranium values in stream sediment and outcrops of uranium-bearing quartz-pebble conglomerates has several possible explanations. First, available data are mainly from areas which contain only mildly radioactive conglomerates and the two major areas of outcrop of radioactive quartz-pebble conglomerate were not extensively sampled, so it may be a sampling problem. Second, deep leaching of conglomerates prior to recent erosion and alluvial transport appears to have removed much of the uranium from outcrops so that the resulting stream sediment would not be expected to contain appreciable uranium. We suspect a thorium assay might be a more useful tool than a uranium assay in trying to identify radioactive anomalies from stream sediment in the Medicine Bow Mountains and Sierra Madre.

Radon

Miller and others (1977) obtained interesting results by determining the radon gas concentrations in 50-ml samples of water stored in glass containers. This work was done in the Arrastre Lake area of the Medicine Bow Mountains where there are known low-grade occurrences of uranium-bearing quartz-pebble conglomerate. The radon study was strictly reconnaissance in that the amount of radon gas dissolved in water was determined on only seven water samples (Miller and others, 1977, p. 20, Table 4).

The radon studies showed that the highest radon values were in areas which contained the most radioactive quartz-pebble conglomerate and that the radon content of waters near occurrences of radioactive quartz-pebble conglomerate is extremely high for dilute waters. According to Miller and others (1977, p. 17-21), the equivalent amount of uranium-238 needed to produce the amount of radon-222 in the waters of the Arrastre Lake area greatly exceeds the uranium content of surface rocks. Inasmuch as the surface samples are leached of uranium, Miller and others (1977) hypothesized that the radon gas might be generated in the subsurface and migrate upwards in ground water.

After the radon study was made of this area, we postulated that uranium-bearing quartz-pebble conglomerates would be encountered by drilling to intersect beds below the zone of weathering. Only one drill hole penetrated the conglomerate layer and it was barren (maximum values were 12 ppm and 36 ppm Th). Obviously there has not been enough exploration in the Arrastre Lake area to test the validity of the radon survey, and we believe that additional exploration is warranted. Nevertheless, other workers have shown that radon anomalies can be generated in areas that do not have significant concentrations of uranium. Andrews and Wood (1972) have demonstrated that significant radon can be concentrated by ground water circulation in fractured rocks despite the fact that the uranium content of the rocks is not much greater than average.

Rock Chips

The most comprehensive geochemical survey done in the Sierra Madre and Medicine Bow Mountains was uranium and thorium analyses of rock chips collected during the current study and also on rock samples collected during prior studies at the University. 1632 rock samples were analyzed in the Sierra Madre and 1324 rock samples were analyzed in the Medicine Bow Mountains for uranium by neutron activation, delayed neutron count at the Oak Ridge Gaseous Diffusion Plant, Oak Ridge, Tennessee.

Random rock sampling has several uses as a prospecting tool. In relatively homogeneous rock masses, such as felsic intrusions, anomalous areas may be outlined and examined in greater detail. And, on a regional scale, random rock sampling is useful in establishing background values of elemental abundances, for use in evaluating anomalies and in defining regional metaelogenic provinces. However, where a great variety of rock types are exposed, as in the Sierra Madre and Medicine Bow Mountains, random geochemical sampling cannot be used alone to outline anomalous areas. Meaningful interpretation of geochemical results requires detailed geologic information on rock types, rock ages, and geologic history.

Table 1.8 shows that in this area rock analyses are not useful in outlining lithologic subdivisions that have greater potential as a source of uranium. Uranium values are no greater in quartzite of the Deep Lake Group than in quartzite of the Phantom Lake Metamorphic Suite, and Libby Creek Group, Archean Granite, or rocks south of the Mullen Creek-Nash Fork shear zone. In fact, when quartzite alone is analyzed, the Phantom Lake Metamorphic suite appears to be a more promising target for uranium exploration than the Deep Lake Group which contains the most significant deposits (Table 1.8).

A comparison of mean uranium values in known conglomeratic members such as the Magnolia Formation, Rock Mountain Conglomerate and Deep Gulch Conglomerate is useful in determining which of these units is the most favorable host rock for uranium (Table 1.8), but these results are biased by selection of samples from radioactive horizons.

The Archean granites of the Sierra Madre and Medicine Bow Mountains do not have uranium or thorium values (Table 1.8)greater than crustal averages which are about 3 to 5 ppm uranium and 10-20 ppm thorium (Rogers and Adams, 1969). These Archean granites are not anomalous in terms of Th/U ratios which are 5.7 for these Archean granites as compared with a range of 3 to 5 for most granites (Rogers and Adams, 1969). We suspect, therefore, that the uranium minerals in quartz-pebble conglomerates were not derived locally, and thus the fact that an Archean granite might be located near a quartz-pebble conglomerate locality does not indicate that this locality is a better area for prospecting.

TABLE 1.8. STATISTICAL SUMMARY OF MEAN URANIUM AND THORIUM VALUES FROM VARIOUS GEOLOGIC UNITS OF THE SIERRA MADRE AND MEDICINE BOW MOUNTAINS.

		Unt			Th	
	Max. Value	Mean	Min. Value	Max. Value	Mean	Min. Value
Medicine Bow Mountains & Sierra Madre						
Archean granitic rocks	6.10	1.97	0.10	45	11	<2
Fine grained quartzites of the Phantom Lake Metamorphic Suite	48.60	3.83	0.10	395	14	1
Quartzites of the Deep Lake Group	18.9	1.71	0.30	42	8	ND
Quartzites of the Lower Libby Creek Group	5.00	1.55	0.20	328	22	1
Mafic intrusives	8.3	0.89	0.0	129	13	<2
Rocks south of the Cheyenne Belt	22.60	2.18	0.0	29	8	<2
Medicine Bow Mountains						
Rock Mountain Conglomerate	194.50	11.80	0.60	86	12	<2
Magnolia Formation	1620.00	74.13	0.50	1143	89	1
Magnolia Formation of the Onemile Creek Area	1620.00	91.38	1.00	1143	108	1
Magnolia Formation of the Onemile Creek Area (Surface)	176.70	25.25	1.50	915	99	3
Magnolia Formation of the Onemile Creek Area (Subsurface)	1620.00	125.90	0.35	1143	109	<2
Sierra Madre						
Deep Gulch Conglomerate	718.00	20.53	0.30	2596	109	<2
Deep Gulch Conglomerate (Surface)	205.50	10.19	0.30	839	94	<2
Deep Gulch Conglomerate (Subsurface)	718.00	26.77	0.40	2596	118	<2
Magnolia Formation	76.90	3.68	0.50	667	46	1

CONCLUSIONS FROM GEOCHEMICAL PROSPECTING

Some of the geochemical studies outlined above are useful tools in locating favorable areas to prospect for uranium, but their usefulness varies with terrane and a variety of other factors. We believe that stream sediment and rock sampling are more useful in the areas studied and that the radon method is not yet fully evaluated.

None of these methods substitute for geologic mapping or geologic exploration concepts. We doubt that any of the uranium-bearing quartz-pebble conglomerates would have been discovered without the application of the quartz-pebble conglomerate model in exploration.

Surface Leaching

It was clear from the beginning of this study that surface outcrops had been leached and oxidized because there was discoloration of quartz-pebble conglomerates (red and brown staining) and, in many localities, numerous cubic cavities could be recognized in quartz-pebble conglomerates that were believed to be voids left after oxidation and dissolution of pyrite. No true gossans have been found although a few quartz-pebble conglomerate samples had limonite (geothite) pseudomorphic after pyrite.

The radioactivity of the quartz-pebble conglomerates was initially attributed to thorium, because we assumed that uranium minerals had been leached from surface outcrops along with pyrite. Therefore we did not believe that assays from surface outcrops would be an adequate measure of the uranium content of the rocks and we suspected that airborne radiometric surveys might not be as useful a tool here as they are in Wyoming basins, where primary uranium minerals are dissolved but are re-deposited as oxidized minerals in the vicinity of sandstone-type deposits. Neither the anions necessary for the formation of insoluble oxidized uranium species (carbonate, sulphate, arsenate, silicate) nor elements such as potassium, calcium, or vanadium necessary to form insoluble carnotite or tyuyamunite are abundant in quartz-pebble conglomerates. Therefore, the uranium in these deposits was thought to be removed from soluble uranium mineral species at the surface and transported in dilute ground and surface waters to the adjacent basins. The fact that precipitation in the mountains is two to three times greater than that of the basins is undoubtedly another factor contributing to the thorough leaching of uranium from the quartz-pebble conglomerate outcrops. What is more, we believe that oxidation and leaching of quartz-pebble conglomerates might have occurred at several periods during the Tertiary (Miller and others, 1977, p. 9) so that the amount of surface leaching is not dependent on the amount of time since the conglomerates were last exposed, which was probably Late Pliocene or Early Pleistocene.

The concept of leaching of uranium from surface outcrops led to the proposal that drilling below the zone of oxidation would be necessary before these quartz-pebble conglomerates occurrences could be adequately evaluated (Miller and others, 1977). We knew of no tried way to predict from surface mineralogical or geochemical studies whether a uranium-bearing mineral species would be present at depth in the quartz-pebble conglomerates.

However, one technique which may be useful in the future in evaluating uranium loss due to leaching and in predicting the presence of appreciable uranium in subcrop is lead isotopes. Lead isotope studies of some granites in central Wyoming (Rosholt and others, 1969; Stuckless, 1979) showed that Pb_{208} and its parent isotope Th_{232} are present in approximately equal amounts whereas Pb_{206} is generally present in excess of the amount necessary for equilibrium with its parent isotope U^{238}. If it is assumed that Pb_{208} and Pb_{206} are essentially immobile, the excess of Pb_{206} as compared with U^{238} can be attributed to uranium loss from weathering. After the initial holes were drilled in the Onemile Creek area, it was determined that unaltered pyrite was present in the subsurface and that Th/U ratios below a depth of about 30 meters were about 1 in most samples as compared with ratios of about 2-6 on the surface. It was obvious from this that leaching of uranium minerals had taken place. Samples were then submitted to geochemists of the United States Geological Survey to determine if lead isotope studies could be used to predict uranium loss and thus aid in making decisions on whether or not to drill in other localities. These studies were completed in 1979 but were not conclusive (F.A. Hills, 1979, personal communication) enough to be used for decision making in drilling programs. We anticipate that this type of study will be continued and may be of great value in prospecting.

One key objective of the drilling program was to try to quantify the depth and degree of surface leaching of uranium in order to make accurate uranium resource estimates. A statistical study was made by Borgman and others (1980) in two key areas, where drilling coverage was most complete: the Onemile Creek area of the Medicine Bow Mountains, and the Carrico Ranch area of the Sierra Madre. The results of the statistical study were surprising in that they showed a striking difference in chemistry between the Sierra Madre and Medicine Bow deposits and they demonstrated that uranium loss was not consistent with depth. By using a multivariate stepwise regression method (Borgman and others, 1980, p. 47), it was determined that leaching was negligible below a depth of about 47 meters. The surprise was that equations at substantial depth ($\sim$ 47 meters) reduce to the relations for no leaching of

Sierra Madre (depth = ∞)

$$U/Th = .56Th^{-.26}$$

Medicine Bow (depth = ∞)

$$U/Th = 4.99Th^{-.34}$$

Note that the exponent for thorium is about the same in both areas, but the multiplier is nine times larger in the Medicine Bow Mountains than in the Sierra Madre. This indicates that most of the radioactivity in the Sierra Madre area is from thorium, whereas the Medicine Bow Mountain area has proportionally much more uranium. In other words, there was far less uranium to leach in the Sierra Madre and a method such as the lead isotope method discussed above, to predict this from study of surface samples might save substantial drilling cost in the future.

A second finding of the leaching study was that the amount of uranium leaching increases with depth to about the first ten meters and then decreases steadily to 50 meters below which leaching is absent. This finding remains speculative largely because drilling was designed to intersect radioactive rocks below about 30 m so that sampling control at shallow depths is poor and the regression equations are essentially based on surface samples and samples from below 30 m. Nevertheless, Borgman and others, (1980, p. 52) found this relationship in scatter plots of uranium versus depth in both the Sierra Madre and Medicine Bow Mountains and they believed it to be real and not a peculiarity of the statistical analysis procedure. Borgman and others (1980, p. 52) suggested that the leaching behavior may be affected by the length of time moisture stays in contact with ore as it percolates downward. That is, surface moisture might move down rapidly at first and then slow down with increasing depth. Other processes may also be involved. For example, two periods of leaching may be involved -- one in the Tertiary and a second in the Pleistocene. In any event, the leaching process is not a simple straight line relationship and it will require additional study to understand it and to devise methods of predicting uranium concentrations at depth.

GEOCHEMISTRY AND SEDIMENTOLOGY

It has been shown by Theis (1979) that there is a relationship between the size of clasts and the percentages of various elements in the quartz-pebble conglomerates of the Blind River uranium district of Canada. The variation in percentages of elements in the quartz-pebble conglomerates reflects the mineralogy and the size of clasts is related to sediment carrying capacity of the streams or rivers of Early Proterozoic time. The concept of hydraulic equivalence (Rittenhouse, 1943) states that grains carried in suspension and deposited as current velocity decreases should have the same settling velocity (Rubey, 1933). That is, large light grains should be deposited along with small heavy grains to maintain hydraulic equivalence, and there should be a relationship between size and specific gravity of heavy minerals deposited with a given size of quartz clasts. If heavy minerals are approximately the same size, a heavy mineral of high specific gravity will be deposited with the larger quartz clasts and a heavy mineral with lower specific gravity will be deposited with smaller quartz clasts. There are a number of other variables involved such as transport by traction vs. transport in suspension, re-entrainment (Hand, 1967), and infiltration (Minter, 1979) but, if a relationship between grain size and mineralogy (reflected in chemistry) can be shown, it certainly suggests that hydraulic factors were operative.

In the Blind River quartz-pebble conglomerates, quartz is the primary light mineral and the principal heavy minerals are pyrite, uraninite, brannerite, zircon, and monazite. When the diameter of quartz clasts is plotted against cerium and lanthanum (elements that reflect the presence of monazite) there is an inverse relationship between the percentage of these elements and the size of quartz clasts with highest concentrations in finer-grained conglomerates (Theis, 1979, p. 11). Inasmuch as monazite is relatively small, has a limited size range, and has a relatively low specific gravity as compared with other heavy minerals, this relationship suggests that changes in monazite concentration reflects different depositional energy conditions. The same inverse relationship between quartz clast size and zirconium (element that reflects the presence of zircon) is noted at Blind River (Theis, 1979, p. 11), and that is undoubtedly controlled by similar hydraulic factors. As further verification of the role of hydraulic factors at Blind River, Theis (1979, p. 12) noted a direct relationship between the uranium content of the conglomerate and quartz clast size and an inverse relationship between thorium content and clast size. This is interpreted as due to the high specific gravity uraninite being associated with larger clast size and lower density thorium-bearing minerals (monazite and thorite) being associated with smaller quartz clasts. These relationships are shown more clearly by a strong correlation between U_3O_8/ThO_2 versus grain size, which shows that this ratio is largest in the coarsest conglomerates.

When the same statistical tests used by Theis were applied to samples from the Carrico Ranch locality of the Sierra Madre and the Onemile Creek locality of the Medicine Bow Mountains no Blind River-type correlation was noted between grain size and key elements such as U, Th, Zr, Ce (Tables 1.9 and 1.10). This lack of correlation is explained by the fact that the specific gravity of uranium-thorium-bearing heavy minerals in southern Wyoming conglomerates (uranothorite, monazite-huttonite, monazite, zircon and coffinite) are all about the same (between 4 and 5) so that variations in depositional energy conditions did not significantly fractionate the heavy minerals. In contrast, Blind River conglomerates contain uraninite (specific gravity 8-10) and zircon and monazite (specific gravity 4-5) and these heavy minerals were effectively fractionated by hydraulic processes.

If we assume that hydraulic factors were operative in formation of the Wyoming deposits, as they must have been at Blind River, it seems probable that the known deposits in Wyoming did not contain uraninite originally. If uraninite had been present, say at Onemile Creek, and subsequently altered to uranium-thorium or uranium silicates, a grain size correlation should be expected. That is, the uranium-rich samples should have come from the coarsest conglomerates. Inasmuch as this is not the case, we suggest that uraninite was not a detrital mineral in the southern Wyoming localities.

TABLE 1.9. MATRIX OF PEARSON CORRELATION COEFFICIENTS FROM THE DEEP GULCH CONGLOMERATE. CORRELATION COEFFICIENTS FOR ELEMENTS WERE COMPUTED USING LOG TRANSFORMATIONS BASED ON DATA FROM 220 SUBSURFACE AND 200 OUTCROP SAMPLES.

Subsurface Samples

	U(nt)	Th	Th/U	Fe	Ti	Y	Zr	K	Ce	La*	Pb*	P
U(nt)		.83	−.05	.65	.02	.85	.81	−.16	.70	.53	.55	.38
Th	.85		.42	.73	.04	.91	.92	−.19	.72	.48	.47	.46
Th/U	.12	.56		.24	.00	.26	.36	−.14	.16	.17	.02	.16
Fe	.72	.65	.17		.11	.74	.71	−.35	.62	.51	.55	.57
Ti	.42	.36	.19	.39		.18	.15	.62	.17	.09	−.01	.47
Y	.86	.83	.30	.65	.40		.89	−.13	.73	.49	.48	.55
Zr	.85	.82	.31	.68	.38	.89		−.17	.73	.57	.49	.45
K	.17	.12	.09	.08	.72	.06	.10		.03	−.03	−.20	−.06
Ce	.50	.42	.13	.41	.21	.61	.56	−.09		.98	.40	.46
La*	.51	.56	.50	.37	−.14	.48	.44	−.27	.94		.39	.47
Pb*	.64	−.27	.37	.63	−.39	.33	.48	.03	.40	.64		.40
P	.63	.59	.19	.75	.40	.60	.61	− 23	.46	.44	.54	

Subsurface Samples

Outcrop Samples

Outcrop Samples

*Approximately 80 analyses were available for La and Pb correlations.

TABLE 1.10. MATRIX OF PEARSON CORRELATION COEFFICIENTS FOR RADIOACTIVE CONGLOMERATES OF THE MAGNOLIA FORMATION FROM THE ONEMILE CREEK AREA. CORRELATION COEFFICIENTS FOR ELEMENTS WERE COMPUTED USING LOG TRANSFORMATIONS.

Subsurface Samples

	U(nt)	Th	Th/U	Fe	Ti	Y	Zr	K	Ce	La	Pb	P	Ca	Nb	V	Max. Peb. Size	Max. Peb. Size, Unit 5	Mean Conc. (ppn)
U(nt)		.70	−.38	.57	.20	.77	.53	.06	.53	.59	.73	.45	−.14	56	−.12	.02	−.07	150.1
Th	.63		.40	.48	.[illegible]1	.89	.73	.09	.83	.84	.74	.45	−.37	.75	−.37	−.21	−.25	157.3
Th/U	−.27	.57		−.13	−.[illegible]1	.16	.27	.07	.41	.14	−.23	.01	−.32	.14	−.33	−.34	−.26	1.64
Fe	.57	.35	−.16		.48	.61	.41	−.04	.41	.52	.55	.54	.01	.46	.24	.20	.24	.237*
Ti	.43	.10	−.33	.52		.31	.33	.41	.22	.28	.07	.43	.17	.30	.65	−.09	−.16	1308
Y	.66	.82	.32	.35	.[illegible]4		.72	.09	.80	.81	.80	.61	−.30	.77	−.21	−.19	−.30	21.4
Zr	.55	.75	.35	.23	.25	.77		.18	.65	.67	.47	.56	−.22	.53	−.14	−.14	−.20	121.1
K	.00	.32	−.40	.36	.74	−.29	−.23		.07	.08	.02	−.03	−.17	.04	.39	−.18	−.17	2.00*
Ce	.51	.88	.54	.23	−.02	.84	.77	−.42		.99	.64	.57	−.35	.70	−.41	−.33	−.41	131.5
La	.45	.90	.58	.31	.00	.80	.75	−.33	1.0		.65	.47	−.28	.80	.07	−.15	−.25	125.6
Pb	.61	.79	.27	.38	.[illegible]3	.57	.50	−.15	.68	.69		.24	−.11	.64	−.11	.03	.05	79.23
P	.69	.64	.07	.55	.29	.73	.61	−.09	.66	.65	.64		.13	.60	.01	.10	−.23	167.0
Ca	.19	.21	.06	.16	.24	.16	.07	.10	.17	.05	.28	.30		−.18	.28	−.02	−.12	.096*
Nb	.23	.57	.47	.13	.08	.55	.44	.36	.57	54	.60	.36	.27		−.18	−.16	−.33	25.9
V	.29	.18	.53	.49	.70	−.10	.06	.67	.22	.04	.08	.22	.17	.40		−.31	.36	33.9
Max. Peb. Size	.27	.11	−.16	.35	.44	.09	−.02	.33	.04	.15	.08	.15	.09	.04	.43			
Mean Conc. ppn	25.3	104.9	4.89	1.67*	1274	10.4	127.8	1.93*	161.8	163.0	63.1	231.1	.076*	22.3	37.03			

*wt %

Outcrop Samples

Outcrop Samples

Surface Samples

SUMMARY OF CONCLUSIONS

RESOURCE EVALUATION AND ECONOMIC POTENTIAL

This project has identified 3860 tons of U_3O_8 with a cut-off of 100 ppm U_3O_8 and 8350 tons ThO_2 with a cut-off of 100 ppm ThO_2 in Precambrian quartz-pebble conglomerate deposits of the Sierra Madre and the Medicine Bow Mountains of Wyoming. There is an estimated .10 probability that the actual resources in the Sierra Madre are two or more times the estimates given. Similarly, it is estimated that there is a probability of .45 that the resources in the Medicine Bow area are twice or more those given. In the Medicine Bow Mountains in particular, large areas that are known to be underlain by radioactive quartz-pebble conglomerates have not been drilled and, although our geologic evaluation suggests that these areas do not have the potential of the Onemile Creek area of the northeastern Medicine Bow Mountains, so much of the Medicine Bow Mountains is covered with glacial drift that it remains a good target for future exploration.

The most important occurrence of uranium identified in the Sierra Madre and Medicine Bow Mountains is a 5 km^2 area referred to as the Onemile Creek area of the northeastern Medicine Bow Mountains. The Onemile Creek prospect which is located about 5 km south of the village of Arlington, contains 1801 tons U_3O_8 and 1106 tons ThO_2. There are a number of horizons of quartz-pebble conglomerate that exceed 2 meters in thickness and average more than 100 ppm uranium, but drilling is inadequate to determine the extent of these beds.

The current economic depression in uranium prices (± $25.00 per pound in 1981) makes it very unlikely that uranium of this grade could be mined by subsurface methods at any time in the near future. However, the uranium in quartz-pebble conglomerates is present as coffinite and uranothorite which are metamict and obviously quite soluble, inasmuch as the uranium has been strongly leached from surface outcrops to a depth of about 50 meters. Therefore, the feasibility of chemical leaching of uranium should be investigated as part of future economic appraisals. The conglomerates are strongly foliated and intensely fractured which would make leaching more feasible.

Thorium reserves are significant in both the Sierra Madre and Medicine Bow Mountains but until a ready market is developed for thorium, perhaps by greater use as a fuel in nuclear reactors, there is insufficient demand for thorium to develop these reserves. Furthermore, large reserves of thorium are present in the United States (Staatz and others, 1979; Staatz and others, 1980) in various other types of deposits which can be mined at lower cost than the quartz-pebble conglomerates of this area.

The detailed sampling and analyses necessary for the determination of the amount of gold in the quartz-pebble conglomerates have not been done in either the Sierra Madre or Medicine Bow Mountains. 314 samples have been analyzed by neutron activation and 77 samples contained > 0.0 ppm gold. The range in gold values was from 0.01 ppm to 10.0 ppm with an average gold content of 0.37 ppm. In the Dexter Peak area of the Sierra Madre one sample collected for studies made by the United States Geological Survey contained 10 ppm gold (Houston, 1979) which was the highest gold value determined in any locality. The Sierra Madre was thus considered a promising area for large sample tests. Two samples from the Carrico Ranch locality weighing 89 and 110 pounds were disaggregated and the heavy fraction analyzed for gold.

The method of analyses took into account grain size and sample sparcity factors (Clifton and others, 1969) and thus the values obtained were considered an accurate measure of the gold content of the sample. These two samples contained < 1 ppm and < 1 ppm gold. Therefore, results obtained to date suggest that the quartz-pebble conglomerates of the Sierra Madre and Medicine Bow Mountains are not promising as a source of gold, but we again emphasize that neither sampling methods nor sample size is adequate for statistically valid gold determinations.

At the present time (1981) none of the areas studied in either the Sierra Madre or Medicine Bow Mountains contains enough uranium, thorium, or gold to be mined at today's prices. Inasmuch as the primary reserves are thorium, which is not in great demand at present, it appears that a combination of all values (uranium, thorium, and gold) is also inadequate to promote development. The deposits, however, do represent substantial reserves for the future.

SUGGESTIONS FOR PROSPECTING

There are two reasons to consider additional prospecting in the Sierra Madre and Medicine Bow Mountains: 1) higher grade deposits of uranium or gold may be found, or 2) a future price change for uranium, gold, or thorium might make low-grade deposits mineable. In addition, this area is part of a uranium province (Houston, 1979; Stuckless, 1979) and quartz-pebble conglomerates are certainly not the only type of uranium deposit that may occur in Precambrian rocks.

Deposit type, localities, and justifications for further prospecting in the Sierra Madre and Medicine Bow Mountains are summarized in Table 1.11.

TABLE 1.11. SUGGESTIONS FOR PROSPECTING.

Medicine Bow Mountains

Type Deposit	**Locality**	**Justification**
Quartz pebble conglomerate	Magnolia Formation from Rock Creek to the confluence of Little Brush Creek and Brush Creek	Surface outcrops scarce and drilling inadequate to evaluate potential
Quartz pebble conglomerate	Magnolia Formation west of confluence of Little Brush Creek and Brush Creek	Arkosic paraconglomerate of lower Magnolia Formation radioactive. Overlying quartz pebble conglomerate not exposed, but can be prospected by shallow drill holes
Quartz pebble conglomerate	Magnolia Formation east of Arrastre Lake	Radon anomalies suggest deposits at depth, but not verified by one drill hole
Quartz pebble conglomerate	Magnolia Formation headwaters of North Fork Rock Creek	Radon anomalies suggest deposits at depth, but not verified by one drill hole
Vein deposits	Felsic igneous rock west of Lewis Lake	Sulphide veinlets in felsic igneous rock contains up to 1000 ppm uranium in several localities (Houston and others, 1979, p. 46)
Unconformity	Unconformity at base of Nash Fork Formation from Rock Creek Ridge to Sourdough Creek	Late Early Proterozoic unconformity. Uranium-bearing sulphide veinlets in felsic igneous rock occur at this unconformity
Tertiary Roll-type deposit	Paleocene and Early Eocene arkosic sandstones east and north of Precambrian outcrops	Uranium leached from quartz pebble conglomerate could be redeposited in Tertiary arkosic sandstone
Uranium in faults	In Shear Zone and related faults north and south of major fracture	Radon, rock chip, and stream sediment anomalies in the vicinity of shear zone and other faults. Probability good that remobilized uranium might be found in late faults inasmuch as quartz pebble conglomerate constitutes a major source for uranium
Graphite	Graphite schist in Phantom Lake Metamorphic Suite and in Nash Fork Formation (Houston and others, 1968)	Graphite good collector for mobilized uranium
Quartz pebble conglomerate	Magnolia Formation from Dexter Peak east to Bridger Peak	Inadequate drilling to test potential—gold possibility
Quartz pebble conglomerate	Jack Creek Quartzite on north side of synclinorium	Inadequate drilling to test potential—gold and thorium best prospects
Unconformity	At southern limit of outcrop of metasedimentary rocks	Unconformity at base of Nash Fork Formation equivalent (Slaughterhouse Formation)
Uranium in faults	In Shear Zone and related faults north and south of this major fracture	Same as in Medicine Bow Mountains
Graphite	In Slaughterhouse Formation	Same as in Medicine Bow Mountains

COMPARISON WITH KNOWN DEPOSITS

Table 1.12 is a comparison of the various features of radioactive conglomerates of southeastern Wyoming with a genetic model for Precambrian uranium-bearing fossil placers which is based on data from known deposits of uranium-, thorium- and gold-bearing quartz-pebble conglomerates. As can be seen by inspection of Table 1.12, both Archean and Early Proterozoic deposits of the Sierra Madre and Medicine Bow Mountains are similar in basic characteristics to the model. All deposits are within the correct age bracket and all were deposited in fluvial environments. The best fit to the model is the Magnolia Formation of the Medicine Bow Mountains, where both source rock and environment of deposition fit the model well. We suspect that these Magnolia Formation deposits will ultimately be of economic interest, although from an economic viewpoint it is unfortunate that a more proximal or more completely reworked facies does not crop out. A major disappointment is the Magnolia Formation of the Sierra Madre, where the beds are probably too distal to be of economic interest, or as discussed above, our stratigraphic correlations between the Sierra Madre and Medicine Bow Mountains may not be entirely correct.

The Deep Gulch Conglomerate of the Sierra Madre is considered a classic example of the wrong source rock. Quartz-pebble conglomerates of the Deep Gulch Conglomerate are better developed and probably a more proximal facies than those of the Magnolia Formation of the Medicine Bow Mountains, but the percentage of uranium-bearing heavy minerals is too low to be of economic interest. We believe that this conglomerate was deposited prior to the formation of uranium-rich Late Archean granites which we view as the primary source of uranium minerals of the Wyoming Province.

TABLE 1.12. COMPARISON OF RADIOACTIVE CONGLOMERATES IN SOUTHEASTERN WYOMING WITH THE GENETIC MODEL FOR FORMATION OF PRECAMBRIAN URANIUM-BEARING CONGLOMERATES; ▭ BOXES INDICATE IMPORTANT DISCREPANCIES.

	MEDICINE BOW MOUNTAINS		SIERRA MADRE	
MODEL (HOUSTON AND KARLSTROM, 1980)	ROCK MOUNTAIN CONGLOMERATE	MAGNOLIA FORMATION	DEEP GULCH CONGLOMERATE	MAGNOLIA FORMATION
I Age constraints a) older than 2000 m.y.	I. Age a) older than 2500 m.y.	I Age a) 2500 - 2000 m.y.	I Age a) older than 2600 m.y.	I Age a) 2600 - 2000 m.y.
II Source area constraints a) Late Archean granites provide U b) greenstone belts provide pyrite, Au	II Source area a) mainly Archean granodioritic gneisses, no Late Archean granite to supply U b) older quartzite and mafic volcanic rocks also present	II Source area a) Late Archean granites only locally present b) gneisses and Phantom Lake Suite rocks were important source rocks.	II Source area a) mainly Th-rich granodioritic to tonalitic gneisses b) some older metasediments and volcanics	II Source area a) mainly Th-rich gneisses b) Phantom Lake Suite rocks were important local sources
III Stratigraphic constraints a) occur within Early Proterozoic-type quartz-rich clastic successions b) may occur above basal Proterozoic unconformity c) may occur in Late Archean volcano-sedimentary successions d) may be related to cyclic stratification sequences reflecting transgressions and regressions e) older than thick Proterozoic iron formations and stromatolitic dolomites	III Stratigraphy a) occurs near base of Archean Phantom Lake Metamorphic Suite b) occurs well below basal Proterozoic unconformity c) Phantom Lake Suite is 60% clastic rocks, 40% metavolcanic rocks d) appears to be a coarsening-upward succession which is overlain by transgressive sands of Bow River Qtzite e) older than Nash Fork dolomites	III Stratigraphy a) basal unit of Proterozoic Deep Lake Group b) directly above basal Proterozoic unconformity c) unconformably overlies Phantom Lake Suite d) base of a fining-upward succession; Deep Lake Group deposition appears to reflect cyclic sedimentation e) older than Nash Fork dolomites	III Stratigraphy a) basal unit of Phantom Lake Suit; overlain by 800m thick Jack Creek Quartzites b) well below basal Proterozoic unconformity c) Phantom Lake Suite is 60% clastic, 40% metavolcanic d) appears to be a coarsening-upwards succession overlain by transgressive quartzite and marble of Jack Creek Qtzite e) older than Slaughterhouse dolomites	III Stratigraphy a) basal unit of Deep Lake Group b) directly above basal Proterozoic unconformity c) unconformably overlies Phantom Lake Metamorphic Suite d) only locally developed, not apparently related to cyclic deposition e) older than Slaughterhouse dolomites
IV Sedimentological constraints a) associated with fan deltas and braided river deposits b) paleocurrents unimodal in given areas but may vary over the depositional basin reflecting different entry points c) mineralization is always related to regional or local unconformities d) regressive depositional conditions are most favorable for concentration and preservation of heavy minerals	IV Sedimentology a) proximal alluvial fan deposition b) no paleocurrent data available; no crossbedding preserved c) contact relationships poorly understood d) rapid deposition then burial by transgressive deposition prohibited extensive concentration of placers	IV Sedimentology a) Conglomerate Member alluvial fans; Quartzite Member braided river deposits b) mainly southwest paleocurrents; minor modes directed west, east, or south possibly reflecting basin-edge influence c) most radioactive conglomerates unconformably overlie Late Archean granite, conglomerates overlying Phantom Lake Suite rocks less radioactive d) fluvial deposition was relatively short-lived, giving way to transgressive marine sedimentation up-section and down-paleoslope	IV Sedimentology a) broad, wet-alluvial fan deposition b) paelocurrent patterns complex due to structural problems; directions are variable but predominantly southerly c) appears to unconformably overlie "basement" gneisses d) alluvial fan deposition was short-lived and of limited lateral extent, giving way to marine deposition laterally and up-section	IV Sedimentology a) perhaps alluvial fan sediments although deposition not well understood b) paleocurrents south and southeast c) presumably unconformably overlies Phantom Lake Suite although contact relationships are complex (units interfinger or are infolded together); no mineralized conglomerates are known d) conglomerates, if present at all, give way rapidly up-section and laterally to marine phyllites of Singer Peak Fm.
V Lithologic characteristics a) drab colors, no redbeds b) sericitic quartz-pebble conglomerates contain the most uranium c) moderately sorted, moderately mature lithologically; the most radioactive conglomerates are the most mature rocks within the fluvial succession	V Lithology a) outcrops oxidized (red); subcrops drab b) sericitic quartz-pebble conglomerate and arkosic pebble to cobble polymictic paraconglomerate c) poorly sorted, matrix supported, arkose to subarkose composition, rock fragments of schist and quartzite, few cleaner quartz-pebble conglomerates	V Lithology a) reddish in outcrop due to oxidized pyrite, greenish in drillcore b) sericitic quartz-granule and -pebble conglomerate; polymictic arkosic paraconglomerate c) moderate to poor sorting; most radioactive rocks are quartz-pebble conglomerates; paraconglomerates are much less radioactive	V Lithology a) rust-red in outcrop, green in subcrop b) sericitic quartz-pebble conglomerates interbedded with quartz-granule conglomerates c) moderately sorted, moderately mature lithologically (subarkose)	V Lithology a) outcrops oxidized, subcrops drab b) granular quartzite and micaceous quartzite; very few pebble conglomerates c) moderately sorted, arkosic to subarkosic
VI Mineralogical characteristics a) pyrite in matrix b) uraninite is commonly the main uranium mineral c) thorium minerals present d) gold may be present e) carbonaceous material may be present	VI Mineralogy a) pyrite occurs in drillcore, scattered throughout the matrix of conglomerates b) *uranium minerals if present, unidentified* c) monazite, zircon d) no gold detected e) ?	VI Mineralogy a) pyrite may be up to 35% of matrix b) coffinite is the only identified U-mineral c) *thorite, monazite, zircon* d) maximum gold was 0.50 ppm e) some carbonaceous material identified	VI Mineralogy a) pyrite may be up to 30% of matrix b) uranium minerals unidentified c) monazite, zircon d) maximum gold value was 10 ppm e) ?	VI Mineralogy a) only scattered pyrite b) uranium minerals unidentified c) few monazite, zircon, and other heavies d) no gold e) ?
VII Tectonic setting a) related to intracratonic rift-valley basins b) related to (younger) continental margin sedimentation	VII Tectonic setting a) alluvial fans may be related to basin-edge deposition b) tectonic setting of Phantom Lake Suite depositional basin unknown	VII Tectonic setting a) intracratonic rift-valley deposition b) lowest units of platform and miogeoclinal sedimentary succession on south margin of Wyoming Province	VII Tectonic setting a) alluvial fans may be related to basin-edge deposition b) tectonic setting of Phantom Lake Suite poorly understood	VII Tectonic setting a) rift-related deposition but only limited subaerial sedimentation b) lowest units of miogeoclinal deposition on south margin of Wyoming Province
VIII Preservation and metamorphism a) metamorphic grade should be no higher than amphibolite facies	VIII Metamorphism a) amphibolite facies	VIII metamorphism a) greenschist and amphibolite facies	VIII Metamorphism a) amphibolite facies	VIII Metamorphism a) greenschist and amphibolite facies

GEOLOGIC RESULTS AND REGIONAL CORRELATIONS

Prior to this investigation, the general geology of Precambrian rocks of the Sierra Madre and Medicine Bow Mountains was as well known as for any large area of the Wyoming Province. During this study, additional stratigraphic, sedimentological, and structural data plus new information on geochronology have allowed us to make great strides in interpreting the geologic history, and we anticipate that explorationists will be able to apply this information in the search for mineral deposits.

The Sierra Madre and Medicine Bow Mountains have the most complete succession of Late Archean and Early Proterozoic metasedimentary and metavolcanic rocks of any locality in the Rocky Mountains and, very likely, in all of North America. We believe that this thick metasedimentary succession is preserved here because this was the site of Early Proterozoic rifting where thick wedges of clastic sedimentary rocks accumulated. This area also appears to be one where collision occurred between island arcs, that developed to the south, and the rifted margin of the Wyoming Province so that the Early Proterozoic metasedimentary rocks may have been carried to a deeper level in the crust at the collision site, and thus preserved.

Figure 1.20 is a summary of the stratigraphy of Late Archean-Early Proterozoic metasedimentary and metavolcanic rocks of the Sierra Madre and Medicine Bow Mountains as compared with key areas of metasedimentary rocks elsewhere in the Wyoming Province and in other parts of North America, Greenland, and Scandinavia. The Sierra Madre-Medicine Bow Mountain section contains up to 16,540 meters of metasedimentary and

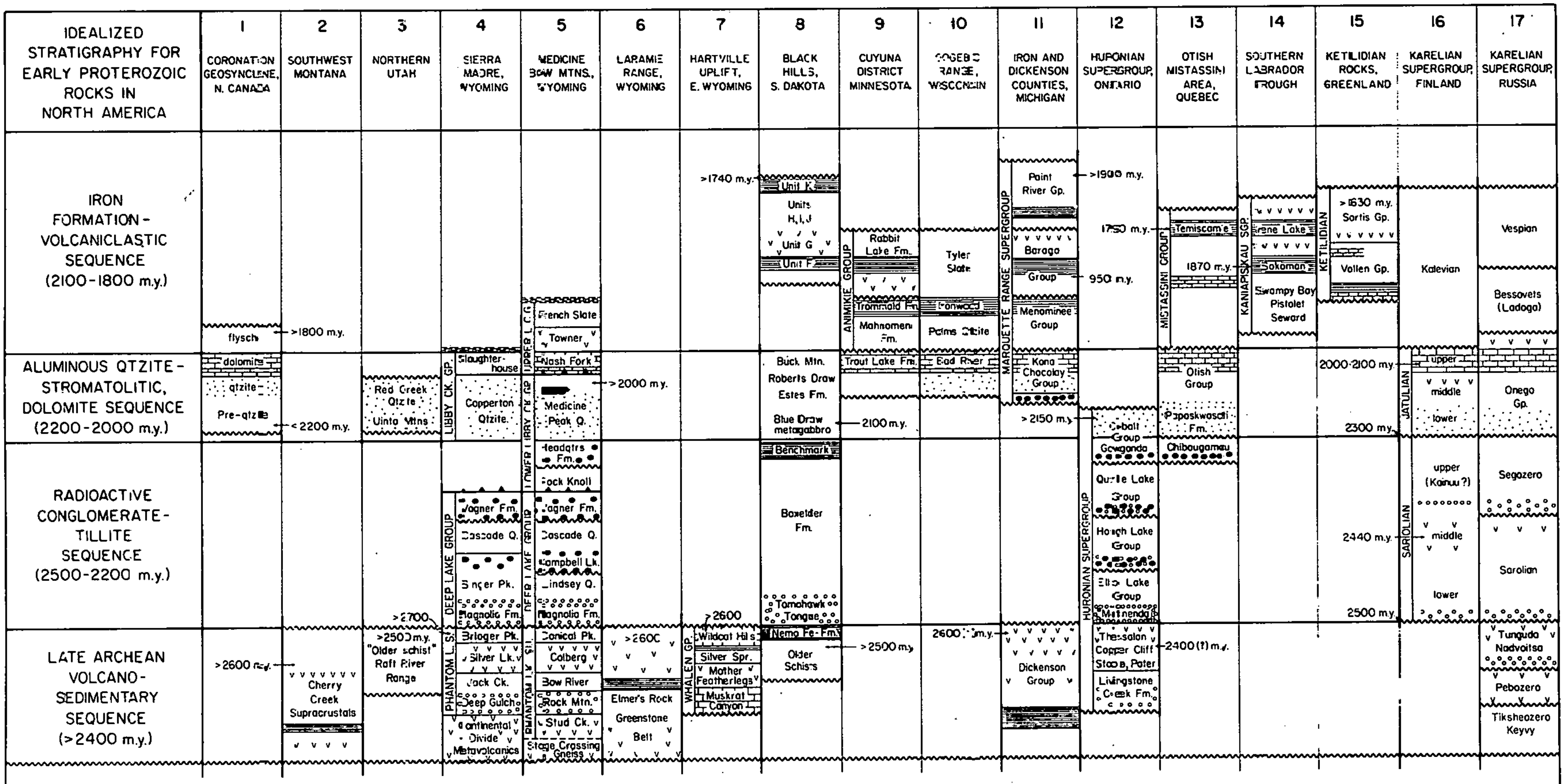


Figure 1.20. Regional lithostratigraphic correlation and idealized stratigraphy of Late Archean and Early Proterozoic supracrustals near the margins of proto-North America. Key lithologies are shown: vvvv = volcanogenic rocks; ⁂ = radioactive quartz-pebble conglomerate; ●•● = diamictite (tillite?); ⋯ = aluminum-rich quartzite; ▤ = stromatolitic dolomite; ☰ = iron formation. References: 1. Hoffman and others (1974) 2. Vitaliano and others (1979) 3. Graff and others (1980) Compton (1975) 4. Graff (1978; 1979) 5. Karlstrom and Houston (1979a; 1979b); Lanthier (1979) 6. Graff and others (1981) 7. Snyder (1980) 8. Redden (1980); Kleinkopf and Redden (1975); Bayley (1970, 1972) 9. Marsden (1972) 10. Sims (1976) 11. James (1958); James and others (1961) 12. Robertson (1976); Roscoe (1969) 13. Roscoe (1973); Chown and Caty (1973); Fryer (1972) 14. Dimroth (1970); Fryer (1971) 15. Bondeson (1970) 16. Meriläinen (1980); Simonen (1960) 17. Salop (1977).

metavolcanic rocks that range in age from about 2800 m.y. to about 1800 m.y. The closest lithologic analog that is time compatible with the entire Sierra Madre-Medicine Bow section, from the base of the Phantom Lake Metamorphic Suite to the top of the Sugarloaf Quartzite of the Lower Libby Creek Group, is the Huronian Supergroup of southern Ontario. Formations that are notably alike are the Matinenda Formation of southern Ontario and the Magnolia Formation of southern Wyoming; the Gowganda Formation of southern Ontario and the Headquarters Formation of southern Wyoming, and the Lorrain Quartzite of the Cobalt Group of southern Onario and the Medicine Peak Quartzite of southern Wyoming (Figure 1.20, Localities 4, 5, and 12). What is more, the Huronian Supergroup and the southeastern Wyoming succession both contain cyclic sedimentary sequences with basal paraconglomerates which suggest similar and probably time-correlative climatic or tectonic events (Young, 1973; Houston and others, 1979; Karlstrom and Houston, 1979b). Formations of the Upper Libby Creek Group of southern Wyoming are very similar lithologically and are time compatible with rocks of the Marquette Range Supergroup of the Lake Superior region. Of particular interest is the correlation between stromatolitic dolomites: Nash Fork Formation (Figure 1.20, Loc. 5); Trout Lake Formation (Figure 1.20, Loc. 9); Bad River Dolomite (Figure 1.20, Loc. 10); Kona Dolomite (Figure 1.20, Loc. 11); and others (Figure 1.20, Loc. 1, 13, 16, 17).

The intriguing aspect of the regional correlation is that the Sierra Madre-Medicine Bow section correlates far better with rocks of the Huronian Supergroup located over 1800 kilometers east of southeastern Wyoming than it does with any of the sections within the Wyoming Province. Our inability to make convincing lithostratigraphic correlations with the Wyoming Province may be attributed to incomplete geochronology and geologic mapping and burial of much of the Precambrian section west of the Mississippi. However, our regional tectonic interpretation also suggests that, while the southern margin of the Wyoming Province and the mid-continent area experienced similar tectonic histories involving rifting followed by collisional orogenesis, much of the rest of the Wyoming Province was a stable cratonic block.

This geologic problem has economic ramifications because, as shown in Figure 1.20, Late Archean-Early Proterozoic mineral deposits such as iron formation and uraniferous quartz-pebble conglomerate are time-bound and exploration for these and other strata-bound mineral deposits is dependent on our understanding of stratigraphy and geologic history. If we assume that the regional stratigraphic correlation (Figure 1.20) is correct, it is necessary to assume that in the area between the north shore of Lake Huron where the Huronian Supergroup rocks were deposited and southeastern Wyoming (or possibly the Black Hills) no rocks of the radioactive conglomerate-tillite sequence (Figure 1.20) were deposited. Either these rocks were removed by erosion or they are not exposed. We will state our current preferences in explaining this problem below, but our interpretations will no doubt be revised as new geochronologic and geologic evidence is compiled in the future.

Late Archean sedimentation and volcanism probably involved microplate movements too complex to interpret regionally with our present knowledge. However, continental and marine rock successions such as the Livingstone Creek Formation and Thessalon Volcanics of southern Canada and the Phantom Lake Metamorphic Group of southeastern Wyoming may have been deposited near continental margins. By Early Proterozoic time, we believe that rifting began at a continental margin that extended from northern Utah to Quebec. The configuration and exact location of the continental margin is unknown, but the main rifted margin was probably at least 100 kilometers south of current exposures of Early Proterozoic metasedimentary rocks on the north shore of Lake Huron and southeastern Wyoming, and was south of exposures of Archean basement in Wisconsin. The rifting episode and accompanying sedimentation probably began about 2400 m.y. ago and continued episodically, perhaps as late as about 1900 m.y. ago. Early sedimentation was continental and radioactive quartz-pebble conglomerates were deposited at this time in fault-bounded rift valleys in southern Canada, perhaps in an area south of what is now Wisconsin, in southeastern Wyoming, and in an aulacogen that may have developed in the Black Hills area. It is quite possible that radioactive quartz-pebble conglomerates were deposited over a far more extensive area than indicated above and that preservation was only possible in rifted areas. Sedimentation continued and, as rifting proceeded, marine incursions took place which were interrupted by periods of continental glaciation -- the most notable of which is recorded in the Gowganda Formation of southern Ontario, Headquarters Formation of southeastern Wyoming, and Chibaugamau Formation of the Otish-Mistassini area (Figure 1.20).

By about 1900-2000 m.y. the rifting was complete and the continental block that was south of the rifted area was transported elsewhere. Marine sedimentation continued along the continental margin and included the development of carbonate banks and widespread carbonate deposition as indicated by the widespread carbonate beds at the top of the aluminous quartzite-dolomite sequence (Figure 1.20). A major difference between the Sierra Madre-Medicine Bow succession and the marine deposits in the Black Hills, the Lake Superior Region of southern Canada, and the Labrador Trough is that marine deposits laid down in the later areas between about 2000 m.y. and 1900 m.y. appear to be essentially intracratonic in the sense that there is recognizable older continental crust on both sides of the metasedimentary basins (Sims and others, 1981; Dimroth, 1972). Furthermore, the "intracratonic" deposits all contain iron formation whereas no significant iron formation is present in the Sierra Madre-Medicine Bow area. We suggest that the only true continental margin deposits of the 2000-1900 m.y. age preserved from Wyoming to southeastern Canada are those thrust over older beds (Upper Libby Creek Group) in the Sierra Madre-Medicine Bow area. Elsewhere (Black Hills, Lake Superior region, Labrator Trough) deposits preserved are in basins that developed within an essentially coherent lithospheric plate.

The Marquette Range Supergroup sedimentation (about 2000-1900 m.y.) of the Lake Superior region has recently been interpreted in a plate tectonic context by Larue and Sloss (1980, p. 1451, Figure 2). They contend (Larue and Sloss, 1980, p. 452), "That the regional characteristics of the Marquette Range Supergroup cannot be readily explained by deposition on a simple planar surface. Instead, considerable evidence exists for sedimentation in separate basins and on intervening platforms, and for repeated basinal subsidence. The eastern upper Michigan sedimentary basins show features more commonly associated with high-angle faulting (extensional tectonics) than with flexural subsidence, including evidence of local sediment sources, lithologic features, and turbidite sedimentation. Because the sedimentary basins in eastern upper Michigan are tens of kilometers in length and show relation to probable extensional tectonics, they are considered to be rift related. In contrast, the Animikie basin (Minnesota) is an elongate southwest-trending basin, hundreds of kilometers in length, which probably deepened toward the southwest. Because it is not known how or why the Animikie basin terminates in the south, it cannot be resolved at this point whether the Animikie basin is an aulacogen or a basin characterized by some other subsidence mechanism and depositional style (for example, a wrench-fault basin). The overall sedimentary history of the Marquette Range Supergroup can perhaps best be explained by sedimentation on and deformation of a rifted passive margin." Essentially Larue and Sloss (1980) propose a depositional history for the Marquette Range Supergroup very much like that proposed by us for the Medicine Bow-Sierra Madre area except that the basins that are preserved in Minnesota and Michigan are within the craton, whereas the Sierra Madre-Medicine Bow Mountains sediments appear to be nearer the actual rifted margin.

If we consider the depositional episode of the Huronian Supergroup (~ 2400-2100 m.y.) and the Marquette Range Supergroup (~ 2100-1900 m.y.) as both related to a major episode of sedimentation along a rifted continental margin, it is reasonable to consider sedimentation in southeastern Wyoming and in the Black Hills as part of the same episode. The primary differences might simply be timing of rifting and sedimentation in rifted basins and the configuration of rift basins with respect to the continental margin. The earliest basins to develop are assumed to be in southeastern Wyoming, on the north shore of Lake Huron and in the Black Hills aulacogen, where rocks of the radioactive conglomerate-tillite sequence are preserved. These rocks are believed to have been deposited in intracratonic rift valleys or in marine settings near the continental margin. Intracratonic rifts developed later on the craton and sedimentary and volcanic rocks of the iron formation-volcaniclastic sequence were deposited in Michigan and Minnesota in these rift-basins and also in the Black Hills aulacogen. There is no record that these rocks were deposited in southeastern Wyoming or on the north shore of Lake Huron. However, sedimentation is thought to have continued south of the current outcrops of Early Proterozoic rocks in southeastern Wyoming and the north shore of Lake Huron, probably along the Early Proterozoic continental margin. The only record of these rocks is the Upper Libby Creek Group rocks which were thrust over the older succession in the Sierra Madre-Medicine Bow area.

Deformation that resulted in cessation of Early Proterozoic sedimentation in the Lake Superior region is referred to as the Penokean orogeny. According to Van Schmus (1976), this deformation, metamorphism and igneous activity took place between 1800-1900 m.y. In the Sierra Madre-Medicine Bow area and in the Black Hills, deformation took place or continued to about 1700 m.y. We believe that these and other 1700-1800 m.y. orogenies in North America were related to closure of intracratonic rifts activated by collision of island arcs and microcontinents with the edge of the rifted Early Proterozoic North American continent. These orogenic events appear to have occurred diachronically from about 1900 to about 1700 m.y., somewhat like the events that constitute the Taconic-Appalachian orogeny of eastern North America. The deformation involved closure of basins within the craton as well as deformation at the continental margin. We suggest that Penokean deformation in the mid-continent and orogenesis in the Black Hills and Labrador Trough represent the former whereas deformation in southern Wyoming represents the later.

This overview of sedimentation does not present an especially promising picture with respect to exploration for additional deposits of radioactive quartz-pebble conglomerates in the United States. It is conceivable that younger deposits of the Marquette Range Supergroup bury older deposits that may contain radioactive quartz-pebble conglomerate in some of the Michigan and Minnesota basins, but we suspect that most quartz-pebble conglomerates were deposited south of present-day outcrops in Minnesota and Wisconsin.

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PART TWO

THE GEOLOGY OF ARCHEAN AND EARLY PROTEROZOIC TERRANES OF THE MEDICINE BOW MOUNTAINS, WYOMING

By: K.E. Karlstrom
R.S. Houston
C.M. Coolidge
A.J. Flurkey
C.K. Sever

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INTRODUCTION

Part 2 is a detailed discussion of the metasedimentary rocks in the Medicine Bow Mountains. Discussions of the stratigraphy and sedimentary features of the metasedimentary rocks is based on data gathered during this investigation and previous studies of stratigraphy and sedimentology, most notably: Blackwelder (1926); Houston and others (1968); Karlstrom and Houston (1979a, 1979b); and Lanthier (1979). These discussions are followed by an interpretation of the depositional history of sedimentary rocks in the Medicine Bow Mountains which employs a plate tectonic model. This section also tries to relate the Wyoming occurrences of uranium-bearing conglomerates to other known occurrences in North America through lithostratigraphic correlation and comparison of tectonic settings. Tectonic history is summarized for the Medicine Bows; this summary considers detailed structural analysis of folding in metasedimentary rocks (in part from Karlstrom and Houston, 1979b) with respect to a plate tectonic model for southeastern Wyoming (modified from Hills and Houston, 1979) which involves continent-island arc collisions at 1700 m.y. and perhaps also near the end of the Archean.

These discussions of the geology of the metasedimentary rocks sets the stage for detailed evaluations of uranium-bearing conglomerates in the Medicine Bow Mountains. Two radioactive units have been indentified: the Rock Mountain Conglomerate and the Magnolia Formation. The former is Archean, not strongly radioactive, and is treated only briefly. The Magnolia Formation, on the other hand, is locally strongly radioactive and is discussed more fully. These conglomerates are the most strongly radioactive conglomerates found so far in the Medicine Bow Mountains and Sierra Madre and are the only rocks which approach economic grades of uranium. Lastly, we summarize the character of the Onemile Creek conglomerates and compare them to known deposits of uranium-bearing conglomerates. In doing so, we examine the validity of existing genetic and exploration models for Precambrian uranium-bearing quartz-pebble conglomerates (summarized by Houston and Karlstrom, 1980) and propose a few modifications of these models.

GEOLOGIC SETTING OF METASEDIMENTARY ROCKS IN THE MEDICINE BOW MOUNTAINS

Precambrian rocks form the core of the Medicine Bow Mountains, a north-trending anticlinal uplift is southeastern Wyoming which was uplifted during the late Cretaceous Laramide Orogeny. These Precambrian rocks range in age from over 2500 m.y. to about 1400 m.y. and preserve a record of over a billion years of geologic history. The Medicine Bow Mountains contain the most complete and best preserved record in western North American of this period of geologic history and, as such, is an important area for studies of Late Archean and Early Proterozoic sedimentation and tectonics in western North America.

Precambrian rocks in the core of the Medicine Bow Mountains, shown in Figure 2.1, are divided into two geologic and geochronologic terranes by the northeast-trending Mullen Creek-Nash Fork shear zone, a zone of cataclastic rocks ranging in width from one to seven kilometers (Houston and McCallum, 1961; Houston and others, 1968; Hills and Houston, 1979). This shear zone, with similar shear zones in the Sierra Madre to the west (Houston and others, 1975; Graff, 1978; 1979) and the Laramie Mountains to the east (Graff and others, (1981), forms the southern boundary of the Archean Wyoming Province, a geochronologic province which encompasses Precambrian rocks in Wyoming and adjacent states (Condie, 1976; Hills and

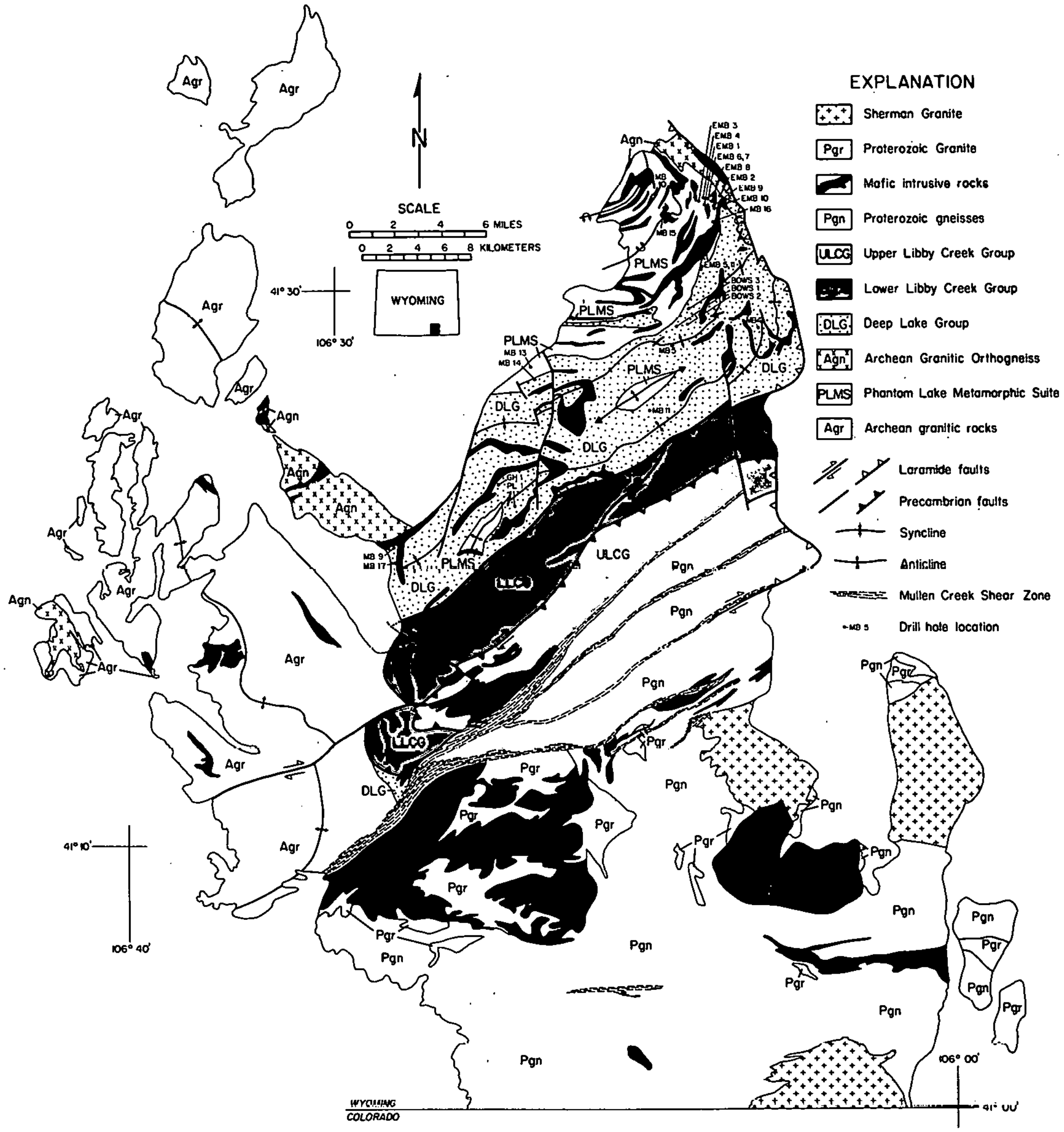


FIGURE 2.1. GENERALIZED GEOLOGIC MAP OF PRECAMBRIAN METASEDIMENTARY ROCKS OF THE MEDICINE BOW MOUNTAINS, WYOMING

Armstrong, 1974; Hills and Houston, 1979; Houston and Karlstrom, 1980). Therefore, rocks exposed in the Medicine Bow Mountains offer the opportunity to study the geologic history along this ancient crustal boundary.

Rocks north of the Mullen Creek-Nash Fork shear zone consist of an Archean gneissic terrane which was strongly metamorphosed about 2500 m.y. ago (Hills and others, 1968) overlain nonconformably by more than 13 km of quartz-rich metasedimentary rocks. Although the age of these metasedimentary rocks is not precisely known, they range in age from Late Archean (more than 2500 m.y. old) to about 1700-1900 m.y. We have divided the metasedimentary rocks into three successions: the Phantom Lake Metamorphic Suite, the lower two units of which are intruded by tonalites and granodiorites of presumed Archean age; the Early Proterozoic Deep Lake Group which unconformably overlies Phantom Lake Suite metasediments and presumed Archean granitic rocks (Houston and others, 1968; Karlstrom, 1977; Karlstrom and Houston, 1979a;1979b); and the Early Proterozoic Libby Creek Group (Blackwelder, 1926; Houston and others, 1968) which is now interpreted to be in thrust fault contact with older units (Lanthier, 1978; 1979).

Metasedimentary rocks south of the shear zone are mainly hornblende and quartzo-feldspathic gneisses with minor sillimanite gneiss and calc-silicate (McCallum, 1964; Hills and Houston, 1979). These gneisses are believed to be partly paragneisses (Houston and others, 1968, p. 58) that are now complexly intruded by a variety of intrusive rocks, most notably layered gabbroic complexes, quartz-diorite, and synorogenic granodiorite. Details of a possible sedimentary origin of the gneisses are uncertain because of the high degree of deformation and amphibolite facies metamorphism which has obliterated primary features. However, the overall compositions of the gneisses are similar to basic to intermediate volcanics and volcanogenic sediments, in marked contrast to the mature siliciclastic metasediments north of the shear zone. Available geologic and geochronologic data permit the interpretation that paragneisses south of the shear zone are island arc-derived sediments and volcanics that were complexly intruded by synorogenic plutonic rocks, strongly deformed, and metamorphosed to amphibolite grade, during a continent-island arc collision about 1700 m.y. ago (Hills and Houston, 1979). The age of gneisses south of the shear zone is reasonably well constrained by geochronologic data: gneisses are known to be older than the 1700-1800 m.y. intrusives which cut them (Hills and others, 1968) and are believed to be younger than about 1900 m.y. because no date older than this has emerged from extensive geochronologic studies of Precambrian rocks in Colorado, Arizona, New Mexico, and southern California (Peterman and others, 1968; Barker and others, 1976; Silver and others, 1977; White, 1978; Condie and Budding, 1979). A detailed discussion of this southern terrane is outside the scope of this study but it is briefly discussed later, in the section on the tectonic history of the Medicine Bow Mountains.

STRATIGRAPHY AND SEDIMENTARY FEATURES OF METASEDIMENTARY ROCKS

As in most complex geologic terranes, stratigraphic nomenclature for metasedimentary rocks in the Medicine Bow Mountains has been modified many times over the years in response to more detailed mapping studies. Table 2.1 compares nomenclatures used by previous workers to nomenclature used in this paper. Use of the term "metamorphic suite" follows amendments to the American code of stratigraphic nomenclature presented by Henderson and others (1980). In this section we present new formal names for lithostratigraphic units within the Phantom Lake Metamorphic Suite. These units are of formational status but contact relationships are often enigmatic and no type sections are defined (see Henderson and others, 1980). The following section describes each stratigraphic unit in the northern Medicine Bow Mountains from oldest to youngest, emphasizing lithofacies distributions and sedimentary structures. Figure 2.2 summarizes our stratigraphy.

ARCHEAN "BASEMENT" ROCKS

Quartzo-feldspathic gneisses

The oldest rocks in the Medicine Bow Mountains are believed to be a heterogeneous assemblage of quartzo-feldspathic gneisses, hornblende gneisses, biotite gneisses, and quartzites which crop out in the western part of the mountains (labeled Agr in Figure 2.1). Lithologic descriptions of these rocks are in Houston and others (1968). This gneissic terrane has not been studied in detail and its history is still poorly understood. Quartzo-feldspathic gneisses have yielded a Rb-Sr whole rock age of 2500 ± 50 (Hills and others, 1968) but this may date the time of metamorphism of the gneisses (Hills and Houston, 1979). However, by

THIS REPORT		KARLSTROM AND HOUSTON (1979 a; b); LANTHIER (1979)		HOUSTON AND OTHERS (1968)		BLACKWELDER (1926)
UPPER LIBBY CREEK GROUP	FRENCH SLATE	LIBBY CREEK GROUP	FRENCH SLATE	LIBBY CREEK GROUP	FRENCH SLATE	FRENCH SLATE
	TOWNER GREENSTONE		TOWNER GREENSTONE		TOWNER GREENSTONE	TOWNER GREENSTONE
	NASH FORK FORMATION		NASH FORK FORMATION		NASH FORK FORMATION	RANGER MARBLE ANDERSON PHYLLITE NASH MARBLE SERIES
LOWER LIBBY CREEK GROUP	SUGARLOAF QUARTZITE		SUGARLOAF QUARTZITE		SUGARLOAF QUARTZITE	SUGARLOAF METAQUARTZITE
	LOOKOUT SCHIST		LOOKOUT SCHIST		LOOKOUT SCHIST	LOOKOUT SCHIST
	MEDICINE PEAK QUARTZITE		MEDICINE PEAK QUARTZITE		MEDICINE PEAK QUARTZITE	MEDICINE PEAK METAQUARTZITE
	HEART FORMATION		HEART FORMATION		HEART FORMATION	HEART METAGRAYWACKE
	HEADQUARTERS FORMATION		HEADQUARTERS FORMATION		HEADQUARTERS SCHIST (included units now mapped as Vagner)	HEADQUARTERS SCHIST (included units now mapped as Vagner)
	ROCK KNOLL FORMATION	DEEP LAKE GROUP	ROCK KNOLL FORMATION	DEEP LAKE FORMATION		DEEP LAKE METAQUARTZITE
DEEP LAKE GROUP	VAGNER FORMATION		VAGNER FORMATION			
	CASCADE QUARTZITE		CASCADE QUARTZITE		marble	
	CAMPBELL LAKE FORMATION		CAMPBELL LAKE FORMATION			
	LINDSEY QUARTZITE		LINDSEY QUARTZITE			
	MAGNOLIA FORMATION		MAGNOLIA FORMATION		quartzite	
PHANTOM LAKE METAMORPHIC SUITE	CONICAL PEAK QUARTZITE	PHANTOM LAKE METAMORPHIC SUITE				
	COLBERG METAVOLCANIC ROCKS		UPPER PHANTOM LAKE METAMORPHIC SUITE		meta-conglomerate	
	BOW RIVER QUARTZITE					
	ROCK MOUNTAIN CONGLOMERATE					
	STUD CREEK VOLCANICLASTIC ROCKS		LOWER PHANTOM LAKE METAMORPHIC SUITE		metavolcanic rocks	
	?—?—?—?—?—?—?					
	STAGE CROSSING GNEISS					

TABLE 2.1. COMPARISON OF STRATIGRAPHIC NOMENCLATURES FOR METASEDIMENTARY ROCKS IN THE MEDICINE BOW MOUNTAINS

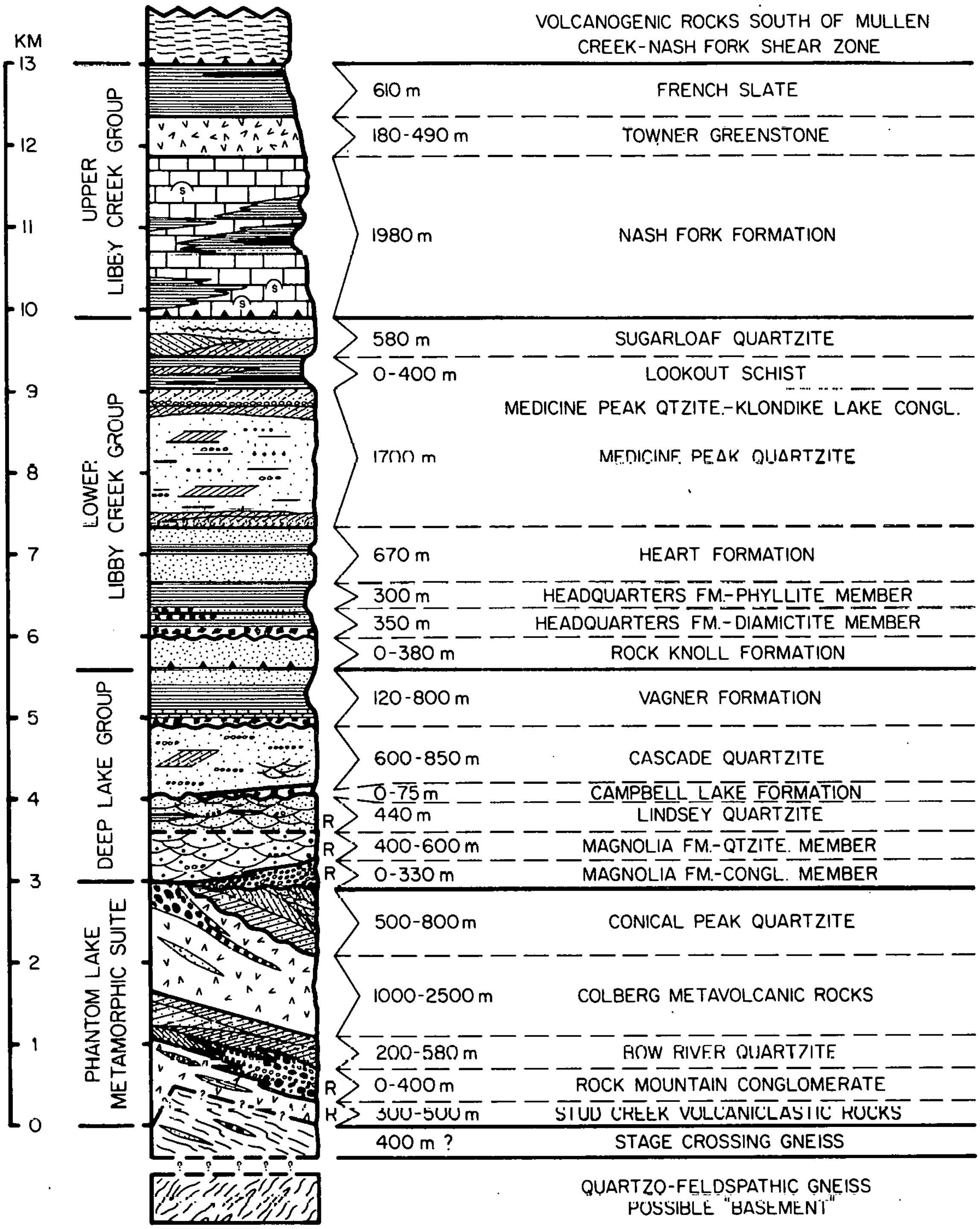
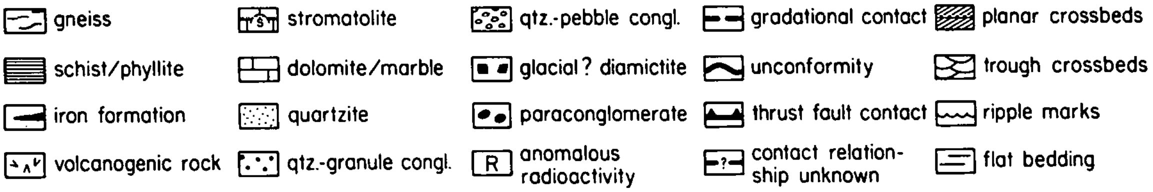


FIGURE 2.2. STRATIGRAPHIC COLUMN OF METASEDIMENTARY ROCKS IN THE MEDICINE BOW MOUNTAINS, WYOMING



analogy to similar gneissic terranes in the Sierra Madre to the west (Divis, 1976; 1977), the Granite Mountains to the north (Peterman and Hildreth, 1978), and the Laramie Mountains to the east (Hills and Armstrong, 1973; Johnson and Hills, 1976); we believe that the gneisses in the Medicine Bow Mountains may record a complex history involving deposition of mafic to intermediate volcanic and volcaniclastic protoliths, perhaps as early as 2950 m.y. ago; metamorphism of the protoliths accompanying intrusion of tonalitic magmas around 2700 m.y. ago; and metamorphism and anatexis of gneisses around 2500 m.y. ago. The important point to this section is that we consider at least part of the gneissic terrane to have been basement which supplied detritus during deposition of Late Archean and Early Proterozoic sediments.

Stage Crossing Gneiss

In the northern Medicine Bow Mountains is a succession of hornblende and biotite gneisses, here named the Stage Crossing Gneiss after the Overland Trail stage crossing of Rock Creek, near Arlington. This unit is interpreted to be Archean because it is crosscut by granite which is similar to the 2400-2500 m.y. old Baggot Rocks granite of the western Medicine Bow Mountains (Hills and others, 1968; Hills and Houston, 1979). The Stage Crossing Gneiss is in close proximity to metasedimentary rocks of the Phantom Lake Metamorphic Suite near Rock Creek (Plate 1) but contact relationships are not well exposed. As shown in Plate 1, we interpret the contact to be a fault contact because of the abrupt change from Phantom Lake quartzites to hornblende gneiss across the north fork of Onemile Creek (NE1/4, Sec. 1, T.18 N., R.79 W. and NW1/4, Sec. 6, T.18 N., R.78 W.). However, there are hornblende schists in the lower Phantom Lake Suite which are similar to hornblende gneisses in the Stage Crossing Gneiss, and we cannot rule out the possibility that the two units may be in part gradational. We believe the Stage Crossing Gneiss to be older than the Phantom Lake Suite because it is in a structurally lower position in the core of French Joes anticlinorium (Plates 1, 3), but we do not know how much older. The Stage Crossing Gneiss might be much older and a remnant of a once more extensive greenstone terrane, or it may represent volcanism and sedimentation in the earliest stages of Phantom Lake Suite deposition.

Petrographic and geochemical data from the Stage Crossing Gneiss are summarized in Table 2.2. Hornblende gneisses which comprise about 80% of the unit, are dominated by hornblende needles with interstitial quartz, plagioclase (An_{53}-An_{60}), and epidote (± sphene, chlorite, magnetite, and rutile). Garnet forms large porphyroblasts in some samples. Biotite gneisses are dominated by plagioclase, quartz, biotite, and muscovite (± kyanite, chlorite, orthoclase, apatite and magnetite). Some of the Stage Crossing Gneiss is definitely metasedimentary: the biotite gneisses locally appear to be medium-bedded and may be metagraywackes (turbidites); and we have found discontinuous outcrops of paraconglomerate, quartzite, and garnet schist. However, the bulk of the unit is hornblende gneiss of unknown origin. The petrology is compatible with volcanic (basalt), plutonic (gabbro), and sedimentary or volcaniclastic (graywacke or tuff) protoliths and all of these rock types may be present in the unit as mapped in Plate 1. The maximum map thickness of the gneiss is about 1200 m. Stratigraphic thickness is unknown but probably much less than this due to repetition by folding.

TABLE 2.2. PETROGRAPHIC DATA FROM THE STAGE CROSSING GNEISS; DATA PARTLY POINTCOUNTS FROM KING (1963); OTHER DATA FROM VISUAL ESTIMATES.

Hornblende gneisses

Sample No.	Amph.	Qtz.	Plag.	Epid.	Sphne.	Opaq.	Garnt.
K26	75.4	14.4	8.9	T	–	1.2	–
K52	85.0	3.9	5.2	4.9	T	T	–
K129	29.1	37.0	24.9	–	T	3.7	5.1
K141	74.0	6.8	12.0	3.0	4.2	T	–
K744	70.3	16.3	9.8	.3	T	3.3	–
K80-14	64	5	30	–	–	1	–
MEAN	66.3	13.9	15.3	1.4	.7	1.5	.9

Biotite gneisses

Sample No.	Plag.	Qtz.	Biot.	Musc.	K-spar	Chlor.	Kyan.	Apat.
K80-2	41	31	8.4	17.7	–	1.3	.5	T
K80-3	48	35	13	–	4	T	–	–

PHANTOM LAKE METAMORPHIC SUITE

The Phantom Lake Metamorphic Suite was defined by Karlstrom and Houston (1979a, 1979b) as the sequence of metasedimentary and metavolcanic rocks which underlies the Deep Lake Group. In those papers we had a poor understanding of the structure and stratigraphy within the sequence (hence the name metamorphic suite) and divided it into a lower part which was dominantly volcanogenic and an upper part which was dominantly quartzite. More detailed work has shown this simplified division to be unusable. In this paper, we define five lithostratigraphic (and in part tectonostratigraphic) units within the Phantom Lake Metamorphic Suite (Table 2.1). This stratigraphy, combined with top and bottom criteria and geometrical analysis of folding, leads to a structural interpretation involving large-scale overturned anticlinoria and synclinoria (Plates 1 and 3). We retain the name metamorphic suite because of numerous uncertainties in our stratigraphic and structural interpretations and the realization that, because of isoclinal folding, amphibolite-facies metamorphism, and rapid facies changes within this sequence, future detailed work may necessitate modification of our stratigraphy. The major features of each unit are summarized below.

Unit 1: Stud Creek Volcaniclastic Rocks

The oldest unit of the Phantom Lake Metamorphic Suite, here named the Stud Creek Volcaniclastic Rocks, crops out in two localities in the core of a large, overturned, doubly-plunging anticlinorium (French Joes anticlinorium of Plate 3). The northern outcrop area, centered near Rock Mountain (Sec. 1, T. 18 N., R. 79 W.) is in the south-plunging part of the structure; the southern outcrop area, centered near Stud Creek (Sec. 15, T. 18 N., R. 79 W.), is in the north plunging and badly faulted part of the fold (Plate 1). Both outcrop areas are characterized by a heterogeneous assemblage of metavolcanic and metasedimentary rocks.

Petrographic data are summarized in Table 2.3 and geochemical analysis of one metabasalt is given in Table 2.7. In a general way, the wide variety of lithologies can be lumped into four groups: Pelitic schists (50%), amphibolitic schists (30%), quartzites and conglomerates (20%), and calcareous rocks (local). Pelitic rocks (Table 2.3) are predominantly biotite- and muscovite-rich quartz schists but there are also schists containing garnet, staurolite, chloritoid, or kyanite. Pelitic rocks were probably tuffs and graywackes. Amphibolitic rocks include plagioclase-rich varieties which were probably basalts and quartz-rich varieties, including garnet amphibole quartz schist, which were probably mafic tuffs. One sample (Table 2.7) has the chemical composition of a tholeiitic basalt (Figure 2.6). Quartzites include fine-grained micaceous quartzites, fuchsitic quartzites, granule conglomerates, and quartz-rich schistose paraconglomerates. Both types of conglomerates are slightly radioactive in the southern outcrop area (twice background gamma radiation and up to 21 ppm U; 29 ppm Th). Calcareous rocks range from impure marble to calcareous pelitic schist and are also radioactive in one isolated area (170 ppm U, 16 ppm Th in one locality near Rock Mountain).

TABLE 2.3. PETROGRAPHY OF THE STUD CREEK VOLCANICLASTIC ROCKS; DATA FROM THE NORTHERN AREA PARTLY POINTCOUNTS FROM KING (1963); OTHER DATA FROM VISUAL ESTIMATES.

	Sample No.	Qtz.	Plag.	Amp.	Epid.	Biot.	Musc.	Calc.	Chl.	Opaq.	Staur.	Garnt.	Hem.	Chltd.
PELITIC SCHISTS	Northern Area													
	K325	26.9	–	–	10.6	34.2	–	26.5	–	T	–	–	–	–
	K526	37.4	–	–	.3	24.9	32.3	2.6	–	2.1	–	–	–	–
	K78-14	80.1	–	–	.4	7.4	9.6	1.4	–	1	–	–	–	–
	K78-87	46	–	–	–	25	3	24	–	2	–	–	–	–
	K78-88	27.6	51	–	T	16	9	–	–	1.4	–	–	–	–
	K537	68.8	–	–	6.8	13	10.6	–	–	.8	–	–	–	–
	Southern Area													
	K80-8	45	12	–	–	8	8	8	–	–	–	–	–	–
	K80-10	55	–	–	14	12	5	4	–	T	5	–	–	–
	K80-11	87	–	–	–	–	12	–	–	1	–	–	–	–
	K78-5	47	50	T	–	–	–	–	2	1	–	–	–	–
	K78-6	48	–	–	–	–	38	–	9	2	3	–	–	–
	K78-7	44	–	–	–	T	56	–	T	T	–	–	–	–
AMPHIBOLE SCHISTS	Northern Area													
	K156	5	1.8	87.9	.2	–	–	–	–	5.1	–	–	–	–
	K157	19.7	13.9	55.8	1	–	–		.2	6.4		–	–	–
	K78-15	30	–	46	–	–	1	4	–	2	–	15	–	2
	Southern Area													
	60-46	44	–	29	23	3	–	1	–	T	–	–	–	–
	60-49	5	41	46	1	–	–	–	–	3	–	–	3	–
	K700	15	–	54	30	–	–	1	–	–	–	–	–	–
	K80-9	34	–	60	1	2	–	.5	.5	2	–	–	–	–

Sedimentary features are poorly preserved in the quartzites, conglomerates and calcareous rocks; bedding is sometimes recognizable and rare crossbeds were seen. Rocks of the Stud Creek Volcaniclastic Rocks probably represent depositional environments ranging from fluvial (as suggested by radioactive conglomerates) to shallow marine (carbonates) but the large variability of rock types, rapid facies changes, the absence of thick and continuous layered graywackes and the absence of pillow basalts suggests to us that subaerial deposition and volcanism may have predominated.

Stratigraphic relationships between facies within the Stud Creek Volcaniclastic Rocks and the overall thickness of the unit are difficult to interpret because of isoclinal folds on every scale, the inferred presence of large strike faults in the southern outcrop area, and the strong superposed F_3 folds in the northern outcrop area (Plate 1). Similarly, stratigraphic relationships with the underlying Stage Crossing Gneiss and overlying Rock Mountain Conglomerate are complicated and may be in part gradational. Nevertheless, we estimate from Plate 1 a maximum exposed stratigraphic thickness ranging from 330 m (northern area) to 500 m (southern area).

Unit 2: Rock Mountain Conglomerate

The Rock Mountain Conglomerate crops out on Rock Mountain and near the extreme northwestern limit of Precambrian outcrop (Secs. 27, 33, T. 19 N., R. 79 W.). The unit is absent in the Stud Creek area as a continuous stratigraphic unit although lithologically similar conglomerates occur as lenses within the Stud Creek Volcaniclastic Rocks

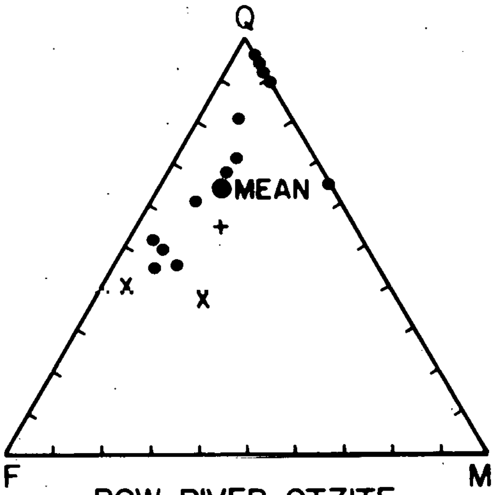
(Plate 1). The conglomerates are locally anomalously radioactive in outcrop (up to 20000 counts per minute or 5 times local background and values up to 270 ppm U, 95 ppm Th) and were sampled extensively and drilled twice near Rock Mountain to test their favorability as a uranium target. The results, discussed in detail in Karlstrom and Houston (1981), indicate that the unit contains thin, lenticular radioactive conglomerates but is not generally a favorable target for uranium mineralization.

Petrographic data for the Rock Mountain Conglomerate are summarized in Table 2.4 and shown graphically in Figure 2.3. The unit is predominantly granular to pebbly muscovitic quartzite with paraconglomerate beds ranging from less than a meter to several hundred meters thick. Paraconglomerates contain stretched clasts of quartz, quartzite, amphibolite and schist in a strongly foliated quartz-muscovite (locally arkosic) matrix. Particularly striking are bright green fuchsitic schist clasts. The paraconglomerate is locally garnetiferous. Contact relationships with the underlying Stud Creek Volcaniclastic Rocks are not exposed but the conglomerates and quartzites in the Stud Creek Volcaniclastic Rocks are similar enough to the Rock Mountain Conglomerate to suggest a gradational (or complexly infolded) relationship. In the extreme northwestern area of Plate 1, the Rock Mountain Conglomerate appears to be intruded by tonalitic rocks of probable Archean age and to be conformably overlain by the Bow River Quartzite. The maximum stratigraphic thickness of the Rock Mountain Conglomerate is about 400 meters in both outcrop areas.

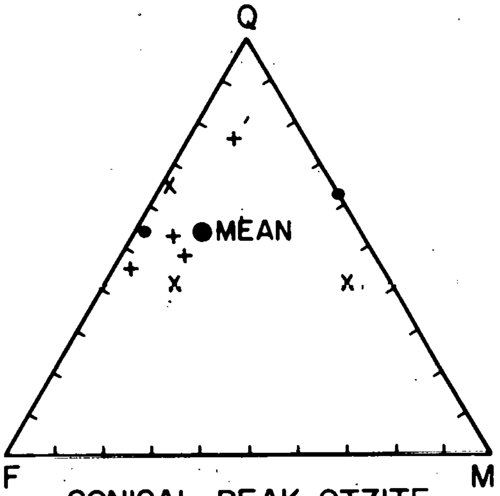
TABLE 2.4. PETROGRAPHY OF THE ROCK MOUNTAIN CONGLOMERATE; MODES FROM VISUAL ESTIMATES.

Sample No.	Qtz.	Musc.	Kspar.	Plag.	Opaq.	Garnt.	Zirc.	Biot.	Chlor.
ROCK MOUNTAIN AREA (EASTERN OUTCROP AREA)									
K240	56.6	21.8	.1	25.1	.2	–	–	–	–
MB15-498	72	25	–	–	3	–	–	–	–
MB15-600	65	32	2	–	1	–	–	–	–
MB15-750	52	28	–	18	2	–	–	–	–
MB15-775	68	30	–	2	T	–	–	–	–
MB15-825	68	20	–	11	1	–	–	–	–
K78-91	83	10	–	–	2	5	T	–	T
K78-92	47	46	–	–	1	–	T	–	6
K78-93	47	48	–	3	2	–	–	–	T
MB10-430	53	27	–	9	2	2	–	–	7
MB10-455	58	39	–	–	T	T	T	3	T
MEAN	60.9	29.7	T	6.2	1.3	.6	T	.3	1.2
FOOTE CREEK AREA (WESTERN OUTCROP AREA)									
SR78-3	60	27	–	–	1	2	–	–	–
SR78-5	49	28	20	–	3	–	–	–	–
K80-5	65	33	–	–	–	2	–	–	–
SR78-16	37	4	16	40	3	–	–	T	T
SR78-17	40	–	8	30	T	–	–	22	T
SR78-18	25	–	–	12	3	plus 60% Amph.			
MEAN (without SR78-18)	50.2	18.4	8.8	16.4	2	.8	–	4.4	T
GRAND MEAN	57.6	26.2	2.8	9.4	1.5	.7	T	2.1	.8

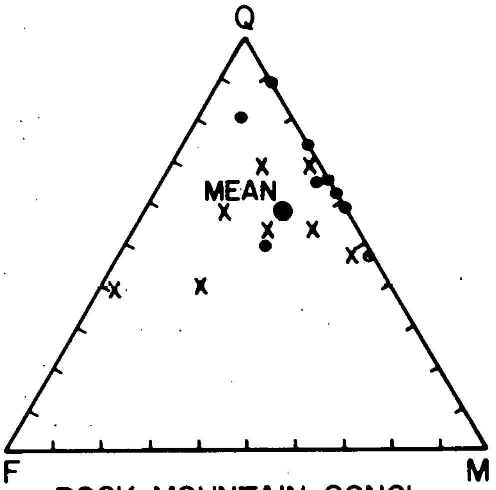
The Rock Mountain Conglomerate was penetrated in two drill holes: MB-10 and MB-15, both near Rock Mountain. In MB-10 pebbly chlorite schist and poorly sorted sericitic quartzite of the uppermost Rock Mountain Conglomerate appear to be gradational with quartzites and thin conglomerates of the overlying Bow River Quartzite. MB-15 penetrated 118 m of a coarsening upwards succession ranging from poorly sorted quartzite at the base to paraconglomerate in the upper half. In general, the Rock Mountain Conglomerate appears to coarsen to the east and upwards.



BOW RIVER QTZITE.



CONICAL PEAK QTZITE.



ROCK MOUNTAIN CONGL.

Figure 2.3. Quartz-feldspar-mica ternary diagrams for sand and granule size fractions of siliciclastic rocks of the Phantom Lake Suite; modes from visual estimates.

- • Kspar > 2/3 total feldspar
- x plag. > 2/3 total feldspar
- \+ plag. = 1/3 - 2/3 total feldspar

The lack of sedimentary structures in the Rock Mountain Conglomerate, poorly exposed contact relationships, and complex structure makes interpreting the depositional environment difficult. However, we interpret the unit to be a prograding alluvial fan deposit (Rust, 1979) on the basis of poor sorting, coarse grain sizes, coarsening upwards successions, the presence of anomalously high radioactivity which may reflect fossil placer accumulation of uranium- and thorium-bearing heavy minerals, and the limited lateral extent of the unit. The source was apparently a nearby, probably fault-bounded, terrane containing mainly older metasedimentary and metavolcanic rocks such as are found in both the Stud Creek Metavolcanic Rocks and the Stage Crossing Gneiss.

Unit 3: Bow River Quartzite

The Bow River Quartzite is a key unit in deciphering the stratigraphy and structure of the Phantom Lake Suite because it contains abundant planar crossbeds and some oscillation ripple marks which indicate the direction of stratigraphic top. The unit is well exposed throughout the northern Medicine Bow Mountains where it defines the limbs of French Joes Anticlinorium (Plates 1 and 3). It also crops out above the Rock Mountain Conglomerate in the extreme northwestern part of Plate 1.

Petrographic data from the Bow River Quartzite are summarized in Table 2.5 and Figure 2.3. The unit contains quartzites (95%), conglomerates (local), biotite and hornblende schists and phyllites (local) and quartz-rich carbonates (local). The quartzites are mainly very fine- to fine-grained (average grain size = .1 to .2 mm), foliated, muscovitic arkoses and subarkoses. Some contain appreciable biotite and amphibole; and many contain zircon and opaque minerals (probably pyrite altered to hematite). The unit ranges from 200 to 580 m thick with an average thickness of about 350 m. The lower contact, with the Rock Mountain Conglomerate, is not exposed in the Rock Mountain area but appears to be gradational in drill hole MB-10. In contrast, the contact between the Rock Mountain Conglomerate and fine-grained Bow River quartzites in the extreme northwestern area of Plate 1 is sharp and conformable.

TABLE 2.5. PETROGRAPHY OF THE BOW RIVER QUARTZITE; AREAS REFERENCED TO PALEOCURRENT ROSE DIAGRAMS IN PLATE 2; MODAL PERCENTAGES FROM VISUAL ESTIMATES.

Sample No.	Qtz.	Kspar.	Plag.	Musc.	Biot.	Chlor.	Amph.	Opaq.	Zirc.	Garnt.	Calc.
ROCK CREEK AREA (northern diagram); mean grain size = .2 mm											
60-53	80	11	–	7	1	–	–	1	–	–	–
SR78-7	50	42	–	6	2	–	–	T	T	T	T
K80-15	45	46	–	5	4	–	–	T	T	–	–
K80-16	33	–	–	–	5	–	60	2	–	–	T
K80-17	61	23	–	8	2	–	–	1	–	–	–
MB10-105	52	30	13	2	1	–	–	–	T	–	2
MB10-155	55	14	13	4	11	–	–	1	–	1	1
MB10-194	45	39	2	3	10	–	–	T	–	1	–
MB10-230	40	4	50	1	4	1	–	T	–	–	–
MB10-280	37	–	40	–	9	13	–	1	–	–	T
MEAN	49.8	21.4	11.8	3.6	4.9	1.6	6.7	.6	T	T	T
CARLSON CREEK AREA; mean grain size = .15 mm											
K78-18	71	16	–	13	–	–	–	T	T	–	–
60-28	68	18	–	13	–	–	–	1	T	–	–
MEAN	70	17	–	13	–	–	–	T	T	–	–
ARRASTRE CREEK AREA(southern diagram); mean grain size = .13 mm											
59-10	92	–	–	8	–	–	–	T	–	–	–
AC-x	95	–	–	5	–	–	–	T	–	–	–
AC-2	65	–	–	25	–	–	–	T	–	–	–
AC-3	90	–	–	8	–	–	–	2	–	–	–
AC-4	92	–	–	3	–	5	–	T	–	–	–
AC-7	90	–	–	10	–	–	–	T	–	–	–
MEAN	87	–	–	9.8	–	1	–	T	–	–	–
GRAND MEAN	64	13.8	6.6	6.7	2.7	.9	3.7	.3	T	T	T

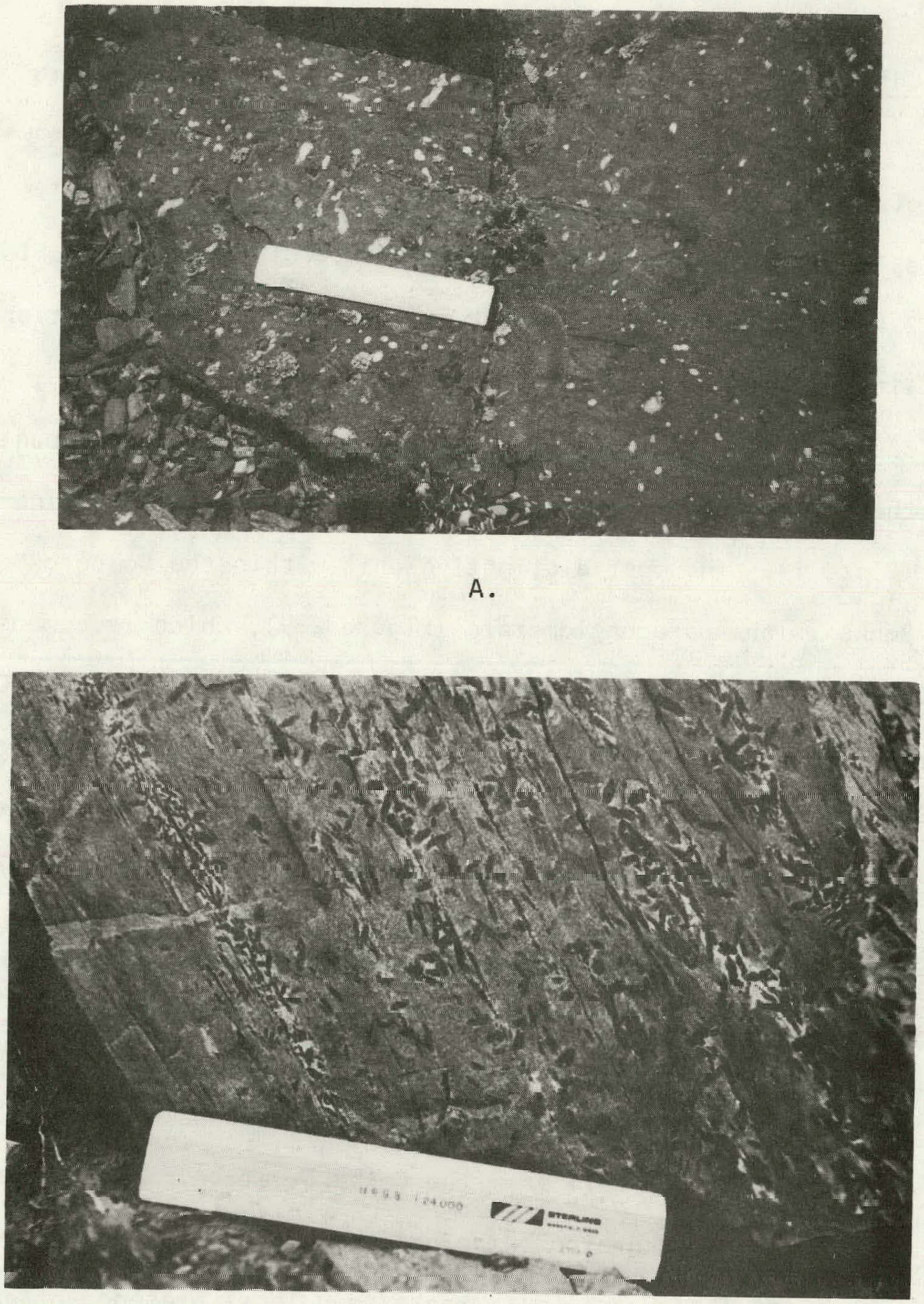
The most prevalent sedimentary structures in the unit are medium- to large-scale planar crossbeds (amplitude about .5-1 m; mean inclination 23°). When combined for the entire unit, these crossbeds yield a bimodal paleocurrent distribution with a prominent mode directed southwest and a secondary mode directed northeast (Plate 2). However, individual outcrop areas show distributions which are somewhat scattered but are dominated by one mode or the other. Several oscillation ripple marks in the unit (Plate 2) confirm a bimodal current pattern but record east-west directed currents. Multiple ripple sets in one outcrop show east-west and north-south directed current oscillations.

