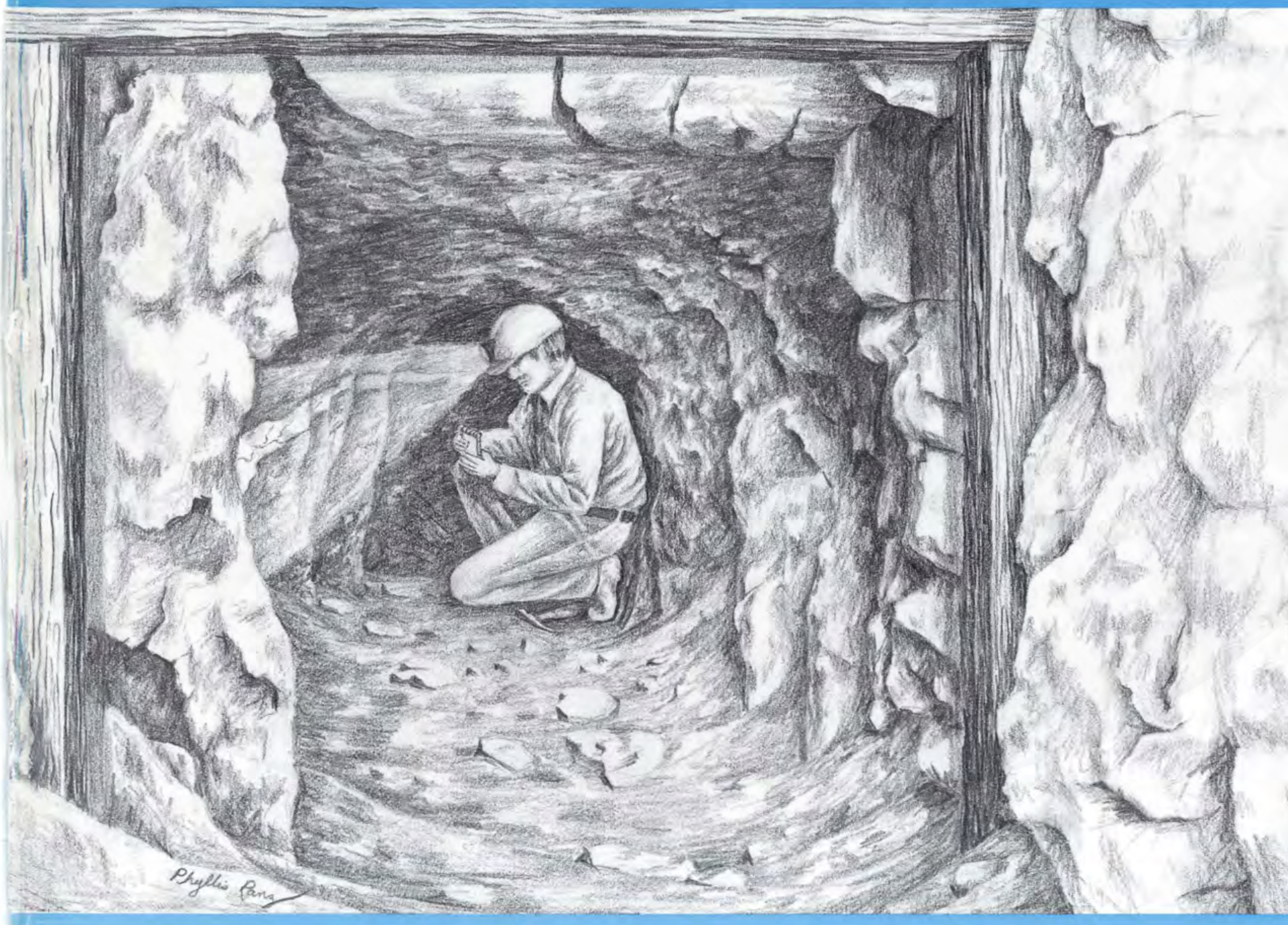
THE GEOLOGICAL SURVEY OF WYOMING
Gary B. Glass, State Geologist

METALLIC AND NONMETALLIC DEPOSITS OF WYOMING AND ADJACENT AREAS, 1983 CONFERENCE PROCEEDINGS

Edited by Sheila Roberts



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Preface

This volume represents the proceedings of a 1983 joint conference of the Geological Survey of Wyoming, the University of Wyoming Department of Geology and Geophysics, and the Wyoming Geological Association, held on the University of Wyoming campus in Laramie. Talks presented on mineral deposits that occur in rocks of Archean to Recent age in Wyoming and adjacent areas provided optimism for future discoveries. These talks also pointed out the need for geochemical, isotopic, and petrographic studies, and detailed geological mapping of the known deposits, so that useful exploration and ore models can be formulated. Recent advances in the understanding of tectonic settings, in particular for the Precambrian, are clearly a step in the right direction (see Houston, this volume).