We interpret most of the Bow River Quartzites to be shallow marine sediments. The generally fine grain sizes suggest low-energy deposition; large-scale planar crossbeds may be sand waves; and the bimodal paleocurrent distribution probably represents ebb and flood tides. This is compatible with interpretations of a tidally-influenced deltaic depositional environment for the unit. We envision a river-fed embayment or estuarine situation for deposition of the Bow River Quartzite with early, fluvial deposition of the Rock Mountain Conglomerate in the northeastern area, close to a tectonically active highlands, giving way to marine deposition to the south and higher in the section. The source area was contributing mainly sedimentary and volcanic detritus during deposition of the Rock Mountain Conglomerate but appreciable K-feldspar in the Bow River Quartzites suggests that granitic rocks also contributed detritus during Bow River Quartzite deposition. Thus, the source area was probably composed of Archean granitic gneisses and older metasedimentary and metavolcanic rocks.

Unit 4: Colberg Metavolcanic Rocks

The Colberg Metavolcanic Rocks include a heterogeneous assemblage of metavolcanic rocks including: amygdaloidal metabasalts (Figure 2.4A), a few poorly preserved pillow basalts, volcaniclastic schists (Figure 2.4B), fragmental metavolcanic rocks ranging in composition from rhyolitic to basaltic, paraconglomerates and thin quartzites. The unit crops out in large areas of the northern Medicine Bow Mountains and in the core of the Arrastre Anticline in the central Medicine Bow Mountains (Plate 1). The most distinctive unit within the Colberg Metavolcanic Rocks is the paraconglomerate (Figure 2.5), which is composed of varying proportions of rounded granite boulders (up to about 50 cm in diameter), quartzite boulders, and stretched mafic volcanic rock clasts in an amphibole, biotite, quartz matrix. The paraconglomerate unit ranges in stratigrahic thickness up to about 400 m, within which the conglomerates themselves are complexly interbedded with volcanic rocks and quartzites. The entire Colberg Metavolcanic unit ranges from about 100 m thick in the northwest part of Plate 1, to a map thickness of 2500 m in the north-central Medicine Bow Mountains.

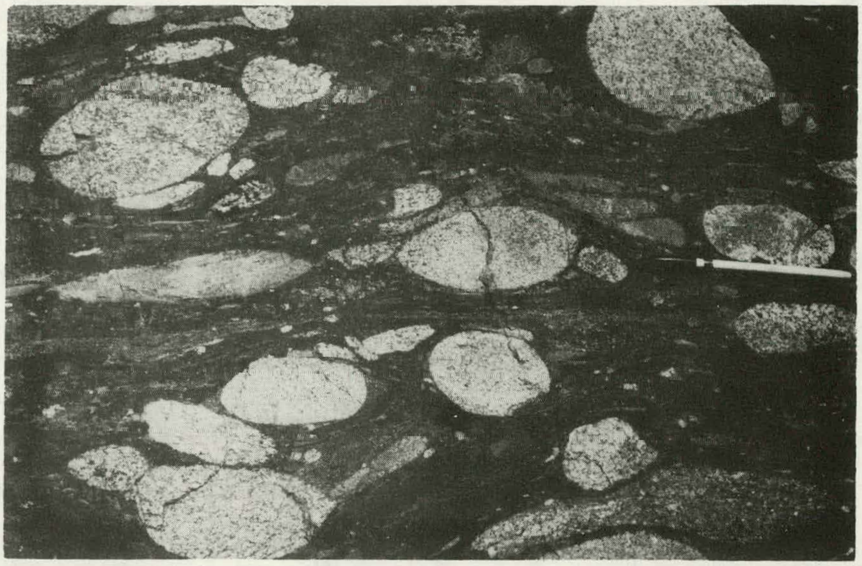
Petrographic data from the Colberg Metavolcanic Rocks are shown in Table 2.6. Fine-grained amphibolites, which were mapped as metabasalt because of their massive character and local presence of amygdules, show a wide compositional range including: plagioclase-amphibole rocks which were probably basaltic flows (e.g. the Arrastre Lake area); quartz-rich amphibolites which probably represent reworked mafic to andesitic tuffs (e.g. the northern area); and amphibole- and biotite-rich quartzites which were probably metagraywackes (e.g. west of Rock



A.

B.

Figure 2.4. Photographs of the Colberg Metavolcanic Rocks: A. amygdules of quartz, with minor epidote and clinozoisite, in metabasalt, S1/2, Sec. 10, T.18 N., R.79 W.; B. amphibole schist from Rock Creek Trail, about 1000 m upstream from confluence of Rock Creek and Deep Creek.



A.

B.

Figure 2.5. Photographs of paraconglomerates from the Colberg Metavolcanic Rocks: A. granite boulder paraconglomerate from Sec. 25, T.18 N., R.79 W.; note that granite boulders are relatively undeformed and mafic schist clasts are strongly stretched parallel to foliation; B. paraconglomerate from Sec. 24, T.18 N., R.79 W. dominated by basaltic clasts.

Creek). Both basaltic rocks and quartz-rich amphibolites locally contain amygdules filled with quartz, epidote, clinozoisite, or calcite (Figure 2.4A). Paraconglomerate and schistose units frequently contain garnet. Plagioclase anorthite determinations made by King (1963) in the northern area range from 49 to 58 (mean of 7 samples was 54).

TABLE 2.6. PETROGRAPHY OF THE COLBERG METAVOLCANIC ROCKS; DATA FROM NORTHERN AREA POINTCOUNTS FROM KING (1963); OTHER DATA FROM VISUAL ESTIMATES.

Sample No.	Amph.	Plag.	Qtz.	Biot.	Epid.	Opaq.	Sphene	Garnt.	Musc.	Zirc.	Probable protolith
Northern Area											
K336	52.5	–	20.1	19.9	7.5	–	–	–	–	–	mafic tuff
K348	64.5	4.8	26.8	–	1.1	2.7	–	–	–	–	mafic tuff
K402	78.2	6.2	12.8	–	1.7	1.2	–	–	–	–	basalt or tuff
K406	38.1	7.4	24.5	19.0	9.0	–	.7	–	–	–	mafic tuff
K413	75.8	11.9	5.3	–	4.5	1.6	.9	–	–	–	basalt
K419	89.0	.5	8.1	–	.8	1.7	–	–	–	–	basalt or tuff
K80-12	–	–	46	7	–	–	–	–	47	T	graywacke
Rock Creek area											
SR78-26A	55	20	20	–	5	T	–	–	–	–	amygdaloidal basalt
SR78-26B	10	–	55	35	T	–	–	–	–	–	graywacke
SR78-27	–	2	84	–	–	3	–	–	3	+3%	subarkose
K78-80	–	–	67	5	–	–	–	–	28		immature sandstone
K78-83	50	20	25	–	5	–	–	–	–		amygdaloidal basalt
K78-84	–	–	60	10	–	2	–	3	25		immature sandstone
Arrastre Lake area											
TS57B	–	65	10	35	–	T	–	5	–		amygdaloidal basalt
TS57C	30	55	10	–	–	5	–	–	+apatite		basalt
TS84	9	72	10	7	–	2	–	–	–		amygdaloidal basalt
TS95	18	64	–	–	5	3	–	–	+10% calcite		amygdaloidal basalt
TS151	25	66	–	5	3	1	–	T	+apatite		basalt
mean of 10 samples of metabasalt from Houston and others 1968	60	8	16	T	T	T	3	–	–	–	plus 6% chlorite
mean of 4 samples of metatuff (?) from Houston and others (1968)	T	7	35	T	T	3	T	T	13	–	plus 26% chlorite

Chemical compositions of metabasalts and metatuffs from the Colberg Metavolcanic Rocks and one sample from the Stud Creek Volcaniclastic Rocks are shown in Table 2.7. Sample 1, from the Stud Creek Volcaniclastic Rocks, has a composition similar to that of an "average" basalt (Cox and others, 1979), but with low total alkali content and therefore a high normative quartz content. Samples 2 and 3 are andesitic, but with more iron and less aluminum than the "average" andesite composition (Cox and others, 1979). The tuffs range in composition from andesitic (sample 4) to rhyodacitic (sample 6) and rhyolitic (sample 5). As shown in Figure 2.6A, an AFM plot of the samples shows an iron enrichment trend characteristic of tholeiitic magmas (Irvine and Baragar, 1971), except that sample 5 is anomalously high in sodium and deficient in magnesium and iron. Figure 2.6B shows that all six samples fall in the tholeiitic field in alkali-silica plots (MacDonald, 1968).

The depositional history of the Colberg Metavolcanic Rocks is not well know. We interpret the unit to be partly marine based on the presence of marine quartzites directly above and below the unit. However, we have found very few pillow basalts and this, combined with the wide variety of lithologies and rapid facies changes in the unit, seem more compatible with subaerial volcanism and volcaniclastic deposition. The depositional environment of the paraconglomerate provides an interesting and unsolved problem. We envision deposition of the Colberg paraconglomerates to have taken place in alluvial or possibly submarine channels and fans which developed adjacent to fault scarps bounding volcanic highlands. If so, the Colberg Metavolcanic Rocks may contain complexly interbedded subaerial and submarine rocks.

TABLE 2.7. GEOCHEMISTRY OF METABASALTS AND METATUFFS FROM THE STUD CREEK AND COLBERG METAVOLCANIC ROCKS; ANALYZED BY STEVE BOESE, UNIVERSITY OF WYOMING.

	METABASALTS			METATUFFS		
	1.	2.	3.	4.	5.	6.
SiO_2	52.9	56.7	55.1	60.3	72.9	67.8
Al_2O_3	13.3	14.0	13.3	12.7	12.2	12.1
CaO	10.42	7.9	7.07	4.97	1.64	0.48
Na_2O	1.61	4.43	3.97	4.28	6.80	4.25
K_2O	0.14	0.30	0.20	0.48	0.10	0.40
Fe_2O_3	11.90	11.75	13.94	14.07	2.42	9.30
TiO_2	0.9	1.0	1.0	1.0	0.5	0.7
MgO	7.44	4.11	5.38	2.19	0.45	2.12
MnO	0.19	0.14	0.19	0.21	0.06	0.08
P_2O_5	0.06	0.17	0.13	0.13	0.21	0.21
Total	98.9	100.5	100.3	100.3	97.3	97.4
ppm U_3O_8	0.1	0.6	0.8	2.2	3.7	2.5
ppm Th	<5	<5	<5	<5	<5	5
			MOLECULAR NORMS*			
Q	12.07	4.64	3.57	13.20	27.61	31.65
OR	0.85	1.80	1.20	2.90	0.60	2.50
AB	5.85	39.90	35.95	39.30	62.35	40.00
AN	34.65	17.48	18.05	14.35	2.53	1.08
Hy	27.52	15.71	23.67	17.89	1.79	16.64
Di	15.36	16.66	13.25	8.06	3.54	—
MT	2.24	2.14	2.54	2.60	0.44	1.76
IL	1.32	1.38	1.4	1.42	0.72	1.02
AP	0.13	0.35	0.27	0.27	0.43	.45
Co	—	—	—	—	—	4.9

1. SR60-25: metabasalt of Stud Creek Volcaniclastic Rocks; north side Carlson Creek, NW¼ Sec. 22, T18N, R79W.
2. SR60-44: andesitic basalt of Colberg Metavolcanic Rocks; 1500 feet S15E of hill 8981, Sec. 3, T18N, R79W.
3. K78-99: amygdaloidal andesitic basalt of Colberg Metavolcanic Rocks; northeast of Colberg cabin; 1400 ft. south of hill 8981, SE¼ Sec. 3, T18N, R79W.
4. SR59-3: andesitic basalt tuff of Colberg Metavolcanic Rocks; northeast of Arrastre Lake, SE¼, NW¼, Sec. 10, T16N, R80W.
5. SR79-8: Colberg rhyolitic tuff; east of Colberg cabin; top of hill 9317, Sec. 10, T18N, R79W.
6. SR79-9: Colberg rhyodacitic tuff; same locality as SR79-8.

*Fe^{+3}/Fe^{+2} approximated at 0.15 (after Cox and others, 1979)

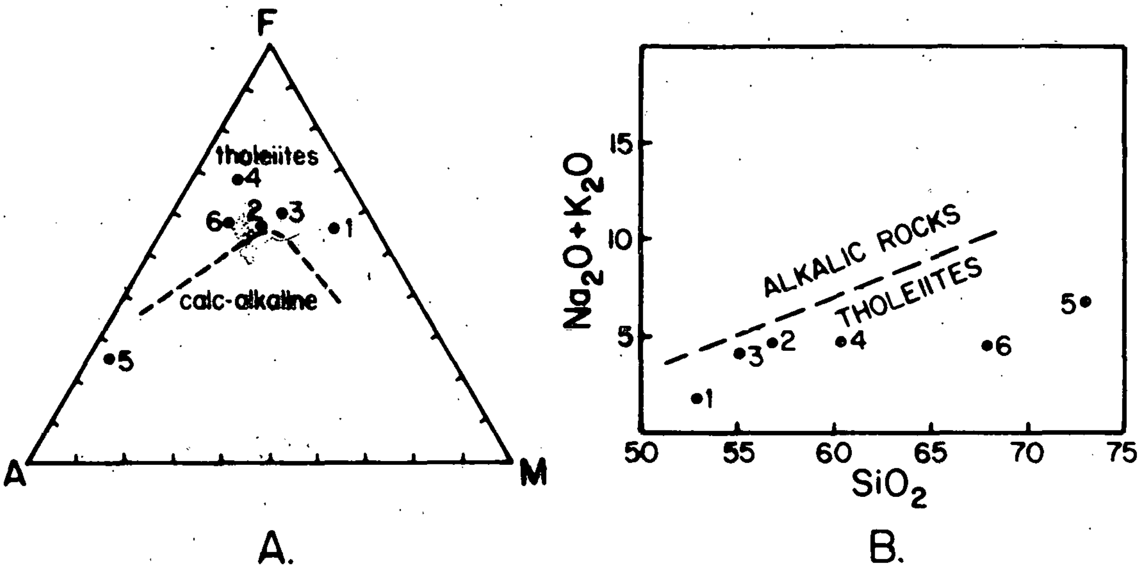


Figure 2.6. Weight percent plots of metavolcanic rocks from the Stud Creek and Colberg Metavolcanic Rocks. A. A = (Na_2O + K_2O); F = (FeO + Fe_2O_3); M = Mgo. The dashed line separates tholeiitic trends (above) from calc-alkaline trends (Irvine and Baragar, 1971). B. Alkalies versus silica. The dashed line separates alkalic rocks (above) from tholeiites (MacDonald, 1968).

Unit 5: Conical Peak Quartzite

The Concical Peak Quartzite is the youngest unit in the Phantom Lake Suite. It occupies the core of the Foote Creek Synclinorium in the northern Medicine Bow Mountains and unconformably underlies the Magnolia Formation of the Deep Lake Group in the northeastern and central parts of Plate 1. Like the underlying Bow River Quartzite, crossbeds in the Conical Peak Quartzite provide key information for stratigraphic and structural interpretations of the Phantom Lake Suite. The dominant lithology is a white, foliated, fine-grained micaceous subarkose. The unit also contains some calcareous quartzite near the

North Fork of Rock Creek. Metabasalts which crop out above the Conical Peak Quartzites near the North Fork of Rock Creek, the Medicine Bow River area near MB-13 and MB-14, and in drill hole EMB-3 are also considered part of the Concical Peak Quartzite.

Petrographic data from the Conical Peak Quartzite are summarized in Table 2.8 and Figure 2.3. The quartzites are very similar to those of the underlying Bow River Quartzite in grain size (uniformly very fine-to fine-grained) but differ compositionally by having more plagioclase in the mode and somewhat less potassium feldspar (Figure 2.3).

TABLE 2.8. PETROGRAPHY OF THE CONICAL PEAK QUARTZITE; AREAS REFERENCED TO PALEOCURRENT ROSES IN PLATE 2; MODAL PERCENTAGES FROM VISUAL ESTIMATES.

Sample No.	Qtz.	Kspar.	Plag.	Musc.	Biot.	Opaq.	Zirc.	Calc.	Garnt.
CONICAL PEAK AREA (n = 33); mean grain size - .1 mm									
59-43	76	7	7	9	–	T	–	–	T
ONEMILE CREEK AREA (n = 11); mean grain size = .2 mm									
K289	53	45	–	1	T	1	T	–	–
K290	41	1.5	7	50	–	T	–	–	–
NORTH FORK ROCK CREEK AREA (n = 31); mean grain size = .1 mm									
60-9	65	8	25	2	–	–	–	–	–
MEDICINE BOW RIVER AREA (n = 50); mean grain size = .2mm									
60-32	62	–	–	36	–	2	–	–	–
SR78 40	40	28	17	9	3	2	1		
SR78-41	48	19	19	9	3	2	T	–	T
MB13-146.2	49	26	11	5	2	2	T	5	–
MB13-380	46	30	21	–	3	1	–	–	–
MEAN	53	18.3	11.9	13.4	1.2	1.1	T	.6	–

One possibility is that the Conical Peak Quartzite was derived from reworking of Bow River sediments with the addition of detrital plagioclase from the underlying Colberg Metavolcanic Rocks. If so, there may be a disconformity between the Colberg Metavolcanic Rocks and the Conical Peak Quartzite, a relationship which is also suggested by drastic thickness changes in the Colberg Metavolcanic Rocks in the northern Medicine Bow Mountains (Plate 1). The maximum exposed stratigraphic thickness of the Conical Peak Quartzite is about 800 m.

As shown in Figure 2.7, crossbedding in the Conical Peak Quartzite is mainly large-scale and planar (amplitude about 1 m; mean inclination 22.6°), but trough crossbeds are also present near Onemile Creek and the Medicine Bow River (Plate 2). The paleocurrent distribution for the entire unit (Plate 2) shows a bimodal, bipolar distribution with currents directed northeast and southwest. These modes are interpreted to represent ebb and flood currents in a shallow marine depositional environment. The unit has many similarities to the underlying Bow River Quartzite and we interpret it to be marine for many of the same reasons: fine grain sizes, large-scale planar crossbeds, and bimodal paleocurrents.

DEEP LAKE GROUP

The Deep Lake Group was defined by Karlstrom and Houston (1979a, 1979b) to include six formations (Table 2.1). However we now consider the uppermost unit, the Rock Knoll Formation, to be part of the overlying Libby Creek Group because it is in depositional contact with the overlying Headquarters Fromation and is now interpreted to be entirely

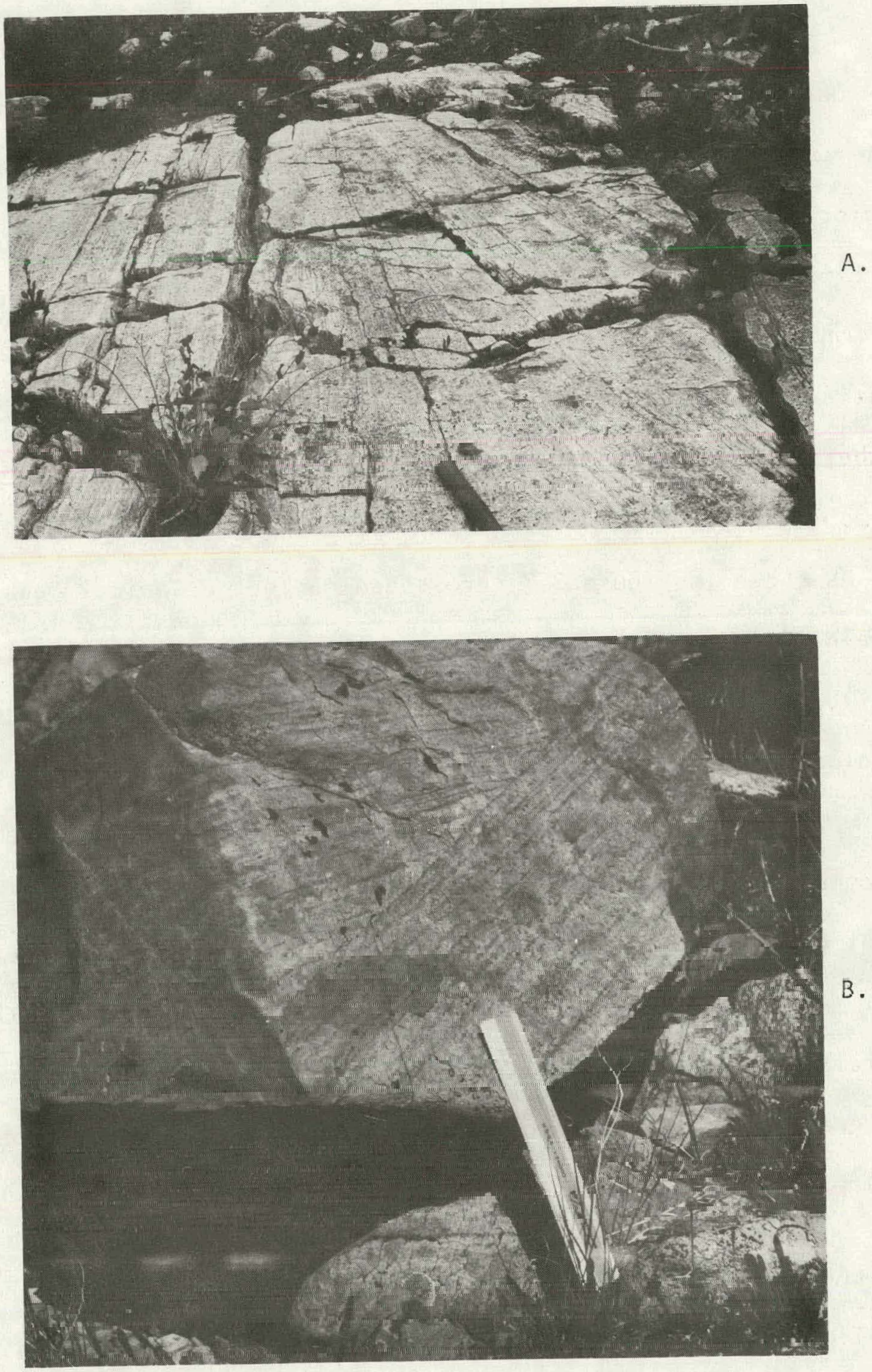


Figure 2.7. Photographs of the Conical Peak Quartzite: A. 1.5 meter thick planar crossbed set is truncated by phyllite layer above (right) and is tangential to phyllite layer below (left); hammer handle is subparallel to foresets; from Sec. 11, T.17 N., R.80 W.; B. large-scale trough crossbedding near Conical Peak, Sec. 4, T.18 N., R.79 W.

in fault contact with the underlying Vagner Formation (Plate 1) on the basis of: breccias in Trail Creek (Sec. 26, T. 17 N., R. 79 W.) and South Brush Creek (Secs. 15, 22, T. 16 N., R. Bow; change in strike of bedding of 45° south of Vagner Lake (Sec. 1, T. 16 N., R. 79 W.); change in dip of bedding by 70% northwest of Rock Creek Knoll (Secs. 27, 34, T. 17 N., R. 79 W.); and a topographic scarp between Reservoir and Vagner Lakes (Lanthier, 1978, p. 26). The basal unit of the Deep Lake Group, the Magnolia Formation, contains the major radioactive conglomerates in the Medicine Bow Mountains. Its stratigraphic, sedimentary, and lithologic characteristics, and its uranium and thorium potential are discussed in detail by Karlstrom and Houston (1981). Other stratigraphic units are discussed in more datail in earlier papers (Karlstrom, 1977; Karlstrom and Houston, 1979a, 1979b). This section reviews briefly the main sedimentary features of each unit and presents our interpretations of the depositional setting of the formations of the Deep Lake Group. The petrography of each of the units is shown in Figure 2.8.

Magnolia Formation

The Magnolia Formation is composed of two members: a basal radioactive Conglomerate Member containing micaceous quartz-pebble conglomerates and arkosic matrix paraconglomerates and a Quartzite Member containing more mature subarkosic and arkosic quartzites and quartz-granule conglomerates (Karlstrom and Houston, 1979a, Figure 2.8, 2.9, and Table 2.9). This subdivision is not always clear cut; the basal conglomerate member is not always present and contact relationships between the two

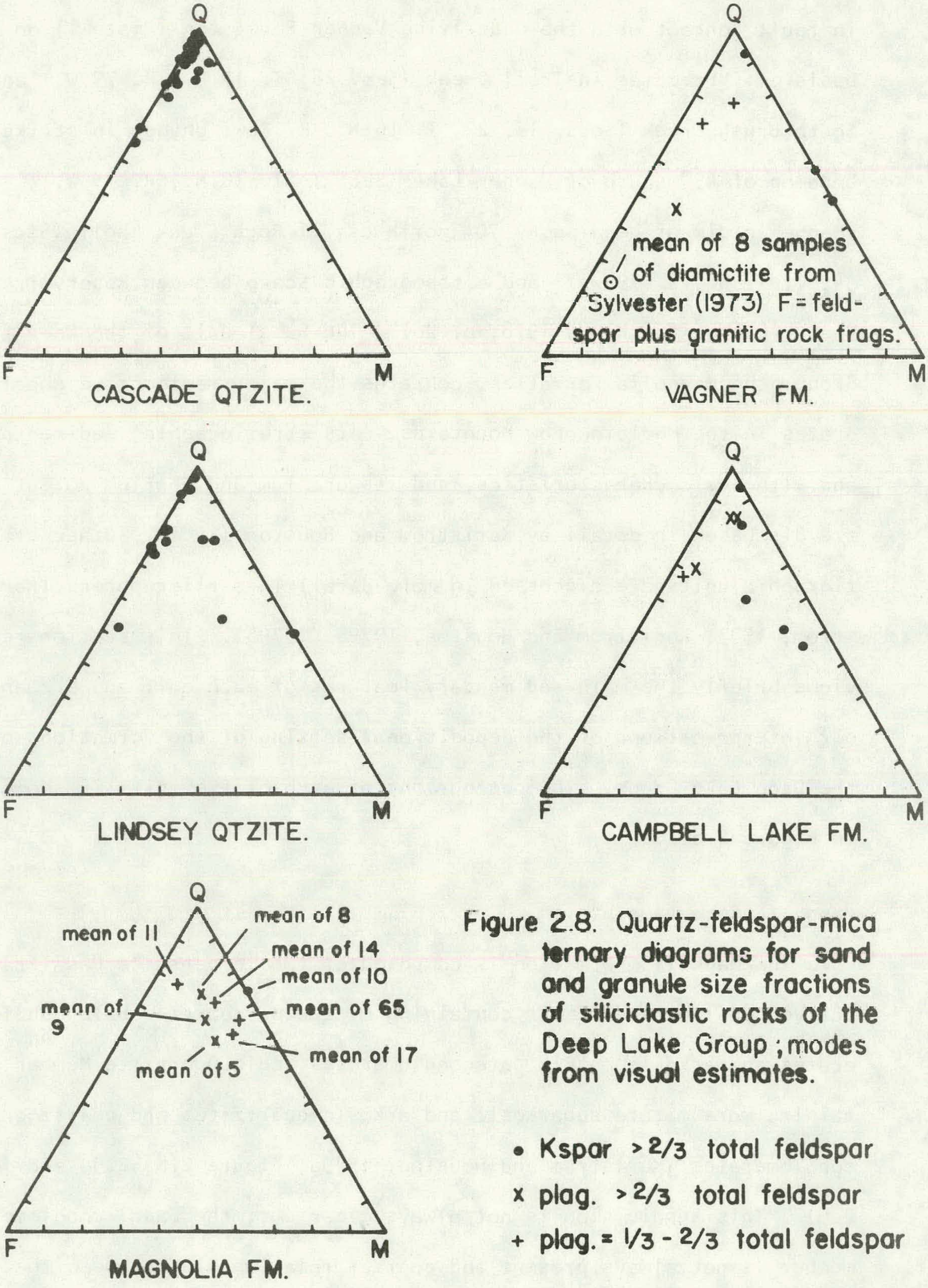


Figure 2.8. Quartz-feldspar-mica ternary diagrams for sand and granule size fractions of siliciclastic rocks of the Deep Lake Group; modes from visual estimates.

- • Kspar >2/3 total feldspar
- x plag. >2/3 total feldspar
- + plag. = 1/3 - 2/3 total feldspar

A.

B.

Figure 2.9. Photographs of the Magnolia Formation, Conglomerate Member, near Threemile Creek, NW1/4, Sec. 19, T.18 N., R.78 W.; A. layer of strongly radioactive quartz-pebble conglomerate; pen points to a square pyrite pseudomorph and is parallel to bedding and S_1 foliation; maximum pebble is 4 x 2.5 cm; B. granite and quartz pebbles and cobbles weathering out in cross-sectional view of arkosic paraconglomerate underlying quartz-pebble conglomerate shown in A; clasts are flattened in foliation.

members appear to be gradational. However, the nomenclature is retained here because it points out a major, and mappable, change in rock-type within the Magnolia Formation.

TABLE 2.9 SUMMARY OF THE PETROGRAPHY OF THE MAGNOLIA FORMATION ARRANGED APPROXIMATELY NORTHEAST (TOP) TO SOUTHWEST (BOTTOM)

MAGNOLIA FORMATION CONGLOMERATE MEMBER

	Qtz.	Qtzte.	K-spar	Plag.	Granite	Musc.	Chlor.	Biot.	Pyrite	Hem	Heavies
ONEMILE CREEK AREA											
Unit 1, mean and range	52	1.5	1.5	1.5	3.8	24	7	10	–	5	
of 9 samples	14-70	0-10	0-14	0-10	0-25	5-47	0-7	0-70	–	0-6.4	
Unit 2, mean and range	51	2	8	7	5	25	.1	.7	.6	.7	
of 7 samples	30-62	0-10	3-25	0-20	0-20	5-45	0-.5	0-3	0-4	0-3	
Unit 3, mean and range	57	included	11	6	–	21	5	–	–	–	
of 2 samples	46-67	w/qtz.	6-16	5-6	0-.75	11-31	1-8	–	–	–	
Unit 4, mean and range	71	4	2.7	.3	–	17	–	–	5	–	
of 4 samples	63-82	1-8	2-4	0-1	–	10-25	–	–	0-20	–	
Unit 5, mean and range	58	1.4	1.3	1	2.8	26	1.5	.4	7.6	–	Apatite
of 43 samples	27-80	0-16	0-15	0-6	0-32	4-50	0-15	0-14	0-30	–	Zircon
Grand Mean of 65 samples	57.9	1.6	1.7	1.8	2.9	24.9	1.2	.26	5.4	.1	
THREEMILE CREEK AREA (EMB-5 & 11)											
Mean and range of	48	10	3.3	2.3	3.3	23.3	2.4	4.2	.6	1.4	1.2 Garnet
17 samples	10-68	0-23	0-18	0-10	0-24	0-75	0-10	0-20	0-5	0-20	Zircon Tourmaline
ARRASTRE ANTICLINE AREA (PL-1)											
Mean and range of	65	6.5	–	.3	.1	21.4	4.1	1	1	–	.5 Zircon
10 samples	45-82	0-25	–	0-3	0-1	1-30	0-15	0-.5	0-5	–	Amphibole
BRUSH CREEK AREA (MB-9)											
Mean and range of	58	4	1.3	2.6	–	16	5.6	2.2	.6	–	Garnet
3 samples	46-62	0-20	0-5	0-5	–	0-20	0-10	0-5	0-2	–	

MAGNOLIA FORMATION QUARTZITE MEMBER

	Qtz.	Qtzte.	K-spar	Plag.	Granite	Musc.	Chlor.	Biot.	Pyrite	Hem	Heavies
THREEMILE CREEK AREA (EMB-11)											
Mean and range of	51	4	4	8	3	20.5	2.1	2	5.4	–	Zircon
5 samples	20-76	0-20	0-10	0-20	0-10	5-50	0-5	0-5	0-15	–	Apatite
MB-4 AREA											
Mean and range of	70	4	9.31	4.3	.6	8.	.6	1.2	–	.4	Zircon
8 samples	37-90	0-20	0-20	0-18	0-5	2-15	0-7	0-5	–	0-2	Amphibole Apatite
NORTH FORK ROCK CREEK CRATER LAKE AREA (MB-11)											
Mean and range of	69	4.3	1.6	4.5	1	12.3	1.8	1.5	–	.9	Amphibole
10 samples	44-93	0-15	0-10	0-10	0-5	3-25	0-5	0-7	–	0-2	Calcite Apatite Zircon
ARRASTRE ANTICLINE AREA (PL-1 & GH-1)											
Mean and range of	66	5	3	8	.3	11.3	3.3	2	.3	.8	Tourmaline
8 samples	54-82	0-20	0-7	0-30	0-2	0-25	0-10	0-5	0-2	0-2	Zircon Calcite

Modal percentages represent entirety of poorly sorted rocks, granule and subgranule matrix for Bimodal conglomerates.

The radioactive Conglomerate Member crops out in several localities as lenticular zones of polymictic paraconglomerate interbedded with quartz-pebble conglomerates and coarse-grained arkosic quartzites. The paraconglomerates are micaceous arkoses and subarkoses with poorly sorted clasts (up to tens of centimeters in diameter) of quartzite, schist, mafic volcanic rocks and, locally, granite. These paraconglomerates are generally only mildly radioactive (up to 50 ppm U, 200 ppm Th, but usually less). In the Onemile Creek area, at the northern limit of Magnolia outcrop, the Conglomerate Member is much more radioactive (up to 176 ppm U, 915 ppm Th in outcrop) and is dominated by muscovitic quartz-pebble conglomerates and coarse-grained muscovitic arkoses (Table 2.9). In this area, the radioactive Conglomerate Member is about 400 m thick and conglomerate layers range in thickness from beds one pebble thick to composite comglomerate zones up to 20 meters thick.

The Quartzite Member is present throughout the Medicine Bow Mountains, averaging about 400 m thick. This unit consists of trough cross-bedded, coarse-grained to granular quartzites which vary in composition from micaceous subarkose to micaceous arkose (Table 2.9). The Quartzite Member is generally less micaceous, less arkosic, finer-grained, and less radioactive (averaging less than 10 ppm U, Th) than the Conglomerate Member.

We interpret the Magnolia Formation to be a fluvial succession for a variety of reasons. The unit overlies rocks of the Phantom Lake Metamorphic Suite with angular unconformity. This unconformity is exposed in the cirque walls of Crater Lake (Sec. 35, T. 18 N., R. 79 W.), where north-striking outcrops of the Quartzite Member overlie tightly folded, east-west striking rocks of the Colberg Metavolcanic Rocks (Plate 1). It is also evidenced by the fact that the Magnolia Formation overlies a variety of units of the Phantom Lake Suite in the central Medicine Bow Mountains and Archean granites and the Stage Crossing Gneiss in the northern Medicine Bow Mountains (Plate 1). This and the presence of pebbles of Archean granite in the northern outcrops suggests a major period of erosion prior to Magnolia deposition and implies that subaerial conditions dominated, at least in the early stages of Magnolia deposition. Lithologic characteristics and the lenticular distribution of the radioactive Conglomerate Member are also consistent with interpretations of a fluvial origin. The unit is poorly sorted, relatively immature, and it contains radioactive minerals which are believed to be heavy mineral placers (Houston and Karlstrom, 1980). The lenticular distribution, combined with the coarse grain sizes, poor sorting, and immature nature of the sediments suggest proximal, high energy fluvial deposition (Rust, 1979). We envision a paleogeography characterized by fairly high relief, fault bounded highlands, and deposition of radioactive conglomerates in subaerial debris flows, alluvial fans, and high energy braided streams.

The Quartzite Member is also interpreted to be fluvial in origin because of relatively poor sorting, abundant trough crossbedding, and gradational (fining-upwards) relationships to the underlying conglomerates. We interpret the quartzites to represent deposition in a well-developed and laterally extensive braided river system because of the lateral persistence of the unit (an outcrop length of about 35 km) and the change to less micaceous and arkosic quartzites from northeast to southwest, down the paleoslope (Plate 2 and Table 2.9).

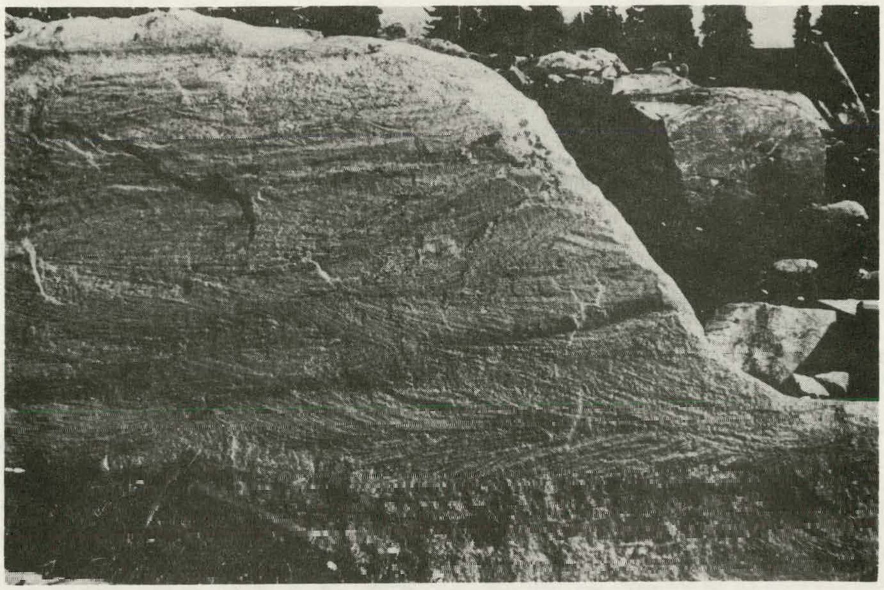
Paleocurrent measurements (Plate 2), although almost entirely from the Quartzite Member, substantiate the fluvial interpretation fairly well. The distribution for the entire Magnolia Formation shows a somewhat dispersed polymodal pattern dominated by southwesterly directed currents. This type of dispersed polymodal pattern is often cited as evidence for intertidal deposition (Hereford, 1977; Klein, 1970) but may also represent distributions in braided stream environments (Ore, 1964; Williams, 1971) or fluvial deposition combining alluvial fan and braided river deposition. For individual areas (Plate 2), some paleocurrent roses show unimodal patterns, others show distinctly polymodal patterns. If the polymodal patterns are real (i.e. if they are not a relict of measurement problems associated with distinguishing trough from planar crossbeds in small, discontinuous outcrops), they may reflect a southwesterly directed major river system, bounded on the northwest by areas of high relief, in which alluvial fans were developed which transported sediment in southeasterly directions, into the trunk river system. By this explanation, the conglomerates of the lower Magnolia Formation represent the basal portions of these alluvial fan deposits. The wide distribution of conglomerate lenses and consistent coarse grain sizes of these conglomerates in a down-current direction (southwest) in the Medicine Bow Mountains is explained in terms of multiple southeasterly directed fan systems built on fault-bounded highlands to the northwest. This interpretation remains tentative because crossbedding is rare in the conglomerates themselves and their paleocurrent distributions are not known.

Lindsey Quartzite

The Lindsey Quartzite includes trough crossbedded quartzarenite and subarkose (Figures 2.8, 2.10, and Table 2.10), which gradationally overlie the Quartzite Member of the Magnolia Formation. Lindsey quartzites are generally medium-grained but also contain scattered pebbles (up to 1 cm in diameter) along planar foreset beds or in small scours, and thin phyllitic layers and partings. The thickness of the Lindsey Quartzite is about 410 m in the central Medicine Bow Mountains and the unit either pinches out or is unconformably onlapped by the Cascade Quartzite to the north (Plate 1). The Lindsey Quartzite is locally anomalously radioactive, especially near its northern outcrop limit.

TABLE 2.10. PETROGRAPHY OF THE LINDSEY QUARTZITE; MODES FROM VISUAL ESTIMATES; K-SPAR COLUMN INCLUDES MUSC. AFTER FELDSPAR; ARRANGED NORTHEAST (TOP) TO SOUTHWEST (BOTTOM).

Sample No.	Qtz.	K-spar	Plag.	Musc.	Biot.	Chlor.	Opaq.	Zirc.	Apat.
MB4-25	93	5	—	2	—	—	—	—	—
MB4-189	77	14	6	2	1	—	T	T	—
60-66	76	7	5	12	—	—	—	—	—
SR78-34	80	18	—	2	—	—	T	—	—
SR78-37	47	—	—	51	T	T	1	1	T
MB6-143	77	10	—	12	1	—	T	—	—
PC-x	96	3		1	1	T	T	T	T
59-1	77	7	—	15	—	—	1	—	—
59-15	93	5	1	1	—	—	T	—	—
TS-22	91	4	5	T	—	—	T	—	—
TS-14A	54	—	17	29	T	—	T	T	—
TS-14B	51	30	16			4	T	—	—
TS-14C	73	20	5	—	—	2	T	—	—
TS-14D	75	22	3	—	—	—	T	—	—
MEAN	76	10	4	9	T	T	T	T	T



A.

B.

Figure 2.10. Photographs of trough crossbedding in the Lindsey Quartzite; A. Sec. 33, T.17 N., R.79 W.; B. Sec. 30, T.18 N., R.78 W.

Trough crossbedding is well preserved in the Lindsey Quartzite (Figure 2.10) and ripple marks are present occasionally. The paleocurrent distribution shown in Plate 2, is strongly unimodal with currents directed southwest and a mean direction (229°) very similar to the Magnolia Formation. This type of unimodal pattern is often cited as evidence for fluvial deposition (Allen, 1967; Potter and Pettijohn, 1977). We interpret the Lindsey Quartzite to be a fluvial unit on this basis, as well as: its gradational relationship to the Magnolia Formation; the abundance of trough crossbeds; and scattered pebbles and phyllite partings which may represent lag gravels and thin overbank sediments respectively.

Campbell Lake Formation

The Campbell Lake Formation is a thin (up to 65 m), discontinuous paraconglomerate -- quartz phyllite sequence which serves as a useful stratigraphic marker in the central Medicine Bow Mountains. The paraconglomerate contains poorly sorted subangular clasts (up to 76 cm in diameter) of granite, quartzite, and phyllite in a poorly sorted micaceous arkose to subarkose matrix (Figure 2.8; Table 2.11). The unit appears to be a debris flow of some type and has been interpreted to be glacial in origin (Karlstrom and Houston, 1979a, 1979b), although we have no definite evidence for glacial deposition and the unit might also represent alluvial fan debris flows (Bull, 1972).

Cascade Quartzite

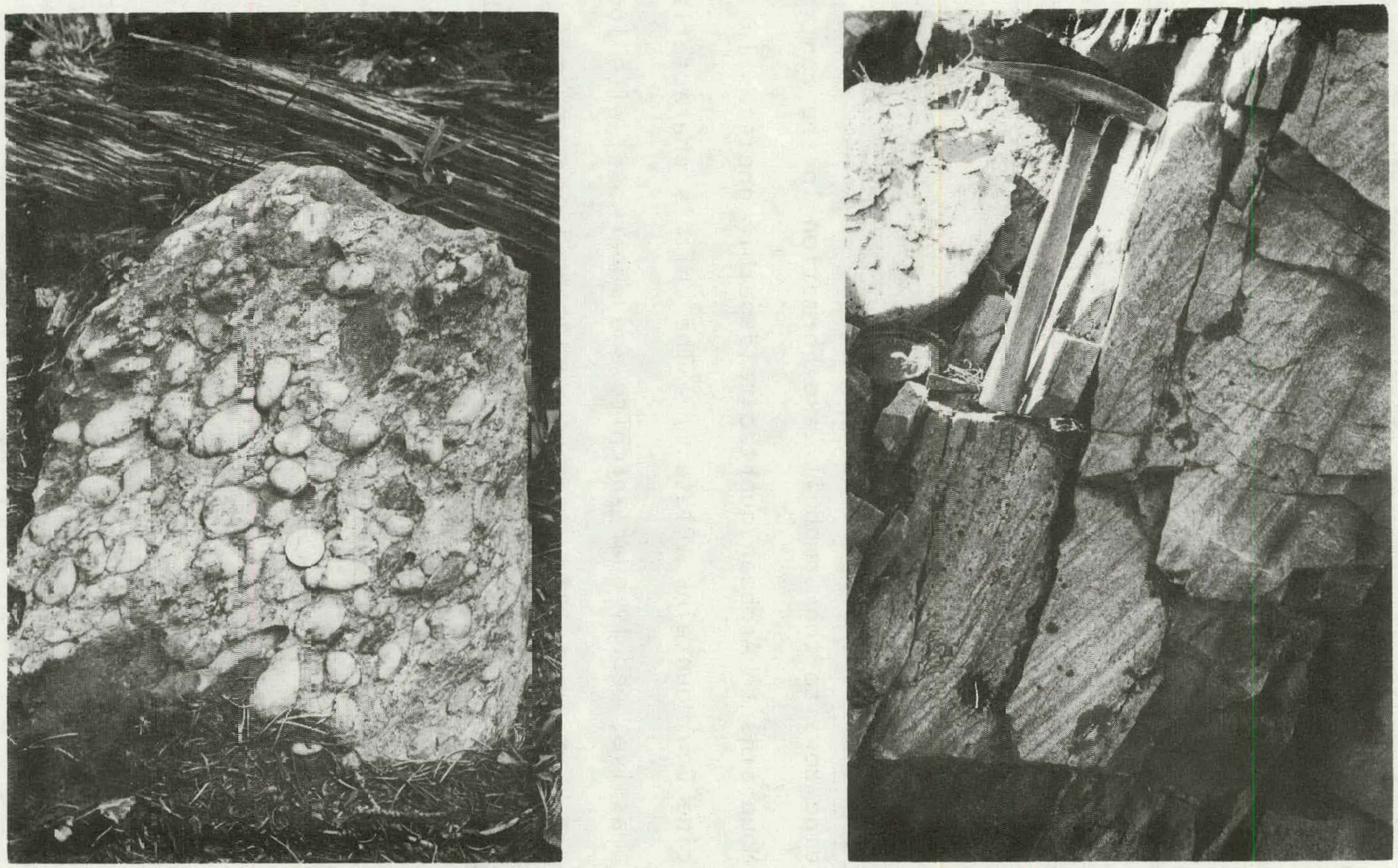
The Cascade Quartzite is the thickest (up to 850 m) and most laterally extensive unit in the Deep Lake Group. It unconformably overlies units ranging in age from Archean granite and Stage Crossing Gneiss in the

TABLE 2.11. PETROGRAPHY OF THE CAMPBELL LAKE FORMATION; MODES FROM VISUAL ESTIMATES: K-SPAR COLUMN INCLUDES MUSC. AFTER FELDSPAR; ARRANGED NORTHEAST (TOP) TO SOUTHWEST (BOTTOM).

Sample No.	Qtz.	Qtzite.	K-spar.	Plag.	Musc.	Biot.	Chlor.	Opaq.	Zirc.	Epid.	Apat.	Labile R.F.
K78-22	84	–	–	8	9	1	–	T	–	–	–	–
TS-371	94	–	5	–	1	–	–	T	–	–	–	–
TS-236	56	–	16	–	19	–	5	2	–	–	–	–
C-520	46	–	7	–	38	–	7	2	–	–	–	–
61-35	51	–	10	–	36	3	–	1	–	–	–	–
TS-128	83	–	6	T	10	1	T	T	T	T	–	–
PC-1A	83	–	–	7	3	3	T	2	T	–	T	–
PC-1B	62	5	13	14	2	2	T	T	T	–	–	2
PC-2B	69	–	6	19	2	4	T	T	–	–	–	–
MEAN	69.8	.6	7	5.3	13.3	1.6	1.3	.8	T	T	T	.2

northern Medicine Bows to the Campbell Lake Formation in the central Medicine Bow Mountains to Archean quartzo-feldspathic gneisses in the southern Medicine Bow Mountains (Plate 1). The unit is characterized by clean, white, massive, pebbly quartzarenite and subarkose, with local arkose (Figure 2.8; Table 2.12). Quartz and black chert pebbles occur in distinct layers, commonly 5-10 cm thick (Figure 2.11). Maximum pebble sizes, shown in Plate 2, for the unit appear to be largest (∿ 3 cm) in the central part of the outcrop belt and smaller (∿ 2 cm) in the northeast and southwest.

Paleocurrent measurements from abundant planar and trough cross-bedding in the Cascade Quartzite show a unimodal distribution about a west-southwest directed mean paleocurrent (248°). The dispersion of paleocurrent vectors is appreciably smaller (L is larger and variance is smaller) than for both the Lindsey Quartzite and Magnolia Formation. This type of small variance unimodal paleocurrent pattern is often considered to be one characteristic of fluvial paleocurrent patterns (Potter and Pettijohn, 1977).



A.

B.

Figure 2.11. Photographs of the Cascade Quartzite: A. pebble layer on bedding surface in Sec. 30, T.18 N., R.78 W.; largest pebble is 6 x 3.5 cm; gray pebbles in lower right are chert; B. 15 cm thick planar crossbed set in Sec. 32, T.18 N., R.78 W.

TABLE 2.12 PETROGRAPHY OF THE CASCADE QUARTZITE; MODES FROM VISUAL ESTIMATES; K-SPAR COLUMN INCLUDES MUSC. AFTER FELDSPAR. ARRANGED NORTHEAST (TOP) TO SOUTHWEST (BOTTOM).

Sample No.	Qtz.	K-Spar	Plag.	Musc.	Biot.	Chlor.	Rock Frags.	Opaq.	Zirc.	Apat.	Epid.	Rutile
Cooper Hill	83	11	2	2	–	–	–	2	–	–	–	–
60-61	90	10	–	–	–	–	–	T	–	–	–	–
60-6	83	17	–	–	–	–	–	T	–	–	–	–
TS-218A	93	5	1	–	T	T	–	1	–	–	–	–
TS-340	97	2.5	T	–	–	–	–	.5	–	–	–	–
60-8	69	29	–	–	1	–	–	1	–	–	–	–
TS-6A	91	8	–	1	–	–	–	T	–	–	–	–
TS-6C	93	4	–	–	–	–	–	3	–	–	–	–
TS-6E	97	2.5	–	2	–	–	–	T	–	–	–	–
TS-6F	96	3	–	–	–	1	–	T	T	–	–	–
TS-6G	89	3	–	3	3	–	–	2	T	–	T	T
TS-10A	86	7	1	–	–	6	–	T	–	–	–	T
PC-2A	96	3	–	1	–	–	–	–	–	–	–	–
PC-3	90	9	–	1	–	–	–	T	–	–	–	–
PC-4	84	7	2	–	–	7	T	T	–	–	–	–
PC-Y	65	28	6	1	–	–	–	T	T	T	–	–
TS-144	79	19	–	2	–	–	T	T	T	–	–	–
TS-146	89	9	1	1	–	–	–	T	T	–	–	–
59-13	83	1.5	15	–	–	–	–	.5	–	–	–	–
MEAN	87	9.4	1.5	.7	.2	.7	T	.5	T	T	T	T

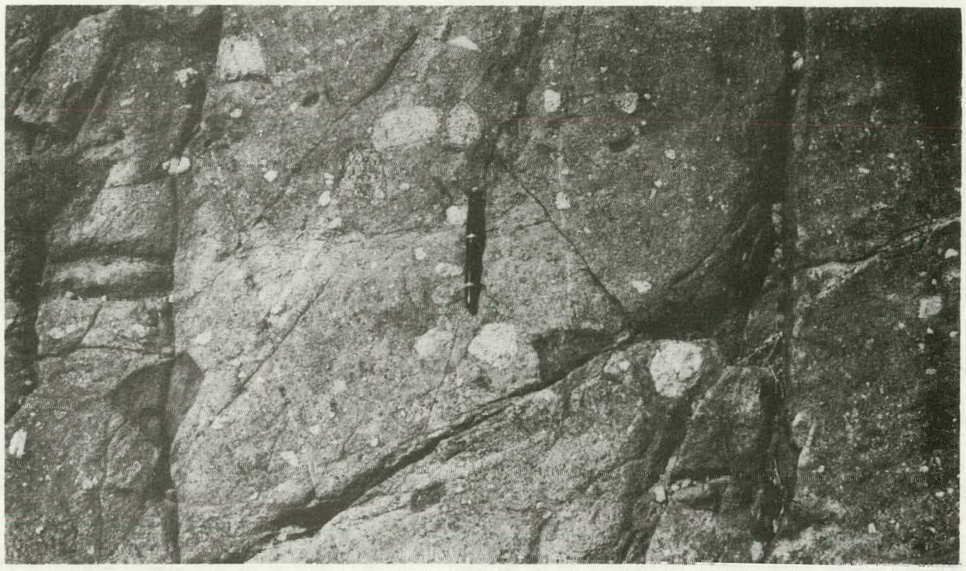
We have interpreted the Cascade Quartzite in previous papers (Karlstrom and Houston, 1979a, 1979b) to be a fluvial unit on the basis of: the low variance unimodal paleocurrent distribution, the presence of both planar and trough crossbedding (interpreted to represent bars and migrating dunes in a river system), layers of well rounded, well sorted pebbles interpreted to be lag gravels, and unconformities at the base and top of the Cascade Quartzite which suggest subaerial conditions before and after Cascade deposition. However, none of these features is restricted to fluvial environments and the lateral continuity of the unit (a strike length of over 50 km in the Medicine Bow Mountains and about 30 km in the Sierra Madre to the west), the mature composition, the relative consistency of maximum pebble sizes in a down-current direction, and the presence of one bimodal (north-south) paleocurrent distribution in the northeastern Medicine Bow Mountains (Plate 2) might also be cited as evidence supporting marine or deltaic deposition. The unimodal paleocurrent distribution for the entire unit and the mature composition are both similar to the Medicine Peak Quartzite (Plate 2) which we consider to be a shallow marine deposit (discussed later). At present, our understanding of the Cascade Quartzite permits either a fluvial or deltaic interpretation or both types of sediments may be present.

Vagner Formation

The Vagner Formation is a paraconglomerate-marble-phyllite-quartzite succession (up to 800 m thick) that unconformably overlies the Cascade Quartzite. This unit was originally included as part of the Headquarters Schist (Table 2.1) because of similarities between paraconglomerates in the Vagner and Headquarters Formations and the proximity of these units to one another in the central Medicine Bow Mountains. However, detailed mapping has shown that the paraconglomerate-marble succession of the Vagner Formation is present in the northeastern Medicine Bow Mountains (Plate 1) and is definitely part of the Deep Lake Group. The proximity of Vagner and Headquarters paraconglomerates is now interpreted to be related to movement on the Reservoir Lake Fault (Plates 1 and 3). The Vagner Formation is an important unit in the Medicine Bow Mountains from several perspectives. Stratigraphically, it represents the early stages of a major transition in sedimentation: from dominantly epicontinental deposition in the Deep Lake Group to dominantly marine deposition of the miogeoclinal Libby Creek Group. Structurally, the marbles and the phyllites in the unit apparently were some of the most ductile rocks in the sequence, localizing strain during major Proterozoic thrust faulting and during Laramide uplift so that the unit is intimately associated with major faults. In the central Medicine Bow Mountains, the Vagner Formation forms the footwall of the Reservoir Lake Fault over most of its length; in the northeastern Medicine Bow Mountains, the unit is in the hanging wall of the Arlington thrust fault of Laramide age.

The basal unit of the Vagner Formation is a polymictic paraconglomerate (or diamictite) with angular and subangular clasts of granite, quartzite, and mafic schist (Figure 2.12A) in a subarkosic matrix (Figure 2.8; Table 2.13). The unit is thin (average thickness about 300 m), but laterally persistent and is interpreted to be a glaciomarine conglomerate on the basis of the presence of dropstone clasts, poor sorting, subangular clasts, faint stratification in some conglomerates, and the geochemistry of the sand-size matrix of the conglomerates (Table 2.14) which is similar to the Gowganda tillites of the Huronian Supergroup and to other Early Proterozoic glacial(?) deposits in North America (Sylvester, 1973; Young, 1970; 1973). Marble and phyllite of the Vagner Formation (Figure 2.12B) provide support for a marine depositional setting.

Limited paleocurrent data from the Vagner Formation (Plate 2) show a change to more westerly directed transport (mean = 256°). One possible explanation for this change involves a highland to the east which could supply detritus. This explanation fits well with our concept of sedimentation in a rift basin during the Early Proterozoic (discussed in detail later).



A.

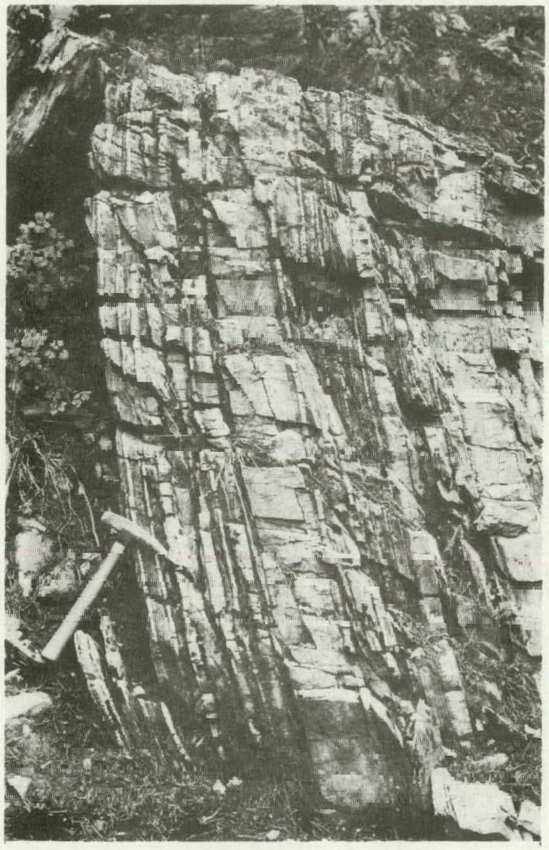


Figure 2.12. Photographs of the Vagner Formation; A. poorly sorted, sub-angular to sub-rounded clasts of white granite, quartzite, and mafic schist in medium-grained argillaceous subarkose matrix of basal diamictite near Reservoir Lake, Sec. 5, T.16 N., R.79 W ; B. marble overlying diamictite near Rock Creek, Sec. 6, T.17 N., R.78 W.

B.

TABLE 2.13 PETROGRAPHY OF THE VAGNER FORMATION; MODES MAINLY FROM VISUAL ESTIMATES; K-SPAR COLUMN INCLUDES MUSC. AFTER FELDSPAR; ARRANGED NORTHEAST (TOP) TO SOUTHWEST (BOTTOM) WITHIN EACH ROCK-TYPE.

Sample No.	Qtz.	K-Spar	Plag.	Calc.	Epid.	Musc.	Biot.	Opaq.	Amph.	Span.	Chl.
SILICICLASTIC ROCKS – SAND AND SILT – SIZE FRACTION											
K662	35.0	–	8.9	–	5.7	–	40.3	T	10.0	T	–
K795	50.2	–	5.7	–	–	5.7	34.4	1.0	–	–	2.3
K78-10	30	–	2	–	5	–	–	2	61	–	–
K78-36	47	–	–	–	–	52	–	1	–	–	–
K78-62	92	–	–	–	–	8	–	–	–	–	–
60-63	48	10	22	9	7	–	4	–	–	–	–
60-65	56	–	–	–	–	40	3	1	–	–	–
TS-367A	70	7	16	–	–	2	5	T	–	–	–
TS-367C	44	–	43	2	–	1	4	6	–	–	–
CLASTIC CARBONATES											
K78-29	36	–	3	43	13	–	5	T	–	–	–
K78-96	62	–	–	23	15	–	–	T	–	–	–
TS-349	46	20% feldspar		9	14	6	6	1	–	–	–
TS-367B	20	–	–	80	–	–	–	–	–	–	–
TS-393	30	25% feldspar		8	37	–	–	–	–	–	–

TABLE 2.14. CHEMICAL COMPOSITIONS OF SAND-SIZE MATRIX FROM GLACIAL (?) PARACONGLOMERATES; DATA FROM SYLVESTER (1973) AND YOUNG (1973).

	1.	2.	3.	4.
SiO_2	74.38	65.15	70.69	65.32
Al_2O_3	9.42	13.54	11.79	14.73
K_2O	2.74	1.90	1.39	1.65
CaO	.48	.80	.34	1.48
Total Fe	4.69	5.55	4.87	5.86
Na_2O	–	–	–	4.50
MgO	–	–	–	3.57

1. Vagner Formation – basal paraconglomerate; mean of 8 samples taken between Dipper and Reservoir Lakes from Sylvester's "basal diamictite unit."
2. Headquarters Formation – basal paraconglomerate; mean of 5 samples taken between Dipper and South Twin Lakes.
3. Headquarters Formation – upper paraconglomerate; mean of 4 samples taken west of Twin Lakes.
4. Gowganda Formation, Huronian Supergroup; mean of 16 samples from north shore of Lake Huron (Young, 1973)

LOWER LIBBY CREEK GROUP

The name Lower Libby Creek Group is used here to refer to the dominantly marine siliciclastic units south of the Reservoir Lake fault and stratigraphically below the Nash Fork Formation (Table 2.1). This new subdivision, Upper versus Lower Libby Creek Groups, is introduced in this report for several reasons. First, the lithologies in these two successions are quite different. Whereas the lower succession is predominantly siliciclastic, the Upper Libby Creek Group contains dolomite, volcanogenic rocks, and black slate. This major change in lithology indicates very different conditions for sedimentation for the two successions. Second, new map interpretations (Plate 1) place a major fault, the Lewis Lake Fault, between the Sugarloaf Quartzite and the Nash Fork Formation on the basis of the disappearance of the Sugarloaf Quartzite in Secs. 9, 10, and 17, T. 16 N., R. 79 W.; the abrupt thinning of the Medicine Peak Quartzite in Sec. 2, T. 16 N., R. 79 W. and various breccia zones along the contact (Lanthier, 1978, p. 26). As discussed later, we interpret the fault to be a rotated thrust fault which implies that the upper Libby Creek Group is allochthonous or para-autochthonous with respect to the Lower Libby Creek Group and may have been deposited in very different tectonic and sedimentary environments. Third, although the exact age relationship between the two successions is not yet clear, there are several indications that the Upper Libby Creek Group may be appreciably younger than the Lower Libby Creek Group. The Gaps Trondjhemite intrudes the lower succession but not the upper (Plate 1) and regional correlations to other metasedimentary successions in North America suggest that the Upper Libby Creek Group may correlate with the 1900-2100 m.y. old

Marquette Range Supergroup of Minnesota whereas the Lower Libby Creek Group may correlate with the 2150(+) m.y. old Huronian Supergroup of Ontario (Houston and Karlstrom, 1980). This is discussed later.

We consider the Lower Libby Creek Group to represent miogeoclinal deposition on the stable south margin of the Wyoming Archean craton. The two lowest units of the Lower Libby Creek Group, as well as the uppermost unit of the Deep Lake Group, are interpreted to be glaciomarine sediments which record a change from continental and epicontinental sedimentation in the Deep Lake Group to miogeoclinal sedimentation in the Libby Creek Group.

Rock Knoll Formation

The Rock Knoll Formation is a quartzite unit which crops out in two places in the hanging wall of the Reservoir Lake Fault: near Rock Creek Knoll in the east and near Twin Lakes to the west (Plate 1). Its maximum exposed stratigraphic thickness is 380 m but the base of the unit in both outcrops is truncated against the Reservoir Lake Fault.

The predominant lithology is a gray, medium-grained, plagioclase arkose (Figure 2.13 and Table 2.15). The unit also contains phyllitic layers and partings up to 30 cm and conglomerates up to 1 m thick. The conglomerates contain quartz, quartzite, and granite clasts. The lithology of Rock Knoll is very different from underlying Deep Lake Group quartzites in that plagioclase is overwhelmingly the dominant feldspar. (Compare Figures 2.8 and 2.13). Plagioclase is also the dominant feldspar in parts of the Vagner Formation and throughout the Lower Libby Creek Group. This suggests to us that the onset of glaciomarine sedimentation in the upper Deep Lake Group and Lower Libby Creek Group was in some way related to the change in composition of the sediments.

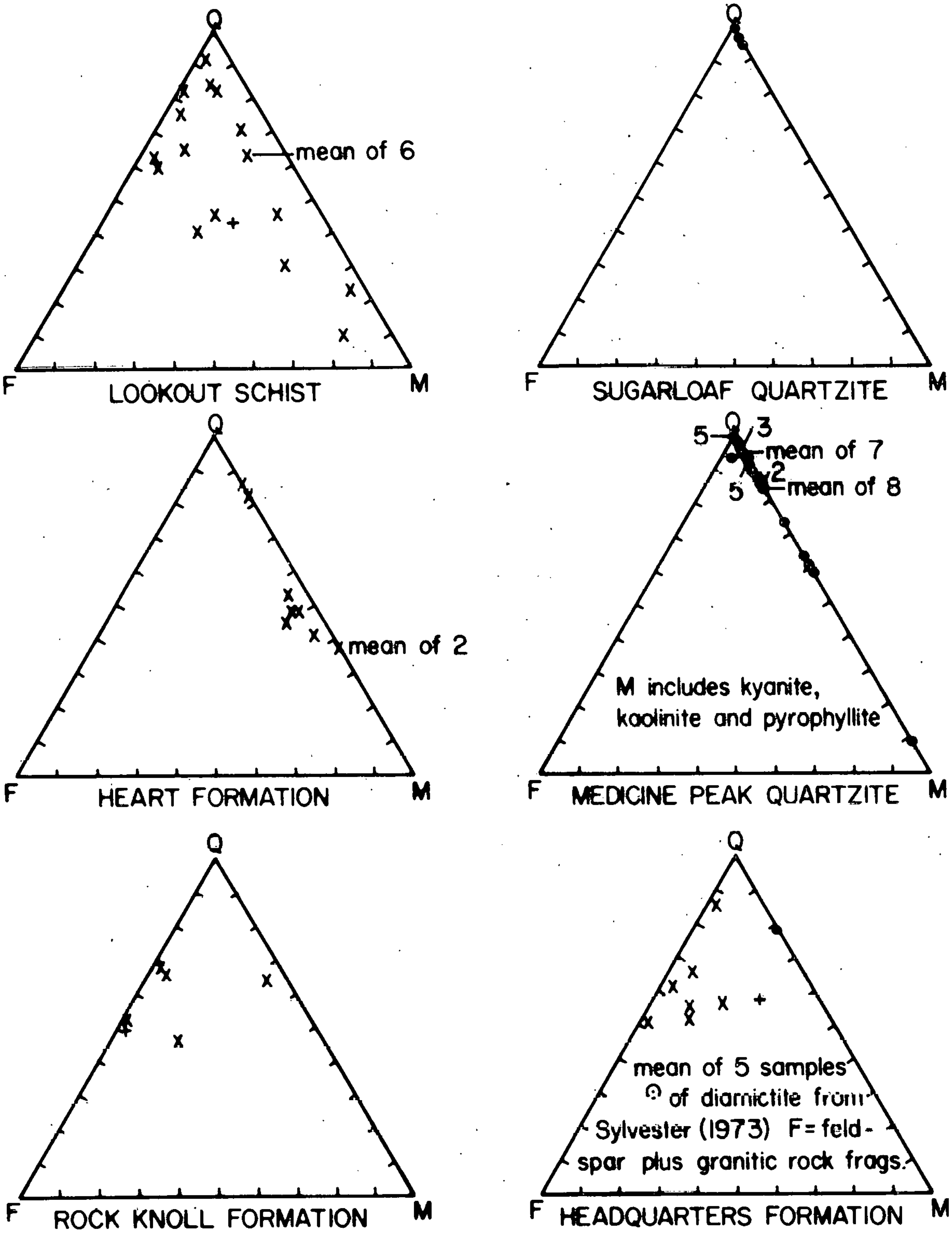


Figure 2.13. Quartz-feldspar-mica ternary diagrams for sand and granule size fractions of siliciclastic rocks of the Lower Libby Creek Group. Medicine Peak data-point counts; other modal data from visual estimates; • kspar > 2/3 total feldspar; x plag. > 1/3 total feldspar; + plag. = 1/3 - 2/3 total feldspar.

TABLE 2.15. PETROGRAPHY OF THE ROCK KNOLL FORMATION MODES FROM VISUAL ESTIMATES: ARRANGED NORTHEAST (TOP) TO SOUTHWEST (BOTTOM).

Sample No.	Qtz.	Plag.	K-spar	Musc.	Biot.	Chlor.	Opaq.	Epid.	Zirc.
TS-334	46	35	1	18	—	—	T	—	—
TS-351	68	21	8	1	1	1	—	—	—
TS-352	66	20	9	3	1	1	T	—	—
TS-28A	48	48	—	1	—	—	1	—	—
TS-28B	65	—	—	T	25	6	T	4	—
TS-1A	53	34	12	1	—	—	T	—	T
MEAN	57.7	26.3	5	4	4.5	1.3	T	.6	T

Paleocurrent data from outcrop areas of the Rock Knoll Formation indicate westerly directed paleocurrents (mean = 280°). These westerly current directions are unique to the Rock Knoll and Vagner Formations and are interpreted as evidence for an eastern source area which apparently was underlain by plagioclase-rich metavolcanic and plutonic rocks, quartzites, and some granites.

Sedimentary structures in the Rock Knoll Formation include ripple marks, planar crossbedding, and clay galls. We postulate glacial-related fluvial and deltaic deposition for the Rock Knoll Formation because of its stratigraphic position and its similarity in composition with the Vagner and Headquarters Formations. The quartzites of the Rock Knoll Formation appear to represent a period of glacial retreat between glaciomarine sedimentation episodes represented by the underlying Vagner and overlying Headquarters Formations.

Headquarters Formation

The Headquarters Formation, as redefined by Karlstrom and Houston (1979a) and Lanthier (1979) includes a lower member (350 m thick) composed of lenticular paraconglomerates, quartzites and schists and an upper member (300 m thick) containing laminated schists and phyllites. Since Blackwelder in 1926, the Headquarters Formation has been interpreted to be glacial or glaciomarine in origin (Houston and others, 1968; Sylvester, 1973; Karlstrom, 1977; Lanthier, 1979; Kurtz and Anderson, 1979). The unit has been correlated with other Early Proterozoic glacial(?) paraconglomerates in North America, most notably the Gowganda Formation of the Huronian Supergroup (Young, 1970, 1973; Houston and others, 1979; Houston and Karlstrom, 1980).

Petrographically, the unit is quite heterogeneous (Table 2.16). The lower member contains several lenses of paraconglomerate (or diamictite) composed of granite, quartzite and schist clasts (average size 4-6 cm, but ranging up to about 1 m in diameter) in a poorly sorted matrix of sand and silt size quartz, plagioclase, K-feldspar, rock fragments, and micas. At one locality, west of Twin Lakes, there are three stacked paraconglomerate units separated by quartzites. In other areas, one or multiple paraconglomerates in the lower member appear to be complexly intercalated with quartzites and schists. Quartzites are plagioclase arkoses, very similar in composition to the Rock Knoll quartzites (Figure 2.13). The upper member of the Headquarters Formation is a biotite, chlorite, quartz phyllite with laminations formed by alternating quartz-rich and mica-rich layers.

Sedimentary structures in the Headquarters Formation include small-scale planar and trough crossbedding, laminations in both paraconglomerate and phyllite units, dropstone structures in paraconglomerates, and climbing ripples. In one outcrop, Bouma A-B-C turbidite sequences were observed associated with paraconglomerates. The paraconglomerates themselves are some of the most interesting sedimentary features of the unit. These conglomerates are massive to slightly stratified, generally non-graded, poorly sorted, and contain isolated to poorly-packed subangular to rounded clasts, a few of which appear to depress underlying strata but to be covered by overlying strata (Figure 2.14) suggesting ice-rafted dropstones (Sylvester, 1973).

TABLE 2.16. PETROGRAPHY OF THE HEADQUARTERS FORMATION MODES FROM VISUAL ESTIMATES.

Sample No.	Qtz. + Qtzite R.F.	Plag.	K-Spar + Granite R.F.	Musc.	Biot.	Chlor.	Opaq.	Labile R.F.	Zirc.	Rutile	Calcite
DIAMICTITES											
TS-353	55	21	5	3	1	4	–	2	–	–	1
TS-26A	50	38	10	2	–	–	T	–	–	–	–
TS-03C	30	30	7	2	–	–	–	–	T	T	–
TS-01B	57	10	6	13	4	5	T	5	–	–	–
Mean of 5 samples from upper diamictite of Sylvester (1973)	30	10	46	4.9	–	–	–	–	–	–	–
QUARTZITES											
TS-354	50	28	9	–	10	2	1	–	–	–	–
C-538	35	5	–	45	5	3	7	–	–	–	–
C-512	65	24	5	5	–	–	1	–	–	–	–
59-59	54	33	2	1	5	4	1	–	–	–	–
TS-01C	84	12	2	2	–	T		–	–	–	–
PHYLLITES											
SR76-7	66	–	–	28	–	4	2	–	–	–	–
C-521	67	–	–	10	–	20	3	–	–	–	–
C-511	40	–	–	51	2	5	2	–	–	–	–
59-56A	84	–	–	7	1	6	2	–	–	–	–
59-56B	50	–	–	25	2	20	3	–	–	–	–
Mean of 2 samples from Sylvester (1973)	36	–	–	33	4	21	6	–	–	–	–
Mean of 2 samples from Houston and Others (1968)	31	T	–	28	4	8	8	plus garnet, sphene, tourmaline			

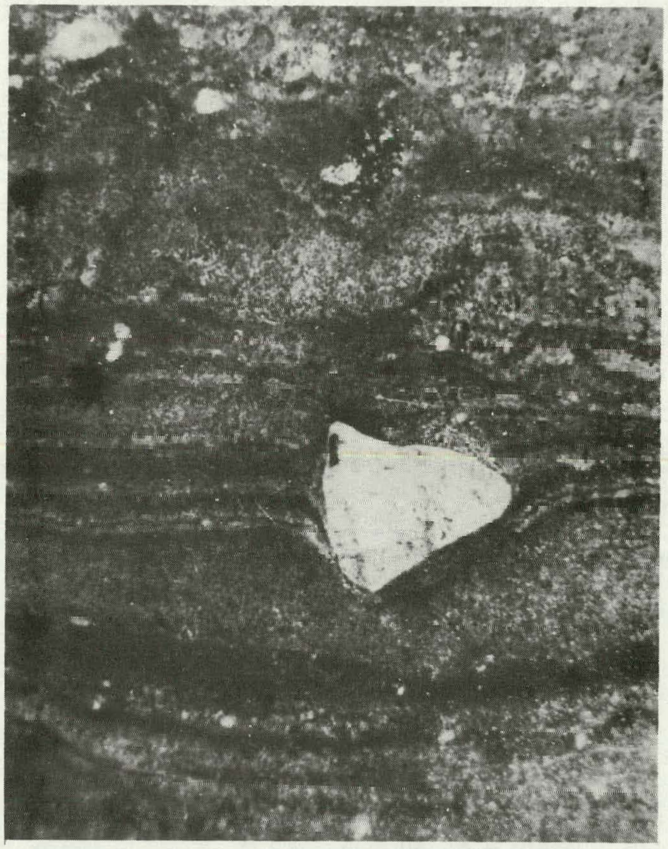


Figure 2.14. Polished slab of diamictite from the Headquarters Formation, near South Twin Lake, central Medicine Bow Mountains; granite clast (actual size) appears to have depressed underlying layers but subsequent layers are truncated against clast, suggesting that clast was dropped into soft sediments from floating iceberg. Photograph by George Sylvester.

These characteristics of the paraconglomerates are similar to descriptions of other glaciomarine paraconglomerates (Young, 1970) and the overall geochemistry of the Headquarters paraconglomerates (Table 2.14) is very similar to those of the glacial Gowganda Formation of the Huronian Supergroup (Young, 1973). This, combined with the association of the paraconglomerates with laminated, presumably marine, phyllites (metasiltsones), the rapid lateral facies changes in the lower member, and the appearance of multiple conglomerate lenses are all consistent with a glaciomarine depositional setting. Prodeltaic mudflows and turbidites might have most of the same features but would not be expected to contain dropstones. Kurtz and Anderson (1979) suggested, by analogy to the Antarctic continental margin sediments, that deposition of the Headquarters Formation took place at some distance from the ice sheet and on at least a moderate slope -- as suggested by evidence for turbidity flow in the paraconglomerates, the presence of laminated pebbly and non-pebbly argillites which both occur on the continental slope off Antarctica, the scarcity of paraconglomerates deposited directly from glacial ice (i.e. tillites) and the abundance of finer-grained rocks in the Headquarters Formation. We prefer prodelta and delta front depositional settings because of the stratigraphic position of the Headquarters Formation; overlying presumed deltaic deposits of the upper Deep Lake Group and conformably underlying quartzites of the Heart Formation which we also interpret to be prodelta and delta front sands.

Heart Formation

The Heart Formation (Blackwelder, 1926; Houston and others, 1968; Lanthier, 1979) lies conformably between the Headquarters Formation and the Medicine Peak Quartzite throughout the map area. The formation is 670 m thick and is predominantly quartzite. A phyllite unit up to 90 m thick is locally present about 400 m above the base of the Heart Formation (Figure 2.2).

The dominant lithology of the Heart Formation is a highly sericitic, very fine- to medium-grained quartzite (Table 2.17 and Figure 2.13). The unit becomes more quartzose near the top. In addition to quartz and sericite; plagioclase, chlorite, biotite, and opaque minerals may be present. Heavy minerals include zircon, tourmaline, sphene, rutile, and apatite (Table 2.17). Lanthier (1978) noticed a micaceous, quartz-pebble conglomerate at one locality. To the southwest, the metamorphic grade increases. Here, the Heart Formation is a biotite schist and may contain kyanite, garnet, and staurolite in addition to the minerals of Table 2.17 (Houston and others, 1968).

TABLE 2.17. PETROGRAPHY OF THE HEART FORMATION; MODES FROM POINT COUNTS (FIRST THREE) AND VISUAL ESTIMATES.

Sample No.	Qtz.	Plag.	Chlor.	Biot.	Musc.	Zirc.	Tour.	Opaq.	Sphn.	Rutile,	Apat.
9	52.1	5.5	—	6.7	33.8	0.6	Tr.	1.2	—	—	—
HL29	82.0	—	—	—	17.7	—	—	Tr.	—	—	—
HL26	86.5	—	4.5	—	8.7	—	—	Tr.	—	—	
SR76-8	45	10	—	3	41	—	Tr.	1	—	—	—
SR76-9	41	5	Tr.	3	50	—	—	1	—	—	—
SR76-10	48	5	3	3	40	—	—	1	—	—	—
Sericitic qtzite Houston and others, 1968)	46	7	15	3	25	—	Tr.	—	Tr.	Tr.	—
Mean of 2 mica schists (Houston and others 1968)	35	—	13	13	32	—	Tr.	3	Tr.	—	Tr.
MEAN	52.3	3.6	5.4	5.0	31.1	Tr.	Tr.	1.1	Tr.	Tr.	Tr.

A variety of sedimentary structures are present in the quartzites of the Heart Formation. These include small-scale planar and trough crossbedding, climbing ripples, interference ripples, symmetric ripples, ball and pillow structures, and graded bedding. The quartzites at the top of the formation are coarser grained and generally massive to plane bedded. The phyllites are well laminated (Lanthier, 1979).

The Heart Formation is interpreted as prodelta and delta front sediments associated with a prograding macrotidal (tide-dominated) delta (Figure 2.15). The laminated phyllites and very fine-grained, argillaceous, feldspathic quartzites of the lower two-thirds of the Heart Formation may represent prodelta bottom-set deposits (Reineck and Singh, 1975, p. 273). The presence of graded bedding, climbing ripples, and ball and pillow structures suggest rapid deposition of sediment temporarily thrown into suspension, perhaps as a result of slumping on the delta front slope (Blatt and others, 1972, p. 131; Reineck and Singh, 1972, p. 77; Elliott, 1978). The coarser-grained, sub-argillaceous, massive quartzites at the top of the Heart Formation represent delta front sediments (Eriksson, 1978, Elliott, 1978) where deposition can occur by grain flow (Blatt and others, 1972, p. 163). The occasional crossbeds and ripple marks in the upper part of the formation indicate shallowing of the water to where waves and tidal currents could affect the sediments (Elliott, 1978). A limited paleocurrent analysis (Plate 2) indicates that sediment transport was primarily to the southwest.

Medicine Peak Quartzite

The Medicine Peak Quartzite conformably overlies the Heart Formation and is conformably overlain by the Lookout Schist. In several localities

TIDAL FLATS : *no deposits of this facies preserved in Libby Creek Group*

DELTA PLAIN : *Cascade Quartzite (?), Medicine Peak Quartzite, Sugarloaf Quartzite*

DELTA FRONT AND DELTA PLAIN:
Headquarters Formation, Heart Formation, Lookout Schist

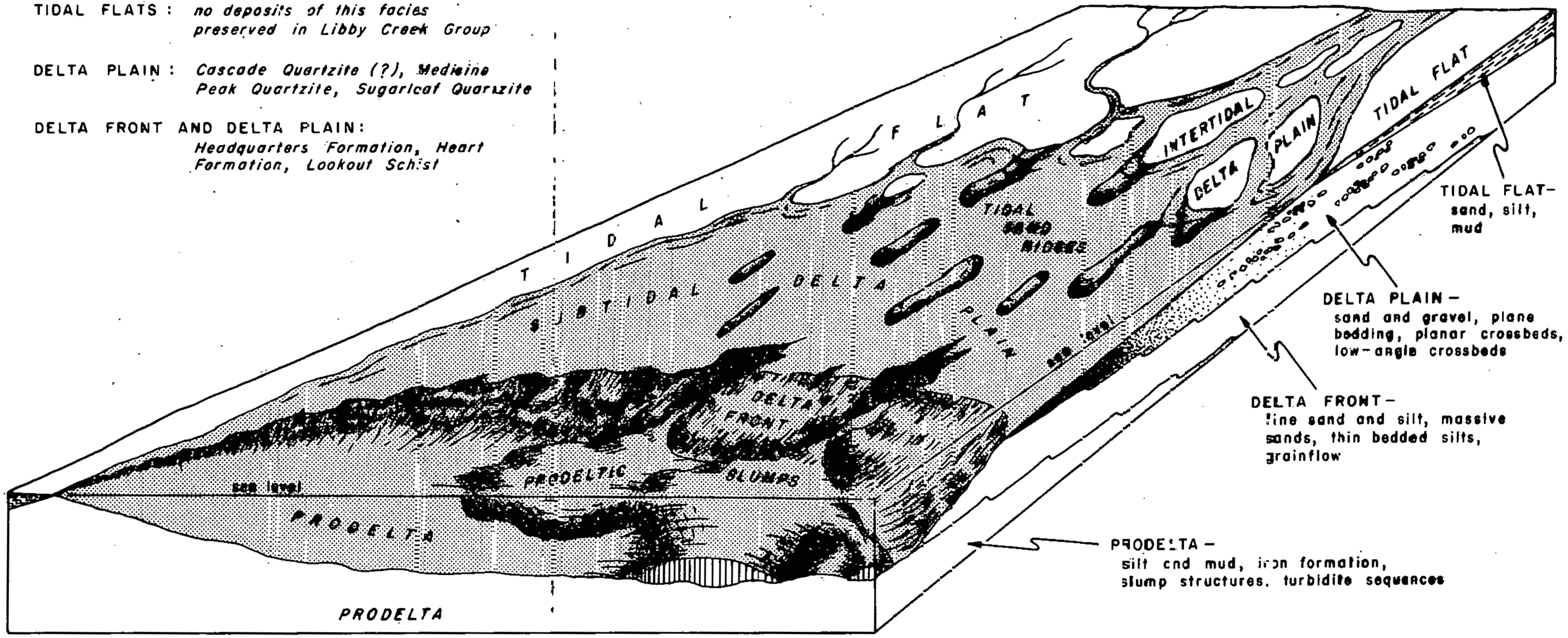


Figure 2.15 Deltaic facies model for deposition of the Lower Libby Creek Group.

it is in fault contact with younger metasediments. The quartzite is strongly folded in the vicinity of the French Creek Syncline (Houston and Parker, 1963). The quartzite is about 1700 m thick (Figure 2.2) and forms an impressive cliff in the Snowy Range. The following discussion is from Flurkey (1981).

The Medicine Peak Quartzite is predominantly a medium- to very coarse-grained quartzite with pebbly zones and layers of quartz-pebble conglomerate. The quartzite ranges from quartzarenite to argillaceous arenite (Figure 2.13). Sericite is common and kyanite, pyrophyllite, kaolinite, zircon, tourmaline, and iron oxide are minor constituents (Table 2.18). Rarely, plagioclase and biotite may be present. The aluminosilicate minerals: sericite, kaolinite, pyrophyllite and kyanite, may be in part an in situ alteration product (diagenetic and/or metamorphic) of feldspar suggesting that part of the unit was more arkosic than at present (Pettijohn and others, 1972, p. 408; Miyashiro, 1973, p. 199). A hematitic quartz-pebble conglomerate, here named the Klondike Lake Conglomerate, lies approximately 125 m below the top of the Medicine Peak Quartzite (Figure 2.2). The conglomerate is nearly 17 m thick and can be traced for about 9 km.

The quartzite is generally medium- to thick-bedded and massive to plane bedded. Planar, tabular crossbedding (average inclination = 23.5°) is the principal cross-stratification type in the Medicine Peak Quartzite. Trough crossbeds, asymmetrical ripples, large-scale low-angle crossbeds, and graded bedding are infrequent. The graded beds are conglomeratic and may show inverse as well as normal grading. Paleocurrent analysis (Plate 2) indicates a dominant southwest, largely unimodal directed current system. A few outcrops are polymodal and one shows only north directed currents.

TABLE 2.18. PETROGRAPHY OF THE MEDICINE PEAK QUARTZITE: MUSCOVITE COLUMN MAY INCLUDE PYROPHYLLITE AND KAOLINITE; KL27 IS FROM KLONDIKE LAKE CONGLOMERATE.

Sample No.	Qtz.	Plag.	Musc.	Biot.	Zirc.	Tour.	Opaq.	Sphn.	Kyanite
LM5	99.0	—	1.0	—	—	—	—	—	—
LM6	96.0	—	4.0	—	—	—	—	—	—
LM7	100		Tr.	—	—	—	—	—	—
TL13	91.7	—	8.3	—	Tr.	—	—	—	—
TL17	63.5		30.0	—	—	—	1.0	—	4.7
TL19	60.1	=	15.3						24.6
KL6	97.0	—	2.7	—	Tr.	Tr.	0.7	—	—
KL7	87.7		12.0	—	Tr.	Tr.	Tr.	—	=
KL8	94.6	—	1.4	—	—	—	4.0	—	—
KL9	8.5	Tr.	83.1	—	—	1.0	6.9	—	—
KL16	94.0	—	5.0	—	Tr	—	Tr.	—	—
KL18	90.4	—	6.6	—	Tr.	—	3.0	—	—
KL20	69.3	—	24.3	—	1.0	2.0	3.3	—	—
KL27	91.7	—	2.3	—	—	—	6.0	—	—
KL32	98.7	—	1.3	—	—	—	—	—	—
LL12	93.9	3.3	2.7	—	—	—	—	—	—
LL13	91.1	Tr.	7.8	Tr.	—	—	—	—	Tr.
LL14	61.0	—	38.1	—	—	Tr.	—	—	—
NG-1	90.8	—	7.9	—	—	—	1.3		
NG-3	89.4	—	10.6	—	—	—	—	—	—
NG-4	87.0	—	12.0	—	Tr.	—	—	—	Tr.
Sec 2	90.9	—	8.1	—	—	—	—	—	1.0
Sec 2-2	91.4	—	8.0	—	—	—	—	—	Tr.
6	86.7	—	12.0	—	—	—	Tr.	—	1.0
Mean of 8 samples from Houston and others (1968)	83	—	4	—	—	Tr.	Tr.	—	10
Mean of 7 samples from Houston and others (1968)	88	—	6	—	Tr.	Tr.	5	Tr.	—
MEAN	84.7	Tr.	9.7	Tr.	Tr.	Tr.	1.6	Tr.	2.9

Available evidence suggests that the Medicine Peak Quartzite was deposited on the subtidal to intertidal portions of a tidally dominated delta plain (Figure 2.15). Stream transported arkosic, argillaceous sands are presumed to have been deposited in the supratidal zone (not presently seen) and reworked by the macrotidal currents in the intertidal zone. The paucity of feldspar in this unit suggests feldspar may have been selectively abraded and reduced in grain size and thus transported with the finer-grained sediments to lower-energy envrionments; i.e. tidal flats or delta front (Field and Pilkey, 1969; Balaz and Klein, 1972).

The bulk of the Medicine Peak Quartzite appears to represent the coarser-grained and quartz-rich sediments that remained in the high-energy intertidal and subtidal environments (Tankard and Hobday, 1977; Button and Vox, 1977; Eriksson, 1979). Large-scale, low-angle crossbeds are interpreted to represent deposition of sand on the subtidal delta as large linear shoals elongate in the direction of tidal currents (Wright and others, 1975; Klein, 1970). Planar and tabular crossbeds presumably represent sand waves formed on the shoals and in the intershoal channels in response to tidal currents. The dominantly unimodal paleocurrent distribution in the Medicine Peak Quartzite suggests that most of the sand waves formed in response to ebb-tidal currents, in keeping with the time-velocity asymmetry of tidal currents (Klein, 1970). North-directed paleocurrents indicate the dominance of flood tides locally. The local bimodal (south- and west-directed) paleocurrents suggest bedforms moving in channels (south-directed) and on shoals (west-directed). The pebbly quartzites and quartz-pebble conglomerates appear to represent channel fill sequences (Eriksson, 1977) in the wide, shallow sub-tidal channels.

The origin of the iron-oxide grains in the Klondike Lake Conglomerate is uncertain but the lack of other heavy minerals suggests a near-by source for the iron oxide grains. A possible location for iron formation deposition is in lagoons on the tidal flats (not seen here) commonly adjacent to tide-dominated deltas (Von Brunn and Hobday, 1976).

Migrating tidal or fluvial channels could occasionally rip up a lagoonal deposit and transport the sediment out into the delta.

Lookout Schist

The Lookout Schist lies conformably between the Medicine Peak and Sugarloaf Quartzites. This unit is about 400 m thick (Figure 2.2) but may be thinner or absent locally due to faulting (Plate 1).

The Lookout Schist is a variable unit consisting of phyllites and subarkosic, arkosic, and argillaceous quartzites (Figure 2.13). Grain size ranges from clay to fine sand. Quartz and sericite are the principal minerals with variable amounts of plagioclase, K-feldspar, chlorite, biotite, tourmaline, calcite and iron oxides (Table 2.19). Tremolite and siderite have also been reported (Houston and others, 1968). Ferruginous schist is present with abundant iron-oxide grains in a matrix of mica and silt-sized quartz and feldspar. Graded bedding in some of these sediments indicates that some of the iron-oxide grains are detrital rather than a chemical precipitate. One thin metadolomite bed and one thick lense of "Medicine Peak-like" quartzite were also observed. The Lookout Schist is strongly deformed (Wilson, 1975; 1977) and has been intruded by many mafic sills.

The principal sedimentary structure in the Lookout Schist is a laminated phyllite with primary quartz-rich and mica-rich layers crosscut by one or more foliations, including probable slaty cleavage (Wilson, 1975). In many localities, these laminae are folded and contorted by syndepositional and/or tectonic deformation. In some places, clastic dikes were injected across the laminae (Wilson, 1975). One syndepositionally deformed bed consists of randomly rotated blocks of fine-grained quartzite in a massive metasiltstone matrix. Small-scale planar crossbedding is common in fine-grained arkosic quartzites, especially near the top and bottom of the unit. These quartzites may also contain hummocky cross-stratification, climbing ripples, and flat-pebble conglomerates. Symmetric ripples, graded bedding, and sole marks (Wilson, 1975) have also been observed in the Lookout Schist. Paleocurrent analysis (Plate 2), primarily from fine-grained quartzites at the top of the unit, shows a bimodel-bipolar, northeast-southwest distribution.

TABLE 2.19. PETROGRAPHY OF THE LOOKOUT SCHIST AND SUGARLOAF QUARTZITE MODES MAINLY FROM POINT COUNTS.

LOOKOUT SCHIST

Sample No.	Qtz.	Plag.	K-spar.	Chlor.	Biot.	Musc.	Zirc.	Tour.	Opaq.	Sphn.	Apat.	Epid,	Garnt.	Carb.
SR77-2	40	33	–	15	–	10	–	–	2	–	–	–	–	–
LM-13	65.5	24.2	–	–	–	10.3	–	–	–	–	–	–	–	–
LM-14	83.4	16.0	–	–	–	Tr.	–	–	–	–	–	–	–	–
LM-16	84.4	8.9	–	–	–	6.6	–	–	–	–	–	–	–	–
TI24	64.1	7.6	–	2.0	–	17.4	–	Tr.	8.4	–	–	–	–	–
KL1	30.8	16.2	–	–	–	52.7	–	Tr.	–	–	–	–	–	–
KL2	44.3	17.3	6.0	–	–	32.3	Tr.	–	Tr.	–	Tr.	–	–	–
KL3	92.0	7.3	–	–	–	0.7	Tr.	–	–	–	–	–	–	–
KL4	45.7	7.7	3.3	–	–	42.7	0.7	–	–	–	–	–	–	–
KL5	83.7	6.7	0.7	–	–	7.7	Tr.	Tr.	1.0	–	–	–	–	–
LL4	53.7	28.5	Tr.	–	–	3.3	Tr.	–	1.5	–	–	–	–	13.1
LL5	43.7	25.7	–	–	–	26.9	–	Tr.	2.1	–	–	–	–	–
LL6	10.3	12.1	–	–	–	76.7	–	Tr.	0.8	–	–	–	–	–
LL8	23.1	3.7	–	–	–	72.6	–	Tr.	–	–	–	–	–	–
LL10	75.6	20.6	Tr.	–	–	3.3	–	–	0.5	–	–	–	–	–
LL11	56.4	28.0	2.8	–	Tr.	5.6	–	–	5.2	–	–	–	–	–
Mean of 6 samples from Houston and others (1968)	53	9	–	7	–	17	Tr.	Tr.	5	Tr.	Tr.	Tr.	Tr.	Tr.
MEAN	55.2	14.4	0.6	Tr.	Tr.	21.4	Tr.	Tr.	2.3	Tr.	Tr.	Tr.	Tr.	0.6

SUGARLOAF QUARTZITE

Sample No.	Qtz.	Musc.	Tour.	Opag.
59-33	99.0	1.0	Tr.	–
SR77-21	97	1	–	2
SR77-22	99	–	–	1
MEAN	98.3	0.7	Tr.	1.0

The depositional environment of the Lookout Schist appears to be similar to that of the Heart Formation (Figure 2.15). The graded and contorted beds, clastic dikes, and laminated phyllites suggest prodelta and delta front deposition (Blatt and others, 1972; Reineck and Singh, 1975; Elliott, 1978). The small-scale planar and hummocky crossbeds, flat-pebble conglomerates, and climbing ripples in the fine-grained arkosic quartzites at the top and bottom of the unit suggest storm and reversing tidal current deposits on the shallow water portion of the delta front (Bourgeois, 1980). The lense of "Medicine Peak-like" quartzite appears to be a delta front distributary channel fill deposit (Elliott, 1978). Carbonates and possible iron formation are interpretted to represent sporadic chemical sedimentation in the prodelta basin. Some of the detrital iron-oxide grains may have the same source as the similar grains in the Klondike Lake Conglomerate.