Gold production from the Archean Wyoming Province has been significant, largely due to the Homestake mine in the northeastern corner of the province. Greenstone belts have not contributed a proportionate share to the total precious metal production from the Archean, but in the future that share could be increased by mineral discoveries stemming from more detailed and intensive exploration in these volcanogenic terrains. In particular, Graff and Hausel (this volume) point to the South Pass-Atlantic City and Lewiston mining districts in the South Pass greenstone belt as areas that warrant further attention.

Near the southern edge of the Wyoming Province within the Proterozoic basement, Conoco Minerals, Incorporated recognized volcanogenic massive sulfides in 1979, in a low-grade metamorphosed terrain (see Schmidt, this volume). Drilling by Conoco defined a thick stratiform zinc-copper-silver deposit with some zones assaying better than two percent zinc. A number of stratiform massive sulfide deposits are now known in this region.

Wyoming is a recognized uranium province. Uranium occurrences are scattered throughout the geologic record, but the greatest interest in the past years has been the Tertiary roll-front deposits. Uranium in these roll-front-type deposits provides Wyoming with the second largest domestic uranium resources. However, a little more than five years ago, Witwatersrand-type, Proterozoic, uraniferous quartz-pebble conglomerates were discovered along the southeastern edge of the Wyoming Province, opening a new frontier for uranium exploration. Scattered gold anomalies have also been detected in these ancient conglomerates (see Hausel, this volume). Harris (this volume) discusses another type of uranium deposit that could open another exploration frontier in Wyoming - the nonconformity type. Several years ago, under more favorable economic conditions, one such deposit was considered for development in the Copper Mountain district of the Owl Creek Mountains.

Diamonds were discovered in a kimberlite diatreme 20 miles south of Laramie in 1975. Before this discovery, very few people considered Wyoming as a place to find diamonds. Today, at least 100 kimberlite diatremes and dikes have been discovered and at least 15 are diamond bearing. Testing by Cominco American, Incorporated showed low-grade diamond mineralization (0.005 to 0.01 carat per ton) with gem-quality to industrial-quality ratios similar to many South African pipes. In Colorado, Superior Minerals Company recovered near-economic grades of 0.2 carat per ton. Diamonds weighing up to a carat have been recovered by both companies. Contributions in this volume by McCallum and Lincoln, Marks and Marrs, and Brink and others discuss aspects of diamond exploration.

Wyoming is the nation's largest bentonite producer. The clay is used principally in drilling muds by petroleum companies, and is also used to pelletize iron concentrates from taconite mining operations. In this volume, Rath discusses the origin of bentonite, how it is mined, and how it is refined.

Following Kentucky, Wyoming is the nation's second largest producer of coal. However, no other state in the United States contains as large coal resources as Wyoming. This important energy mineral is investigated by Rich (this volume). In his paper, the origin of some Wyoming coals are compared to the accumulation of organic debris in modern wetlands.

Papers and extended abstracts on shear zone-hosted base and precious metals, classification of iron formations, sandstone-hosted silver-copper-zinc deposits, Stillwater platinum reef deposits, mineralization associated with alkalic intrusive complexes, porphyry copper-molybdenum, trona, and zeolites are included. We appreciate the contributions of the authors and wish to thank everyone who participated in the 1983 symposium.

Abstracts for the 1983 symposium were published in Geological Survey of Wyoming Public Information Circular 19, *Genesis and exploration of metallic and nonmetallic mineral and ore deposits of Wyoming and adjacent areas.*

Laramie
1986

W. Dan Hausel and Ray E. Harris
Symposium Chairmen

Wyoming Precambrian Province

-example of the evolution of mineral deposits through time?

Robert S. Houston

Introduction

The suggestion that environments favorable for the formation of mineral deposits have changed through time as the planet evolved has become a popular and reasonable thesis with economic geologists and petrologists in recent years (Watson, 1973; Anhaeusser, 1981; Meyer, 1981). If processes that generate mineral deposits differ through time, or if environments of deposition were not the same in the past as they are today, mineral exploration should be guided accordingly. The Wyoming Province, which includes the State of Wyoming and adjacent parts of Montana, South Dakota, Idaho, and Utah, is an excellent area to use as an example of such evolution because of the remarkably complete geologic section preserved, from Archean to Cenozoic. Inasmuch as this symposium emphasizes metallic mineral deposits, these deposits, which are chiefly of Precambrian age in this area, will be the main ones treated in this review.

Evolution of Archean deposits

Archean rocks of the Wyoming Province underlie an area as far east as the Black Hills, south roughly to the Wyoming-Colorado and Wyoming-Utah State Lines, west to the Albion Range of Idaho and Utah, and north to the Little Rocky Mountains of Montana (Plate 1, back pocket). The Archean is divided into Early (greater than 3,400 million years [m.y.] ago), Middle (3,400 - 3,000 m.y. ago), and Late (3,000 - 2,500 m.y. ago). The Wyoming Province includes rocks representative of each of these Archean eras.

Archean rocks are generally divided into three types (Windley, 1977): granulite terrains composed of gneiss with inclusions of high-rank metasedimentary and metavolcanic rocks; gneiss-granite terrain of amphibolite metamorphic grade; and greenstone belts. In Wyoming the most ancient rocks are the granulite terrains and the youngest are believed to be the greenstone belts, but this age relationship is certainly not true in all Archean cratons.

Early and Middle Archean core area

There is a core area of the Wyoming Province that includes the Beartooth Mountains, Bighorn Mountains, Owl Creek Mountains, and northern Wind River Range that may contain the most ancient rocks of the Province (Plate 1). Ages in excess of 3,400 m.y. (Henry and others, 1982) have been suggested for metasedimentary rocks of this region, and there are a number of examples of ages greater than 3,000 m.y. that are believed to date metamorphic events. This is also a region where evidence for early episodes of granulite facies metamorphism has been best documented (Worl, 1968; Worl, personal communication, 1981; Reid and others, 1975; Henry and others, 1982), but where subsequent tectonic and metamorphic events have been so severe that the early history is largely destroyed (Reid and others, 1975), and much of the area is now of lower metamorphic rank.

Remnants of metasedimentary and metavolcanic rocks are present in this gneiss-granite complex, but they are usually in small masses (Worl, 1968). The exceptions are large areas of metasedimentary and metavolcanic rocks in the western Beartooth Mountains (Reid and others, 1975; Page, 1977; and Casella and others, 1982). Experience in similar areas (Anhaeusser, 1981) indicates that mineral deposits are most likely to be in the metasedimentary-metavolcanic units. These units might include chromite in ultramafic bodies; iron formation; and nickel sulfide and associated platinum group metals in mafic bodies.

If most of this core area was subject to granulite facies metamorphism at some early stage in its history, the rocks of the region may also be impoverished in such elements as K, Rb, Th, and U (Tarney, 1976), and any mobile mineral concentrations have probably been destroyed. It is certainly inappropriate to condemn the entire area for mineral exploration, since included in the area are younger intrusives such as the Stillwater Complex that have substantial promise for economic mineralization. However, using the evolutionary approach, the gneiss terrain and included metasedimentary-metavolcanic rocks are not as promising for exploration as other parts of the Wyoming Province.

Middle and Late Archean granite-gneiss terrain and felsic intrusions

Middle and Late Archean rocks of the Wyoming Province include four distinct terrains: (1) a gneiss-granite basement; (2) greenstone belts; (3) a terrain having characteristics intermediate between the volcanic succession-dominated greenstone belt and typical Early Proterozoic metasedimentary rocks (Houston and Karlstrom, 1979); and (4) felsic intrusions that invade these successions. The gneiss-granite basement is like that of the Early Archean except that it is primarily amphibolite facies and has yielded no evidence of an early granulite facies event. The tonalitic, trondhjemitic, and granitic gneisses of the gneiss-granite terrain are not promising hosts for mineralization, and the level of erosion of the granite is such that it is not a promising exploration target, except perhaps for skarn deposits near contacts with calcium-rich rocks.