Sugarloaf Quartzite

The Sugarloaf Quartzite conformably overlies the Lookout Schist but is in fault contact with the overlying Nash Formation (Plate 1). The quartzite is at least 580 m thick but is commonly thinner or absent due to faulting (Lanthier, 1979). The abrupt disappearance of this thick unit along strike in two places (Plate 1) is one of the main justifications for new map interpretations which show major thrust faults in the Libby Creek Group.

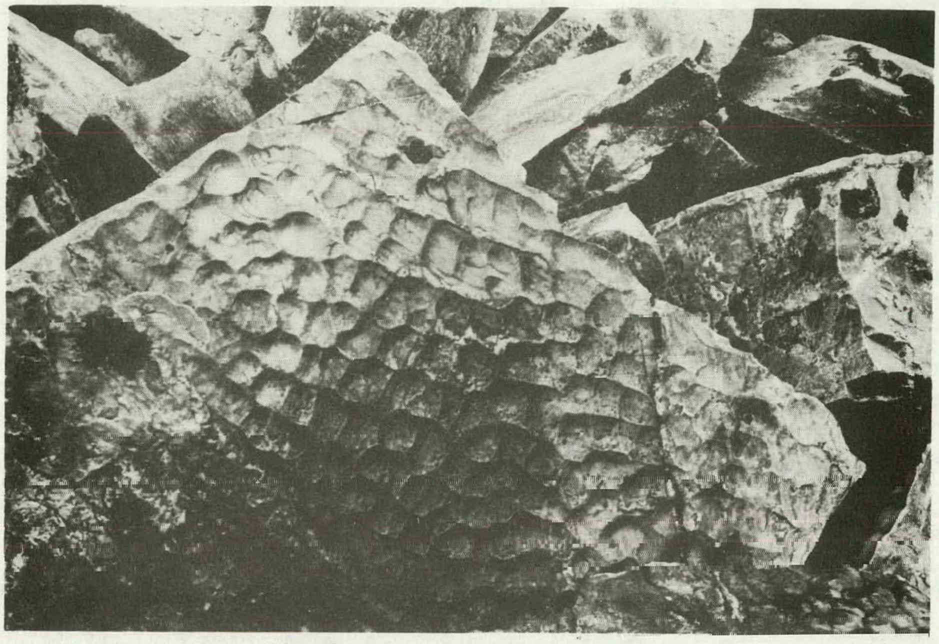
The Sugarloaf Quartzite is a monotonously white, medium-grained quartzarenite (Figure 2.13). Sericite, tourmaline, and opaque minerals are present in minor amounts (Table 2.19). We observed thin beds of hematitic quartzite near the upper end of the unit and Blackwelder (1926) reported a quartz-pebble conglomerate in a similar stratigraphic position.

The dominant sedimentary structure is medium- to thin-bedded plane beds. Other common sedimentary structures include small-scale planar and trough crossbeds, and symmetric, interference, and climbing ripples (Figure 2.16). A few larger-scale planar crossbeds were observed. Paleocurrent analysis of crossbedding (Plate 2) indicates that sediment transport was dominantly to the west-southwest. Oscillation ripples record north-south and east-west current sense (Plate 2).

The maturity of the sediment and the abundance of small-scale sedimentary structures, especially oscillation ripples, suggests that the Sugarloaf Quartzite was deposited in a shallow marine environment. The shallowing of the water is probably in response to progradation of a delta system as in the Medicine Peak Quartzite (Figure 2.15). The abundance of oscillation ripples, however, suggests that wave activity was becoming more important as a sediment mover at the expense of ebb-tidal currents (Reineck and Singh, 1975).

UPPER LIBBY CREEK GROUP

The Upper Libby Creek Group is here defined to include the three uppermost formations of the metasedimentary succession in the Medicine Bow Mountains: Nash Fork Formation, Towner Greenstone, and French Slate. A major change occurred in the sedimentation pattern between the siliciclastic (deltaic) succession of the Lower Libby Creek Group and the more varied lithologies of the (open marine) Upper Libby Creek Group. The



A.

B.

Figure 2.16. Photographs of the Sugarloaf Quartzite; A. interference ripple marks from SW1/4, Sec. 31, T.17 N., R.78 W.; B. trough cross-bedding from Sec. 35, T.16 N., R.80 W.

change in sedimentation is probably a reflection of evolving tectonic conditions (discussed later).

Nash Fork Formation

As discussed earlier, the Nash Fork Formation is in fault contact with the Lower Libby Creek Group along the Lewis Lake Fault. The upper contact is not exposed (Houston and others, 1968, p.36) but the Nash appears to be structurally conformable with the overlying Tower Greenstone (Plate 1). The unit is at least 1980 meters thick (Figure 2.2) but originally may have been thicker because of lost stratigraphic section due to faulting.

The Nash Fork Formation consists predominantly of tan metadolomite with thick lenses of black phyllite. Rarely, thin beds of quartzite, metachert, flat-pebble conglomerate, and iron formation may be present. The most striking sedimentary feature of the metadolomite is stromatolitic bioherms (Figure 2.17) which occur in a variety of shapes and sizes (Knight and Keefer, 1966, Knight, 1968).

The metadolomite consists primarily of dolomite and quartz with variable amounts of plagioclase, muscovite, phlogopite, tremolite, talc, apatite, and opaque minerals (Table 2.20). The phyllites consist of muscovite, chlorite, biotite (and/or phlogopite), zoisite, sphene, tremolite, quartz, dolomite, and pyrite and other opaque minerals (Table 2.20). Pyrite, graphite, and hematite have been observed in a small lense of iron formation within a larger, black phyllite lense (Houston and others, 1968).

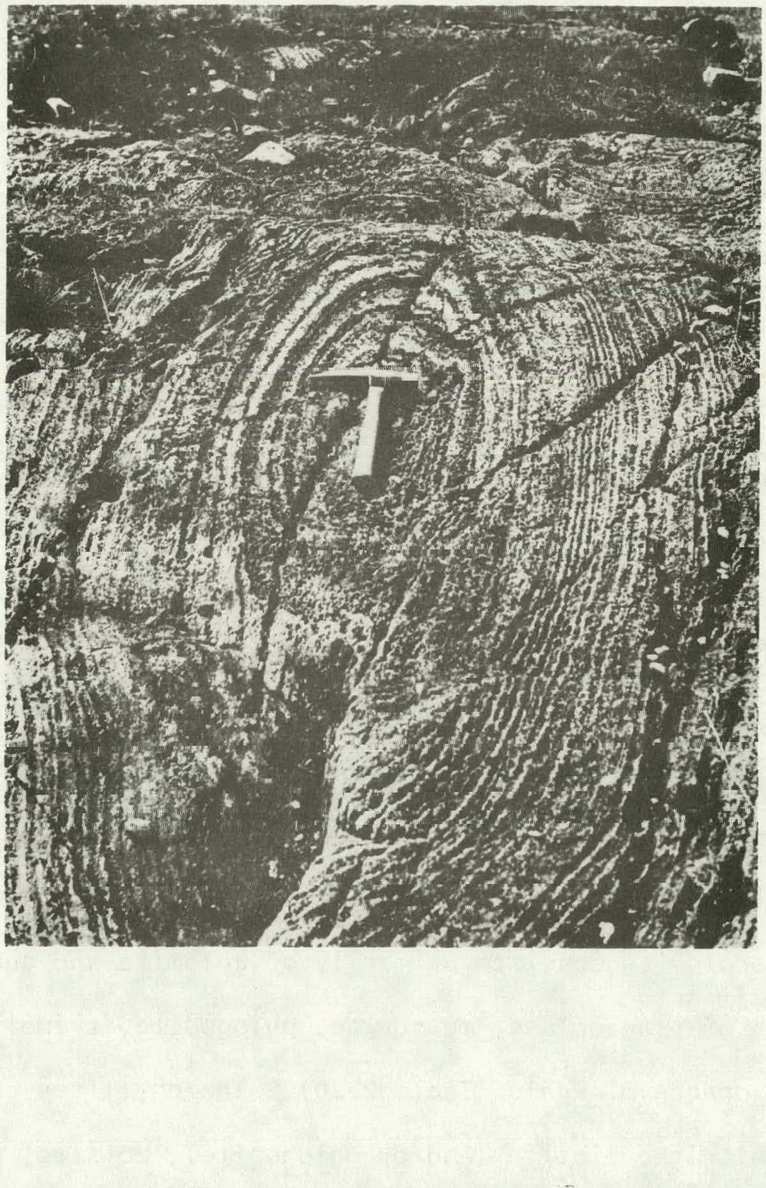


Figure 2.17. Stromatolite in Nash Fork Formation in locality near Prospect Lake, central Medicine Bow Mountains.

Knight and Keefer (1966) and Knight (1968) mapped about 150 bioherms and three reefs in the metadolomite. The reefs are about 60 m thick and perhaps a kilometer in length and consist of many bioherms. The bioherms range from one to about thirty meters thick and from three to about thirty meters long (Knight, 1968). Knight (1968) subdivided them into several types and forms. No specific organisms have been identified from these stromatolites. Some of the bioherms show evidence of erosion. Crossbedding and flat-pebble conglomerates are other sedimentary structures found in the dolomites (Houston and others, 1968; Knight, 1968).

TABLE 2.20. PETROGRAPHY OF THE UPPER LIBBY CREEK GROUP MODES FROM VISUAL ESTIMATES.

NASH FORMATION

	Sample No.	Qtz.	Plag.	Biot. Phlog.	Chlor.	Musc.	Pry.	Epid.	Opaq.	Sphn.	Carb.	Amph.	Talc	Apat.
"CARBONATES"	C-546	70	—	—	—	3	—	—	2	—	10	15	—	—
	C-555	40	30	1	—	7	—	—	2	—	30	—	—	—
	59-37	35	5	—	—	—	—	—	Tr.	—	35	—	20	Tr.
	59-39	10	5	—	—	5	—	—	—	—	80	—	—	—
	60-57	15	—	—	—	Tr.	—	—	—	—	84	—	—	—
	MEAN	34.0	8.0	Tr.	—	3.0	—	—	0,8	—	47.8	3.0	4.0	Tr.
SLATES	59-34	45	—	45	—	—	Tr.	—	—	—	10	—	—	—
	59-38	55	—	—	5	33	—	—	7	—	—	—	—	—
	60-83	20	—	1.0	20	2	Tr.	13	—	5	15	15	—	—
	MEAN	40.0	—	18.3	8.3	11.7	Tr.	4.3	2.3	1.7	8.3	5.0	—	—

TOWNER GREENSTONE

Sample No.	Qtz.	Plag.	Chlor.	Epid.	Opag.	Sphn.	Carb.	Amp.
Houston and others (1968)	15	—	—	Tr.	12	—	15	5.7
59-36	—	5	15	7	—	Tr.	—	73

FRENCH SLATE

Sample No.	Qtz.	Biot.	Musc.	Opag.
59-31	10	1	74	15
59-32	60	—	35	5
59-35A	50	—	40	10
MEAN	40	Tr.	49.7	10

Fenton and Fenton (1939) suggested that the bioherms grew in clear, shallow water. The abundant siliciclastic particles indicate that at least part of the time there was a considerable influx of clastic debris. Current activity is also shown by crossbedding, flat-pebble conglomerates, and erosional scours on the bioherms. The presence of pyrite and graphite in the phyllites implies restricted circulation for at least part of the time (James, 1966), suggesting variable currents. The influx of clastic sediments, erosional features, and possible variable currents suggests to us that the Nash Fork Formation was deposited on intertidal flats. Similar features have been described from Lower Proterozoic carbonates in South Africa (Button and Vos, 1977; Beukes, 1977).

Towner Greenstone

The Towner Greenstone is reported to be conformable with the overlying French Slate (Childers, 1957, p. 27). This formation is 180 to 490 meters thick (Houston and others, 1968, p. 37) but appears to pinch out to the southwest (Plate 1), perhaps due to faulting.

The Towner Greenstone consists of massive to schistose amphibolite with several small lenses of coarse-grained sandstone and fine-grained quartzites which may be meta-cherts. The greenstones consist primarily of actinolite with variable amounts of chlorite, albite, epidote, carbonates, opaque minerals and sphene (Table 2.20). No diagnostic genetic features have been observed although the sandstone lenses suggest an extrusive origin (Houston and others, 1968) and the cherts (?) suggest marine deposition. An intrusive origin cannot be ruled out.

French Slate

The French Slate overlies the Towner Greenstone and is truncated by the Mullen Creek-Nash Fork shear zone along its entire upper contact. Its present map thickness is about 610 m. This formation consists primarily of laminated black ferruginous slate and phyllite. The laminae appear primary and consist of layers containing roughly equal amounts of muscovite, chlorite, quartz, and opaque minerals alternating with quartz-rich layers with minor muscovite and chlorite. Metacrysts of biotite and pyrite are common (Houston and others, 1968). We have observed two thick lenses of hematitic iron formation in the upper part of the unit. The hematite is in a matrix of very fine-grained quartzite (metachert). The French Slate has been complexly folded and crenulated by movement along the shear zone.

The lack of coarse clastics and the presence of iron formation imply that the French Slate was deposited in a low-energy environment. The thickness of the unit suggests that deposition was in a deep marine or prodelta basin (Reineck and Singh, 1975).

INTRUSIVE IGNEOUS ROCKS NORTH OF SHEAR ZONE

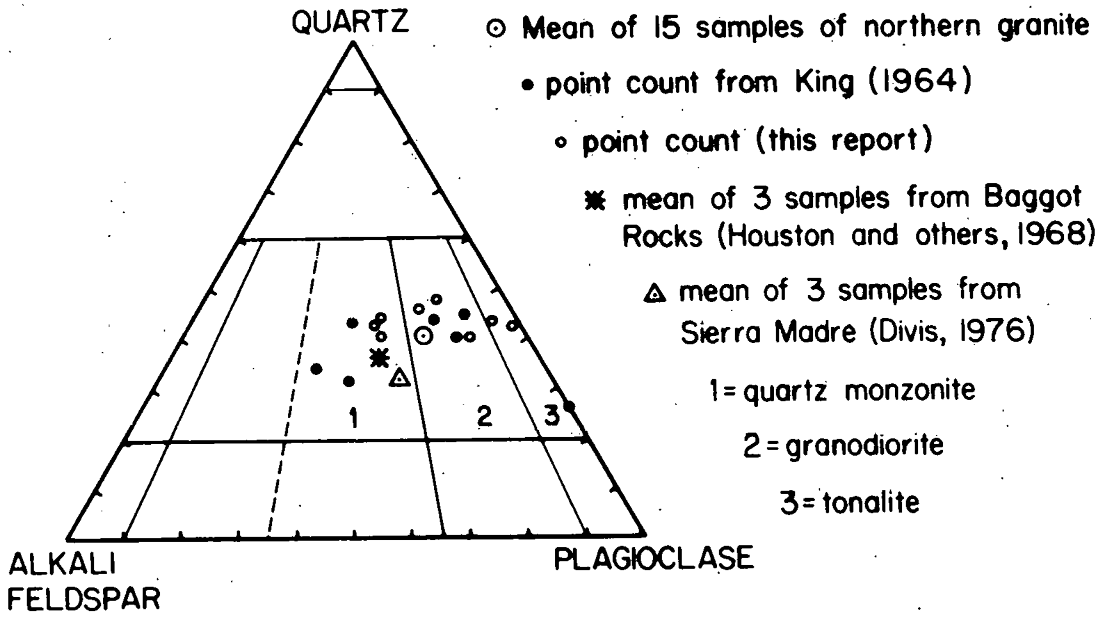
Baggot Rocks-type Granite

The oldest known intrusive rocks in the Medicine Bow Mountains are granites and granitic gneisses which crop out near Baggot Rocks in the western Medicine Bow Mountains; near Lincoln Park in the western part of Plate 1; and in the northern Medicine Bow Mountains, near Arlington. In the Baggot Rock area, these granitic gneisses crosscut Archean quartzofeldspathic gneisses and have yielded an age of 2425 ± 50 m.y. (Hills and Houston, 1979). These rocks were named the Baggot Rocks Granite by Houston and others (1968) and we use the term Baggot Rocks-type granite for the Lincoln Park and Arlington granite bodies as well because of the similarities in field relationships and geochemistry. Neither of the later two bodies has been adequately dated.

In all of these areas the granites are cross-cutting in detail but, in large scale, are sill-like bodies which generally are conformable with foliation in Archean gneiss or metasedimentary rocks. The granites contain foliated, sheared, and massive varieties which are complexly interleaved among themselves and with the surrounding country rock and locally they appear to grade into the country rock through migmatized zones.

Petrographically the granites are heterogeneous. Figure 2.18 shows compositions ranging from quartz monzonite to tonalite. (Presumably similar Baggot Rocks-type granite in the Sierra Madre (Divis, 1977) ranges from quartz monzonite to granodiorite.) All of these granites contain modal muscovite (mean = 6%; range up to 15%) and many (9 of 15) contain modal garnet. These minerals would appear to indicate relatively high aluminum content for these rocks. Chemical compositions and normative mineralogies for Baggot Rocks-type granites in the Medicine Bow Mountains are shown in Table 2.21. Aluminum content does not appear to be as high as we would have guessed from mineralogy; they ranged from peralkaline to peraluminous (Carmichael and others, 1974, p. 31). This variation probably reflects metamorphic conditions, where alkalies are mobile but aluminum is not, rather than any primary igneous properties of the granites.

We tentatively interpret the Baggot Rocks-type granites in the Medicine Bow Mountains to be S-type granites (derived from a sedimentary source) on the following evidence (following Chappel and White, 1974; and Miller and Bradfish, 1980): gradational contacts with country rock, association with some migmatites and pegmatites, presence of peraluminous minerals muscovite and garnet, relatively low sodium (true only for the Arlington granite body), relatively high silica, high Sr_{87}/Sr_{86} ratio for the Baggot Rocks Granite, and the somewhat scattered nature of isochrons from both the Baggot Rocks and Arlington areas (Hills and others, 1968) which may indicate melting from a heterogeneous source.



MAFIC MINERALS

	$\bar{x}$	range	fraction of samples with mineral
muscovite	6%	T-14	15/15
biotite	.5%	0-4	4/15
garnet	.2%	0-2	9/15
epidote	trace	0-1.3	4/15
magnetite	.3%	0-1	8/15

Figure 2.18. Modal compositions of Baggot Rocks-type granites.

TABLE 2.21 WHOLE-ROCK GEOCHEMISTRY OF GRANITIC ROCKS FROM THE MEDICINE BOW MOUNTAINS; ANALYST STEVE BOESE, UNIVERSITY OF WYOMING; 2. FROM HOUSTON AND OTHERS (1968)

	1.	2.	3.	4.	5.
SiO_2	70.7	70.66	73.5	72.3	67.2
Al_2O_3	12.6	13.45	13.1	13.6	13.7
CaO	0.81	1.67	1.36	0.12	1.84
Na_2O	4.26	3.45	5.44	2.72	4.17
K_2O	3.32	5.13	2.40	9.22	2.91
TiO_2	0.4	0.41	0.0	0.0	0.6
Fe_2O_3	4.21	1.88	1.36	0.49	4.17
MnO	0.05	–	0.02	0.01	0.07
MgO	1.39	0.73	0.32	0.05	2.25
P_2O_5	0.14	–	0.04	0.01	0.25
H_2O	–	1.36	–	–	–
Total	98	98.74	97	99	97
U (ppm)	–	–	0.9	0.4	2.3
Th (ppm)	–	–	22	<5	15
Q	25.96	24.44	27.41	21.96	21.75
Or	20.20	31.35	14.6	55.50	17.75
Ab	39.30	31.95	49.85	19.95	38.60
An	3.25	6.23	4.20	–	7.78
Mt	0.59	0.27	0.20	–	0.59
Il	0.58	0.58	–	–	0.72
Di	–	1.88	2.08	0.08	–
Hy	8.91	3.60	1.64	0.78	11.25
Cor	0.93	–	–	–	1.04
Ap	–	–	0.10	0.16	0.53
*Sm	–	–	–	1.40	–
Ol	–	–	–	–	–
Ac	–	–	–	0.16	–

1. SR79-20: Peraluminous quartzo-feldspathic gneiss from Troublesome Ridge: center Sec. 24, T17N, R82W
2. Baggot Rocks Granite (Houston and others, 1968) — subaluminous
3. SR59-30: Baggot Rocks — type (subaluminous) granite from Lincoln Park area; center Sec. 3, T16N, R81W
4. K-261: Baggot Rocks type (peralkaline) granite near Arlington, NW¼, NE¼, Sec. 31, T19N, R78W
5. SR60-55: Peraluminous tonalite near Foote Creek: SE¼, SE¼, Sec. 28, T19N, R79W

*Sm = "sodium metasilicate" Na_2SiO_3

The Arlington granite body is of particular interest here because it intrudes the Stage Crossing Gneiss and Unit 2, the Rock Mountain Conglomerate, of the Phantom Lake Suite but appears to be unconformably overlain by, and a source of detritus for, the Magnolia Formation of the Deep Lake Group. Thus, because we assume the granite is Archean by analogy to the Baggot Rocks Granite, we believe the Phantom Lake Metamorphic Suite to be older than 2400-2500 m.y. and the Deep Lake Group to be younger. Also, this body, and ones like it, appear to have been the source of granite rock fragments and K-feldspar found in the Magnolia Formation of the Onemile Creek area. Was it also a source for the uranium? Our limited geochemical data (12 samples) on the Arlington granite body (Table 2.22) show a mean of 2.5 ppm U (range 0.2 to 8.8) and 17 ppm Th (range 22 to 49) which is not anomalously high compared with "average granite" compositions of about 4 ppm U, and 18 ppm Th (Gabelman, 1975; Nishimori and others, 1977). Nevertheless, the area of outcropping granite is small and more radioactive phases of Late Archean granite may have been present farther north. As examples, granite in the Potato Creek area of the northern Laramie Range contain up to 96 ppm U, 535 ppm Th and the Baggot Rocks granite of the western Medicine Bow Mountains was reported by Charlton (written communication, 1980) to contain up to 5.5 ppm U and 610 ppm Th in outcrop.

Gabbroic Intrusive Rocks

All Precambrian rocks in the Medicine Bow Mountains, with the exception of the 1400 m.y. old Sherman Granite, are cross-cut by mafic sills or dikes of gabbroic composition which have been entirely or partly converted to amphibolite (see Houston and others, 1968 for a discussion of the petrography of these rocks). In the Archean quartzo-feldspathic gneiss terrane of the western Medicine Bow Mountains (Figure 2.1), some of these dikes are folded with the gneisses into north- and northwest-trending folds (Houston and others, 1968). This fold system has not been observed in overlying metasedimentary rocks (see section on structural analysis) so that these mafic bodies are interpreted to be Archean in age. The ages of mafic intrusives which cross-cut the metasedimentary rocks are less well constrained. However, we postulate at least two ages of intrusions, in addition to the Archean intrusions, based mainly on geologic relationships.

TABLE 2.22. U, Th CONTENT OF BAGGOT ROCKS-TYPE GRANITES IN THE ARLINGTON AREA, NORTHERN MEDICINE BOW MOUNTAINS.

Sample No.	U (ppm)	Th (ppm)
K-261	0.4	<5
157010	2.3	15
157138	2.7	19
157157	0.8	<2
160065	5.4	—
160133	0.2	12
160142	0.5	11
160150	0.3	45
160151	8.8	49
160154	1.0	5
711050	4.9	15
711041	3.0	11
MEAN	2.53	17
Average granite (Gabelman, 1975)	3.6	19.0

Gabbroic intrusives which cross-cut the Phantom Lake Suite and Deep Lake Group are generally large (kilometers long and several hundred meters wide) phacolithic bodies which we loosely refer to as sills but which cross-cut F_2 folds of bedding. Intrusives which cross-cut the Lower Libby Creek Group are less abundant and are mainly small dikes (Plate 1). Large sills reappear in the Upper Libby Creek Group and in areas close to the Mullen Creek-Nash Fork shear zone, some of which cross-cut cataclastic foliation in the shear zone. It is impossible to say with certainty, but structural evidence, discussed in detail later, suggests that most of the large sills in the Phantom Lake Suite and Deep Lake Group were emplaced before movement on the Reservoir and Lewis Lake Faults, and the Mullen Creek-Nash Fork shear zone. On geologic grounds, the most likely time of emplacement would have been during a rifting event, discussed in detail later, which preceded and accompanied formation of the Libby Creek Group miogeoclinal basin. In contrast, many of the sills and dikes in close proximity to (and south of) the shear zone appear to have post-dated movement on the shear zone (Plate 1).

Geochemical data, shown in Table 2.23, from the various gabbroic intrusions suggests a difference between the sills in the Deep Lake Group and Phantom Lake Suite and the intrusions in the Libby Creek Group. Chemical analyses of five samples from four separate intrusive bodies in the Deep Lake Group show them all to be quartz normative tholeiitic gabbros (Cox and others, 1979). Figure 2.19 shows them to be closely related to one another and distinctly tholeiitic both on the alkali versus silica plot (Figure 2.19A) and on the AFM diagram (Figure 2.19B; this figure also includes some partial analyses of other intrusives in the Deep Lake

Group). The tholeiitic iron enrichment trend is comparable to that of Hawaiian or Mid-Atlantic basalts (Irvine and Baragar, 1971).

Two samples from intrusives cutting the Phantom Lake Suite have also been analyzed, and are included in Figure 2.19. They are tholeiitic gabbros similar to those of the Deep Lake Group.

Analyses of seven samples from intrusives in the Libby Creek Group (Table 2.23) also fall in the general class of gabbros but they show a higher average alkali content and lower average silica content than the Deep Lake Group and Phantom Lake Suite sills. On the alkali versus silica plot of Figure 2.19A they fall clearly outside the tight field of the Deep Lake Group samples. Three of these samples plot as alkalic rocks (MacDonald, 1968) despite containing normative hypersthene. On the AFM plot the Libby Creek Group samples again fall outside the field of the Deep Lake Group samples. The iron enrichment trend, although not well defined by the limited data, is intermediate between those of tholeiitic and calc-alkaline rocks (Irvine and Baragar, 1971). These rocks are similar to rocks described by Jakes and Gill (1970) as "island arc tholeiites," in that while they are normative tholeiites they have anomalously high total alkali contents, particularly Na, and they show too much iron enrichment to be properly called calc-alkaline.

Analyses of three samples from gabbroic intrusives in the Archean gneiss are also shown in Table 2.23. The wide variation in their normative compositions and their scattered distribution in Figure 2.19 indicates a different, more complex history than any of the other groups considered.

TABLE 2.23 GEOCHEMISTRY OF MAFIC INTRUSIVE ROCKS OF THE MEDICINE BOW MOUNTAINS. SAMPLES 1, 6, AND 11 ARE FROM HOUSTON AND OTHERS, 1968. REMAINDER ANALYZED BY STEVE BOESE. WHERE NO FeO VALUE IS GIVEN TOTAL Fe IS REPORTED AS Fe_2O_3, AND Fe_2O_3/FeO IS ASSUMED TO BE 0.15 FOR NORMATIVE CALCULATIONS (COX AND OTHERS, 1979).

	ARCHEAN ROCKS			PHANTOM LAKE SUITE		DEEP LAKE GROUP					LIBBY CREEK GROUP						
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
SiO_2	50.20	52.90	50.80	52.70	49.10	51.73	51.50	51.00	53.0	53.30	48.91	49.00	51.00	52.20	50.20	46.00	49.50
Al_2O_3	12.24	7.20	19.60	13.10	18.90	15.64	12.90	14.60	13.80	14.80	14.51	14.70	17.60	17.30	13.00	14.90	12.90
Fe_2O_3	2.61	11.42	10.10	9.39	10.32	2.55	14.38	9.65	10.09	11.70	4.39	10.56	10.04	10.88	11.71	12.60	13.33
FeO	7.84	–	–	–	–	7.98	–	–	–	–	8.81	–	–	–	–	–	–
MgO	9.82	16.63	5.73	10.34	6.29	6.51	6.65	8.73	8.13	6.84	5.12	7.77	3.41	4.00	7.06	6.78	6.86
CaO	12.04	7.09	9.71	11.48	9.66	11.26	10.42	9.83	11.31	10.21	7.76	10.27	8.45	4.87	8.18	10.08	9.71
Na_2O	1.39	1.02	3.16	1.42	2.55	2.07	2.06	1.84	1.67	2.04	2.74	2.13	3.21	3.65	3.35	2.29	1.59
K_2O	0.40	0.06	0.43	0.13	0.31	0.39	0.19	0.47	0.26	0.39	1.61	0.83	1.31	2.29	0.47	0.22	0.34
TiO_2	0.76	0.40	1.00	0.50	0.40	0.73	1.10	0.40	0.60	0.60	1.95	0.90	0.90	1.00	1.30	1.00	1.00
P_2O_5	0.06	0.04	0.11	0.03	0.08	0.06	0.08	0.05	0.04	0.05	0.16	0.09	0.08	0.12	0.10	0.09	0.09
MnO	0.16	0.20	0.13	0.18	0.14	0.15	0.21	0.15	0.16	0.17	0.14	0.17	0.13	0.16	0.15	0.17	0.20
H_2O^+	1.56	–	–	–	–	0.77	–	–	–	–	2.73	–	–	–	–	–	–
H_2O^-	0.38	–	–	–	–	0.53	–	–	–	–	0.55	–	–	–	–	–	–
CO_2	0.99	–	–	–	–	0.01	–	–	–	–	0.47	–	–	–	–	–	–
TOTAL	100.45	96.96	100.74	99.27	97.75	100.38	99.49	96.72	99.06	100.01	99.85	96.42	96.13	96.47	95.52	94.13	95.52
								MOLECULAR NORMS									
Q	2.46	2.29	–	3.54	–	4.03	3.21	2.06	3.70	4.48	2.09	–	0.93	–	–	–	4.10
Or	2.31	0.35	2.50	0.80	1.90	2.35	1.00	3.00	1.50	2.30	9.50	9.10	8.00	14.10	9.99	1.40	2.15
Ab	11.71	9.30	28.10	12.85	23.40	18.95	19.00	17.00	15.00	18.50	23.08	19.90	30.00	34.05	31.55	22.05	15.25
Am	25.90	15.10	37.70	29.15	40.08	32.85	26.30	31.30	29.80	30.43	22.51	29.23	31.30	24.32	19.88	31.85	28.93
Hy	22.30	52.18	7.48	29.19	21.32	18.91	25.98	10.03	26.50	25.30	16.20	22.58	17.43	23.30	20.00	14.36	28.32
Di	27.23	19.08	15.02	22.44	6.84	19.04	20.70	14.80	21.10	16.44	12.07	18.60	9.40	–	17.62	16.96	17.64
Ol	–	–	6.47	–	4.32	–	–	–	–	–	–	1.63	–	0.72	4.17	9.84	–
Cor	–	–	–	–	–	–	–	–	–	–	–	–	–	0.27	–	–	–
Mag	3.80	1.56	1.35	1.28	1.43	2.72	2.01	1.34	1.50	1.59	6.35	1.47	1.41	1.53	1.67	1.83	1.94
Il	1.40	0.06	1.38	0.70	0.56	1.04	1.60	0.60	0.80	0.84	3.71	1.30	1.32	1.44	1.90	1.48	1.48
Ap	–	0.08	0.24	0.06	0.16	0.13	0.18	0.10	0.80	0.11	0.36	0.21	0.18	0.27	0.21	0.21	0.21

1 Houston and Others, 1968, P. 51, No. 2, from an orthoamphibolite cutting quartzo-feldspathic Archean gneiss in NE¼, Sec. 3, T.13N, R.82W.
2 SR58-21, amphibolite from a large sill cutting quartzo-feldspathic Archean gneiss; 2770'E, 4224'N from the SW Cor., Sec. 5, T.15N, R.81W.
3 SR58-6, from a large gabbroic intrusion in quartzo-feldspathic Archean gneiss; 1075'E, 4205'N from the SW Cor. Sec. 15, T.15N, R.82W.
4 *SR60-24, from a gabbroic sill cutting the Colberg Metavolcanic Rocks of the Phantom* Lake Suite, 5400'E, 3050'N *from the SW Cor.* Sec. 31, T.18N, R.79W.
5 SR60-52, from a large gabbroic sill cutting the Conical Peak Quartzite of the Phantom Lake Suite; 5100'E, 1610'N from the SW Cor. Sec. 33, T.19N, R.79W.
6 Houston and Others, 1968, P. 51, No. 3, from a large gabbroic sill cutting the Lindsey Quartzite of the Deep Lake Group; 2415'E, 3560'N from the SW Cor. Sec. 15, T.16N, R.80W.
7 SR60-33, from a gabbroic sill cutting the Cascade Quartzite of the Deep Lake Group; 1390'E, 2760'N of the SW Cor. Sec. 13, T.17N, R.80W.
8 SR61-38, from a gabbroic sill cutting the Cascade Quartzite of the Deep Lake Group; 220'E, 200'S from the NW Cor. Sec. 35, T.17N, R.80W.
9 SR59-20, from a gabbroic sill cutting the Magnolia Formation of the Deep Lake Group; 4385'E, 3490'N from the SW Cor. Sec. 23, T.16N, R.81W.
10 GSW59-2, from the same large gabbroic sill as sample 6, cutting the Lindsey Quartzite of the Deep Lake Group, 4385'E, 540'N from the SW Cor. Sec. 10, T.16N, R.80W.
11 Houston and Others, 1968, P. 51, No. 1, from a diabase dike cutting the Medicine Peak Quartzite of the Libby Creek Group, in the NW¼ Sec. 26, T.16N, R.80W.
12 SR77-32, from a gabbroic dike cutting the Sugarloaf Quartzite of the Libby Creek Group; 400'E, 3900'N from the SW Cor. Sec. 19, T.16N, R.79W.
13 SR77-28, *from a gabbroic dike cutting the Sugarloaf Quartzite* of the Libby Creek Group. 380'E, 3200'N from the SW Cor. Sec. 19, T.16N, R.79W.
14 SR77-36, from a gabbroic dike cutting the Nash Formation of the Libby Creek Group; 4600'E, 1090'N from the SW Cor. Sec. 18, T.16N, R.79W.
15 SR77-23, from a small gabbroic dike cutting the Sugarloaf Quartzite of the Libby Creek Group; 3935'E, 2415'N from the SW Cor. Sec. 23, T.16N, R.78W.
16 SW-1, from a large gabbroic sill cutting the Nash Formation of the Libby Creek Group. 3670'E, 1700'N from the SW Cor. Sec. 19, T.15N, R.80W.
17 SW-3, from a gabbroic sill cutting the Medicine Peak Quartzite of the Libby Creek Group; 4115'E, 2240'N from the SW Cor. Sec. 26, T.15N, R.81W.

In summary, we believe that geological and geochemical data suggest the presence of more than one tectonic setting for emplacement of Proterozoic mafic intrusive rocks and, most probably, more than one episode of intrusion. From geologic considerations, and by analogy to other Precambrian terranes in the Wyoming Province, (Stueber and others, 1976) the ages of intrusive events in the Medicine Bow Mountains may well be close

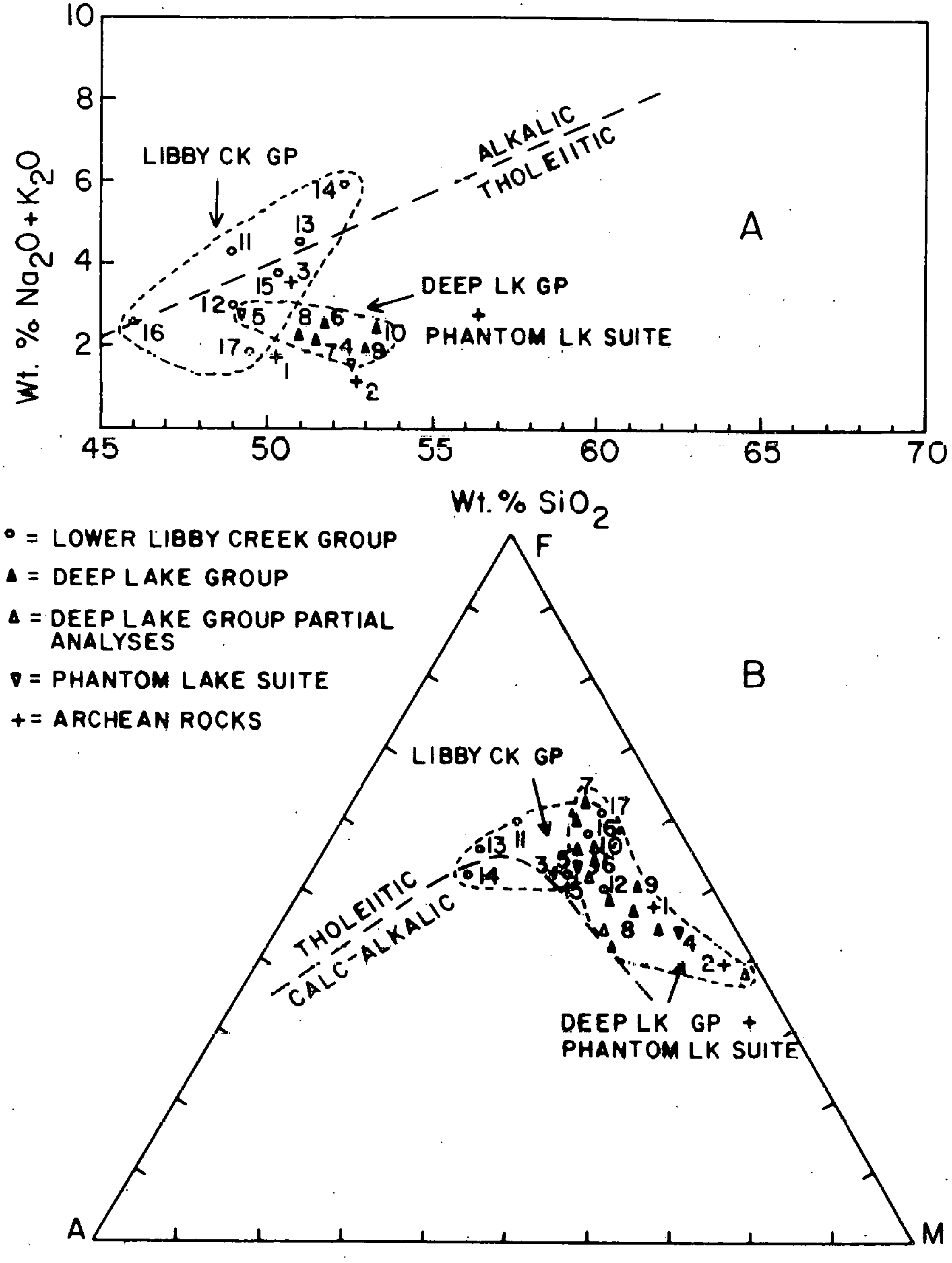


Figure 2.19. Geochemistry of gabbroic intrusives from the Medicine Bow Mountains; numbered points referenced to Table 2.23. A. $Na_2O + K_2O$ versus silica; the alkalic rock-tholeiite boundary is from MacDonald (1968). B. AFM plot; A = wt. % $Na_2O + K_2O$, F = wt. % $Fe_2O_3 + FeO$, M = wt. % Mgo; tholeiitic-calc alkalic rock boundary is from Irvine and Baragar (1971).

to 2500 m.y., 2100-2200 m.y., and 1700 m.y. We are inclined to believe that most of the large sills in the Deep Lake Group and Phantom Lake Suite will prove to be penecontemporaneous with the 2100 m.y. old episode which is also the age of the Blue Draw Metagabbro in the Black Hills (Redden, 1980) and the 2150 m.y. Nipissing Diabase of the Lake Huron area of the Canadian Shield (Fairbairn and others, 1969), whereas many of the mafic bodies cutting the Upper Libby Creek Group and rocks south of the shear zone will prove to be younger -- about 1600-1700 m.y. As discussed in the next section, most of the dikes in the Lower Libby Creek Group may be about the same age as or only slightly younger than the sills in the Deep Lake Group and Phantom Lake Suite. If so, the observed chemical differences might suggest differing composition of source magmas during a protracted rifting event.

Gaps Trondhjemite

The Gaps Trondhjemite (Gaps "Granite" of Houston and others, 1968) is a leucocratic plagioclase-rich granitic rock found at several localities in the Medicine Bow Mountains (Plate 1) and the Sierra Madre. It generally crops out within, or is spatially associated with, gabbroic intrusive bodies which cross-cut the Lower Libby Creek Group in the Medicine Bow Mountains. (It also is believed to cross-cut the upper Deep Lake Group in the Sierra Madre.) Bodies of Gaps Trondhjemite range in size from a few meters to several hundred meters in diameter.

As shown in Table 2.24, Gaps Trondhjemite consists primarily of plagioclas (An_8 to An_{34}) with variable amounts of quartz, muscovite (or sericite), and opaque minerals. Biotite, chlorite, carbonate, and epidote are occasionally present. Several of the samples contain abundant myrmekite. Much of the plagioclase is albite twinned and has undergone partial alteration to sericite. Muscovite and hematite are common in some fractures while carbonate has replaced some of the plagioclase. Uraninite has also been found in fractures in the Gaps Trondhjemite and some samples of fractured rock have yielded 1000 ppm U. Figure 2.20 shows that the Gaps Trondhjemite ranges in composition from tonalite to diorite (following Streckeisen, 1973) but we use the term trondhjemite because most samples are leucocratic quartz diorites without any alkali feldspars, whose plagioclase is oligoclase or albite (Barker and others, 1976).

TABLE 2.24 PETROGRAPHY OF THE GAPS TRONDHJEMITE FROM THE MEDICINE BOW MOUNTAINS AND SIERRA MADRE; MODES FROM POINT COUNTS; SIERRA MADRE DATA FROM SCHUSTER (1972).

	Qtz.	Plag.	Musc.	Biot.	Epid.	Carb.	Opaq.	Chl.	An content of Plag.
MEDICINE BOW MOUNTAINS									
HL3	11.9	80.7	0.4				7.1		
GG	23.9	67.8	5.3				3.0		9
60-85	24.7	53.2	16.1				6.0		
60-85(B)	23.3	46.5	21.4		0.6		8.2		
SR-77-20	5.1	40.2		40.2	Tr	2.6	12.0		
SR-77-29	23.5	66.9	4.4		2.2	2.9			10
SR-77-42	29.3	56.9	10.6				3.2		8
SR-77-48	5.6	49.1	2.8			10.2	2.3		13
SIERRA MADRE									
AFCP-1	20	60	10				8	2	13
AHA-272	5.0	80.6		0.2			8.4	5.8	34
AHA-275	2.8	68.7		1.4			8.5	18.6	33
AHA-300	9.2	42.2					8.2	40.4	28
AHA-302	2.6	46.5				0.1	9.3	41.5	32
Grand Mean	14	61	5	3	Tr	1	6	8	

Chemical analyses of the Gaps Trondhjemite are shown in Table 2.25. Compared to associated gabbros (Table 2.23), the trondhjemite is enriched in silica, sodium and potassium (Figure 2.21) and titanium, and depleted in magnesium (slightly), manganese, and calcium. In spite of these differences, the AFM plot (Figure 2.21) shows that the trondhjemite follows the same iron enrichment trend as the Libby Creek Group gabbros: one which lies close to the line (Irvine and Baragar, 1971) separating tholeiitic from calc-alkalic rocks. This, combined with the intimate spatial association in outcrops, suggests that the Gaps Trondhjemite is genetically related to (felsic differentiates of) at least some of the gabbroic bodies which cross-cut the Libby Creek Group.

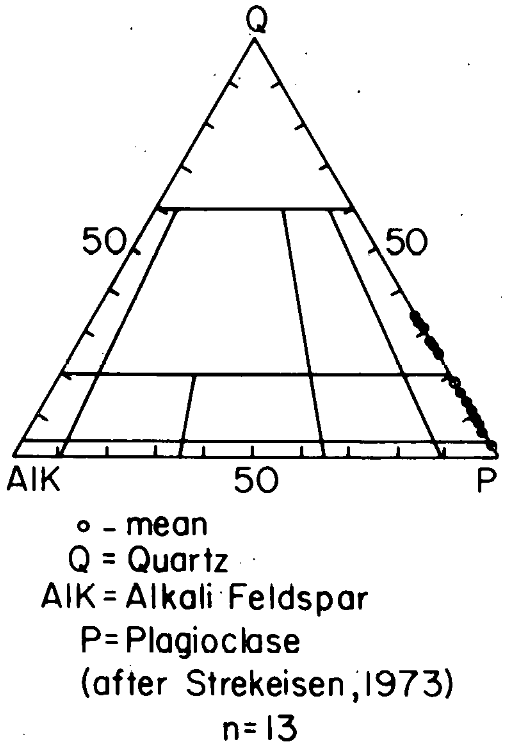


Figure 2.20. QAP ternary diagram for the Gaps Trondhjemite

Attempts to date the trondhjemite have yielded equivocal results. Hills and Houston (1979) reported a date of 1755 m.y. (± 215 m.y.) with an initial ^{87}Sr/^{86}Sr of 0.723. The ^{87}Rb/^{86}Sr ratio ranges from 1.1 to 4.8 (Hills, personal communication, 1980). The initial ^{87}Sr/^{86}Sr ratio is anomalously high and the origin of the excess radiogenic strontium is uncertain (Hills and Houston, 1979). More recent attempts at dating the trondhjemite suggest an age of 2000 m.y. (Hedge, personal communication, 1980).

The new 2000 m.y. date appears to indicate that the Gaps Trondhjemite and at least some of the Libby Creek Group gabbros are significantly older than the inferred 1700-1800 m.y. collision and may have been rift-related intrusions. It is interesting to note, here, that the Gaps Trondhjemite is mineralogically similar to plagiogranite differentiates found in some ophiolites (Coleman and Peterman, 1976). Relative to their associated gabbros, the Gaps Trondhjemite and oceanic plagiogranites are both enriched in silica, sodium, and titanium and depleted in calcium, magnesium, and manganese. However, oceanic plagiogranites also show low Rb/Sr and initial Sr^{87}/Sr^{86}, which the Gaps does not. However, this difference in chemistry might also reflect emplacement of the Gaps Trondhjemite into a thick sedimentary succession.

Perhaps the best explanation of available data is that the sills in the Deep Lake Group and Phantom Lake Suite are 2100-2200 m.y. old tholeiitic intrusions related to early rifting while the Gaps Trondhjemite and some co-genetic gabbros are slightly younger (2000 m.y.), but still rift-related gabbros which were emplaced oceanward and at higher tectonic levels. This explanation suggests that rifting was a protracted event

TABLE 2.25 GEOCHEMISTRY OF THE GAPS TRONDHJEMITE FROM THE MEDICINE BOW MOUNTAINS AND SIERRA MADRE; ANALYSTS: STEVE BOESE (1-3) AND ERIC SCHUSTER (1972; 4-7).

	MEDICINE BOWS		SIERRA MADRE				
	1	2	3	4	5	6	7
	SR-77-20	SR-77-48	AFCP-1	AHA-272	AHA-275	AHA-300	AHA-302
SiO_2	48.5	54.9	64.3	57.71	52.58	58.00	66.39
Al_2O_3	12.0	12.7	10.9	15.27	15.81	13.23	10.26
Fe_2O_3	17.1	3.58	14.84	12.86	16.04	13.20	9.45
MgO	3.65	2.22	0.53	3	5	5	3
CaO	2.93	6.49	0.84	0.27	0.35	0.08	3.83
Na_2O	4.71	7.92	5.44	5.49	4.81	1.00	1.06
K_2O	2.77	0.21	0.63	0.96	1.01	1.24	2.03
TiO_2	4.2	1.3	1.3	2.18	2.38	0.97	0.64
MnO	0.09	0.25	0.02	0.03	0.04	0.07	0.11
P_2O_5	0.82	--	0.25	--	--	--	--
Total	96.77	90.00	99.05	97.8	98.0	93.7	96.8
			MOLECULAR NORMS				
Q	--	--	18.16	9.57	3.31	27.24	34.12
Or	17.35	1.35	3.90	5.85	6.15	8.1	12.95
Ab	44.75	66.28	50.95	50.8	44.45	18.1	10.25
An	3.60	--	3.60	1.40	1.80	0.45	18.55
Hy	0.90	--	18.82	22.16	31.70	32.18	20.22
Di	4.88	24.24	0.34	--	--	--	1.56
Mt	2.45	--	2.10	1.8	2.25	1.98	1.38
Il	6.18	1.96	1.88	3.12	3.42	1.48	0.96
Ap	1.82	--	.26	--	--	--	--
Ol	18.06	--	--	--	--	--	--
Cor	--	--	--	5.29	6.93	10.48	--
Ac	--	1.87	--	--	--	--	--
Ne	--	4.31	--	--	--	--	--

1. SR77-20: Subaluminous biotitic diorite, along Highway 130, NE¼, SW¼, Sec. 25, T16N, R80W
2. SR77-48: Peralkaline diorite, near the Gap, SE¼, SW¼, Sec. 8, T16N, R79W
3. AFCP-1: Subaluminous red trondhjemite, near Copperton, SE¼, NW¼, Sec. 31, T14N, R86W
4. AHA-272: peraluminous gray diorite, SE¼, NE¼, Sec. 26, T14N, R86W
5. AHA-275: peraluminous gray diorite, SE¼, NE¼, Sec. 26, T14N, R86W
6. AHA-300: peraluminous chloritic quartz diorite, NE¼, SE¼, Sec. 30, T14N, R85W
7. AHA-302: metaluminous chloritic quartz diorite, NE¼, SE¼, Sec. 30, T14N, R85W

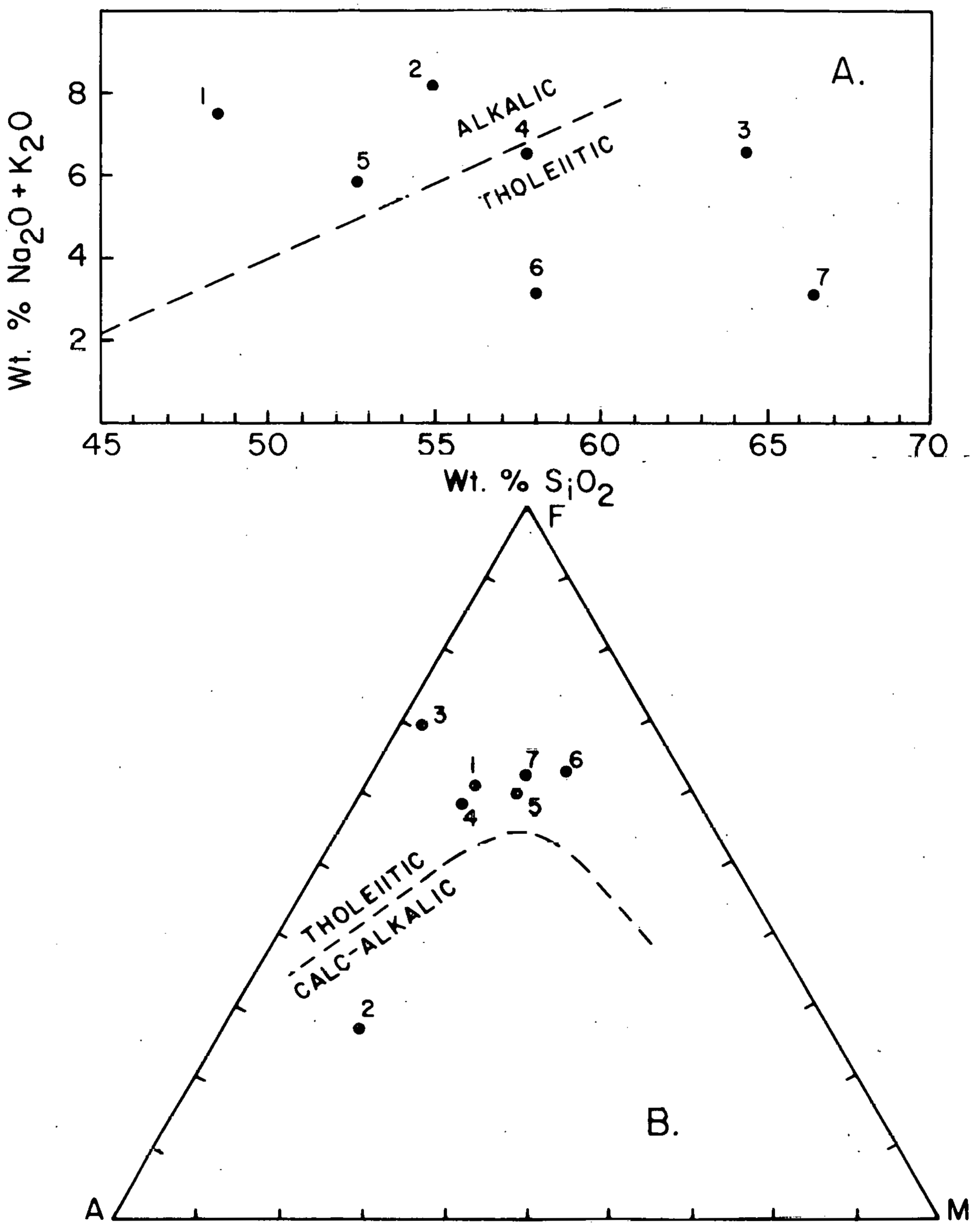


Figure 2.21. Geochemistry of the Gaps Trondhjemite; numbered points referenced to Table 2.25. **A.** Alkali-silica plot; the alkalic rock-tholeiitic boundary is from MacDonald (1968). **B.** AFM plot; A = Na_2O + K_2O; F = Fe_2O_3 + FeO; M = MgO; tholeiitic-calc alkalic boundary is from Irvine and Baragar (1971).

involving either an evolving upper mantle magma source or multiple (heterogeneous) sources. Later, 1700 m.y., gabbroic intrusives must also be present because the shear zone and 1700-1800 m.y. volcanogenic rocks south of the shear zone are cross-cut by mafic bodies.

SUMMARY OF DEPOSITIONAL ENVIRONMENTS

Archean Sedimentation

The oldest rocks within the metasedimentary terrane of the northern Medicine Bow Mountains, the Stage Crossing Gneiss, is an amphibolitic sequence which probably contains metamorphosed mafic flows, intrusions, and tuffaceous sediments in addition to thin quartzites and conglomerates. This unit is still only poorly understood but might be a remnant of a once more extensive greenstone succession in the northern Medicine Bow Mountains. It has many lithologic and structural similarities to the Elmers Rock greenstone belt of the central Laramie Mountains (Graff and others, 1981) and to the Continental Divide Metavolcanic Rocks of the Sierra Madre (Karlstrom and Houston, 1981). Also like these areas, the Stage Crossing Gneiss is complexly intruded by, and infolded with, sills of granitic rocks of presumed Archean age.

The depositional setting for these and other Archean greenstone successions in the Wyoming Province is not known. Obviously, volcanism was widespread and possibly marine deposition predominated as is evidenced by local pillow basalts and marbles in the Sierra Madre and Laramie Mountains.

The Phantom Lake Metamorphic Suite, in the earliest stages of deposition, was also dominated by volcanism and volcaniclastic sedimentation.

Contact relationships with the structurally lower Stage Crossing Gneiss are ambiguous and the two sequences may be closely related in time (or actually gradational) or they may be separated by considerable hiatus. In the Medicine Bow Mountains, we lean slightly towards the former view, that the Stage Crossing Gneiss and overlying Stud Creek Volcaniclastic Rocks were related to the same protracted volcano-sedimentary depositional episode.

The Stud Creek Volcaniclastic Rocks are interpreted to contain predominantly subaerial tuffs and flows with minor fluvial conglomerates and shallow marine clastic carbonates. These rocks are overlain, in the northern Medicine Bow Mountains, by a mappable conglomerate succession, the Rock Mountain Conglomerate, containing small areas of radioactive conglomerate. This unit is only of local extent and probably was being deposited in prograding alluvial fans adjacent to fault scarps in the northern part of the basin at the same time that volcaniclastic rocks of the Stud Creek assemblage were being deposited to the south.

The Bow River Quartzite is a fine-grained subarkosic quartzite which we interpret to be shallow marine sands which transgressed northward across the older units. In places in the northern Medicine Bow Mountains, the lower Bow River Quartzite appears to be gradational with the underlying Rock Mountain Conglomerate suggesting fluvial-influenced marine deposition in the north, in some type of river-fed embayment.

The Colberg Metavolcanic Rocks represents renewed widespread volcanic activity, probably dominantly in a subaerial environment. Volcanism was dominated by tholeiitic basalt and andesite flows accompanied by deposition of andesitic (and volumetrically minor rhyolitic) tuffs, quartzites, and paraconglomerates which represent debris flows in alluvial or submarine channels and fans, possibly adjacent to volcanic highlands. The paraconglomerates contain boulders of granite so the tectonic environment contained a granitic source area.

The Conical Peak Quartzite represents a return to deposition of marine sands except now the source contained appreciable mafic volcanic rock, as suggested by the increase in detrital plagioclase in the quartzite.

Proterozoic Deposition of the Deep Lake Group

Archean sedimentation of the Phantom Lake Suite was followed by intrusion of granitic sills in several areas of the Medicine Bow Mountains accompanied by regional metamorphism about 2500 m.y. ago. This orogenic episode, contemporaneous with the Kenoran Orogeny of the Superior Province, marks the end of the Archean (permobile) tectonic regime (Dewey and Burke, 1973) and the beginning of a tectonic regime more like that operating today, with large, stable continental blocks. The newly stabilized Wyoming craton was uplifted and eroded sometime after 2500 m.y. and provided detritus for deposition of the Deep Lake Group.

The Deep Lake Group consists of three broad sedimentary sequences or cycles separated by disconformities (Houston and Karlstrom, 1979b): a fining-upwards fluvial assemblage (Magnolia and Lindsey Formations); a mixed fluvial and marine succession (Campbell Lake and Cascade Quartzite); and the beginnings of extensive glaciomarine deposition (Vagner Formation) which also is represented in the Lower Libby Creek Group.

The earliest Proterozoic sediments are in the Conglomerate Member of the Magnolia Formation. These rocks contain uranium-bearing quartz-pebble conglomerates and are interpreted by us to represent alluvial fans which developed on fault scarps bounding a braided trunk river system which flowed southwest. The Quartzite Member is thicker and more laterally continuous than the Conglomerate Member and is interpreted to be the braided trunk river deposits. The Lindsey Quartzite gradationally overlies the Magnolia Formation and represents continued, but lower energy, fluvial sedimentation. This entire sequence, Magnolia through Lindsey, is a fluvial fining-upward stratification sequence (Cycle 1 of Karlstrom and Houston, 1979b).

Deposition of paraconglomerates of the Campbell Lake Formation (which are interpreted to be debris flows) represents interruption of the fining-upwards trend. This was probably caused by movement along basin-bounding faults and deposition on alluvial fans. (It might also represent a first, and minor, pulse of glacial or glaciomarine sedimentation.) The Campbell Lake Formation and Cascade Quartzite form a sedimentary package (Cycle 2 of Karlstrom and Houston, 1979b) which unconformably overlies the Lindsey Quartzite in the central Medicine Bow Mountains, the Magnolia Formation to the northeast, and Archean rocks farther northeast and southwest (Plate 1). The Cascade Quartzite is interpreted to include both fluvial and shallow marine deltaic deposits which were laid down in a southwest-northeast trending basin.

The Vagner Formation unconformably overlies the Cascade and is interpreted to be a glaciomarine deposit which appears to be part of a larger glaciomarine sequence that includes the overlying Rock Knoll and Headquarters Formations. On structural grounds, we believe there is a major fault separating the autochthonous Vagner from allochthonous or para-autochthonous Rock Knoll and Headquarters. However, in spite of probable appreciable tectonic transport along this fault, units above and below appear to represent very similar depositional environments. Therefore, the Vagner will be discussed along with overlying units in the next section.

Proterozoic Deposition of the Lower Libby Creek Group

The Lower Libby Creek Group contains two major sedimentary assemblages: a glaciomarine succession (Rock Knoll and Headquarters Formations) and a deltaic succession (Heart Formation, Medicine Peak Quartzite, Lookout Schist, and Sugarloaf Quartzite) which records marine regressions and transgressions across a major delta system. The two are obviously closely related and, in fact, similar deltaic sedimentation in a small ocean basin may go back as far as Cascade deposition in the Deep Lake Group. The ice apparently advanced down the river and delta system which formed the Cascade depositional basin and then advanced and retreated several times during Vagner, Rock Knoll, and Headquarters deposition. Final retreat of the glaciers (ice cap or ice shelf) sometime before or during Heart deposition is probably related to combination of climatic change toward warmer conditions and continued rifting and opening of the small ocean basin in which the Lower Libby Creek Group miogeoclinal sediments were deposited. Isostatic effects due to advance and retreat of glacial ice may have been important controls of marine transgressions and regressions throughout Lower Libby Creek Group deposition but these effects cannot be assessed with present information.

Glaciomarine sedimentation of the Vagner, Rock Knoll and Headquarters Formations was quite complex, and the rapid facies changes indicate highly changeable sedimentary conditions, as might be expected in a glacial environment. The basal paraconglomerate of the Vagner is reasonably continuous and contains dropstones which are interpreted to be debris dropped from floating icebergs. Subsequent units in the Vagner include clearly marine carbonates and phyllites. Rock Knoll deposition of fluvial and deltaic sands is interpreted to be related to glaciomarine processes because it is sandwiched between the Vagner and overlying Headquarters. Headquarters deposition included several paraconglomerates, probably related to advance of glacial ice (Sylvester, 1973), plagioclase-rich arkoses, and laminated silts. Turbidite structures in some paraconglomerates indicate deposition on a slope and Kurtz and Anderson (1979) suggested that the Headquarters Formation had many similarities to continental slope deposits off Antarctica. We prefer a nearer-shore environment because we believe that both the underlying Cascade and overlying Heart are related to deltaic sedimentation. Therefore, we postulate an ice-filled fjord-like environment with continued deltaic sedimentation and prodelta slumping and turbidity, but now influenced by glacial mass-movement processes.

Deltaic sedimentation of the Heart, Medicine Peak, Lookout and Sugarloaf (Figure 2.15) represents marine regression, transgression, and renewed regression. The Heart Formation is interpreted to be a prodelta and delta front deposit; seas regressed prior to deposition of the Medicine Peak Quartzite which represents mainly intertidal delta plain deposits; seas transgressed to deposit the Lookout Schist which is similar to the Heart and is interpreted to be mainly delta front and prodelta deposits; and the Sugarloaf represents another regression and re-establishment of shallow water delta plain sedimentation.

Proterozoic Deposition of the Upper Libby Creek Group

The Upper Libby Creek Group represents a major change from the deltaic sedimentation in the Lower Libby Creek Group. Again, we infer a major thrust fault between the two sequences (Plate 1) so we don't know how much time or distance separated the two. The lower unit in the Upper Libby Creek Gruop is the Nash Fork Formation, a dolomite-phyllite unit of undoubted marine origin. This is followed by the Towner Greenstone, which may be a series of submarine flows (although no pillows have been found), and the French Slate, which was a deep water, black shale unit. Thin and discontinuous iron formations crop out in both the Nash and the French Slate.

We interpret the Upper Libby Creek Group to represent open marine sedimentation. The Nash Fork Formation is believed to be intertidal because of the high clastic quartz content and the presence of stromatolites. The French Slate is interpreted to represent off-shore, deep water clastic sedimentation. In general, the Upper Libby Creek Group appears to represent continued, but slower, subsidence of the continental shelf following rifting (Dickinson, 1974). The appearance of volcanic (?) rocks of the Towner Greenstone and deposition of black shales may be related to island arc volcanism to the south during Lower Libby Creek Group deposition but this is conjectural.

MODELS FOR DEPOSITION OF METASEDIMENTARY ROCKS IN THE MEDICINE BOW MOUNTAINS

In going from descriptions of mapped units and sedimentary features to discussion of depositional environments, we made a large interpretive step. This section takes the even larger, and wobblier step of proposing models that try to unify our stratigraphic and sedimentological observations and interpretations into a coherent picture of early Precambrian sedimentation. We emphasize two aspects of early Precambrian history.

1. Several aspects of the sedimentary and tectonic record in southern Wyoming suggest cratonic margin sedimentation and plate tectonics during the Early Proterozoic. The best lines of evidence are that Early Proterozoic Libby Creek Group metasediments have many similarities to miogeoclinal, passive plate margin successions and that these miogeoclinal sediments are juxtaposed against highly deformed late Early Proterozoic eugeoclinal rocks and synorogenic intrusives across a major shear zone. This geologic discontinuity, plus profound geochronologic discontinuity, across this shear zone suggest craton-island arc collisional orogenesis about 1700 m.y. ago (discussed later; also see Hills and Houston, 1979; Houston and Karlstrom, 1980). Working backwards from the inferred collision, and from sedimentary evidence, the metasedimentary succession is interpreted to represent progressively deeper water sedimentation on a passive, trailing continental edge. Paleocurrents indicate that fluvial and deltaic sedimentation took place on a persistent southwest paleoslope, which is parallel to the present margin and, inferencially, the Proterozoic margin of the Wyoming Province. Paleocurrents parallel to

the cratonic margin appear to require a southern highland bounding Early Proterozoic miogeoclinal sedimentation. By this line of reasoning, and the presence of thick tholeiitic sills in the Deep Lake Group, we postulate intracratonic rifting and subsidence of the continental shelf as the tectonic setting for Deep Lake and Libby Creek Group deposition.

2. Early Precambrian sedimentation the world over appears to record important evolutionary trends in tectonic regimes, atmospheric and hydrospheric composition, and biologic activity. Three trends are important to this discussion. First, about 2500 m.y. ago there was a change from permobile Archean tectonics, characterized by abundant volcanism and widely distributed deformations, to an Early Proterozoic tectonic regime characterized by larger, stable cratonic blocks and concentration of deformation along cratonic margins. Second, there was a change sometime during the Early Proterozoic from anoxygenic atmosphere and hydrosphere, in which detrital uraninite and pyrite were stable and transportable, to more oxygenated surface conditions under which these heavy minerals could not be transported and concentrated. Third, large volumes of iron formation accumulated during the late Early Proterozoic on a scale not seen before on earth. The appearance of these Proterozoic iron formations is generally believed to be related to atmospheric and/or biospheric evolution. In addition to these world-wide evolutionary trends, many Early Proterozoic metasedimentary successions in North America, including the Medicine Bow Mountain succession, record evidence of glacial activity, perhaps due to a continental-scale glaciation. We propose that the world-wide evolutionary character of Early Precambrian sedimentation and tectonics combined with continent-wide climatic and

tectonic events in North America created an "idealized stratigraphy" for Late Archean and Early Proterozoic metasedimentary successions throughout North America which involved sequencial deposition of several key lithologies: volcano-sedimentary units, uranium-bearing conglomerate, tillite, aluminous quartzite, stromatolitic dolomite, and iron formation. These lithologies, in approximately this order, are recognizable in many places in North America, especially along the southern margin of what we consider to be a large proto-North American continent, and provide means for lithostratigraphic correlations of Early Proterozoic metasedimentary rocks in North America.

PLATE TECTONIC MODEL FOR LATE ARCHEAN AND EARLY PROTEROZOIC SEDIMENTATION IN SOUTHEASTERN WYOMING

This section develops a plate tectonic model for deposition of metasedimentary rocks of the northern Medicine Bow Mountains and Sierra Madre. If the orogenesis 1700 m.y. ago involved cratonic margin-island arc collision (discussed later), what was the nature of the cratonic margin and what was the pre-collision history of that margin?

Plate margin sedimentation in southeastern Wyoming may go back as far as Late Archean deposition of the Phantom Lake Metamorphic Suite, although this is still conjectural. The Phantom Lake Suite as a unit is somewhat anomalous with respect to most other Late Archean supracrustal successions in the Wyoming Province in that it contains a significant percentage of mature siliciclastic sediments, including radioactive conglomerates and thick marine quartzites, instead of being dominated by volcanogenic rocks. The shape of the Phantom Lake Suite depositional basin is not known but very similar rocks are found in the Medicine Bow Mountains and Sierra Madre suggesting that the basin had a minimum area something like is shown in Figure 2.22A. Proximal, alluvial fan sediments, containing radioactive conglomerates are found in the northeast Medicine Bow Mountains and northwest Sierra Madre and these are interpreted to be basin edge deposits related to faulting along basin-bounding highlands. A few isolated outcrops of Phantom Lake Suite rocks crop out in the gneissic terrane between the Medicine Bow and Sierra Madre depositional centers so we show a continuous basin in Figure 2.22A connecting the two ranges. Paleocurrent data from the Medicine Bow Mountains show bimodal, northeast-southwest directed paleocurrents which we interpret to be ebb and flood tides on a southwest paleoslope in a dominantly marine setting. Paleocurrent data from the Sierra Madre are sparse but the fluvial conglomerates appear to have been derived from the north.

Figure 2.22A also shows that the present southern boundary of the Wyoming Province has three other outcrop areas of Archean supracrustals along it: the Elmer's Rock greenstone belt of the Laramie Range (Graff and others, 1981), the Walen Group of the Hartville Uplift (Snyder, 1980), and un-named, probable Archean schists in the Black Hills (Redden, 1980). Existing geochronologic data indicate that these successions are all Late Archean (older than about 2600 m.y.) although they are not necessarily the same age. The Hartville succession contains thick carbonates interbedded with volcanics and, like the Phantom Lake Suite, is not a typical greenstone succession. The Elmer's Rock succession, however, is a greenstone belt and we are faced with the possibility of broadly contemporaneous deposition of marine clastics (Phantom Lake Suite), volcanics

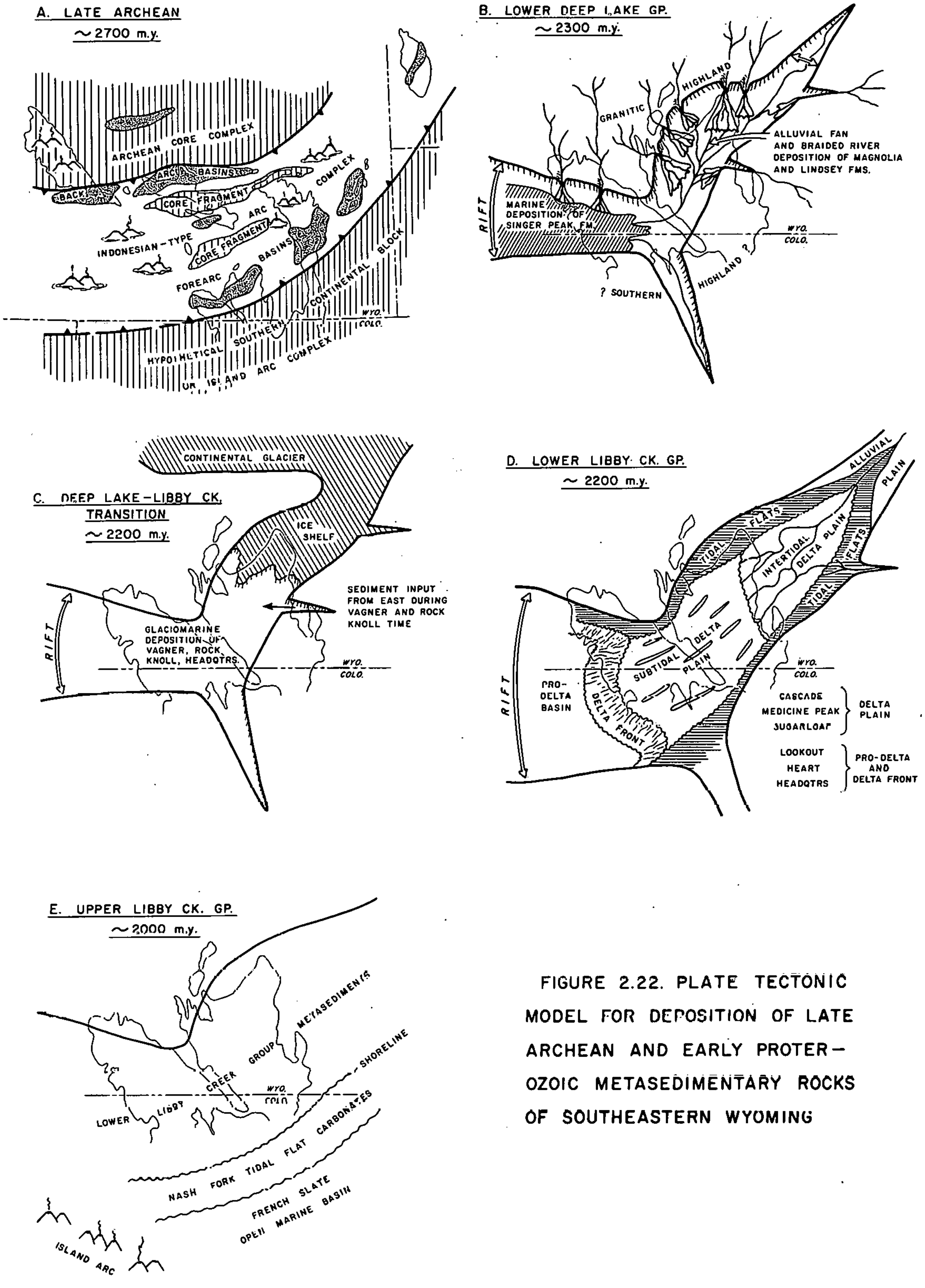


FIGURE 2.22. PLATE TECTONIC MODEL FOR DEPOSITION OF LATE ARCHEAN AND EARLY PROTEROZOIC METASEDIMENTARY ROCKS OF SOUTHEASTERN WYOMING

(Elmer's Rock), and carbonates (Hartville) within a relatively small area. One possibility for such rapid change in tectonic and sedimentary regimes is a fore-arc setting, which can contain fairly small depositional basins and diverse tectonic settings (Hamilton, 1979). This possibility is presented in Figure 2.22A simply as food for thought. The basins might, equally well, be back-arc basins but we do think there are several attractions to a model which postulates deposition of southeastern Wyoming, Late Archean supracrustals south of a composite arc system. This model provides some explanation for the observed age differences between a core area and marginal areas of the Wyoming Province; it provides a mechanism for formation of Late Archean uranium-rich granites of the Granite and northern Laramie Mountains by partial melting of older fragments of Archean felsic crust; it helps explain the wide diversity of Archean styles of sedimentation in the Wyoming Province, some representing back-arc or intra-arc volcano-sedimentary basins, others representing fore-arc or cratonic margin basins; it provides a mechanism for pervasive Late Archean deformation, related to closing of small basins during arc-arc and craton-arc collisions; and it is philosophically satisfying to postulate that the later, Early Proterozoic, plate margin processes of rifting followed by collisional orogenesis took place along or parallel to older zones of crustal weakness.

A hypothetical southern continental block or Archean arc complex south of the Late Archean basins of southwestern Wyoming is shown in Figure 2.22A for several reasons. Primarily, it provides a block to rift away during our postulated Early Proterozoic rifting (although a small fragment of continental crust analogous to Baja California would do as well). Also, an Archean arc provides an explanation for nappe-like Archean folds in Phantom Lake Suite rocks and underlying "basement" gneisses which have axial planes that dip steeply to the north and could have formed during an arc-arc collision across a north-dipping subduction zone in the Late Archean.

Deep Lake Group sedimentation represents an appreciable change in tectonic and sedimentary regimes from the Phantom Lake Suite. The basal unit, the Magnolia Formation, is interpreted by us to be fluvial sediments, both alluvial fans and braided river deposits, which were deposited unconformably on Late Archean granites and folded Phantom Lake Suite rocks. The hiatus represented by the angular unconformity is of unknown duration. Fluvial sedimentation appears to have continued through at least the upper Lindsey Quartzite and probably some of the Cascade is also fluvial. As cartooned in Figure 2.22B, we postulate an intracratonic fault-bounded basin, probably an incipient rift system, for Magnolia deposition. By Lindsey time, the western (Sierra Madre) part of the basin appears to have opened enough to permit marine deposition of the correlative Singer Peak Formation. The aulacogen shown in the southern block, like the southern block itself, is strictly hypothetical.

The rifting story during upper Deep Lake Group and lowest Libby Creek Group deposition is complicated by advance of continental-scale glaciers down the Medicine Bow sedimentary basin and perhaps across the newly forming continental edge (Figure 2.22C). Vagner through Headquarters deposition appear to be related to advance and retreat of this ice sheet. One interesting aspect of this succession is the west-directed paleocurrents of the Vagner and Rock Knoll Formations which are our only direct evidence for the existence of the southeastern rift block. These paleocurrents plus the increase of plagioclase in the sediments, and decrease in K-spar relative to underlying rocks of the Deep Lake Group suggests that a mafic to intermediate volcanic or plutonic highland to the east was a principal source of detritus during glacial deposition.

Retreat of glacial ice in upper Headquarters time re-established a major delta system similar to one which may have existed earlier, in Cascade time. This delta was fed from the northeast and persisted throughout Lower Libby Creek Group deposition (Figure 2.22D and 2.15). It is these rocks which lend most support to the rifting model. This thick, mature, deltaic succession was deposited by paleocurrents directed southwest (and northeast) parallel to older fluvial paleocurrents and parallel to the continental margin as inferred from the major shear zone bounding the Wyoming Province. Similar deltaic sedimentation parallel to continental margins is believed to have taken place in the early stages of rifting of South America and Africa (Burke, 1976, mentions deltas formed in the Gabon and Cuanza grabens, which became "successful" rift arms, leaving the Benue Trough as a failed arm). Deltaic sedimentation parallel to a continental margin is also taking place in the Gulf of California today (Meckel, 1975). Either of these appears to us to be a reasonable analog for Lower Libby Creek Group sedimentation and we stress that we have no idea whether the southern block was of continental or micro-continental proportions.

As shown in Figure 2.22E, by Upper Libby Creek Group time, rifting had progressed to the open ocean stage, and deltaic sedimentation had ceased in southeastern Wyoming. Nash Fork clastic carbonates and shales are interpreted to be intertidal sediments deposited on Lower Libby Creek Group deltaic sands. The French Slate is interpreted to be open marine basin shales. Island arc sedimentation may have been taking place to the south at about the same time as, or slightly later than, Upper Libby Creek Group deposition and it was these island arcs that, we believe, collided with southeastern Wyoming about 1700 m.y. ago.

EVOLUTION OF EARLY PRECAMBRIAN SEDIMENTATION IN NORTH AMERICA

In considering the record of over a billion years of Late Archean and Early Proterozoic sedimentation in southern Wyoming, it is clear that sedimentary and tectonic regimes in southern Wyoming had important similarities to those which existed elsewhere in North America. Archean sedimentation of the Phantom Lake Metamorphic Suite was characterized by an abundance of tectonism, volcanic activity, and volcano-clastic sedimentation. Earliest Proterozoic sedimentation in the Deep Lake Group was continental, dominantly fluvial, and was characterized by anomalously radioactive quartz-pebble conglomerates containing detrital pyrite, and uranium and thorium minerals. The upper Deep Lake and lowest Libby Creek Groups record evidence of a period of glaciomarine sedimentation which marks the transition from fluvial deposition in the Deep Lake Group to miogeoclinal deposition of the Lower Libby Creek Group and which appears to be approximately contemporaneous with the change from anoxygenic environments, evidenced by detrital pyrite and uranium minerals, to oxygenic environments, evidenced by detrital hematite and the absence of radioactive conglomerates. The Upper Libby Creek Group represents deeper water miogeoclinal sedimentation and, perhaps, the earliest evidence of volcanic activity associated with contemporaneous or slightly younger island arc sedimentation to the south.

Evolutionary changes such as these in tectonic regimes, climatic conditions, and the chemistry of the atmosphere and hydrosphere, as recorded in key lithologies, forms the basis for an attempt to correlate the Late Archean and Early Proterozoic metasedimentary rocks of the Medicine Bow Mountains with other Late Archean and Early Proterozoic sequences in the Wyoming Province and elsewhere in North America. This correlation, shown in Figure 2.23, is modified slightly from one we presented earlier (Houston and Karlstrom, 1980) and is presented again because we feel it has profound implications for the early history of North America and for models of formation of Early Proterozoic-type uranium-bearing fossil placers.

We divide the Late Archean and Early Proterozoic metasedimentary successions in North America into four broad groups: 1. The volcano-sedimentary sequence includes Late Archean supracrustal successions which contain an appreciable percentage of non-volcanic rocks such as quartzite, radioactive quartz-pebble conglomerate and carbonate interbedded with the volcanic and volcaniclastic rocks. Some of these successions have many similarities to typical greenstone belt successions but most appear to represent a transition between true greenstone belts and Proterozoic-type clastic successions.

2. The radioactive conglomerate-tillite sequence is Early Proterozoic and is characterized by fluvial uranium-bearing quartz-pebble conglomerates and presumed glacial deposits. The validity of these lithologies for lithostratigraphic correlation is twofold. First, uranium-

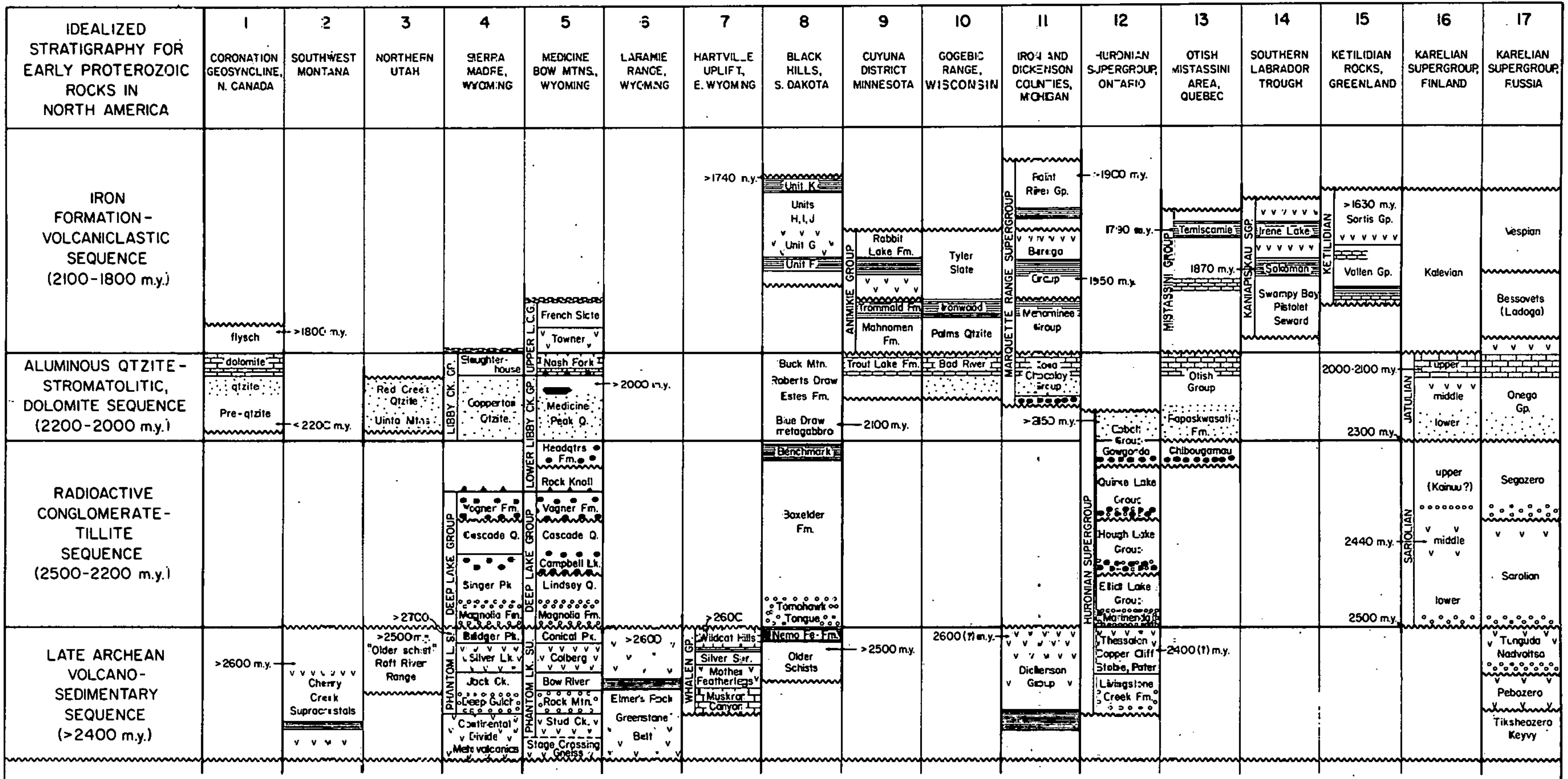


Figure 2.23. Regional lithostratigraphic correlation and idealized stratigraphy of Late Archean and Early Proterozoic supracrustals near the margins of proto-North America. Key lithologies are shown: vvvv = volcanogenic rocks; ⁘ = radioactive quartz-pebble conglomerate; ●•● = diamictite (tillite?) ⁘ = aluminum-rich quartzite; ▤ = stromatolitic dolomite; ☰ = iron formation. References: 1. Hoffman and others (1974) 2. Vitaliano and others (1979) 3. Graff and others (1980) Compton (1975) 4. Graff (1978; 1979) 5. Karlstrom and Houston (1979a; 1979b). Lanthier (1979) 6. Graff and others (1981) 7. Snyder (1980) 8. Redden (1980); Kleinkopf and Redden (1975); Bayley (1970; 1972) 9. Marsden (1972) 10. Sims (1976) 11. James (1958); James and others (1961) 12. Robertson (1976); Roscoe (1969) 13. Roscoe (1973); Chown and Caty (1973); Frye (1972) 14. Dimroth (1970); Fryer (1971) 15. Bondeson (1970) 16. Meriläinen (1980); Simonen (1960) 17. Salop (1977).

bearing conglomerates reflect deposition in anoxygenic surface environments (generally believed to exist on earth before 2200 m.y.; Roscoe, 1973) and deposition in intracratonic fluvial environments (rift valleys?) draining uranium-rich granitic source terranes which, we believe, became widespread only after 2500 m.y. Second, Early Proterozoic glacial rocks all over North America may represent a continental glaciation (Young, 1969; 1973) and, if so, are contemporaneous.

3. The aluminous quartzite-stromatolitic dolomite sequence is characterized by these lithologies in miogeoclinal successions which tend to be thicker and more continuous than units of the underlying sequences. These units contain hematite instead of pyrite suggesting more oxygenated surface conditions such as existed after about 2200 m.y. (Roscoe, 1973) and they formed in shallow marine miogeoclinal sedimentary environments.

4. The iron formation-volcaniclastic succession contains mainly eugeoclinal clastic sediments plus major Proterozoic banded iron formations. The restriction of major banded iron formations to the Early Proterozoic has long been recognized (Cloud, 1968; James, 1966,1969; Veizer, 1976) and generally is ascribed to atmospheric and hydrospheric evolution (Cloud, 1968; Eugster, 1969; Drever, 1974). In North America, banded iron formations also may be related to the presence, between 2100 and 1800 m.y., of small rift-generated ocean basins favorable for iron formation deposition.

The lithostratigraphic correlation of Early Proterozoic sequences shown in Figure 2.23 has important implications with respect to the development of North America. We emphasize that this correlation is strictly valid only for North America (if it is valid at all) because we are combining regional (tectonic and climatic) effects with worldwide (atmospheric, biospheric, hydrospheric and tectonic) evolutionary trends. Figure 2.24 shows the known extent of Archean rocks in North America and this nucleus of the continent is referred to as proto-North America in the following discussion.

It is interesting to note that localities of the radioactive conglomerate-tillite sequence of Figure 2.23 crop out in several places along the south margin of this nucleus (Figure 2.24), suggesting that the Wyoming Province and Superior Province were part of a larger, semi coherent cratonic mass shortly following the end of the Archean, and that platform and miogeoclinal clastic rocks were being deposited on a continental trailing edge somewhat analogous to the Atlantic coast of North America today. The remarkable correlation of formations between the Deep Lake Group and the Huronian Supergroup (Young, 1973; Houston and others, 1977; Graff, 1979; Houston and Karlstrom, 1980), can be understood in this context; it would be similar to comparing and correlating post-Permian sections along the rifted east coast of North America. In fact, we envision a rift origin for the south margin of proto-North America sometime about 2200-2100 m.y. ago which was preceded by fluvial deposition of radioactive conglomerates in small, intracratonic, graben-like rift-valley basins. The glacial sequences near the top of the Huronian and Deep Lake sections may actually be time-equivalent sections deposited during a continental-wide glacial episode (Young, 1973). This is evidence that the two areas experienced similar climatic histories in the Early Proterozoic which also supports the hypothesis that the Wyoming and Superior Provinces were part of a larger, semi-coherent continental block at the end of the Archean. This idea is also supported by available paleomagnetic data which shows a similar apparent polar wandering path for the Superior and Wyoming Provinces (Cavenaugh and Seyfert, 1977; Irving, 1979).

The aluminous quartzite-stromatolitic dolomite sequence is more widely distributed than the radioactive conglomerate-tillite sequence. It is found in isolated areas near the southern margin of proto-North America, from Wyoming to Greenland (and in Finland and Soviet Karelia), and in the Coronation Geosyncline along the western margin. If this correlation is valid, it suggests that both the southern and western margins of proto-North America were trailing edges of a large proto-continent between 2300 and 2000 m.y. Both the thicknesses (southeastern Wyoming--6-7 km; mid-continent, adding Huronian and Lower Marquette Range--8 km; Karelia--2.4 km; Coronation geosyncline--3-5 km) and lithologies in this succession appear to be comparable with the clastic wedge and carbonate/shale assemblages found in more modern miogeoclines (King, 1969; Dickinson, 1974) and we suggest that rifting processes were much the same in the Early Proterozoic and in Phanerozoic time. The aluminous quartzites and associated clastics appear to represent rapid subsidence of the continental shelf during Dickinson's (1974) narrow ocean stage of rifting due to thermal contraction (Bott, 1976); carbonates and shale successions represent open ocean environments and much slower shelf subsidence, perhaps due to sediment loading.

There are some difficulties with this interpretation: one pertains to the mid-continent region along the south margin of proto-North America; a second pertains to the correlation along the western margin of

EXPLANATION FOR FIGURE 2.24

▧	Grenville and Keweenawan rocks (~1.0 b.y.)
▥	Early and Middle Proterozoic eugeoclinal rocks
■	Early Proterozoic miogeoclinal rocks
░	Archean terranes rejuvenated during Hudsonian orogeny
▒	Archean rocks
········	Positive gravity anomaly
——— - -	Known or inferred boundaries of geologic provinces
⊥⊥⊥⊥	Electrical conductivity anomaly
2.5-3.2	Range of radiometric ages in billions of years
③	Numbered localities in Figure 2.23

LABELED LOCALITIES

BI - Belcher Islands; BH - Black Hills; CG - Coronation Geosyncline; CS - Cape Smith Foldbelt; FR - Front Range; HG - Hurwitz Group and Montgomery Lake sediments; HS - Huronian Supergroup; K - Ketilidian Supracrustals; KR - Keweenawan Rift basalts and sediments; LR - Laramie Range; LT - Labrador Trough; MB - Medicine Bow Mountains; MG - Mistassini Group; MR - Marquette Range Supergroup; NACP - North Atlantic Central Plains conductivity anomaly; NF - Nelson Front; OM - Otish Mountains; S - Shuswap Complex; SL - Sakami Lake; SM - Sierra Madre; U - Umanek area; WL - Wollaston Lake foldbelt.

REFERENCES

Geology - King (1979, 1976); Bell (1970); Bondeson (1970); Bridgwater and others (1973); Condies (1976); Dimroth (1972); Henderson and Pultertaft (1967); Hoffman (1973); Houston and others (1968); Kleinkopf and Redden (1975); Money and others (1970); Roscoe (1969).
Geophysics - Alabi and others (1975); Gibb and Thomas (1977); Horner and Hasegawa (1978); Kent and Simpson (1973); Kreary (1976); Lidiak (1971); Thomas and Gibb (1977).
Geochronology - Duncan (1978); Goldich and Hedge (1974); Goldich and others (1966); Hurst and others (1975); Hills and others (1968); Hills and Armstrong (1974); Hills and Houston (1979); King (1976); Moorbath and others (1972); Peterman and Hedge (1968); Price and Douglas (1972); Van Schmus (1976).

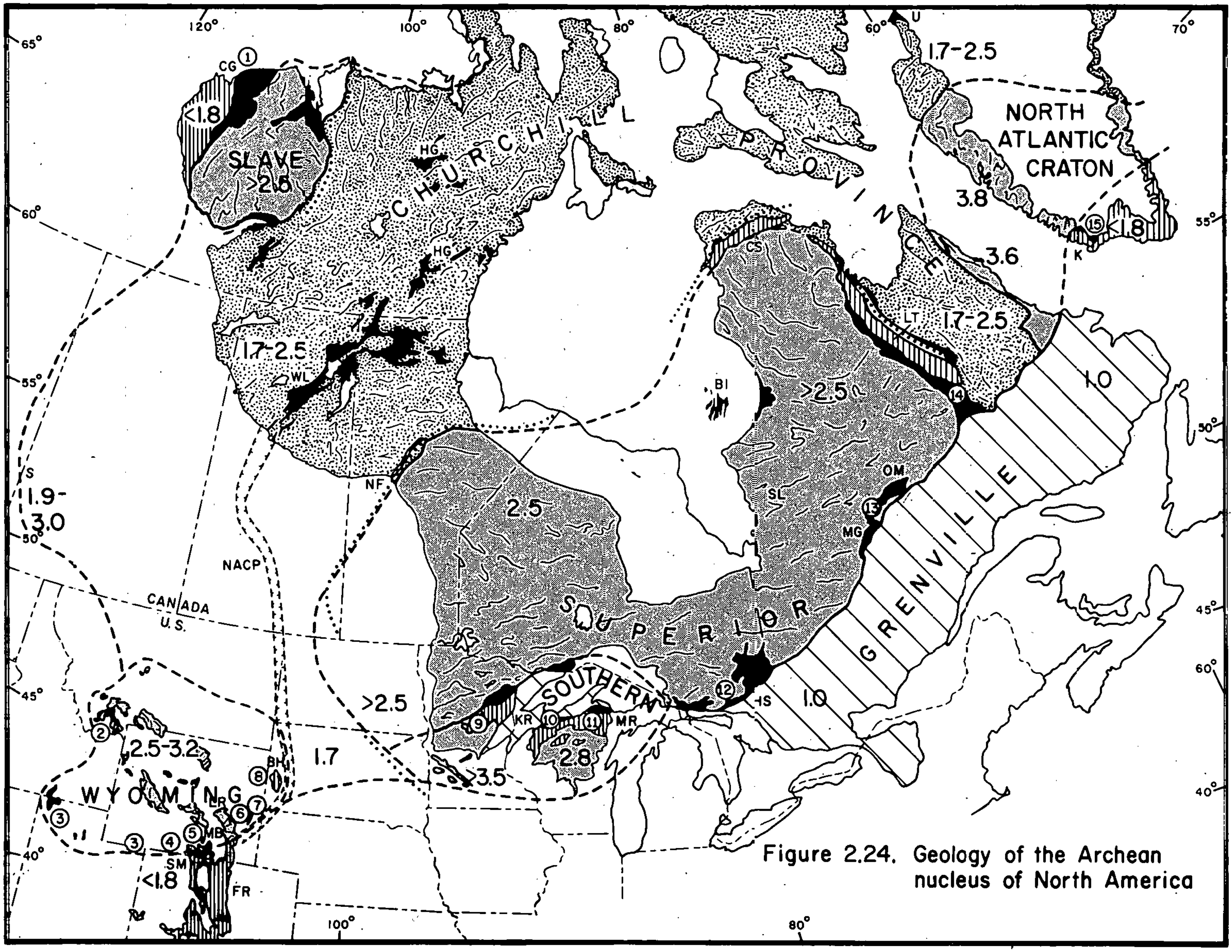


Figure 2.24. Geology of the Archean nucleus of North America

our proto-North America.