At least three subdivisions of felsic intrusions of the Wyoming Province can be made. These are:

(1) Middle Archean felsic intrusions (older than 2,900 m.y.) of the northern Bighorn Mountains and northern Wind River Range.

(2) Late Archean felsic intrusions (dated at about 2,600 to 2,800 m.y.) found in the western Beartooth Mountains, in the northern Yellowstone Park area, and in the south central Wind River Range (Lewis Lake batholith). Such granite may invade large areas of the gneiss-granite basement of the eastern Beartooth Mountains (Mueller and others, 1982).

(3) Late Archean to earliest Proterozoic granite (ranging in age from about 2,450 m.y. to 2,600 m.y.) found throughout the southern part of the Wyoming Province including the Black Hills, Owl Creek Mountains, Granite Mountains, Wind River Range, Seminoe Mountains, Freezeout Hills, northern Laramie Range, northern Medicine Bow Mountains, and northern Sierra Madre (Plate 1).

No major mineral deposits have been found in or associated with the Middle Archean or the 2,600 to 2,800+ m.y. old Late Archean felsic intrusions, but the 2,450 to 2,600 m.y. old Late Archean to Early Proterozoic granites are believed to have major economic significance. These Late Archean to Early Proterozoic granites are pink, uranium-, thorium- and potassium-rich granites that appear to have acted as a source of uranium from the Archean to the present (Houston, 1979; Stuckless, 1979) and may have been a direct or indirect source of uranium in the Tertiary deposits of the Wyoming Province.

Late Archean, uranium-bearing granites are present in other Precambrian shield areas, and their development may be related to the initial evolution of rigid subducting plates or to the development of a thick enough crust to allow melting at its base (Condie, 1982). Whatever their origin, these pink granites are the key to the uranium industry of the Wyoming Province.

Middle and Late Archean "greenstone belts"

Geologists specializing in the Precambrian tend to group volcanic rock - dominated successions of unlike character into the greenstone belt classification, and there is an implication that greenstone belts all have certain uniform characteristics. This generalization is necessary if we are to discuss this type of terrain in a reasonable space. The term greenstone implies a low metamorphic rank (greenschist facies), and most rocks of greenstone belts, although commonly deformed, are of low rank, and in many cases have well-preserved primary texture and structure. Greenstone belts generally have a thick lower volcanic sequence that makes up three quarters or more of the total thickness of the rock succession. This volcanic sequence is overlain by a sedimentary sequence dominated by graywacke (Anhaeusser, 1971).

The character of the volcanics seems to have changed through time. In older greenstone belts, ultramafic rocks are abundant in the lower part of the volcanic pile. In younger belts, basalt dominates and volcanic cycles may be present in the upper part of the volcanic pile that are like calc-alkaline suites and range in composition from basalt to rhyolite. Various combinations of these types are possible, and the proportion of sedimentary rocks as well as the maturity of the sedimentary rocks may be greater in the younger belts.

The maturity and character of the greenstone belt of any given area is of economic importance because the nature of the mineralization is thought to be tied to the stage of evolution of the greenstone belt (Anhaeusser, 1981). For example, where greenstone belts are old or have thick ultramafic successions, nickel deposits may be present in the ultramafics. In addition, chromite, chrysotile asbestos, gold, and iron deposits might also be present in the ultramafic successions (Meyer, 1981). In contrast, Canadian greenstone belts, which are generally younger than those of South Africa, have much higher proportions of basalt, and may be characterized by cyclic basalt-rhyolite sequences, typically containing abundant volcanogenic stratiform sulfide deposits (Sangster, 1972; Goodwin, 1979).

Wyoming "greenstone belts" are in the Owl Creek, southern Wind River, Granite, Seminoe, Laramie, and Casper mountains (Plate 1). As can be seen from the discussions of stratigraphy below, the term greenstone belt may not be an appropriate designation for the supracrustal successions in such areas as the southern Wind River Range, Granite Mountains, Owl Creek Mountains, and possibly even Casper Mountain. In addition these successions are largely amphibolite facies and can certainly not be designated as classic or typical greenstone belts. I will use "greenstone belt" in quotes as a temporary designation to separate these supracrustals from those I consider true greenstone belts.