The tectonic settings for deposition of the Huronian Supergroup of Ontario and the Marquette Range Supergroup and equivalent strata of Minnesota and Michigan have received considerable attention in recent years. Sims (1976), Morey and Sims (1978), and Sims and others (1980) have maintained that the Early Proterozoic metasediments of Minnesota and Michigan were deposited in an intracratonic setting along a boundary (the Great Lakes tectonic zone of Sims and others, 1980) between Archean granite-greenstone terrane to the north and Archean gneissic terrane to the south. They point out that deposition in Early Proterozoic basins in both the Lake Huron and Lake Superior areas was related to extensional tectonics along and parallel to this boundary but they find no evidence that rifting progressed to the point of development of oceanic crust in either area. Further, they view compressional tectonics during the Penokean Orogeny (1850-1900 m.y.) in terms of thermal contraction of the continental block, not plate tectonics.

Other workers (Van Schmus, 1976; Cambray, 1977, 1978; and Larue and Sloss, 1980) have advocated plate tectonic models for both Early Proterozoic deposition in the Lake Superior region and Penokean deformation. Cambray (1977, 1978) and Larue and Sloss (1980) have argued, from sedimentary evidence, that the Marquette Range Supergroup and correlative rocks were deposited in several sedimentary basins on a rifted, south margin of proto-North America and were deformed during the Penokean Orogeny by collision of the Archean block to the south (central Minnesota and Wisconsin) with proto-North America across a south dipping subduction zone.

We believe that the argument of intracratonic deposition of the Huronian and Lake Superior metasedimentary rocks is not a sufficient argument against plate tectonic processes, nor is the absence of demonstrable oceanic crust. Continental rifts, after all, are intracratonic to start with and oceanic crust is consumed during collisional orogenesis. Our view is that the tectonic settings for both the Huronian and Lake Superior metasedimentary rocks are well explained by plate tectonic models. The Huronian Supergroup appears to represent an area along the south margin of proto-North America where rifting proceeded in a relatively uncomplicated way, similar to along the south boundary of the Wyoming Province in southern Wyoming. Early deposition of fluvial rocks, including radioactive conglomerates, was in a small, fault-bounded, intracratonic rift-valley; later deposition of the Cobalt Group was in a miogeoclinal setting as the rift proceeded to narrow ocean stages. The trend of the main rift-margin may have passed from the Huronian area, south of the Archean gneissic terrane of central Minnesota and Wisconsin.

The Lake Superior metasedimentary rocks are slightly younger than the Huronian Supergroup and appear to represent deposition in a series of rift-related basins (either intracratonic or narrow ocean basins) formed during diachronic rifting of a small continental fragment containing the gneissic terrane away from the main part of proto-North America. In our view, both the Huronian and Lake Superior metasedimentary rocks are an expression of a protracted extensional event which took place between about 2400 and 2000 m.y. ago, all along the southern areas of proto-North America and the different depositional styles in the various basins reflects differences in the way rifting proceeded in different areas along the continental margin. Likewise, we view the Penokean Orogeny in terms of plate tectonic models, as an expression of collisional orogenesis along the south margin of proto-North America and closing of these rift-related basins. (This is discussed in more detail later.)

A second difficulty with our interpretation that proto-North America was much the same shape in the Early Protoerozoic as now, stems from the fact that some workers (Burke and others, 1976; Seyfert and Cavanaugh, 1978; Gibb and Thomas, 1977) have argued that the Slave Province was separate from the Superior Province until they collided about 1750 m.y. ago. If this is so, the correlation suggested in Figure 2.23 would be fortuitous. However, such an interpretation is open to question. The most recent interpretations of paleomagnetic data (Irving and McGlynn, 1976; Roy and LaPointe, 1976, Irving, 1979) show that all existing data can be reconciled with a single apparent polar wandering path for proto-North America. Furthermore, Irving (1979) suggested that the great lengths of Proterozoic polar wandering tracks implied unimpeded drift trajectories and inferencially, large cratons in the Early Proterozoic. However, it should be emphasized that the hypothesis of a single semi-coherent proto-North America does not preclude minor relative motions between cratons (Irving, 1979; Briden, 1976).

The distribution of the iron formation-volcaniclastic sequence, shown in Figures 2.23 and 2.24, has several interesting implications. First, in the Labrador Trough and the Black Hills, these rocks occupy fold belts with structural grains at high angles to the proposed southern margin of proto-North America. These areas may well represent rifting and deposition of volcanics and iron formation in a Red Sea-type environment. These are interpreted by us to be aulacogens which opened enough to produce at least some oceanic crust. A similar interpretation can be put forth for the mid-continent metasediments and iron formations except there, incipient rifting was sub-parallel to the margin of the continent--in approximately the same position as Keewenawen incipient rifting in the late Precambrian.

As stated earlier, this interpretation (see also Larue and Sloss, 1980: Cambray, 1977, 1978) suggests that Early Proterozoic rifting was complex and involved formation of rift-related basins at some distance behind the true south margin of Archean North America.

In summary, from the Early Proterozoic sedimentary record we suggest the following: 1. The Wyoming Province and Superior Province were in roughly their present relative positions by the end of the Archean. The Slave Province and North Atlantic Craton may also have been part of this protocontinent but their tectonic histories are more open to question.

2. Early Proterozoic platform and miogeoclinal sedimentation of radioactive conglomerates, quartzites, tillites, and stromatolitic dolomites and associated rocks, took place 2500 and 2000 m.y. on the rifted south margin of proto-North America, in environments which reflect stages of rifting. Early depositon of fluvial radioactive conglomerates reflects fault-bounded intracratonic rift-valley basins; clastic wedge sediments reflect rapid subsidence of the continental shelf in narrow oceans; carbonates and shales represent open ocean conditions.

3. Volcanic, volcaniclastic, and iron formation sedimentation toward the end of the Early Proterozoic (2100-1800 m.y.) records a period of incipient rifting within proto-North America, with formation of small ocean basins.

4. The Hudsonian and related orogenies reflects compressional, convergent margin tectonics on three sides of proto-North America. Island arc collisions have been proposed for the Coronation Geosyncline (Hoffman, 1973), in southern Wyoming and Colorado (Hills and Houston, 1979; and next section), and in the Ketilidian Mobile belt of Greenland (Bridgwater and others, 1973). A similar story of collisional tectonism has been proposed for the Penokean foldbelt of the mid-continent which involved a small Archean microplate (central Minnesota and Wisconsin) caught up in a convergent zone which involved south-dipping subduction along the north margin of the microplate (Cambray, 1977, 1978). These collisional orogenies all around proto-North America are either the cause or a manifestation of a general compressive regime for proto-North America which caused closing of incipient rifts (e.g. Wollaston Lake foldbelt, mid-continent sedimentary basins) and aulacogens (e.g. Labrador Trough, Black Hills, Athapuscow and Bathurst aulocogens) and jostling of the more stable Archean blocks. We envision the Penokean deformation in the mid-continent region, and deformation in the Black Hills and Labrador Trough, to represent this type of closing of incipient ocean basins and we view the Hudsonian orogeny of the Churchill Province in terms of jostling of cratonic blocks in a general compressive system caused by island arc collisions around the margins of the proto-continent. This is not drastically different from Burke and Dewey's (1973) basement reactivation hypothesis except we do not envision large movements between cratons and suggest that the driving force for the Hudsonian orogeny was related to the island arc-continent collisions causing closing of incipient rifts and jostling of cratonic blocks within an already semi-coherent lithospheric plate.

TECTONIC SETTING OF METASEDIMENTARY ROCKS

Detailed study of the structural geology of metasedimentary rocks in the Medicine Bow Mountains was undertaken for several reasons. First, in a complex geologic terrane such as this, understanding of stratigraphy and structure are mutually dependent, each on the other, and geologic mapping demands that both be investigated simultaneously. Second, knowledge of fold geometries is prerequisite to paleocurrent analysis and provenance studies because bedding and crossbedding must be restored to inferred pre-folding positions to determine paleocurrent direction. Third, we were interested in trying to reconstruct the tectonic setting and tectonic history of these metasedimentary rocks as related to their position near the south margin of the Archean Wyoming Province.

This section attempts to explain the observed structural features in the Medicine Bow Mountains and regional structural features of the southern Wyoming Precambrian in terms of plate tectonics and current understandings of an evolving North America: with widespread tectonism and volcanism in the Archean which terminated in a period of orogenesis and cratonization about 2500 m.y. ago. This was followed by an Early Proterozoic (2500-1700 m.y.) tectonic regime characterized by rifting of larger and thicker cratonic blocks and deposition of platform and miogeoclinal sediments, including radioactive conglomerate, on the stable margins of these cratonic blocks. At about 1700 m.y., North America experienced widespread orogenesis which, in southern Wyoming, appears to have taken the form of one or more craton-island arc collisions (discussed below). This model of collisional orogenesis in southern Wyoming is summarized by Hills and Houston (1979) and successfully accounts for the regional geologic and geochronologic discontinuities across the Mullen Creek-Nash Fork shear zone and for the detailed structural features and fold chronologies described below.

Detailed structural studies of the metasedimentary rocks in the Medicine Bow Mountains have been presented by Houston and others (1968), Wilson (1975, 1977), Karlstrom (1977), and Karlstrom and Houston (1979b). This section is presented as an up-date and revision of structural interpretations presented in the latter paper and includes new data from the northern Medicine Bow Mountains. Table 2.26 is a comparison of fold chronologies used in this report and in previous papers. The most important new additions are evidence for a fold system in the Phantom Lake Metamorphic Suite which predated deposition of the Deep Lake Group and our interpretation that large gabbroic sills in the Deep Lake Group and Phantom Lake Suite were emplaced prior to an episode of thrust faulting and F_1 folding of the Libby Creek Group. In this report we emphasize the differences in structural style and tectonic history between the Libby Creek Group, which we now view as allochthonous or para-autochthonous and the underlying Phantom Lake and Deep Lake sequences which we view as autochthonous. We attempt to explain the observed structures in terms of Archean and Proterozoic tectonic events along the southern margin of the Wyoming Archean craton. Summaries of the depositional history of the metasedimentary rocks are presented earlier and in Houston and others, 1968; Karlstrom and Houston, 1979a; and Lanthier, 1979.

Plate 3 is a tectonic map of the northern Medicine Bow Mountains which shows the major faults and folds, sketches of the geometry of minor structures, and stereonet plots of structural elements from 25 subareas.

TABLE 2.26. COMPARISON OF CHRONOLOGIES OF MAJOR FOLDING EVENTS IN METASEDIMENTARY ROCKS

This Report	Karlstrom and Houston (1979b)	Wilson (1975, 1977)
F_1 - isoclinal overturned folds in Phantom Lake Suite		
F_2 - open, shallow plunging NE to E-W trending folds in Deep Lake Group; co-axial refolding of F_1 in Phantom Lake Suite.	F_1 - E-W folds in Deep Lake Group and Phantom Lake Suite.	
Intrusion of large gabbroic sills in Deep Lake Group and Phantom Lake Suite and dikes in Libby Creek Group		
F_3 - isoclinal reclined folds in Phantom Lake Suite and Deep Lake Group in northeastern Medicine Bow Mountains only		
F_1 - thrust faulting in Libby Creek Group and rotation to near vertical attitudes possibly contemporaneous with F_3 in Deep Lake Group.	F_2 - NE folds in Deep Lake and Libby Creek Groups and thrusting in Libby Creek Group.	F_1 - development of slaty cleavage in Lookout Schist. F_2 - development of cleavage parallel to bedding in Lookout Schist.
	Intrusion of gabbroic sills and dikes in all three sequences	Intrusion of dikes in Libby Creek Group
F_4/F_2 - northwest warping and development of crenulation in Deep Lake/Libby Creek Groups	F_3 - northwest warping and development of crenulation.	F_3 - asymmetrical crenulation cleavage in Lookout Schist.
F_5/F_3 - vertical rotation in Deep Lake/Libby Creek Groups and strike-slip movement on shear zone	F_4 - vertical rotation and strike-slip motion on shear zone.	Vertical axis warping and metamorphism of Lookout Schist.
Intrusion of mafic dikes and sills crosscutting shear zone.	Intrusion of mafic dikes and sills crosscutting shear zone.	

This type of geometrical analysis of folding, using stereographic projections (following Turner and Weiss, 1963), is ideal for representing megascopic folds in the metasedimentary rocks because original bedding is usually recognizable and defines a unique form surface. Most of the subareas in Plate 3 are roughly cylindrical with respect to one fold system; others show evidence of more than one fold system. The following discussion points out the main structural features shown in Plate 3, preceding from oldest to youngest. Then the structural and sedimentary evidence is synthesized into a Precambrian tectonic history of the Medicine Bow Mountains, involving a craton-island arc collision about 1700 m.y. ago. The last part of the paper points out the regional implications of this plate tectonic model in terms of concepts of continental accretion and Precambrian plate tectonics.

ARCHEAN STRUCTURAL FEATURES

Archean Gneisses

The western half of the Medicine Bow Mountains is made up of a series of Laramide blocks cored by Archean quartzo-feldspathic gneisses (Figure 2.1). The structure within these blocks is complex (Houston and others, 1968; Karlstrom and Houston, 1979b) but, on a regional scale, the gneisses are folded into a series of antiforms and synforms which vary in trend from north to northwest and which plunge gently northwest and southeast. Subarea 23 (Plate 3) shows a northwest plunging upright fold system and subarea 24 shows a southeast plunging overturned fold (both limbs dip northeast). This north- to northwest-trending fold system is not present in the metasedimentary rocks and probably predates deposition of the Phantom Lake Metamorphic Suite. These structures formed either before or during the tectono-thermal event about 2500 m.y. ago which affected basement rocks throughout the southern Wyoming Province (Houston and Karlstrom, 1980; Condie, 1976).

Archean gneisses were also deformed during the Proterozoic. Subarea 25 (Plate 3) shows a subvertical fold axis which is similar to the folds generated in metasedimentary rocks adjacent to the Mullen Creek-Nash Fork shear zone (subareas 18, 19, 20). This is good evidence that the gneissic basement was plastically deformed during the 1700 m.y. deformation in areas close to the shear zone and suggests that the rocks, exposed today, were at deep crustal levels around 1700 m.y. ago. Northeast-trending shear zones in the gneisses (Houston and others, 1968) probably also reflect Proterozoic deformations.

Phantom Lake Metamorphic Suite--F_1 Folding

New data from the northern Medicine Bow Mountains show that rocks of the Phantom Lake Metamorphic Suite record several deformational events and are now folded into tight to isoclinal folds, often overturned (axial planes dipping northwest), with variable trend and plunge (Plate 3). The earliest deformation(s) predated deposition of the Deep Lake Group, as shown by the angular unconformity seen in map pattern (Plate 3) and in the cirque walls of Crater Lake (Sec. 35, T.18 N., R.79 W.). These early folds will be designated F_1--the earliest fold system identified in the metasedimentary rocks. These folds probably formed about 2500 m.y. ago, contemporaneous with intrusion of granitic magmas in the northeastern Medicine Bow Mountains and near Baggot Rocks. Evidence for this interpretation is that the granites intrude the Stage Crossing Gneiss and Phantom Lake Metamorphic Suite in the northeast and pegmatite veins are folded with S_1 foliation in the lower Phantom Lake Suite (see Figure 2.25 and sketches of mesoscopic folds in Sec. 25, T.19 N., R.79 W. of Plate 3) implying that granitic material was injected parallel to S_1 in the metasediments probably during late stages of folding. The granite body in the northeast is discussed in more detail below.

Subareas 2, 3, and 4 in the Phantom Lake Suite record evidence of F_1 folding although the distributions of poles to bedding and foliation are dispersed due to later, Proterozoic, folding. Subarea 2 shows a south plunging fold in bedding with minor fold axis lineations parallel to the megascopic fold axis defined by the girdle of poles to bedding. Subarea 3 is one megascopic isoclinal syncline plunging south and southwest, here named the Foote Creek Synclinorium. Here, the original trend of S_1 was probably northeast (as shown in the stereonet) whereas S_1 and B_1 now have variable orientations due to later folding. Subarea 4 shows a girdle of poles to bedding with a maximum defining the position of S_1 foliation: trending northeast and dipping steeply northwest. Minor fold axis lineations in subareas 3 and 4 lie along the S_1 foliation or plunge west, parallel to the superposed F_3 fold axis.

There are numerous faults parallel to S_1 in the Phantom Lake Suite and these probably were first active during F_1 folding. These faults exhibit large vertical movements--both normal and reverse, which caused repetition of the major anticlinorium axis in subarea 4 (French Joes Anticlinorium of Plate 3). This type of fault could easily form in anisotropic sediments in response to the same stresses that caused F_1 folding. Like the F_1 fold axis, the faults have been refolded by Proterozoic

Figure 2.25. Granite crosscutting the Stage Crossing Gneiss in the northern Medicine Bow Mountains, west side of Rock Creek, Sec. 25, T.19 N., R.79 W. Granite was apparently folded into F_1 folds and injectd parallel to S_1 foliation suggesting that intrusion of granite took place during deformation of the gneisses.

deformation (as in Sec. 1, T.18 N., R.79 W.).

In the central Medicine Bow Mountains, evidence for a pre-Deep Lake fold system in rocks of the Phantom Lake Suite is not as clearly preserved and both sequences appear to be folded together co-axially into F_2 folds (compare subarea 10 in the Phantom Lake Suite with subarea 11 in the Deep Lake Group). Farther southwest, there appears to be weak evidence for east-west trending folds in the Phantom Lake Suite rocks in the core of the Arrastre Anticline (subarea 14) but not in flanking Deep Lake Group rocks (subarea 13).

Baggot Rocks-Type Granite

Granitic gneisses crosscut Archean quartzo-feldspathic gneisses in the Baggot Rocks area and metasedimentary rocks of the Stage Crossing Gneiss and Phantom Lake Metamorphic Suite in the northeastern Medicine Bow Mountains, near Arlington. The Baggot Rocks Granite is a phacolithic body located in the hinge of a north-plunging antiform in Archean gneisses. Foliation in the granite generally is parallel to foliation in the gneiss but the granite definitely crosscuts the gneiss in some areas (Houston and others, 1968). The granite body near Arlington is less well exposed. However, it too is generally sill-like, conformable to S_1 foliation in the metasedimentary and metavolcanic rocks, but in detail crosscuts the metasedimentary rocks. The granite locally appears to post-date the development of S_1 foliation and granitic veinlets are folded, with S_1, into F_3 folds. However, some granites are not foliated so we interpret the intrusion episode to have been contemporaneous and during the late stages of F_1 folding of the metasedimentary rocks.

Unfortunately, only a small area of granite is exposed in the northern Medicine Bow Mountains; granite outcrops are bounded by the Arlington thrust fault of Laramide age and Phanerozoic sedimentary rocks, so we don't know the original extent of the Arlington granite body. However, the intensity of F_3 folding adjacent to the granite (discussed below), and the absence of any positive gravity anomaly (Houston and others, 1979, unpublished data) which might indicate a continuation of metavolcanic rocks to the north, suggests that there may have been a sizeable granitic dome to the north which was a buttress against which the metasediments were deformed during Proterozoic orogenies and which may have been a structural and topographic highland which bounded the Deep Lake Group depositional basin. If so, outcrops of granite now exposed in the northern Medicine Bow Mountains may be the slightly granitized margin of a larger granitic body.

As discussed earlier, we interpret the Baggot Rocks-type granites to be S-type granites (Chappel and White, 1974) which were derived from partial melting of the older quartzo-feldspathic gneiss and intruded as sills and phacoliths on a variety of scales.

PROTEROZOIC STRUCTURAL FEATURES

Deep Lake Group--F_2 Folding

The earliest Proterozoic deformation is recorded in rocks of the Deep Lake Group, which unconformably overlies Archean granites in the northern Medicine Bow Mountains. This deformation produced open, upright, northeast-trending folds, designated F_2, which are approximately co-axial with F_1 folds. These folds can best be seen in subareas 8, 9, 11, 12, 13, and 14, which lie along a single anticline-syncline system in the Deep Lake Group that traverses the entire Medicine Bow Mountains (Arrastre Anticline-Sand Lake Syncline of Plate 3). The trend of this fold system varies from east-west to north-northeast and plunges are shallow (less than 30 degrees) to the southwest and northeast. The relationships of F_2 to F_1 can best be seen in the Crater Lake area (Sec. 35, T.18 N., R.79 W.) where open folds in the Deep Lake Group are coaxial with tight to isoclinal folds in the unconformably underlying Phantom Lake Suite and in subareas 12 and 11 where F_2 and F_1 folds are coaxial but one limb of the fold in the Phantom Lake Suite is overturned whereas folds in the Deep Lake Group are upright (Plate 3). Apparently the Deep Lake Group was deposited on a folded Phantom Lake Suite surface; then continued folding formed open basin and dome type folds in the Deep Lake Group while more tightly appressing folds in the Phantom Lake Suite. The types of folding seen in the Deep Lake Group and the accompanying longitudinal faults are typical of continental platform folding seen in areas like the Gulf Coast today (Hobbs and others, 1976).

Timing of F_2 folding is constrained by time of deposition of the Deep Lake Group which is involved in F_2 folds and time of intrusion of the large gabbroic sills and dikes which crosscut F_2 folds and do not possess an S_2 foliation (Karlstrom and Houston, 1979b). This folding, therefore, took place after 2500 m.y., the inferred age of the northern granites, and before 2150 m.y., the inferred age of gabbroic sills. However, as shown in Table 2.27 and reported by Karlstrom and Houston (1979b), it remains possible that F_2 folding may have taken place as late as 1700 m.y. if the gabbroic sills in the Deep Lake Group are also 1700 m.y. old.

Deep Lake Group--F_3 and F_4 Folding

Rocks of the Phantom Lake Suite and Deep Lake Group retain evidence of a later fold system--F_3. The deformation was intense enough in the northern Medicine Bow Mountains that in subareas 1 and 5, F_3 has obscured earlier structures and is the dominant fold system seen in mesoscopic folds and in stereographic plots of bedding and foliation (Plate 3). In subarea 1, poles to bedding and foliation in the Phantom Lake Suite plot together in a great circle girdle defining a west-plunging fold axis which is subparallel to minor fold axis lineations and mineral lineations. The fold axis is perpendicular to the maximum of poles to bedding and foliation so F_3 is reclined. Subarea 5 shows a similar west-plunging, reclined F_3 fold system in the Deep Lake Group. However in subarea 5A, the maximum of poles to bedding includes both upright and overturned beds providing evidence for an earlier, F_2, fold system.

North-trending, west-plunging, reclined F_3 folds are present in subareas 1, 3, 5, 6, 7, and 8--all in areas close to the Laramide boundaries of Precambrian outcrop (Plate 3). Because of this spatial relationship, we have wondered whether this fold system might be Laramide in age. However, the plastic deformation seen in mesoscopic F_3 structures suggests to us that the folding was Precambrian in age and we prefer to believe that the intense development of F_3 in these areas and the (much later) propagation of Laramide faults both reflect an older Precambrian crustal weakness, possibly related to the margin between a large granitic dome or "highland" to the north and a Proterozoic sedimentary basin to the south.

Several mesoscopic folds in the northern Medicine Bow Mountains show a later northwest-trending foliation, which crenulates F_3 folds (Plate 3), and similar northwest-trending folds and foliation are present in the central Medicine Bow Mountains, where F_2 folds and gabbroic intrusions are gently warped. We refer to this northwest-trending crenulation schistosity and the warping of megascopic structures as F_4. F_4 may represent a modest change in orientation of principal stresses during the same deformational event which produced F_3 folds or it may represent a later deformation. Certainly both the development of west-plunging reclined folds in the northern Medicine Bow Mountains and the development of broad folds and northwest-trending foliation in the central Medicine Bow Mountains post-dated the intrusion of gabbroic sills and the stress fields needed to produce both types of folds are roughly similar (i.e. northeast-southwest compression). Thus, for simplicity, we interpret F_3 and F_4 to represent a somewhat heterogeneous deformation, probably related to Proterozoic deformations seen in the overlying Libby Creek Group.

Libby Creek Group--Thrust Faulting

Structural style in the Libby Creek Group is so different from that of the Deep Lake Group and Phantom Lake Suite that we will discuss it separately before trying to relate structures in the Libby Creek and Deep Lake Groups to one another and to an interpretive overview of the tectonic history of the Medicine Bow Mountains.

The Libby Creek Group strikes northeast and stands on end, with dips near vertical at the Mullen Creek-Nash Fork shear zone (subarea 20 of Plate 3) and 50 to 60 degrees southeast near the contact with the Deep Lake Group (subarea 15). The overall structure of the Libby Creek Group can best be visualized as the steep limb of a large south-facing monoclinal flexure (Plate 3). Perhaps the most important aspect of recent structural interpretations of the Libby Creek Group (Lanthier, 1978; Karlstrom and Houston, 1979b) is the presence of several major reverse faults which roughly follow the Libby Creek Group-Deep Lake Group contact (Reservoir Lake Fault of Plate 3) and bedding contacts within the Libby Creek Group (Lewis Lake Fault of Plate 3) and cause tectonic attenuation of the Libby Creek Group section. Evidence for such faults, in addition to breccias and fault scarps along the fault trace, is that massive quartzite sections of both the Medicine Peak (1700 m thick) and Sugarloaf (600 m thick) quartzites abruptly pinch to zero thickness, then reappear at normal thickness, along strike (Plate 1). Also, major transverse faults in the Lower Libby Creek Group terminate at the inferred thrusts and are not present either in footwall rocks of the Deep Lake Group or younger hanging wall rocks of the Upper Libby Creek Group (Plate 1). We believe that tectonic attenuation of this magnitude, as well as differing structural styles between hanging wall and footwall suggest large amounts of movement and we postulate that these reverse faults were originally thrust faults that have subsequently been rotated to steep attitudes.

There are three main fold systems recorded in rocks of the Libby Creek Group. The main one, and presumably the earliest, rotated the Libby Creek Group rocks and the inferred thrust faults contained therein to subvertical attitudes. This structure is referred to as F_1 for the Libby Creek Group and is reflected in plots of poles to bedding in subareas 17, 18, and 20. A secondary deformation, which may have been only slightly later than the first caused flexuring about a northwest-southeast trend. This fold system F_2, is best seen in subareas 15 and 18. A third fold system, F_3, seen in areas close to the Mullen Creek-Nash Fork shear zone, rotated earlier structural elements about a near vertical axis in response to left-lateral, strike-slip movement on the shear zone. This rotation is seen in subarea 19, where steeply plunging fold axis lineations and quartz c-axis are redistributed on a small circle about a vertical fold axis (Houston and Parker, 1963), indicating flexural slip re-folding of a pre-existing lineation. This last deformation may also have produced crenulations in subareas 18 and 20.

The best clue to the timing of deformations in the Libby Creek Group is that rocks within the Mullen Creek-Nash Fork shear zone (subarea 21) and rocks south of the shear zone (subarea 22) show structural features similar to those of the Libby Creek Group suggesting similar tectonic histories. Hills and Houston (1979), from geochronologic data, showed that the last major movements across the shear zone took place around 1700 m.y. ago contemporaneous with metamorphism of the Libby Creek Group, and they interpreted both in terms of an island arc-craton collisional orogeny. As discussed in more detail later, we believe that all three deformations observed in the Libby Creek Group were related to this 1700 m.y. orogeny. We infer that subhorizontal thrusting took place first in the miogeoclinal Libby Creek Group followed by rotation of the Libby Creek Group and the thrusts to near vertical position in this area and metamorphism of the sediments to amphibolite grade. Then the northwest warping and vertical rotations were formed during a change from compressional to strike-slip movements in an oblique collision zone.

Intrusion of Gabbroic Rocks

Interpretation of the timing of deformations in the Libby Creek Group relative to that of the underlying metasedimentary successions depends to a large degree on evidence from the gabbroic intrusive rocks which crosscut all three sequences of metasedimentary rocks. In the Phantom Lake Suite and Deep Lake Group, these intrusives are large bodies (kilometers long and several hundred meters wide) which were emplaced along bedding and fault planes and are generally concordant to bedding. These can loosely be referred to as sills. In the Libby Creek Group the gabbroic rocks are less abundant and are mainly smaller crosscutting dikes except in areas adjacent to the Mullen Creek-Nash Fork shear zone, where large sills reappear.

Large sills in the lower metasedimentary sequences are phacolithic bodies which, in detail, crosscut F_1 and F_2 folds (Plate 1, Secs. 10, 11, T.16 N., R.80 W.). They are folded by F_3 and F_4 and locally possess north and northwest-trending S_3 and S_4 foliations. The direction of opening of several of the sills suggests that they were emplaced under a stress field with extension in a northwest-southeast direction (northeast-southwest compression) which is approximately the same stress field needed to generate F_3 (Karlstrom, 1977; Karlstrom and Houston, 1979b). Thus, the sills were intruded after F_2 folding in the Deep Lake Group and before or in the early stages of F_3 folding. Tholeiitic sills (Houston and others, 1968) of this type are commonly interpreted as evidence for extensional tectonics which accompanies rifting (May, 1971; Windley, 1977; Burke, 1980) and we interpret these sills in this light--as evidence for northwest-southeast rifting along the south margin of the Wyoming Province.

If this is true, the sills must be older than the postulated 1700 m.y. collisional orogeny and may be analogous to the 2150 m.y. old Nipissing diabase of the Canadian Huronian Supergroup and the 2100 m.y. old mafic sills in the Black Hills of South Dakota (Redden, 1980).

Some of the gabbroic dikes in the Libby Creek Group may also be related to rifting processes. This is supported by gradational and apparently mutually crosscutting contact relationships between some mafic dikes and the Gaps Trondhjemite which has yielded a 2000 m.y. date (Carl Hedge, personal communication, 1980). However, these intrusives are more alkaline than the tholeiitic sills in the underlying successions (Figure 2.19) and we speculate that some of the dikes in the Libby Creek Group, while still related to the 2100-2000 m.y. rifting event, may be slightly younger than the large tholeiitic sills. Other dikes in the Libby Creek Group and large sills and dikes near the Mullen Creek-Nash Fork shear zone may be associated with the 1700 m.y. orogeny. This interpretation is supported by the higher concentration of mafic bodies close to the shear zone and by evidence that mafic dikes and sills crosscut the major reverse faults as well as the Mullen Creek-Nash Fork shear zone (Plate 1).

Thus, we postulate at least two ages of mafic intrusions: a compound episode about 2100-2000 m.y. which produced both tholeiitic sills and more alkaline dikes at higher structural levels, and an episode related to collision about 1800-1700 b.y. ago, and we hypothesize a somewhat different interpretation of the timing of deformations than was presented in earlier papers (Karlstrom and Houston, 1979b). As shown in Table 2.26, we now interpret the intrusion of sills in the Deep Lake Group and Phantom Lake Suite and some dikes in the Libby Creek Group to have pre-dated the F_1 folding in the Libby Creek Group. Additional evidence for this interpretation comes from Lanthier (1978) who shows large sills in the upper Deep Lake Group which are crosscut by the basal thrust fault (the Reservoir Lake Fault of Plate 3). By this interpretation, F_3 folding in the northern Medicine Bow Mountains either preceded or was contemporaneous with F_1 folding and thrust faulting in the Libby Creek Group; and F_4 folding in the lower sequences would be contemporaneous with F_2 folding in the Libby Creek Group (Table 2.26).

LARAMIDE STRUCTURAL FEATURES

Before proceeding to a Precambrian tectonic synthesis, it is useful briefly to examine the complexities and uncertainties introduced by Laramide uplift. The best documented Laramide structural features include: the Arlington-Corner Mountain thrust which bounds the Medicine Bow Mountain uplift on its east side; several other west-dipping thrusts which bound the east sides of blocks of Archean gneiss in the western Medicine Bow Mountains; and at least two right-lateral faults with up to several kilometers of displacement which traverse the metasedimentary section and offset the Arlington-Corner Mountain thrust (Plate 3). There are also numerous smaller faults which involve Mesozoic and Tertiary rocks and we presume that most of the faults shown in Plate 3 moved during the Laramide orogeny. In fact, we interpret the pattern of fault-bounded blocks seen in Plate 3 to be a basement controlled but Laramide reactivated system of faults and we envision the mechanism of Laramide uplift to be brittle failure along a myriad of such faults and microfractures. If so, observed map displacements in Precambrian rocks often will be a result of both Precambrian and Laramide movements which need not show the same relative motions and which may not be distinguishable. Here, again, the gabbroic intrusives are sometimes useful in establishing the time of deformation. Faults which have been intruded by gabbroic rocks clearly moved in the Precambrian prior to the episode of magma intrusion while faults which offset gabbroic rocks may be Laramide faults.

SYNTHESIS--A PRECAMBRIAN TECTONIC HISTORY OF THE MEDICINE BOW MOUNTAINS

The following is an attempt to understand the observed stratigraphic and structural relationships in the Medicine Bow Mountains in terms of plate tectonic models. We feel that a model for southeastern Wyoming involving Archean permobile tectonics related to microplate tectonics followed by rifting, miogeoclinal (passive margin) sedimentation, and island arc-continent collision in the Early Proterozoic has numerous advantages. It fits existing geologic and geochronologic data very well (Hills and Houston, 1979) and it provides a framework to better understand and test future data. Figure 2.26 and Table 2.27 list what we consider to be "facts" so that critical readers and future workers in the area can distinguish fact from inference in the following discussion. Figure 2.27 is a series of cartoons depicting our model for the tectonic history of southeastern Wyoming.

ARCHEAN SEDIMENTATION AND TECTONICS

The oldest rocks in the Medicine Bow Mountains are the quartzo-feldspathic gneisses in the western part of the range. These rocks have yielded a metamorphic age of 2500 m.y. (Hills and Houston, 1979). The history of these rocks is poorly known but analogies with similar gneissic terranes in the Granite Mountains (Peterman and Hildreth, 1978), the Laramie Mountains (Graff and others, 1978) and the Sierra Madre (Graff, 1978; Divis, 1976, 1977) suggest a complex history, shown in Figure 2.26, involving: deposition of volcano-sedimentary protoliths to the gneisses possibly prior to 2900 m.y.; regional metamorphism around 2900-3000 m.y.; intrusion of tonalitic magmas accompanying folding and regional metamorphism around 2700 m.y.; and intrusion of granitic magmas accompanying

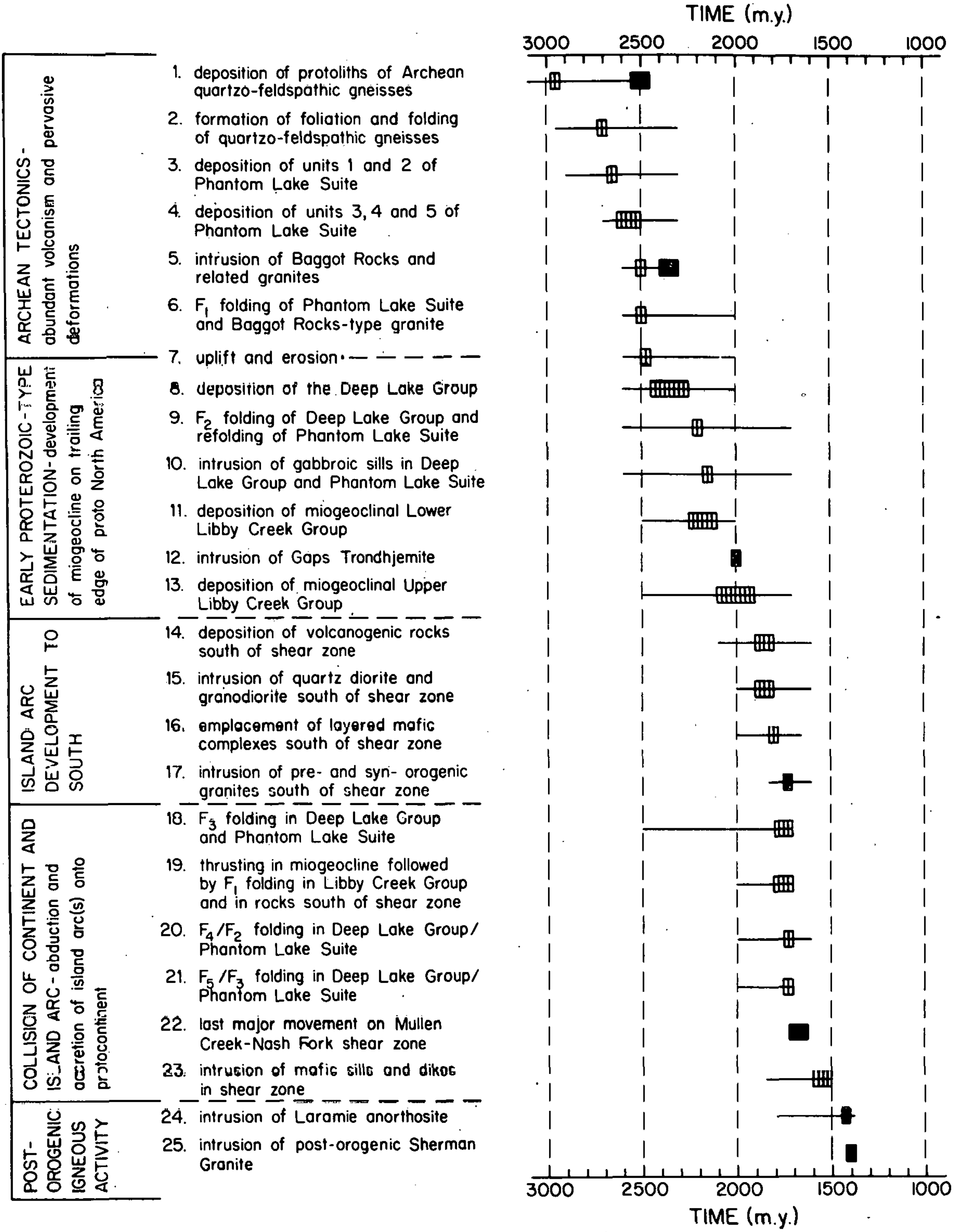


Figure 2.26. Precambrian geologic history of the Medicine Bow Mountains. solid box=geochronological data, showing analytical error; vertical lines=preferred interpretation; solid line=possible range of timing of event. See Table 2.27 for references and justification of dates shown.

regional metamorphism around 2500 m.y. (Hills and Houston, 1979). Of these, the only event positively recognized in the gneissic terrane of the Medicine Bow Mountains is the tectono-thermal event at 2500 m.y. and the intrusion of the synorogenic Baggot Rocks Granite (Hills and others, 1968). However, structural data indicate that the gneissic basement was deformed before deposition of the Phantom Lake Metamorphic Suite, which is also intruded by presumed 2500 m.y. old synorogenic Baggot Rocks-type granites, suggesting pre-2500 m.y. tectono-thermal events in the gneisses.

Deposition of both the Stage Crossing Gneiss and the Phantom Lake Metamorphic Suite is presumed to have taken place on a basement gneiss terrane similar to that of the western Medicine Bow Mountains although metavolcanics and metasediments are not in direct contact with the quartzo-feldspathic gneisses. As discussed earlier, the Stage Crossing Gneiss is a volcanogenic succession, perhaps on the order of .5 km thick. This was followed by .5 km of volcaniclastic rock, up to one km of arkosic paraconglomerate (locally radioactive) and fine-grained biotitic quartzite, 1-2 km of volcanic rocks, and .8 km of fine-grained biotitic quartzite. Deposition appears to have taken place in a relatively small sedimentary basin (Figure 2.22A) perhaps bounded on several sides by gneissic basement.

The earliest deformation of the Phantom Lake Suite was probably broadly contemporaneous with intrusion of granitic rocks near the end of the Archean (Figure 2.26). Formation of S_1 foliation in the Stage Crossing Gneiss and Phantom Lake Suite appears to have preceded or accompanied granitic intrusions as evidenced by the facts that pegmatites are folded with S_1 and the granitic body is locally foliated and generally conformable

TABLE 2.27. REFERENCES AND JUSTIFICATIONS FOR MAXIMUM, MINIMUM, AND PREFERRED AGES OF GEOLOGIC EVENTS SHOWN IN FIGURE 2.26.

	MAXIMUM	MINIMUM	PREFERRED
1. Deposition of protoliths of Archean quartzo-feldspathic gneisses.	3100, from oldest rocks known in the area (Peterman and Hildreth, 1978).	2500±50, from metamorphic date on quartzo-feldspathic gneisses (Hills and others, 1968)	2950, from the presence of pre-2700 m.y. gneisses in the Sierra Madre (Hedge, personal communication, 1979) and Laramie Mountains (Johnston and Hills, 1976) and presence of 2900 metamorphic episodes in both these ranges.
2. Formation of foliation and folding of quartzo-feldspathic gneisses.	2950, from oldest metamorphic date in the area.	2300, because deformation preceded deposition of lower Phantom Lake Suite.	2700, from evidence for regional metamorphism and intrusion of synorogenic tonalites in the Sierra Madre at about 2700 (Hedge, personal communication, 1979).
3. Deposition of units 1 and 2 of Phantom Lake Suite.	2900, because Wyoming Province greenstone belts are mainly 2700-2900 (Houston and Karlstrom, 1980) and are similar to lower Phantom Lake Suite.	2300, from intrusive contacts with Baggot Rocks-type granite.	2650, from analogy to Laramie Range (Graff and others, 1981), and Atlantic City (Bayley and others, 1973) greenstone belts and from fact that deposition of the lower Phantom Lake Suite postdated folding in the basement gneiss terrane.
4. Deposition of units 3, 4, and 5 of Phantom Lake Suite.	2700, because upper Phantom Lake Suite is more quartz-rich than typical greenstone belts.	2300, from presumed intrusive relations to Baggot Rocks-type granite.	2650-2500, from structural and depositional(?) continuity with lower Phantom Lake Suite.
5. Intrusion of Baggot Rocks and related granites.	2505±100, from Sierra Madre (Divis, 1977).	2265±70 from Medicine Bows (Hills and others, 1968).	2500, from prevalence of 2500 m.y. old granites in the Wyoming Province and elsewhere in the world (Houston and Karlstrom, 1980).
6. F_1 folding of Phantom Lake Suite and Baggot Rocks-type granite.	2600, because Baggot Rocks-type granite is involved in folding.	2000, because F_1 pre-dated deposition of the Deep Lake Group.	2500, because F_1 is interpreted to be a late orogenic folding episode associated with intrusion of the Baggot Rocks-type granite.
7. Uplift and erosion.	2600, because Baggot Rocks-type granite was eroded.	2000, because basal Deep Lake Group contains granite clasts.	2450-2500, shortly after metamorphism and intrusion of granites and coincident with a period of world-wide cratonic stabilization and erosion (Windley, 1977).
8. Deposition of Deep Lake Group.	2600, age of underlying granite.	2000, age of Gaps Trondhjemite (Hedge, personal communication, 1979).	2450-2250, by analogy to the lower Huronian Supergroup and other Early Proterozoic sequences containing uranium-bearing fossil placers (Roscoe, 1973, Houston and Karlstrom, 1980).
9. F_2 folding of Deep Lake Group and re-folding of Phantom Lake Suite.	2600, because Deep Lake Group is involved in folding.	1700, because folding pre-dated intrusion of large gabbroic sills.	2200, just prior to rifting and intrusion of gabbroic sills. F_2 is believed to have resulted from sediment loading and stresses due to rifting.
10. Intrusion of gabbroic sills in Deep Lake Group and Phantom Lake Suite and dikes in Libby Suite and dikes in Libby Creek Group.	2600, because sills intrude Deep Lake Group.	1700, sills appear to be cut by shear zone.	2150, by analogy to Nipissing diabase (Fairbairn and other, 1969) and probability that thrust faults in the Lower Libby Creek Group crosscut sills.
11. Deposition of miogeoclinal Lower Libby Creek Group.	2600, sedimentation was contemporaneous or later than that of upper Deep Lake Group.	2000, age of intrusive Gaps Trondhjemite.	2250-2100 by analogy to upper Huronian Supergroup (Young, 1973), by presence of hematite rather than pyrite in quartzites (Roscoe, 1973), and interpretation that sediments were deposited on a rifted continental margin.
12. Intrusion of Gaps Trondhjemite.	2600, intrudes Lower Libby Creek Group.	1755, poor quality Rb-Sr date reported by Hills and Houston (1979).	2000, preliminary 2000 m.y. Rb-Sr whole rock date (Carl Hedge, personal communication, 1980).
13. Deposition of miogeoclinal Upper Libby Creek Group.	2500, probably younger than Deep Lake Group.	1700, because sediments are cut by shear zone.	2100-1900, by analogy to the Marquette Range Supergroup (Van Schmus, 1976).

TABLE 2.27 (continued) REFERENCES AND JUSTIFICATIONS FOR MAXIMUM, MINIMUM, AND PREFERRED AGES OF GEOLOGIC EVENTS SHOWN IN FIGURE 2.26.

	MAXIMUM	MINIMUM	PREFERRED
14. Deposition of volcanogenic rocks south of shear zone.	2000, oldest date known from anywhere south of the shear zone in North America is about 1800 m.y. (Barker and others, 1976; Silver and others, 1977). Oldest date in the area is a questionable 1940±70 m.y. date on granodiorite which crosscuts volcanogenic sediments in the Sierra Madre (Divis, 1977).	1600, minimum age of crosscutting syn-orogenic granites (Hills and Houston, 1979).	1800-1900, deposition must have pre-dated the major orogenic activity at 1700 and volcanogenic rocks are crosscut by large layered mafic complexes similar to the 1780 m.y. gabbro complex in the southern Sierra Madre (Snyder and Hedge, 1978).
15. Intrusion of quartz diorite and granodiorite south of shear zone.	2000, from questionable 1940±70 m.y. date on granodiorite in the southern Sierra Madre (Divis, 1977).	1600, from minimum age of intrusive syn-kinematic granites (Hills and others, 1968).	1800-1900, from interpretation that quartz-diorites represent batholithic intrusions of the encroaching island arc which should be approximately the same age as volcanogenic sediments.
16. Emplacement of layered mafic complexes south of shear zone.	2000, the Lake Owen mafic complex may crosscut the Keystone quartz diorite which is similar to granodiorite in the Sierra Madre which could be as old as 1940±70 m.y. (Divis, 1977).	1700, the Mullen Creek complex is crosscut by the granite of Horse Creek (Hills and Houston, 1979).	1800, from date of 1780 on a similar layered mafic complex in the southern Sierra Madre (Snyder and Hedge, 1978).
17. Intrusion of pre- and syn-orogenic granites south of shear zone.	1725±120 from granite of Horse Creek (Hills and Houston, 1979).	1600, from various dates on granites in Sierra Madre and Medicine Bows (Hills and Houston,1979).	1720, from Hills and Houston (1979), as an estimate of the average of the most reliable dates (1730±15; 1725±120; 1775±?).
18. F_3 folding in Deep Lake Group and Phantom Lake Suite.	2500, large sills in Deep Lake Group are folded by F_3.	1700, must be same age or older than thrusts in Libby Creek Group because northeast trend of thrusts crosscuts warping in the Deep Lake Group F_2 structures which may be related to F_3.	1800-1700, F_3 interpreted to be result of stresses generated by thrust faulting and transmitted through the sedimentary pile; localization of folding may have been related to proximity to a large anisotropic block of granite to the north.
19. Thrusting in miogeocline followed by F_1 folding in Libby Creek Group and in rocks south of shear zone.	2000, volcanogenic rocks south of shear zone are interpreted to be involved in thrusting.	1700, thrusting preceded major strike-slip movement on shear zone.	1800-1700, interpreted as the first effects of island are collison and obduction; probably preceded major metamorphism of Libby Creek Group at 1700 m.y. (Hills and others, 1968).
20. F_4/F_2 folding in Deep Lake Group/Libby Creek Group.	2000, thrust faults are folded by by F_4/F_2.	1600, folding probably occurred during orogenic activity.	1750-1700, folding reflects NE-SW compression, perhaps generated by change in orientation of stresses in an oblique collision zone.
21. F_5/F_3 folding in Deep Lake Group/Libby Creek Group.	2000, F_5/F_3 folding post-dates F_4/F_2.	1700, folding of French Creek Syncline and development of crenulations preceded major strike-slip movement on shear zone.	1750-1700, Wilson (1975) reported that the major metamorphism of the Lookout Schist, which was dated by Hills and others (1968) to be at 1700 m.y., was later than the development of F_3 crenulations in Libby Creek Group.
22. Last major movement on Mullen Creek-Nash Fork shear zone.	1840, Horse Creek granite (1725±100) is crosscut by the shear zone.	1600, post-tectonic Red Granite in the Sierra Madre (1645±50-1635±20) crosscuts the shear zone (Hills and Houston, 1979; Divis, 1977).	1725-1635, from Hills and Houston, (1979), strike-slip movement of the shear zone is interpreted to represent a change from the compressive stresses generated during obduction to transform motion towards the end of the end of the orogenic episode.
23. Intrusion of mafic sills and dikes in shear zone.	1840, some mafic dikes crosscut the shear zone and appear to have invaded folded thrust faults in the French Creek Syncline area.	1500, the age of some pegmatites south of the shear zone (Hills and others, 1968).	1600-1500, these dikes are interpreted to be late- and post-orogenic intrusions.
24. Intrusion of Laramie anorthosite.	1800, anorthosite could be temporally related to layered mafic complexes south of the shear zone.	1385, the anorthosite is intruded by the Sherman Granite.	1435±15, date on syenite which crosscuts but is probably co-genetic with the anorthosite (Subbarayudu, 1975).
25. Intrusion of post-orogenic Sherman Granite.	1430±60, from the Sherman Granite of the Laramie Range (Subbarayudu, 1975).	1380±30, from the Colorado Front Range (Peterman and others, 1968).	1385±30, from the southern Medicine Bow Mountains (Hills and others, 1968).

with S_1. We visualize a tectonic regime where deformation proceeded contemporaneous with and perhaps as a consequence of intrusion of granitic diapirs which originated by partial melting of quartzo-feldspathic gneisses (and perhaps in part Phantom Lake Suite metasediments), moved upward in response to gravitational instabilities caused by density differences, and were emplaced mainly as sills and phacolithic bodies at higher structural levels. The processes may have been similar in many respects to the formation of gneissic core complexes in the Cordilleran infrastructure (Best and others, 1974; Reesor, 1970; Davis and Coney, 1979). F_1 folding in the Phantom Lake Suite resulted in the formation of tight, northeast-trending, northwest dipping, overturned folds with shallow plunges. Numerous high angle faults, both normal and reverse, developed parallel to the axial plane probably during this deformation.

Proto plate tectonic models for deposition and deformation of the Phantom Lake Metamorphic Suite, while not necessarily compelling, might be helpful in understanding several regional relationships. Archean sedimentary sequences which may be approximately the same age (older than 2700 m.y.) occur in the Sierra Madre, Medicine Bow Mountains, Laramie Range (Graff and others, 1981), Hartville Uplift (Snyder, 1980) and perhaps the Black Hills (Redden, 1980). The configuration of outcrop areas (Figure 2.22A) combined with the differences in lithologies suggest that each of these areas represents different volcanic and sedimentary conditions and possibly different sedimentary basins. These volcano-sedimentary successions may well be different ages but, even so, it is interesting to note that each of these basins is located along a line parallel to major Proterozoic plate boundaries and parallel to a line of greenstone belts in central Wyoming. As shown in Figure 2.22A, both parallelisms might be understood using a proto plate tectonic model involving an Archean island arc complex: the line of small Archean sedimentary basins near the south margin of the Wyoming Province may represent differing contributions of volcanism and sedimentation, i.e. differing tectonic conditions, along a cratonic margin or in a fore-arc basin setting and the inboard line of greenstone belts may represent back-arc basins or intracratonic sedimentation and volcanism related to rifting. (See also Condie, 1972, for a plate tectonic model of the Atlantic City greenstone terrane.) By this model, the Late Archean U- and K-rich granites of the Granite Mountains (Stuckless, 1979) and northern Laramie Range (Karlstrom and Houston, 1981), which we believe were the major source of uranium for southeastern Wyoming fossil placers, may represent areas where partial melting of mantle rocks in subduction complexes was contaminated by contributions from fragments of older continental crust (Figure 2.22A) so that the resulting batholithic granites were enriched in U and K. This situation may have been analogous to formation of highly silicic and potassic granites of the Malay Peninsula of the Indonesian region (Hamilton, 1979, p. 71).

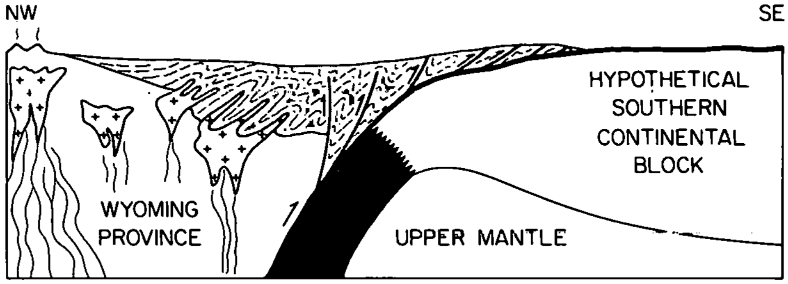
The details of such a plate tectonic model for the Late Archean of the Wyoming Province are engimatic, but we envision tectonic complexities (in time and space) similar to those described by Hamilton (1979) in the Indonesian region. In fact, we believe that the Indonesian area may be analogous in many ways to Archean processes of formation of continental crust and interaction of microplates. This type of proto plate tectonic model for formation of the Wyoming Archean craton suggests: 1) Archean crust within the Province was experiencing tectonism at about the same time as the margins of the Province due to interaction of microplates; 2) stabilization of the Wyoming Archean cratons near the end of the Archean was related to cessation of tectonism along the margins of microplates; 3) the southern boundary of a Late Archean island arc system became the site of the Early Proterozoic cratonic boundary, on which miogeoclinal rocks were deposited. Figure 2.27A is a schematic cross-section illustrating one possible tectonic model for Late Archean deformation and plutonism in southeastern Wyoming.

EARLY PROTEROZOIC-TYPE SEDIMENTATION--RIFTING AND DEVELOPMENT OF A MIOGEOCLINE ON A TRAILING EDGE OF PROTO-NORTH AMERICA

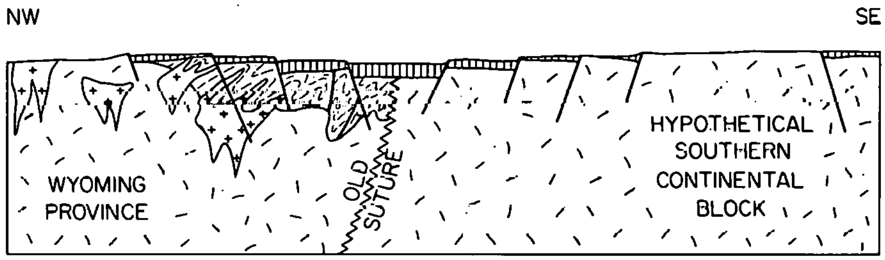
Deposition of the Deep Lake Group took place mainly in fluvial depositional environments sometime after the F_1 deformation, as evidenced by the angular unconformity and the difference in structural styles between the Deep Lake Group and the underlying Phantom Lake Suite. The basal Deep Lake Group contains a uranium- and pyrite-bearing quartz-pebble conglomerate which was deposited unconformably on the Phantom Lake Metamorphic Suite and on presumed Archean granites in the northern Medicine Bow Mountains. This conglomerate indicates a period of erosion of an uplifted Archean granitic terrane to the north and the persistence of relatively stable sedimentary systems near the south margin of the Wyoming Province sometime after 2500 m.y. ago. As discussed earlier, and shown in Figure 2.22B, deposition throughout the lower Deep Lake Group was dominated by fluvial sedimentation in presumably intracratonic, fault-bounded basins. Deposition of the upper Deep Lake Group involved deltaic and glacio-marine sedimentation.

F_2 folding, seen mainly in rocks of the Deep Lake Group, consisted of concentric, open, basin and dome folding and contemporaneous faulting. The folds trend northeast and plunges are shallow to the northeast and southwest. The presence of disharmonic (soft-sediment?) folds in the upper Deep Lake Group suggests that this deformation commenced during deposition of the upper Deep Lake Group (Karlstrom, 1977). This is consistent with our current interpretation that F_2 folding resulted from differential subsidence and faulting within the epicontinental sedimentary sequence. As shown in Figure 2.27C, this folding is interpreted to be related to rifting of the cratonic margin and sediment loading due to deposition of the overlying miogeoclinal sediments of the Libby Creek Group. In addition to forming open folds in the Deep Lake Group, F_2 caused nearly co-axial refolding of F_1 folds in the underlying Phantom Lake Suite.

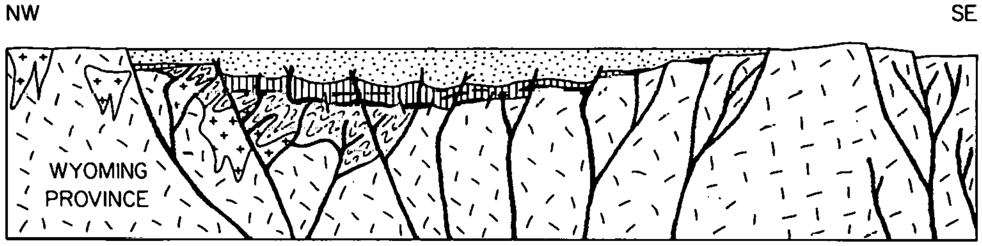
The next event was the intrusion of gabbroic sills in the Phantom Lake Suite and Deep Lake Group (Figure 2.27C). This intrusion episode definitely post-dated the development of F_2 folds but the timing of intrusion of gabbroic sills relative to the time of deposition of the Libby Creek Group is not well constrained. However, we presume that the gabbroic sills were related to an episode of rifting and extensional tectonics along the cratonic margin so we consider the sills to be about the same age as deposition of miogeoclinal rocks of the Libby Creek Group. Some gabbroic dikes in the Lower Libby Creek Group (Plate 3) may well be the same age, but a higher-level expression of the sills in the Deep Lake Group and Phantom Lake Suite. Other dikes and sills, especially in the Upper Libby Creek Group and rocks south of the shear zone probably



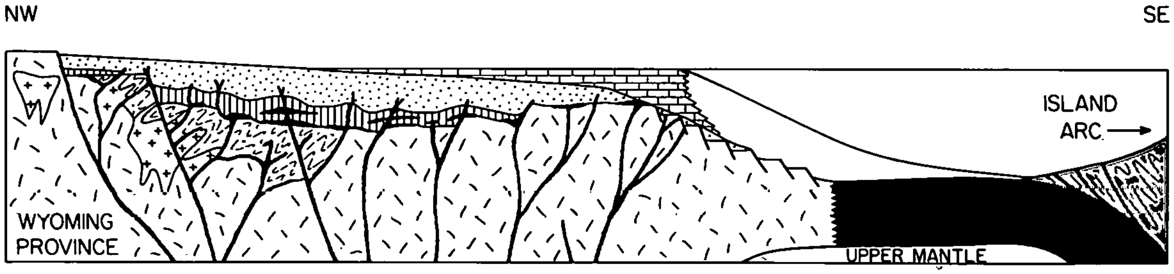
A. LATE ARCHEAN OROGENESIS ~2500 m.y.: SOUTHEAST VERGING DEFORMATION IN PHANTOM LAKE SUITE (F_1); EMPLACEMENT OF SYNOROGENIC GRANITE SILLS AND PHACOLITHS.



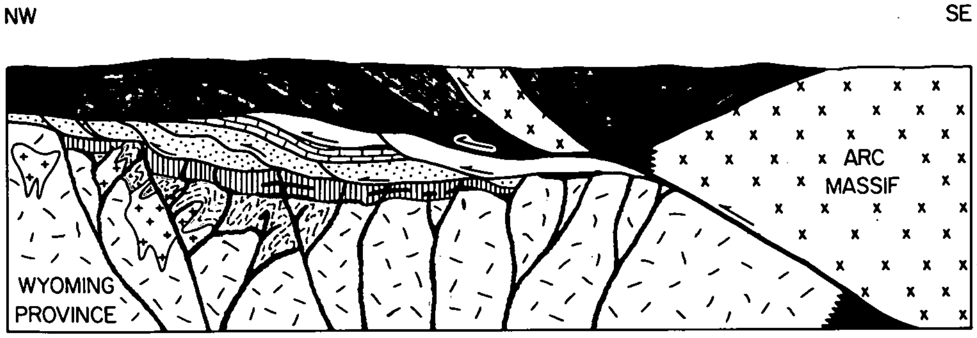
B. RIFT VALLEY SYSTEM ~2300 m.y.: UPLIFT AND RIFTING OF STABILIZED ARCHEAN-CRATON; FLUVIAL DEPOSITION OF LOWER DEEP LAKE GROUP, INCLUDING RADIOACTIVE CONGLOMERATE.



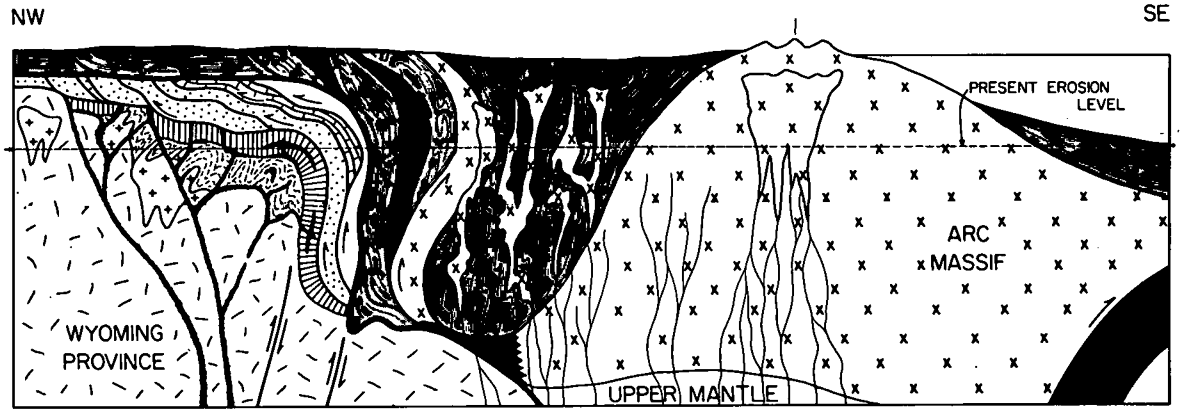
C. PROTO-OCEANIC GULF ~2200 m.y.: DELTAIC DEPOSITION OF UPPER DEEP LAKE AND LOWER LIBBY CREEK GROUPS; FOLDING OF DEEP LAKE GROUP (F_2); INTRUSION OF THOLEIITIC SILLS.



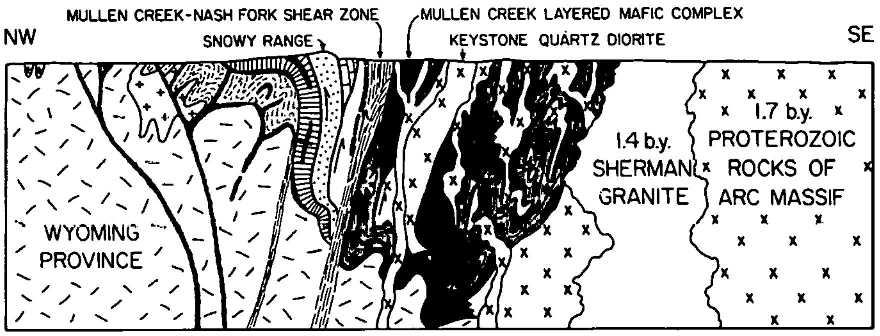
D. OPEN OCEAN ~2000 m.y.: CARBONATE/SHALE DEPOSITION IN UPPER LIBBY CREEK GROUP; APPROACH OF ISLAND ARC FROM SOUTH.



E. CONTINENT-ISLAND ARC COLLISION ~1800 m.y.: OBDUCTION OF ISLAND ARC; THRUSTING IN LIBBY CREEK GROUP AND FOREARC VOLCANOGENIC ROCKS.

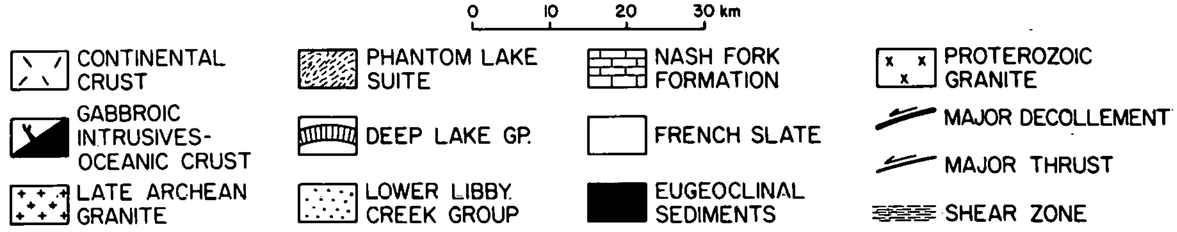


F. FLIP IN DIRECTION OF SUBDUCTION ~1700 m.y.: ROTATION OF BEDS AND THRUSTS; CONTINUED THRUSTING; F_1 AND F_2 FOLDING IN LIBBY CREEK GROUP; F_3 AND F_4 FOLDING IN DEEP LAKE GROUP AND PHANTOM LAKE SUITE; MAXIMUM METAMORPHISM.



G. STRIKE SLIP MOVEMENT ON SHEAR ZONE ~1600 m.y.; INTRUSION OF POST-OROGENIC SHERMAN GRANITE ~1400 m.y.; UPLIFT DURING LARAMIDE OROGENY ~70 m.y.; EROSION TO PRESENT TOPOGRAPHY.

FIGURE 2.27. PLATE TECTONIC MODEL FOR THE PRECAMBRIAN TECTONIC HISTORY OF THE MEDICINE BOW MOUNTAINS, WYOMING.



represent a later (1700 m.y.) intrusive event associated with the Proterozoic collisional orogeny because many of these gabbroic bodies postdate the F_1 folding of the Libby Creek Group which we interpret to have taken place during this orogeny (Figure 2.26). In the absence of supporting geochronological and geochemical data, this interpretation remains speculative. From regional considerations, it seems plausible to equate the gabbroic sills in the Medicine Bow Mountains to the similar, 2150 m.y. old, Nipissing Diabase of the southern Canadian Shield and to the 2100 m.y. Blue Draw Metagabbro of the Black Hills (Redden, 1980). All of these may be related to the same rifting event along the southern margin of proto-North America. 2100 m.y. mafic intrusions have also been reported from the Bighorn Mountains (Stueber and others, 1976) and the Beartooth Mountains (Reid and others, 1975) suggesting that this may have been a regional event even away from the southern continental margin.

Sedimentary evidence form the Libby Creek Group, discussed earlier, suggests that the Libby Creek Group sequence was deposited almost entirely in shallow marine depositional environments (Figures 2.22C and D). The Lower Libby Creek Group contains glaciomarine sediments and aluminous quartzites which have been interpreted to be correlative to similar units in the upper Huronian Supergroup of Canada (Young, 1973; Houston and others, 1979; Graff, 1979; Houston and Karlstrom, 1980) while the Upper Libby Creek Group contains stromatolitic dolomite, mafic volcanics, slates, and some iron formation and is similar to rocks of the Marquette Range Supergroup. We believe that the Libby Creek Group represents a period of miogeoclinal deposition in southern Wyoming comparable to the time span represented in the upper Huronian and lower Marquette Range

Supergroups combined (i.e. 2200-1900 m.y.). This idea is supported by a preliminary date of 2000 m.y. on the Gaps Trondhjemite (Hedge, personal communication, 1980) which intrudes the Lower Libby Creek Group but is not known to crosscut the Upper Libby Creek Group. This relationship, combined with the rapid change in the character of the sediments, local fault breccia along an otherwise poorly exposed contact, and analogy to the Great Lakes area suggest that the Lower and Upper Libby Creek Groups are separated by a major thrust fault (Plate 3) and that the Upper Libby Creek Group is allochthonous or para-autochthonous with respect to the Lower Libby Creek Group. The tectonic position of the Gaps Trondhjemite appears to substantiate this interpretation. It crops out as several small bodies (up to 30,000 square meters) which were intruded into the Gap fault (Houston and others, 1968) and were later sheared and brecciated by recurring movements of the fault. The Gap fault appears to be older than the thrust faulting because it terminates, both to the north and south, against major reverse faults (Plate 3). This supports the interpretation of a major thrust fault which brought Upper Libby Creek Group rocks up against Lower Libby Creek Group rocks sometime after 2000 m.y. ago (Figure 2.26 and Table 2.27). The tectonic significance of the Gaps Trondhjemite is not well understood but it appears to be temporally related to emplacement of nearby metagabbroic rocks and its composition and chemistry are similar to plagiogranites found associated with mafic intrusions near the mid-Atlantic ridge (Coleman and Peterman, 1975) so we postulate that it was emplaced in late rifting stages.

ISLAND ARC DEVELOPMENT TO THE SOUTH

The geology of the rocks south of the shear zone is somewhat less well understood than that of the rocks to the north because of the higher degree of deformation and metamorphism. A chronology of events is more difficult to establish because many geologic processes appear to have been going on nearly simultaneously within a relatively narrow bracket of Precambrian geologic time (1900-1600 m.y.). We feel that this wide variety of geologic events and the resulting heterogeneity of rock assemblages can best be explained in terms of development of one or more island arcs--where volcanism, volcaniclastic sedimentation, and calc-alkaline plutonism would be occurring nearly simultaneously. Another explanation for the rather narrow range of radiometric dates south of the shear zone, of course, would be that a 1800-1700 m.y. metamorphic episode obliterated traces of an earlier geologic history. However, we find this unconvincing in light of the vast number of radiometric dates available from Precambrian rocks in Colorado, New Mexico and Arizona and the absence of any date older than about 1800 m.y. (Peterman and others, 1968; Silver, 1977).

The oldest rocks south of the shear zone in the Medicine Bow Mountains appear to be a series of hornblende- and plagioclase-rich gneisses which are compositionally similar to basalts, andesites and graywackes. We interpret these rocks to be metamorphosed volcanics and volcaniclastic sediments. These gneisses yield ages of up to 1800-1900 m.y. (Hills and Houston, 1979) and are intruded by both the Keystone quartz diorite and the Lake Owen layered mafic complex which are believed to be about 1800 m.y. old from analogy to similar rocks in the southern Sierra Madre (see

Table 2.27). All of these units are, in turn, crosscut by synorogenic granites which yield ages of about 1730 m.y. (Hills and Houston, 1979). The quartz diorite and the granitic plutons are interpreted to be parts of an island arc batholithic complex; the layered mafic complex, and three similar layered complexes to the west, are interpreted to be pieces of dismembered oceanic crust, perhaps from the floor of a marginal basin, which were obducted during the island arc collision.

COLLISION OF CONTINENT AND ISLAND ARC--OBDUCTION AND ACCRETION OF THE ISLAND ARC(S) ONTO THE NORTH AMERICAN PROTOCONTINENT

The rest of the structural events shown in Figure 2.26 with the exception of post-orogenic igneous activity, are interpreted to be related to collision and obduction of an island arc terrane to the south margin of the Wyoming Province about 1700-1800 m.y. ago. The most compelling evidence for such an interpretation is the profound geologic and geochronologic discontinuity across the Mullen Creek-Nash Fork shear zone, a zone of cataclastic rocks from one to seven kilometers wide in the central Medicine Bow Mountains (Houston and McCallum, 1961; Houston and others, 1968).

The first effect of the orogeny, recorded in the metasedimentary rocks north of the shear zone, appears to have been sub-horizontal thrust faulting in the miogeoclinal Libby Creek Group (Figure 2.27E). This interpretation is somewhat open to question because all major faults in the Libby Creek Group are now sub-vertical. However, the patterns of tectonic attenuation seen in map pattern, where the Medicine Peak and Sugarloaf Quartzites completely disappear, then reappear, along strike, are reminiscent of cross sections parallel to major thrust belts and would seem to be much easier to accomplish with low-angle than high-angle faults.

To rotate the thrusts to their present vertical orientation (Figure 2.27G), we envision a flip in direction of subduction during the collision (Figure 2.27F). During the initial stages of the orogeny, the subduction zone dipped to the south, as indicated by the abundant synorogenic granitic intrusives south of the Mullen Creek-Nash Fork shear zone and the absence of such intrusives to the north. However, it is widely believed that continental crust cannot be subducted to any substantial extent (McKenzie, 1969; Dewey and Bird, 1970) so at some stage in the collision, continued compressive stresses would have to be accommodated by initiation of a new subduction zone to the south, perhaps accompanied by unsuccessful attempts to subduct the southern block beneath the northern block (Figure 2.27F). This situation, and resulting rotation of structural elements, is described by Roeder (1973) as a common occurrence in orogenic zones and a mechanism for formation of Alpine-type root zones.

We do not have good control on how much movement may have been associated with the thrust faults within the Libby Creek Group, but several lines of evidence suggest that transport distances were not huge. Depositional environments in the upper Deep Lake Group and Lower Libby Creek Group were both glaciomarine so that the Reservoir Lake Fault does not juxtapose vastly different sedimentary sections; similarly the Lewis Lake Fault which separates the Sugarloaf and Nash Fork Formations brings shallow marine dolomites and deeper water slates against deltaic quartzites. This need not represent large lateral movements although our regional correlation suggests the Lower Libby Creek Group may correlate with the

Huronian Supergroup and the Upper Libby Creek Group has more similarities to the Marquette Range Supergroup.

In contrast, we visualize very large (hundreds of kilometers) lateral transport distances across a master decollement (now represented by the Mullen Creek-Nash Fork shear zone) during the initial obduction of the island arc and underthrusting of the continental block (Figure 2.27E). This thrusting resulted in juxtaposition of rocks from different crustal levels, as shown by the differences in deformation and metamorphic facies across the shear zone (Hills and Houston, 1979).

Geochronological data from the southern Wyoming Province appears to permit (if not support) the concept of large lateral transport distances across a master decollement. Peterman and Hildreth (1978) presented evidence for a geochronologic boundary in the northern Laramie Mountains and the Granite Mountains which separated a zone of Archean (primary) K-Ar mineral dates in the core of the Wyoming Province from a 150 km wide zone of 1800-1400 reset dates near the southern margin of the Wyoming Province and they suggested that the reset dates represent the time when Precambrian rocks which had been buried to depths of more than 10 km were elevated and cooled through the 350 degree isotherm necessary to reset the K-Ar system. Hills and Houston (1979) used this evidence to postulate that burial of this outer zone of the Wyoming Province took place during obduction of the island arc and partial subduction of the craton. This implies at least 150 kilometers of thrusting across the shear zone. This amount of thrusting of crystalline rocks is the same order of magnitude postulated for the Piedmont Province of the Appalachians, based on recent COCORP seismic data (Cooke and others, 1979). In fact, we visualize many similarities between the Medicine Bow orogeny and the Taconic orogeny of the southern Appalachians except that, in the Medicine Bow Mountains we are now looking at much deeper crustal level corresponding to a view of the southern Appalachians with the upper 10 km or so eroded off; hence without the foreland fold and thrust belt corresponding to the Valley and Ridge Province (Figure 2.27E, F, G). In this context, the geologic relationships seen in the Medicine Bow Mountains may be an example of a deeply eroded orogenic root zone or suture zone.

Isostatic uplift of the partially subducted continental crust and the overlying obducted volcanogenic sequences and foreland thrust belt would have taken place over several hundred million years (1700-1400 m.y.) and possibly provided a source of sediment for the Late Precambrian Uinta Mountain Group to the west (Peterman and Hildreth, 1978). If this interpretation is valid, we are now looking at an island arc-continent suture as it appears at depths of about ten kilometers, as estimated by considering the thickness of obducted material (overburden) necessary to raise temperatures to the 350 degree C. temperatures necessary to open the K-Ar isotopic system, assuming normal geothermal gradients (Peterman and Hildreth, 1978). This estimate of 10 kilometers agrees with estimates of temperature and pressure conditions of metamorphism in the Libby Creek Group. Flurkey (1981) observed co-existing kyanite and pyrophyllite in the Medicine Peak Quartzite which, from thermodynamic considerations, indicate minimum pressures of 3 kb and temperatures of 425 degrees C--which, under normal geothermal gradients, correspond to a depth of about 10 km.

There are several generations of folding which we interpret to be related to this collisional orogeny--indicating a suitably complex structural history for the orogenesis. F_3 folding in the Phantom Lake Suite and Deep Lake Group may reflect stresses transmitted through the sedimentary pile during collision and thrusting and concentration of folding and strain near the northern edge of the sedimentary basin--adjacent to a massive, anisotropic granitic basement block. By this interpretation, F_3 and F_4 (the northwest warping of F_2 folds in the Deep Lake Group) as well as F_2 (northwest warping in the Libby Creek Group) all represent northeast-southwest directed compressive stresses caused by oblique collision.

The last movements along the suture zone would then represent a change from thrusting to strike-slip movement (Figure 2.27G), again in response to accentuation of the oblique components of the collision. The folds generated during this episode, F_3 in the Libby Creek Group, are characterized by a vertical fold axis and vertical rotation of pre-existing fabric elements. It is significant that F_3 is also present in Archean gneisses near the suture zone indicating that the basement was plastically deformed during the orogeny, as would be expected at these deep crustal levels. Even more significant is the fact that rocks within the Mullen Creek-Nash Fork shear zone and rocks south of the shear zone exhibit the same distribution of s-surfaces seen in the Upper Libby Creek Group. We interpret this as strong evidence that all of these rocks were deformed together during the island arc collision.

The timing of the main orogenic episode in southern Wyoming (Figure 2.26) is well constrained by geochronologic data (see Hills and Houston, 1979, for a summary of available radiometric dates). Major strike-slip movement on the shear zone is bracketed between 1730, the age of granites which are cut by the shear zone and 1640 m.y., the age of a granite which intrudes the shear zone (Hills and Houston, 1979); major metamorphism in the Libby Creek Group took place around 1710 m.y. ago (Hills and others, 1968); and major regional metamorphism in Colorado took place around 1710 m.y. ago (Hedge and others, 1967).

POST- OROGENIC IGNEOUS ACTIVITY

The last major igneous activity in the Medicine Bow Mountains was intrusion of the Sherman Granite at about 1385 m.y. This event was post-orogenic in the sense that it was not accompanied by regional deformation and metamorphism. This activity was part of a regional "anorogenic" event (Emslie, 1978; Silver and others, 1977; Bridgwater and Windley, 1973) which affected much of the southwestern United States. In the Laramie Mountains, intrusion of the Sherman Granite appears to have occurred at about the same time as emplacement of the Laramie anorthosite (Figure 2.26). This leads us to speculate that the granites, and the nearly contemporaneous anorthosite, may be a deep level analog of the bimodal basalt-rhyolite association observed in the American Cordillera which is believed (Lipman and others, 1971, 1972; Snyder and others, 1976) to reflect a change from subduction-related volcanism to volcanism related to extensional tectonics. This is consistent with Emslie's (1978) interpretation of Late Precambrian rifting in North America.

REGIONAL TECTONIC MODELS

CONTINENTAL ACCRETION

The concept of continental accretion for the North American continent has had both advocates (Engle, 1963) and critics (Sims, 1976). Whether or not the concept has merit over all of geologic time, we believe that the model presented herein is strong evidence in support of continental accretion of North America from 1800 to 1400 m.y. One of the most important lines of evidence for island-arc-continent collision in southern Wyoming, rather than continent-continent collision or some mechanism of intracratonic deformation, is that no rocks older than about 1900 m.y. are known south of the proposed suture zone. This applies to Precambrian rocks in Colorado, New Mexico, Arizona, southern California, and northern Sonora and it suggests that the entirety of this Precambrian terrane was added on to the North American proto continent during a series of continent-island arc collisions in the Early Proterozoic, much in the same way as the western Cordillera has been added on to North America in the Mesozoic and Cenozoic. This implies that the southwestern Precambrian terrane ought to exhibit extreme structural complexities due to superposed compressional and transform (microplate) tectonics and that the zone ought to contain cryptic sutures, parallel to the continental margin (northeast) which get progressively younger to the south.

Warner (1978) coined the term "Colorado Lineament" for a northeast-trending zone of faults which transects the southwestern Precambrian (and post-Precambrian terrane) and is bounded on the north by the Mullen Creek-Nash Fork shear zone. He interpreted the lineament to be a major wrench fault system in North America, analogous to the San Andreas system, which formed adjacent to the southeastern margin of proto-North America. We agree with Warner's interpretation--that the Colorado Lineament is a major crustal feature in the western North American Precambrian, that the zone formed near the margin of Proterozoic North America, and that the zone is now characterized by northeast-trending faults which exhibit strike-slip movement. However, we believe that the Colorado Lineament represents compressional tectonics (thrusting) fundamentally and transform motion (wrench faulting) only secondarily. We view the Colorado Lineament as a zone of accreted oceanic and island arc-type crust which became continental crust during 1800-1600 m.y. collisional orogenies and we view the major shear zones within the Colorado Lineament such as the Colorado Mineral Belt to be the probable locations of cryptic sutures which later experienced transform movement.

EARLY PROTEROZOIC PLATE TECTONICS

Another question widely debated among Precambrian geologists is how far back in the earth's history plate tectonic processes operated (Dewey and Spall, 1975; Windley, 1977). Hoffman's (1973) work in the Coronation Geosyncline of the Slave Province is widely cited as evidence that the Wilson Cycle of the generation of geosynclines and orogenic belts by the opening and closing of ocean basins (Dewey and Spall, 1975) was operating in North America by about 1700 m.y. ago. Our model for southern Wyoming supports such an interpretation and raises some questions about the regional tectonic picture around 1700 m.y. ago.

From a regional viewpoint, one of the most conspicuous and important geologic "events" in the history of the North American continent was the widespread orogenesis which took place about 1800-1700 m.y. ago. The scale of this orogenic event is reminiscent of the Archean tectonothermal event at 2500 m.y. (Houston and Karlstrom, 1980) except that by 1700 m.y. deformation was more strongly located into linear or arcuate zones.

There appear to be two main types of orogenies during this 1700 m.y. "event". The first type, which affected by far the largest areas, involved "basement reactivation" (Dewey and Burke, 1973), where Archean basement terranes were deformed and metamorphosed strongly enough to obliterate Archean structural trends and nearly completely reset isotopic systems. Examples of this type of orogeny are the Hudsonian orogeny of the Churchill Province (Davidson, 1972) and the Nagssutoquidian mobile belt of Greenland (Bridgwater and others, 1973). Within the Churchill Province or near its margins are a variety of more linear orogenic or geocynclinal zones (Figure 2.24) such as the Wollaston Lake Foldbelt (Ray and Wanless, in prep.), the Circum-Ungava Geosyncline (Dimroth and others, 1970), the Nelson Front (Gibb, 1968; Davidson, 1972), the Thelon Front (Wright, 1967; Gibb and Thomas, 1977), and the Black Hills orogenic belt (Goldich and others, 1966). Plate tectonic models have been proposed for each of these linear orogenic zones (see references cited above) and for the Churchill Province as a whole (Dewey and Burke, 1973).

The second type of orogeny in the 1700 m.y. event involved proven Archean crust on only one side of the belt. These have been interpreted to be continent-island arc collisions: the Coronation Geosyncline of the Slave Province (Hoffman and others, 1974), the Medicine Bow orogeny in southern Wyoming (this paper and Hills and Houston, 1979), and the Ketilidian mobile belt of southern Greenland (Bridgwater and others, 1973). This second type of orogeny and its possible regional significance has not been fully appreciated because the interpretation of the Medicine Bow orogeny as a plate tectonic phenomenon is relatively recent (first reported by Hills and Armstrong, 1974) and reconstructions of proto-North

America rarely consider a combined Greenland and North America (Figure 2.24) which almost certainly were together in the Archean, and probably the Early Proterozoic, because of the geologic continuity of the Nain Province and the North Atlantic craton (Hurst and others, 1975).

We have presented earlier an interpretation, based on lithostratigraphic correlation of sedimentary sequences in North America, that the Early Proterozoic North America was similar in size to the Precambrian nucleus of the continent seen today (Figure 2.24). Thus, we believe that the widespread nature of 1700 m.y. orogenesis and the localization of deformation along linear and arcuate zones have two rather obvious implications. First, the entire North American proto-continent appears to have been in a compressional stress field during the orogenic interval (1800-1600 m.y.). Such a tectonic regime could have been produced, but was certainly manifested, by collision of island arcs with the present west, south and east margins of the proto-continent (Figure 2.24). Second, deformation was concentrated along the boundaries of cratonic blocks. These boundaries probably mainly took the form of incipient rifts and small ocean basins and rifts, and collision of small intra-plate cratonic blocks in response to a general compressive regime in North America. This idea is similar to Dewey and Burke's (1973) basement reactivation hypothesis except we do not envision large movements between cratonic blocks, In keeping with Irving's (1979) assessment that, to date, Precambrian paleomagnetic data from North America are most simply interpreted in terms of a single polar wandering path for Proterozoic North America.

SUMMARY

Precambrian metasedimentary rocks of the northern Medicine Bow Mountains, Wyoming preserve a record of over a billion years of geologic history (>2700 to 1700 m.y.). The oldest unit, the Stage Crossing Gneiss, contains hornblende schists and gneisses of probable volcanic origin, which are in fault contact with structurally overlying rocks of the Phantom Lake Metamorphic Suite. The Phantom Lake Metamorphic Suite is more than 3 km thick and contains tightly folded and complexly faulted amphibolite facies metavolcanic rocks (60%) and metasedimentary rocks (40%). Field relationships and sedimentary structures permit a tentative stratigraphy for the Phantom Lake Metamorphic Suite as follows: Stud Creek Volcaniclastic Rocks (metatuffs?), Rock Mountain Conglomerate (arkosic paraconglomerate), Bow River Quartzite (fine-grained quartzite), Colberg Metavolcanic Rocks (metabasalt, paraconglomerate), and Conical Peak Quartzite (fine-grained quartzite). The Stage Crossing Gneiss and the lower Phantom Lake Metamorphic Suite units are intruded by Late Archean (?) granitic rocks.

The Proterozoic Deep Lake Group unconformably overlies various units of the Phantom Lake Metamorphic Suite and the Late Archean (?) granite. The Deep Lake Group (>2.5 km thick) is divided into the Magnolia Formation (radioactive quartz-pebble conglomerate), Lindsey Quartzite (trough crossbedded quartzite), Campbell Lake Formation (paraconglomerate, phyllite), Cascade Quartzite (pebbly quartzite), and Vagner Formation (paraconglomerate, marble).

The Libby Creek Group overlies the Deep Lake Group and Archean (>2500 m.y.) quartzo-feldspathic gneisses. The lower contact may originally have been an unconformity but breccias, truncated units, changes in strike and dip of beds, and topographic scarps indicate faulting, with appreciable but unknown reverse movement. The Libby Creek Group (7.5 km thick) is divided into a 4.5 km thick lower part (Lower Libby Creek Group) containing quartzites and a 3 km thick upper part (Upper Libby Creek Group) containing dolomite and slate, separated by a second major reverse fault (magnitude of displacement unknown) which is evidenced by the abrupt lateral disappearance, and re-appearance, of the 580 m thick Sugarloaf Quartzite. The Upper Libby Creek Group is truncated above by the Mullen Creek-Nash Fork shear zone.

Quartzites of the Phantom Lake Metamorphic Suite are mainly shallow marine, as indicated by large-scale planar crossbeds, ripple marks, fine-grain sizes, lithofacies patterns, and NE-SW bipolar paleocurrents. Volcanogenic rocks are believed to be mainly subaerial, as suggested by the absence of pillow basalts and rapid facies variations.

Proterozoic sedimentation was dominated by mature, quartz-rich clastic sedimentation on a miogeoclinal platform under stable tectonic conditions. Deposition was generally transgressive. Fluvial, then deltaic sedimentation of the Deep Lake Group and Lower Libby Creek Group took place on a persistent southwest paleoslope, with sediment transport subparallel to the inferred south boundary of the Wyoming Archean craton. This, and a few west-directed paleocurrents, suggest a southern highland bounding fluvial and deltaic sedimentation. We postulate intracratonic rifting as a tectonic setting for Deep Lake Group and Lower Libby Creek Group deposition. The Upper Libby Creek Group represents open marine conditions, presumably following rift-separation of two continental blocks. Tholeiitic sills intrusive into the Phantom Lake Metamorphic Suite, Deep Lake Group, and Libby Creek Group are also thought to be rift-related, as is the 2000 m.y. old Gap Trondhjemite which intrudes the Lower Libby Creek Group.

Regional lithostratigraphic correlations suggest similar tectonic settings and ages (?) for the Deep Lake Group-Lower Libby Creek Group and the Huronian Supergroup of Ontario; and for the Upper Libby Creek Group and Marquette Range Supergroup and equivalent rocks of the Lake Superior region. We suggest that Early Proterozoic metasedimentary successions all along the southern boundary of the Archean nucleus of North America are rift-related; and we suggest a rift-valley model for deposition of Early Proterozoic radioactive quartz-pebble conglomerates.

Geometrical analysis of folding shows different tectonic styles and histories for the Phantom Lake Metamorphic Suite, Deep Lake Group, and Libby Creek Group. The Phantom Lake Metamorphic Suite was folded into tight to isoclinal anticlinoria and synclinoria (F_1) prior to Deep Lake Group deposition; the Deep Lake Group was folded into concentric, doubly plunging anticlines and synclines (F_2) prior to intrusion of tholeiitic sills; and the Libby Creek Group forms a northeast-striking, steeply south-dipping homoclinal succession. Highly deformed eugeoclinal rocks south of the Mullen Creek-Nash Fork shear zone contain NE-striking, subvertical foliation, parallel to bedding in the Libby Creek Group, suggesting that beds in the Libby Creek Group were rotated to vertical attitudes during the same deformation which formed foliation in rocks south of the shear zone.

Hills and Houston (1979) proposed a 1700 m.y. continent-island arc collision to explain geologic and geochronologic discontinuities across the Mullen Creek-Nash Fork shear zone. Structural data from metasedimentary rocks support this interpretation. F_1 folding of the Phantom

Lake Metamorphic Suite is believed to be an Archean deformation and F_2 folding of the Deep Lake Group probably took place in early stages of rifting. Major reverse faults separating the Deep Lake Group from the Lower Libby Creek Group and the Lower Libby Creek Group from the Upper Libby Creek Group are interpreted to be rotated thrust faults. This is supported by differing tectonic styles between the autocthonous Deep Lake Group and the para-autochthonous or allocthonous Libby Creek Group, and by transverse faults and mafic intrusive bodies which appear to be truncated against the "thrusts".

We propose that subhorizontal thrusting in the miogeocline took place in early stages of the collisional orogenesis; that the Mullen Creek-Nash Fork shear zone represents a master decollement that brought island arc rocks over miogeoclinal rocks; and that the shear zone and bedding and thrusts in the Libby Creek Group were rotated to steep attitudes during changes in direction of subduction during collision of two buoyant blocks. Broad, northwest folds (F_4 in the Deep Lake Group, F_2 in the Libby Creek Group) of bedding, "thrust" faults, and earlier fold traces, and vertical rotation (F_5 in the Deep Lake Group, F_3 in the Libby Creek Group) of pre-existing structures, are interpreted to represent stresses generated in later stages of an oblique collision.

URANIUM-BEARING CONGLOMERATES OF THE MEDICINE BOW MOUNTAINS

Numerous stratigraphic horizons within the metasedimentary succession in the Medicine Bow Mountains contain anomalously radioactive rocks. These include the Stud Creek Volcaniclastic Rocks, Rock Mountain Conglomerate, Magnolia Formation, Lindsey Quartzite, Cascade Quartzite, and Medicine Peak Quartzite (Figure 2.2). This section discusses these radioactive rocks, from oldest to youngest. The Magnolia Formation is the only one with significant volumes of radioactive conglomerate. Thus, the majority of this section focuses on the sedimentological, petrographic, and geochemical characteristics of the Magnolia Formation in the Medicine Bow Mountains. At the end of this section we compare the Magnolia conglomerates to other occurrences of Precambrian uranium-bearing fossil placers in the world and we discuss the geologic setting and history of the Magnolia Formation in the context of genetic models for the formation of Precambrian uranium-bearing quartz-pebble conglomerates (Houston and Karlstrom, 1980).

PHANTOM LAKE METAMORPHIC SUITE

There are two units in the Phantom Lake Metamorphic Suite which contain anomalously radioactive conglomerates: the Stud Creek Volcaniclastic Rocks and the Rock Mountain Conglomerate. Conglomerates in the Stud Creek Volcaniclastic Rocks occur in Secs. 11 and 15, T. 18 N., R. 79 W. as lenticular outcrops of granular quartzite, quartz-granule conglomerates and isolated outcrops of arkosic paraconglomerate. All are interleaved with amphibolites and pelitic schists. These conglomerates register two to three times background gamma radiation in outcrop and have yielded values up to 21 ppm U and 57 ppm Th. We interpret these conglomerates as thin fluvial sands, gravels, and occasional alluvial debris flows which were deposited at about the same time as the Rock Mountain Conglomerate to the north, but in a depositional setting dominated by presumed subaerial volcanism and volcaniclastic sedimentation, with only occasional fluvial sedimentation. One granular zone in the northern part of Sec. 15 is about 1000 m long and 100 m thick. However, most of the conglomerates are thinner and are too lenticular and low-grade to be of potential economic interest for uranium and thorium.

Rock Mountain Conglomerate

The Rock Mountain Conglomerate crops out in two areas of the northern Medicine Bow Mountains (Figure 2.28 and Plate 1). The western area, near Foote Creek, is a 200 m thick layer of muscovitic quartz-granule conglomerate (with rare boulder paraconglomerate near the base) which is continuous for a lateral distance of 4 km before being faulted out to the north and covered by Phanerozoic rocks to the south. This zone contains assays on the order of 5 ppm U and 8 ppm Th and is of little economic interest.

The eastern outcrop area, near Rock Mountain, is more promising for uranium and thorium. This area contains thick (300-400 m) arkosic paraconglomerates, with stretched clasts of quartzite, quartz and mafic schist interbedded with muscovitic quartz-granule and quartz-pebble conglomerates. Surface assays of paraconglomerates yielded up to 270 ppm U and 57 ppm Th while typical values were about 5-50 ppm U and 5-40 ppm Th.

The Rock Mountain Conglomerate is thick enough near Rock Mountain, and surface values high enough, that two holes were drilled to test subsurface uranium content of the conglomerates (see Volume 2). MB-10 was drilled from the bottom of Rock Creek toward two of the most radioactive outcrops we had found, the lower outcrop on the east slope of Rock Creek yielded up to 97 ppm U and 95 ppm Th and the upper one yielded up to 270 ppm U and 57 ppm Th (Figure 7, Volume 2). The subsurface extension of the beds in the lower outcrop contained maximum values of only 3.9 ppm U and 9 ppm Th. Unfortunately, the hole was terminated before reaching the subsurface extension of beds of the upper outcrop because the intrusive ultra-mafic body which separates the two outcrops

(Figure 7, Volume 2) is wider at depth than on the surface and the drill used was unable to go to the depths (1400-1500 feet) which we believe would have been required to hit the conglomerates.

Drill hole MB-15, on the top of Rock Mountain penetrated about 122 meters of arkosic paraconglomerate. Most assays from the paraconglomerate were less than 10 ppm U and 10 ppm Th. However, one sample yielded 194 ppm U and 86 ppm Th. This sample was directly adjacent to a 15 meter thick metagabbro body.

The structure and the stratigraphy of the Rock Mountain area is exceedingly complicated and very few primary sedimentary features are preserved in the Rock Mountain Conglomerate. Therefore, the sedimentary history of the unit remains somewhat obscure. On the basis of lithology and limited distribution of the unit, we interpret it to be an alluvial fan succession of debris flows and braided stream deposits derived from a northerly source area and which prograded south into the Phantom Lake Suite depositional basin. By this interpretation, the conglomerates of the Stud Creek Volcaniclastic Rocks to the south are distal equivalents of the Rock Mountain Conglomerate. This interpretation also suggests that the Rock Mountain Conglomerate was deposited in fluvial depositional environments which are favorable for the concentration of heavy minerals as placers. Assays of several hundred ppm U in conglomerates suggest that uranium-bearing minerals may be present in placer concentrations in the conglomerates but surface sampling and subsurface information indicate that zones containing these grades of uranium are widely scattered, thin, and lenticular. Therefore, we believe that the Rock Mountain Conglomerate is unlikely to contain appreciable uranium and thorium mineralization, although we can't rule out the possibility of finding subcrops on Rock Mountain containing somewhat higher uranium grades and thicker conglomerate zones.

It is interesting to note that the highest uranium assays in several cases in the Rock Mountain area came from samples without correspondingly high thorium values, (270 ppm U, 57 ppm Th; 97 ppm U, 75 ppm Th; and 190 ppm U, 86 ppm Th from paraconglomerates; and 170 ppm U, 16 ppm Th from calcareous pelitic schist of the Stud Creek Volcaniclastic Rocks). Archean conglomerates in the Sierra Madre are appreciably richer in thorium than uranium because, we believe, Archean gneisses contained more thorium minerals to contribute as detritus for conglomerate placers. Thus, the high uranium content of the Rock Mountain samples seems anomalous. Either the source terrane for the Rock Mountain Conglomerate contained higher uranium values than the Sierra Madre Archean source terrane or the uranium in the Rock Mountain Conglomerate has been remobilized and reconcentrated. The latter seems likely because, as shown in Figure 7 of Volume 2, the most uraniferous samples are all from outcrops near contacts with mafic intrusive bodies. We speculate that uranium was remobilized out of generally low-grade conglomerates during the episode of intrusion of mafic magmas and reconcentrated adjacent to the contacts with mafic intrusives.

MAGNOLIA FORMATION

The Magnolia Formation crops out in isolated areas in a northeast-southwest trending zone (Figure 2.28) extending some 40 km from the Arlington area, in the extreme northeast area of Plate 1 to near Brush Creek, in the southwest part of Plate 1. The Magnolia Formation is subdivided into two members: a lower Conglomerate Member which crops out in discontinuous lenses up to 330 m thick and unconformably overlies folded Phantom

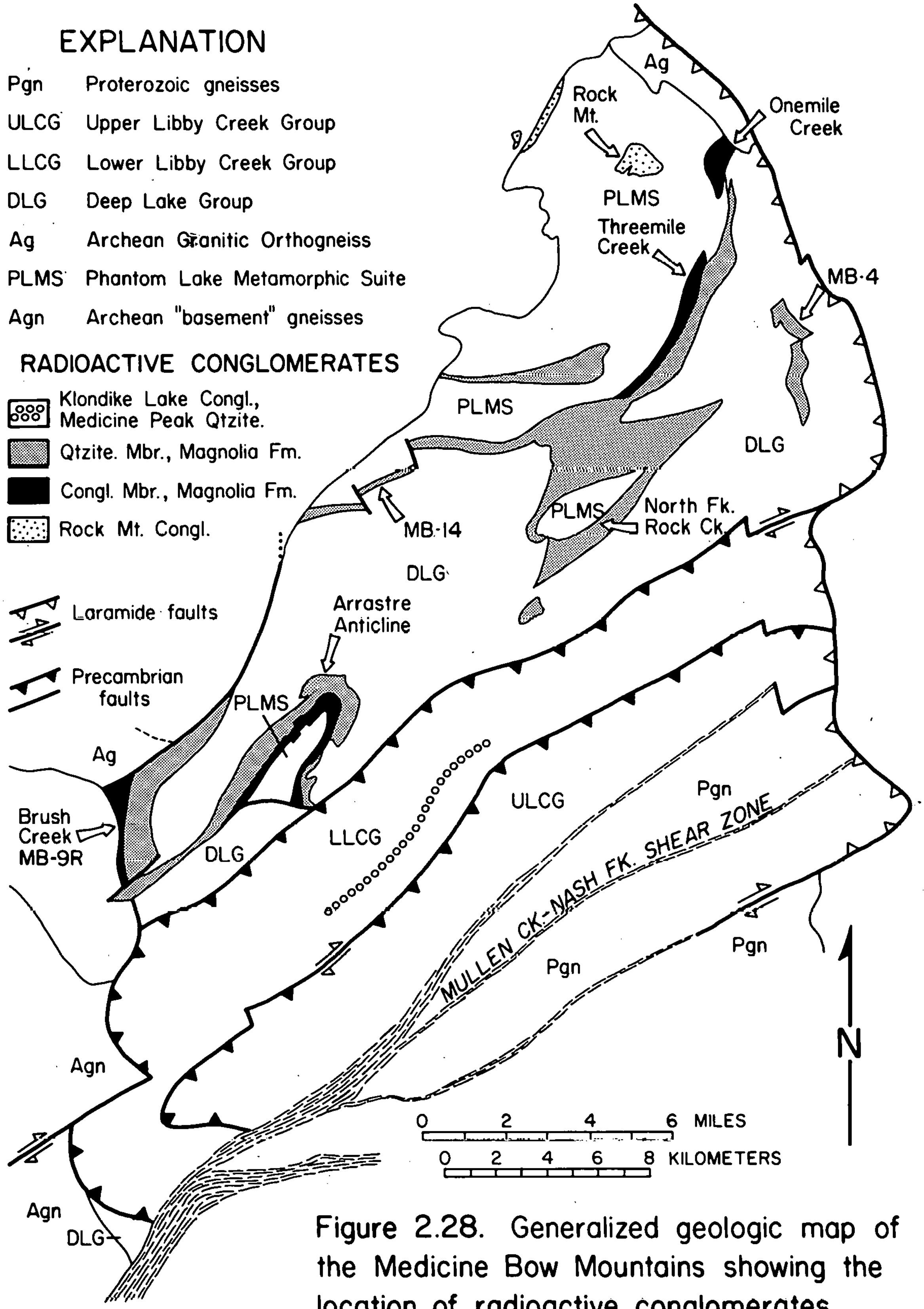


Figure 2.28. Generalized geologic map of the Medicine Bow Mountains showing the location of radioactive conglomerates.

Lake Suite rocks or Archean granite; and an upper Quartzite Member which forms a continuous sheet some 500 m thick which either unconformably overlies Phantom Lake Suite rocks or gradationally overlies the Conglomerate Member.

Figure 2.28 shows the main outcrop areas of the Magnolia Formation and Table 2.9 summarizes the petrography of the unit in various areas. The following discussion of the Magnolia Formation proceeds from northeast to southwest, approximately down the Magnolia paleoslope (Plate 2).

Radioactive Conglomerates of the Onemile Creek Area

The Onemile Creek area in the northern Medicine Bow Mountains (Figure 2.28) contains the most radioactive quartz-pebble conglomerates known within the Magnolia Formation of the Medicine Bow Mountains and is one of two areas in southern Wyoming known to contain Precambrian fossil-placer uranium mineralization of possible economic interest (the other being the Carrico Ranch-Deep Gulch area of the northwestern Sierra Madre). Therefore, an appraisal of the distribution and grade of uranium mineralization in this key area is one cornerstone in evaluating uranium reserves in quartz-pebble conglomerates in southern Wyoming.

This section presents our interpretation of the geologic setting of the radioactive conglomerates -- their areal distribution, structure, and stratigraphy. It also discusses the lithology, distribution of lithofacies, and sedimentary features of the conglomerates, and presents geochemical and radiometric data on the abundance of uranium and thorium in surface and subsurface samples. The result is an interpretation of the extent and variability of uranium and thorium mineralization and an estimate of the subsurface distribution of mineralized zones. This information, ultimately, is used in the resource estimate presented in Volume 3 of this report.

General Stratigraphy of the Onemile Creek Area

As shown in Figure 2.29, the Onemile Creek area contains four major units. An Archean basement terrane containing the Stage Crossing Gneiss, Phantom Lake Metamorphic Suite, and Late Archean granite is overlain with angular unconformity by the Magnolia Formation; the Magnolia is unconformably overlain by the Cascade Quartzite; and everything is cross-cut by gabbroic sills and dikes.

The Archean terrane in this area is represented by hornblende gneiss, metabasalt, quartzite, and granitic gneiss. These units are discussed in detail earlier.

The Magnolia Formation unconformably lies on various units of the basement complex. The angular nature of the unconformity is seen in the area of drill hole EMB-3 (Figure 2.29) where the basal contact of the Magnolia is perpendicular to the strike of bedding in Phantom Lake Suite quartzites. The Magnolia Formation is about 400 m thick in the Onemile Creek area and consists mainly of radioactive, muscovitic, granular quartzite and pyritic pebble conglomerates. Detailed discussion of the lithology and internal stratigraphy of the Magnolia Formation is presented later.

The Cascade Quartzite overlies the Magnolia Formation with disconformity or mild angular unconformity (C-C', Figure 2.30). The Cascade

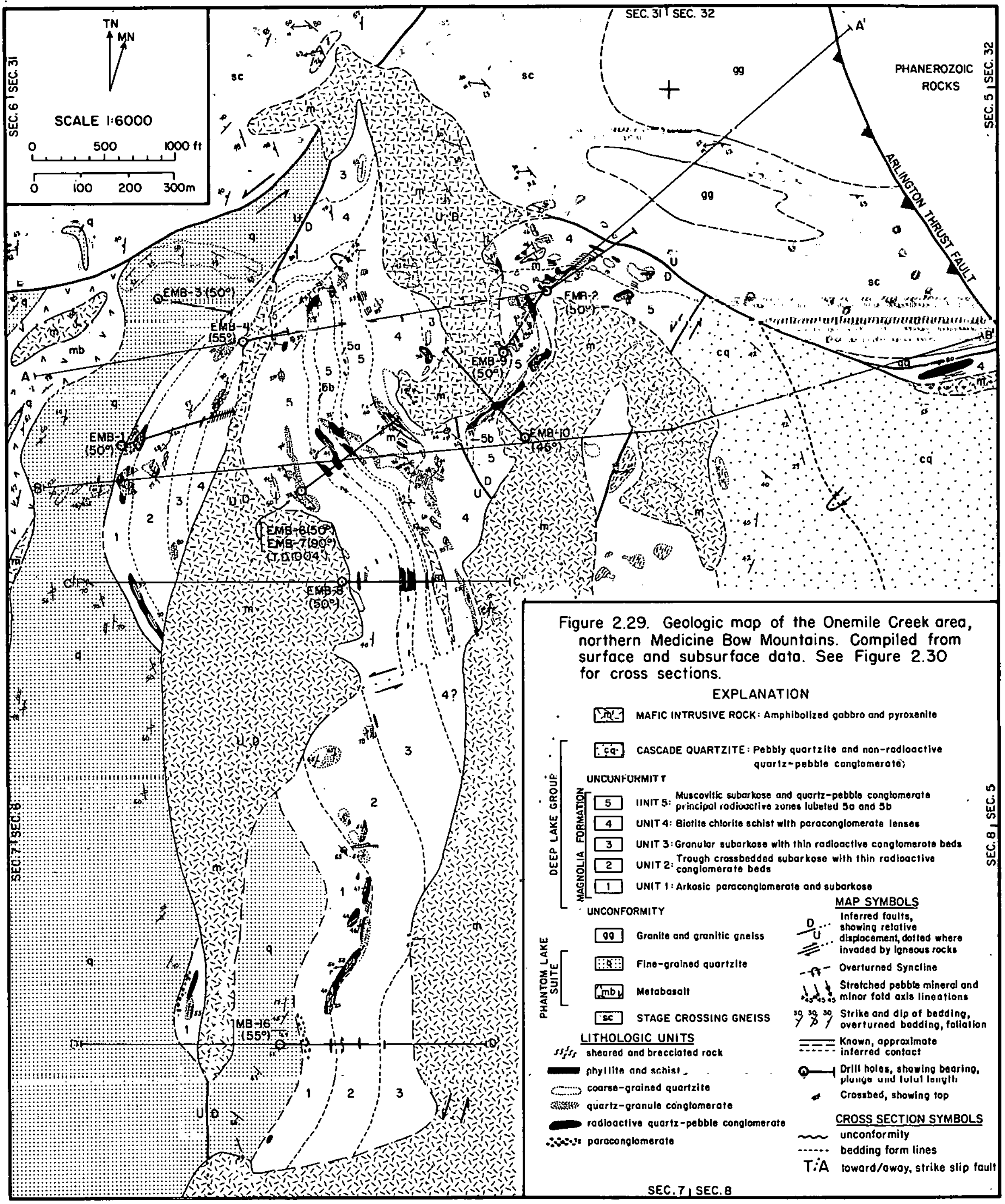


Figure 2.29. Geologic map of the Onemile Creek area, northern Medicine Bow Mountains. Compiled from surface and subsurface data. See Figure 2.30 for cross sections.

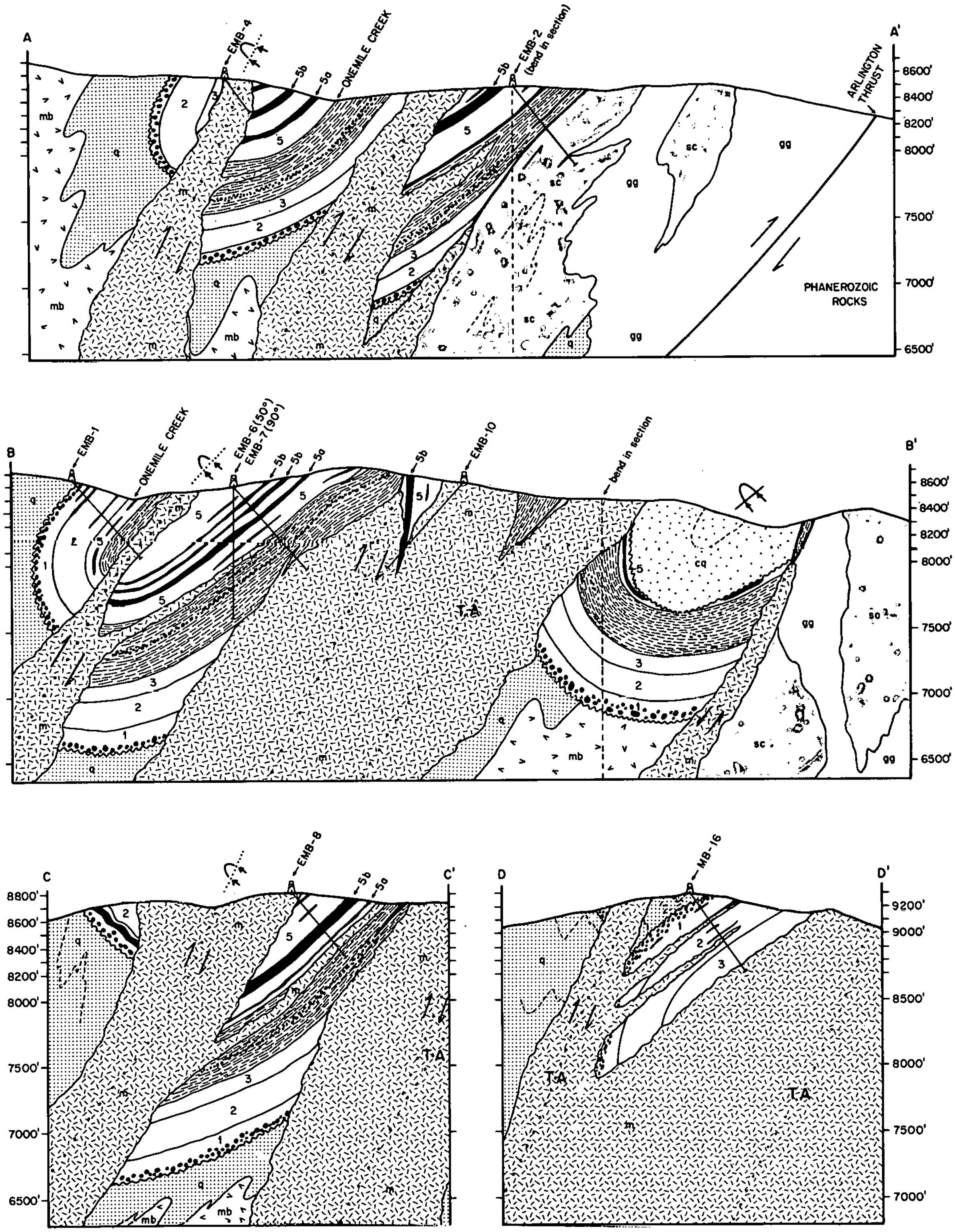


Figure 2.30. Cross sections and drill sections from the Onemile Creek area.

Quartzite is on the order of 400 m thick in the Onemile Creek area and consists of white, massive, medium-grained pebbly quartzite and non-radioactive, non-muscovitic quartz-pebble conglomerates. The unit does not contain the abundant black chert pebbles which characterize the Cascade in the central Medicine Bow Mountains but, otherwise, has similar lithologies.

Amphibolitized mafic intrusive rocks cross-cut older metasedimentary units. These bodies represent gabbroic to ultramafic magmas which were intruded along faults and bedding contacts (Figure 2.30). These rocks are also discussed in detail in earlier sections of the report.

Structure of the Onemile Creek Area

As shown in cross-section A-A' of Figure 2.30, the major structure in the Onemile Creek area is a faulted and intruded overturned syncline in the Magnolia Formation. This syncline is documented by top and bottom critera in the Magnolia quartzites (Figure 2.29). Underlying the folded Magnolia Formation rocks are isoclinally folded rocks of the Phantom Lake Metamorphic Suite which are shown in A-A' of Figure 2.30 to be part of a major synclinorium. Cross sections B-B' and C-C' (Figure 2.30 each show two faulted synclines but, as shown on the geologic map (Figure 2.29), these are interpreted to be the same fold, offset left laterally along a north-northeast-trending fault.

Despite detailed surface mapping and ten drill holes in the Onemile Creek area, there are still uncertainties in the interpretation given above and shown in Figures 2.29 and 2.30. The main structural questions involve the intensity of folding in the Phantom Lake Suite and the style of folding in the Magnolia Formation. The second question is more pertinent to the estimation of the extent of mineralized conglomerates and involves uncertainty regarding the plunge of the major folds in the Magnolia Formation. A-A' of Figure 2.30,and Figure 2.31A show a gently plunging fold axis for the major syncline. This interpretation is based on the macroscopic pattern of the Magnolia Formation which can be traced over a distance of 10 km to the southwest, into an area containing open, upright anticlines and synclines with subhorizontal fold axes.

The only problem with this interpretation is that all the minor fold axes mapped in the Onemile Creek area plunge west. Admittedly, most of these have northwest-striking axial planes and can be explained in terms of a later fold system, which is well documented by stereonet plots of bedding, foliations and lineations (see below). However, some have north-striking axial planes and imply that the major structure may be a reclined syncline, plunging steeply west, instead of an inclined syncline plunging gently south (terminology is that of Turner and Weiss, 1963). Both possibilities are presented in Figure 2.31, the major difference being the plunge of the syncline in the Onemile Creek area and the fact that, if the reclined syncline shown in Figure 2.31B is used, mineralized conglomerates in the Onemile Creek area could extend to appreciably greater depths than are shown in Figure 2.30. We used the inclined syncline interpretation of Figure 2.31A in constructing the cross-sections because it gives a more conservative estimate of the subsurface extent of the conglomerates and because it is the simplest, and most easily visualized explanation of regional relationships.

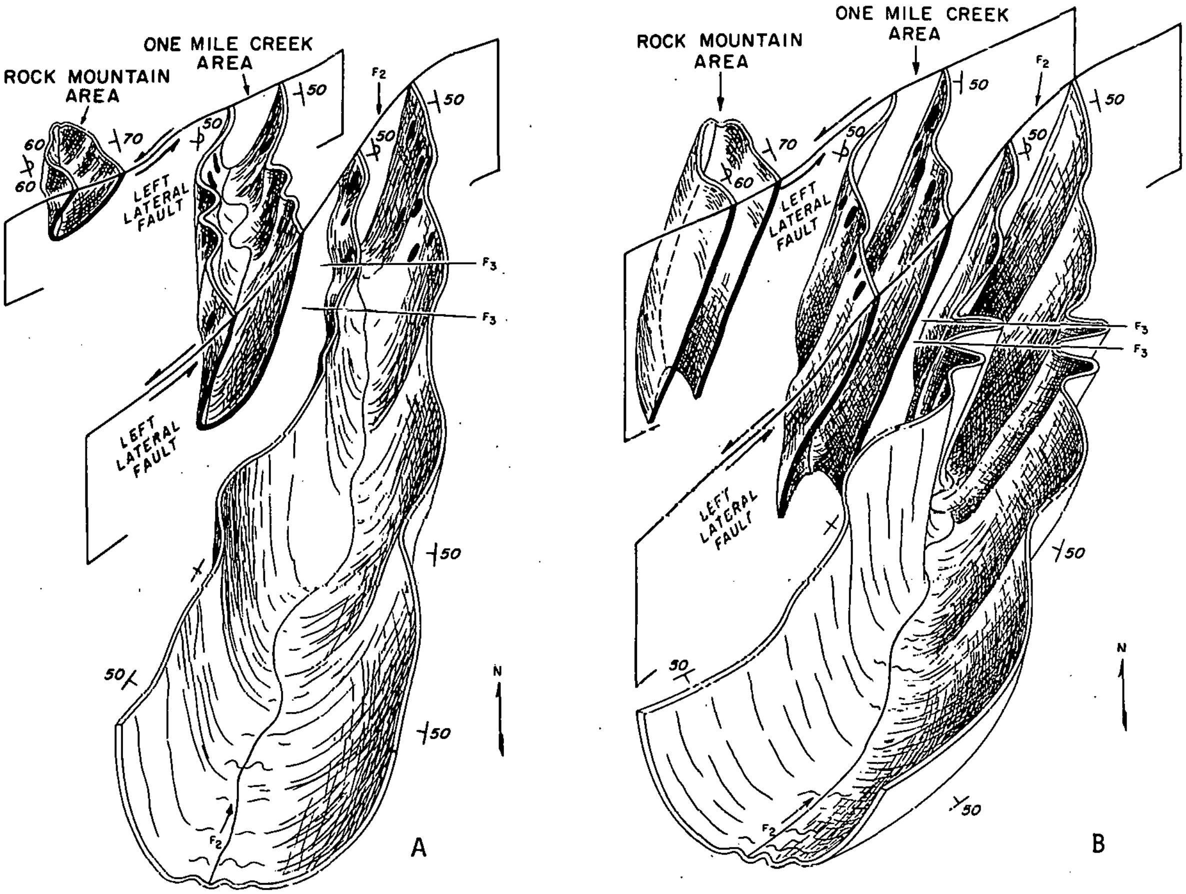


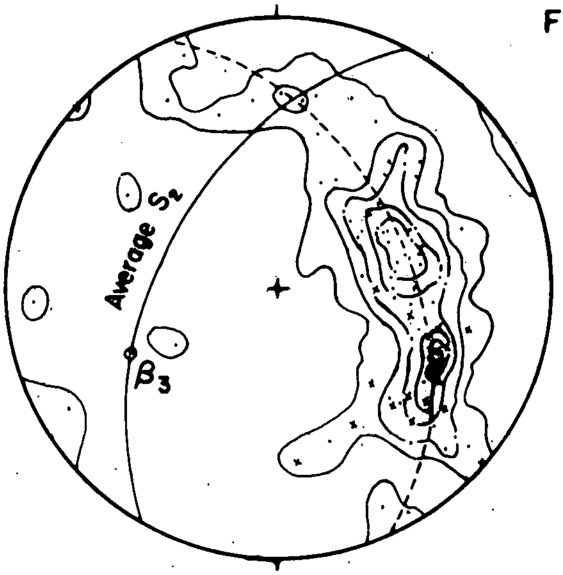
Figure 2.31. Stereograms of two possible interpretations of the large-scale structure of the Magnolia Formation in the northern Medicine Bows, neglecting complications caused by north-trending, west-dipping reverse faults and large mafic intrusive bodies.

A. Favored interpretation shows gently south-plunging inclined syncline in north passing into gently southwest-plunging normal syncline to south. F_3 increases in intensity to north.

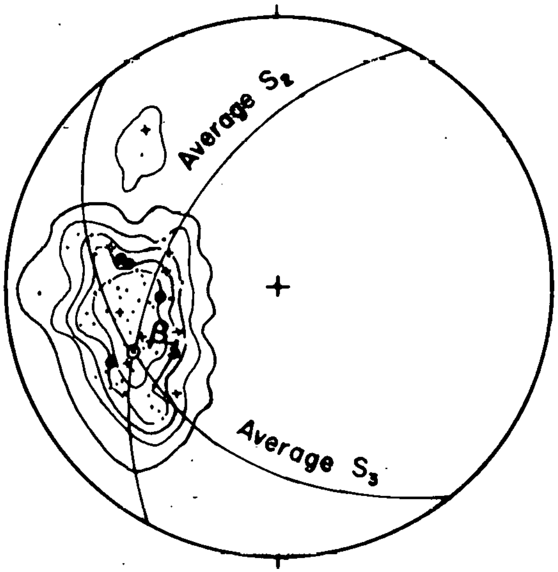
B. Alternate interpretation shows west-plunging reclined syncline in north passing into gently southwest-plunging normal syncline to south.

Stereographic projections of mesoscopic fabric elements in the Magnolia Formation are shown in Figure 2.32. The only indication of the prominent north-northeast trending syncline is in the plot of the poles to bedding (Figure 2.32A) which shows that up-right and overturned beds plot together in a maximum defining an isolinal fold whose axial plane is labeled

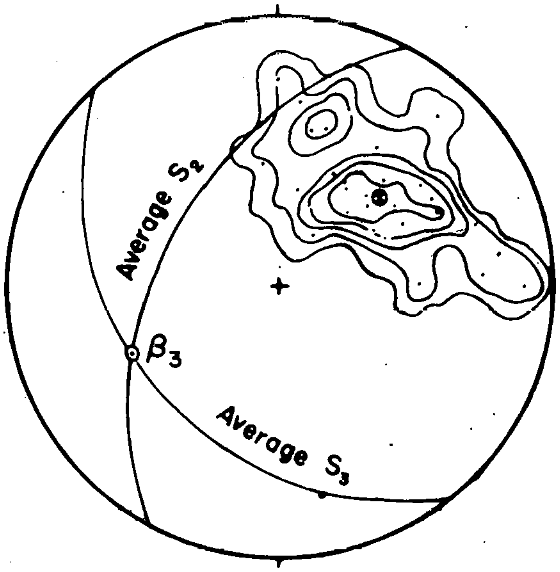
Figure 2.32. Lower hemisphere equal-area projections of structural elements from the Magnolia Formation near Onemile Creek. Contoured at 1,3,5,7,9 % per 1% area.



A. 96 poles to bedding. Overturned beds (×) and right-side-up beds (·) have the same orientation and define an isoclinal fold with axial plane labeled S_2. Dashed line is girdle of poles to bedding defining a later generation fold with axis β_3.



B. 51 stretched pebble (·), 9 minor fold axis (+) and 4 mineral (•) lineations. All lineations are subparallel to β_3 with minor scatter along S_2 and S_3.



C. 30 poles to S_3 foliation. β_3 formed at the intersection of the F_3 axial plane (average S_3) and the average orientation of bedding and F_2 axial plane (average S_2) on the earlier isoclinal fold.

"average S_2". This fold is referred to as an F_2 fold because there is evidence for at least one earlier fold system, F_1, in the underlying Phantom Lake Suite. Figure 2.32A also shows that poles to bedding have been redistributed onto a great-circle girdle about a later, F_3, fold axis (labeled B_3).

Lineations in the Mangolia Formation of the Onemile Creek area are shown in Figure 2.32B. Minor fold axes, stretched pebble lineations in the conglomerates, and mineral lineations in the mafic intrusive rocks are all subparallel to B_3, defined from Figure 2.32A. Most of the minor fold axes have axial planes parallel to the S_3 foliation (Figure 2.32C) and those are clearly late F_3 folds. However, several have north-striking axial planes and these are more readily explained in terms of a system of F_2 reclined folds (Figure 2.31B). If so, F_2 and F_3 would be co-axial west-plunging fold systems, the former isoclinal and the later more open. Stretched-pebble lineations are enigmatic and could represent: 1) lineations nearly perpendicular to the axis of an inclined, gently plunging F_2 fold (Figure 2.31A); 2) lineations parallel to the fold axis of the later, west-plunging F_3 folds, or most probably, 4) a combination of (3) with (1) or (2).

Figure 2.32C shows the orientation of the S_3 foliation. In outcrop, this foliation cross-cuts both bedding and S_2 foliation. It affects rocks of the Phantom Lake Suite, Magnolia Formation, Cascade Quartzite, and mafic intrusive rocks indiscriminantly in the Onemile area. The fact that it affects mafic intrusive rocks is evidence that the F_3 fold system post-dates the major episode of intrusion of gabbroic sills and dikes.

As shown in Figure 2.29, the Onemile Creek area contains a network of faults with varying orientation and varying sense of movement. All faults shown have been invaded by gabbroic magma and are, therefore, older than the intrusion episode. The oldest faults in Figure 2.29 are two north-trending reverse faults. The western one passes just east of EMB-4 and reverse movement on this fault is believed to have partly removed the overturned limb of the F_2 syncline (Figure 2.30). The eastern one passes west of EMB-10 and is responsible for splitting off and repeating part of the upright limb (A-A' of Figure 2.30). We believe that movement on these faults was penecontemporaneous with F_2 folding.

The next oldest faults are two left-lateral fault systems. The southern one, seen in cross sections B-B' and C-C' (Figure 2.30) trends north-northeast and offsets the axis of the major F_2 syncline about 2 km. This fault has been invaded by a large sill which reaches a map thickness of 1.2 km southwest of the area of Figure 2.29. This fault system is shown in Figure 2.29 to have a splay which passes between EMB-8 and MB-16, disrupting the continuity of the main radioactive conglomerates. A second major left-lateral fault occurs north of A-A'. This fault truncates the F_2 syncline. The northwest-trending fault just north of EMB-2 is shown in A-A' and B-B' (Figure 2.30) to be a normal fault bringing older rocks up on the north side and down-dropping the lower units of the Magnolia Formation. This fault may also have a component of left-lateral movement.

Discussion of Structural Controls of Mineralization

A discussion of the chronology of folding events in terms of regional tectonic relationships is presented in earlier sections of this paper. Here, it is of obvious importance to ask which aspects of the structural setting of the Onemile Creek area might have been important in the formation of radioactive quartz-pebble conglomerates.

There are two aspects which deserve attention. First, the Onemile area is one of two areas in the Medicine Bow Mountains where the Magnolia Formation is in close proximity to a "basement" terrane containing granitic rocks. In most areas to the southwest, conglomerates in the Magnolia Formation unconformably overly thick sections of the Phantom Lake Metamorphic Suite containing quartzite and metabasalt. In light of the fact that the source of detrital uranium and thorium minerals in Precambrian quartz-pebble conglomerates is believed to be Late Archean granites (Roscoe, 1969; Pretoriums, 1976; Houston and Karlstrom, 1980) it is tempting to speculate that the proximity of the Onemile Creek conglomerates to granitic rocks may partly explain the relatively high uranium and thorium contents of the Magnolia Formation in this area. In other areas, although conglomerates were deposited by similar fluvial processes, there appears not to have been as good a source of radioactive minerals.

The second aspect which deserves discussion is our interpretation that the Onemile Creek area may have been near the margin of the Deep Lake Group depositional basin. There are several lines of evidence supporting this interpretation. First, the Onemile Creek area is the northernmost limit of the Deep Lake Group, as seen today. Second, the presence of onlapping unconformities, one below the Magnolia Formation and one below the Cascade Quartzite suggests proximity to basin-bounding highlands (Salop, 1977, p. 21). Third, the intensity of faulting and of F_2 and F_3 folding in the Onemile area, compared to areas to the south, may reflect higher strain in the thinner sedimentary pile at the edge of the depositional basin. In view of the very active syndepositional tectonism that is hypothesized for fan-delta formation in the Witwatersrand Sequence of South Africa (Brock and Pretorius, 1964, Pretorius, 1976), it is plausible to suggest that some of the faults in the Onemile Creek area (e.g. the northwest trending normal fault in Figure 2.29) may represent reactivations of old syndepositional faults which were active along the margin of the depositional basin. Incremental movements along such faults, during deposition, plays a large role in causing prolonged fluvial reworking of conglomerates on alluvial fans and fan deltas. This reworking mechanically concentrates the uranium, gold and other heavy minerals in the matrix of mature quartz-pebble conglomerates (Minter, 1976). Thus, the position of the Onemile Creek area near a tectonically active margin of the Deep Lake depositional basin may help explain the presence of radioactive conglomerates in this area.

Detailed Stratigraphy of the Magnolia Formation in the Onemile Creek Area

Plate 4 shows six stratigraphic sections of the Magnolia Formation in the Onemile Creek area and summarizes pebble size information, petrography, and radiometric characteristics of the Onemile Creek conglomerates and quartzites. These sections were compiled from drill core and surface measured sections and are arranged from more proximal (left) to more distal (right), assuming a northerly source for the clastic sediments. The dominant lithology in the Onemile Creek area is coarse-grained subarkosic quartzite but beds of boulder paraconglomerate, phyllite, quartz-granule conglomerate, and highly radioactive quartz-pebble conglomerate are also present and locally abundant.

Figure 2.29 shows five lithostratigraphic units in the Magnolia Formation of the Onemile Creek area. This stratigraphic interpretation aids in outlining the complex structure in the Onemile Creek syncline and is a useful simplification in reporting ore reserve estimates. However, it is predicated on the assumption that phyllites of Unit 4 are laterally continuous and correlatable over the Onemile Creek area. An alternative interpretation, presented in Figure 2.33 and discussed later, views the phyllites and other lithofacies in the Onemile Creek area of the Magnolia Formation as discontinuous and laterally variable units. Ore reserve estimates, discussed in Volume 3, were done for small segments of rock around the fold (roughly following Plate 4) and are valid regardless of which stratigraphic interpretation is used. The biggest difference between the two interpretations is in the total reconstructed thickness of the Magnolia Formation in this area and in interpretations of depositional environments. We favor the former interpretation of laterally continuous stratigraphic units which suggests an overall thickness of 400 m and deposition on an extensive wet alluvial fan or in a braided river system. The latter interpretation, of rapidly changing lithofacies (Figure 2.33) suggests an overall thickness of 204 m and deposition on a smaller alluvial fan system. The following discussion of Plate 4 points out the characteristics of the major lithofacies in each stratigraphic section.

Unit 1 is composed of a basal arkosic paraconglomerate overlain by gravelly subarkose. Basal conglomerates are exposed near EMB-1 and MB-16 (Figure 2.29). These are poorly packed conglomerates with cobbles (and rare boulders up to 35 cm in diameter) of quartz, quartzite, and granite, in a matrix of laminated, muscovitic and biotitic subarkose. The basal paraconglomerates appear to be discontinuous, forming thin lenses or sheets tens of meters wide and up to 2 m thick. As shown in Plate 4, the basal paraconglomerates are only mildly radioactive; maximum assays of 240 ppm U and 210 ppm Th were found in MB-16.

Unit 2 is coarse-grained to granular muscovitic subarkose. The lower boundary is gradational and is placed just above the last major conglomerate of Unit 1. In the vicinity of EMB-1 (column 2 of Plate 4) there are two distinct parts of Unit 2. The lower part is a poorly sorted granular subarkose which, in outcrop, shows partings of one centimeter or less which are low amplitude, small scale trough crossbeds with trough depths of several centimeters, widths of 10-40 cm, and lengths of a meter or more. This laminated section changes abruptly to a thicker bedded (10 centimeter beds) quartz-granule conglomerate which is also trough crossbedded. Unit 2 also contains thin (several cm thick) muscovitic quartz-pebble conglomerates which extend laterally tens of meters and contain up to 130 ppm U and 99 ppm Th in the EMB-1 area. Near MB-16, Unit 2 is more arkosic and contains more highly radioactive conglomerates (up to 1620 ppm U and 212 ppm Th). Conglomerates in both areas contain pebbles of quartz, quartzite, and granite, in an iron-stained muscovitic matrix. Drilling has shown that the iron stain is from oxidized pyrite.

Unit 3 is similar to Unit 2, a coarse-grained to granular muscovitic, trough crossbedded subarkose with scattered, slightly radioactive subarkosic conglomerates. Trough crossbeds are similar to those described in Unit 2. The subdivision is based on the presence of clean quartz-pebble conglomerates within more massive quartzites in Unit 3. Near EMB-1, these conglomerates contain up to 88 ppm U and 150 ppm Th in drill hole. Near MB-16, Unit 3 conglomerates contain up to 30 ppm U and 26 ppm Th.

Unit 4 contains biotite- and chlorite-rich phyllites and schists with beds of paraconglomerate, fine- to medium-grained biotitic quartzite, and occasional quartz-pebble conglomerates. Unit 4 phyllites were encountered in EMB-1, EMB-6, EMB-7, and EMB-2 and are used as a stratigraphic marker in Plate 4.

Unit 5 contains the major, highly radioactive, quartz-pebble conglomerates in the Onemile Creek area. Conglomerates occur within thick zones (labeled 5a and 5b in Plate 4) of muscovitic quartz-pebble conglomerate interbedded with coarse-grained to granular, muscovitic quartzite. All contacts are gradational and lithologic designations shown in Plate 4 reflect the dominant lithology within intimately mixed sections of quartz-pebble conglomerate, quartz-granule conglomerate, and coarse-grained quartzite. Individual radioactive quartz-pebble conglomerates are sheet-like and range in thickness from one or two pebbles thick to beds over a meter thick. Individual beds appear to be continuous for several hundred meters. Some conglomerates are distinctly bimodal, others are poorly sorted with a complete range in grain size from sand to pebbles (Plate 4). We have noticed no systematic grading within individual conglomerate beds. Some beds fine upward, others fine downward or laterally.

Clast size in conglomerates was studied by measuring the ten largest pebbles in outcrop or the single largest pebble in drill core. Clasts are severely stretched so we converted pebble ellipsoids to equivalent spheres for presentation in Plate 4. Where only two sizes of pebble ellipsoids were measurable, the third axis was assigned a value based on the ratios of the axes of mean pebble ellipsoids (strain ellipsoids?) from each outcrop. As shown in Plate 4, we found no obvious trends in maximum pebble size over the Onemile Creek area. Clasts ranged in size from granules to cobbles 7.5 cm in diameter. However, most of the radioactive conglomerates are small pebble conglomerates with pebbles one to three centimeters in diameter. Pebbles are stretched and well rounded. Many conglomerates contain 100 percent quartz pebbles; others contain up to 20 percent quartzite and granite in with quartz. The quartzite clasts are fine-grained and slightly micaceous and resemble quartzites of the underlying Phantom Lake Suite. Granite clasts appear to have been derived from the body similar to the Late Archean granite immediately to the north of the Onemile Creek area (Figure 2.29). No mafic rocks or phyllitic clasts were found in Unit 5. Packing density of pebbles is hard to estimate because of stretching and flattening of pebbles during deformation. Many conglomerates in Unit 5 are obviously pebble supported, others may not have been.

The matrix of Unit 5 conglomerates consists of rounded to subangular quartz granules, up to 30 percent muscovite and sericite, up to 5% feldspar granules, and local granules of quartzite and granite. In some samples the matrix may contain up to 30 percent very fine-grained, angular, quartz grains forming a trimodal distribution with pebbles and granules. The most radioactive conglomerates contain up to 30 percent pyrite in the matrix. Pyrite occurs as sand size grains (up to 2 mm) which may be rounded grains with euhedral overgrowths, euhedral crystals, or irregular masses filling space between quartz grains. In conglomerate outcrops, pyrite is nearly completely oxidized to hematite and limonite but cubic pseudomorphs or vugs are frequently preserved.

Quartzites of Unit 5 which are interbedded with the quartz-pebble conglomerates are generally moderately sorted and similar in composition to the matrix of nearby conglomerates. Quartzites are generally much poorer in pyrite, usually containing only one to two percent.

The composition of the conglomerates is quite variable both laterally and vertically. Plate 4 shows that feldspar content of Unit 5 is low in EMB-6-7, and EMB-10 but is higher in EMB-8, and EMB-9-2. This variation in feldspar content occurs on smaller scales also. Near EMB-10, the feldspar content of the matrix of conglomerates varies from zero to seven percent, with the more feldspathic conglomerates containing more granite pebbles. In general, it appears that the cleaner conglomerates, i.e. the ones with less feldspar tend to be the most radioactive.

Abundant faults, and faults invaded by gabbroic intrusions, make it difficult to reconstruct the true thickness of the stratigraphic section. Correlating units as we have done in Plate 4, the total thickness is about 400 meters. However, another possible interpretation, is that all lithofacies are lenticular and not correlatable from one area to another. Figure 2.33 uses the same sections as in Plate 4 but they are positioned so that the thickness of the exposed section at each location represents nearly the entire thickness of the Magnolia Formation, about

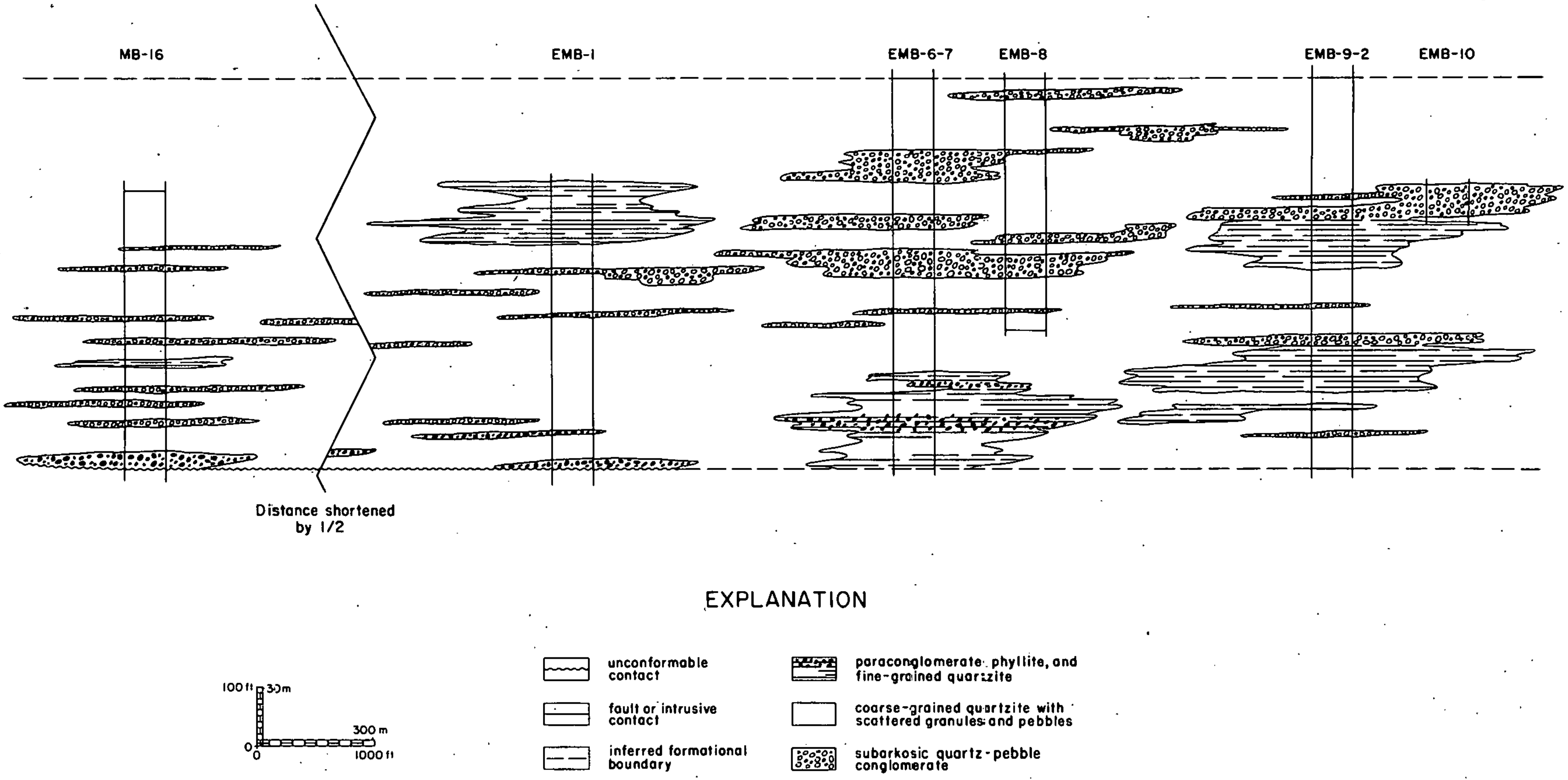


Figure 2.33. Reconstructed stratigraphic columns of the Magnolia Formation in the Onemile Creek area, this interpretation shows rapid facies changes, and total formational thickness approximately equal to maximum exposed thickness (204 m).

204 m. Because of complicated structure and discontinuous outcrops, we cannot say that either of our interpretations is correct. We prefer the one shown in Plate 4 because the relatively great (30 meter) thickness of phyllitic zones suggests to us fairly extensive (on the order of kilometers at least) lateral dimensions for these fine-grained units.

Sedimentology of the Magnolia Formation in the Onemile Creek Area

The Magnolia Formation of the Onemile Creek area is characterized by a complex mixture of lithofacies: horizontally bedded clast-supported gravels, muddy matrix-supported gravel, trough crossbedded sand, horizontally bedded sand, and silts and muds. These facies are all common in braided river environments (Miall, 1977; 1978) and we interpret the intercalated character of these lithofacies within a several km^2 area to be strong evidence for braided stream deposition of the Magnolia Formation.

More specifically, available evidence suggests braided stream deposition in the mid-fan, but mainly in more distal portions of a wet alluvial fan system; i.e. one formed by perennial stream flow (Schumm, 1977). Evidence for alluvial fan deposition includes the rapid facies changes in the unit, the change in character of the Magnolia Formation to the south (discussed below) and the presence of poorly sorted, micaceous paraconglomerates which we interpret to be debris flows. Debris flows are common in many alluvial fan environments (Rust, 1979, p. 12). Evidence for mid-fan deposition is the presence of radioactive quartz-pebble conglomerates which may represent sheet gravels, braid bar deposits, and channel fill deposits in which heavy mineral placers were concentrated by continued sediment reworking in active parts of the fan system (Minter, 1976, 1978; Smith and Minter, 1980). The relatively thin lenses of conglomerate, dominance of small pebble conglomerates, and the absence in significant quartzites of the most dense heavy minerals -- gold and uraninite, appear to indicate more distal fluvial environments than those inferred for fossil-placer conglomerates in the Witwatersrand Sequence (Minter, 1978) and the Elliot Lake area of Canada (Roscoe, 1969). The presence of thick phyllite sections in the Onemile Creek area also appear to require either distal depositional environments or low fan profiles and fairly stable positioning of the main stream channel, which would allow accumulation of thick sections of fine-grained material. Possible low fan profiles, the thin and laterally extensive character of conglomerate layers, the absence of scour features, low amplitude of trough crossbeds, and the relative scarcity of paraconglomerates (debris flows) all suggest deposition on a wet alluvial fan where deposition was dominated by sheet-floods.

Mineralogy of Radioactive Conglomerates in the Onemile Creek Area

The matrix of radioactive quartz-pebble conglomerates in the Onemile Creek area contains a variety of minerals. As shown in Plate 4, quartz, feldspar, and mica are by far the most abundant. A plot of these major constituents (Figure 2.8) shows the rocks to be argillaceous subarkoses and quartzarenites.

A number of heavy minerals are also concentrated in the matrix of the conglomerates. Pyrite is by far the most abundant heavy mineral and may constitute up to 30 percent of the matrix of conglomerates (5a of column EMB-9-2 in Plate 4). As shown in Plate 4, pyrite (labeled opaques - 0) occurs in significant proportions only in conglomerates of Unit 5. Pyrite typically occurs as euhedral grains and as aggregate grains (Figure 2.34). Some pyrite grains have partly rounded shapes (Figure 2.34B), others have rounded core grains with secondary euhedral overgrowths (Figure 1.15). This suggests that the original pyrite grains may have been rounded detrital grains and that the present euhedral grain shapes reflect recrystallization during amphibolite facies metamorphism. The presence of pyrite in microfractures (Figures 2.34A and 1.15) both parallel and perpendicular to bedding is also probably related to metamorphism.

Other non-radioactive heavy minerals include ilmenorutile, zircon, apatite, galena (mainly radiogenic), chalcopyrite, bornite, marcasite, sphene, ilmenite, columbite, magnetite, anatase, rutile, and spessartine garnet (Desborough and Sharp, 1979, p. 38). Ilmenorutile is fairly common and occurs as spherical or elliptical grains with 20-30 percent interstitial quartz (Figure 2.34B). Rounded and subrounded zircons are also common but are volumetrically minor because of small grain sizes. Some zircons appear metamict and others are, in fact, fine aggregates which may be metamorphic in origin. Garnet and apatite grains are generally well rounded and are considered to be detrital in origin.

Chalcopyrite and bornite are uncommon but are present as small irregular masses in the matrix of conglomerates or as inclusions in pyrite grains. Marcasite is also uncommon and occurs as an alteration product of pyrite. Galena occurs as small masses in pyrite and as minute crystals in uranium-thorium minerals. Magnetite, ilmenite, and columbite are very uncommon minerals in the heavy mineral suite and are probably of detrital origin. Rutile is present as deep red, rounded grains

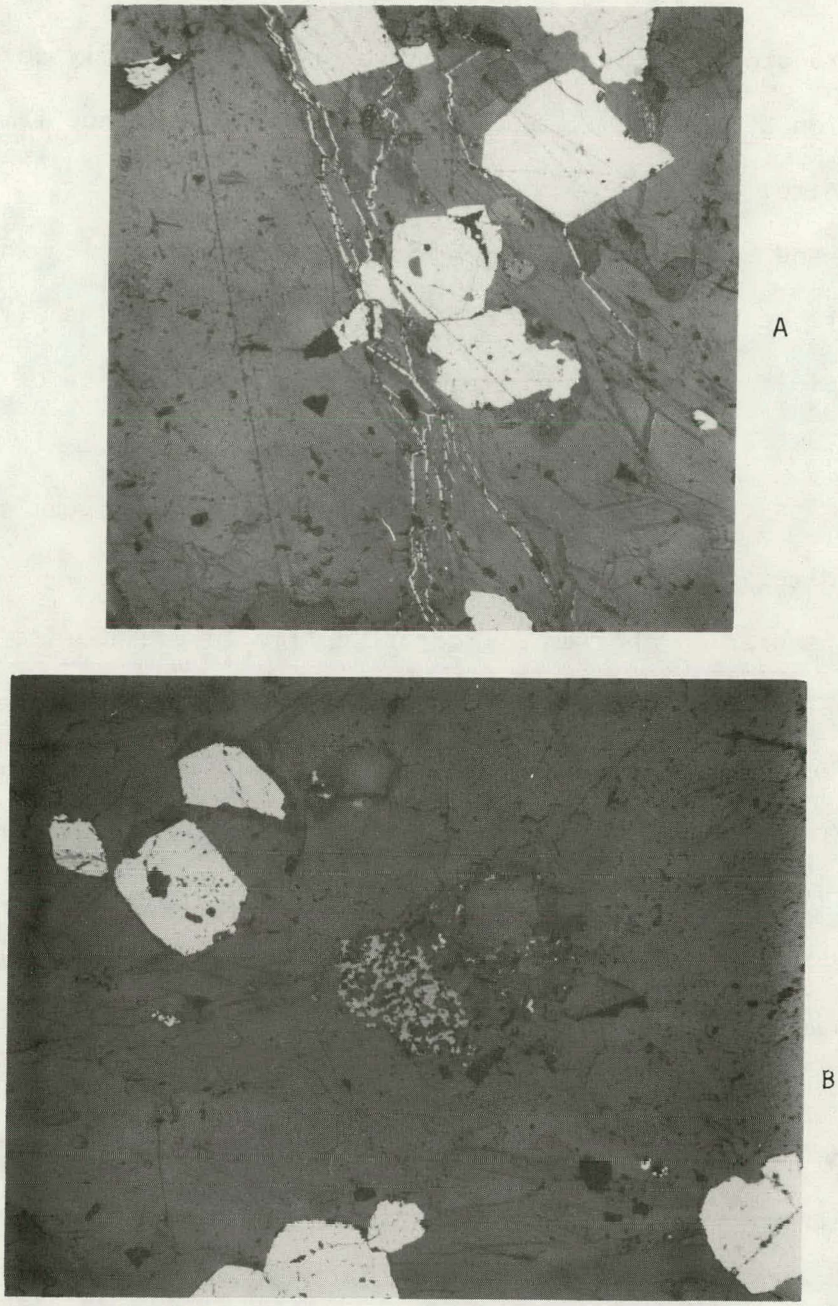


Figure 2.34. Photomicrographs of heavy minerals from the Magnolia Formation of Onemile Creek. A. Euhedral pyrite grains, pyrite aggregates, and microfractures healed with pyrite (white). B. skeletal grain of ilmenorutile (center) consisting of a complex intergrowth of ilmenotutile (intermediate gray) and quartz (dark gray). White pyrite grains are subhedral or partly rounded. Gray fibrous crystals are muscovite. Ilmenorutile grain is .28 mm long.

and as fine aggregates with anatase. These aggregates probably developed through alteration of titanium-bearing minerals such as sphene, ilmenorutile and ilmenite.

Desborough and Sharp (1979, p. 38) also identified graphite in one sample. Graphite occurs as poorly crystallized, rounded grains with inclusions of coffinite. The graphite contains sulfur and may be of organic origin.

Radioactive heavy minerals are coffinite, thorite, thorogummite, monazite, huttonite(?), and zircon. According to Desborough and Sharp (1979, p. 39) the uranium-thorium silicates coffinite and thorite vary widely in composition. The coffinite contains 45-61 weight percent uranium and up to several weight percent thorium and lead whereas the thorite contains 30-45 weight percent thorium and up to several percent uranium, lead, and yttrium. These uranium and thorium minerals are in masses that may have been rounded originally but now appear recrystallized and to have developed irregular borders or overgrowths (Figure 1.16).

Some of the uranium and thorium in the quartz-pebble conglomerate is present in zircon and monazite. Some zircon appears metamict or recrystallized and most of the monazite is believed to be recrystallized and therefore a metamorphic mineral. These two minerals contain up to six weight percent thorium (Desborough and Sharp, 1979, p. 38).

Minute crystals of uranium and thorium silicate (coffinite) are present in graphite and in nickel-cobalt-iron sulphide grains.

Uranium and Thorium Content of Radioactive Conglomerates in the Onemile Creek Area

Table 2.28 summarizes the uranium and thorium content of the most radioactive conglomerates encountered in drill holes in the Onemile Creek area. Values from EMB-6, 7, 8, and 10 are from Unit 5; values from MB-16 are from Units 1 and 2. Background values of muscovitic, granular and pebbly quartzites throughout these units is about 10-30 ppm U and 10-25 ppm Th. Maximum uranium values range up to 1620 ppm U over 6 cm in MB-16 and 1380 ppm U over 60 cm in EMB-6. Maximum thorium values were 970 ppm Th over 6 cm in EMB-6 and 730 ppm over 30 cm in EMB-6. Th/U ratios are generally less than one and some are as low as .19 reflecting the fact that Onemile Creek Conglomerates are appreciably richer in uranium than thorium. Mean U and Th values (Table 2.28) were computed for selected intervals by calculating weighted means from radioactive conglomerates (with thicknesses shown) combined with large thicknesses of quartzites with "background" U, Th values. This compilation presents a more conservative view of the uranium resources in the Onemile Creek than the one predicted in Volume 3. Even though there appears to be appreciable tonnages of uranium in the conglomerates the mineralization may be too disseminated for effective underground mining. Table 2.28 suggests that the most radioactive conglomerates tend to be fairly thin so that average uranium grades are generally low over mineable thicknesses. Possible exceptions are in EMB-6 where a nine-foot interval averages 411 ppm U (236-245 feet) and where a thirteen-foot interval averages 285 ppm U (290-303 feet); and in MB-16 where one ten-foot interval averages 135 ppm U (380-390 feet).

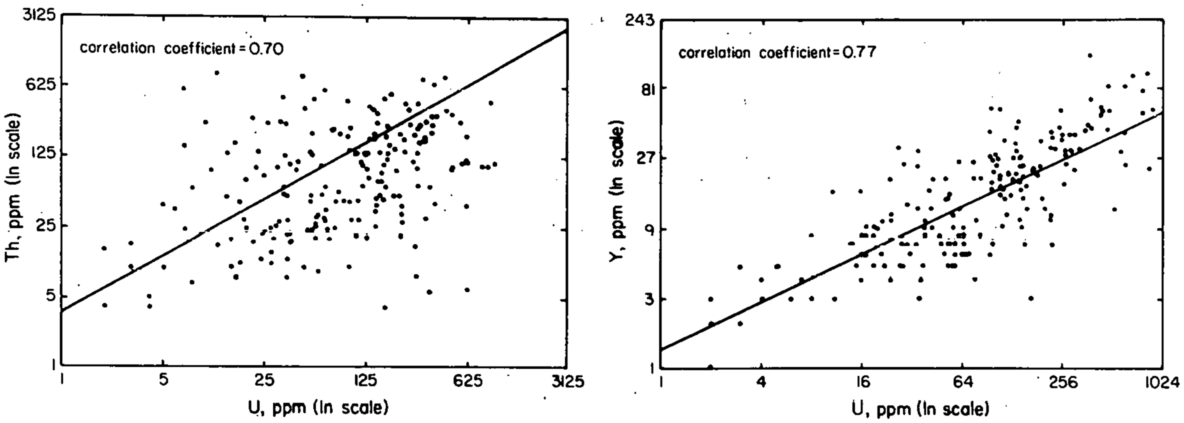
TABLE 2.28. ESTIMATE OF URANIUM AND THORIUM CONTENT OF RADIOACTIVE UNITS IN THE ONEMILE CREEK AREA.

DRILL HOLE	INTERVAL FT	BACKGROUND Th	BACKGROUND U	MEAN Th	MEAN U	MAXIMUM Th	THICKNESS OF CONG., FT	MAXIMUM U	THICKNESS OF CONG., FT
EMB-6	0-130	23	13	24	18	170	.5	110	1
	130-162	23	23	53.8	69.5	620	.8	580	.5
	162-236	23	23	23	23.5	47	.7	23	.2
	236-245	23	23	151	411	200	2	1400	2
	245-290	23	23	23	23	29	1	23	.2
	290-303	23	23	228	285	730	1	909	2
	303-388	23	23	24	25	160	.5	310	.5
MEAN Th/U					1.06			.33	
EMB-7	0-188	23	20	27	22	450	.3	260	.2
	188-195	23	30	82	88	430	1.0	430	1
	195-309	23	30	25	32	300	.7	190	.7
	309-350	23	30	60	56	620	.5	330	.2
	350-372	23	30	23	30	51	.3	34	.3
	372-406	23	30	41	53	970	.2	1000	.3
	406-509	23	30	32	41	560	.2	690	.2
	509-792	23	30	24	30	610	.2	490	.2
MEAN Th/U					.91			1.14	
EMB-8	75-288	12	18	15	21	390	.5	298	.5
	288-369	12	25	25	35	480	.2	310	1
	369-395	12	25	59	60	370	1	280	1
	395-481	12	25	14	26	350	.2	280	.2
MEAN Th/U					.58			1.30	
EMB-10	661-681	No Background Samples		27	49	88	.2	460	.2
					.55			.19	
MB-16	63-176	6	8	19	15	210	1	240	1
	176-209	10	11	19	34	65	1	78	1
	209-323	10	11	14	16	210	1	270	1
	323-343	10	11	18	69	110	1	870	1
	343-380	10	11	11	15	24	1	58	1
	380-390	10	11	22	135	110	.2	1600	.2
	390-667	10	10	16	10	26	.2	55	.2
MEAN Th/U					1.11			.35	
GRAND MEAN Th/U					.84			.51	

Geochemistry of Radioactive Conglomerates in the Onemile Creek Area

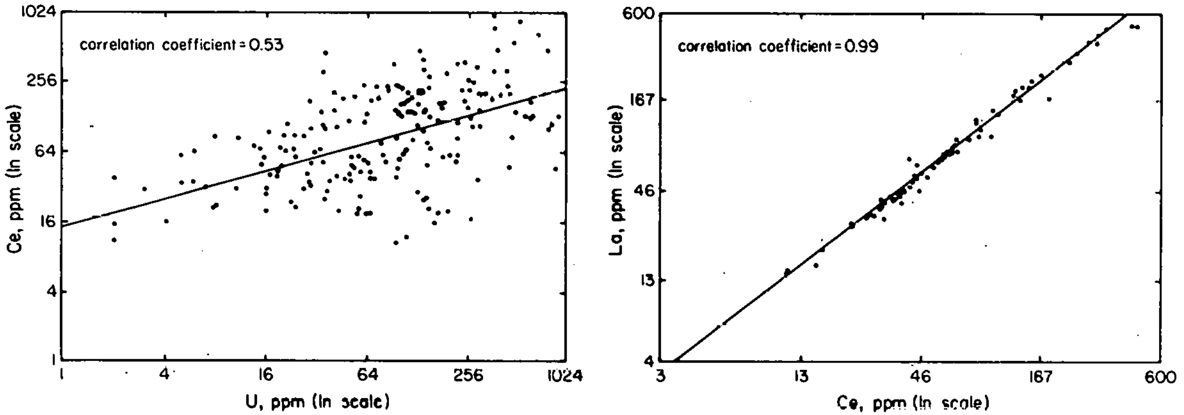
Correlations between element concentrations and between elements and maximum grain size were calculated for radioactive conglomerates of the Onemile Creek area to gain more information on mineralogy of the conglomerates and to examine the relationship between uranium content of conglomerates of a sedimentary parameter-grain size. Pearson correlation coefficients were calculated for 14 elements as well as the ratio Th/U and maximum pebble size (Table 2.29). As shown in Table 2.29 separate matrices were generated for surface conglomerates (118 samples) and subsurface conglomerates (209 samples from EMB-6, 7, 8, 9, and 10 and MB-16). Correlations were determined using a lognormal distribution for each element. This study was patterned after Theis' (1979) study of the Elliot Lake uranium deposits of southern Ontario and shows major differences between the Magnolia Conglomerates and the Elliot Lake ore rocks.

As shown in Figure 2.35A and Table 2.29, uranium correlates moderately well with thorium (correlation coefficient = .70 for drill-core samples) which agrees with optical and microprobe identification of uranium- and thorium-bearing phases such as uranothorite, thorogummite, and possible xenotime. The presence of xenotime is also suggested by the good correlations between uranium/thorium and yttrium (Figure 2.35B,F). The moderate correlations (0.5) of uranium with zirconium and cerium (Table 2.29) suggest that some uranium is present in the lattices of zircons and monazites. (Figure 2.35D shows that Ce is strongly correlated with La which indicates the presence of monazite). Monazites also contain appreciable thorium as shown by the good correlation (.83) between



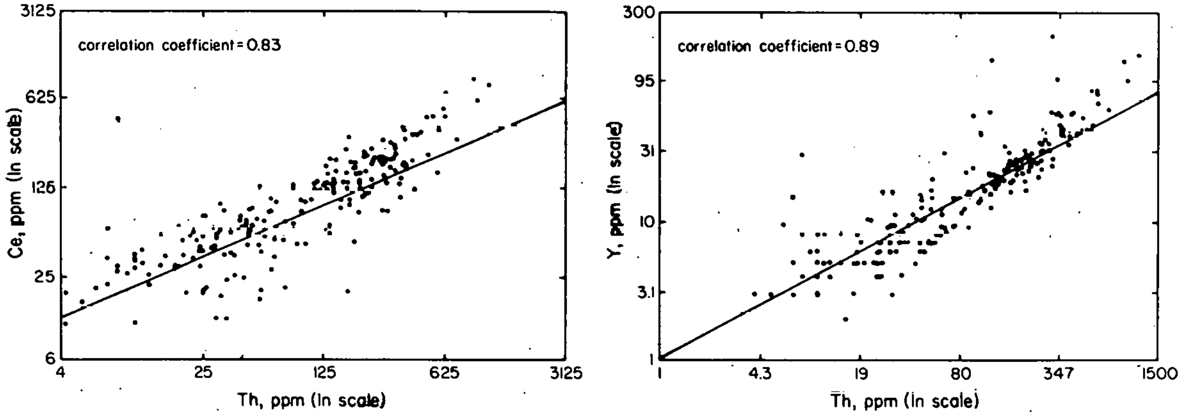
A. U plotted against Th.

B. U plotted against Y.



C. U plotted against Ce.

D. Ce plotted against La.



E. Th plotted against Ce.

F. Th plotted against Y.

Figure 2.35. Correlations between selected elements from radioactive conglomerates of the Magnolia Formation of the Onemile Creek area.

Th and Ce (Figure 2.35E). The poor correlation (.20) between uranium and titanium indicates that brannerite, if present, is uncommon. The scatter in the U-Th plot (Figure 2.35A) suggests the presence of a uranium-bearing phase that does not contain thorium. Coffinite was identified by microprobe studies (Desborough and Sharp, 1979, p. 32) and is the most likely candidate.

Uranium correlates very poorly (0.02 in drill core) with maximum pebble size in the Onemile Creek conglomerates. This is in marked contrast to the Elliot Lake conglomerates where Theis (1979, p. 12) showed a reasonably good correlation (0.47) between these two parameters. This poor correlation in the Onemile Creek conglomerates suggests that present uranium content is not related to grain size and inferrentially to stream velocity and depositional energy. This may reflect the low specific gravity (∿ 5) of uranium-bearing phases in the Onemile Creek conglomerates which would be less easily fractionated by sedimentary processes than denser uraninite grains (specific gravity ≲ 10) of the Elliot Lake conglomerates. Alternatively, it may reflect remobilization of uranium during metamorphism or inadequacies in determining maximum grain size in drill core. Note that the correlation improves somewhat (0.27) for outcrop samples, where it was possible to more accurately characterize maximum pebble size. The lack of correlation between uranium and grain size, even if it is related to sedimentary features of the conglomerate, does not refute the field observation that the highest radioactivity is associated with conglomeratic layers. It may merely indicate that small pebble conglomerates are as likely (or more so) to be highly radioactive as cobble-bearing paraconglomerates. Uranium also

TABLE 2.30. MATRIX OF PEARSON CORRELATION COEFFICIENTS FOR RADIOACTIVE CONGLOMERATES OF THE MAGNOLIA FORMATION FROM THE ONEMILE CREEK AREA. CORRELATION COEFFICIENTS FOR ELEMENTS WERE COMPUTED USING LOG TRANSFORMATIONS.

Subsurface Samples

Outcrop Samples

Surface Samples

	U(nt)	Th	Th/U	Fe	Ti	Y	Zr	K	Ce	La	Pb	P	Ca	Nb	V	Max. Peb. Size	Max. Peb. Size, Unit 5	Mean Conc. (ppn)
U(nt)		.70	−.38	.57	.20	.77	.53	.06	.53	.59	.73	.45	−.14	56	−.12	.02	−.07	150.1
Th	.63		.40	.48	.11	.89	.73	.09	.83	.84	.74	.45	−.37	.75	−.37	−.21	−.25	157.3
Th/U	−.27	.57		−.13	−11	.16	.27	.07	.41	.14	−.23	.01	−.32	.14	−.33	−.34	−.26	1.64
Fe	.57	.35	−.16		.48	.61	.41	−.04	.41	.52	.55	.54	.01	.46	.24	.20	.24	.237*
Ti	.43	.10	−.33	.52		.31	.33	.41	.22	.28	.07	.43	.17	.30	.65	−.09	−.16	1308
Y	.66	.82	.32	.35	.14		.72	.09	.80	.81	.80	.61	−.30	.77	−.21	−.19	−.30	21.4
Zr	.55	.75	.35	.23	.25	.77		.18	.65	.67	.47	.56	−.22	.53	−.14	−.14	−.20	121.1
K	.00	.32	−.40	.36	.74	−.29	−.23		.07	.08	.02	−.03	−.17	.04	.39	−.18	−.17	2.00*
Ce	.51	.88	.54	.23	−.02	.84	.77	−.42		.99	.64	.57	−.35	.70	−.41	−.33	−.41	131.5
La	.45	.90	.58	.31	.00	.80	.75	−.33	1.0		.65	.47	−.28	.80	.07	−.15	−.25	125.6
Pb	.61	.79	.27	.38	.13	.57	.50	−.15	.68	.69		.24	−.11	.64	−.11	.03	.05	79.23
P	.69	.64	.07	.55	.29	.73	.61	−.09	.66	.65	.64		.13	.60	.01	.10	−.23	167.0
Ca	.19	.21	.06	.16	.24	.16	.07	.10	.17	.05	.28	.30		−.18	.28	−.02	−.12	.096*
Nb	.23	.57	.47	.13	.08	.55	.44	.36	.57	54	.60	.36	.27		−.18	−.16	−.33	25.9
V	.29	[illegible]	.53	.49	.70	−.10	.06	.67	.22	.04	.08	.22	.17	.40		−.31	.36	33.9
Max. Peb. Size	.27	[illegible]	−.16	.35	.44	.09	−.02	.33	.04	.15	.08	.15	.09	.04	.43			
Mean Conc. ppn	25.3	104.9	4.89	1.67*	1274	10.4	127.8	1.93*	161.8	163.0	63.1	231.1	.076*	22.3	37.03			

*wt %

correlates poorly with potassium (0.06) in the Onemile Creek conglomerates suggesting that uranium content is not related to the maturity of the sediments and highly radioactive conglomerates are as likely to be arkoses and argillaceous quartzites as quartzarenites.

The ratio Th/U does not correlate particularly well with anything (Table 2.29). However, the difference in mean Th/U values of outcrop samples (4.89) compared to drill core samples (1.65) indicates appreciable surface leaching of uranium.

Theis (1979, p. 11) reported that uranium and lead had a strong correlation (0.92) which suggested that nearly all lead in the Elliot Lake conglomerates was radiogenic. The poorer (0.73) correlation in the Onemile Creek conglomerates suggests that whereas most of the lead is probably radiogenic, there also appears to be some primary lead mineral associated with the heavy mineral suite. This agrees with petrographic observations of galena inclusions in pyrite grains.

Perhaps the most significant finding of this study of element to element correlations is that the geochemistry of the Magnolia Formation in the Onemile Creek area is very different from that of the ore conglomerates from Elliot Lake. In the Onemile Creek conglomerates uranium occurs in a variety of minerals: uranothorite, thorogummite, coffinite, monazite, and zircon whereas in Elliot Lake, the main uranium minerals are uraninite and brannerite. Also, the positive correlation between grain size and uranium content in the Elliot Lake rocks (Theis, 1979) does not appear to hold for the Onemile Creek conglomerates. This may be due to different mineralogies of the heavy mineral suites, different sedimentary conditions, as reflected by the overall finer grain sizes in the Onemile Creek conglomerates compared to the Elliot Lake rocks, or it may in some way reflect metamorphic changes in the Onemile Creek conglomerates, deformation of pebbles, or inadequate characterization of maximum pebble size by our limited measurements of drill core.

Magnolia Formation excluding the Onemile Creek area

Outside the Onemile Creek area radioactive conglomerates crop out along the northwest limb of the Sand Lake Syncline in the drainage area of Threemile Creek (Figure 2.28). Granite and quartzite boulder, amphibole and biotite matrix, paraconglomerates of the Colberg Metavolcanic Rocks grade into arkosic paraconglomerates of the Magnolia Formation. Immediately above the arkosic paraconglomerates are radioactive quartz-pebble and quartz-granule conglomerates with gamma-ray values up to 20 times background levels and up to 51 ppm U and 130 ppm Th. Outcrops of conglomerate strata extend south along the Sand Lake Syncline for about 5 km (Fiqure 2.28) where they are truncated by an east-west trending normal fault (see Plate 1). At its southern extension the Conglomerate Member is a polymictic arkosic paraconglomerate containing up to 20 ppm U and 190 ppm Th in outcrop, and is overlain by quartz-granule conglomerates.

Two holes were drilled in the Threemile Creek area (EMB-5 and EMB-11) and three were drilled in the southern section (Bows -1, -2, and -3) by companies in the private sector. Subsurface information confirmed the gradational nature of the contact between the Magnolia Formation and the Colberg Metavolcanic Rocks in the Threemile Creek area.

In both areas the Conglomerate Member consists of basal, slightly radioactive, arkosic paraconglomerate. The upper portion of this member, and the basal section of the Quartzite Member, contain discontinuous lenses of highly radioactive quartz-pebble and quartz-granule conglomerates. Values of up to 365 ppm U and 344 ppm Th were obtained from core in the Threemile Creek area and highs of 110 ppm U and 41 ppm Th occurred in the Bows holes. Subsurface extension of the most radioactive conglomerate in the Threemile Creek area indicates a downdip continuation of the radioactive zone for at least 300 meters. Farther south conglomerates are lower grade and less continuous. The overlying quartzites are fine-grained to granular and locally trough crossbedded.

The clasts in these conglomerates are up to 10 cm in diameter and consist of quartz, fine-grained quartzite, chloritic quartzite, granite, granite gneiss, sericite schist, and chlorite schist. In the southern section, basal conglomerates contain abundant mafic and some rhyolitic clasts. Most individual conglomerate layers contain both quartz and quartzite clasts; sorting is poor or bimodal. The matrix composition is commonly chlorite, biotite schist, or chlorite, biotite-rich quartzite, and often contains appreciable garnet (Table 2.9). The overlying micaceous quartzites are fine- to coarse-grained rocks containing megacrysts of chlorite and biotite. It is interesting to note, also, that the highest percentages of pyrite occur within the lower Quartzite Member (up to 15 percent in EMB-11) and correspond to radiometric peaks.

Hole MB-4 was drilled on the east limb of the Sand Lake Syncline where a small section of the Magnolia Formation is exposed (Figure 2.28).

Scattered outcrops of quartz-pebble conglomerate occur in this area and contain up to 13 ppm U and 485 ppm Th. Here, the Magnolia Formation unconformably overlies the Bow River Quartzites of the Phantom Lake Metamorphic Suite and consists entirely of very coarse-grained to granular, micaceous subarkose with scattered pyrite (see Table 2.9). No highly radioactive conglomerates are present; the highest uranium and thorium values are 14 ppm and 30 ppm. The upper contact with the Lindsey Quartzite appears to be gradational in this area.

The central area of the Medicine Bow Mountains includes outcrops of the Magnolia Formation south and west of the Threemile Creek area, and northwest of the Arrastre Anticline. Outcrops here are widely scattered but, in general, the basal section consists of quartz-granule conglomerates which are only mildly radioactive. The only location where basal quartz-pebble conglomerates are found is near the North Fork of Rock Creek where they unconformably overlie metabasalts and quartzites of the Phantom Lake Metamorphic Suite. Surface samples contain up to 3.2 ppm U, and 13 ppm Th and register gamma-ray values that are 6 times background radioactivity. These values are not high, but large radon anomalies in ground waters were reported for this area (Miller, 1980) suggesting greater concentrations of uranium in the subsurface. The Bendix Corporation drilled hole MB-11 in this area, in which no basal conglomerates were encountered. Instead, 180 feet of quartz-granule conglomerate directly overlay fine-grained quartzites of the Conical Peak Quartzite . Maximum U and Th values are 14 ppm and 36 ppm. The presence of radioactive quartz-pebble conglomerates on the surface suggests that the radon anomalies may be accounted for by similar highly radioactive but discontinuous beds in the subsurface. However, no such beds were intersected by drill hole MB-11.

In the Crater Lake area, southwest of the Threemile Creek area, outcrops of basal Magnolia Formation consist of quartz-granule conglomerate which unconformably overlie the Colberg Metavolcanic Rocks. Maximum uranium and thorium values here are only 1 ppm U and 26 ppm Th respectively, and no holes were drilled in this area. Southwest of Crater Lake, in the area of MB-14, quartz-granule conglomerates contain 1.8 ppm U and 16 ppm Th in outcrop. In the area south of the three Bows holes, outcrops are scarce but a hole was drilled by the Bendix Field Engineering Corporation (MB-5) to a depth of 1429 feet. A 13 foot interval of quartz-pebble conglomerate was encountered near the top of the Magnolia Formation, with a maximum of 79 ppm U and 32 ppm Th. No conglomerates of the Conglomerate Member were encountered.

Thus for the most part in the central portion of the Medicine Bow Mountains, the basal section of the Magnolia Formation consists of quartz-granule conglomerates which are considered to be part of the Quartzite Member. Clasts in these conglomerates consist of quartz, quartzite and occasionally granite. The matrix is generally a micaceous quartzite similar to surrounding quartzites and contains rounded feldspar grains, including perthitic and chessboard plagioclase.

The Magnolia Formation crops out again north of Arrastre Lake along a northeast trending anticline shown in Figure 2.28. In this area, lenticular beds of radioactive quartz-pebble conglomerate are found within a thick sequence of arkosic paraconglomerates. These conglomerates can be traced for about 1000 meters on the east limb of the anticline with maximum U and Th values of 8 ppm and 38 ppm, and radiation values (22,000 cpm) which are 3.5 times background levels. The Conglomerate Member does not crop out on the west side of the anticline. Two holes were drilled in this area, PL-1 on the east flank of the anticline and GH-1 on the west flank of the structure.

On the east flank of the anticline the Conglomerate Member consists of about 633 feet of polymictic arkosic paraconglomerate which unconformably overlies the Colberg Metavolcanic Rocks. The conglomerates are only mildly radioactive with subsurface maximum uranium values of 11 ppm and thorium values of 30 ppm. No highly radioactive quartz-pebble conglomerates were encountered. As in the Threemile Creek area, the radon content of dilute ground water was found by Miller and others (1977) to be significantly higher than can be accounted for by surface rocks. It is possible that laterally extensive, but low-grade paraconglomerates, and isolated, lenticular radioactive quartz-pebble conglomerates are the cause of the anomalous radon centent of ground water. The overlying rocks are very fine- to coarse-grained quartzites containing siltstone lenses, trough crossbeds and fining-upward sequences. These quartzites reach a thickness of at least 750 feet. Drill hole Pl-1 was terminated at a depth of 1205 feet, still in the Quartzite Member of the Magnolia Formation; no radioactive conglomerates were cored.

The Quartzite Member on the west flank of the anticline is predominantly a medium- to coarse-grained subarkose containing many individual thin conglomerate lenses as well as a few thin phyllitic layers. Bed forms include local trough crossbedding with fining upward sets, and ripples which occur primarily near the base.

The Conglomerate Member in the Arrastre Anticline area contains clasts up to 10 cm in diameter of quartz, quartzite, chloritic quartzite, some radioactive pyritic quartzite near the base, and some mafic and granite clasts near the top. The scarcity of granite clasts and of feldspar is noticeable in this section (see Table 2.9). Rarely the conglomerates contain up to 10% pyrite. The matrix is generally muscovitic and chloritic and often contains biotite megacrysts. The Quartzite Member contains scattered lenses of conglomerate and contains noticeably more feldspar than the Conglomerate Member (see Table 2.9); the feldspar is predominantly plagioclase and generally occurs as rounded grains. The quartzites are poorly sorted, muscovitic, chloritic, and biotitic, containing numerous megacrysts of biotite similar to quartzites of the east limb of the Arrastre Anticline, and to quartzites in the vicinity of EMB-5 (central area) and the Threemile Creek area. Small amounts of pyrite and hematite are also present in the quartzite member near the Arrastre Anticline.

The most southwestern outcrop of the Magnolia Formation in the Medicine Bow Mountains occurs in the area of Brush Creek, (Figure 2.28). As in other areas, outcrops are widely scattered, but quartz-pebble conglomerates and arkosic paraconglomerates are strongly radioactive (up to 5 times background). Interbedded granule, quartz-pebble, and paraconglomerates similar to those in the Arrastre Anticline area and in drill hole Bows-1 were drilled to a depth of 372 feet in drill hole MB-9R. These rocks are not very radioactive, with maximum values of 78 ppm U and 52 ppm Th. These conglomerates are interbedded with biotitic quartzite, phyllitic graywackies, and biotite-chlorite schists. Below the conglomerate zone is a section of micaceous quartzite and chlorite-biotite schist which is tentatively identified as the Conical Peak Quartzite of the Phantom Lake Metamorphic Suite. The conglomerates in the Brush Creek area contain clasts of quartz, quartzite, chlorite schist, and garnet schist; no granite clasts were observed. The matrix and quartzites in this section are muscovitic, chloritic, and biotitic containing lesser amounts of pyrite and hematite.

Regional trends from northwest to southwest, recognized in the Magnolia Formation of the Medicine Bow Mountains are: 1) a decrease in the maturity of conglomerates with paraconglomerates dominating over quartz-pebble conglomerates in the southwest, 2) a decrease in the number of granite clasts and K-feldspar content and an increase in the number of mafic clasts and plagioclase content 3) a slight increase in the amount of biotite and chlorite in the matrixes of conglomerates and overlying quartzites (except for Unit 1 at Onemile Creek), and 4) a decrease in radioactivity. This systematic change in character is in agreement with our view of the depositional environment for the Magnolia

Formation. We envision the presence of three or four separate but coalescing fans with different source areas, as indicated by differences in clast and quartzite composition. We attribute the higher radioactivity of the Onemile Creek area to a more granite-rich source area and to a greater accumulation of the more mature quartz-pebble conglomerates.

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PART THREE:

THE GEOLOGY OF ARCHEAN AND EARLY PROTEROZOIC TERRANES OF THE SIERRA MADRE, WYOMING

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INTRODUCTION

Part 3 is a detailed discussion of Precambrian rocks in the northern Sierra Madre which emphasizes the geologic history and uranium potential of Late Archean and Early Proterozoic metasedimentary rocks north of the Cheyenne Belt. The discussion of stratigraphy and sedimentary features, presented first, is a modification of stratigraphic interpretations of Graff (1978, 1979), and incorporates more recent mapping, sedimentological studies, and subsurface information. The basis for our proposed stratigraphic subdivisions is a new structural interpretation (Plates 5 and 6) which involves a major refolded synclinorium in the Archean Phantom Lake Metamorphic Suite unconformably overlain by a generally south-dipping Early Proterozoic section containing the Deep Lake and Libby Creek Groups which is greatly attenuated (relative to correlative strata in the Medicine Bow Mountains) due to movement along thrust faults.

As in the Medicine Bow Mountains, we interpret the Proterozoic depositional and tectonic history of the metasedimentary rocks in terms of rifting, plate margin sedimentation, and collisional orogenesis. These interpretations lean heavily on data from the Medicine Bow Mountains. Archean geologic history, as recorded in metasedimentary and metavolcanic rocks of the Phantom Lake Metamorphic Suite remains enigmatic despite detailed mapping by many workers and excellent exposures over large areas of the central Sierra Madre. Structural complexities in these rocks will need to be unraveled more fully before detailed stratigraphic studies of the Archean metasedimentary rocks are possible.

Stratigraphic and structural studies of the entire metasedimentary succession were necessary to help understand the nature of radioactive quartz-pebble conglomerates in the Sierra Madre. The main radioactive rocks, the Deep Gulch Conglomerate, crop out near the base of the Archean Phantom Lake Suite in the northwestern Sierra Madre. These rocks occupy part of a large overturned limb of the major synclinorium in the northwest Sierra Madre and, aside from the fact that they are overturned, can be dealt with as structurally simple, dipping strata. Attempts were made to study in detail the relationship of uranium and thorium mineralization to sedimentary features such as lithofacies, grain size, thickness of conglomerate horizons, and sedimentary structures. Results of these studies from the Deep Gulch Conglomerate demonstrate a fluvial, fossil-placer origin for the mineralization. Geochemical and mineralogical studies of the Deep Gulch Conglomerate indicate a predominance of thorium over uranium (with values ranging up to 720 ppm U and 2600 ppm Th) and the presence of monazite and huttonite(?) as the primary radioactive phases.

The Proterozoic Magnolia Formation, which contains highly radioactive rocks in the Medicine Bow Mountains, was drilled in several locations in the Sierra Madre. No significant mineralization was found.

GEOLOGIC SETTING OF METASEDIMENTARY ROCKS IN THE SIERRA MADRE

The Sierra Madre is a Precambrian-cored anticlinal uplift of Laramide age about ten kilometers southwest of the Medicine Bow Mountains which forms the northerly extension of the Park Range in northern Colorado (Figure 3.1). Precambrian rocks in the core of the range have been subdivided by Houston and others (1975) into three principal terranes shown in Figure 3.1: 1) a northern terrane containing Archean gneisses and granites; 2) a central terrane consisting of the Late Archean Phantom Lake Metamorphic Site and the Early Proterozoic Deep Lake and Libby Creek Groups; and 3) a southern terrane composed of late Early Proterozoic (1800-1700 m.y.) volcanogenic gneisses, syn-orogenic granites and layered gabbroic complexes. The central and southern terranes are separated by a shear zone which is believed to be the western extension of the Mullen Creek-Nash Fork shear zone in the Medicine Bow Mountains and part of the Cheyenne Belt, a major Proterozoic crustal boundary in southern Wyoming (Houston and others, 1979). The boundary between the northern and central terranes is variably marked by an unconformity or intrusive contact.

Early studies of the Precambrian rocks of the Sierra Madre were done by Hague (1877), Knight (1909), and Spencer (1904). Blackwwelder (1935) attempted to correlate the metasedimentary rocks of the Sierra Madre with his measured sections in the Medicine Bow Mountains. More recently, detailed studies of small areas in the Sierra Madre were undertaken by Short (1958), Wied (1961), Merry (1963), Ferris (1964), Lackey (1965), DeNault (1967), Huang (1970), Miller (1971), Ridgely (1971), Schuster (1972), and Hughes (1973). Ebbett (1970) produced a preliminary outcrop map of the central part of the range which was expanded into a more detailed map by Houston and others (1975). Graff (1978) completed outcrop mapping of the central Sierra Madre and his maps form the basis for new interpretations of Sierra Madre stratigraphy and structure shown in Plate 5. Gwinner (1979) extended Graff's outcrop maps into the northernmost part of the Sierra Madre. Regional syntheses for the Sierra Madre have been attempted by Divis (1976; 1977) and Houston and Ebbett (1977).

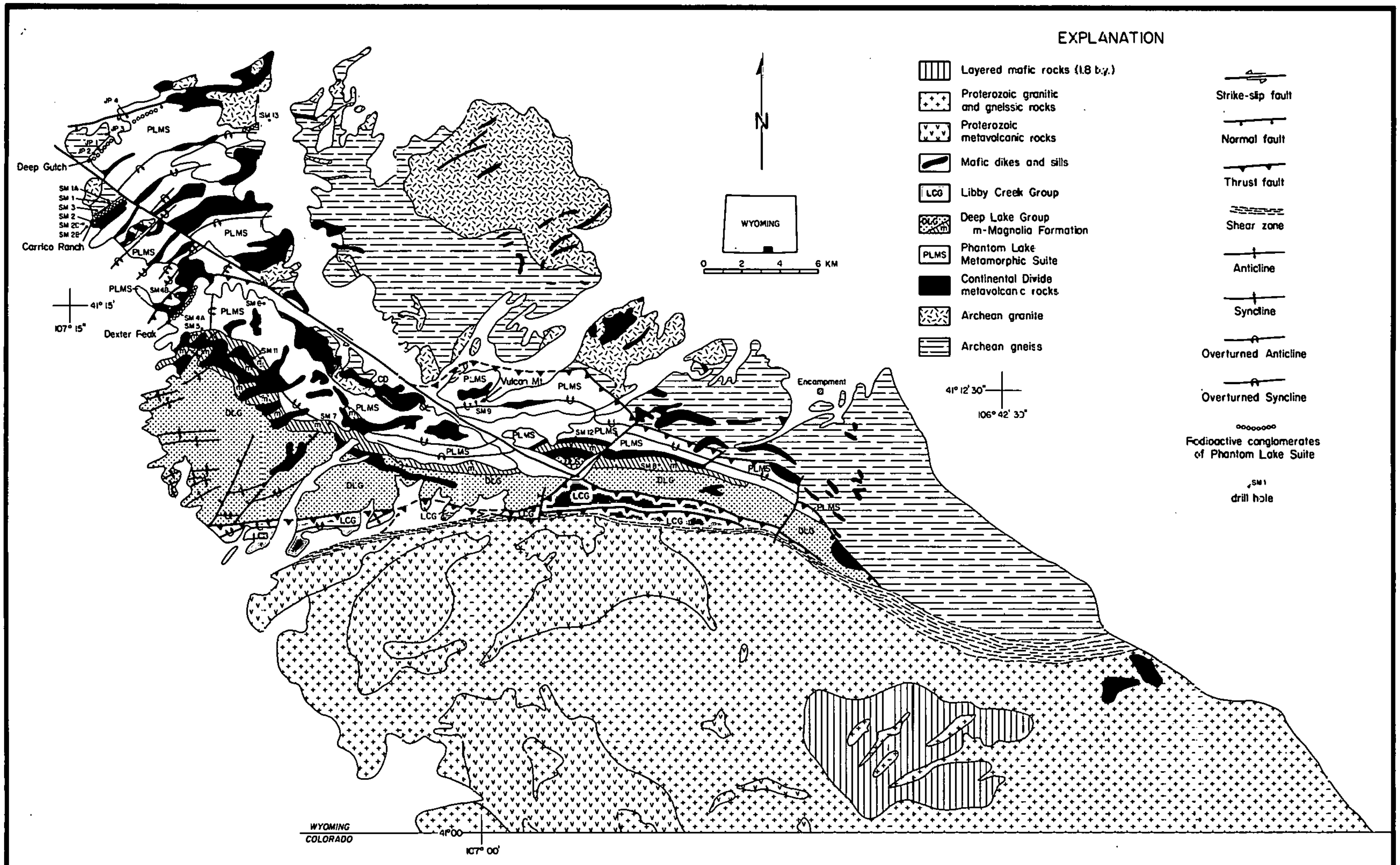


FIGURE 3.1. GENERALIZED GEOLOGIC MAP OF SIERRA MADRE, WYOMING

Recent studies on the economic geology of the Sierra Madre were done by Houston and others (1975, 1977, 1979) and Graff (1978, 1979). Geochronologic data are presented by Divis (1976, 1977) and Hills and Houston (1979).

STRATIGRAPHY AND SEDIMENTARY FEATURES OF METASEDIMENTARY ROCKS IN THE SIERRA MADRE

ARCHEAN "BASEMENT" ROCKS

The oldest rocks of the Sierra Madre are a complex of gneisses, metasedimentary rocks, and metavolcanic rocks. The contact relationships between these various rock types are poorly understood but the "basement" terrane probably includes supracrustal strata and intrusive rocks of several ages. These strata are grouped into three map units in Plate 5: biotite-plagioclase gneiss, Archean metasedimentary and metavolcanic rocks (undifferentiated), and Continental Divide Metavolcanic Rocks. All three are intruded by Late Archean (2700-2500 m.y.) granitic rocks and are therefore older than 2700 m.y.

Biotite-plagioclase Gneiss

Volumetrically, the most important of the oldest rocks are the biotite-plagioclase gneisses. This unit crops out along the northern flank of the range. It is a hybrid unit consisting of foliated, gray, biotite gneiss; pink, quartzo-feldspathic gneiss; thin quartzite pods; and amphibolites. The pink, quartzo-feldspathic gneisses commonly show intrusive characteristics and will be discussed under Archean granitic rocks. The quartzites and amphibolites will be discussed in the next section. The remainder of this section will discuss the dominant unit, the gray, biotite-plagioclase gneiss.

The well-foliated, gray gneisses consist of sodic plagioclase (An_{15-38}), quartz, biotite, and muscovite. Minor amounts of K-feldspar, chlorite, epidote, and hornblende are present as well as trace amounts of apatite, sphene, zircon, rutile, and opaques (Table 3.1; note that Hughes (1973) data are somewhat anomalous and may be due to misidentification of untwinned plagioclase as orthoclase). Many of the plagioclase grains have been partially altered to sericite and epidote. More detailed petrology can be found in Ferris (1964) and Miller (1971). The gray gneisses are primarily tonalites and trondhjemites (Figure 3.2 and Divis, 1976). They are too rich in alkalies to indicate a trend on an AFM diagram (Figure 3.3) but are chemically very similar to other Archean gray gneisses that are part of a calc-alkaline suite (Tarney, 1976).

The origin of the gray gneiss is obscure. A sedimentary origin is suggested by the lenses and pods of quartzite within the gneiss, the interleaved quartzo-feldspathic and biotite-rich layers in drill hole SM-2D, and the apparently gradational contact between some of the metasedimentary layers and the gneiss (Hughes, 1973). Divis (1976) suggested that the gneisses had a dacitic volcanic and volcaniclastic origin. Windley (1977) summarized the principal arguments concerning the origin of Archean gray gneisses. It is apparent that more research is required. The augen gneisses of Table 3.1 and Figure 3.2 probably represent a contact metamorphic phase of the gray biotite-plagioclase gneiss (Divis, 1976).

Archean Metasedimentary and Metavolcanic Rocks (Undifferentiated)

This map unit consists of quartzites, metagraywackes, marbles, amphibolites, metatuffs, chlorite-actinolite schists, and paragneiss. Outcrops range from a few square kilometers to thin lenses and

TABLE 3.1. PETROGRAPHY OF ARCHEAN BIOTITE GNEISSES AND AUGEN GNEISSES FROM THE SIERRA MADRE. DATA FROM DIVIS (1976) = *; FERRIS (1964), MILLER (1971), HUGHES (1973), AND POINT COUNTS.