The stratigraphy of Wyoming "greenstone belts" is not well established, and classic successions, ranging from metavolcanics to metasedimentary rocks up-section, have been established in only two areas: the Seminoe Mountains (Klein, 1982) and the Laramie Range (Graff and others, 1981; 1982). In the Seminoe Mountains, a basal volcanic succession about 10,600 feet (3,300 meters) thick, consisting of around 90 percent tholeiitic basaltic volcanics and about 10 percent ultramafics, is overlain by about 2,200 feet (700 meters) of metasedimentary rocks including about 480 feet (150 meters) of iron formation. In the Elmers Rock greenstone belt of the central Laramie Range, the lower three quarters of the greenstone belt succession is mafic volcanics, including ultramafic flows and sills near the base and andesite and rhyolite near the top, which is overlain by metasedimentary rocks consisting of graywacke, paraconglomerate, quartzite, marble, and iron formation. On Casper Mountain (Gable and Burford, 1982), the "greenstone belt" succession is not complete, but includes a lower group of feldspathic schists and quartzite that may have been graywacke, overlain by a succession of serpentinized ultramafic rocks that contain podiform chromite.

The "greenstone belt" of the southern Wind River Range is similarly only partially preserved. The lower part of the Wind River "greenstone belt," which Bayley and others (1973) believed to be mafic volcanics, is represented by inclusions in granite that invades the base of the succession. The upper part consists of a lower quartzite, iron formation, and pelitic schist facies, overlain by a middle section of marine basaltic metavolcanics, and an upper succession of metasedimentary rocks containing local beds of tuff and flows. The Owl Creek "greenstone belt" is not well enough mapped to summarize the stratigraphy, but it does include a thick amphibolite mass that may be metavolcanics and successions of metasedimentary rocks including iron formation (Condie, 1967; Duhling, 1969; Granath, 1975). The same situation applies to the "greenstone belt" of the Granite Mountains (Peterman and Hildreth, 1978; Houston, 1973).

Geochronologic studies on the volcanic rocks of the Wyoming Province "greenstone belts" are sorely needed so that the relationship between isolated belts can be established. Information to date suggests that most "greenstone belts" may be 100 to 300 m.y. older than the 2,450 to 2,600 m.y. old Late Archean granites that intrude them. There are four dates on felsic volcanics of the "greenstone belts": 2,755 $\pm$ 101 m.y. on metadacite of the Owl Creek "greenstone belt" (Mueller and others, 1981); 2,729 $\pm$ 62 m.y. on metarhyolite located in a small mass of greenstone belt-type rocks near Garrett, Wyoming, in the central Laramie Range (Z.E. Peterman, personal communication, 1984); 2,860 $\pm$ 80 m.y. on metamorphism of supracrustals of the Granite Mountains (Peterman and Hildreth, 1978); and 2,800 $\pm$ 100 m.y. on metamorphism of metagraywacke and metavolcanic rocks of the Wind River "greenstone belt" (Z.E. Peterman, personal communication, 1982). These dates suggest that the "greenstone belts" of the Owl Creek Mountains, Wind River Range, and Granite Mountains are older than those of the central Laramie Range, but there is a large enough uncertainty in the dates to make this interpretation questionable. I suspect that the "greenstone belts" are not the same age, and it is probably not even safe to regard them all as Late Archean at this time. Nonetheless, the Wyoming Province "greenstone belt" successions are more like the younger greenstone belt type of the Canadian Shield; certainly the proportion of ultramafic rocks is too low for these rocks to equate to older South African types.

From an economic viewpoint, the significance of the above classification is that this type of greenstone belt is more likely to have volcanogenic stratiform sulfide deposits than nickel-platinum ores, but, to date, no volcanogenic deposits of economic significance have been found. Certainly, more mapping needs to be done before the Wyoming "greenstone belts" can be fully evaluated. Cyclic volcanic successions containing rhyolite are the more favored hosts for volcanogenic sulfide deposits in the Canadian Shield, and although cyclic successions have not been identified in Wyoming, many of the "greenstone belts" do have felsic volcanic rocks.