				BIOTITE GNEISS					Secondary (after plagioclase)							An.
																content
Sample no.	Qtz.	Plag.	K-Spar	Musc.	Chl.	Biot.	Epid.	Hnbl.	Seri.	Epid.	Ap.	Sph.	Opaq.	Zir.	Rt.	of Plag.
SM2D-425	34.0	45.3	–	7.3	–	5.3	0.6	–	7.3	–	–	–	–	–	–	–
SM2D-468	25.8	17.1	–	3.9	Tr	41.5	8.3	–	2.0	1.0	–	–	–	–	–	–
SM2D-506.4	26.8	31.1	–	6.1	1.8	27.4	1.8	–	4.9	–	–	–	–	–	–	–
SM35	37.0	53.3	–	3.0	2.4	–	0.6	–	3.6	–	–	–	–	–	–	–
SM36	25.3	60.8	–	–	4.2	7.2	1.2	–	–	1.2	–	–	–	–	–	–
SM47	38.6	25.9	1.0	1.5	Tr	13.7	8.6	9.1	0.5	1.0	–	–	–	–	–	–
SM48	24.7	50.6	Tr	16.1	–	4.0	4.0	–	0.6	–	–	–	–	–	–	–
SM55	25.0	39.1	17.3	–	–	9.0	–	–	9.0	0.6	–	–	–	–	–	–
SM100	55.3	27.0	–	8.8	–	6.9	1.9	–	–	–	–	–	–	–	–	–
BG101*	26.4	58.0	–	5.6	–	6.3	3.7	–	–	–	–	–	–	–	–	An25
BG150*	25.9	63.2	–	4.8	–	6.1	Tr	–	–	–	–	–	–	–	–	An25
BG157*	24.2	58.7	–	16.3	–	Tr	0.8	–	–	–	–	–	–	–	–	An23
MEAN (12)	30.8	44.2	1.5	6.1	0.7	10.6	2.6	0.8	2.3	0.3	–	–	–	–	–	–
Ferris; Mean (10)	34	41	8	3	Tr	12	2	–	–	–	Tr	Tr	Tr	–	–	An15-38
Miller; Mean (4)	35.5	22.1	2.0	Tr	Tr	20.2	10.0	7.5	–	–	Tr	Tr	Tr	Tr	Tr	An28-31
Hughes; Mean (4)	29	17	39	Tr	–	10	–	–	–	–	Tr	–	Tr	Tr	–	–
							AUGEN GNEISS									
QF-110*	33.7	46.5	15.6	1.7	2.5	–	–	Tr	–	–	–	–	–	–	–	An12
QF-111*	30.9	49.5	19.6	Tr	–	–	–	Tr	–	–	–	–	–	–	–	An8
QF-152*	27.8	42.3	12.3	7.6	8.4	Tr	1.6	–	–	–	–	–	–	–	–	An12
MEAN (3)	30.8	46.1	15.8	3.1	3.6	Tr	0.5	Tr	–	–	–	–	–	–	–	–
Hughes	21	15	50	3	12	Tr	–	Tr	Tr	Tr	Tr	–	–	–	–	–

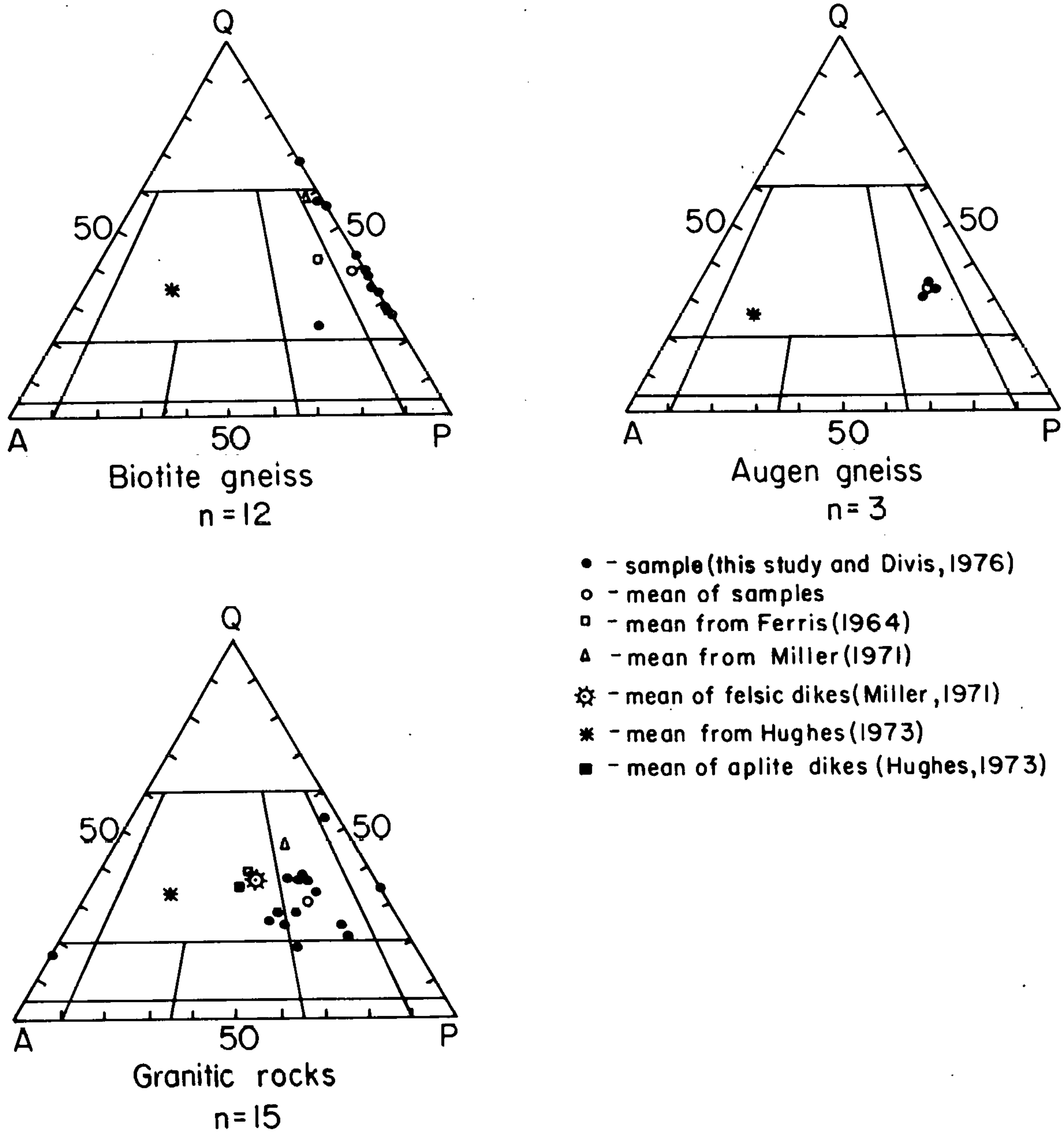
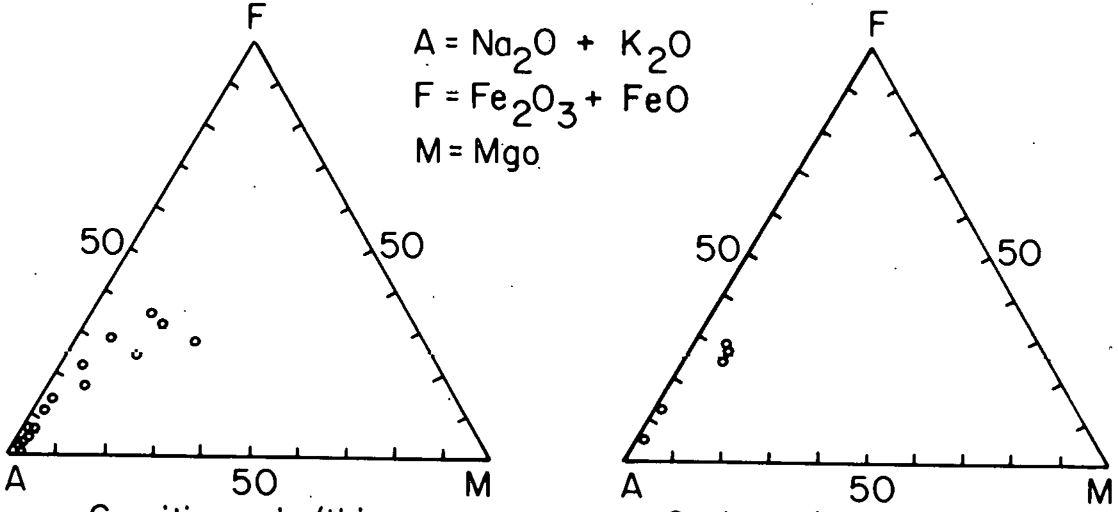


Figure 3.2. Quartz (Q), alkali feldspar (A), and plagioclase (P) ternary diagrams for gneisses and granitic rocks (after Streckeisen, 1973).

pods a few meters across. Gradational contacts are common where more than one lithic type is present in any one area. Primary structures are rarely preserved.



Granitic rocks (this study and Divis, 1976) n=15

Gneisses (from Divis, 1976) n=5

Figure 3.3. AFM ternary diagrams for Archean gneisses and granitic rocks from the Sierra Madre.

Petrographic data from a few samples is shown in Table 3.2. The quartzites are arkosic, usually medium-grained with plagioclase the dominant feldspar (Figure 3.4). One sample contains significant amounts of mafic minerals and is interpreted to be of tuffaceous or volcaniclastic origin. The two mafic (ultramafic?) samples have an unknown origin as no primary structures are present. Marbles are present in paragneiss and quartzites adjacent to the Continental Divide Metavolcanic Rocks. Paraconglomerates are present at Deep Gulch. Metagraywacke and paragneiss are prevalent at the northern end of the Sharp Hill Quadrangle.

TABLE 3.2. PETROGRAPHY OF ARCHEAN METASEDIMENTARY ROCKS UNDIVIDED; MODES FROM VISUAL ESTIMATES.

Sample no.	Qtz.	Plag.	K-Spar.	Musc.	Chlor.	Biot.	Epid.	Opaq.	Sph.	Act.	Hnbl.	Talc	Rock name
SM13-263	5	—	—	—	10	—	—	Tr	—	—	85	—	Amphibolite
SM34	65	25	—	10	—	—	—	—	—	—	—	—	Quartzite
SM52	55	40	—	5	—	—	—	—	—	—	—	—	Quartzite
SM49	70	10	10	10	—	—	—	—	—	—	—	—	Quartzite
SM39	45	35	—	8	—	5	5	—	—	2	—	—	Metatuff
SM97	—	—	—	—	50	—	—	—	1	30	—	19	Schist

Most of these rocks have an undoubted sedimentary origin but their relationship with surrounding strata is unclear. Locally, these rocks have very gradational contacts with the gray, biotite gneisses and the Continental Divide Metavolcanic rocks. In at least one locality (Deep Gulch) an intrusive contact is apparent with the gray gneiss. In areas where primary structures are preserved (as in the northeast corner of the Divide Peak Quadrangle) these quartzites may be infolded remnants of the Jack Creek Quartzite of the Phantom Lake Metamorphic Suite. It is probable that more than one age of deposition is represented by rocks of this map unit.

Continental Divide Metavolcanic Rocks

The Continental Divide Metavolcanic Rocks (a new name) consists primarily of fine-grained amphibolite, metabasalt, and hornblende gneiss but also includes quartzite, schist, and marble. These rocks are best exposed in the headwaters of South Spring Creek, North Spring Creek, and Jack Creek. Hornblende gneiss adjacent to Savery Creek is also interpreted to be part of this unit. These rocks lie stratigraphically between the Phantom Lake Metamorphic Suite and Archean Metasedimentary and metavolcanic rocks (Plate 5).

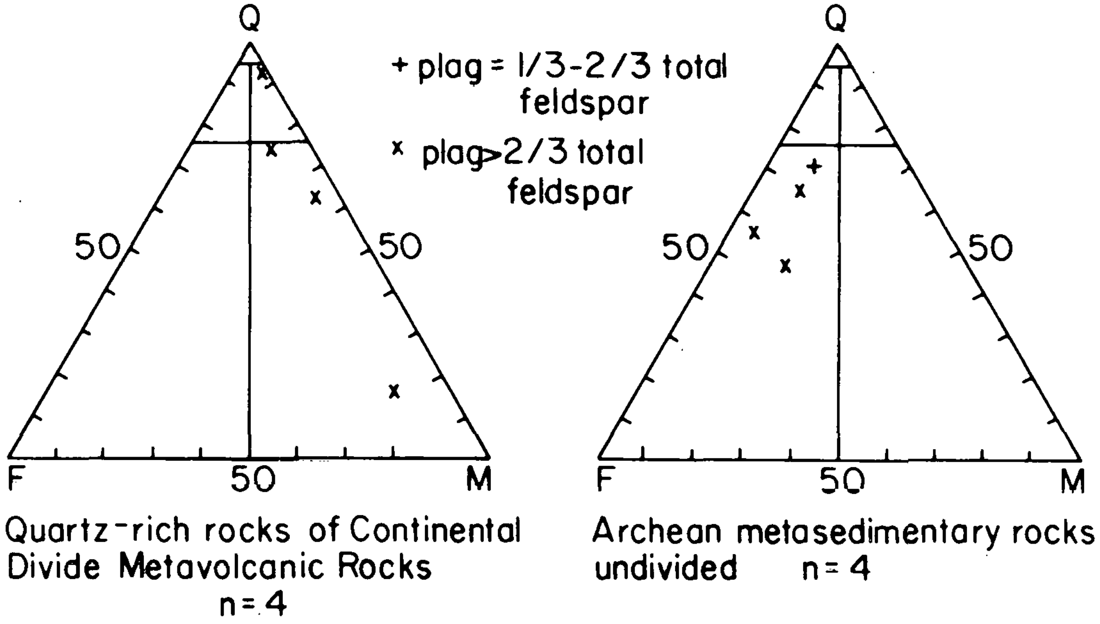


Figure 3.4. Quartz (Q), feldspar (F), and mica + chlorite (M) ternary diagrams for miscellaneous Archean metasedimentary rocks (after Folk, 1968)

Petrographic data of several lithic types is given in Table 3.3. The metabasalts consist primarily of hornblende with quartz, quartz-plagioclase, and epidote. Two volcaniclastic samples contain approximately equal amounts of quartz and hornblende suggesting derivation from both silicic and mafic rocks. Metasediments include a biotite schist and subargillaceous quartzite (Figure 3.4). Feldspar is scarce in most rocks of this unit. The mineralogy of the samples reflect epidote-amphibolite to amphibolite facies of metamorphism (Miyashiro, 1973).

TABLE 3.3. PETROGRAPHY OF THE CONTINENTAL DIVIDE METAVOLCANIC ROCKS; MODES FROM VISUAL ESTIMATES.

Samp. no.	Qtz.	Plag.	Chlor.	Biot.	Musc.	Epid.	Gnt.	Opaq.	Carb.	Hnbl.	Sph.
METABASALTS											
TS-241	15	1	–	3	–	5	–	–	–	75	1
TS-245	17	5	Tr	–	–	10	–	1	2	65	–
TS-183	5	10	Tr	–	–	–	–	–	–	85	–
MEAN (3)	12	5	Tr	1	–	5	–	Tr	1	75	Tr
VOLCANICLASTIC ROCKS											
SM6-443.4	40	2	1	20	–	–	1	–	2	34	–
TS-184	44	5	5	3	–	–	1	2	–	40	–
MEAN (2)	42	3.5	3	11.5	–	–	1	1	1	47	–
METAPELITES AND QUARTZITES											
SM6-458	89	1	–	2	4	–	–	1	3	–	–
SM6-530	15	10	5	63	–	–	–	2	5	–	–
MEAN (2)	52	6	2	32	2	–	–	2	4	–	–

Sedimentary structures have not been observed in this unit. Pillow basalts have been observed in two localities and along with interbedded thin marbles, imply a marine origin for at least part of the unit. The pillows coupled with structural data suggest that the metabasalt is younger than some of the Archean metasediments and paragneiss in which it is in contact. The Continental Divide Metavolcanic Rocks are about 360 meters thick but structural complexities make this figure questionable.

PHANTOM LAKE METAMORPHIC SUITE

The Phantom Lake Metamorphic Suite was named by Karlstrom and Houston (1979a, 1979b) and extended to the Sierra Madre by Graff (1979). In this paper we propose major revisions in the stratigraphy of this unit in both the Sierra Madre and the Medicine Bow Mountains. Several ambiguities remain in stratigraphy and structure, and for this reason we retain the term metamorphic suite. As shown in Figure 3.5, the Phantom Lake Metamorphic Suite in the Sierra Madre is subdivided into the Jack Creek Quartzite, Silver Lake Metavolcanic Rocks, and Bridger Peak Quartzite, each of which is described below.

Jack Creek Quartzite

The Jack Creek Quartzite is modified from the Jack Creek Formation of Graff (1978). It is herein defined to include all of the metasedimentary rocks at the base of the Phantom Lake Metamorphic Suite (beneath the Silver Lake Metavolcanic Rocks). It consists predominantly of quartzite but contains lenses of phyllite, marble, paraconglomerate, quartz-pebble conglomerate, and metagraywacke. The Deep Gulch Conglomerate is a pyritic and radioactive, quartz-pebble conglomerate facies at the base of the Jack Creek Quartzite. It is exposed on the west and north edges of the Divide Peak Synclinorium, and is discussed in detail later. Excellent exposures of the Jack Creek Quartzite can be found in Jack Creek Canyon and the cirque above North Spring Creek Lake (Sections 5, 8, and 9, T. 14 N., R. 86 W). Figure 3.6 is a measured section from the Carrico Ranch area (Sec. 7, T. 15 N., R. 87 W). Because of faulting, this is probably not a complete section. Drill holes JP1, 2, 3, and 4, SM-1, 1A, 2, 2D, and 3 intersected the Deep Gulch Conglomerate of the Jack Creek Quarzite. Drill hole SM-6 intersected a non-conglomeratic basal section of the quartzite. Drill holes SM-4A and 4B intersected (we believe) the upper contact of the unit. The Jack Creek Quartzite is about 700 meters thick in the Divide Peak Synclinorium and is about 300 meters thick near South Spring Creek Lake.

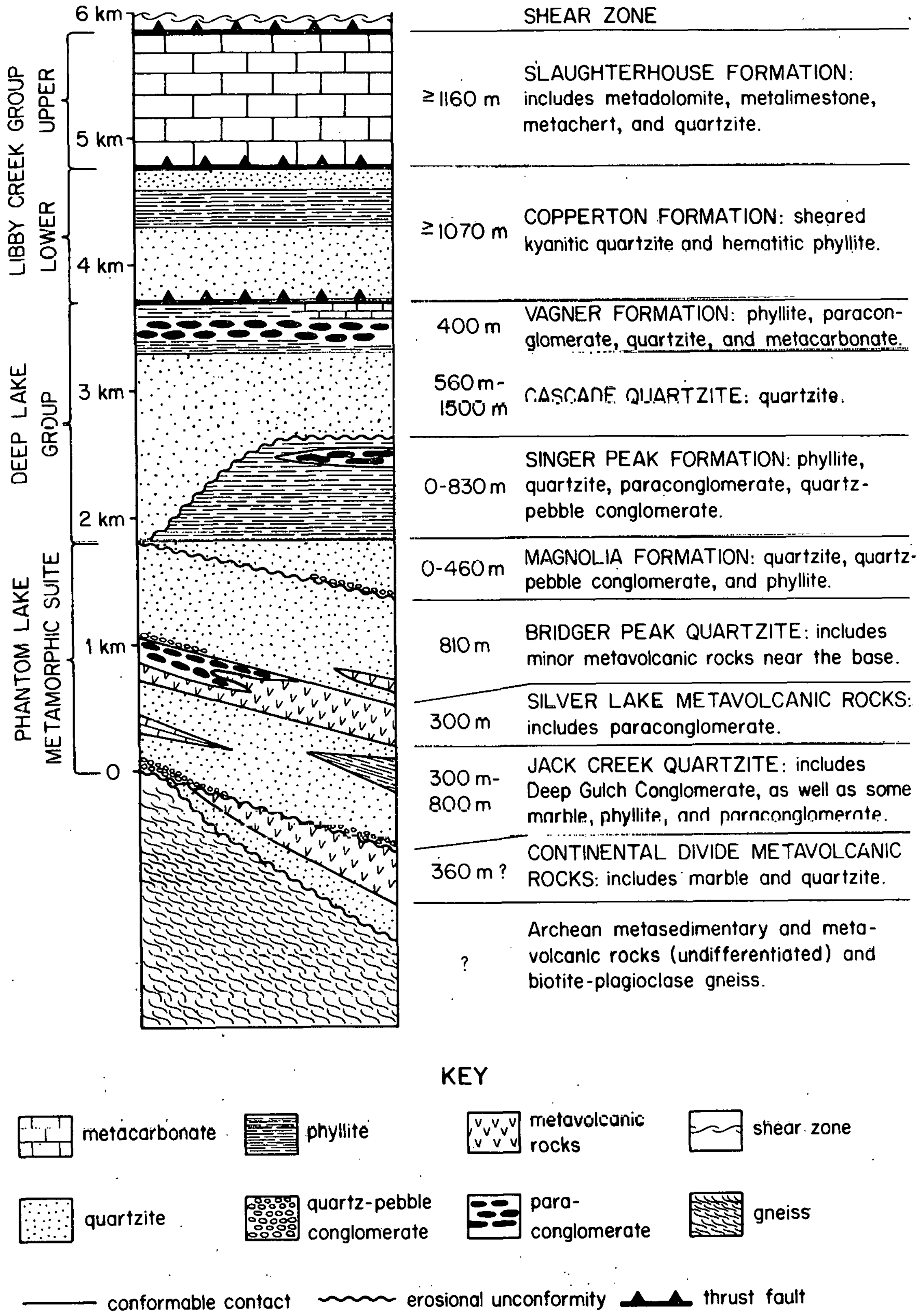


Figure 3.5. Stratigraphic column for Precambrian rocks in the Sierra Madre, Wyoming.

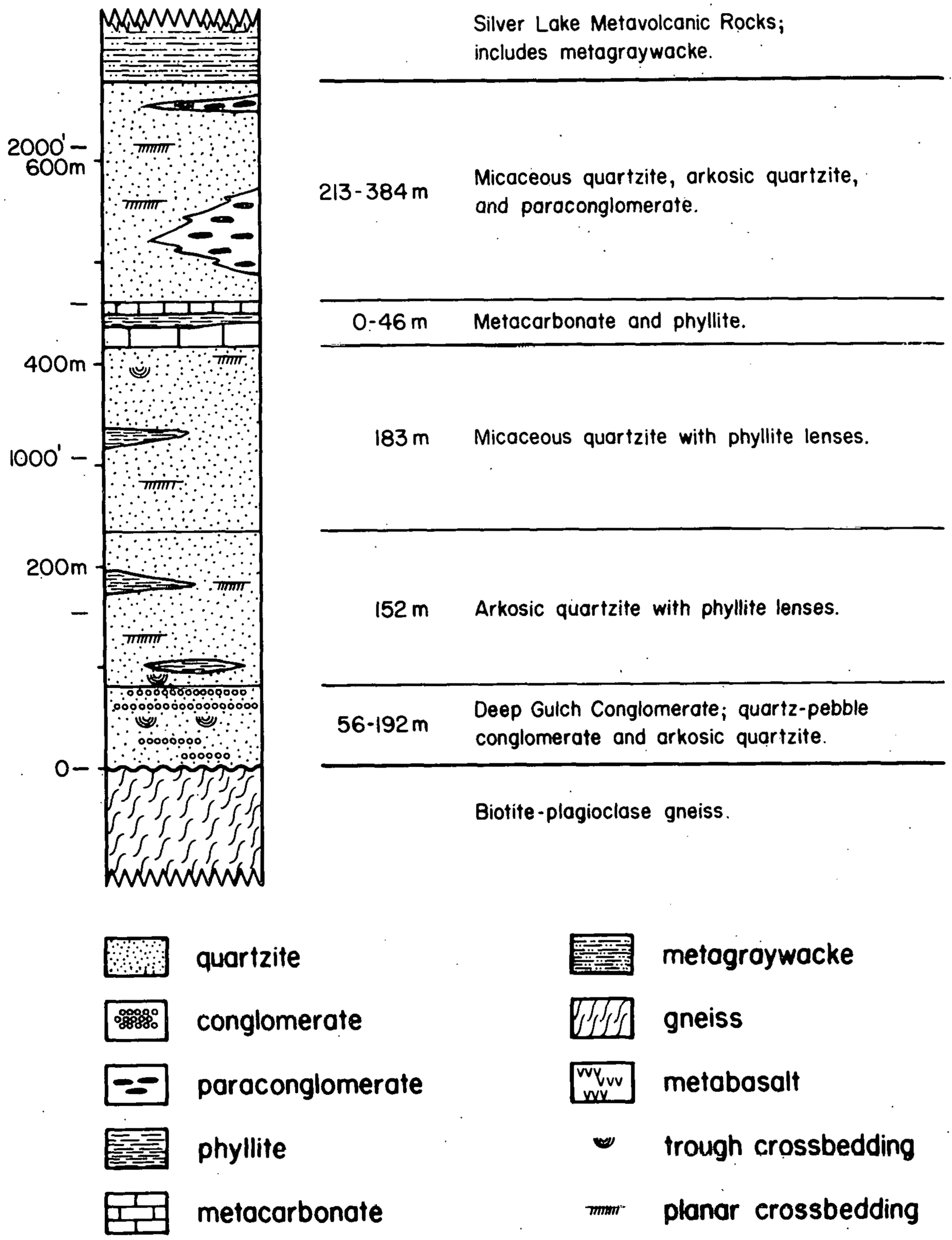


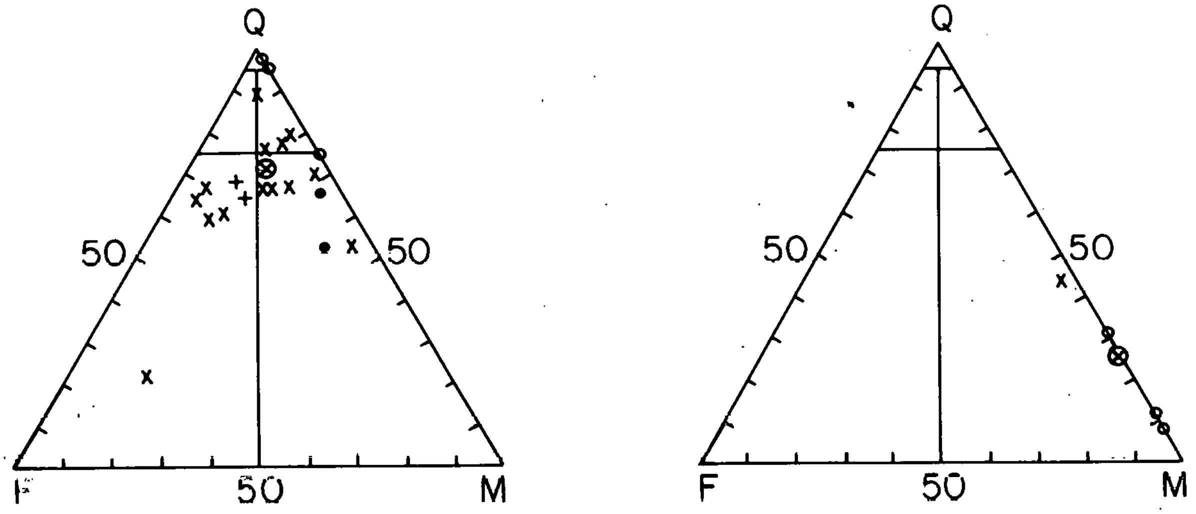
Figure 3.6. Stratigraphic section for the Jack Creek Quartzite in the Carrico Ranch area, northwestern Sierra Madre.

Petrographic data are shown in Table 3.4. The quartzites are generally arkosic, argillaceous, or sub-argillaceous (Figure 3.7). Plagioclase is the dominant feldspar with very little K-feldspar. Muscovite and biotite are usually present with generally minor amounts of chlorite, epidote, carbonate, and garnet. Metapelites consist primarily of muscovite, chlorite, or biotite with quartz and garnet. Plagioclase, epidote, staurolite, and kyanite may be present locally (Table 3.4). Metacarbonates are commonly very siliceous. This siliceous material is probably detrital and is arkosic (Figure 3.7). The carbonates are unusual in that K-feldspar is abundant. Jack Creek rocks are generally of the epidote-amphibolite facies of metamorphism in the Divide Park Synclinorium and amphibolite facies in the Singer Peak and Bridger Peak quadrangles (Miyashiro, 1973).

Sedimentary structures are well preserved over much of the outcrop area of the Jack Creek Quartzite. Trough crossbedding is prominent in the Deep Gulch Conglomerate with tabular planar crossbedding dominant elsewhere. Plane beds are common and wedge-shaped and herringbone cross-stratification have been observed. Paleocurrent directions were measured and attempts were made to restore bedding to pre-folding attitudes (following Ramsey, 1961). However, the complex nature of the folding leaves doubt as to the veracity of the absolute directions; it is probable that rotation about a plunging concentric fold axis is too simplistic in these rocks. Field relations indicate that relative paleocurrent directions within a close cluster of outcrops are correct. A dominant north-northeast paleocurrent trend is suggested for the sediments of the Jack Creek Quartzite above the Deep Gulch Conglomerate (Figure 3.8). This

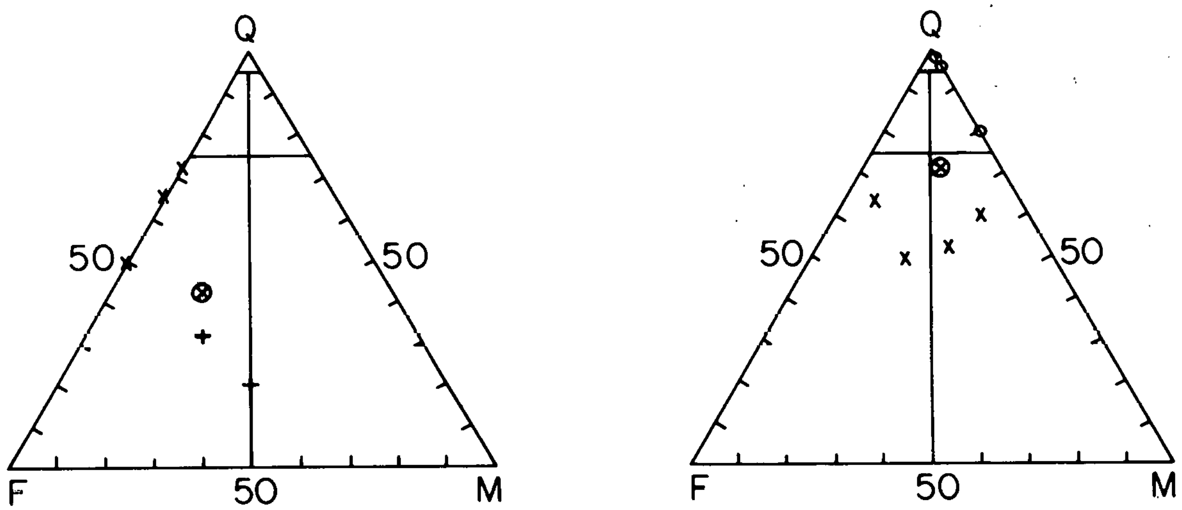
TABLE 3.4. PETROGRAPHY OF THE JACK CREEK QUARTZITE OF THE SIERRA MADRE; WITHOUT THE DEEP GULCH CONGLOMERATE; MODES FROM VISUAL ESTIMATES.

Sample no.	Qtz.	Plag.	K-spar.	Musc.	Chlor.	Biot.	Epid.	Carb.	Opaq.	Trem.	Other
QUARTZITES											
SM-41	65	30	–	–	5	–	–	–	–	–	
SM-73	65	10	10	15	–	–	–	–	–	–	
SM-89	68	20	5	–	1	5	–	–	–	–	
SM-96	94	–	–	5	–	–	–	–	–	–	
Meta-1	90	5	–	5	–	–	–	–	Tr	–	
TS-205	20	55	–	–	–	15	10	Tr	–	–	
TS-207	60	30	–	8	–	2	–	–	–	–	
TS-209	55	25	–	5	–	5	–	10	–	–	
SM4A-99.5	68	15	–	15	–	2	–	–	–	–	
SM4A-133.5	69	10	–	20	Tr	1	–	–	–	–	
SM4A-156.5	79	5	–	10	3	3	–	–	–	–	
SM4A-311.5	76	10	–	10	–	2	2	Tr	–	–	
SM4A-321	80	3	–	15	1	1	–	Tr	–	–	
SM4A-497.5	68	15	–	10	2	5	–	Tr	–	–	
SM4B-184	53	5	–	40	–	2	–	–	–	–	
SM4B-241.2	60	3	–	15	5	2	–	Tr	Tr	–	
SM6-103.1	96	–	–	3	–	1	–	–	–	–	
SM6-165.5	67	10	10	10	–	1	–	2	–	–	
SM6-299.5	66	–	–	10	2	10	–	10	1	–	1% Tour.
SM6-310	65	–	5	15	–	15	–	Tr	–	–	Ap, Zir.
SM6-340	51	–	10	35	–	3	–	–	1	–	Tr. Zir.
MEAN (21)	67	12	2	12	1	4	1	1	Tr	–	
CALCAREOUS ROCKS											
SM-90	3	2	–	2	–	4	–	89	Tr	–	
Meta-2	29	10	1	–	–	–	35	10	–	15	
Meta-3	25	15	20	5	–	15	–	20	–	–	
Meta-4	20	10	10	–	–	–	15	35	–	10	
Carb-1	20	10	–	–	Tr	Tr	–	70	Tr	10	
Carb-2	10	10	10	10	–	10	–	50	Tr	–	
MEAN (6)	18	10	7	3	Tr	5	8	46	Tr	6	
METAPELITES											
SM-43	10	–	–	–	83	5	–	–	2	–	
SM-69	40	5	–	25	10	5	–	–	–	–	15% Garnet
SM-70	27	–	–	60	2	–	–	–	1	–	5% Gnt., 5% Staur.
SM6-178.7	5	Tr	–	–	15	44	5	–	–	–	30% Gnt., 1% Kyan.
MEAN (4)	20	1	–	21	28	14	1	–	1	–	



Jack Creek quartzites
n=21

Jack Creek metapelitic rocks
n=4



Siliceous fraction Jack Creek carbonates
n=5

Bridger Peak Quartzite
n=7

o - no feldspar
• - plagioclase < 1/3 total feldspar
+ - plagioclase >1/3,< 2/3 total feldspar
x - plagioclase >2/3 total feldspar
⊗ - mean

Figure 3.7. Quartz (Q), feldspar (F), and mica+ chlorite (M) ternary diagrams for the Jack Creek and Bridger Peak Quartzites (after Folk, 1968)

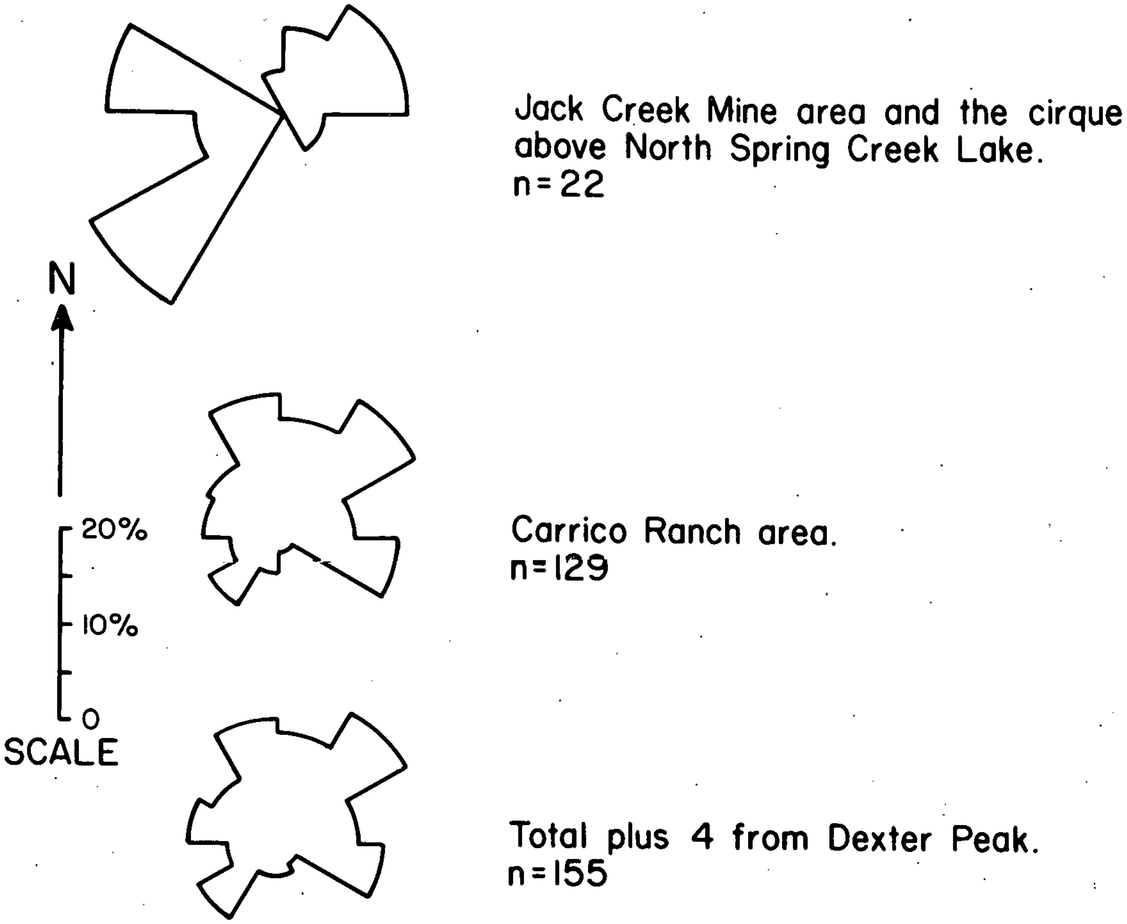


Figure 3.8. Paleocurrent rose diagrams for the Jack Creek Quartzite, excluding Deep Gulch Conglomerate.

northeast direction is similar to that of correlative sediments of the Bow River Quartzite in the Medicine Bow Mountains (discussed earlier).

In the Carrico Ranch area, the Jack Creek Quartzite consists of the 100 m thick Deep Gulch Conglomerate overlain by approximately 500 meters of relatively clean, fine- to medium-grained, sericitic quartzites (Figure 3.6). Beds of phyllite and metacarbonate are locally present. The metacarbonate is commonly coarse-grained, contains abundant clastic detritus (including a chert-pebble conglomerate) and appears crossbedded in places, all suggesting deposition in shallow water. One structure resembling organic bioherms was noted. Petrographic data is lacking for the quartzites of this section but field evidence suggests that these sediments are much more mature than the fluvial sediments of the basal Deep Gulch Conglomerate. The interbedded phyllites and metacarbonates and the broadly dispersed paleocurrent distribution suggest marine deposition for these rocks.

On-strike with the metacarbonate and only a few hundred meters to the southwest, the Jack Creek Quartzites consist of interbedded coarse-grained arkosic quartzite and paraconglomerate. Here, paleocurrents suggest a westerly source. The rapid facies change from coarse-grained, poorly sorted paraconglomerates, possibly debris flows (Bull, 1972), to marine carbonates suggests deposition in a fan-delta setting (Wescott and Ethridge, 1980).

The Jack Creek Quartzite in the cirques above North Spring Creek Lake and the upper end of Jack Creek Canyon does not contain a fluvial conglomerate facies at its base. Rather, the quartzites here are fine-grained, sericitic, and calcareous. Several large-scale crossbeds and herringbone cross-stratification are present along with abundant smaller-scale, planar crossbeds. Paleocurrent distribution is bimodal northeast-southwest (Figure 3.8). Other metasediments in this area include well-laminated, fine-grained, graded quartzites; thinly bedded, alternating beds of metacarbonate and very fine-grained quartzite (metachert?); and phyllites. These lithologies suggest deposition on a shallow marine shelf under the influence of tidal currents.

Silver Lake Metavolcanic Rocks

The Silver Lake Metavolcanic Rocks is expanded from Graff's (1978) Silver Lake Conglomerate to include his Spring Lake Volcanics. We have adopted this stratigraphic nomenclature because of the interbeddded relationship between these two units in outcrop. The Silver Lake Metavolcanic Rocks consist of mafic (and ultramafic?) metavolcanic rocks, granite-boulder paraconglomerate, biotite schist, metagraywacke, metatuff, quartzite, and metacarbonate. This unit is well exposed in the cliffs between South Spring Creek Lake and Silver Lake. Generally poor exposures and complex structure make it difficult to measure the thickness of this unit accurately. Over most of the study area this unit appears to be about 300 meters thick. The paraconglomerates appear to be thickest just south of Bridger Peak while the metagraywacke and tuffaceous rocks are thickest on the north flanks of the Divide Peak Synclinorium. Rapid lateral and vertical facies changes are common in this unit. Various levels of this unit were intersected by drill holes Sm-7, 9, 11, and 12 (see Volume 2 for generalized lithologic logs).

Petrographic data is given in Table 3.5. Mafic and ultramafic rocks are dominated by either actinolite or hornblende. Chlorite, garnet, and talc commonly are present in the ultramafic rocks while chlorite, biotite and plagioclase are common constituents of the mafic rocks. Several of

TABLE 3.5. PETROGRAPHY OF THE SILVER LAKE METAVOLCANIC ROCKS FROM THE SIERRA MADRE; MODES FROM VISUAL ESTIMATES

METABASALTS

Sample no.	Qtz.	Plag.	Chlor.	Biot.	Epid.	Opaq.	Garn.	Carb.	Act.	Hnbl.	Talc.	Stilp.	Sph.
AFCR 13-2	3	—	5	3	1	1	—	Tr	—	87	—	—	—
AFCR 13-3	27	—	1	—	2	—	5	—	—	65	—	—	—
AFCR 13-4	1	—	14	—	—	—	—	—	85	—	—	—	—
AFCR 7-2	2	1	—	—	3	1	—	—	—	93	—	—	—
AFDG -3	—	—	15	—	—	—	—	—	85	—	—	—	—
AFDG -4	5	—	—	—	—	2	20	—	—	73	—	—	—
SM-6	7	—	—	—	8	Tr	—	—	—	85	—	—	—
SM-10	10	10	—	—	2	2	1	10	—	65	—	—	—
SM-12	—	—	40	—	—	—	—	—	40	—	20	—	—
SM-14	—	—	5	—	—	—	—	—	95	—	—	—	—
SM-32	—	—	45	—	—	Tr	—	—	45	—	10	—	Tr
TS-147	5	10	5	15	—	—	—	—	65	—	—	—	—
TS-151	13	2	—	—	—	—	—	—	75	—	—	10	—
TS-172	—	35	—	1	—	—	—	—	—	64	—	—	—
SM9-289	20	—	—	—	15	—	—	Tr	60	—	—	—	5
SM11-1004.8	15	15	Tr	5	—	—	Tr	—	—	65	—	—	—
MEAN (16)	7	5	8	2	2	Tr	2	1	34	37	2	1	Tr

PARACONGLOMERATES

Sample no.	Qtz.	Plag.	Musc.	Chlor.	Biot.	Epid.	Opaq.	Gt.	Carb.	Act.	Hnbl.	Stilp.	Chltd.
SM-84	40	10	—	1	1	—	1	2	—	45	—	—	—
TS-222	59	20	—	—	15	1	—	—	—	—	5	—	—
TS-158	5	—	—	—	—	10	—	—	—	—	35	—	—
TS-159	20	20	—	2	5	—	—	—	—	52	—	1	—
TS-225	40	15	5	4	—	20	1	—	—	—	15	—	—
TS-295	50	20	15	—	5	6	—	—	—	—	4	—	—
TS-297	2	—	20	41	15	—	2	—	—	—	10	—	10
TS-300	15	—	—	2	75	3	—	3	2	—	—	—	—
TS-318	45	15	—	3	—	—	2	—	—	35	—	—	—
TS-134	84	—	—	5	5	—	—	1	—	—	5	—	—
TS-136	72	10	—	1	10	—	Tr	—	5	—	2	—	—
TS-133	39	20	—	—	Tr	—	—	1	5	—	35	—	—
TS-137	57	15	3	3	10	2	—	5	Tr	—	5	—	—
SM7-391.8	30	—	—	5	50	Tr	—	10	5	—	—	—	—
SM7-442.2	53	5	—	—	10	—	1	Tr	1	30	—	—	—
SM9-219.5	45	45	—	5	3	—	Tr	Tr	2	—	—	—	—
SM9-501.5	49	20	—	20	7	—	1	—	3	—	—	—	—
SM9-678.3	87	5	—	1	5	1	1	—	—	—	—	—	—
MEAN (18)	4[illegible]	12	2	5	12	2	Tr	1	1	9	9	Tr	Tr

METAGRAYWACKES AND METATUFFS

Sample no.	Qtz.	Plag.	K-spar.	Musc.	Chlor.	Biot.	Epid.	Opaq.	Garn.	Carb.	Hnbl.
SM12-589	52	30	—	15	3	—	—	—	—	—	—
SM-8	75	—	10	Tr	—	15	—	—	—	—	—
SM-82	30	—	—	30	—	1	—	4	25	—	—
SM-83	50	—	—	15	1	—	—	—	34	—	—
SM-86	30	48	—	—	1	15	1	1	4	—	—
AFDG-1	55	20	—	—	—	10	10	—	—	—	5
AFDG-2	55	20	—	—	—	—	15	—	—	—	10
SM11-729	70	16	—	Tr	1	10	1	—	1	1	—
SM11-862	40	20	—	—	—	5	30	—	5	—	—
SM-16	59	20	—	Tr	10	5	Tr	1	—	—	5
SM-30	50	35	—	—	3	3	1	5	—	—	3
AFCR 18-1	60	—	—	16	3	11	—	—	10	—	—
TS-299	60	5	—	3	8	10	1	6	5	—	5
SM7-276.2	30	4	—	—	—	1	—	—	—	—	65
AFDG-5	48	5	—	—	1	—	1	—	—	10	35
SM12-640	55	15	5	20	—	5	—	—	—	—	—
SM12-637	78	5	10	5	—	2	—	—	—	—	—
TS-128-B	25	35	—	—	15	—	—	—	—	25	—
TS-175	25	60	—	3	1	3	—	—	3	5	—
MEAN (19)	50	18	1	6	2	5	3	1	5	2	7

METAPELITES

Sample no.	Qtz.	Plag.	Musc.	Chlor.	Biot.	Epid.	Opaq.	Garn.	Carb.	Act.
SM-[illegible]1	30	20	40	4	4	—	2	—	—	—
TS-239	40	5	29	15	5	3	1	—	2	—
SM5-573	30	12	1	5	35	15	—	—	2	—
SM7-334.5	44	10	—	—	35	—	—	1	—	10
SM7-559	72	5	—	—	15	5	—	3	Tr	—
SM9-255.5	73	5	—	1	20	—	1	Tr	—	—
SM9-430	60	10	2	15	5	—	3	5	—	—
SM9-596	55	8	—	—	35	—	2	—	Tr	—
MEAN (8)	50	9	9	5	19	3	1	1	Tr	1

MARBLE

Sample no.	Qtz.	Musc.	Biot.	Carb.
SM-[illegible]4	70	Tr	Tr	30

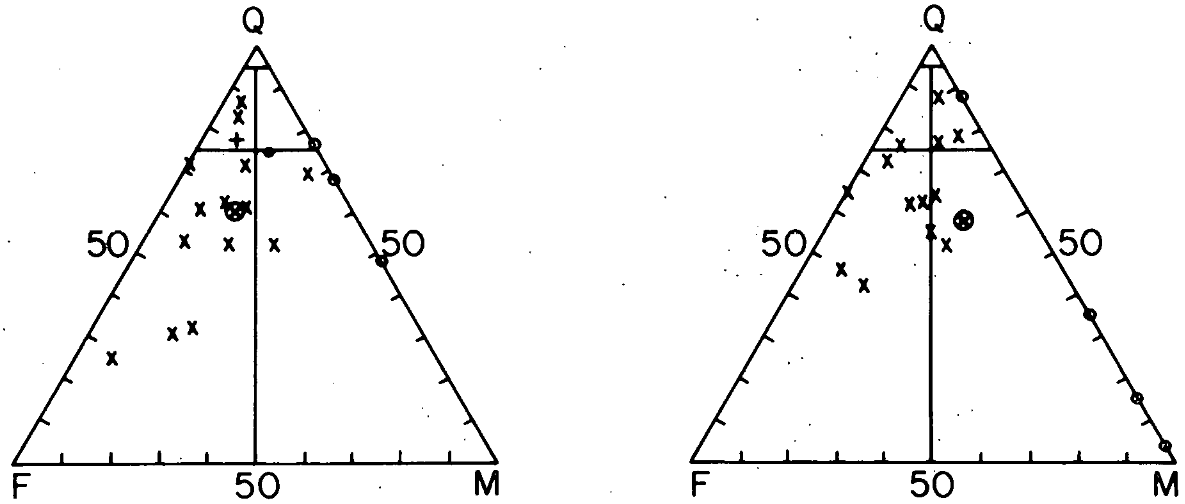
the samples contain considerably more quartz than one would expect for a mafic volcanic rock. We consider these rocks to be volcaniclastic in origin.

The Silver Lake paraconglomerates (Figure 3.9) consist of metasedimentary, granitic, and metavolcanic clasts (up to boulder-size) in an amphibolitic matrix. In contrast, the paraconglomerates in the underlying Jack Creek Quartzite are volumetrically less important, contain few metavolcanic clasts, and have a metapelitic matrix. Quartz, plagioclase, muscovite, chlorite, biotite, epidote, and amphibole are the most abundant constituents of the Silver Lake paraconglomerates. The siliceous fraction of these rocks is usually arkosic (Figure 3.10).

Pelitic rocks of the Silver Lake Metavolcanic Rocks consist of a mixture of mica, chlorite, quartz, and plagioclase (Table 3.5). These rocks are generally not as feldspathic as most of the Silver Lake rocks. Epidote, garnet, and actinolite may be abundant locally. The one "marble" examined petrographically is actually a very calcareous quartzite (Table 3.5).

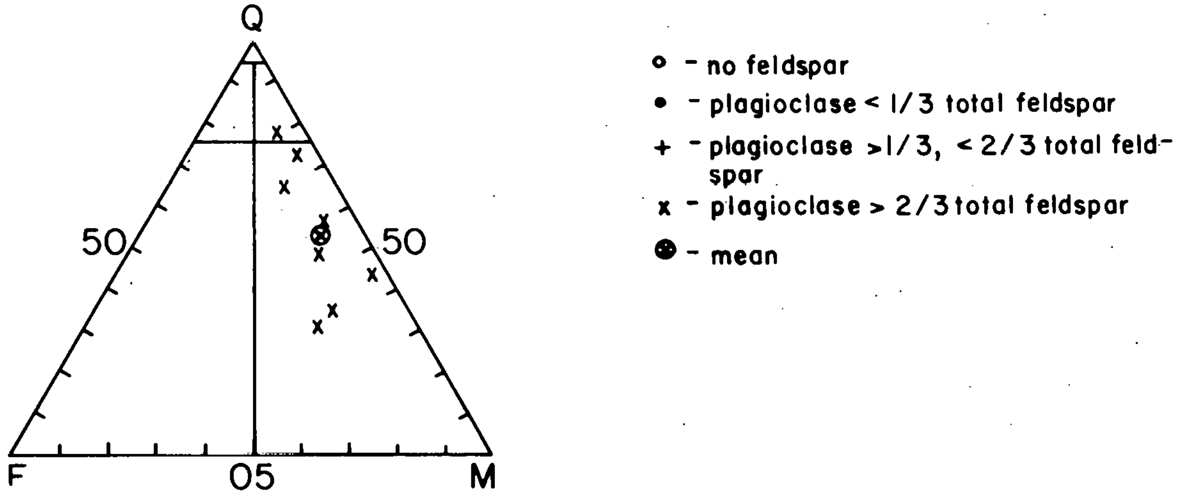
A variety of tuffaceous or volcaniclastic metasedimentary rocks are present in the Silver Lake Unit. The contain abundant quartz, feldspar, and mica as well as occasional chlorite, epidote, garnet, carbonate, and hornblende (Table 3.5). These rocks are generally very arkosic (Figure 3.10). Most are poorly sorted and some have a distinct bimodal grain size distribution with large, nearly euhedral, feldspars in a finer-grained arkosic matrix. Two samples (Sm-16 and SM-30) have a very fine-grained siliceous matrix with a few coarse crystals of

Figure 3.9. Photograph of Silver Lake Paraconglomerate in the Sierra Madre.



Silver Lake metagraywackes, quartzites, metatuffs, and volcaniclastic metasedimentary rock
n=19

Silver Lake paraconglomerates
n=17



Silver Lake metapelitic rocks
n= 8

Figure 3.10. Quartz (Q), feldspar (F), and mica+ chlorite (M) ternary diagrams for Silver Lake metasedimentary rocks (after Folk, 1968)

hornblende and feldspar. We consider these to represent metamorphosed dacitic welded tuffs (SiO_2 content is about 69%).

Few primary sedimentary or igneous structures have been observed in the Silver Lake rocks. We envision the environment as having been scattered mafic volcanic centers shedding lava flows and tuffaceous rocks into surrounding basins. Rapid facies changes and the poorly sorted nature of the sediments suggests rapid deposition, perhaps as alluvial fans, fan deltas, mud flows, or turbidity currents. The variety of clasts in the paraconglomerates implies uplift of pre-existing granitic and sedimentary strata, probably along fault scarps, in conjunction with volcanism.

Bridger Peak Quartzite

The Bridger Peak Quartzite is here restricted to Divis' (1976) usage. We interpret this unit to lie in the core of a recumbent, isoclinal syncline throughout the study area (Plate 5). This is the youngest unit of the Phantom Lake Metamorphic Suite and the top is not exposed. Over most of the study area, the quartzite is about 800 meters thick. The quartzite is well exposed at Bridger Peak and Vulcan Mountain and for several miles east of these two localities (Plate 5). This unit consists primarily of fine-grained quartzite with some phyllites. A few metavolcanic rocks and metacarbonates and at least one locality of pyritic quartz-pebble conglomerate are present near the base.

The quartzites are quite variable, ranging from quartz arenites to argillaceous and arkosic quartzites (Figure 3.7). Quartz, plagioclase, and muscovite are the most abundant species with minor amounts of chlorite and biotite (Table 3.6). A quartz-pebble conglomerate is present on Vulcan

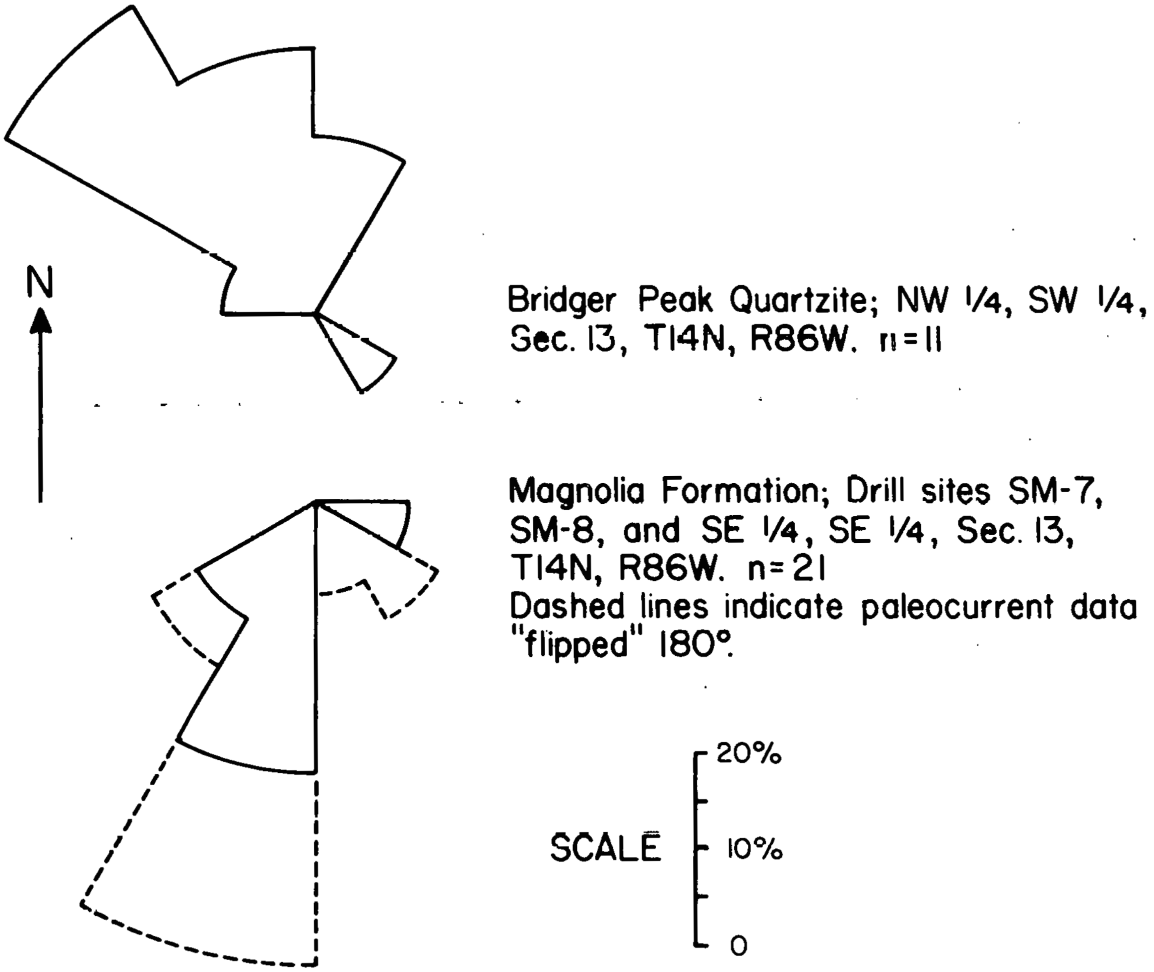
TABLE 3.6. PETROGRAPHY OF THE BRIDGER PEAK QUARTZITE OF THE SIERRA MADRE; MODES FROM VISUAL ESTIMATES

QUARTZITES

Sample no.	Qtz.	Plag.	Musc.	Chlor.	Biot.	Opaq.	Gar.	Carb.
TS-117	99	—	—	1	—	—	—	—
TS-119	99	—	—	1	—	—	—	—
TS-121	64	30	1	5	—	—	—	—
SM9-169	64	—	10	1	5	20	—	—
SM9-163	60	10	25	—	5	—	—	—
SM5-674.6	50	30	15	—	5	—	—	Tr
SM5-708	50	20	20	2	5	—	Tr	3
MEAN (7)	69	13	10	1	3	3	Tr	Tr

MISCELLANEOUS

Sample no.	Qtz.	Plag.	K-spar.	Musc.	Chlor.	Biot.	Epid.	Opaq.	Carb.	Act.	Hnbl.	Rock name
TS-114	15	—	—	—	—	—	55	Tr	—	30	—	Metabasalt
TS-115	5	—	—	—	Tr	—	55	—	—	40	—	Metabasalt
TS-122	28	—	—	—	70	2	—	Tr	—	—	—	Schist
AFMi 1	40	5	15	5	—	—	—	—	35	—	—	Marble
SM5-511.3	10	15	—	—	5	—	2	—	Tr	—	68	Volcaniclastic



Bridger Peak Quartzite; NW 1/4, SW 1/4, Sec. 13, T14N, R86W. n=11

Magnolia Formation; Drill sites SM-7, SM-8, and SE 1/4, SE 1/4, Sec. 13, T14N, R86W. n=21
Dashed lines indicate paleocurrent data "flipped" 180°.

Figure 3.11. Paleocurrent rose diagrams for the Bridger Peak Quartzite and Magnolia Formation.

Mountain containing up to twenty percent pyrite (Sample SM 9-169, Table 3.6), and up to 35 ppm U and 40 ppm Th. The metavolcanic rocks consist primarily of epidote, amphibole, and quartz (Table 3.6).

Little is known about the depositional environment of the Bridger Peak Quartzite. The few sedimentary structures observed were medium- to large-scale planar crossbeds and small-scale trough crossbeds. Paleocurrent analysis is limited and suggests that in one area sediment transport was dominantly northwest (Figure 3.11). This analysis is subject to the same structural difficulties as in the Jack Creek Quartzite. The pyritic quartz-pebble conglomerate at Vulcan Mountain is lithologically similar to the fluvial sediments of the Deep Gulch Conglomerate suggesting that the basal part of the Bridger Peak Quartzite may be partly fluvial in origin. The genesis of the bulk of the unit remains uncertain. The more mature texture and mineral content suggests waning volcanism and tectonic activity during deposition of the quartzite.

LATE ARCHEAN GRANITIC INTRUSIVES

A variety of granitic rocks (mainly granite and granodiorite, Figure 3.2) intrude the Archean strata. Three different types have been identified in the field but petrology shows no major differences between them (Table 3.7). Several different ages are probably represented.

Ferris (1964), Miller (1971), and Hughes (1973) described pink quartz-feldspathic gneisses complexly interwoven with the gray biotite-plagioclase gneisses. These gneisses are not differentiated on Plate 4. Xenoliths of gray gneiss are present in many areas in the pink gneisses. Intrusive and gradational contacts are present between the two gneisses. Similar

TABLE 3.7. PETROGRAPHY OF ARCHEAN GRANITIC INTRUSIVE ROCKS FROM THE SIERRA MADRE. DATA FROM POINT COUNTS; MILLER (1971); FERRIS (1964); HUGHES (1973); AND DIVIS (1976) = *

									after Plag.							An. of
Sample no.	Qtz.	Plag.	K-spar	Musc.	Biot.	Epid.	Hnbl.	Carb.	Seri.	Epid.	Ap.	Zir.	Opaq.	Sph	Chl.	Plag.
SM-53	49.2	37.6	2.8	9.4	—	0.6	—	—	0.6	—	—	—	—	—	—	—
SM-56	26.7	30.4	11.8	1.2	19.9	6.8	0.6	—	2.5	—	—	—	—	—	—	—
SM-99	34.3	40.0	15.6	1.2	2.5	3.8	0.6	—	1.9	—	—	—	—	—	—	—
SM13-341	27.2	32.4	12.1	8.7	13.9	—	—	1.7	4.1	—	—	—	—	—	—	—
AFCR1-1	19.8	45.7	12.3	—	Tr	4.9	13.0	—	1.2	3.1	—	—	—	—	—	—
AFCR7-1	15.0	38.8	10.6	3.1	10.3	5.6	10.6	—	5.6	—	—	—	—	—	—	—
AFSC-1	30.6	28.3	16.0	2.3	15.5	1.0	—	1.4	1.8	3.2	—	—	—	—	—	—
144	24.5	47.9	—	1.2	20.2	5.5	—	—	—	—	—	—	—	—	—	—
QF159*	32.3	51.2	16.5	Tr	—	Tr	—	—	—	—	Tr	—	—	—	—	An8
AG015*	17.3	55.1	27.6	—	—	Tr	—	—	—	—	—	—	—	—	—	An2
AG064*	23.8	49.5	26.7	Tr	—	Tr	—	—	—	—	—	—	—	—	—	An10
AG067*	19.2	52.6	27.4	0.8	—	Tr	—	—	—	—	—	—	—	—	—	An8
AG104*	28.1	44.2	27.7	—	—	—	—	—	—	—	Tr	—	—	—	—	An6
AG106*	28.8	50.0	21.2	Tr	—	Tr	—	—	—	—	Tr	—	—	—	—	An10
AG140*	26.3	43.1	29.5	1.1	—	—	—	—	—	—	Tr	—	—	—	—	—
MEAN (15)	26.9	43.2	17.2	2.0	5.5	1.9	1.7	0.2	1.2	0.3	Tr	—	—	—	—	—
MEAN OF 4 ORTHOGNEISSES FROM MILLER																
	31.2	26.2	11.2	21.5	Tr	1.2	—	—	—	—	Tr	—	Tr	Tr	Tr	An30-32
MEAN OF 4 ORTHOGNEISSES FROM FERRIS																
	30	27	2.5	0.5	2.3	—	—	—	—	—	—	—	—	—	—	An12-34
MEAN OF 4 ORTHOGNEISSES FROM HUGHES																
	29	18	45	5	6	Tr	Tr	—	—	—	Tr	Tr	Tr	—	—	—
MEAN OF 3 FELSIC DIKES FROM MILLER																
	30.3	26.0	21.7	15.0	Tr	3.3	—	—	—	—	—	Tr	Tr	Tr	Tr	An28-31
MEAN OF 3 FELSIC DIKES FROM HUGHES																
	30	25	24	14	Tr	Tr	—	—	—	—	—	Tr	Tr	—	—	—

rocks are also present as felsic (aplitic) dikes. Thes rocks are primarily granites. Their age is not known but they are probably older than the Phantom Lake Metamorphic Suite.

Intrusive into the Phantom Lake Metamorphic Suite is the Spring Lake Granodiorite (Plate 5). The composition of two samples (AFSC-1 and AFCR 7-1) is shown in Tables 3.7 and 3.8. Charlton (personal communication, 1980) reports that much of the unit is a tonalite. This intrusive is nearly 2700 m.y. old (Hedge, personal communication, 1980). The granodiorite body near North Spring Lake clearly cuts across the foliation of the Continental Divide Metavolcanic Rocks and is itself foliated. At the Carrico Ranch area, the granodiorite is folded with the Phantom Lake Metamorphic Suite.

The third granite type in the Sierra Madre is the red or "Baggot Rocks-type" granite. These rocks include granite, granodiorite, quartz monzodiorite, and trondhjemite (Figure 3.2 and Table 3.7). Baggot Rocks Granite in the Medicine Bow Mountains and the northeast Sierra Madre ranges in age from 2325 m.y. to 2560 m.y. (Divis, 1976, Hills and Houston, 1979). Some of the red granites on the north flanks of the Divide Peak Synclinorium appear to be older than the Phantom Lake Metamorphic Suite and, as such, are not related to the true Baggot Rocks Granite. These rocks appear to contain more biotite and less K-feldspar than the younger red granites (Table 3.7). The older granodiorites contain considerably more iron and magnesium than the Baggot Rocks Granite (Table 3.8). Figure 3.3 is an AFM plot of all Archean granitic rocks from the Sierra Madre and Medicine Bow Mountains. Together they indicate a calc-alkaline trend but the disparity in ages may be great enough to negate all the data being plotted together.

TABLE 3.8. GEOCHEMISTRY OF GRANITIC ROCKS FROM THE SIERRA MADRE. ANALYSTS ALLAN DIVIS (1976) 1-6; AND STEVE BOESE, UNIVERSITY OF WYOMING 7-11.

	1	2	3	4	5	6	7	8	9	10	11
	AGO15	AGO67	AG140	AG106	QF152	QF159	SM53	SM54	AFCR1-1	AFCR7-1	AFSC-1
SiO_2	71.71	72.47	73.65	75.15	71.44	75.88	72.6	75.5	62.7	62.6	69.2
Al_2O_3	15.00	15.21	15.37	14.17	15.58	13.88	13.3	11.4	15.5	14.1	12.9
Fe_2O_3	0.37	0.58	0.13	0.73	2.13	0.68	2.85	1.01	4.70	4.33	3.39
MgO	0.05	0.15	0.18	0.10	0.50	0.01	0.73	0.15	1.74	3.74	1.55
MnO	0.001	0.001	0.014	0.013	0.030	0.007	0.03	0.03	0.08	0.08	0.05
CaO	1.14	1.19	0.64	1.13	1.50	1.02	1.57	0.43	5.64	2.82	2.26
Na_2O	5.85	5.26	4.05	5.02	3.65	5.12	3.54	4.78	6.11	4.47	3.72
K_2O	4.78	4.52	4.87	3.28	3.97	2.63	2.61	3.59	1.40	2.86	3.98
TiO_2	0.31	0.05	0.13	0.03	0.32	0.04	0.40	0.0	0.50	0.60	0.3
P_2O_5	0.06	0.04	0.12	0.10	0.23	0.11	0.10	0.06	0.22	0.23	0.13
Co_2	0.15	0.18	0.11	0.04	0.14	0.08	–	–	–	–	–
H_2O	0.50	0.30	0.65	0.22	0.55	0.45	–	–	–	–	–
Total	99.92	99.93	99.91	99.98	100.04	99.91	98	98	98	96	97
MOLECULAR NORMS											
Q	16.77	20.56	27.77	29.21	28.13	31.74	34.85	32.14	7.75	12.14	23.8
Or	28.00	26.54	28.81	19.23	23.75	15.68	16.00	21.95	8.30	17.50	24.3
Ab	52.06	46.92	36.40	44.71	33.18	46.37	32.95	42.30	54.90	41.40	34.45
An	0.54	4.48	1.70	4.66	5.12	3.87	7.45	–	10.73	10.25	6.9
Hy	0.14	0.49	1.42	1.06	3.45	0.76	5.27	1.20	3.81	14.64	6.94
Di	–	–	–	–	–	–	–	1.44	12.72	2.08	3.04
Mt	0.25	0.36	0.12	0.09	0.25	0.08	0.41	–	0.65	0.62	0.48
Il	0.43	0.07	0.18	0.04	0.45	0.06	0.58	–	0.70	0.86	0.44
Ap	0.12	0.08	0.25	0.21	0.49	0.23	0.21	0.13	0.45	0.51	0.29
Ol	–	–	–	–	–	–	–	–	–	–	–
Co	–	0.01	3.07	0.69	4.80	1.00	2.28	–	–	–	–
"Sm"	–	–	–	–	–	–	–	0.48	–	–	–
Ac	–	–	–	–	–	–	–	0.36	–	–	–
C	0.38	0.45	0.28	0.10	0.36	0.20	–	–	–	–	–

1-6: Baggot Rocks – type Granite (see Divis, 1976, p. 49)
7: Red trondhjemite near Jack Creek, NW¼ SE¼, Sec. 32, T16N, R86W
8: Red "granite" near Almeda Creek, NE¼, NE¼, Sec. 28, T16N, R86W
9: Red granite north of Carrico Ranch, SE¼, SW¼, Sec. 1, T15N, R88W
10: Gray (Spring Lake) Granodiorite, crosscuts Silver Lake metavolcanic rocks south of Carrico Ranch; SW¼, SE¼, Sec. 7, T15N, R87W
11: Gray (Spring Lake) Granodiorite in North Spring Lake Cirque; SE¼, SW¼, Sec. 4, T14N, R86W

DEEP LAKE GROUP

The Deep Lake Group in the Sierra Madre (Figure 3.5) consists of the Magnolia Formation, Singer Peak Formation, Cascade Quartzite, and Vagner Formation. This is probably not a complete section as the upper contact is a thrust fault. Graff (1978) placed several thrust faults in the Deep Lake Group but we feel that erosional unconformities best explain the thickness variations (Plate 5).

Magnolia Formation

The Magnolia Formation was defined in the Medicine Bow Mountains (Karlstrom and Houston, 1979a, 1979b) and first described in the Sierra Madre by Graff (1979). In the Sierra Madre, this unit consists of slightly radioactive conglomerates (up to 27 ppm U, 110 ppm Th), quartzites, and phyllites. It ranges in thickness from 460 meters just south of Dexter Peak to zero south of Encampment (Plate 5). This variation in thickness may reflect a combination of several factors: the unevenness of the depositional surface beneath the Magnolia; erosion prior to deposition of the Cascade Quartzite; and Proterozoic faulting.

Several problems exist concerning the Magnolia Formation. As pointed out by Charleton of Resource Associates of Alaska (written communication, 1980) the Magnolia is similar lithologically to Phantom Lake Suite quartzites (compare Tables 3.4 and 3.9 and Figures 3.7 and 3.12) and coarse-grained quartzites, like those of the Magnolia are complexly interleaved with the underlying Bridger Peak Quartzite and Silver Lake Metavolcanic Rocks. What is more, there is no basal conglomerate above an obvious regional-scale unconformity in the Sierra Madre like those mapped in the Medicine Bow Mountains. Nevertheless, we have retained the Magnolia Formation for several reasons. First, we consider the

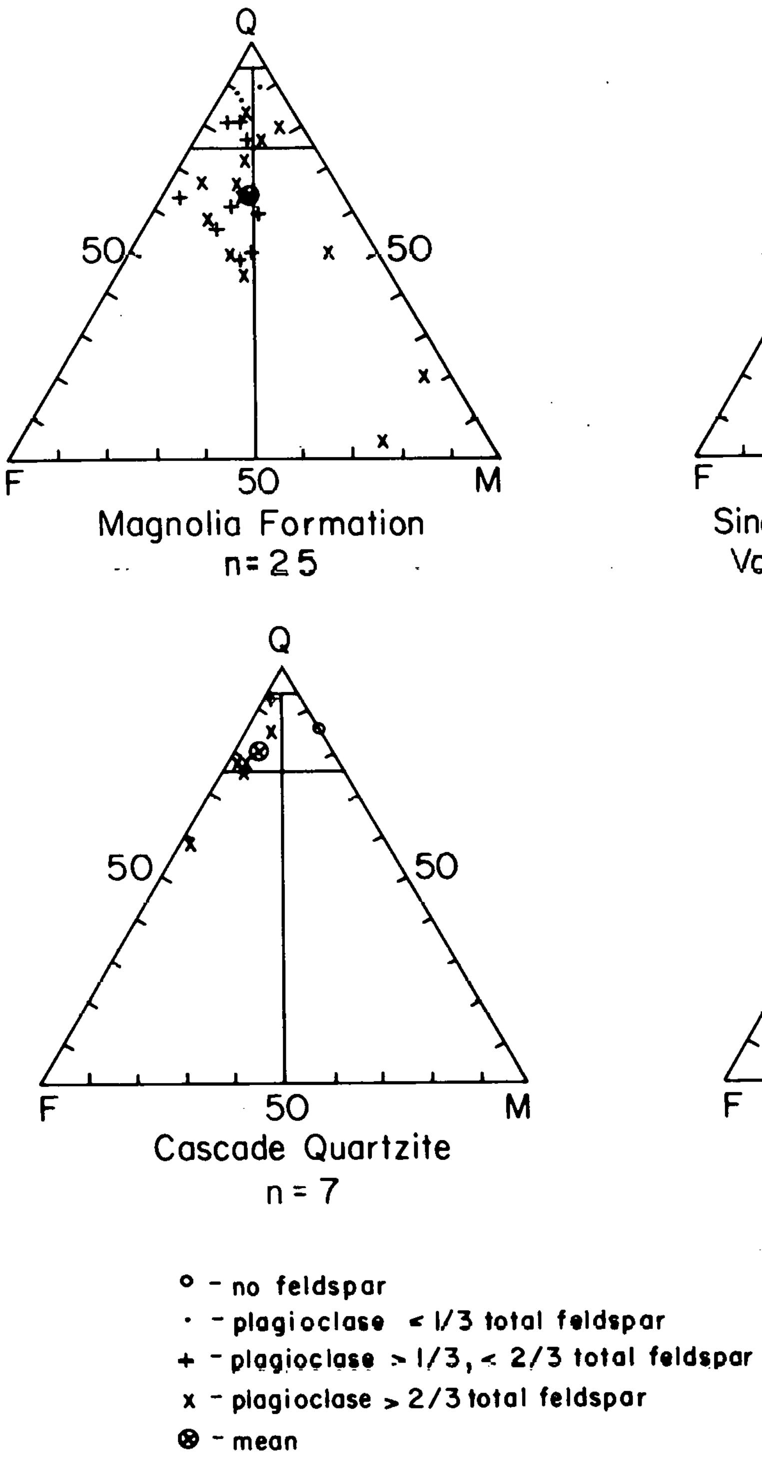


Figure 3.12. Quartz (Q), feldspar (F), and mica+chlorite (M) ternary diagrams for Proterozoic metasedimentary rocks, Sierra Madre, Wyoming (after Folk, 1968)

TABLE 3.9. PETROGRAPHY OF THE MAGNOLIA FORMATION FROM THE SIERRA MADRE; MODES FROM VISUAL ESTIMATES.

Sample no.	Qtz.	Plag.	K-spar.	Musc.	Chlor.	Biot.	Carb.	Epid.	Opaq.	Apat.	Zirc.
TS-100	80	5	9	5	—	1	—	—	Tr	—	—
TS-111	80	5	9	5	—	1	—	—	Tr	—	—
TS-112	75	5	9	10	—	1	—	—	Tr	—	—
TS-126	59	10	10	20	—	1	—	—	—	—	—
TS-179	84	10	—	5	1	Tr	—	—	—	—	—
TS-229	89	—	10	1	—	—	—	—	—	—	—
TS-230	88	—	10	2	—	—	—	—	—	—	—
SM5A-180	48	10	—	35	5	—	Tr	—	2	—	—
SM8-124	65	25	—	3	1	3	—	3	—	—	—
SM8-153	5	20	—	5	65	—	5	—	—	Tr	—
SM8-307.7	20	5	—	74	1	—	—	—	—	—	—
SM8-383.8	80	5	—	10	—	5	—	—	—	—	—
SM8-487.5	46	10	20	20	—	3	1	—	—	—	—
SM8-515.8	60	15	19	3	Tr	1	2	—	—	—	—
SM8-527.5	50	10	15	20	1	4	Tr	—	—	—	—
SM8-581	60	10	15	14	—	1	Tr	—	—	—	—
SM8-623	63	20	—	10	—	1	5	—	1	—	—
SM11-240.5	65	15	5	10	1	3	1	—	—	—	—
SM11-330	56	30	—	2	2	10	—	—	—	—	—
SM11-400	71	15	—	10	2	—	1	—	1	—	Tr
SM11-503.9	54	20	10	10	1	5	—	—	—	—	—
SM12-126	77	10	—	10	—	3	—	—	—	—	—
SM12-156	45	30	—	—	—	25	—	—	—	—	—
SM12-195	50	30	—	18	1	1	—	—	—	—	—
SM5-393	88	—	5	5	1	—	1	—	—	—	—
MEAN (25)	62	13	6	12	3	3	1	Tr	Tr	Tr	Tr

lithologic similarities to reflect derivation of Magnolia detritus from underlying units and we feel that coarse grain sizes and trough crossbeds serve to distinguish the Magnolia from underlying quartzites. Second, interfingering relationships may represent in-folding of Proterozoic and Archean units, with a relationship also observed in the Medicine Bow Mountains. Third, the lack of basal conglomerate above a profound unconformity in the Sierra Madre, we feel, is explained in terms of rapid transgressional deposition. Fluvial deposition in the Magnolia Formation was spatially and temporally rather restricted and was quickly superceded by marine deposition of the Singer Peak Formation. Structural data in the area of drill hole SM-5 (south of Dexter Peak in Plate 5) suggests the presence of an angular unconformity between the Singer Peak Formation and the Phantom Lake Suite and, quartzites separating the two units which we map as Magnolia Formation appear to be folded with the Singer Peak Formation, not the Phantom Lake Suite.

The principal sedimentary structure in the Magnolia Formation is small-scale, trough crossbeds. Paleocurrent analysis (Figure 3.11) indicated a bimodal north-south current distribution, however the predominance of trough crossbeds strongly suggests a unimodal distribution. The bimodal distribution is probably an artifact produced by rotation on a stereonet. In Figure 3.11B we flipped the north trending mode by 180 degrees. We interpret the Magnolia Formation to have been deposited by braided streams in a rift basin (see discussion of the Magnolia Formation of the Medicine Bow Mountains).

Singer Peak Formation

The Singer Peak Formation (Graff, 1979) consists predominantly of phyllites with minor quartzites and a paraconglomerate layer near the top. It interfingers with the Magnolia Formation in the Dexter Peak area but the contact is more obscure elsewhere. The thick phyllite section in the Magnolia at Bottle Creek may be an eastern remnant of the Singer Peak Formation. The Campbell Lake Formation of Graff (1979) is here included with the Singer Peak Formation. The Singer Peak Formation is correlative with the Lindsey Quartzite and the Campbell Lake Formation of the Medicine Bow Mountains. This unit is about 831 meters thick in the west and thins due to erosion, facies change, or faulting to zero in the east.

The phyllites consist primarily of muscovite, or chlorite, and quartz, plagioclase, biotite, garnet, epidote, chloritoid, and andaulsite (Table 3.10). One conglomeratic quartzite layer well within the formation consists almost entirely of quartz and plagioclase (Sample AFSP-1). The mineralogy suggests that these rocks reached epidote-amphibolite facies of metamorphism (Miyashiro, 1973).

No sedimentary structures have been observed in the Singer Peak Formation. We interpret it to be an offshore marine facies of fluvial/glacial units in the Medicine Bow Mountains.

Cascade Quartzite

The Cascade Quartzite was first defined in the Medicine Bow Mountains (Karlstrom and Houston, 1979a, 1979b) and then extended to the Sierra Madre (Graff, 1979). The unit is dominated by quartzite with layers of quartz-pebble and black chert-pebble conglomerate. The quartzite is

TABLE 3.10. PETROGRAPHY OF THE UPPER DEEP LAKE AND LIBBY CREEK GROUPS FROM THE SIERRA MADRE; MODES FROM VISUAL ESTIMATES

SINGER PEAK FORMATION

	Qtz.	Plag.	Musc.	Chlor.	Biot.	Chloritoid	Epid.	Garn.	And.
TS-254	50	15	—	30	—	—	2	3	—
TS-288	3	Tr	91	—	3	3	—	—	—
AFSP-1	73	25	1	1	—	—	—	—	—
AFSP-2	54	5	35	—	5	—	—	—	1
MEAN (4)	45	11	32	8	2	1	Tr	1	Tr

CASCADE QUARTZITE

	Qtz.	Plag.	K-spar	Musc.	Chlor.	Biot.	Carb.	Opaq.
TS-103	79	20	—	1	—	—	—	—
TS-104	77	20	—	3	—	—	—	—
TS-105	77	15	5	3	—	—	—	—
TS-106	85	—	—	15	—	—	—	—
TS-127	59	40	—	1	—	—	—	—
TS-138	83	10	—	—	5	—	2	Tr
TS-162	92	3	3	—	—	2	—	—
MEAN (7)	79	15	1	3	1	Tr	Tr	Tr

VAGNER FORMATION

	Qtz.	Plag.	Musc.	Chlor.	Opaq.
TS-131	43	35	20	1	1

COPPERTON QUARTZITE

	Qtz.	Plag.	Musc.	Kyan.	Opaq.	
TS-261	97	—	2	1	—	
AFCP-2	90	2	7	—	1	"Lookout schist"
AFCP-3	39	1	30	—	30	"Lookout schist"
MEAN (3)	75	1	13	Tr	10	

SLAUGHTERHOUSE FORMATION

	Qtz.	Musc.	Chlor.	Carb.	
TS-128A	96	3	1	—	Quartzite
TS-129	10	—	—	90	Marble
TS-139	99	1	—	—	Metachert
MEAN (3)	68	1	Tr	30	

about 560 meters thick in the center of the map area and thickens to the east and west. Maximum thickness is at least 1500 meters but structural complexities and the lack of marker beds make it difficult to determine the exact thickness. The quartzite was deposited over an erosional unconformity and, in the eastern Sierra Madre, it is the basal unit of the Deep Lake Group.

The Cascade Quartzite is primarily composed of quartz, plagioclase, and muscovite with minor amounts of K-feldspar, chlorite, biotite, and carbonate (Table 3.10). Several thin sections examined show evidence of shearing. The few thin sections examined were primarily subarkoses (Figure 3.12).

Planar and trough crossbedding and graded bedding have been observed in this unit (Graff, 1978) but detailed sedimentologic analysis has not been made. Its similarity to the Cascade Quartzite of the Medicine Bow Range suggests a similar deltaic origin.

Vagner Formation

The Vagner Formation (Karlstrom and Houston, 1979a, 1979b, and Graff 1979) is a heterogeneous unit consisting of quartzites, phyllites, angular-clast paraconglomerates (diamictites), and metacarbonates. This unit is at least 400 meters thick but nowhere is its top exposed. In most localities it is in contact with the Quartzite Peak or Hidden Treasure thrust faults (Plate 6). The one thin section examined is similar to the Singer Peak Formation (Table 3.10 and Figure 3.12). No sedimentary structures have been observed (Graff, 1978) but the unit is very similar to the glaciogenic rocks of the Vagner Formation in the Medicine Bow Mountains.

LIBBY CREEK GROUP

The Libby Creek Group of Houston and others (1968) was modified by Lanthier (1979) and is now subdivided into a lower siliciclastic part and an upper-carbonate-phyllite part (this report, Table 2.1). Contacts in both the Medicine Bow Mountains and Sierra Madre are tectonic: the Lower Libby Creek Group is bounded by thrust faults and the Upper Libby Creek Group has a thrust at its base and is bounded above by shear zones of the Cheyenne Belt. In the Sierra Madre, the Lower Libby Creek Group is represented by the Copperton Formation; the Upper Libby Creek Group by the Slaughterhouse Formation (Figure 3.5).

Divis (1976, 1977) mistakenly identified almost all of the metasedimentary rocks in the Sierra Madre to be Libby Creek Group. Graff correlated the Copperton Formation with the Rock Knoll Formation and the Slaughterhouse with the Headquarters. We correlate the Copperton mainly with the Medicine Peak because both are aluminous quartz arenites and the Slaughterhouse with the Nash Fork Formation because both are dolomite-phyllite units (Figure 1.6).

Copperton Formation

The Copperton Formation is modified from Graff (1979) and is here correlated to the Medicine Peak Quartzite, Lookout Schist, and Sugarloaf Quartzite of the Medicine Bow Mountains (Figure 1.6). Its base is the Quartzite Peak Fault and its top is the Hidden Treasure Fault (Plates 5 and 6). Maximum exposed thickness of the formation is about 1070 meters.

The formation is predominantly a coarse-grained, highly sheared, kyanite-bearing quartzite. It is essentially a quartz arenite (Table 3.10 and figure 3.12). Near the old town site of Copperton there are several hundred meters of ferruginous phyllite that are very similar to the Lookout Schist (Table 2.9). These phyllites are overlain by more quartz arenites. Three small-scale planar crossbeds were observed in fine-grained quartzites near the top of the phyllite section. These crossbeds yielded a biomodal northeast-southwest paleocurrent direction, similar to the Lookout Schist of the Medicine Bow Mountains.

Slaughterhouse Formation

The Slaughterhouse Formation (Divis, 1976, Graff, 1979) is exposed between the Hidden Treasure Fault below and the main shear zone of the Cheyenne Belt above (Plate 6). This unit is composed of metacarbonate, quartzite, phyllite, and metachert (Table 3.10). It is at least 1160 meters thick and is correlated with the Nash Fork Formation of the Medicine Bow Range (Figure 1.6) although none of the stromatolites characteristic of the Nash Fork have been seen in the Slaughterhouse Formation. This unit was probably deposited in a shallow marine environment.

PROTEROZOIC ROCKS SOUTH OF THE CHEYENNE BELT

A complex assemblage of metasedimentary and metavolcanic rocks (amphibolites through felsic tuffs) crop-out south of the shear zone (Lackey, 1965). Neither the thickness nor the stratigraphy of these rocks is known. These rocks were deformed approximately 1700 m.y. and presumably were deposited during the Early Proterozoic (Hills and Houston, 1979). Intrusive into these rocks is the Encampment River Granodiorite and the Sierra Madre

Granite (Lackey, 1965, Divis, 1976), both of which are late Early Proterozoic (Hills and Houston, 1979).

GABBROIC INTRUSIVE ROCKS

As shown in Plate 5, all Precambrian rocks in the Sierra Madre are crosscut by mafic sills and dikes of basaltic composition. As in the Medicine Bow Mountains, we interpret these bodies to represent several episodes of intrusion of mafic magma. Divis (1976), on the basis of petrology and field relations, subdivided the mafic intrusives into three types: garnet amphibolites, ophitic amphibolites, and syn-granitic diabases. Divis (1976) then suggested that the garnet amphibolites are Archean, the ophitic amphibolites are Early Proterozoic, and the syn-granitic diabases are slightly younger than the Sierra Madre Granite (1700 m.y.).

This interpretation fits fairly well our observations on field relationships and geochemistry of the gabbroic intrusives. Some intrusives (Divis' syn-granitic diabases) crosscut the shear zone and 1700 m.y. rocks south of the shear zone are definitely latest Early Proterozoic. Table 3.11 shows mean values of a large number of whole-rock analyses of gabbroic intrusives in rocks north of the shear zone. These intrusives are plotted on alkali versus silica and AFM diagrams in Figure 3.13.

All of the rocks are distinctly tholeiitic and three possible subgroups can be distinguished. The large sill cutting gneissic basement (number 1), and the sill cutting the Continental Divide Metavolcanic Rocks (number 3) are ultramafic, characterized by low alkali contents. These rocks may be related to Divis' garnet amphibolites, and may be Archean. Intrusives in the Phantom Lake Suite and Deep Lake Group,

TABLE 3.11. GEOCHEMISTRY OF MAFIC INTRUSIVE ROCKS FROM THE SIERRA MADRE, REPORTED AS MEANS. DATA FROM HOUSTON AND OTHERS, 1977. TOTAL Fe REPORTED AS FE_2O_3. FE_2O_3/FEO ASSUMED TO BE 0.15 FOR NORMATIVE CALCULATIONS (COX AND OTHERS, 1979).

	1	2	3	4	5	6	7	8
SiO_2	50.65	51.17	50.59	52.36	54.76	50.49	50.42	50.17
Al_2O_3	7.53	13.14	6.28	13.30	12.99	12.64	11.64	14.08
Fe_2O_3	12.07	12.06	12.51	11.67	10.65	11.86	12.11	12.88
MgO	15	6	12	9	6	9	8	7
CaO	7.41	10.24	6.41	9.82	9.89	11.71	11.33	9.06
Na_2O	0.69	1.75	0.46	2.36	1.53	1.58	1.60	3.30
K_2O	0.11	0.38	0.28	0.56	0.44	0.22	0.54	0.64
TiO_2	0.43	0.79	0.40	1.00	0.81	0.86	0.95	1.37
P_2O_5	—	—	—	—	—	—	—	—
MnO	0.28	0.19	0.26	0.17	0.19	0.21	0.20	0.17
Total	93.6	95.4	89.2	100.2	97.0	98.25	96.4	98.7
				MOLECULAR NORMS				
Q	3.94	7.21	11.08	0.03	11.65	1.04	2.53	0.00
OR	0.70	2.30	1.90	3.30	2.75	1.35	3.35	3.85
Ab	6.55	16.20	4.70	21.20	14.40	14.50	15.00	30.15
An	18.10	27.73	16.17	24.03	28.63	27.38	24.35	22.10
Hy	51.67	19.27	47.55	28.62	21.57	27.16	24.32	10.60
Di	16.68	19.84	16.03	19.84	18.40	25.76	27.72	19.00
Ol	—	—	—	—	—	—	—	10.59
Cor	—	—	—	—	—	—	—	—
Mag	1.73	6.89	1.94	1.59	1.47	1.47	1.71	1.79
Il	0.64	0.58	0.64	1.40	1.18	1.16	1.38	1.94

1 mean of 20 samples from large gabbroic to pyroxenitic sill (3A-3) cutting Archean gneisses in NE¼, SE¼, Sec. 7, T14N, R84W.

2 mean of 11 samples from small gabbroic sill (3A-2) cutting Archean gneisses in SE¼ Sec. 7, T14N, R84W.

3 mean of 8 samples from large amphibolite sill (3G) cutting rocks of Continental Divide Metavolcanic Rocks in NW¼, NW¼, Sec. 10, T14N, R86W.

4 mean of 47 samples from small gabbroic sill (3A-1) cutting the Bridger Peak Quartzite of the Phantom Lake Suite in NW¼ Sec. 18, T14N, R84W.

5 mean of 8 samples from small gabbroic to pyroxenitic sill (3D) cutting Bridger Peak Quartzite of the Phantom Lake Suite in NW¼, SW¼, Sec. 14, T14N, R86W.

6 mean of 33 samples of large gabbroic to pyroxenitic sill (3F) cutting Silver Lake Metavolcanic Rocks of the Phantom Lake Suite in SE¼, SW¼, Sec. 13, T14N, R86W.

7 mean of 12 samples of small gabbroic sill (3B) cutting Vagner Formation of Deep Lake Group in SW¼, NE¼, Sec. 30, T14N, R86W.

8 mean of 16 samples of small gabbroic sill (3E) cutting Slaughterhouse Formation of Libby Creek Group in NW¼, Sec. 26, T14N, R86W.

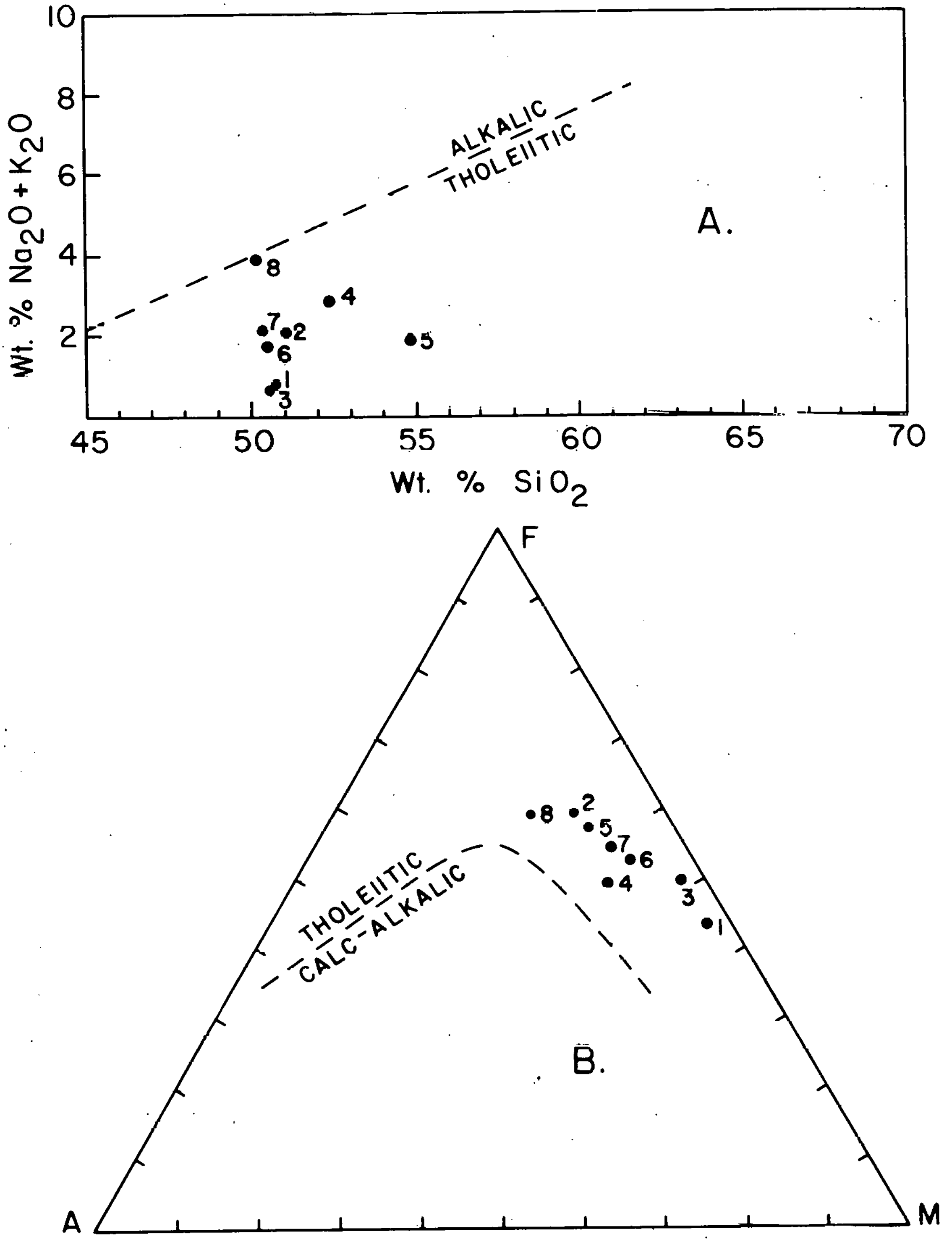


Figure 3.13. Geochemical plots of gabbroic rocks in the Sierra Madre; numbers referenced to Table 3.11. **A.** Alkali-silica plot (dashed line from MacDonald, 1968); **B.** A ($Na_2O + K_2O$), F ($Fe_2O_3 + FeO$), M (MgO) diagram (dashed line from Irvine and Baragar, 1971).

and small sills cutting gneissic basement (numbers 2, 4, 5, 6, 7 in Table 3.11) show a strong similarity on both plots and are similar in composition to presumed rift-related Early Proterozoic intrusives in the Medicine Bow Mountains.

Sample 8, the mean of 16 samples from small sills cutting the Slaughterhouse Formation of the Libby Creek Group, is apparently unrelated to any of the other samples. It has the highest alkali content, the lowest silica content, and contains normative olivine. It is chemically similar to dikes which cut the Libby Creek Group in the Medicine Bow Mountains which are presumed to be related to late-stages of rifting.

GAPS TRONDHJEMITE

Intrusive into the Copperton Formation near the defunct town of Copperton (Plate 5) is a red, ferruginous, sheared trondhjemite which appears to be gradationally related to an amphibolitic dike. This rock is similar to trondhjemites found near the Gap in the Medicine Bow Mountains. In addition, several small bodies of aplite are present in a large metagabbro body crosscutting the Cascade Quartzite in the eastern part of Plate 5 which are also similar to the Gaps Trondhjemite (Graff, personal communication, 1980). The petrography of these rocks is shown in Figure 2.20. We consider these rocks in the Sierra Madre, like the Gaps Trondhjemites in the Medicine Bow Mountains to be differentiates of the gabbroic dikes in the Libby Creek Group.

TECTONIC HISTORY OF THE SIERRA MADRE

This section summarizes the structural features of the metasedimentary rocks of the Sierra Madre, compares them to those of the Medicine Bow Mountains, and then tries to relate them to a plate tectonic model for southeastern Wyoming. As in the Medicine Bow Mountains, Precambrian rocks of the Sierra Madre can be divided into three major structural terranes: a complexly deformed Archean terrane, concentrically folded rocks of the Deep Lake Group, and a zone of vertical bedding foliation and fold axes which affects rocks of several ages and is related to the Cheyenne shear zone.

Plate 6 is a tectonic map of the Sierra Madre which summarizes the major folds and faults and displays lower hemisphere, equal area projections of fabric elements: bedding, foliation, and lineations.

GEOMETRY OF FOLDS IN ARCHEAN ROCKS

For this discussion, Archean rocks are divided into two groups. One consists of the Phantom Lake Metamorphic Suite and the other is a basement terrane, referred to as the northern complex in Plate 6, containing quartzo-feldspathic gneisses, interleaved supracrustal rocks, (including the Continental Divide Metavolcanic Rocks), and granitic intrusives.