Greenstone belt successions are also hosts for gold as a trace element in volcanogenic sulfide deposits and exhalative-type iron formation, as well as in veins that may represent material mobilized from the other types of deposits during metamorphism (Boyle, 1979). The "greenstone belts" of Wyoming might offer more promise for gold if the proportion of ultramafic rocks were greater in the volcanic succession. The most important Archean gold province is in South Africa (Auhaeusser, 1976) where ultramafic rocks are abundant in greenstone belts. Geochemical studies in those rocks (Crockett, 1974) show that gold content decreases from ultramafic to felsic rocks. However, gold, is found in all Archean greenstone belts regardless of the stage of evolution. In the Wyoming Province, gold has been found in both iron formation (Hausel, 1982; Hausel and Harris, 1983) and vein-type deposits (Spencer, 1916).

In summary, the "greenstone belts" of the Wyoming Province probably offer the greatest potential for Archean mineralization of any part of the Province. Iron formation is the only mineral deposit of current economic interest that has been identified in any of the belts. Whether this is a function of less than fortuitous exposure, inadequate mapping and exploration drilling, or the stage of evolution of the belts is yet to be determined. It is obvious that much could be gained from detailed mapping and geochronological studies in virtually all "greenstone belt" areas. Hausel (1982) described mineral prospects in the Wyoming portion of the "greenstone belts".

Middle and Late Archean transitional terrain

The third Late Archean terrain includes sequences of metasedimentary and metavolcanic rocks having characteristics transitional between the greenstone belt and the Early Proterozoic metasedimentary successions. These terrains are located along the margin of the Wyoming Province in southwest Montana, northern Utah and southern Idaho, and southern Wyoming (Plate 1).

The Phantom Lake metamorphic suite (Houston and others, 1986) of the northern Sierra Madre and Medicine Bow Mountains of southern Wyoming is a good example of a succession of this type. The basal Phantom Lake metamorphic suite is cut by 2,700 m.y. old granite (Hedge *in* Karlstrom and others, 1981, p. 68) and is thus Archean, but the exact age of all rocks of the succession is unknown. The basal beds of the Phantom Lake metamorphic suite are fluvial and include radioactive quartz-pebble conglomerate of the Blind River type that unfortunately is thorium rich rather than uranium rich. The presence of this well-developed fluvial succession suggests that by the time of deposition of the Phantom Lake metamorphic suite, cratonization had proceeded to the extent that substantial river systems could develop on Archean land masses. The fluvial successions grade upward into marine sedimentary rocks which are, in turn, overlain by marine and continental volcanic successions. The uppermost beds of the Phantom Lake metamorphic suite are marine quartzites. The rocks of the Phantom Lake metamorphic suite are more like those of a mature island arc such as Japan than they are greenstone belt successions.

A somewhat similar Archean succession has been described in the Hartville uplift of Wyoming by Snyder (1980), but here marine successions such as stromatolitic dolomite and iron formation are more prevalent than in the Phantom Lake metamorphic suite. Archean metasedimentary rocks with greater amounts of mature marine sedimentary rocks than volcanics are also present in the Albion and Raft River Ranges of Idaho and Utah, on the southeastern margin of the Great Salt Lake, and in the western margin of the Wasatch Range, and are represented by the Red Creek Quartzite of the northeast Uinta Mountains. However, perhaps aside from the successions of southern Wyoming, the most extensive area underlain by the above type of Archean metasedimentary rocks is in southwestern Montana where metasedimentary rocks are present in the Ruby Range, Blacktail Range, Tobacco Root Mountains, Gallatin Range, and Madison Range. In order of abundance, the metasedimentary succession includes quartzofeldspathic gneiss, hornblende gneiss, anthophyllite-gedrite gneiss, quartzite, aluminous schists, iron formation, and marble (Vitaliano and Cordua, 1979).

Perhaps the most striking feature of this metamorphic rock succession is the low percentage of rocks interpreted as volcanic compared with rocks interpreted as sedimentary (Vitaliano and Cordua, 1979). According to Vitaliano and Cordua (1979), the stratigraphy of this complexly deformed succession is not well enough known to determine its depositional environment, but one possibility is that the entire succession is part of a single sequence derived from an eastern land mass. The clastic succession, dominated by feldspathic gneisses and concentrated in the east, may have graded westward into finer clastic material that was deposited along with chemically precipitated carbonate and iron formation, possibly on a shallow platform (Vitaliano and Cordua, 1979).