The oldest pervasive structural feature in the Sierra Madre is a gneissic foliation (S_1) in the northern complex rocks and a corresponding penetrative axial plane schistosity in the Phantom Lake Metamorphic Suite of the central Sierra Madre. Both units have similar fold styles (compare subarea 7 with 11 and subarea 4 with 14) and they appear to have been strongly deformed together so that, in most areas, we cannot prove an age difference based on structural relationships. However, a few areas retain indications that the northern complex is a basement which has undergone a longer and more complex structural history than the Phantom Lake Suite. In the Deep Gulch area, Sec. 36, T. 16 N, R. 88 W, paraconglomerates interbedded with "basement" gneisses contain folded and highly stretched clasts whereas overlying conglomerates of the Deep Gulch Conglomerate are relatively undeformed. On ridge number one, Carrico Ranch, Sec. 12, T. 15 N, R. 88 W, an angular discordance exists between the strike of foliation in the gneisses and the basal Phantom Lake Suite contact. The northern complex also contains interleaved gneisses and remnants of supracrustal rocks which suggest a longer or more intense deformational history. The details of the earlier deformations which may have produced these early features in the northern complex are obscure due to the intensity of the folding (F_1 of this report) which produced the penetrative S_1 foliation.

S_1 foliation in the northern complex is a gneissosity that strikes east-west in the central Sierra Madre and dips both north and south at shallow angles. Subarea 1 has gently north-dipping foliation folded into broad north-trending (F_3) folds which are reminiscent of similar north-trending Archean structures found in subarea 26 of the western Medicine Bow Mountains (Plate 3) some 30 km to the east. Subareas 3 and 4 contain maxima of poles to foliation which define gently south-to southwest-dipping planes. Subareas 5, 6, and 21 have foliations which generally dip gently north and, in subareas 7 and 8, foliation in the northern complex dips at steeper angles (∿ 45°) northwest, probably reflecting later (F_2) deformations.

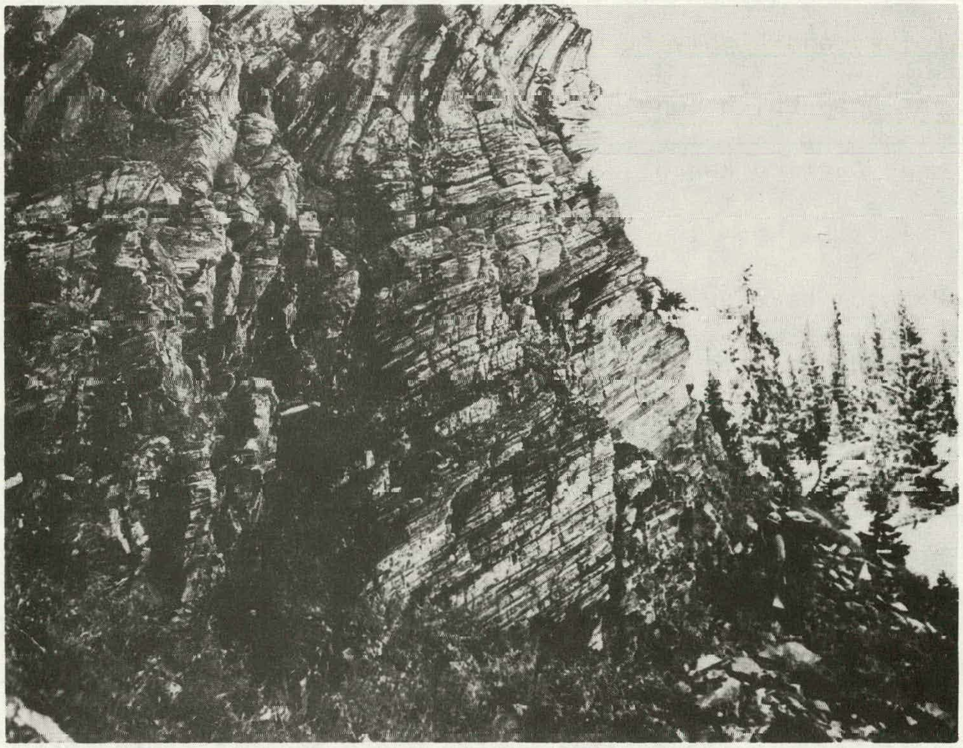


Figure 3.14 Photograph showing hinge of mesoscopic F_1 fold in the South Spring Creek Lake area of the Central Sierra Madre, Sec. 11, T.14 N., R.86W., looking west.

Similar geometries of planar structures are seen in the Phantom Lake Suite. In subarea 14, bedding is parallel to S_1 schistosity and both dip gently (∿ 30°) south. In this subarea and immediately to the north, in rocks of the Continental Divide Metavolcanic Rocks of subarea 4, we have documented mesoscopic and macroscopic recumbent isoclinal folds, or nappes, in the metasedimentary rocks (Figure 3.14). Because of similar geometries between subareas 14 and 4, these nappes are also believed to be present in rocks of the northern complex in the central Sierra Madre and we interpreted the subhorizontal S_1 gneissosity in the northern complex to be related to nappe-like folding.

The nappe system in the Phantom Lake Suite in the central Sierra Madre is named the Rudefeha Synclinorium (Plate 6) because: we believe the Phantom Lake Suite to be a synclinal supracrustal succession which lies on an older northern complex "basement"; our present stratigraphic interpretation (Plate 5) within the Phantom Lake Suite suggests a general synclinal configuration; and top and bottom criteria in the metasedimentary rocks are generally consistent with such a structure. However, there must be many more megascropic folds in the central area than we have mapped and shown in Plates 5 and 6.

Mapping in the west and northwest Sierra Madre reveal equally complex isoclinal fold systems. Plates 5 and 6 show that these structures can be viewed as a westward extension of the Rudefeha Synclinorium and that the overall structural style of the Phantom Lake Suite is that of a warped synclinorium that trends east-west in central Sierra Madre, north-south in the western Sierra Madre, and northeast-southwest in the northwest Sierra Madre (Figure 3.15). In subarea 13 (Plate 6), S_o and S_1 dip gently west and except for this change in trend have the same geometric pattern (a simple maxima) seen in subarea 14.

The Rudefeha Synclinorium and four or more intralimb folds east of the trace of the main axial plane of the synclinorium appear to be truncated to the north against the Vulcan Mountain Fault (Plate 6). A similar synclinorium, the Divide Peak Synclinorium (Figure 3.15) appears to the northwest and our stratigraphic interpretation (Plate 5) suggests that this is a northwestern extension of the warped Rudefeha Synclinorium. Bedding and S_1 foliation generally strike northeast in subareas 9, 10, 11, and 12 although there is also re-distribution of poles about later west- and northwest- plunging fold axes. As mentioned earlier, the similarity in fold geometries between subareas 5 and 10 and between subareas 7 and 11 and 8 and 12 indicate that the Phantom Lake Metamorphic Suite and the northern complex gneisses were deformed together into the (F_1) Divide Peak Synclinorium.

F_1 folding appears to have been synkinematic with intrusion of the Spring Lake Granodiorite which crosscuts an F_1 fold in the area of North Spring Creek Lake but also possesses an F_1 foliation in this area and is involved in F_2 folds in the Carrico Ranch area of the northwestern Sierra Madre. This suggests that F_1 folding took place about 2700 m.y. ago.

Subsequent deformations in the northwestern Sierra Madre formed northwest (F_2) and northeast (F_3) intralimb folds. F_2 is the prominant

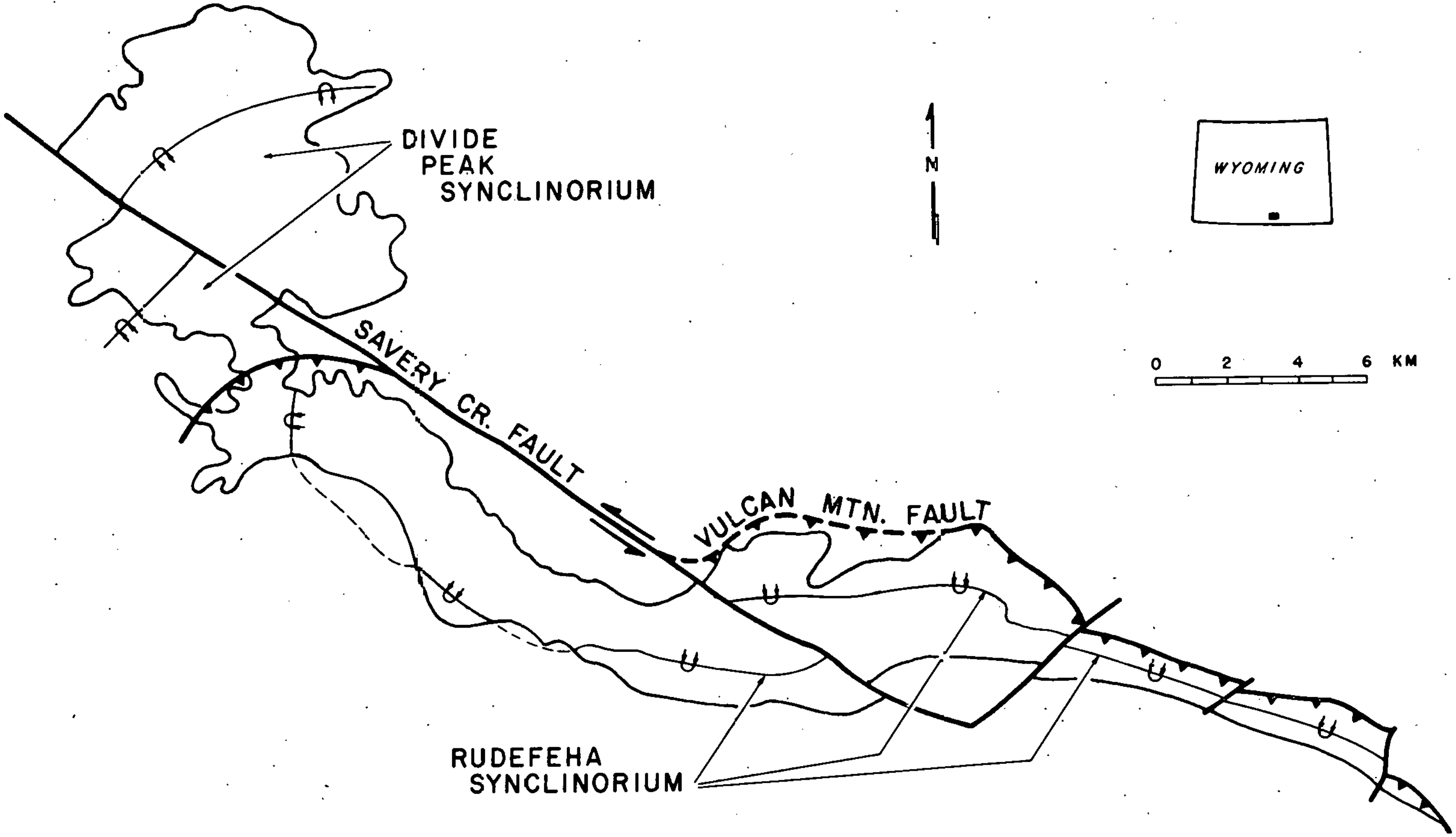


Figure 3.15. Generalized structural interpretation of the Phantom Lake Metamorphic Suite as a warped synclinorium.

mesoscopic fold structure seen in the northwest Sierra Madre, as shown by lineations in subareas 7, 8, 9, 11, 12. This F_2 folding is also responsible for the warping of the early F_1 synclinorium in the Phantom Lake Suite and the gneissosity in the northern complex. Evidence for this interpretation is that maxima of poles to S_o and S_1 from subareas 3-7, 9, 10, 13-15, fall on a girdle defining a west- to northwest-plunging (F_2) fold axis (Figure 3.16). F_2 also caused redistribution of poles to S_o and S_1 surfaces in many of the subareas of the northern complex.

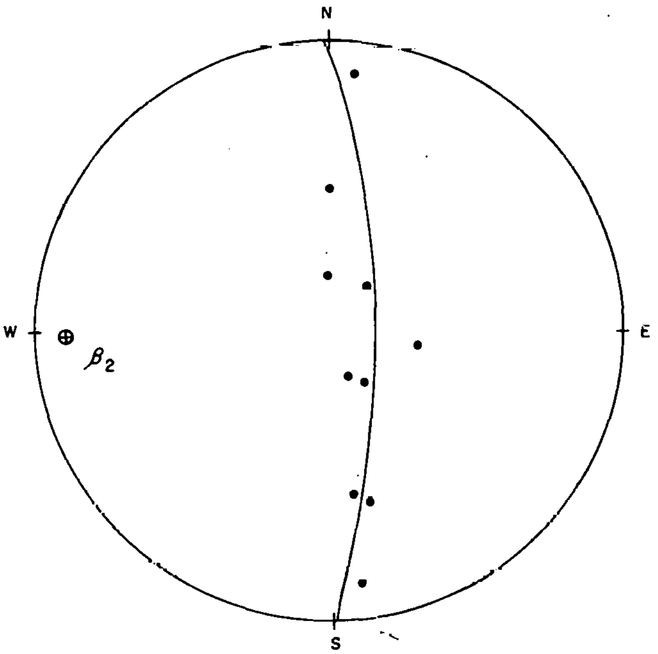


Figure 3.16. Synoptic diagram of centers of maxima of poles to surfaces from subareas 3-7, 9, 10, 13-15 of Plate 6. Maxima fall on a girdle defining a west plunging F_2 fold axis.

F_3 is best developed in subareas 11 and 12 of the northwest Sierra Madre. These intralimb folds plunge moderately (∿ 20°) north-northeast and can be seen near Deep Gulch, Sec. 36, T.16N., R.88W., (Figure 3.17).

Figure 3.17. F_3 chevron folds near Deep Gulch. These folds are on the overturned limb of an F_1 fold.

GEOMETRY OF FOLDS IN PROTEROZOIC ROCKS

Seven anticlines and synclines are present in the Proterozoic Deep Lake Group of the southwestern Sierra Madre (subareas 16 and 17 of Plate 6; and Spencer, 1904, p. 25). We refer to these folds as F_4 although they could be contemporaneous with F_2 folds. F_4 folds plunge about 20-30 degrees west-southwest and are generally open concentric folds, with interlimb angles of less than 90 degrees. These folds become overturned as one approaches the Cheyenne shear zone. F_4 folds are intruded by gabbroic sills and phacoliths and this, combined with their concentric style and northeast trend makes them appear to be analogous to F_2 folds in the Deep Lake Group of the Medicine Bow Mountains (Table 3.12).

A major structural dichotomy separates the Deep Lake Group from the Phantom Lake Metamorphic Suite. This is particularly evident in the area of Dexter Peak (compare subareas 13 and 16, Plate 6) and is some of our best evidence for an unconformity between the Magnolia Formation of the Deep Lake Group and the Phantom Lake Metamorphic Suite. This dichotomy is less obvious to the east due to overprinting of younger fold systems.

The youngest structural pattern (here called F_5) is vertical bedding and foliation and subvertical fold axes seen adjacent to the Cheyenne shear zone. This pattern affects rocks of several ages: Archean, northern complex rocks in subarea 2, Phantom Lake Suite rocks in subarea 15, Deep Lake Group rocks in subarea 18, Libby Creek Group rocks in subarea 19, and rocks south of the shear zone in subarea 20. These subvertical structural patterns are similar to the F_3/F_5 structures seen in the Libby Creek Group/

Deep Lake Group rocks close to the shear zone and in rocks south of the shear zone in the Medicine Bow Mountains (Table 3.12).

TABLE 3.12. COMPARISON OF TECTONIC EVENTS BETWEEN THE SIERRA MADRE AND MEDICINE BOW MOUNTAINS.

MEDICINE BOW MOUNTAINS	SIERRA MADRE
	F_1 — formation of nappes in Phantom Lake Metamorphic Suite and northern complex gneisses — and synkinematic intrusion of Spring Lake Granodiorite $\sim$2700 m.y.
F_1 — isoclinal folding of Phantom Lake Suite, possible contemporaneous with intrusion of Baggot Rocks — type granite	F_2 — warping of Phantom Lake Suite synclinorium and development of west — plunging structures, possibly contemporaneous with intrusion of Baggot Rocks — type granite $\sim$2500 m.y.
	F_3 — north-northeast — trending intralimb chevron folds in Phantom Lake Suite
F_2 — northeast — southwest concentric folding of Deep Lake Group, followed by intrusion of gabbroic sills	F_4 — northeast — southwest concentric folding of Deep Lake Group, followed by intrusion of gabbroic sills
F_3 — west — plunging folds in Deep Lake Group	
F_1 — thrusting in Libby Creek Group, rotation to vertical attitudes, development of vertical foliations in rocks south of shear zone	F_5 — thrusting, rotation of thrusts to vertical attitudes, development of vertical foliation in rocks south of shear zone, major metamorphism
F_4 /F_2 — northwest warping in Deep Lake Group/Libby Creek Group	
F_5 /F_3 — vertical rotations associated with movement on shear zone	

FAULTS

There are a variety of faults representing different styles of movement and ages of deformation in the Sierra Madre. Only the most important or prominent faults are shown on Plate 6. Many more are undoubtedly present but are obscured by younger rocks.

One of the more prominent faults (although completely covered by Cenozoic sediments, Plate 5) is the Savery Creek Fault. The fault trends northwest-southeast from the northwest boundary of the range to Cow Creek where it is offset by an Early Proterozoic normal fault. The Savery Creek Fault may continue southeast but it is overlain by rocks in the hanging wall of the Quartzite Peak Fault. The Savery Creek Fault shows approximately 2600 meters of left-lateral movement of the Archean rocks of the Divide Peak Synclinorium; but Deep Lake Group strata are displaced only a few hundred meters. The Rudefeha Synclinorium as well as other F_1 structures are also displaced, placing the major movement of the fault between 2700 m.y. and deposition of the Deep Lake Group sometime after 2500 m.y. Post-Deep Lake Group left-lateral movement is interpreted to be a post-Archean re-activation along the earlier fault trace. Vine and Prichard (1959) also reported substantial Tertiary movement and a few meters of late Cenozoic, vertical movement along this fault at the western edge of the study area (Plate 6).

The Vulcan Mountain Fault is a large reverse or thrust fault at the base of the Phantom Lake Metamorphic Suite in the eastern part of the range. This fault merges with the Savery Creek Fault for part of its length. In the Dexter Peak area the Vulcan Mountain Fault separates the Divide Peak Synclinorium from the rest of the Phantom Lake rocks. The Vulcan Mountain Fault is intruded by metagabbro and offset by several normal faults. One of these intrusive bodies forms a large sill which is offset by the Savery Creek Fault, suggesting that the Vulcan Mountain Fault is probably older than the Savery Creek Fault. A maximum age for the Vulcan Mountain Fault can be determined from the youngest rocks it displaces -- the 2700 m.y. old Spring Lake Granodiorite.

Several normal faults have been mapped in the range (Plate 6). Three of those faults cut the Deep Lake Group rocks in the southwest part of Plate 6, terminate against the Quartzite Peak Fault and have been intruded by mafic rocks. Hence, these faults are probably Early Proterozoic in age. Three others cut the Libby Creek Group and two of them also displace the Sierra Madre Granite. These faults are Middle Proterozoic or younger. One large Archean normal fault has been inferred on the northernmost edge of the Divide Peak Synclinorium, where part of the section is repeated. It has been intruded by a large mafic sill similar to the large Archean sills which have been offset by the Savery Creek Fault. Another Archean (?) normal fault is on the east edge of the Spring Lake Granodiorite intrusion in subarea 12.

Two large faults, the Quartzite Peak Fault and the Hidden Treasure Fault (Plate 6) lie at the base of the Copperton and Slaughterhouse Formations respectively. These faults were mapped as major reverse faults by Graff (1978). We agree, and by analogy to similar faults in the Medicine Bow Mountains, consider them to be rotated, originally subhorizontal, thrust faults. Evidence for this interpretation is that the faults are generally parallel to bedding and that they attenuate the stratigraphic succession. More than two kilometers of the Libby Creek Group and up to 1 kilometer of the Deep Lake Group are missing in the eastern Sierra Madre due to movement along these faults. The major shear zone in the Sierra Madre, the Cheyenne shear zone, (Plate 6) appears to be the western extension of the Mullen Creek Nash Fork shear zone of the Medicine Bow Mountains. Major geologic and geochronologic discontinuity across this shear zone here, as in the Medicine Bow Mountains, suggests large movements across this fault zone.

SUGGESTED SEQUENCE OF EVENTS

1. Deposition before 2700 m.y. ago of protoliths of biotite-plagioclase gneiss (may include Archean metasedimentary and metavolcanic rocks undivided and the Continental Divide Metavolcanic Rocks).
2. Deformation of biotite-plagioclase gneiss and intrusion of quartzo-feldspathic gneisses as well as some older red granites, age perhaps 2900 m.y.
3. Deposition of the Phantom Lake Metamorphic Suite, age > 2700 m.y.
4. F_1 deformation and intrusion of the Spring Lake Granodiorite, age 2700 m.y.
5. Movement on Vulcan Mountain Fault, age 2500-2700 m.y.
6. F_2 and F_3 deformation and intrusion of Baggot Rock-type granite, age 2500 m.y.
7. Intrusion of large mafic sills, age 2500 m.y.
8. Major movement on Savery Creek Fault, age 2500 m.y.
9. Deposition of Deep Lake Group, age 2000-2500 m.y.
10. F_4 folding, normal faulting, and intrusion of mafic rocks, age 2000 m.y. to 2500 m.y.
11. Deposition of Lower Libby Creek Group, age 2000 to 2500 m.y.
12. Intrusion of Gaps Trondhjemite and related mafic rocks, age 2000 m.y.
13. Deposition of Upper Libby Creek Group, age 1700 to 2000 m.y.
14. Thrust faulting in Libby Creek Group and in rocks south of shear zone, age 1760 to 2000 m.y.
15. Last major movement on the shear zone, age 1640 to 1735 m.y.
16. Intrusion of Sierra Madre Granite, age 1000 to 1700 m.y.

COMPARISON OF TECTONIC EVENTS IN THE SIERRA MADRE AND MEDICINE BOW MOUNTAINS

Table 3.12 summarizes our correlation of tectonic events between the Sierra Madre and Medicine Bow Mountains. Archean events are better documented in the Sierra Madre and because of more complete preservation of the Libby Creek Group, Proterozoic tectonic events are better documented in the Medicine Bow Mountains.

The most striking parallels between the two ranges are fold styles in Proterozoic rocks: 1) the similar NE-SW concentric folds in the Deep Lake Group, which in both ranges, are crosscut by gabbroic intrusives; 2) the presence of major reverse faults (thrusts) at the base of the lower and upper Libby Creek Groups in both ranges; and 3) the fact that Libby Creek Group rocks, rocks south of the shear zone, and all older rocks adjacent to the shear zones exhibit subvertical bedding and foliation trends which strike parallel to the shear zones. These similarities indicate similar structural histories for Proterozoic metasedimentary rocks in both ranges.

TECTONIC MODEL FOR DEFORMATION OF METASEDIMENTARY ROCKS IN THE SIERRA MADRE

The similarities in structural styles between Proterozoic metasedimentary rocks in the Sierra Madre and Medicine Bow Mountains strengthens our interpretation that Proterozoic deformations in southern Wyoming were related to regional stresses generated by plate margin tectonics: rifting followed by collisional orogenesis about 1700 m.y. ago. This model is discussed in detail in earlier sections on the Medicine Bow

Mountains (see Figure 2.22 and 2.27) and need not be reiterated here except to note some important differences in tectonic setting between the two ranges.

Trend of the Cheyenne shear zone changes from northeast, in the Medicine Bow Mountains, to west-northwest in the Sierra Madre, a bend of some 120-140 degrees. This change in strike combined with sedimentary evidence for miogeoclinal (passive plate margin) deposition and the presence of presumably rift-related tholeiitic sills cutting the metasediments, suggests to us that the present trend of the shear zone is the approximate original shape of the rift-formed margin of the southern Wyoming province. This shape is reminiscent of triple junctions so we postulate a third, and failed arm of the rift trending south into the hypothetical rift-block (Figure 2.22). This interpretation helps explain the preservation of Proterozoic metasedimentary rocks. The Deep Lake and Libby Creek Group are preserved in the successful arms of the rift but are absent near the triple junction, which would have been a topographic high during rifting (Burke, 1980) and consequently would not have accumulated as much sediment as the grabens along the rift arms. Also, the area of the triple junction forms a south-protruding salient to the southern Wyoming Province and it seems reasonable to postulate that deformation would have been most intense here during impingement of the island arc(s) about 1700 m.y. ago and less intense in the more protected rift basins to the northeast and to the west.

Structural data also support the hypothesis that the area between the two ranges was a salient which experienced more intense deformation then the adjacent areas. The Deep Lake Group in the Sierra Madre is thickest and best preserved in the extreme western part of the range and it is only in this area that the Deep Lake Group rocks retain open, concentric folds similar to folds in the Medicine Bow Mountains which we believe to have formed during rifting. However, in the central and eastern areas of the Sierra Madre, Deep Lake Group rocks are tectonically thinned and overprinted by vertical bedding and foliation patterns which appear to be related to collisional orogenesis in the Medicine Bow Mountains. Similarly, Libby Creek Group rocks are thickest in the western Sierra Madre and disappear to the east as one approaches the triple junction. In addition, Archean gneissic rocks in the area of the triple junction (salient) in both the Sierra Madre and Medicine Bow Mountains exhibit structural styles (Plates 3 and 6) which are identical to those found in the Libby Creek Group and in rocks south of the major shear zones. This is strong evidence that the Archean block between the two ranges was deformed plastically in response to the same orogenic stresses which formed vertical foliations in rocks south of the shear zone, and rotated Libby Creek Group rocks, and the inferred subhorizontal thrust faults, to vertical positions. All of these structures are compatible with the hypothesis of collisional orogenesis about 1700 m.y. ago.

DETAILED DESCRIPTION OF URANIUM-BEARING ROCKS IN THE SIERRA MADRE DEEP GULCH CONGLOMERATE

STRATIGRAPHY OF THE DEEP GULCH CONGLOMERATE

The Deep Gulch Conglomerate is an informal lithologic unit within the lower Jack Creek Quartzite. The unit is named for the exposures in Deep Gulch located in Sec. 36, T. 16 N., R. 88 W. and Sec. 31, T. 16 N., R. 87 W. The best exposures are in the NE¼ Sec. 12, T. 15 N., R. 88 W. (Ridge 1, Carrico Ranch). This section discusses the stratigraphy and lithofacies variation of the uraniferous Deep Gulch Conglomerate and other informal units of the lower Jack Creek Quartzite.

The Deep Gulch Conglomerate is defined to include the radioactive metasedimentary rocks which directly overlie either Archean gneissic "basement" rocks or non-radioactive biotite quartzite in the northwestern Sierra Madre. The upper contact of the Deep Gulch Conglomerate is the uppermost quartz-pebble conglomerate directly overlying the highly radioactive conglomerates within Unit 3 (Figure 3.23); this conglomerate may or may not be radioactive. The lower contact is believed to be: 1) an angular unconformity between the radioactive metasedimentary rocks and biotite plagioclase gneiss, or 2) a disconformity between the radioactive metasedimentary rocks and slightly older medium-grained biotitic quartzites.

In the Carrico Ranch area this unconformity between the Deep Gulch Conglomerate and gneissic "basement" is intruded by metagabbroic sills and is not directly observable. However, foliation in the underlying gneiss strikes N 25 E whereas bedding in the radioactive metasediments strike N 60 E (Figure 3.18). At Deep Gulch the contact relationships are well exposed but still enigmatic. Here, gneissic rocks grade "up" into biotitic quartzites which appear to be in sharp contact with muscovite-bearing, radioactive, subarkose and arkose of the Deep Gulch Conglomerate. The lower contact of the Deep Gulch Conglomerate is mapped at this contact between biotitic and muscovitic quartzites. Farther northeast, in the Manning Ranch area the lower contact is not exposed.

As shown in the index map of Plate 7, the Deep Gulch Conglomerate crops out in the northwest Sierra Madre in an arcuate trend (about 13 km long) from Carrico Ranch to Joes Park. Similar conglomerates crop out to the south, on Dexter Peak, and to the east, on Divide Peak, but these conglomerates are not discussed here.

"Basement" Gneiss and Metasediments

The "basement" in the northwestern Sierra Madre is composed of paragneiss, amphibolites and paraconglomerates (Graff, 1978; Gwinner, 1979; Divis, 1976). The term basement is used here to denote the rocks which we believe unconformably underlie the Phantom Lake Metamorphic Suite. The paragneiss is composed of quartz, biotite and plagioclase and may represent metamorphosed volcaniclastic sediments (Divis, 1977). This lithology constitutes the major rock type of the basement.

Amphibolites are medium-grained amphibole schists which are laterally discontinuous in outcrop and may represent either metavolcanics or metamorphosed mafic sills. Paraconglomerates are also laterally discontinuous and consist of deformed (stretched and folded) granitic cobbles and pebbles in a quartz-rich amphibolitic matrix.

Biotite Quartzite

The oldest metasedimentary unit above the "basement" is a lenticular biotite quartzite unit which underlies the Deep Gulch Conglomerate in three areas: drill holes SM-2D, JP-3, and the Deep Gulch area.

EXPLANATION

m — **MAFIC IGNEOUS ROCKS:** Mafic sills and dikes ranging from massive gabbro and diabase to foliated amphibolite.

INTRUSIVE CONTACT

4 / 4a — **UNIT 4:** Arkose to subarkose, pink, medium- to coarse-grained; planar-crossbedded, massive, well-sorted. Underlain by massive, pink, well-sorted, medium-grained arkose (4a).

DEEP GULCH CONGLOMERATE

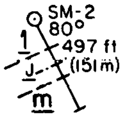
3a / 3 — **UNIT 3:** Subarkose, light pink, very coarse-grained to granular; plane-bedded. Includes 15 to 75 cm thick beds of hematite stained, pyritic, radioactive, quartz-pebble conglomerate. Overlain by plane-bedded coarse-grained to granular subarkose. Individual conglomerate beds are labeled G through L. I, J and K are strongly radioactive.

2 / 2m — **UNIT 2:** Arkose to subarkose, muscovitic, light gray, coarse-grained to granular; abundant small-scale trough crossbeds, slightly pyritic. Includes thin beds of quartz-pebble (qp) and arkosic (acgl) conglomerates. Interbedded with light gray very muscovitic coarse-grained subarkose (2m). Mildly radioactive conglomerates are labeled E and F.

1 / 1m — **UNIT 1:** Arkose to subarkose, muscovitic, light gray, coarse-grained to pebbly; poorly sorted, scattered K-feldspar pebbles. Includes thin beds of quartz-granule (qg), quartz-pebble (qp) and arkosic (acgl) conglomerates. Interbedded with very muscovitic, light gray coarse-grained arkose (1m). Slightly radioactive conglomerates are labeled A through D. C is a moderately radioactive K-feldspar and quartz-pebble conglomerate.

UNCONFORMITY

bqg — **BIOTITE QUARTZ PLAGIOCLASE GNEISS:** medium grained.

——— — —	Lithologic contact, dashed where uncertain	— — U/D — —	Inferred fault with relative movement
———	Outcrop boundary	Λ ; V	Fining-; coarsening-upward sequences
…………	Conglomerate horizon	▨	Planar crossbeds
.-.-.-.-.-	Conglomerate horizon projected to map surface.	⌒⌒⌒	Trough crossbeds
36 ; 42	Strike and dip of bedding; overturned	4	Sample transects
48	Strike and dip of foliation		Drill hole with bearing, plunge, total depth, and lithologic contacts projected to map surface.

Explanation for the plane-table map and cross section of Ridge 1, Carrico Ranch (figures 3.18 and 3.19).

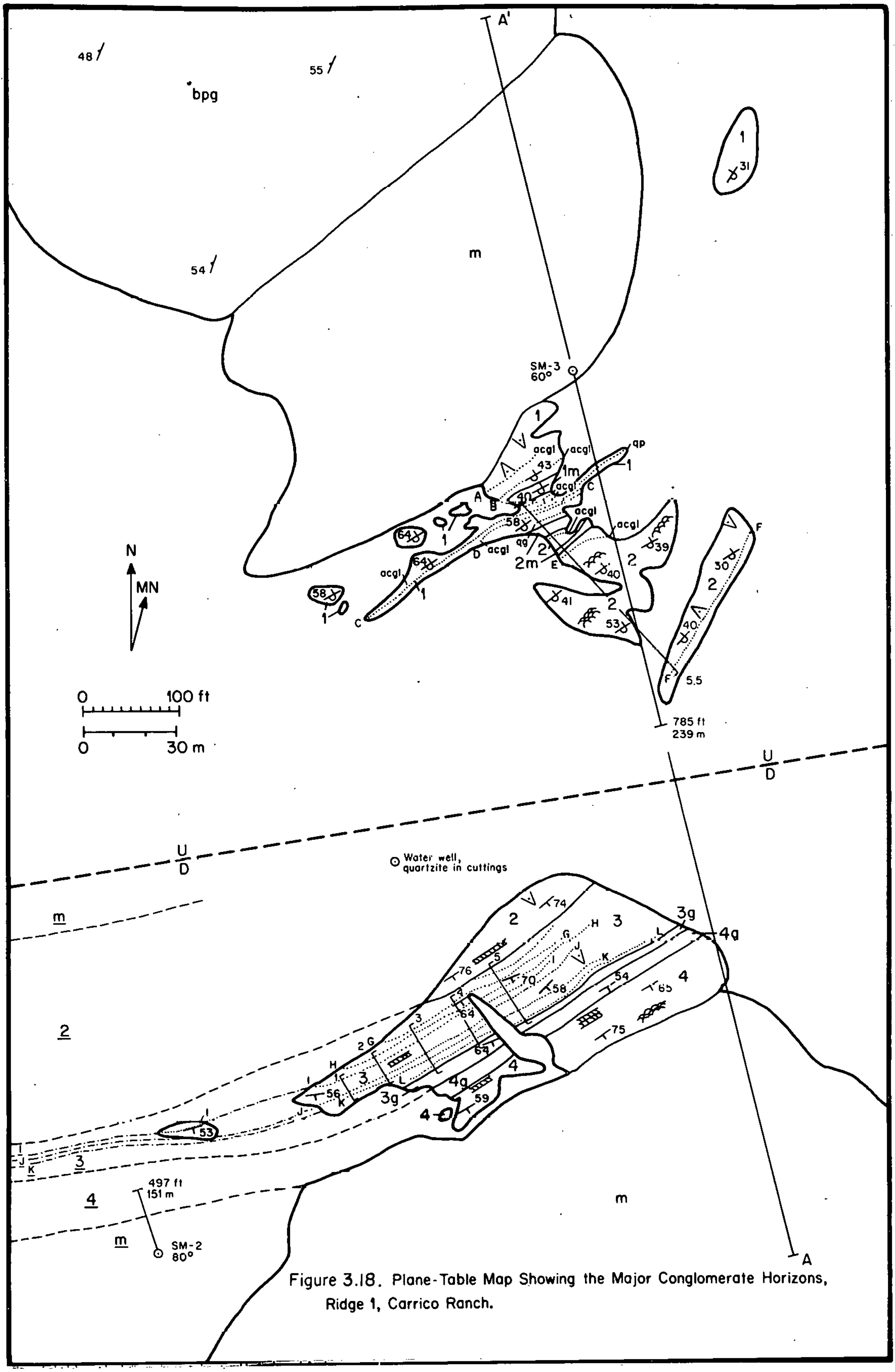


Figure 3.18. Plane-Table Map Showing the Major Conglomerate Horizons, Ridge 1, Carrico Ranch.

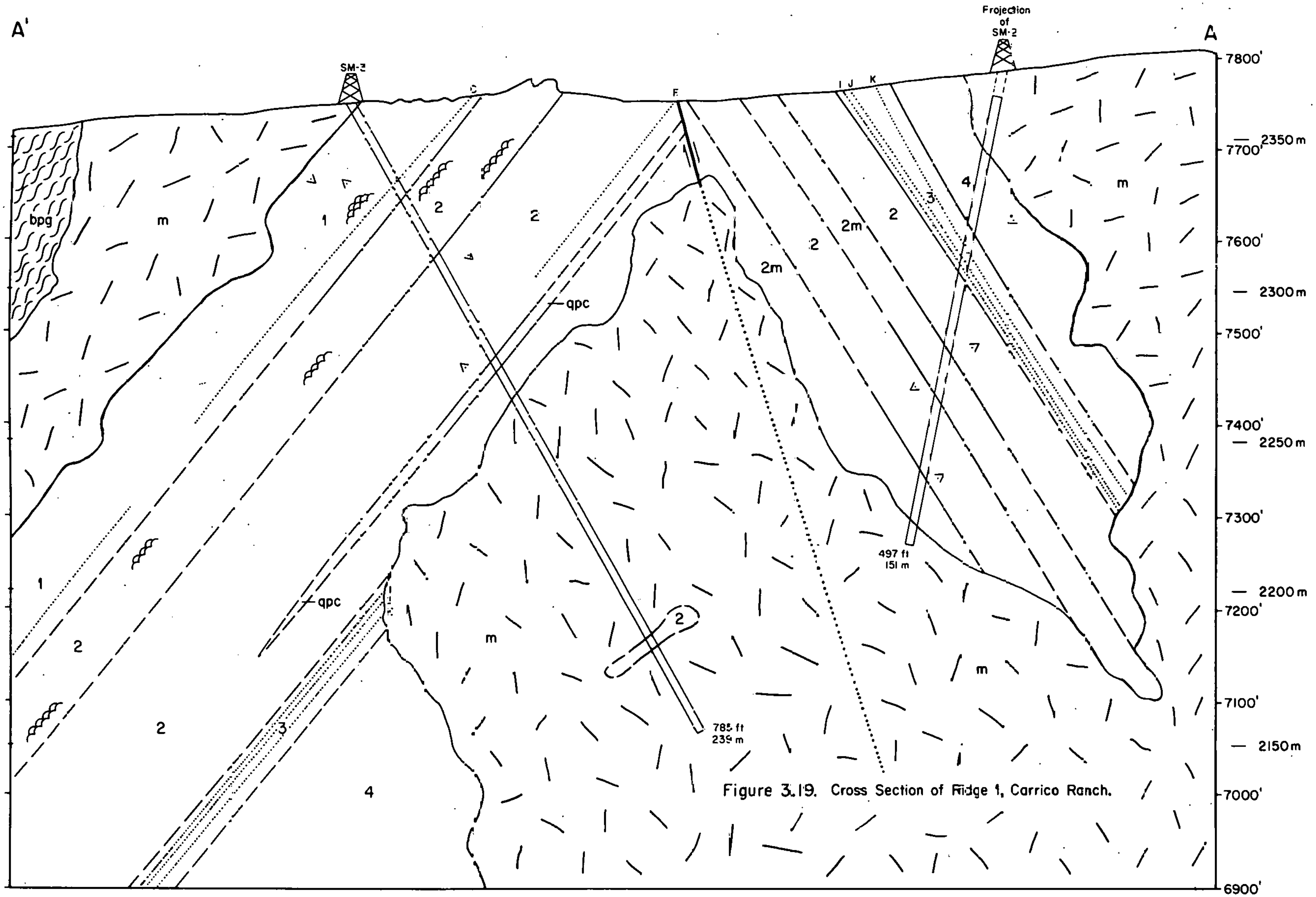


Figure 3.19. Cross Section of Ridge 1, Carrico Ranch.

Table 3.13 presents the characteristics of the biotite quartzite in outcrop, drill core and thin section. The unit is more closely related to the Deep Gulch Conglomerate than to basement rocks. It is not as severely deformed as the basement and it appears to be composed of sediment derived from the underlying paragneiss. However, the biotite quartzite appears to be separated from the Deep Gulch Conglomerate by a disconformity, which is evidenced by its lenticular distribution and the sharp contrast in lithologies and radioactive signature across the contact in Deep Gulch and in drill holes SM-2D and JP-3.

Deep Gulch Conglomerate

The Deep Gulch Conglomerate is divided into three informal units, each of which is distinguished by petrography, sedimentary bedforms and radioactivity. The main features of each unit observed in outcrops are summarized in Table 3.13

Unit 1. Unit 1 is a poorly sorted muscovitic arkosic to subarkosic conglomerate, interpreted to be a granitic gruss which was transported a short distance and has been reworked by fluvial processes. K-feldspar clasts which range from 8 to 15 mm in diameter occur in Unit 1 and the occurrences of such large K-feldspar grains may indicate that the source area was nearby (Pittman, 1969). The first five meters of sediment at Ridge 1, Carrico Ranch above the gabbroic sill (Figure 3.23) are very poorly sorted and contain only one or two recognizable sedimentary features. Above this zone sorting becomes better and fining- and coarsening-upward sequences, lag gravels and rare trough crossbedding are recognized; these bedforms suggest fluvial deposition (Collinson, 1978b).

TABLE 3.13 DESCRIPTION OF FOUR INFORMAL UNITS OF THE DEEP GULCH CONGLOMERATE AND LOWER JACK CREEK QUARTZITE. THE DESCRIPTIONS ARE PRIMARILY FROM RIDGE 1, CARRICO RANCH. NOT ALL FEATURES ARE PRESENT IN EACH OUTCROP.

Unit 3

Hematite-stained, sericitic, pyritic, oligomictic, radioactive quartz-pebble conglomerates:

1) medium-grained to granular planar crossbeds;
2) fining-and coarsening-upward sequences in conglomerate horizons;
3) multiple conglomerates may occur in one conglomerate horizon;
4) thickness of conglomerates varies from 2 to 70 cm;
5) number of conglomerate horizons varies from 3 to 15 per outcrop;
6) anomolous radioactivity 3 to 18 times background in conglomerate horizons;
7) rounded to euhedral pyrite grains in conglomerates;
8) rare trough crossbeds;
9) sorting varies from very poor to moderately-sorted;
10) loss of mica in non-conglomerate beds compared to conglomerate beds

Unit 2

Coarse-grained to granular micaceous subarkose, trough crossbedded in the lower portion:

1) lenticular radioactive lag gravels:
2) lag gravels are hematite-stained and 3-5 times background radioactivity;
3) trough crossbeds fine-upward, with granular bases 2-3 times background radioactivity;
4) thickness of trough crossbeds varies from 4 to 20 cm;
5) decrease in the number of trough crossbeds upsection;
6) increase in the number of quartz-pebble conglomerates and fining sequences upsection;
7) fining sequences are variable in thickness;
8) size and percentage of K-feldspar decreases upsection;
9) clast size increases upsection.

Unit 1

Medium-grained to granular micaceous arkose and subarkose:

1) hematite-stained lenticular lag gravels;
2) angular K-feldspar pebbles up to 15 mm in diameter;
3) radioactivity 3-6 times background in lag gravels;
4) fining-upward and -downward sequences of variable thickness;
5) rare trough crossbedding;
6) poorly to very poorly sorted

Unit bq (bq, bqs on map explanations and core logs)

Dark gray, medium-grained to granular micaceous subarkose:

1) non-radioactive;
2) poorly sorted;
3) fining and cleaning sequences;
4) rare trough crossbedding;
5) muscovite exceeds biotite in thin section;
6) rare K-feldspar grains

Lag gravels vary greatly in their lateral and downdip extent. In outcrop (Figures 3.18, 3.20, and 3.21) lag gravels vary from 1 m to 40 m in width and are up to 4 cm thick, these lag gravels often form the base of a fining-upward sequence. Downdip extent of the lag conglomerates is difficult to determine; outcrop relief tends to be only 1-3 m high in Unit 1. Examination of radioactive conglomerates from drill hole SM-3 provided a possible correlation with conglomerate C at the surface. Known downstream extent, from Figure 3.19, of conglomerate C is 36 m with an inferred extent of 140 m. Conglomerate C is also the thickest conglomerate in Unit 1 with thickness ranging from 5 to 30 cm.

Unit 2. Unit 2 is predominately a trough crossbedded subarkose with thin conglomerate lenses formed by lag gravels (up to 4 cm thick) which separate individual trough crossbedded sequences that range from 1 to 10 m thick (Figure 3.23). The lower contact of Unit 2 is defined as the point when trough crossbedding replaces fining sequences as the dominant bedform. The upper contact is placed 2 m below the first highly radioactive conglomerate of Unit 3. This arbitrary definition was established to include within Unit 3 all marginally radioactive conglomerates. K-feldspar grains of Unit 2 are smaller than those in Unit 1, decreasing in average size from 12 to 18 mm in diameter. Grains in Unit 2 decrease in size upsection from 10 mm to 6 mm. Trough crossbedding also occurs less frequently upsection at Ridge 1, Carrico Ranch and at Deep Gulch. Plane-bedded conglomerates become the dominant bedform near the top of Unit 2 as shown in Figure 3.23. Trough crossbeds may be present at the top of fining-upward sequences.

EXPLANATION

m — **MAFIC IGNEOUS ROCKS:** Mafic sills and dikes ranging from massive gabbro and diabase to foliated amphibolite.

INTRUSTIVE CONTACT

5a / 5 — **UNIT 5:** Phyllite, black, with boudinaged quartz veins. Contains lenses of pink, medium-grained arkose (5a).

4m / 4 / 4p — **UNIT 4:** Arkose and subarkose, pink, medium-grained; planar-crossbedded, massive, well-sorted. Intertongues with poorly-sorted muscovitic subarkose (4m) and thin lenses of black phyllite (4p).

DEEP GULCH CONGLOMERATE

3 — **UNIT 3:** Subarkose, light pink, very coarse-grained to granular; plane-bedded. Includes 15 to 75 cm thick beds of hematite-stained, pyritic, radioactive quartz-pebble conglomerate. Individual conglomerate beds are labeled G through L; J and K are strongly radioactive.

2 — **UNIT 2:** Arkose to subarkose, muscovitic, light gray, coarse-grained to granular; abundant small-scale trough crossbeds, slightly pyritic. Includes thin beds of quartz-granule (qg), quartz-pebble (qp) and arkosic (agcl) conglomerates. Mildly radioactive conglomerates are labeled C through F, C is the most radioactive

1 / 1m — **UNIT 1:** Arkose, light gray, medium- to coarse-grained; poorly-sorted. Interbedded with muscovitic arkose (1m). Includes thin quartz-pebble (qp) and arkose (acgl) conglomerates. Slightly radioactive conglomerates are labeled A and B.

———— – –	Lithologic contact, dashed where uncertain	42	Strike and dip of foliation
———— – –	Outcrop boundary, dashed where uncertain	76	Strike and dip of faults and quartz veins
		8	Bearing and plunge of minor fold axis
....	Conglomerate horizon	Λ	Fining-upward sequences
U ———— – – D	Fault, showing relative displacement, dashed where approximately located		Planar crossbeds
44	Strike and dip of overturned bedding		Trough crossbeds
SM-1 ⊙ Vertical 626 ft (191m)	Drill hole with plunge angle and total depth.	8	Sample transects

Explanation for the plane-table map and cross section of Ridge 2, Carrico Ranch (figures 3.20 and 3.21).

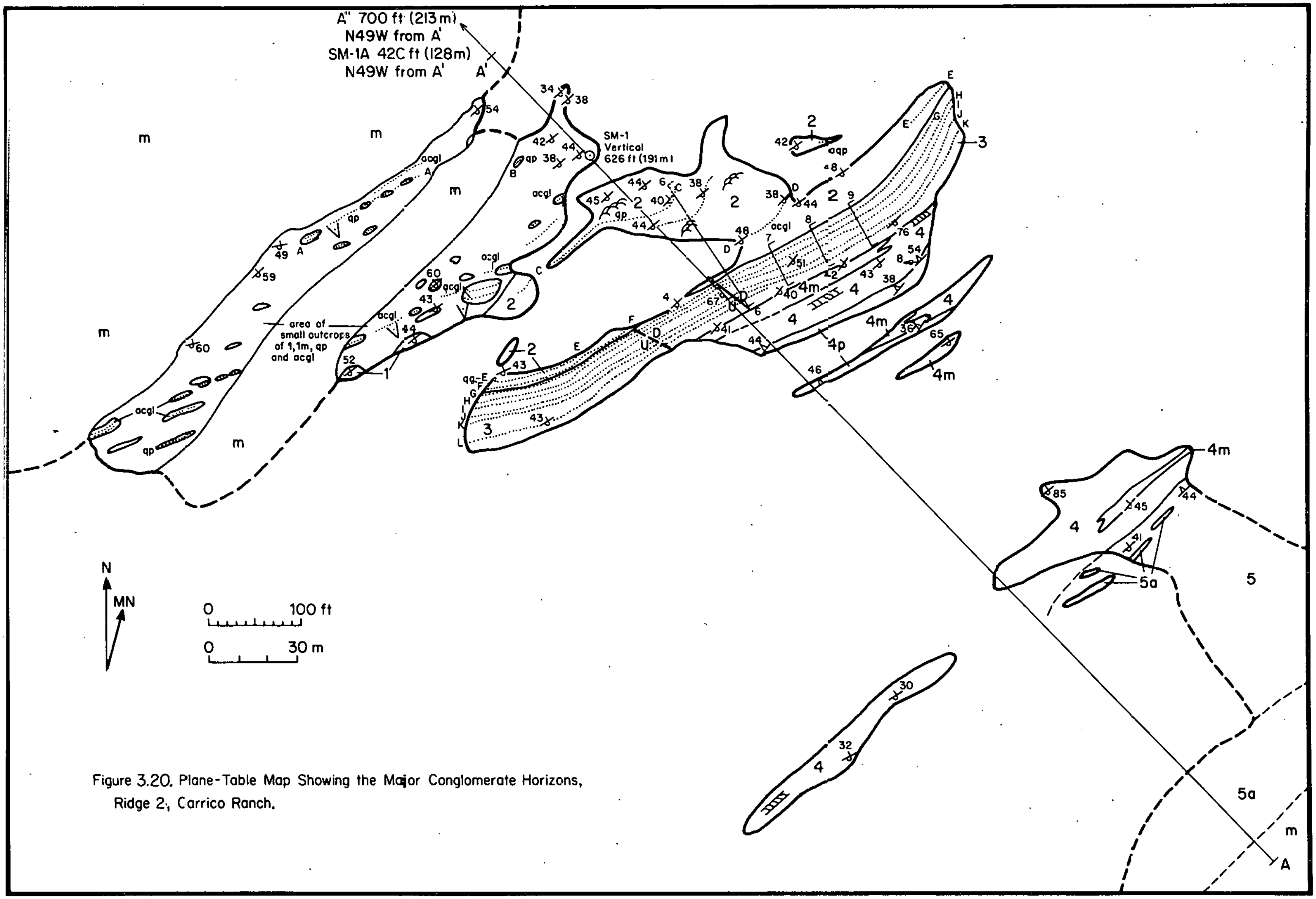


Figure 3.20. Plane-Table Map Showing the Major Conglomerate Horizons, Ridge 2, Carrico Ranch.

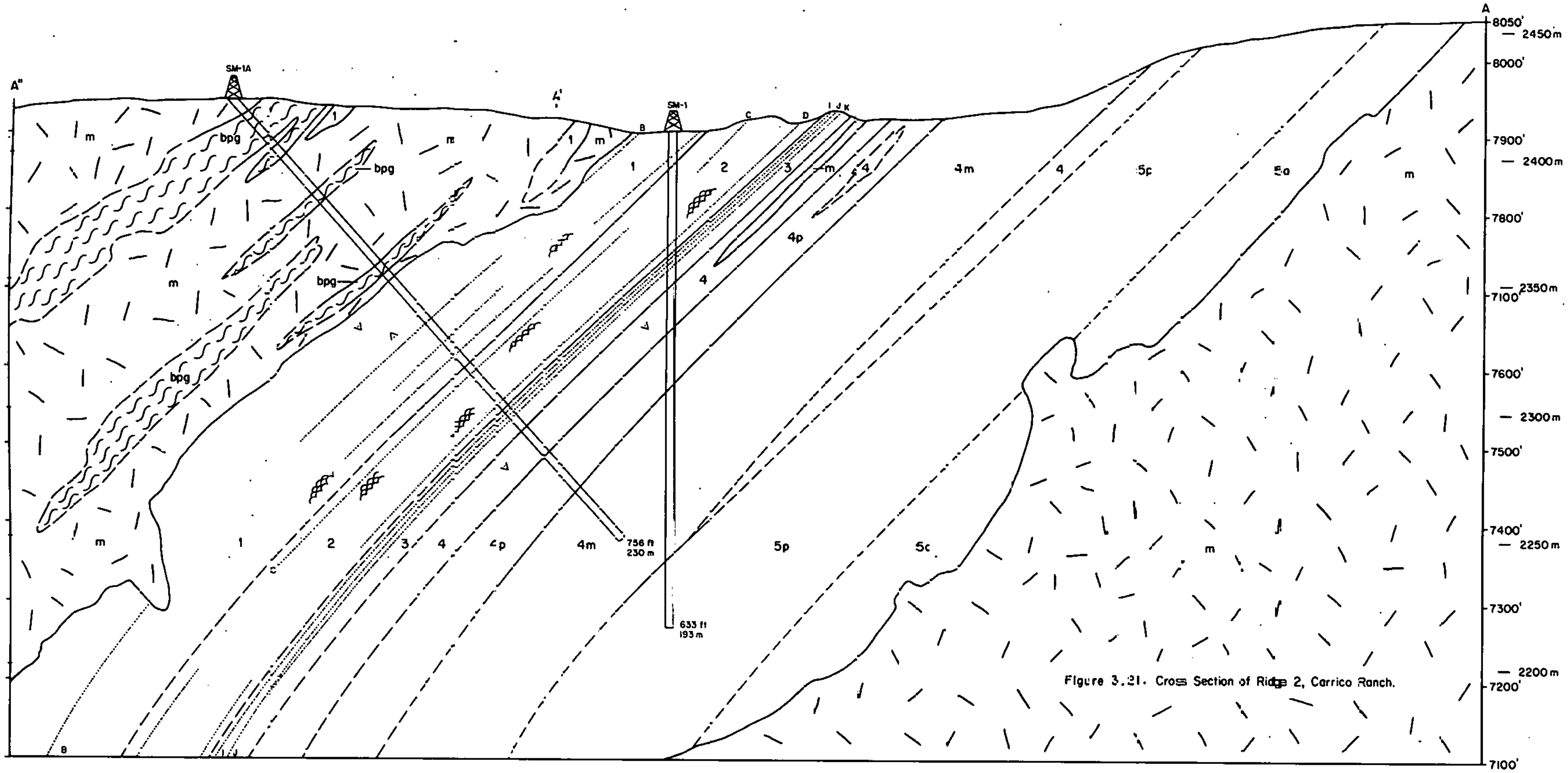


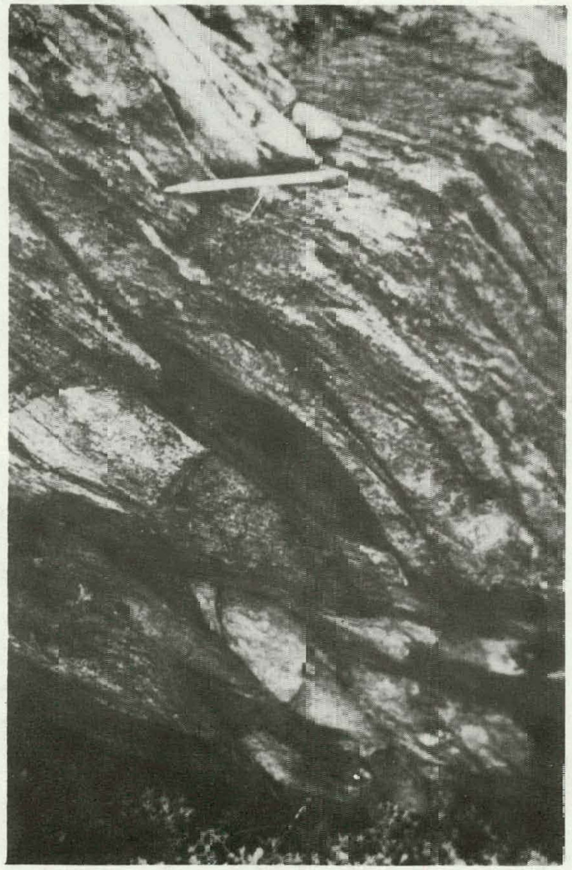
Figure 3.21. Cross Section of Ridge 2, Carrico Ranch.

Trough crossbeds within Unit 2 are variable in their amplitude and downstream extent. Figure 3.22 shows typical crossbeds; in Figure 3.22A, a solitary trough is shown, the trough is well exposed and is about 6 cm thick. Figure 3.22B shows the crosscutting (festoon) nature of the troughs and illustrates the great variability in amplitude (2-30 cm) and width (0.25 to 1.0m) of the troughs.

The thin, lenticular conglomerates which separate trough crossbedded sequences (Figures 3.18 and 3.23) have an unknown downdip extent. In outcrop the conglomerates can be traced for 10 m at Ridge 2, Carrico Ranch; correlation of conglomerates in outcrop with those intersected in drill core proved to be too tenuous to be of use.

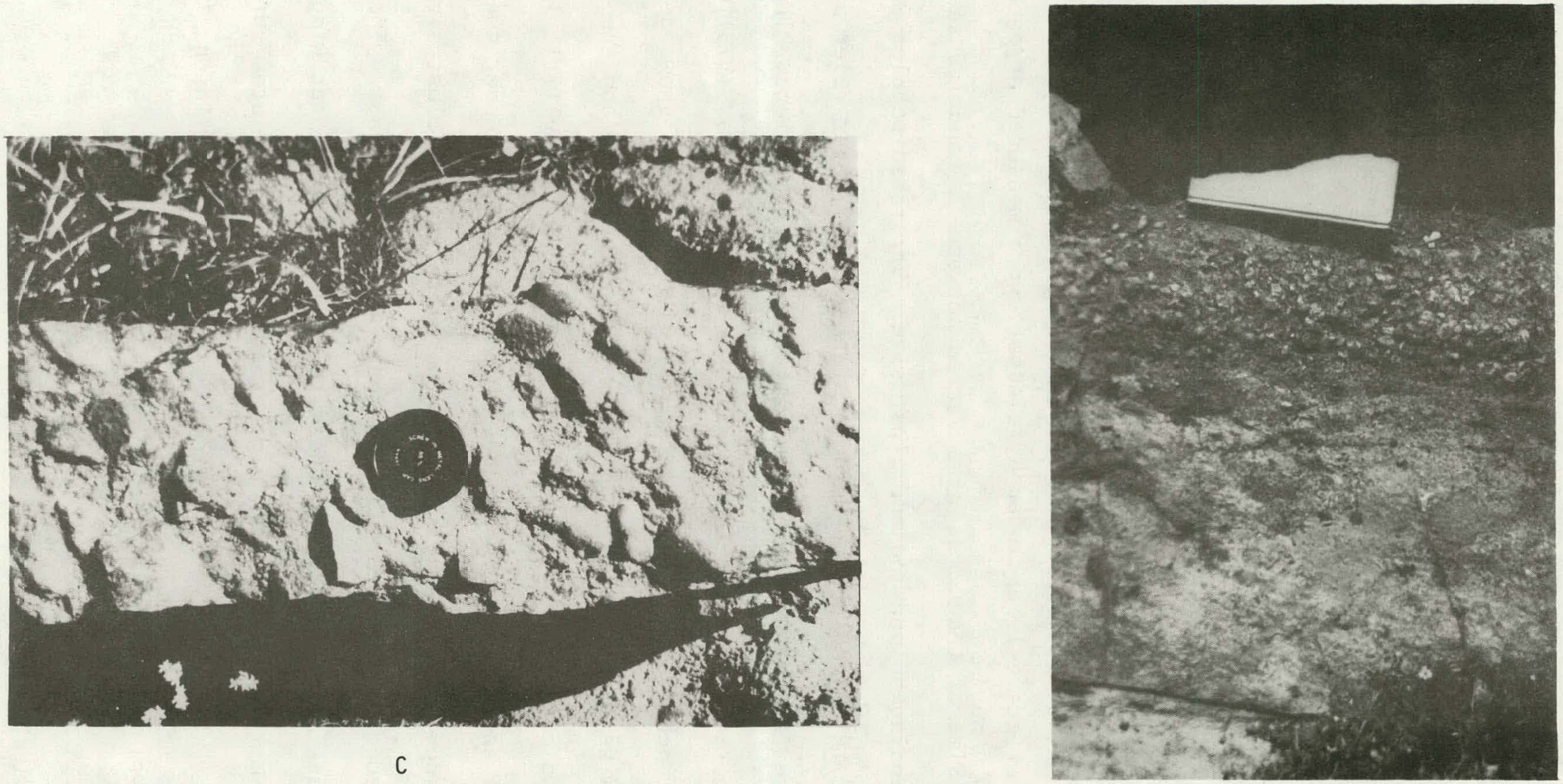
Plate 7 shows Units 1 and 2 combined in the Manning Ranch area. This grouping of Units 1 and 2 in this area was necessary because of: 1) the lack of outcrop, 2) the paucity of sedimentary structures in the available outcrop and drill core, 3) the lack of distinguishing K-feldspar size (10-15 mm), and 4) the uncertainty of where the basal Deep Gulch Conglomerate contact is in this area.

Unit 3. Unit 3 contains the most radioactive, thickest, and laterally continuous quartz-pebble conglomerates of the Deep Gulch Conglomerate and the only conglomerate horizons of potential economic interest (Figure 3.22). Plate 7 shows the lateral variation in the thickness of Unit 3, the continuity of individual conglomerate beds and the variation of clast size. Figures 3.22, 3.25, and 3.26 all present data collected from the quartz-pebble conglomerates at Ridge 1, Carrico Ranch.



A

B

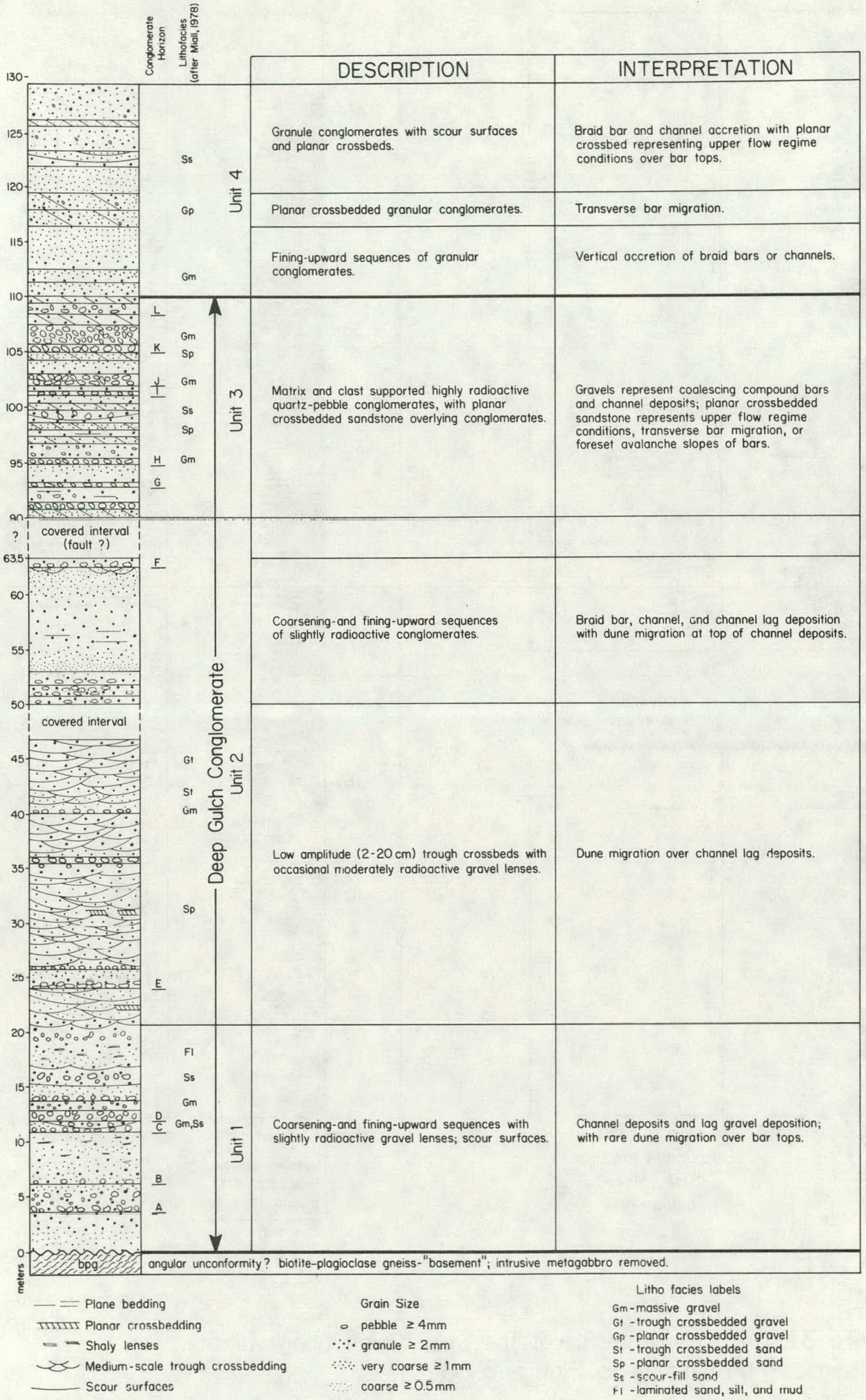


C

D

Figure 3.22. Primary sedimentary structures of the Deep Gulch Conglomerate. A - overturned trough crossbed, Unit 2, Ridge 1, Carrico Ranch, pemcil for scale. B - overturned crosscutting (festoon) trough crossbeds, Unit 2, Ridge 1, Carrico Ranch, notebook is 5 inches wide. C - conglomerate horizon K(?), maximum size of quartz-pebbles occurs in this horizon, Unit 3, Ridge 2, Carrico Ranch, lens cover is 5.2 cm in diameter. D - conglomerate horizon J, note the basal scour surface into the underlying sandstone, Unit 3, Ridge 2, Carrico Ranch, notebook is 5 inches wide.

Measurement of the largest clast sizes in radioactive conglomerate horizons indicates that clast size decreases along the strike, northeast and southwest, from Ridge 2, Carrico Ranch as shown in Plate 1. Clasts from the most radioactive conglomerate horizon in each outcrop were measured to provide data for the study of the sedimentology of mineralized conglomerates to help reconstruct depositional models. Also, Theis (1979) reported that clast size is related to mineralization of the conglomerate horizon and both features appear to be related to the depositional energy. Between 125 and 300 clasts larger than 6 mm (this effectively excluded K-feldspar) were measured at each sample location. Size of the sample area varied between 0.5 and 1.0 square meters depending on packing and clast size. Stretched pebbles were encountered in all outcrops; using a standard millimeter ruler pebbles were measured to the nearest whole number in two dimensions. The 50 maximum-apparent diameters (long axis of a pebble, Theis, 1979) of the largest pebbles were used to compile Plate 7, which shows decreasing clast size along strike from Ridge 2, Carrico Ranch and the occurrence of two conglomerate successions separated by a phyllitic layer, at Ridges 1 and 2 at Manning Ranch. Clast size of the upper conglomerate sequence increases slightly to the northeast and clast size of the lower conglomerate decreases markedly to the northeast. The variation in clast size and the increased thickness of Unit 3 at Ridges 1 and 2, Carrico Ranch relative to other areas shown in Plate 7 are interpreted as the result of coalescing fluvial fans (Collinson, 1978b; Blunck, 1965; Bull, 1977).



DESCRIPTION	INTERPRETATION
Granule conglomerates with scour surfaces and planar crossbeds.	Braid bar and channel accretion with planar crossbed representing upper flow regime conditions over bar tops.
Planar crossbedded granular conglomerates.	Transverse bar migration.
Fining-upward sequences of granular conglomerates.	Vertical accretion of braid bars or channels.
Matrix and clast supported highly radioactive quartz-pebble conglomerates, with planar crossbedded sandstone overlying conglomerates.	Gravels represent coalescing compound bars and channel deposits; planar crossbedded sandstone represents upper flow regime conditions, transverse bar migration, or foreset avalanche slopes of bars.
Coarsening-and fining-upward sequences of slightly radioactive conglomerates.	Braid bar, channel, and channel lag deposition with dune migration at top of channel deposits.
Low amplitude (2-20 cm) trough crossbeds with occasional moderately radioactive gravel lenses.	Dune migration over channel lag deposits.
Coarsening-and fining-upward sequences with slightly radioactive gravel lenses; scour surfaces.	Channel deposits and lag gravel deposition; with rare dune migration over bar tops.
angular unconformity? biotite-plagioclase gneiss-"basement"; intrusive metagabbro removed.	

Figure 3.23. Measured stratigraphic section and paleoenvironmental interpretation of the Deep Gulch Conglomerate, Ridge 1, Carrico Ranch.

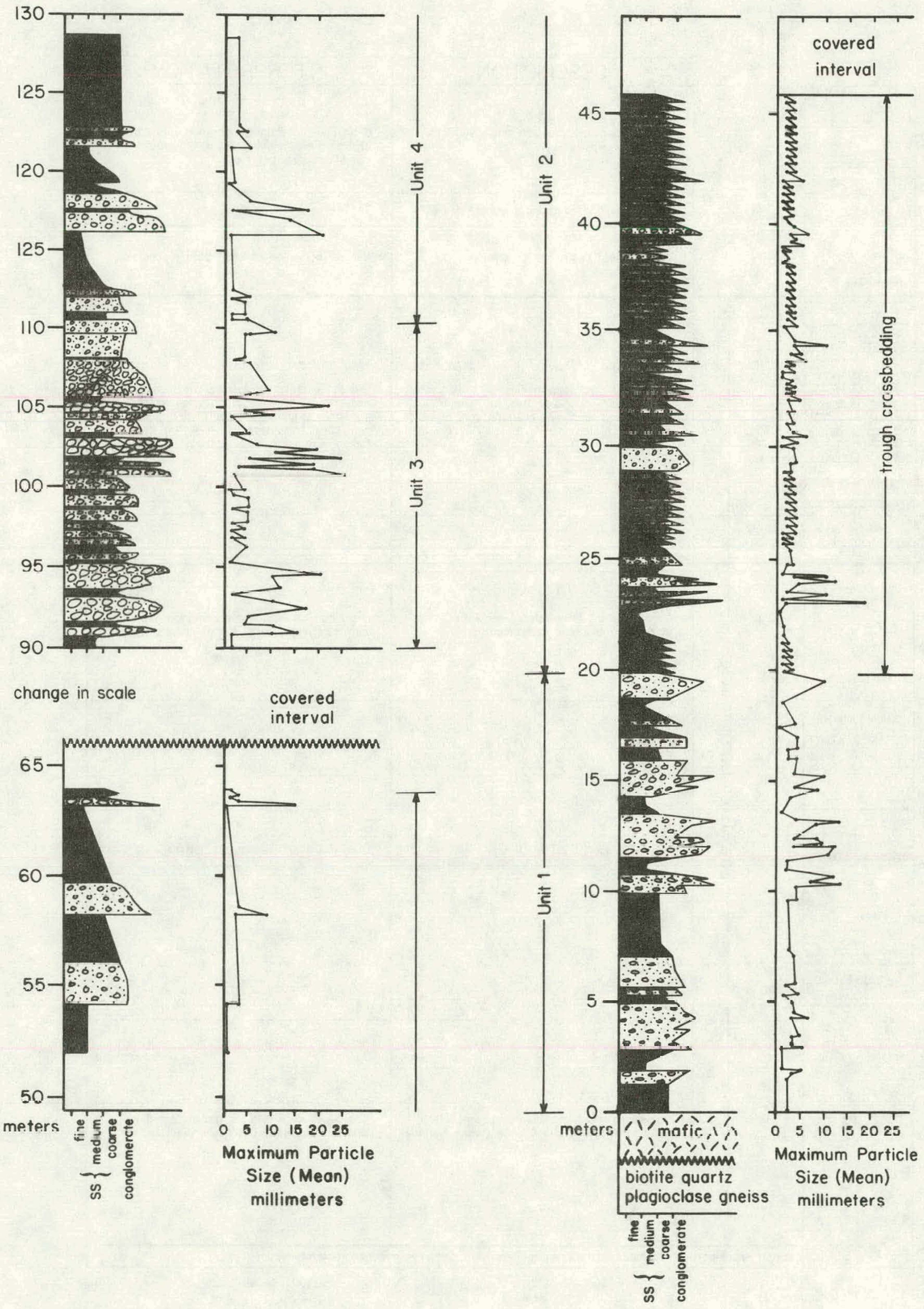


Figure 3.24. Grain size profile of the Deep Gulch Conglomerate. This section corresponds to Figure 3.23.

Vertical Variation of the Deep Gulch Conglomerate

In stratigraphic section, the Deep Gulch Conglomerate consists of one distinct coarsening-upward sequence which is accompanied by a change in sedimentary bedforms and decrease of K-feldspar clast size. Figure 3.24 shows the maximum particle size for Ridge 1, Carrico Ranch. The diagram illustrates that the trough crossbedded sequence (Unit 2) is at the base of a coarsening-upward sequence that culminates in the conglomerate horizons of Unit 3.

Variation of sedimentary structures is illustrated in Figures 3.23 and 3.24. Unit 1 is dominated by coarsening and fining-upward sequences with rare trough crossbedding. Unit 2 is characterized by crosscutting (festoon) trough crossbedding which grades vertically into fining sequences with poorly developed lag gravels at their base and trough crossbeds near the top of the sequence. Unit 3 is dominated by horizontally-bedded conglomerate horizons, planar crossbedded sands and gravels, and plane-bedded sands.

Decrease in the grain size of K-feldspar and the amount of K-feldspar occurs upsection in the Deep Gulch Conglomerate. Unit 1 often has arkose lag gravels with K-feldspar grains up to 15 mm in diameter, Unit 2 has a gradual decrease in K-feldspar size from 10 mm to less than 6 mm, the amount of sand-size and finer-grained K-feldspar increases upsection also. Unit 3 contains K-feldspar up to 8 mm but grains larger than 4 mm are rare, the majority of K-feldspar is concentrated in the matrix of the conglomerates and is coarse-grained or finer. Lithologically both Units 2 and 3 are subarkoses as shown by the ternary digrams in Figure 3.27. These vertical variations can be explained in terms of a coalescing fluvial fan model for the deposition of the Deep Gulch Conglomerate (Heward, 1978; Spearing, 1975).

Units 4 and 5. Units 4 and 5 which are used in Plate 7, geologic maps, cross-sections and drill hole lithologic logs directly overlie the Deep Gulch Conglomerate. These units were only identified on detailed plane-table maps of the northwestern Sierra Madre and are not widely recognized subdivisions of the Jack Creek Quartzite. Unit 4 is a massive fining-upward non-radioactive sequence consisting of subarkosic quartz-granule conglomerates (with scattered quartz-pebbles) grading upward to medium- to coarse-grained, moderately sorted subarkose and occasional quartzarenite. K-feldspar in this unit is finer grained than in Unit 3 (less than 0.5 mm). Unit 5 is a dark gray, fine-grained quartz phyllite with boundinaged quartz veins and veinlets which forms lenses within Unit 4 quartzites. Unit 5 crops out at Ridge 2, Carrico Ranch and also occurs in drill hole SM-2D.

SEDIMENTOLOGY OF THE DEEP GULCH CONGLOMERATE

The Deep Gulch Conglomerate is interpreted as a braided river deposit on a coalescing fluvial fan. Support for the fluvial origin is: probable angular unconformity with underlying biotite plagioclase gneiss and biotite quartzite; lenticular lithofacies; abundant trough crossbedding; poor to moderate sorting; and coarsening and fining-upward stratification sequences similar to those in braided fluvial lithofacies models (Miall, 1977, 1978; Steel, and others 1977). Support for braided fluvial sedimentation is that pre-Mid-Paleozoic time was dominated by braided river deposition (Cotter, 1978), due to the absence of vegetation which stabilized river banks and controlled sediment runoff (Long, 1978), in addition to the lithofacies models. Deposition probably took place in a "wet" fluvial fan environment rather than the "dry" alluvial fan environment common in southwestern U.S.A. today. Schumm (1977, p. 259) distinguished wet fluvial fans from dry alluvial fans by the concentration of heavy minerals at several stratigraphic levels on the fan rather than dispersion which is common in dry alluvial fans, the braiding of sediments with well-developed bedforms instead of the near random distribution of erosion and deposition in dry alluvial fans, and the scarcity of mudflows (debris flows). The Deep Gulch Conglomerate exhibits all these characteristics of wet fluvial fans.

Braided Stream Environment

Before considering the sedimentological characteristics of the Deep Gulch Conglomerate in detail, it is appropriate to briefly review some of the processes and characteristics of braided river systems and the distribution of heavy minerals within those systems. The following summary of braided river environment is primarily from Miall (1977, 1978), and Smith and Minter (1980).

Braided rivers are typified by shifting bars and channels, high width/depth ratios, relatively steep slopes and low channel sinuosities. The scale of the bars and channels is an indication of the size of the fluvial system. The Bramaputra River of India contains sand waves up to 1-2 km in length (Coleman, 1969) while smaller streams have braid bars up to tens of meters in length and often less (Boothroyd and Ashley, 1975). Due to the widely fluctuating flow in braided rivers, velocity, depth and sediment grain size are quite variable. Gravel in braided streams tends to be transported in bursts, according to local flow conditions, moving a few meters or tens of meters, and then becoming a new temporary bar. Complex bar deposits (compound bars of Smith and Minter, 1980) and channel fill, which may represent multiple periods of deposition and erosion, are more likely to be preserved than bars formed by depositional processes alone, (unit bars of Smith, 1974). These compound bars may be formed by the "vertical overriding or lateral coalescence of smaller braid bars, gravel sheets on small-scale bedforms" (Smith and Minter, 1980). When these bars and adjacent channels are a stable feature within a river system, they may be important in concentrating and preserving heavy minerals; because as the bars are reworked by currents, heavy minerals will become concentrated (trapped) within the coarser bed material. The braid bars, a general term for any bar in braided rivers, form in channels and are localized by secondary currents (Leopold and Wolman, 1957) or are initiated by build up from gravel sheets (Hein and Walker, 1977). These bars and channels are only temporary sites of gravel and heavy mineral deposition, and are termed local point sources, from which minerals may be concentrated further or dispersed. These local point sources while actually temporary in the river system may become permanent if the system is aggrading.

Bedforms which often occur with compound and longitudinal bars are planar crossbeds (tabular and non-tabular) with possible tangential lower contacts (Smith, 1971, 1974; Boothroyd and Numendahl, 1978). These are interpreted to represent transverse bar migration and slipface migration of longitudinal bars (Williams and Rust, 1969; Rust, 1975). Fining-upward sequences within bars have been well documented (Miall, 1977). Small scale planar crossbeds and plane beds overlying bar tops have been interpreted by Rust (1972) to have formed in response to the shallowing of water, which results in a transition from low flow regime (trough crossbeds and ripples) to upper flow regime bedforms.

Paleocurrent measurements in braided river systems may yield a great diversity (Williams and Rust, 1969). Trough crossbeds yield the lowest variance while planar crossbeds yield the highest variance. The directional variability tends to vary with the scale of the bedform being measured (similar scale yields less variance) and the discharge level of the stream; high discharge equals lower variance due to lesser effect of bar topography (Miall, 1977).

Vertical profile models of the braided river system described by Rust (1978) and Miall (1977, 1978), reflect the most commonly occurring facies associations within this depositional environment. Six vertical profiles are proposed by Miall (1978) which encompass many of the variations in braided river systems from proximal-gravelly, to distal-sandy, environments. We conclude that the sedimentary features of the Deep Gulch Conglomerate fit most closely with the Donjek and Scott models of Miall (1977, 1978).

Paleocurrent Analysis

Paleocurrent data from the Deep Gulch Conglomerate and quartzite units immediately overlying the Deep Gulch Conglomerate are shown in Plate 7. These data are somewhat equivocal for several reasons. The biggest problem is related to the complexity of folds in the area. Structural information for unfolding the folds (following Ramsey, 1961) was obtained from structural analysis of folding using stereographic projection. These stereographic plots show a northwest-plunging reclined and isoclinal fold system with at least two other superposed but less prominent, intralimb fold systems, we are not sure that the reclined fold system is the earliest system and we did not consider the effect of strain. As a consequence, our unfolding procedures are at best only an approximation of the true orientation of the beds; any errors in estimating the plunge of the folds will cause very large errors in the direction of the inferred paleocurrents, especially in the overturned limbs, where all our data lie (Ramsey, 1961).

Initial results after unfolding were quite discouraging, and yielded bimodal and diverse patterns from one outcrop area to another. Diverse directions can be accounted for in an alluvial fan-braided river system (Selley, 1968; Howard, 1966; Miall, 1977) but bimodal currents would not be expected, especially because most measurements were from the axes of trough crossbeds. We decided to assume a unimodal distribution, an assumption that is consistent with other sedimentary evidence which strongly suggests fluvial sedimentaton, and to take those paleocurrents which, after rotation, plunged less than 10° in the direction opposite to the majority of paleocurrents and "flip" them 180° (dashed lines in the rose diagrams of Plate 7 show what percentage were flipped). Using this correction of data, the results are more believable than before flipping and are discussed below.

The synopsis of diagram of 117 paleocurrent measurements, from the Carrico Ranch area (Plate 7) shows a unimodal pattern about a south-southeast directed mean paleocurrent (Plate 7). Individual outcrop areas cluster about this mean direction except at Ridge 1, Unit 3 where there is a wide dispersion. This dispersion may result from measuring planar crossbeds in this outcrop, which may represent transverse bars or slip facies on longitudinal bars, and thus vary up to 90 degrees from the "true" paleocurrent direction (Potter and Pettijohn, 1977).

The synopsis diagram for the Manning Ranch and Mill Creek areas was compiled from 58 planar and trough crossbed paleocurrent measurements whose mean direction is west-southwest. This diagram and the individual area diagrams show a wide dispersion about the mean, also probably due to the large number of planar crossbeds that were measured. The 17 trough crossbeds measured at Joes Park are unimodal and directed to the south.

The source area for the Deep Gulch Conglomerate appears to have been to the north-northeast. We believe the diversity of the paleocurrents is real and can be explained in terms of a coalescing fluvial fan-braided stream depositional model.

Stratification Sequence

As shown in Figures 3.23 and 3.24, the stratigraphic section of the Deep Gulch Conglomerate contains one coarsening-upward sequence (Unit 2 through Unit 3) and two fining-upward sequences (Units 1 and

4). Coarsening-upward sequences reflect uplift of the basin-margin which manifests itself as coarsening-upward sedimentation in the depositional basin (Steel, and others, 1977). Rust (1979b) described fining-upward sequences as the result of establishment of equilibrium conditions after the initial uplift of a highland, Units 1 and 4 are interpreted in this light. Unit 1 is composed of the initial sediments transported off a granitic highland and unconformably deposited on the basement and older biotite quartzites; further uplift produced the aggrading sequence of trough crossbeds in Unit 2 and the prograding Unit 3 fluvial fan sediments. A return to equilbirium conditions is seen in Unit 4, as it fines upward.

Unit 1. Sedimentary structures present in Unit 1 (Table 3.13) reflect the accretion of sediments in channels and small braid bars. Coarsening-upward sequences are interpreted as surge deposits (Smith, 1970), while fining-upward sequences are the result of vertical accretion in waning flow conditions or bar build up (Rust, 1972, 1978; Williams and Rust, 1969; Miall, 1977). Channel lag gravels contain quartz and K-feldspar pebbles, these are usually thin, from 1 to 2 pebbles thick and up to 35 cm thick in conglomerate C at Ridge 1, Carrico Ranch. The width of the channel lag in outcrop varies from 5 to 60 m and this is considered to be a minimum value for the original width of the channels (Cant and Walker, 1976; Smith, 1970; Miall, 1970). The lag gravels are lenticular and some portions are eroded away by overlying lag gravels, the most radioactive lag gravels (those shown in cross sections and plane-table maps) are seldom eroded.

Cross sections of the ridges and drill holes, Figures 3.18, 3.19, 3.20, and 3.21 show the correlation of the outcrop lag gravels to those identified in the drill core. Downdip minimum length of lag gravels appears to vary from 2 m to 30 m. The great variability of the dimensions of bodies of lag gravel is most likely due to the anastomizing pattern of braided rivers and the variability of stream flow. The identification of bar sequences in Unit 1 is tenuous, the only exposure of possible braid bars is in the upper 5 m of Unit 1, at Ridge 1, Carrico Ranch and Ridge 2, Manning Ranch. These are identified by their fining-upward sequences with rare trough crossbeds at the top.

Unit 2. Trough crossbedding is the major bedform present in Unit 2. Lag gravels and fining-upward sequences become the dominant sedimentary structures in some areas (Figures 3.23 and 3.24). Cross-cutting trough crossbeds (festoon, pi cross-stratification of Allen (1963) have been interpreted as megaripple or dune migration in fluvial channels (Miall, 1977) often the amplitude of the trough crossbeds decreases upsection. Occasional lag gravels outline the base of channels and some of the gravels may represent small scale braid bars which formed in the channels. The thick sequences of trough crossbedding separated by lag gravels in the Deep Gulch Conglomerate indicate that the stream was aggrading and channels were relatively stable. The evidence for stability and aggradation is that few episodes of non-deposition are recognized. Erickson (1978) describes scouring and lag gravels as the markers for such periods of non-deposition.

The fining-upward sequences at the top of Unit 2 are interpreted as aggrading shallow braid bars containing occasional lag gravel deposits. The thick fining-upward sequence at the top of Unit 2 (Figure 3.23 from 55 to 62 m) is interpreted as a channel deposit developed during waning flow conditions as might occur if a bar, which progressively reduced water flow, was developed upstream.

The increased thickness of Unit 2 at Ridges 1 and 2, Carrico Ranch (Plate 7) is interpreted as a function of increased sedimentation in the apex of alluvial fan (Blunck, 1965, Bull, 1972, 1977).

Unit 3. Unit 3 which contains the thickest and most radioactive conglomerates is interpreted to have formed in a coalescing braided fluvial fan system and was dominated by thick accumulations of gravel and subordinate finer grained planar cross beds and plane-bedded sands . Plate 7 shows that the apparent-maximum diameter of pebbles, in Unit 3, decreases laterally along strike northeast and southwest of Ridge 2, Carrico Ranch. This pattern of change of clast size in Unit 3 is similar to patterns of decreasing pebble size in alluvial fans investigated by Blunck (1965) and Miall (1970). The maximum pebble size is believed to occur at the apex of the fan with the pebble size deceasing laterally.

Two varieties of conglomerate sequences in Unit 3 are depicted in Figures 3.25 and 3.26. Figure 3.25 is the single conglomerate horizon typified by conglomerate "J", Carrico Ranch area. Conglomerate "J" appears to represent a single depositional event, whose base is a scour surface (Figure 3.22C). Planar crossbeds and fining-upward sequences are in intimate association with the conglomerate horizon. The underlying and overlying planar crossbeds are interpreted as transverse bars some of which may have migrated over compound bars and other braid bars. The braid bars are recognized by their fining-upward nature and lack of other recognizable bedforms. This conglomerate horizon is the result of the formation of a clast-supported conglomerate similar to those described by Boothroyd and Ashley (1975) in the Scott and Yama glacial outwash fans. When compared to conglomerate I in Figure 3.26 the relatively low uranium and thorium values indicate that the channel gravel was not significantly reworked, had little new sediment added to it, and did not entrap significant amounts of heavy minerals. These conditions may be due to rapid deposition of, and burial by, overlying transverse and braid bars.

Conglomerate horizon "I" is representative of a horizon containing multiple conglomerates and is shown in Figure 3.26. This horizon is composed of multiple lenticular conglomerates, sandy conglomerates and pebbly sandstone. Other sedimentary structures associated with this horizon are planar crossbeds, fining and occasionally coarsening-upward sequences, plane-bedded sand, and rare trough crossbeds. This type of conglomerate horizon is interpreted as a compound (longitudinal) bar. The entire bar is radioactive (Figure 3.26) and contains well packed and sorted conglomerates having the highest thorium and uranium values of any conglomerate horizon studied. The lenticular nature, the rapid change in facies of the sediments, and the bedforms within this type of conglomerate horizon and also Unit 3 are comparable to changes in character of quartz-pebble conglomerates of the Witwatersrand deposits (Smith and Minter, 1980). In the Witwatersrand such changes are described as a response to the rapid flow velocity fluctuations which are typical of proximal braided streams. When compared to

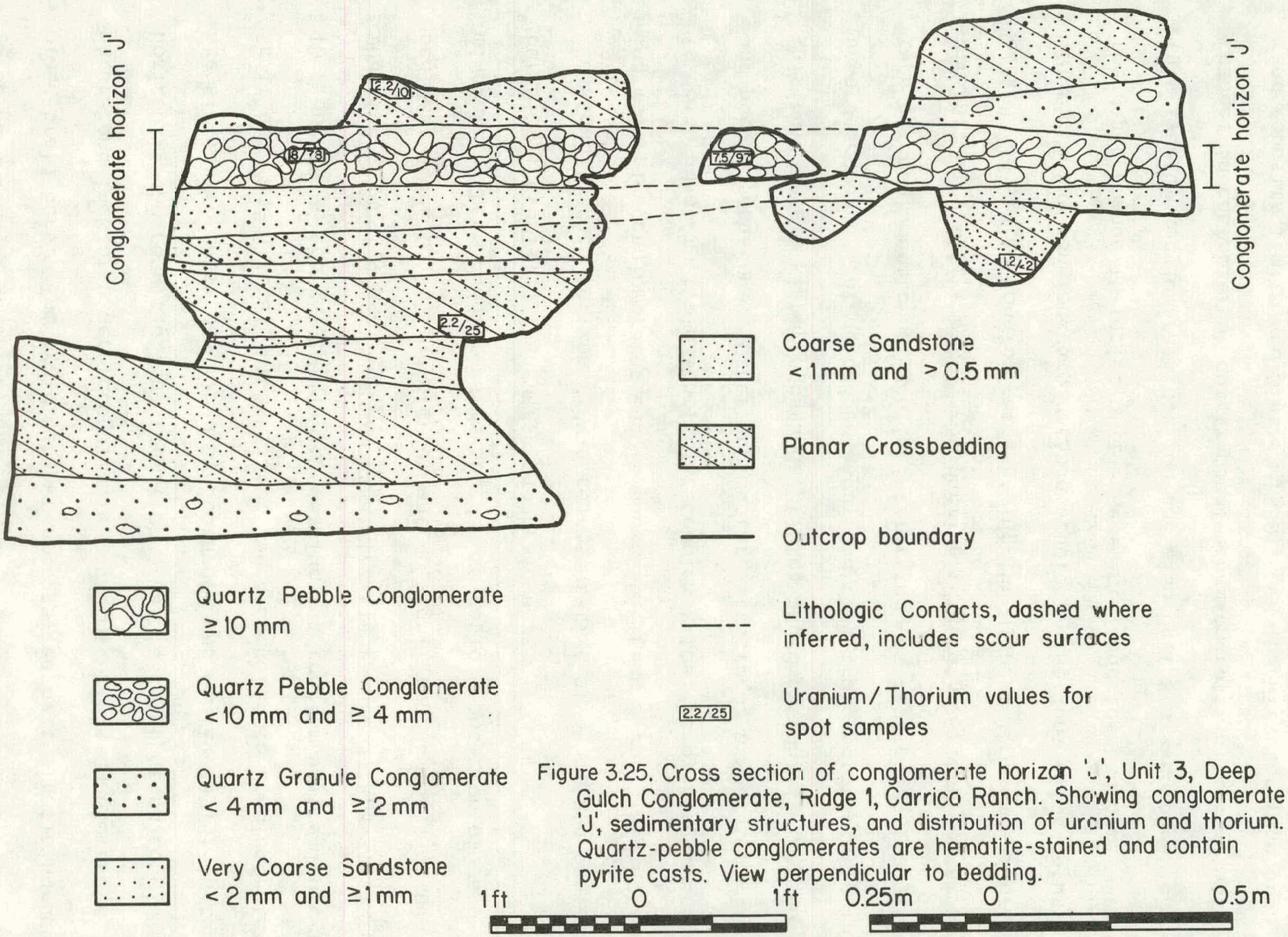


Figure 3.25. Cross section of conglomerate horizon 'J', Unit 3, Deep Gulch Conglomerate, Ridge 1, Carrico Ranch. Showing conglomerate 'J', sedimentary structures, and distribution of uranium and thorium. Quartz-pebble conglomerates are hematite-stained and contain pyrite casts. View perpendicular to bedding.

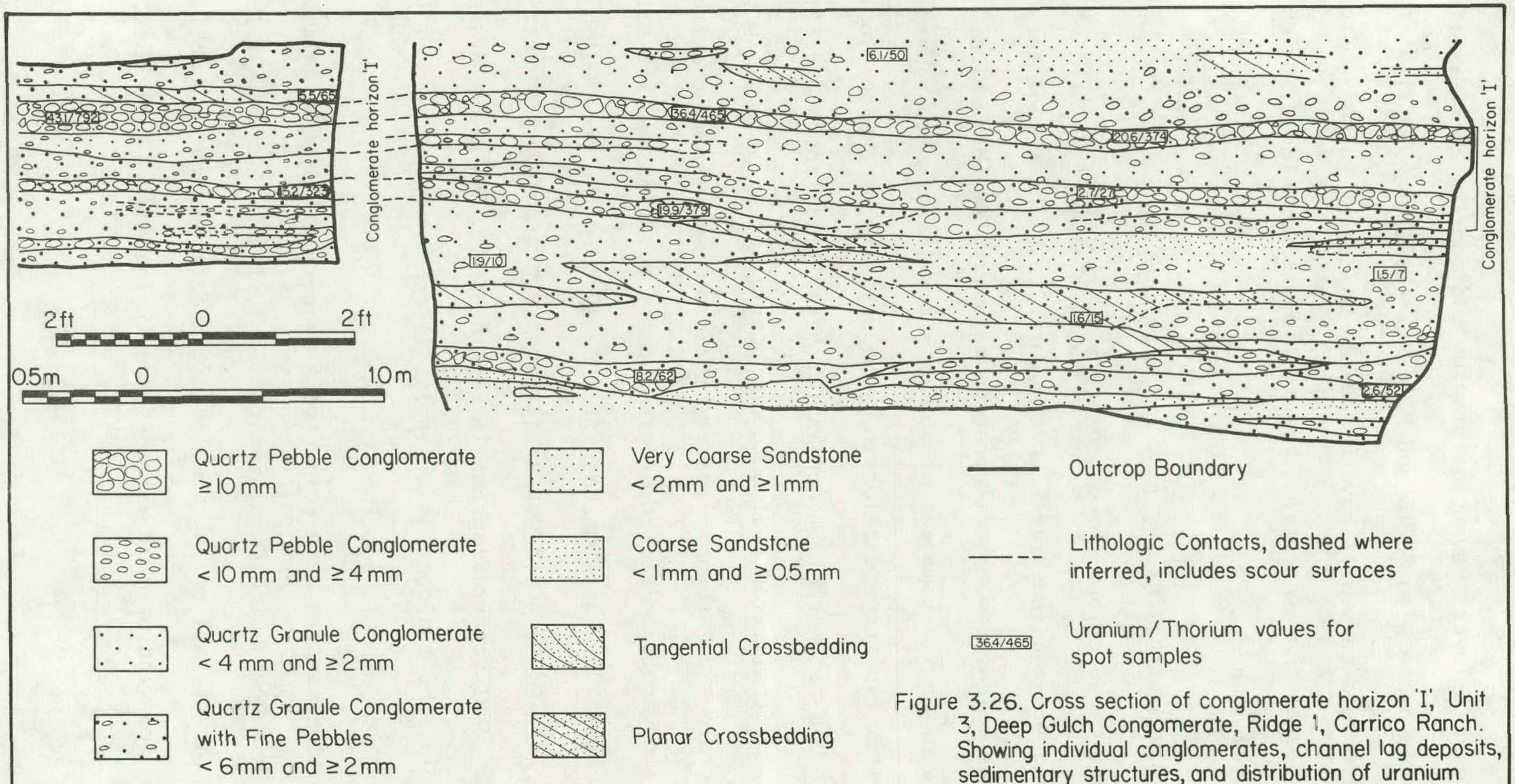


Figure 3.26. Cross section of conglomerate horizon 'I', Unit 3, Deep Gulch Congomerate, Ridge 1, Carrico Ranch. Showing individual conglomerates, channel lag deposits, sedimentary structures, and distribution of uranium and thorium deposits. Quartz-pebble conglomerates are hematite-stained and contain pyrite casts. View perpendicular to bedding.

similar interlayered and underlying conglomerate layers the upper and lower conglomerate layers have high uranium and thorium concentrations. This is in accordance with Smith and Minter's (1980) observation that long-lived compound bars will have a greater chance of being reworked and hence, probably have greater concentrations of uranium, thorium and other heavy minerals.

PETROLOGY AND MINERALOGY OF THE DEEP GULCH CONGLOMERATE

This section describes the petrology and mineralogy of the Deep Gulch Conglomerate and lower Jack Creek Quartzite. These sediments have undergone epidote-amphibolite facies metamorphism (Miyashiro, 1973) and are primarily composed of quartz and feldspar with recyrstallized muscovite, biotite and chlorite. The majority of the phyllosilicate minerals are metamorphic minerals recrystallized from an argillaceous matrix. Overall, the sediments are poorly sorted and immature; the average grain size is approximately 1.25 mm. We believe that the sand-size grains are detrital in origin and are not a product of metamorphism. Petrographic data on each of the units in the Deep Gulch Conglomerate and lower Jack Creek Quartzite are presented on quartz-feldspar-mica (QFM) tenary diagrams in Figure 3.27. The heavy mineral suite, including the radioactive minerals, is considered in detail in the latter portion of this section.

Thin section samples were collected from each drill hole which intersected the Deep Gulch Conglomerate (SM-1, 1A, 2, 2D, and 3; JP-1, 2, 3 and 4). Nearly all of the samples are quartz-granule or pebble conglomerates from the base of well defined fining-upward sequences.

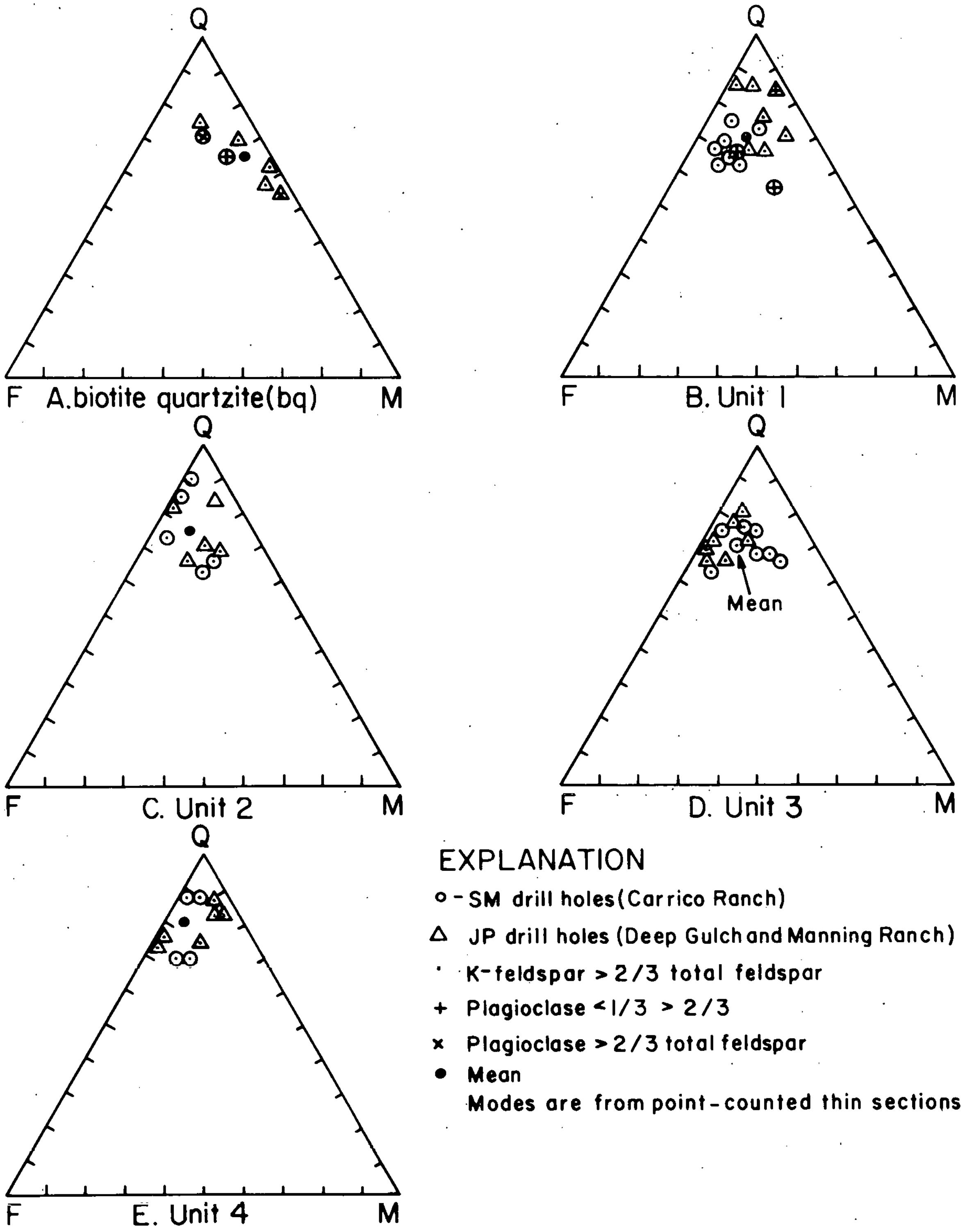


Figure 3.27. Quartz (Q), feldspar (F), and mica + chlorite (M) ternary diagrams for subsurface samples from the Deep Gulch Conglomerate and lower Jack Creek Quartzite (after Folk, 1974).

Five varieties of quartz are common in Unit 3 and are grouped together under the quartz classification in Figure 3.27. The varieties are: undulose quartz, nonundulose quartz, recrystallized quartz clasts with up to a dozen distinct crystals, quartz cement (when it could be identified), and quartzite (fine-grained) rock fragments. Some quartz grains are stretched, whereas feldspar grains are not as stretched. Sutured quartz grain boundaries are common.

Five common varieties of feldspar are recognized in the samples: microcline, perthite, antiperthite (perthites are of the interlocking and replacement variety, Deer and others, 1976), orthoclase, and plagioclase. The feldspar grains are generally angular and commonly exhibit deformed Carlsbad, albite or polysynthetic twinning. Alteration of the feldspar is minimal although some grains are cloudy and some grains are partly altered to sericite. Plagioclase constitutes a small portion of the feldspar present and the grains are angular and generally less than 0.5 mm in diameter. Plagioclase composition varies between An_5 (albite) and An_{33} (oligoclase-andesine). Some clear fine-grained albite and microcline grains may be metamorphic minerals. K-feldspar overgrowths on grains and vein fillings are present in a few samples. Granite rock fragments are rare in the Deep Gulch Conglomerate sediments.

Mica, along with plagioclase (except in Unit bq), is concentrated in the matrix and is either muscovite, biotite or chlorite. Muscovite is the major mica present with biotite and chlorite relatively minor constituents. Pyrite grains are ubiquitous in the matrix of the granule and pebble conglomerates.

Unit bq. Figure 3.27A shows that the composition of the gray biotite quartzite unit cored in drill holes JP-1, JP-3 and SM-2D varies between a very micaceous subarkose (greater than 10% mica and chlorite), and subarkose. Muscovite comprises 84% and biotite 16% of the mica in the unit. Plagioclase An_{20-33} is the dominant feldspar in three of the samples. The gray color, abundant mica and lack of large K-feldspar clasts contrast sharply with Unit 1 (Table 3.13). The source of the detritus is believed to be the underlying biotite plagioclase gneiss, which is composed of sodic plagioclase (An_{15-38}), quartz, biotite, and muscovite.

Unit 1. K-feldspar clasts characterize the sediments of Unit 1. Large K-feldspar pebbles in excess of 10 mm in diameter are common and about 40 percent of the pebbles are coarser than 2 mm. In general, perthite and microcline are the most common varieties of K-feldspar. Plagioclase (An_{15-33}) is restricted to grains which are less than 0.5 mm in diameter, and albite grains are rare. The sorting of the conglomerates in Unit 1 is poor with K-feldspar grains generally larger than the quartz grains, except in the quartz-pebble conglomerates.

Figure 3.27B shows the QFM composition of 16 samples of Unit 1. As shown, the composition varies from an arkose to subarkose. The matrix of the conglomerates is composed of quartz (medium- to fine-grained), muscovite, and feldspar, with plagioclase comprising 31% of the matrix feldspars and only 18% of the total feldspar present. Muscovite accounts for 64% of the total mica content with biotite comprising the remaining 36%.

Unit 2. The ternary QFM diagram for Unit 2 (Figure 3.27C) shows that the percentage of feldspar has decreased from Unit 1, and Unit 2 quartzites vary in composition between quartzarenite and subarkose. The size of the K-feldspar grains has also decreased from those in Unit 1. Most of the K-feldspar clasts are less than 2 mm (granule), although larger clasts are not uncommon, especially in lag gravels. The percentage of plagioclase has decreased from 18% in Unit 1 to 13% in Unit 2. As in Units 1 and 3, polycrystalline quartz clasts are common, but these are not interpreted as quartzites. The percentage of mica is quite variable, from 1 to 19% and muscovite is the major mica. Biotite has decreased to 15% of the mica in Unit 2.

Unit 3. The composition of Unit 3 ranges from subarkose to arkose (Figure 3.27D and Table 3.14). Although the amount of plagioclase in the matrix of the quartz-granule and pebble conglomerates has decreased to a maximum of 2 percent (Table 3.14), the total amount of feldspar has not changed significantly because the amount of K-feldspar has increased. However, the size of K-feldspar grains has decreased. The samples selected from Unit 3 are not the most radioactive of the unit, but are approximately 5 to 7 times the background radiation.

Quartzite rock fragments are more common in Unit 3 than in the other units. Black and green chert fragments are also observed in Unit 3. The amount of quartzite increases as the percentage of pyrite increases. Quartz is the most common mineral present in the matrix with K-feldspar, muscovite, pyrite, plagioclase and other heavy minerals occurring in decreasing abundance.

TABLE 3.14 PETROGRAPHY OF UNIT 3, DEEP GULCH CONGLOMERATE. PERCENTAGES ARE FROM POINT-COUNTED THIN SECTIONS OF QUARTZ-PEBBLE CONGLOMERATES.

Drill Hole Depth	Qtz.	Qtzite	Plag.	Perthite	Micro-cline	Ortho-clase	Musc.	Chlor.	Biot.	Opaq.	Zircon
SM-1 175.4	61	6	—	1	7	3	23	—	—	1	—
SM-1 175.7	45	24	—	—	10	3	10	—	—	7	1
SM-1 192.8	67	1	—	9	3	—	17	—	—	2	—
SM-1 207	66	6	Tr	8	—	11	7	—	—	Tr	Tr
SM-1A 541.6	60	13	Tr	9	3	8	6	—	—	1	—
SM-2 149	53	13	2	7	8	12	6	—	—	Tr	Tr
SM-2 150.1	57	10	Tr	Tr	6	6	17	—	—	4	Tr
SM-2 187	63	14	1	10	Tr	3	Tr	9	Tr	Tr	Tr
JP-1 408.7	70	2	Tr	2	21	—	2	—	—	2	—
JP-1 410	56	25	—	—	13	—	3	—	—	5	Tr
JP-1 412	61	15	—	Tr	19	—	5	—	—	1	—
JP-1 416.5	62	6	1	—	23	—	4	—	—	2	Tr
JP-1 421	71	1	1	1	19	—	4	—	—	3	Tr
JP-2 311	61	4	1	Tr	11	9	9	—	—	3	1
JP-4 284	62	9	1	Tr	9	9	11	—	Tr	2	Tr

Opaque minerals, primarily pyrite, constitute up to 7 percent of the samples that were examined in thin section. In rock slabs and polished thin sections opaque minerals constitute up to 30 percent of the rock (Figures 3.28 and 3.29). Pyrite grain morphology varies greatly, from euhedral to subrounded equant and elongate grains.

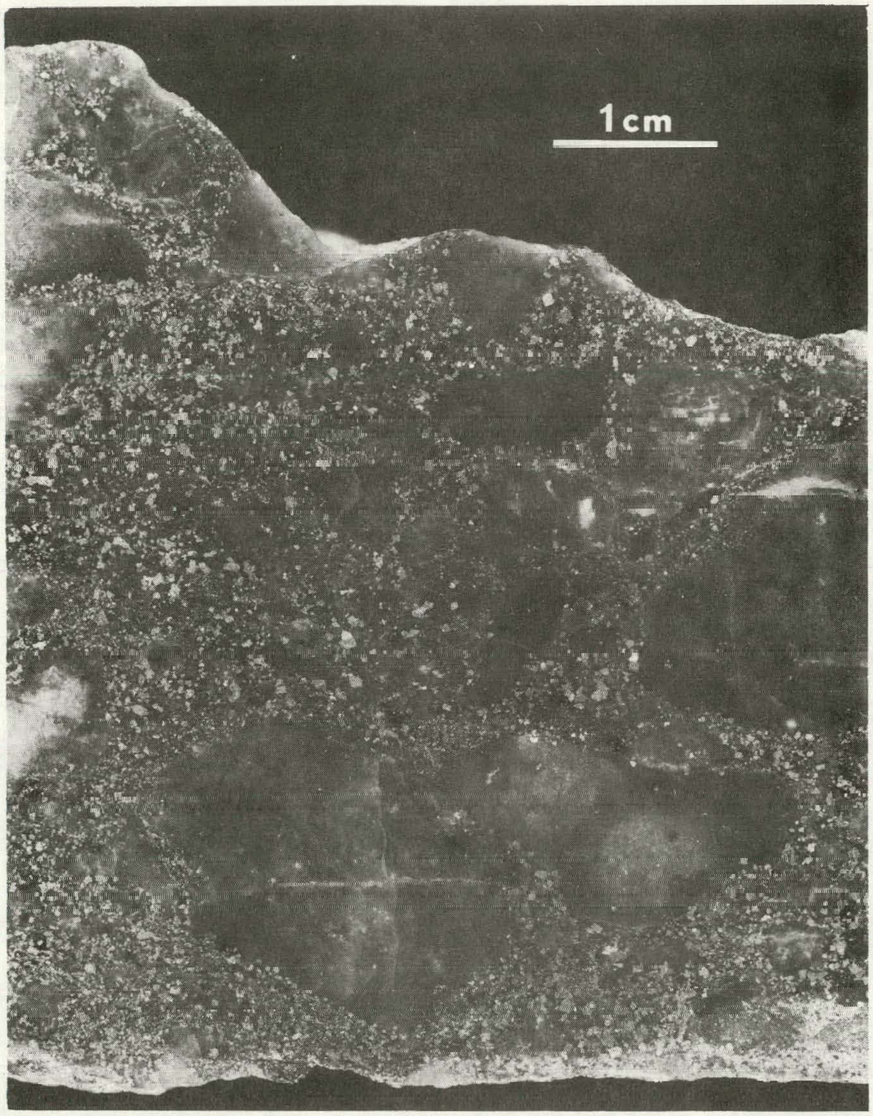


Figure 3.28. Photograph of basal quartz-pebble conglomerate from drill hole SM-2, 165.5 ft. The small bright euhedral to subhedral pyrite grains are concentrated in the matrix. The large euhedral pyrite have probably been recrystallized. The large gray and white clasts are quartz and quartzite pebbles. This conglomerate is correlated with conglomerate K on the surface and has the highest thorium assay of any sample in the Deep Gulch Conglomerate (2600 ppm Th, 220 ppm U; sample number 158460). Maximum pebble size this conglomerate is 35 mm and mean size of the pyrite grains is 0.44 mm.

Aggregates of pyrite grains are also present (Figures 1.14 and 1.17). Pyrite grain size varies from 0.01 to 2.0 mm in diameter. Rare veinlets of pyrite up to 2 mm long, were observed between pyrite grains, and pyrite grains surrounding the veinlets have been recrystallized to euhedral habit.

The heavy mineral suite from the radioactive quartz-pebble conglomerates in Unit 3 consists of radioactive and non-radioactive minerals. Table 3.15 is a list of these minerals. Pyrite is the most common heavy mineral associated with the suite. The majority of heavy minerals are concentrated at the base of clast-supported conglomerates with some heavy minerals also concentrated at the base of matrix-supported conglomerates. Radioactive minerals include zircon, the most abundant mineral and lesser amounts of monazite, monazite-huttonite(?), and huttonite(?). The shape of zircon, monazite, monazite-huttonite(?), and huttonite(?) minerals is subrounded. This grain morphology may be either a metamorphic texture due to recrystallization or a feature of the detrital origin of these minerals.

TABLE 3.15 HEAVY MINERAL SUITE FROM UNIT 3, DEEP GULCH CONGLOMERATE.

pyrite	non-radioactive minerals
ilmenorutile	
apatite	
marcasite	
sphene	
ilmenite	
anatase	
rutile	
garnet	
zircon	radioactive minerals
monazite	
monazite-huttonite(?)	
huttonite(?)	

Figure 3.29 Photograph of a horizontally bedded quartz-pebble conglomerate from drill hole SM-1, 190.7 ft. Pyrite grains are concentrated in dark bands outlining the bedded planes. Individual pyrite grains are bright and reflective while the quartz grains are grey and feldspar grains are white. Quartz and pyrite grains are not as large as those in Figure 17. Maximum size of the quartz grains is 18 mm and mean size of the pyrite grains is 0.29 mm. Assay values for this sample are 170 pp in Th and 57 ppm U (sample number 711461).

GEOCHEMISTRY OF THE DEEP GULCH CONGLOMERATE

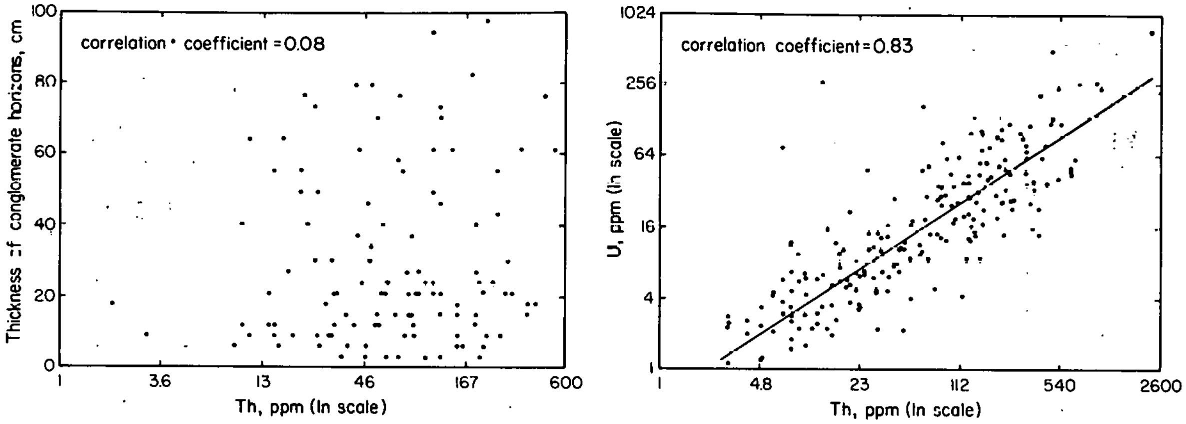
Introduction

This section contains two parts: the first discusses the distribution of the radioactive elements uranium and thorium in the context of fluvial fossil-placer accumulations; the other presents the correlation matrices of outcrop and subsurface concentrations of eleven selected elements, and the correlation of quartz-pebble and pyrite size to selected elements. Sampling procedures are described in Kratochvil (1981).

Disbribution of Thorium and Uranium Related to Sedimentary Bedforms

When examined in outcrop the mineralization in the Deep Gulch Conglomerate was found to be directly related to the occurrence of sedimentary bedforms. These bedforms and the mineralization form a sympathetic relationship in terms of the depositional energy and sorting processes required to create them. The least mineralized bedforms are those developed in the low flow regime and maximum mineralization occured in bedforms representative of upper flow regime conditions (Harms and Fahnstock, 1965). Thorium is the element which we use to characterize the concentration of the radioactive elements, uranium and thorium, because it has not been significantly leached from the outcrop. The use of thorium for this purpose is reasonable because the correlation between uranium and thorium is 0.85 in outcrop and 0.83 in the subsurface (Table 3.17 and Figure 3.30).

The relationship of conglomerate horizon thickness to mineralization is shown in Figure 3.30. The correlation coefficient of 0.08 indicates that mineralization is unrelated to the thickness of the conglomerate horizon. Figures 3.25 and 3.26 show two conglomerate horizons, I and J at Ridge 1, Carrico Ranch, and the geochemical analyses for thorium and uranium values of selected bedforms. The thorium values range from 25 ppm in planar crossbeds to 792 ppm in the well sorted upper conglomerate of horizon I.

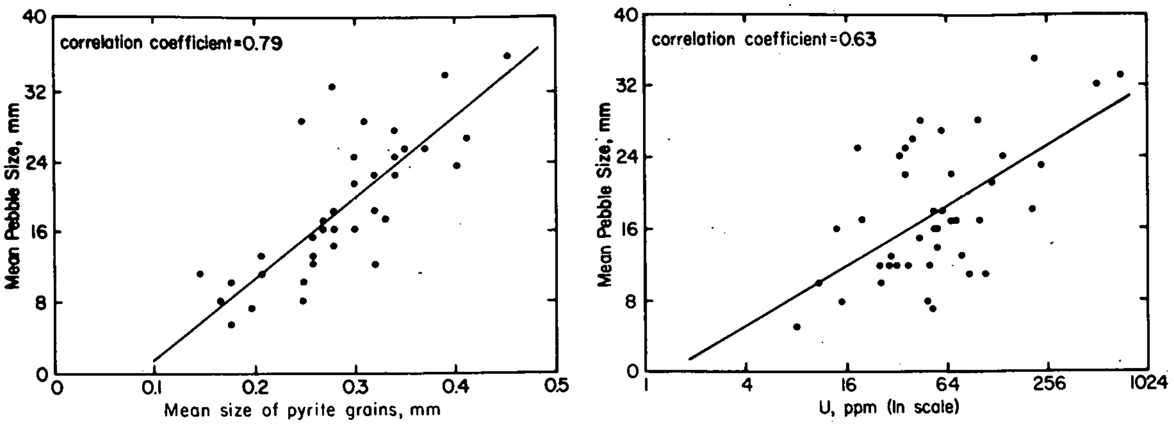


A. Thickness of outcrop conglomerate horizons plotted against Th.

B. U plotted against Th.

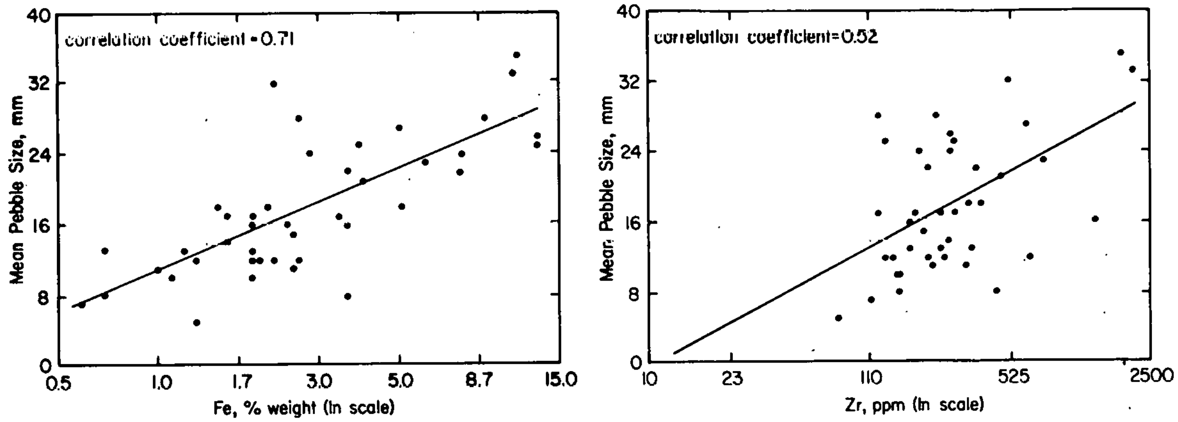
Figure 3.30. Thickness of outcrop conglomerate horizons and U concentrations plotted against Th concentrations from the Deep Gulch Conglomerate.

Planar crossbeds have a maximum value of 25 ppm thorium in layers of maximum grain size on a foreset slope. The base of trough crossbeds have a maximum thorium content of 138 ppm, and lag gravels which separate and define channel migrations in the trough crossbedded sequence of Unit 2, have a maximum value of 256 ppm thorium. Poorly sorted pebbly granular conglomerates, that may represent braid bar or channel accretion, have a maximum value of 50 ppm, which may be a low value because the base of the conglomerate was not sampled.



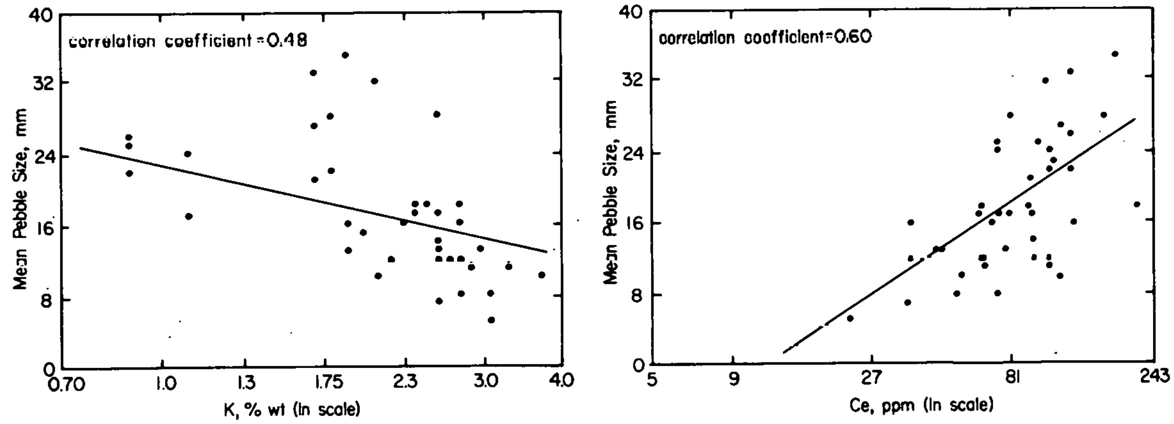
A. Mean size of pyrite grains plotted against mean pebble size.

B. U plotted against mean pebble size.



C. Fe plotted against mean pebble size.

D. Zr plotted against mean pebble size.



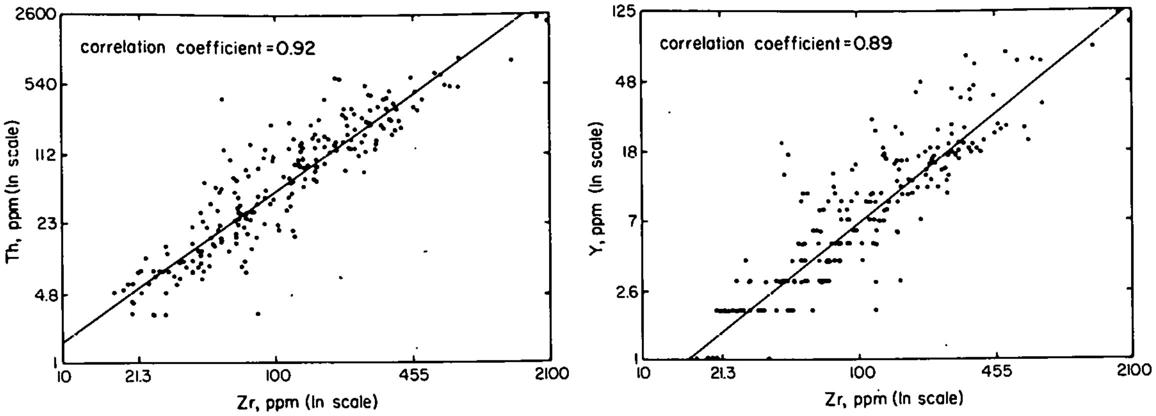
E. K plotted against mean pebble size.

F. Ce plotted against mean pebble size.

Figure 3.31. Mean size of pyrite grains, U, Fe, Zr, K, Ce concentrations plotted against mean pebble size from the Deep Gulch Conglomerate.

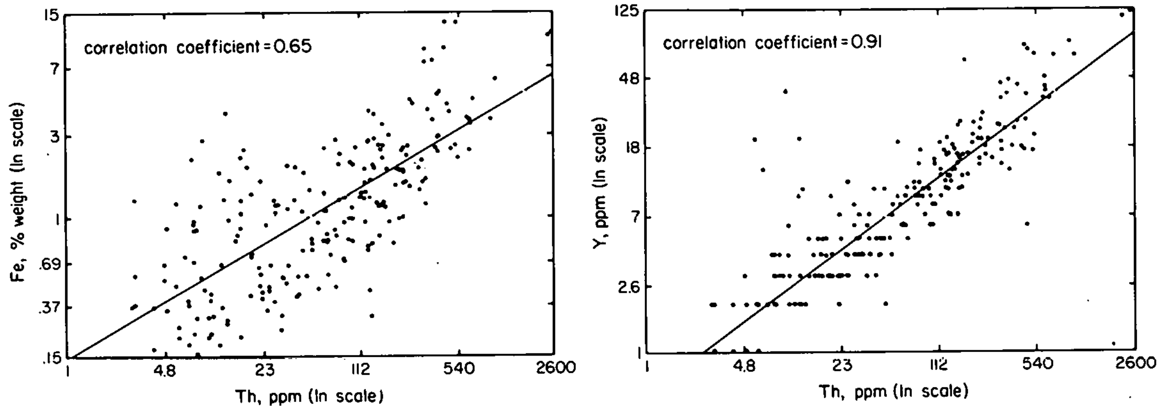
The single conglomerate horizon represented by conglomerate horizon I in Figure 3.25 has a maximum concentration of 97 ppm thorium. This horizon is a moderately sorted clast-supported conglomerate. The 97 ppm thorium value is from the base of the conglomerate, and as such should be representative of the highest value present. Horizon I (Figure 3.26) is interpreted as compound bars which were often reworked by currents, increasing the probability of heavy mineral entrapment in the coarse bed materials (Smith and Minter, 1980). The uppermost and lowermost conglomerates in horizon I contain the highest thorium values of any of the bedforms sampled. Maximum thorium concentration of 792 ppm is present in the upper conglomerate of horizon I. This horizon is the thickest, most laterally continuous, and has the maximum grain size of any conglomerate in horizon I. Conglomerate horizon K (Figure 3.23), the thickest, coarest and most laterally continuous conglomerate horizon of the Deep Gulch Conglomerate (Plate 7), has a maximum thorium value of 1100 ppm at the base of the conglomerate.

The above hierarchy of bedforms as related to the thorium content is no doubt reflective of the depositional environment and hydraulic conditions under which they were deposited and is similar to that described in other fluvial fossil placer deposits (Smith and Minter, 1980; Pieneer, 1963; Roscoe, 1969). Quasi-permanent bedforms in the fluvial system which have undergone reworking of sediments by currents concommitant with the addition of new sediments offer the greatest potential for mineralization (Smith and Minter, 1980). In the Deep Gulch Conglomerate, the conglomerate horizons of Unit 3, interpreted to be compound (longitudinal) bars were such quasi-permanent bedforms in the fluvial system which deposited the Deep Gulch Conglomerate.



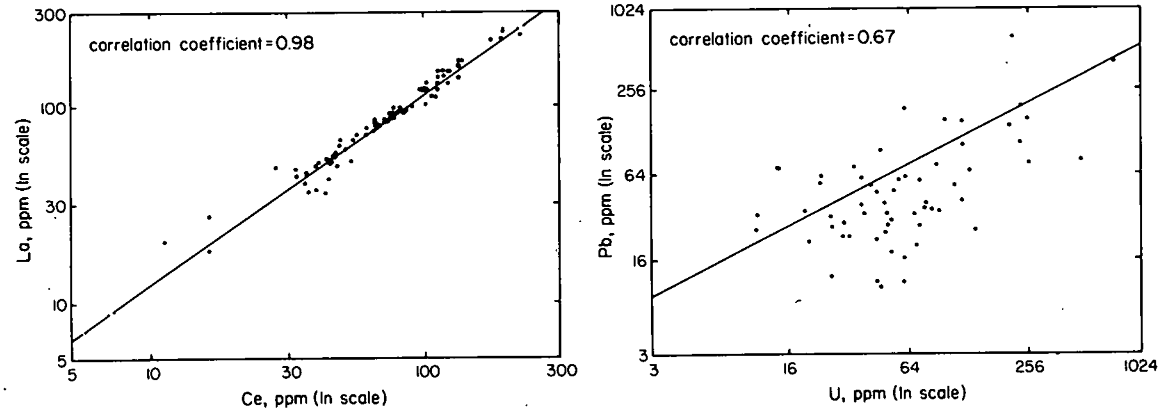
A. Zr plotted against Th.

B. Zr plotted against Y.



C. Th plotted against Fe.

D. Th plotted against Y.



E. Ce plotted against La.

F. U plotted against Pb.

Figure 3.32. Zr, Th, Ce and U concentrations plotted against Th, Y, Fe, La and Pb concentrations from the Deep Gulch Conglomerate.

Pebble Size Compared with Concentrations of Selected Elements

In the Blind River-Elliot Lake area of Canada, Theis (1979) has demonstrated that there is a relationship between quartz-pebble size, mineralogy, and chemistry of ore-bearing conglomerate. Theis (1979) shows that the high specific gravity mineral, uraninite, is most abundant in the maximum quartz-pebble size conglomerates, whereas, the lower specific gravity minerals, zircon and monazite, are enriched in conglomerates of smaller quartz-pebble size. This feature was thought to be related to hydraulic factors (Theis, 1979, p. 10) and was viewed as evidence that uraninite, zircon and monazite are detrital minerals. This relationship of grain-size to mineralogy is reflected in the chemistry of the Blind River-Elliot Lake Conglomerates. For example, the amount of uranium (reflecting uraninite) has a positive correlation with pebble size whereas the amount of cerium (reflecting monazite) is inversely related to pebble size (Theis, 1979, p. 11).

Table 3.16 and Figure 3.31 show plots and correlation coefficients of quartz-pebble size pyrite grain size, and the concentration of selected elements. Pyrite grain size has a sympathetic moderate correlation with the size of quartz pebbles suggesting that hydraulic factors played a part in the deposition of pyrite grains in the conglomerates. This relationship is also reflected in the percentage of iron increasing with increasing quartz-pebble size. Plots of other element concentrations versus quartz-pebble size do not show strong correlation. Zirconium, cerium, and uranium concentrations increase with increasing quartz-pebble size although correlation coefficients are moderate at best (Figure 3.31). The relationship between zirconium, cerium and uranium with quartz-pebble size is not the same as that present in the Blind River-Elliot Lake rocks where zirconium and cerium both show an inverse relationship to quartz-pebble size (Theis, 1979, p. 11-12). We suspect this difference is related to the hydraulic conditions at the time of deposition and mineralogy of the two deposits. Potassium shows an inverse relationship with pebble size similar to that found in the Canadian conglomerates (Theis, 1979, p.13). This probably reflects an increase in the feldspar and phyllosilicate content of the sediment in the quartz-pebble size decreases (Theis, 1979, p. 13).

TABLE 3.16 MATRIX OF PEARSON CORRELATION COEFFICIENTS FOR SUBSURFACE SAMPLES, PEBBLE, AND PYRITE SIZES OF THE DEEP GULCH CONGLOMERATE. 44 CHEMICAL ANALYSES, PYRITE, AND PEBBLE SIZES OF THE MOST RADIOACTIVE QUARTZ-PEBBLE CONGLOMERATES WERE USED TO CALCULATE THE CORRELATION COEFFICIENTS. CORRELATION COEFFICIENTS FOR ELEMENTS WERE COMPUTED USING LOG TRANSFORMATIONS.

		Pyrite													
	12%	25%	Mean	Mean Pebble Size	U(nt)	Th	U/Th	Fe	Ti	Y	Zr	K	Ce	La*	Pb*
Coarsest 12% of Pyrite size	1.00	.96	.91	.77	.45	.50	−.22	.63	−.19	.45	.46	−.42	.51	.36	.33
Coarsest 25% of Pyrite size	.96	1.00	.95	.78	.47	.56	−.25	.66	−.27	.47	.52	−.49	.49	.34	.31
Mean Pyrite size	.91	.95	1.00	.80	.39	.56	−.26	.73	−.33	.43	.47	−.55	.51	.35	.29
Mean Pebble size	.77	.78	.80	1.00	.60	.64	−.08	.70	−.34	.55	.50	−.53	.63	.45	.51
U(nt)	.45	.47	.39	.60	1.00	.64	.33	.33	−.10	.76	.66	−.22	.47	.29	.50
Th	.50	.56	.59	.64	.64	1.00	−.39	.69	−.06	.86	.87	−.29	.53	.23	.37
U/Th	−.22	−.25	−.26	−.08	.33	−.39	1.00	−.35	−.09	−.16	−.25	.01	−.04	.11	.15
Fe	.63	.66	.73	.70	.33	.69	−.35	1.00	−.26	.44	.48	−.56	.59	.50	.42
Ti	−.19	−.27	−.33	−.34	−.10	−.06	−.09	−.26	1.00	.14	.12	.82	.03	.06	.04
Y	.45	.47	.43	.55	.76	.86	−.16	.44	.14	1.00	.82	−.05	.45	.21	.35
Zr	.46	.52	.47	.50	.66	.87	−.25	.48	.12	.82	1.00	−.08	.55	.31	.36
K	−.42	−.49	−.55	−.53	−.22	−.29	.01	−.56	.82	−.05	−.08	1.00	−.24	−.26	−.13
Ce	.51	.49	.51	.63	.47	.53	−.04	.59	.03	.45	.55	−.24	1.00	.80	.46
La*	.36	.34	.35	.45	.29	.23	.11	.50	.06	.21	.31	−.26	.80	1.00	.45
Pb*	.33	.31	.29	.51	.50	.37	.15	.42	.04	.35	.36	−.13	.46	.45	1.00

Correlation Coefficients of Selected Elements

The relationship between certain elements is very much as expected from the mineralogy of the conglomerates and is similar to that present at Blind River-Elliot Lake in Canada. For example cerium and lanthanum have very high correlation coefficients because both elements are primarily in the mineral monazite (Table 3.17and Figure 3.32). Lead and uranium correlate as might be expected if lead is largely a decay product of uranium but not as well as the two elements correlate in Canada (Theis, 1979).

TABLE 3.17. MATRIX OF PEARSON CORRELATION COEFFICIENTS FROM THE DEEP GULCH CONGLOMERATE. CORRELATION COEFFICIENTS FOR ELEMENTS WERE COMPUTED USING LOG TRANSFORMATIONS BASED ON DATA FROM 220 SUBSURFACE AND 200 OUTCROP SAMPLES.

Subsurface Samples

	U(nt)	Th	Th/U	Fe	Ti	Y	Zr	K	Ce	La*	Pb*	P
U(nt)		.83	−.05	.65	.02	.85	.81	−.16	.70	.53	.55	.38
Th	.85		.42	.73	.04	.91	.92	−.19	.72	.48	.47	.46
Th/U	.12	.56		.24	.00	.26	.36	−.14	.16	.17	.02	.16
Fe	.72	.65	.17		.11	.74	.71	−.35	.62	.51	.55	.57
Ti	.42	.36	.19	.39		.18	.15	.62	.17	.09	−.01	.47
Y	.86	.83	.30	.65	.40		.89	−.13	.73	.49	.48	.55
Zr	.85	.82	.31	.68	.38	.89		−.17	.73	.57	.49	.45
K	.17	.12	.09	.08	.72	.06	.10		.03	.03	−.20	−.08
Ce	.50	.42	.13	.41	.21	.61	.56	−.09		.98	.40	.46
La*	.51	.56	.50	.37	−.14	.48	.44	−.27	.94		.39	.47
Pb*	.64	−.27	.37	.63	−.39	.33	.48	.03	.40	.64		.40
P	.63	.59	.19	.75	.40	.60	.61	−.23	.46	.44	.54	

Outcrop Samples

*Approximately 80 analyses were available for La and Pb correlations.

On the other hand, the moderate to strong correlations between certain elements is initially surprising. For example zirconium correlates fairly well with uranium, thorium, yttrium and cerium, and thorium correlates with uranium, iron, yttrium, zirconium and cerium

(Table 3.17 and Figures 3.30 and 3.32). The source of the zirconium should be almost entirely from zircon, and the fact that the best correlations of zirconium are with thorium, yttrium and uranium suggest that zircon contains these elements, or it is possible that zircon is distributed in the same manner in the conglomerates as other minerals which contain uranium, thorium, and yttrium. Thorium and uranium are probably present in a number of minerals and thus correlate with several elements.

The radioactive heavy minerals in the Deep Gulch Conglomerate deposits (Table 3.15) all have about equal specific gravity (4 to 5) and all are roughly correlated with the pebble-size. Therefore when pebble size increases most of the elements found exclusively in the heavy minerals (uranium, thorium, iron, yttrium, zirconium, and cerium) also increase in concentration. Under these circumstances a moderate correlation might be expected for such elements as zirconium, thorium, yttrium, and uranium.

Our interpretations are prejudiced toward a placer origin for uranium and thorium. We must note that the correlations between elements such as zirconium and uranium might also be used as evidence for other than a placer origin for the uranium and thorium.

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EXPLANATION

☐	PHANEROZOIC ROCKS UNDIVIDED

UNCONFORMITY OR THRUST FAULT CONTACT

sg	SHERMAN GRANITE: pink coarse-grained granite, 1400 m.y.

INTRUSIVE CONTACT

m	MAFIC INTRUSIVE ROCKS: metagabbro, local metapyroxenite; probably several ages present (Archean to Middle Proterozoic); small dikes and sills are black

EARLY PROTEROZOIC 2500-1700 m.y.

<1900 m.y.

ROCKS SOUTH OF MULLEN CREEK-NASH FORK SHEAR ZONE

INTRUSIVE CONTACT

cmg	CATACLASTIC MIGMATITE: augen gneiss, gray layered gneiss, local mylonite
cag	CATACLASTIC BIOTITE AUGEN GNEISS
gn	QUARTZ-MONZONITE GNEISS: probably orthogneiss
hgn	HORNBLENDE PARAGNEISS: metabasalt, metagraywacke, metatuff, calc-silicate

MULLEN CREEK-NASH FORK SHEAR ZONE CONTACT

LIBBY CREEK GROUP

fs	FRENCH SLATE: black phyllite and slate, local iron formation near top
tg	TOWNER GREENSTONE: green chlorite schist and chlorite amphibole schist
nf	NASH FORK FORMATION: metadolomite, stromatolitic metadolomite, black slate

THRUST FAULT CONTACT

gt	GAPS TRONDHJEMITE: sheared and brecciated metatrondhjemite

INTRUSIVE CONTACT

slq	SUGARLOAF QUARTZITE: massive quartzarenite with planar crossbeds, ripple marks
ls	LOOKOUT SCHIST: laminated schist with local iron formation, clastic dikes
mpq	MEDICINE PEAK QUARTZITE: planar crossbedded, kyanite-bearing quartzarenite
h	HEART FORMATION: arkosic quartzite and chlorite biotite phyllite
hd	HEADQUARTERS FORMATION: diamictite, arkosic quartzite, chlorite biotite quartz schist

UNCONFORMITY

rk	ROCK KNOLL FORMATION: quartzite, phyllite, quartzite-pebble conglomerate

THRUST FAULT CONTACT

DEEP LAKE GROUP

v	VAGNER FORMATION: diamictite, marble, biotite quartz schist, quartzite

UNCONFORMITY

cq	CASCADE QUARTZITE: quartzarenite with quartz and black chert pebbles, planar and trough crossbedded
cl	CAMPBELL LAKE FORMATION: diamictite, chlorite biotite quartz schist

UNCONFORMITY

lq	LINDSEY QUARTZITE: trough crossbedded quartzite with phyllite partings
mg	MAGNOLIA FORMATION: radioactive quartz-pebble conglomerate, arkosic paraconglomerate, very coarse-grained quartzite

ANGULAR UNCONFORMITY

ARCHEAN >2500 m.y.

g	GRANITIC ORTHOGNEISS: foliated granite, granodiorite and tonalite

INTRUSIVE CONTACT

PHANTOM LAKE METAMORPHIC SUITE

5 mb	UNIT 5: Conical Peak Quartzite: fine-grained muscovitic quartzite, local *metabasalt* (mb)
4 q	UNIT 4: Colberg Metavolcanic Rocks: metabasalt, metatuff, metagraywacke, paraconglomerate (circles), quartzite (q)
3	UNIT 3: Bow River Quartzite: fine-grained muscovitic quartzite, planar crossbeds
2	UNIT 2: Rock Mountain Conglomerate: arkosic paraconglomerate, coarse-grained quartzite
1 q	UNIT 1: Stud Creek Volcaniclastic Rocks: metagraywacke, metatuff, hornblende gneiss, quartzite (q), locally fuchsitic, marble

FAULT CONTACT

scgn	STAGE CROSSING GNEISS: hornblende and biotite paragneiss

CONTACT RELATIONSHIP UNKNOWN

qfgn	QUARTZO-FELDSPATHIC GNEISS possible basement, >2500 m.y.

Contact, long dashed where appoximately located, short dashed where inferred

Reverse fault of Laramide age; long dashed where approximately located; short dashed where projected beneath post-Laramide cover; sawteeth on upper plate; arrows show relative Laramide and post-Laramide horizontal movement

Major reverse fault of Precambrian age; dashed where inferred; arrows show relative horizontal movement; sawteeth on upper plate

Fault, long dashed where approximately located, short dashed where inferred; bar and ball on downthrown side; arrows show relative horizontal movement; open symbols indicate inferred Laramide or post-Laramide movement

Fault, intruded by mafic sill or dike

Fault (shear or breccia) zone

20 20 Plunging anticline; plunging syncline; shows approximate trace of axial plane and plunge of fold axis; dotted where intruded by mafic sill or covered

35 35 Overturned anticline; overturned syncline; shows approximate trace of axial plane, and trend and plunge of fold axis; dotted where intruded by mafic sill or covered

35 Dip and strike of bedding

70 Dip and strike of overturned bedding

Strike of vertical bedding

70 Dip and strike of foliation

Strike of vertical foliation

35 25 Bearing and plunge of minor fold axis; mineral lineation

EMB-4 60° Inclined drill hole; shows name, bearing, plunge, and total length of hole

MB-5 1429 Vertical drill hole; shows name and total depth in feet

GEOLOGIC MAP OF PRECAMBRIAN METASEDIMENTARY ROCKS OF THE MEDICINE BOW MOUNTAINS, WYOMING

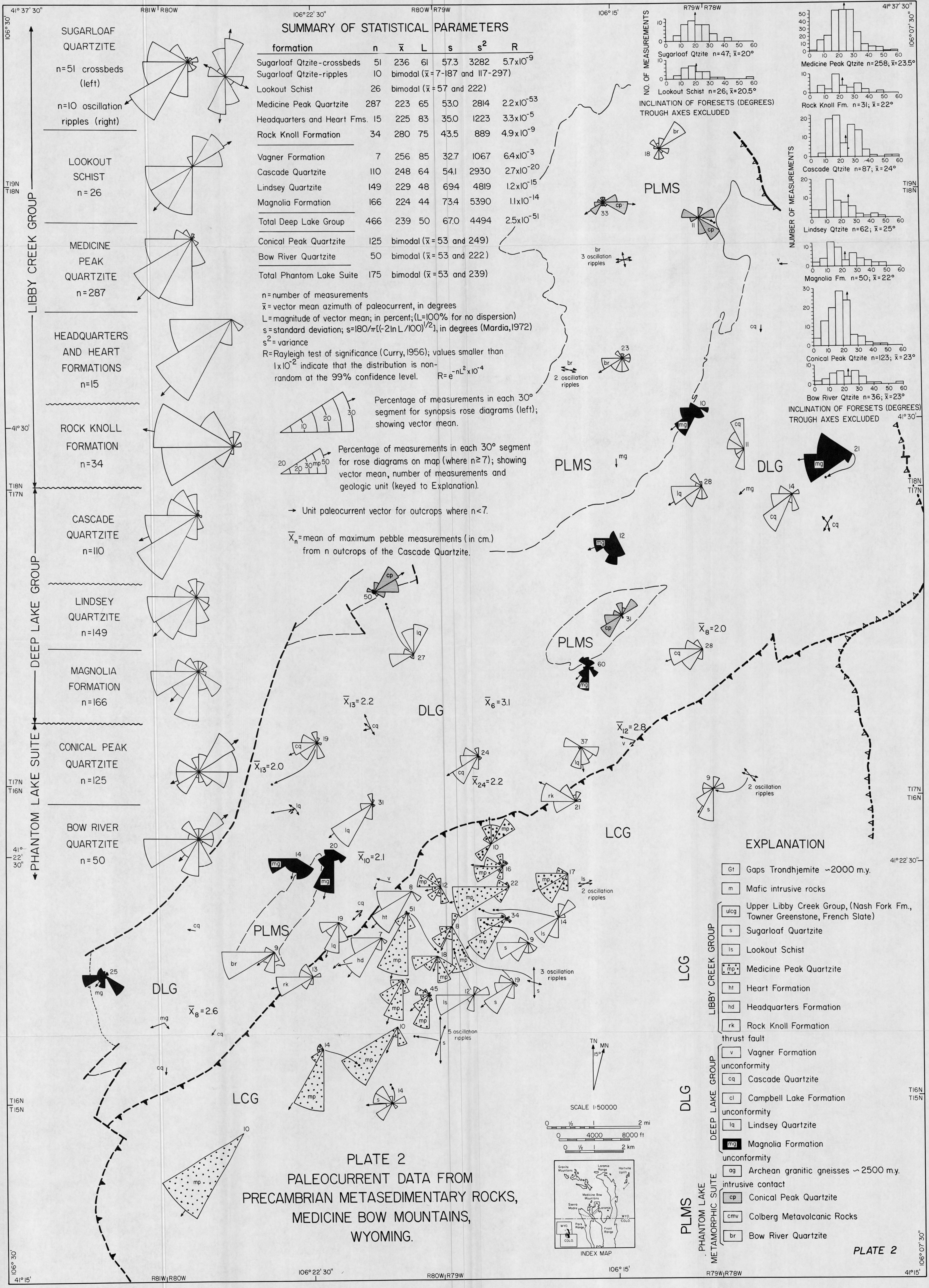
COMPILED BY: R.S. HOUSTON and K.E. KARLSTROM
GEOLOGY BY: K.E. KARLSTROM, R.S. HOUSTON, L.R. LANTHIER and M.E. McCALLUM

SCALE 1:50000

0 1 2 3 4 miles
0 5000 10000 15000 feet
0 1 2 3 4 km

PLATE 1

PLATE 1



SUMMARY OF STATISTICAL PARAMETERS

formation	n	x̄	L	s	s^2	R
Sugarloaf Qtzite-crossbeds	51	236	61	57.3	3282	5.7×10^{-9}
Sugarloaf Qtzite-ripples	10	bimodal (x̄ = 7-187 and 117-297)				
Lookout Schist	26	bimodal (x̄ = 57 and 222)				
Medicine Peak Quartzite	287	223	65	53.0	2814	2.2×10^{-53}
Headquarters and Heart Fms.	15	225	83	35.0	1223	3.3×10^{-5}
Rock Knoll Formation	34	280	75	43.5	889	4.9×10^{-9}
Vagner Formation	7	256	85	32.7	1067	6.4×10^{-3}
Cascade Quartzite	110	248	64	54.1	2930	2.7×10^{-20}
Lindsey Quartzite	149	229	48	69.4	4819	1.2×10^{-15}
Magnolia Formation	166	224	44	73.4	5390	1.1×10^{-14}
Total Deep Lake Group	466	239	50	67.0	4494	2.5×10^{-51}
Conical Peak Quartzite	125	bimodal (x̄ = 53 and 249)				
Bow River Quartzite	50	bimodal (x̄ = 53 and 222)				
Total Phantom Lake Suite	175	bimodal (x̄ = 53 and 239)				

n = number of measurements
x̄ = vector mean azimuth of paleocurrent, in degrees
L = magnitude of vector mean; in percent; (L=100% for no dispersion)
s = standard deviation; $s = 180/\pi[(-2 \ln L/100)^{1/2}]$, in degrees (Mardia, 1972)
s^2 = variance
R = Rayleigh test of significance (Curry, 1956); values smaller than 1×10^{-2} indicate that the distribution is non-random at the 99% confidence level. $R = e^{-nL^2 \times 10^{-4}}$

Percentage of measurements in each 30° segment for synopsis rose diagrams (left); showing vector mean.

Percentage of measurements in each 30° segment for rose diagrams on map (where n≥7); showing vector mean, number of measurements and geologic unit (keyed to Explanation).

→ Unit paleocurrent vector for outcrops where n<7.

$\overline{X}_n$ = mean of maximum pebble measurements (in cm.) from n outcrops of the Cascade Quartzite.

EXPLANATION

Gt	Gaps Trondhjemite ~2000 m.y.
m	Mafic intrusive rocks

LCG — LIBBY CREEK GROUP

ulcg	Upper Libby Creek Group, (Nash Fork Fm., Towner Greenstone, French Slate)
s	Sugarloaf Quartzite
ls	Lookout Schist
mp	Medicine Peak Quartzite
ht	Heart Formation
hd	Headquarters Formation
rk	Rock Knoll Formation

thrust fault

DLG — DEEP LAKE GROUP

v	Vagner Formation

unconformity

cq	Cascade Quartzite
cl	Campbell Lake Formation

unconformity

lq	Lindsey Quartzite
mg	Magnolia Formation

unconformity

ag	Archean granitic gneisses ~2500 m.y.

intrusive contact

PLMS — PHANTOM LAKE METAMORPHIC SUITE

cp	Conical Peak Quartzite
cmv	Colberg Metavolcanic Rocks
br	Bow River Quartzite

PLATE 2
PALEOCURRENT DATA FROM PRECAMBRIAN METASEDIMENTARY ROCKS, MEDICINE BOW MOUNTAINS, WYOMING.

PLATE 2

EXPLANATION

MAP SYMBOLS

ssz	Rocks south of Mullen Creek-Nash Fork shear zone
cgn	Cataclastic gneiss of Mullen Creek-Nash Fork shear zone
ulcg	Upper Libby Creek Group
gt	Gaps Trondhjemite (~2000 m.y.)
llcg	Lower Libby Creek Group
dlg	Deep Lake Group
g	Archean granitic rocks (~2500 m.y.)
plms	Phantom Lake Metamorphic Suite
scgn	Stage Crossing Gneiss
qfgn	Quartzo-feldspathic gneiss (possible basement; >2500 m.y.)

Contact between major rock groups

Subarea boundary within major rock groups

Laramide; Precambrian thrust faults

Laramide fault; shows relative horizontal movement; bar and ball on downthrown side

Precambrian fault; shows relative horizontal movement; bar and ball on downthrown side

Plunging anticline; plunging syncline; dotted where intruded by mafic intrusive rocks

Overturned anticline; overturned syncline; shows bearing and plunge of fold axis

Shear zone

Map view of geometry of structural elements in mesoscopic folds (from field sketches)

STEREONET SYMBOLS

2
$30 \pi S_0$

Lower hemisphere equal area projections of contoured poles to bedding (πS_0) and foliation (S_1) showing subarea number and number of measurements

.5-1%	1-3%	3-5%	5-7%	7-9%	>9%

Concentration of poles per 1% area

Minor fold axis lineation

Mineral lineation

Stretched pebble lineation

Crenulation lineation

Pole to S_3 foliation

Pole to S_4 foliation

Inferred position of girdle of poles to surfaces (S_0 and S_1)

β_2 Inferred macroscopic fold axis defined by girdle of poles to surfaces; subscript shows generation of folding

S_0 Axial plane or average position of S-surface; subscript shows generation of folding

INDEX MAP OF STRUCTURAL SUBAREAS

ROCKS SOUTH OF SHEAR ZONE: SUBAREA 23

SHEAR ZONE ROCKS: SUBAREA 22

UPPER LIBBY CREEK GROUP: SUBAREA 21

GAPS TRONDHJEMITE: SUBAREA 17

LOWER LIBBY CREEK GROUP: SUBAREAS 16,18,19,20

DEEP LAKE GROUP: SUBAREAS 6,7,8,9,11,12,13,15

PHANTOM LAKE METAMORPHIC SUITE: SUBAREAS 1,2,3,4,10,14

STAGE CROSSING GNEISS: PART OF SUBAREA 1

ARCHEAN GNEISSES: SUBAREAS 24,25,26

SCALE 1:50000

0 1 2 3 4 miles

0 5000 10000 15000 feet

0 1 2 3 4 km

TECTONIC MAP OF THE NORTHERN MEDICINE BOW MOUNTAINS, WYOMING

By KARL E. KARLSTROM, 1980

Includes data from: Houston and others (1968); Karlstrom (1977); Karlstrom and Houston (1979b); Lanthier (1978); Wilson (1975); Wilson (1977); Houston and Parker (1963); McCallum (1964).

PLATE 3

PLATE 3

EXPLANATION

MAGNOLIA FORMATION

- m MAFIC INTRUSIVE ROCK
- 5 UNIT 5: Muscovitic subarkose and quartz—pebble conglomerate principal radioactive zones labeled 5a and 5b.
- 4 UNIT 4: Biotite chlorite schist with paraconglomerate and quartzite lenses.
- 3 UNIT 3: Granular subarkose with thin radioactive conglomerate beds.
- 2 UNIT 2: Trough crossbedded subarkose with thin radioactive conglomerate beds.
- 1 UNIT 1: Arkosic paraconglomerate and subarkose.

UNCONFORMITY

- pl PHANTOM LAKE SUITE
- sc STAGE CROSSING GNEISS

- ▤ phyllite or schist with paraconglomerate lenses
- ▨ metagraywacke
- ▨ quartzite
- ▨ quartz-granule conglomerate
- ▨ quartz-pebble conglomerate
- ▨ paraconglomerate
- ～ unconformable contact
- ═ fault or intrusive contact
- ‐‐‐ gradational contact
- ◡ trough crossbeds
- ☰ laminated with small trough crossbeds

Histogram showing average percent quartz, feldspar, mica and opaques for unit or subunit. Percents represent entirety of poorly sorted rocks; granule and subgranule matrix for bimodal conglomerates

5a
P/B
(N)
Q F M O
58 12 17 3

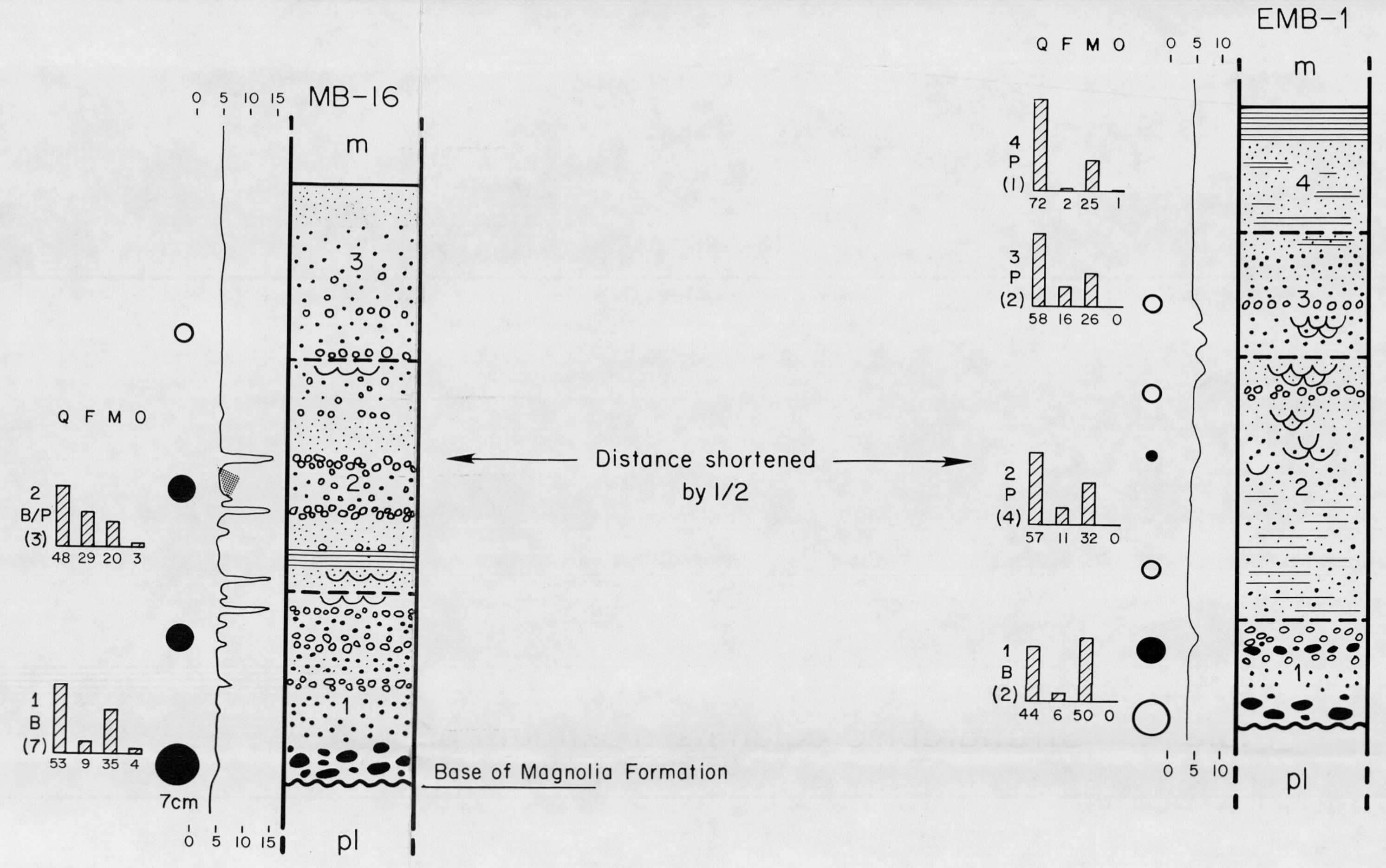
percent for particular column
number of samples averaged
P=poorly sorted B=bimodally sorted
P/B=mostly poorly B/P=mostly bimodal

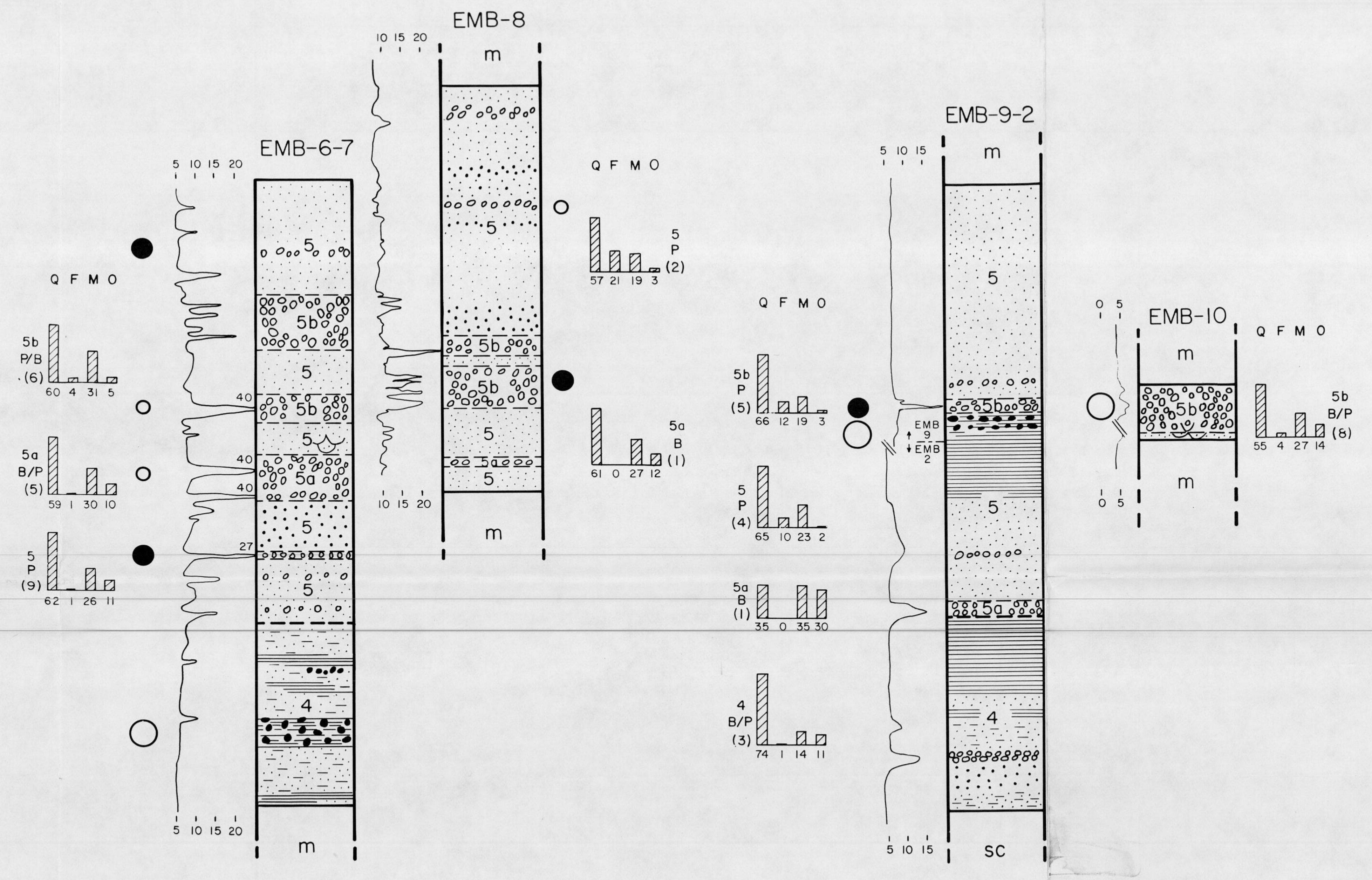
Q = quartz and quartzite
F = k-feldspar, plagioclase, and granite
M = muscovite, sericite, chlorite, and biotite
O = all opaques, primarily pyrite and hematite.

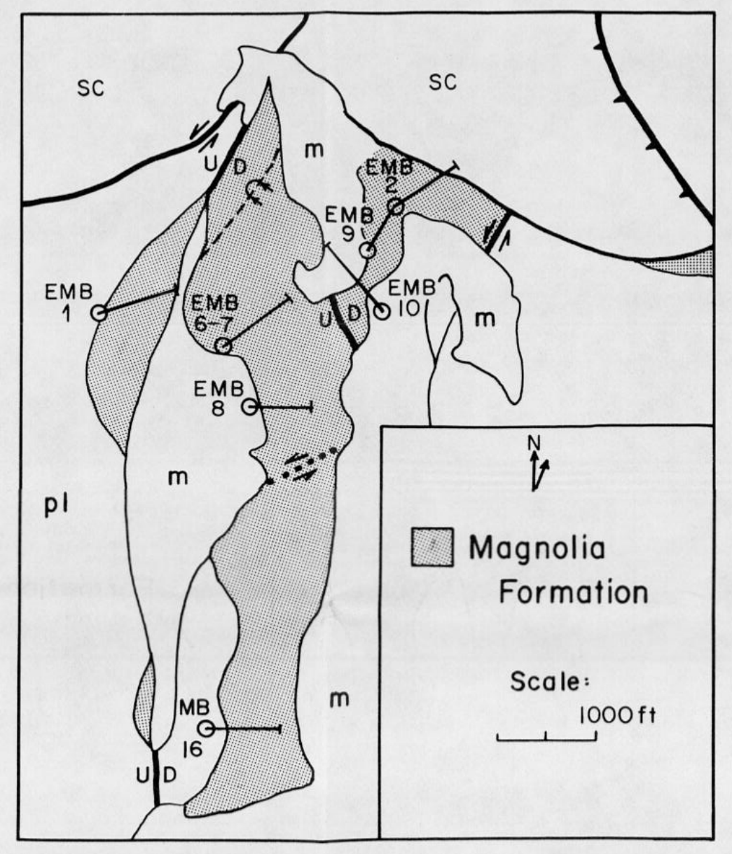
- • 5-1cm
- ○ >1-2cm
- ● >2-3cm
- ○ 3-5cm
- ● >5cm with value written below circle 7cm

Circles represent maximum pebble size for conglomerates in units or subunits.

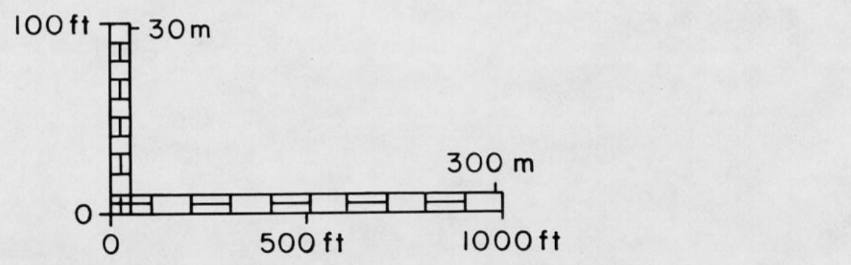
Radiometric scale in CPM x 1000







Index map of Onemile Creek Area showing locations of stratigraphic columns.



Distances measured from center to center of columns

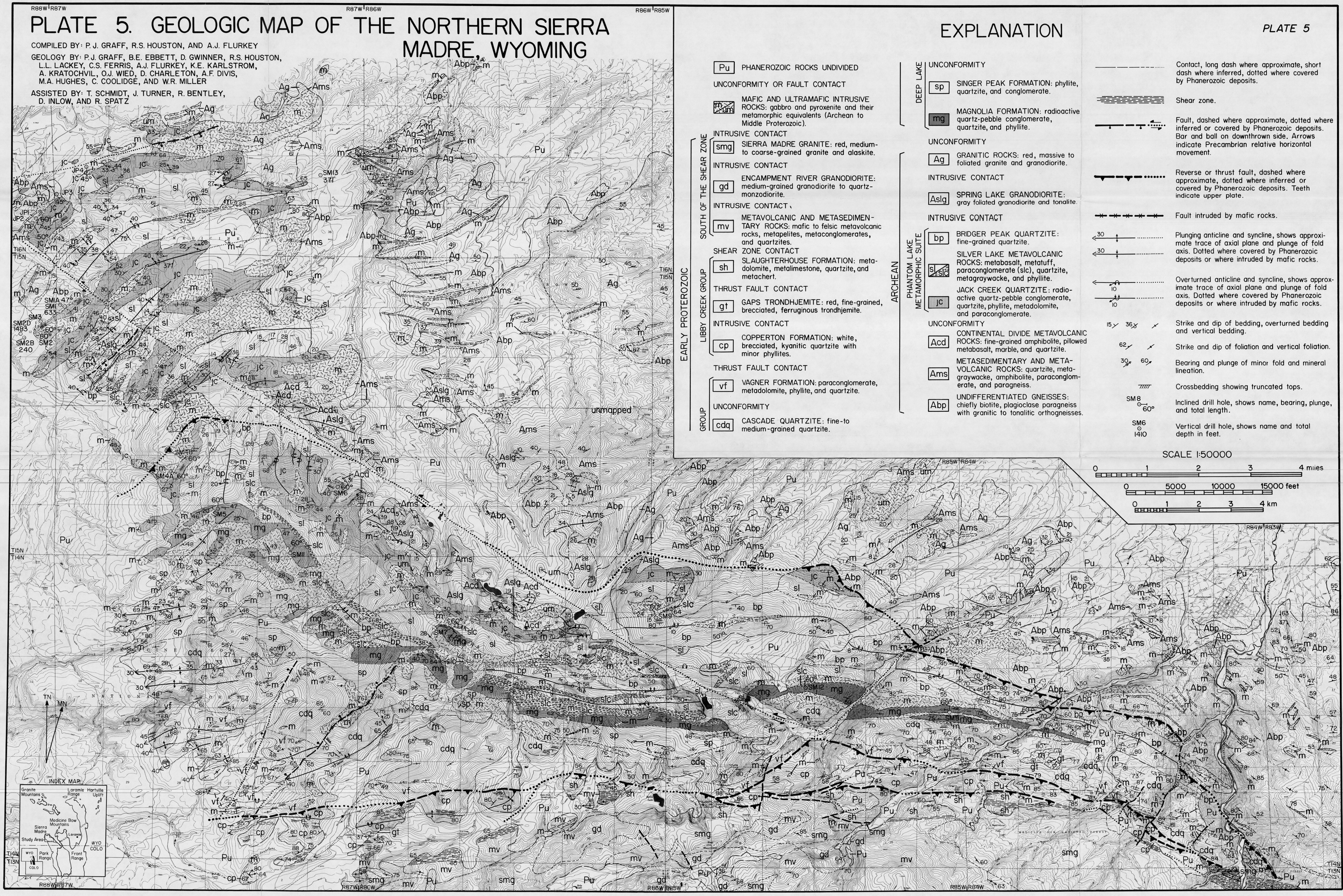
Plate 4. Stratigraphic columns, maximum pebble size, petrography, and gross gamma logs for the Magnolia Formation in the Onemile Creek Area, Medicine Bow Mountains. Data were compiled from drill core and measured sections, gross gamma logs from surveys on core using hand-held McPhar(TV-1A) spectrometers.

PLATE 5. GEOLOGIC MAP OF THE NORTHERN SIERRA MADRE, WYOMING

COMPILED BY: P. J. GRAFF, R.S. HOUSTON, AND A.J. FLURKEY

GEOLOGY BY: P.J. GRAFF, B.E. EBBETT, D. GWINNER, R.S. HOUSTON, L.L. LACKEY, C.S. FERRIS, A.J. FLURKEY, K.E. KARLSTROM, A. KRATOCHVIL, O.J. WIED, D. CHARLETON, A.F. DIVIS, M.A. HUGHES, C. COOLIDGE, AND W.R. MILLER

ASSISTED BY: T. SCHMIDT, J. TURNER, R. BENTLEY, D. INLOW, AND R. SPATZ



EXPLANATION

PLATE 5

Pu PHANEROZOIC ROCKS UNDIVIDED

UNCONFORMITY OR FAULT CONTACT

m / um MAFIC AND ULTRAMAFIC INTRUSIVE ROCKS: gabbro and pyroxenite and their metamorphic equivalents (Archean to Middle Proterozoic).

INTRUSIVE CONTACT

EARLY PROTEROZOIC

SOUTH OF THE SHEAR ZONE

smg SIERRA MADRE GRANITE: red, medium- to coarse-grained granite and alaskite.

INTRUSIVE CONTACT

gd ENCAMPMENT RIVER GRANODIORITE: medium-grained granodiorite to quartz-monzodiorite.

INTRUSIVE CONTACT

mv METAVOLCANIC AND METASEDIMENTARY ROCKS: mafic to felsic metavolcanic rocks, metapelites, metaconglomerates, and quartzites.

SHEAR ZONE CONTACT

LIBBY CREEK GROUP

sh SLAUGHTERHOUSE FORMATION: metadolomite, metalimestone, quartzite, and metachert.

THRUST FAULT CONTACT

gt GAPS TRONDHJEMITE: red, fine-grained, brecciated, ferruginous trondhjemite.

INTRUSIVE CONTACT

cp COPPERTON FORMATION: white, brecciated, kyanitic quartzite with minor phyllites.

THRUST FAULT CONTACT

GROUP

vf VAGNER FORMATION: paraconglomerate, metadolomite, phyllite, and quartzite.

UNCONFORMITY

cdq CASCADE QUARTZITE: fine-to medium-grained quartzite.

DEEP LAKE

UNCONFORMITY

sp SINGER PEAK FORMATION: phyllite, quartzite, and conglomerate.

mg MAGNOLIA FORMATION: radioactive quartz-pebble conglomerate, quartzite, and phyllite.

UNCONFORMITY

ARCHEAN

Ag GRANITIC ROCKS: red, massive to foliated granite and granodiorite.

INTRUSIVE CONTACT

Aslg SPRING LAKE GRANODIORITE: gray foliated granodiorite and tonalite.

INTRUSIVE CONTACT

PHANTOM LAKE METAMORPHIC SUITE

bp BRIDGER PEAK QUARTZITE: fine-grained quartzite.

sl / slc SILVER LAKE METAVOLCANIC ROCKS: metabasalt, metatuff, paraconglomerate (slc), quartzite, metagraywacke, and phyllite.

jc JACK CREEK QUARTZITE: radioactive quartz-pebble conglomerate, quartzite, phyllite, metadolomite, and paraconglomerate.

UNCONFORMITY

Acd CONTINENTAL DIVIDE METAVOLCANIC ROCKS: fine-grained amphibolite, pillowed metabasalt, marble, and quartzite.

Ams METASEDIMENTARY AND METAVOLCANIC ROCKS: quartzite, metagraywacke, amphibolite, paraconglomerate, and paragneiss.

Abp UNDIFFERENTIATED GNEISSES: chiefly biotite, plagioclase paragneiss with granitic to tonalitic orthogneisses.

Contact, long dash where approximate, short dash where inferred, dotted where covered by Phanerozoic deposits.

Shear zone.

Fault, dashed where approximate, dotted where inferred or covered by Phanerozoic deposits. Bar and ball on downthrown side. Arrows indicate Precambrian relative horizontal movement.

Reverse or thrust fault, dashed where approximate, dotted where inferred or covered by Phanerozoic deposits. Teeth indicate upper plate.

Fault intruded by mafic rocks.

30 30 Plunging anticline and syncline, shows approximate trace of axial plane and plunge of fold axis. Dotted where covered by Phanerozoic deposits or where intruded by mafic rocks.

10 10 Overturned anticline and syncline, shows approximate trace of axial plane and plunge of fold axis. Dotted where covered by Phanerozoic deposits or where intruded by mafic rocks.

15 36 Strike and dip of bedding, overturned bedding and vertical bedding.

62 Strike and dip of foliation and vertical foliation.

30 60 Bearing and plunge of minor fold and mineral lineation.

Crossbedding showing truncated tops.

SM8 60° Inclined drill hole, shows name, bearing, plunge, and total length.

SM6 1410 Vertical drill hole, shows name and total depth in feet.

SCALE 1:50000

0 1 2 3 4 miles

0 5000 10000 15000 feet

0 1 2 3 4 km

EXPLANATION

EARLY PROTEROZOIC	sc	SOUTHERN COMPLEX: includes metavolcanic and metasedimentary rocks, mafic intrusives, the Encampment River Granodiorite and Sierra Madre Granite.
	ulcg	UPPER LIBBY CREEK GROUP: Slaughterhouse Formation
	llcg	LOWER LIBBY CREEK GROUP: Copperton Formation
	dlg	DEEP LAKE GROUP
ARCHEAN	plms	PHANTOM LAKE METAMORPHIC SUITE
	nc	NORTHERN COMPLEX: includes Continental Divide Metavolcanic Rocks, metasedimentary and metavolcanic rocks undifferentiated, paragneiss and orthogneiss.

Shear zone, dashed where inferred or covered.

Reverse or thrust fault, dashed where approximate, dotted where inferred or covered.

Fault, dashed where approximate, dotted where inferred or covered. Bar and ball on downthrown side. Arrows indicate Precambrian relative movement. Open circles indicate Late Tertiary movement.

30 40 Plunging syncline and anticline shows approximate trace of axial plane and plunge of fold axis. Dotted where covered by Phanerozoic deposits or where intruded by mafic rocks.

30 40 Overturned syncline and anticline shows approximate trace of axial plane and plunge of fold axis. Dotted where covered by Phanerozoic deposits or where intruded by mafic rocks.

Boundary of subarea.

20
$81\pi S_1$

Lower hemisphere equal area projections of contoured poles to bedding (πS_0) and foliation (S_1) showing subarea number and number of measurements.

.5-1%	1-3%	3-5%	5-7%	7-9%	>9%

Concentration of poles per 1% area

----- Inferred position of girdle of poles to surfaces (S_0 and S_1)

+ Mineral lineation.

· Minor fold axis lineation.

β_2 ● Inferred macroscopic fold axis defined by girdle of poles to surfaces; subscript shows generation of folding.

S_0 Axial plane or average position of S-surface; subscript shows generation of folding.

PLATE 6

TECTONIC MAP OF THE NORTHERN SIERRA MADRE, WYOMING

Andrew J. Flurkey, 1980

includes data from: Graff (1978), Gwinner (1979), Ferris (1964), Lackey (1965), Wied (1960), Divis (1976), Hughes (1973), and Miller (1971).

additional data collected by: A. Flurkey, K. Karlstrom, R.S. Houston, A. Kratochvil, C. Coolidge, J. Turner, T. Schmidt, D. Inlow, R. Bentley, and R. Spatz.

Area	n	$\bar{x}$(°)	L(%)	s(°)	s^2	R
Carrico Ranch	117	191	74	44	1936	1.5×10^{-28}
Ridge 1, Unit 2	35	167	94	20	400	3.7×10^{-14}
Ridge 1, Unit 3	34	241	58	59	3481	1.1×10^{-5}
Ridge 2, Unit 2	31	195	95	18	324	7.1×10^{-13}
Ridge 3, Units 2 & 3	17	179	89	28	784	1.4×10^{-6}
Manning Ranch	32	276	69	49	2401	2.4×10^{-7}
Ridge 1, Unit 4	10	319	72	47	2209	5.6×10^{-3}
Ridges 2&3, Units 1&3	22	260	80	38	1444	7.7×10^{-7}
Manning Ranch and Mill Creek	58	248	54	64	4096	4.5×10^{-8}
Mill Creek	26	206	61	57	3249	6.3×10^{-5}
Joes Park	17	181	81	38	1444	1.4×10^{-5}

n = Number of measurements

$\bar{x}$(°) = Vector mean azimuth of paleocurrent

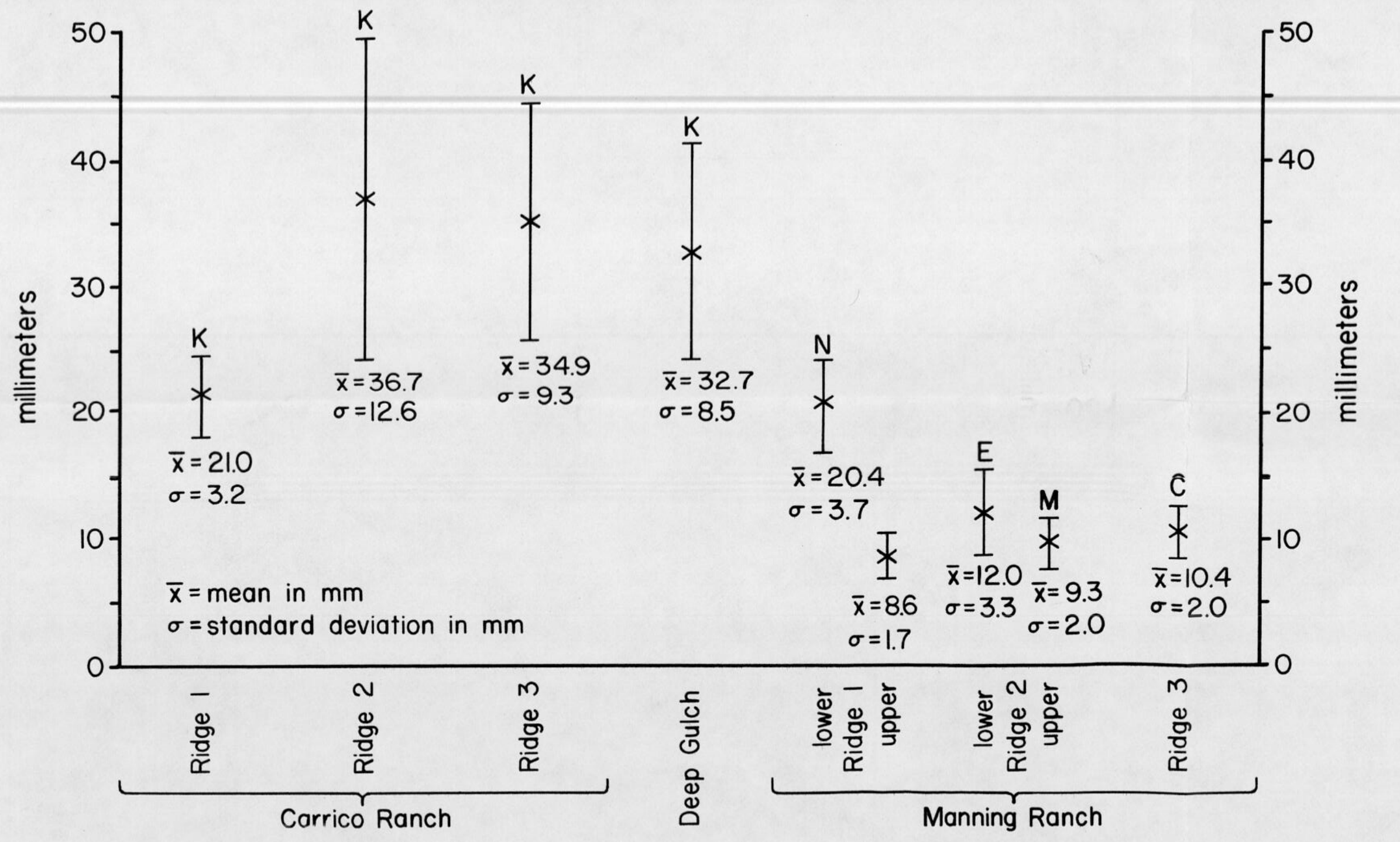
L(%) = Magnitude of vector mean: this is a measure of dispersion (L = 100% for no dispersion)

s(°) = Circular standard deviation (Mardia, 1972, p.24) $s = 180/\pi(-2 \ln L/100)^{1/2}$

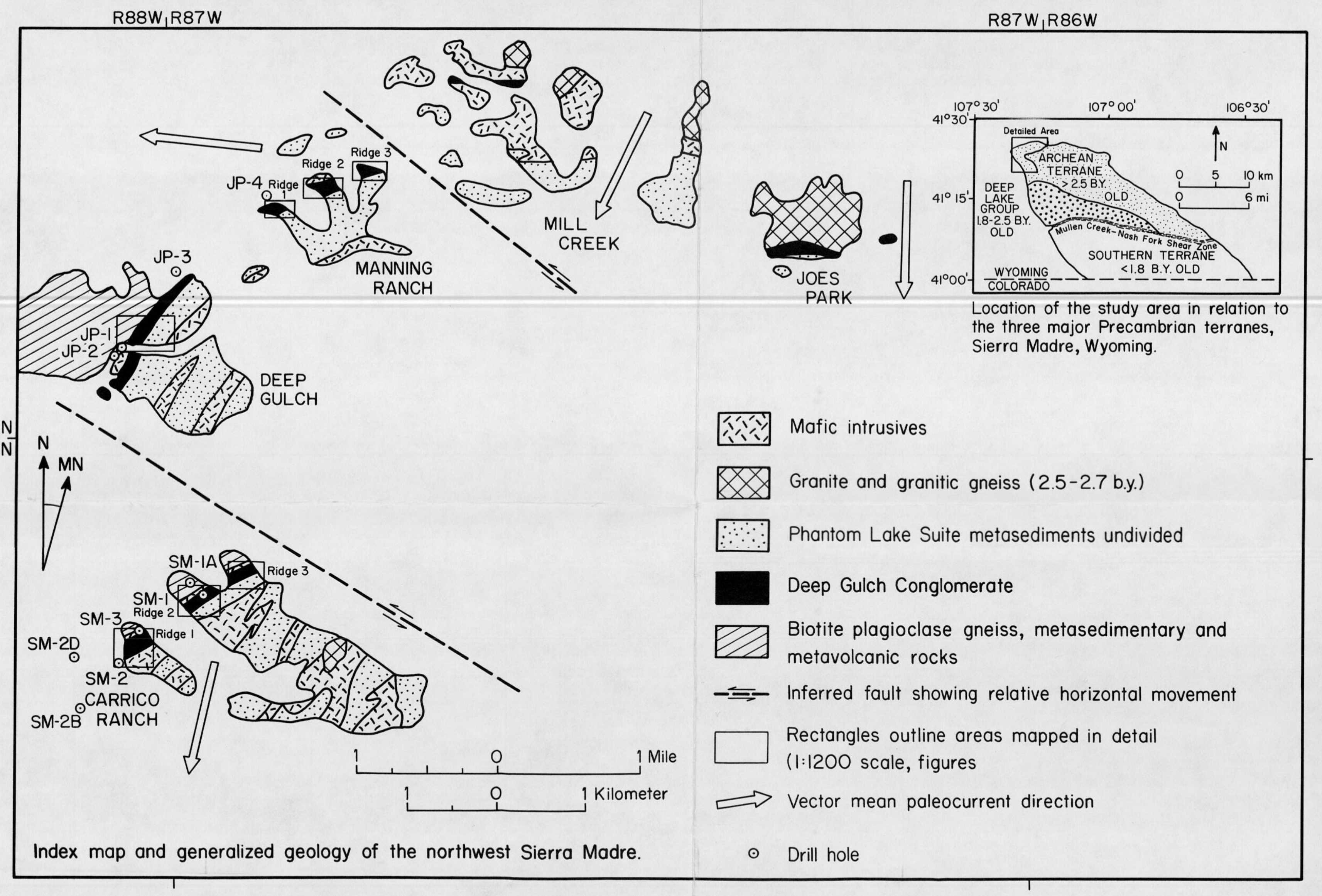
s^2 = Variance

R = Rayleigh test of significance (Curray, 1956, p.125); values less than 5×10^{-2} means that there is less than one chance in 20 that the sample represents a randomly oriented population.

Statistical analysis of paleocurrents from the northwest Sierra Madre, Wyoming.

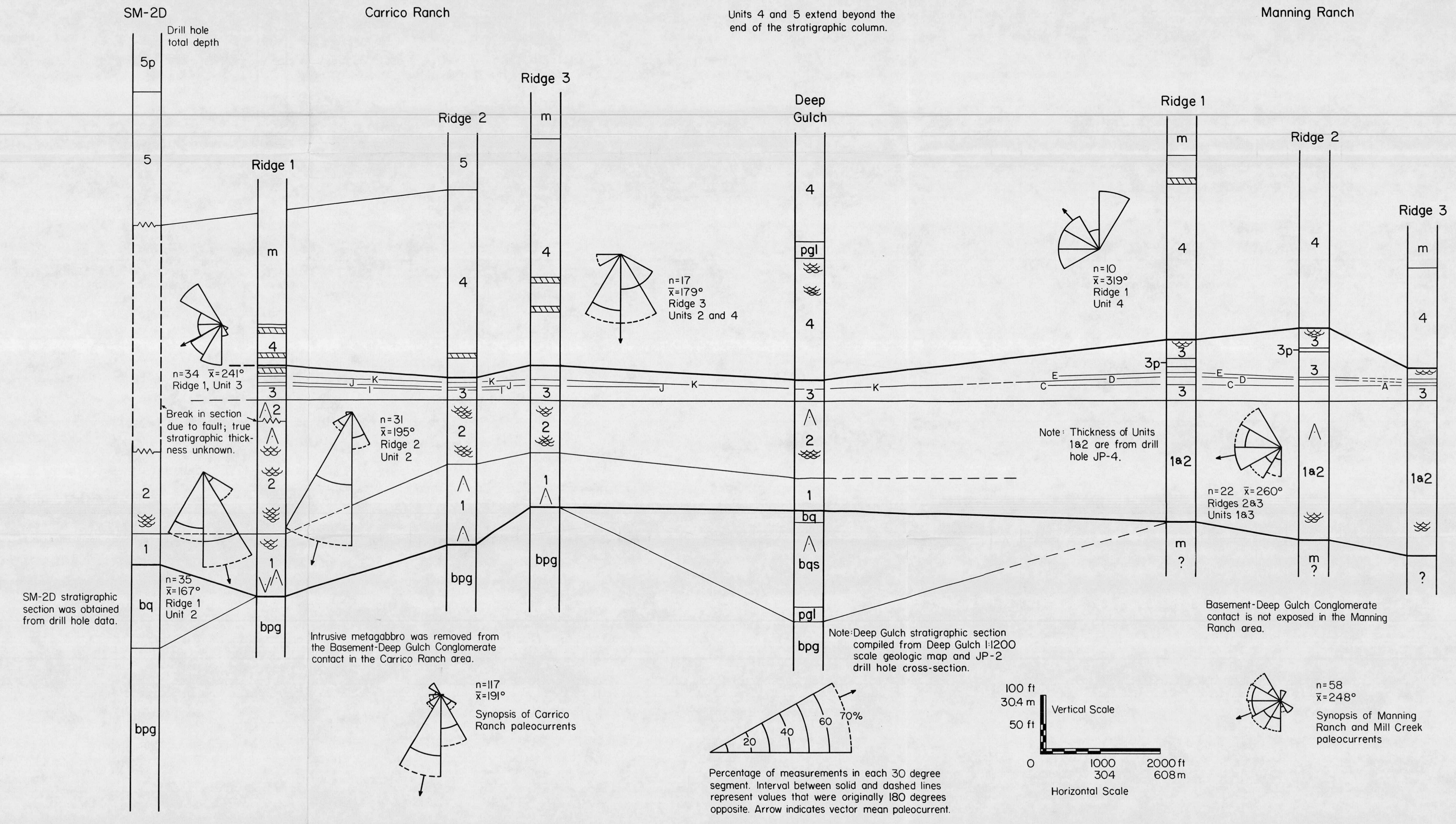


Mean and standard deviation of the maximum apparent diameter of pebbles in the most radioactive quartz-pebble conglomerates. Letters designate conglomerate horizon measured. Ridges 1 and 2 Manning Ranch have upper and lower conglomerates.



Location of the study area in relation to the three major Precambrian terranes, Sierra Madre, Wyoming.

Index map and generalized geology of the northwest Sierra Madre.



Units 4 and 5 extend beyond the end of the stratigraphic column.

Break in section due to fault; true stratigraphic thickness unknown.

SM-2D stratigraphic section was obtained from drill hole data.

Intrusive metagabbro was removed from the Basement-Deep Gulch Conglomerate contact in the Carrico Ranch area.

Synopsis of Carrico Ranch paleocurrents

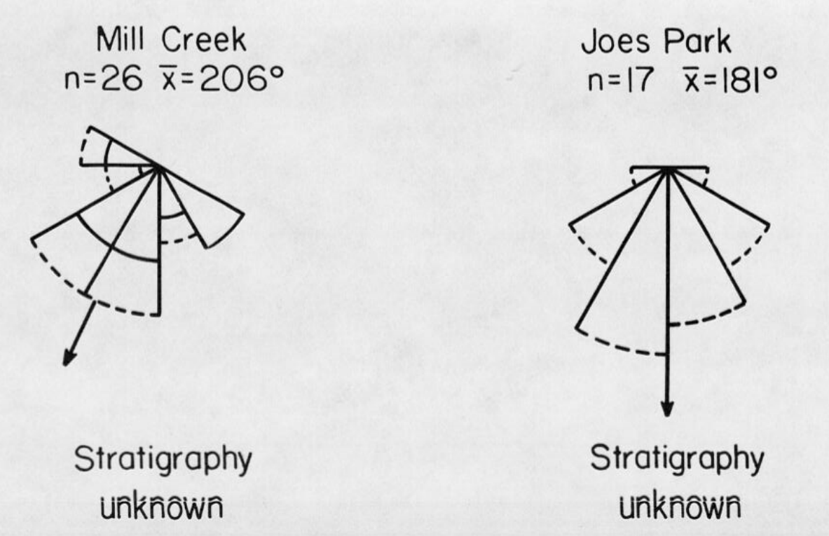
Note: Deep Gulch stratigraphic section compiled from Deep Gulch 1:1200 scale geologic map and JP-2 drill hole cross-section.

Note: Thickness of Units 1&2 are from drill hole JP-4.

Basement-Deep Gulch Conglomerate contact is not exposed in the Manning Ranch area.

Synopsis of Manning Ranch and Mill Creek paleocurrents

Percentage of measurements in each 30 degree segment. Interval between solid and dashed lines represent values that were originally 180 degrees opposite. Arrow indicates vector mean paleocurrent.



Stratigraphy unknown

Stratigraphy unknown

Plate 7. Stratigraphic sections, paleocurrent analysis, and pebble size analysis of the Deep Gulch Conglomerate, northwestern Sierra Madre, Wyoming. Stratigraphic sections were compiled from detailed geologic maps (scale 1:1200) and drill hole data. The heavy lines outline the Deep Gulch Formation. For explanation of symbols and descriptions of units refer to the detailed geologic maps and drill hole lithologic logs.

PLATE 7

Anthony Kratochvil August, 1980