The above interpretation may be an oversimplification. For example, Erslev (1983), working in the southern Madison Range of southwest Montana, sees an older gneissic succession that he interprets as a volcanic pile with interbedded sediments, overlain by a metasedimentary suite that he interprets as having been deposited on a distal shelf environment of a passive continental margin. Erslev's interpretation is, in some respects, a return to earlier interpretative schemes that postulated an older gneissic succession, the Pony Series (Tansley and others, 1933) and a younger Cherry Creek Group (Runner and Thomas, 1928) for the Precambrian of southwest Montana.

The above non-greenstone or transitional Archean successions are among the most intriguing of all rocks of the Wyoming Province because they do not fit any of the classic models for typical Archean rocks. In fact, some of these rocks more closely resemble the Grenville Supergroup than an Archean greenstone succession.

A curious aspect of the mineralization in the transitional succession of southwest Montana is that one of the few known mining districts having economic potential for metals is the Jardine-Crevasse Mountain area. This is a gold-tungsten district described as having vein and replacement type deposits (Seager, 1944), but having deposits with characteristics of volcanogenic stratiform deposits in and associated with iron formation. Perhaps, as suggested by Erslev (1983), volcanism was more prevalent in southwest Montana than suspected.

It is not possible to fully evaluate the mineral potential for rocks like these until their age and depositional environment are better established. I suspect that most of these rocks are Late Archean, but both the Phantom Lake metamorphic suite and the southwest Montana successions are older than 2,700 m.y. and might well be 300 to 400 m.y. older (James, 1981). Certainly we cannot prove that these rocks in widely separated areas are of the same age. They do lie along the margin of the Province and probably developed after a major episode of cratonization and at a time when volcanism was less prevalent than before.

In two areas, southwest Montana (Cohenour and Kopp, 1980) and southern Wyoming (Karlstrom and others, 1981), placer deposits of possible economic interest have been found in these transitional successions. Both are thorium rich rather than uranium rich, which suggests that uranium-rich source rocks had not developed at the time the placers formed. There may be a potential for gold in these placers or in veins and fractures developed during metamorphism of these rocks. To date, the most important secondary mineralization has been copper deposits in the rocks of the Sierra Madre (Phantom Lake metamorphic suite) that may be derived from gabbroic intrusions (Spencer, 1904) and thus are not genetically related to the host rocks. Talc deposits in marble of southwest Montana are important economically and may be present elsewhere, but they may have originated during metamorphic events that were not the same in all areas. These transitional-terrain rocks do contain economically significant iron formation (Hartville uplift), and the potential is present for the development of iron mines in other areas (James, 1981).

Evolution of Proterozoic deposits

The Proterozoic is divided into Early (2,500 to 1,600 m.y. ago), Middle (1,600 to 900 m.y. ago), and Late (900 to 570 m.y. ago). The most critical rocks of Proterozoic age in the Wyoming Province, from an economic viewpoint, are Early Proterozoic successions of the Sierra Madre and Medicine Bow Mountains of southern Wyoming and the Black Hills of South Dakota (Plate 1). These Early Proterozoic successions represent the next step in planetary evolution and include sedimentary rock types quite similar to those of the Phanerozoic. In other parts of the planet, miogeoclinal equivalents of these rocks are noted for gold and uranium placers and extensive thick beds of iron formation -- their deposition being related to the stage of evolution of atmospheric oxygen (Houston and Karlstrom, 1979). In the Wyoming Province, uranium-, thorium-, and gold-bearing placers have been found in these rocks in all three areas mentioned above (Graff and Houston, 1977; Houston and others, 1977; Hills, 1979), but, so far, no deposits of economic interest have been located. The fluvial placers of the Black Hills and Medicine Bow Mountains are substantially richer in uranium than those of the Late Archean and were most likely formed after the emplacement and denudation of uranium-rich Late Archean granite.

There is evidence in southern Wyoming (Karlstrom and others, 1981) and the Black Hills (Redden, 1980) that miogeoclinal sediments of Early Proterozoic age were deposited in rifted areas that finally widened to allow sedimentation in deep water. In the Black Hills the rift may have ultimately developed into a northwest-striking aulacogen, in which deeper water sedimentation accompanied by volcanism resulted in the formation of eugeoclinal facies rocks such as graphitic slate, marine volcanics, and iron formation. The graphitic slates, schists, and quartzites of the Homestake mine area of the northern Black Hills are hosts for the most significant primary gold deposits in the United States. Although there is a long-standing debate as to whether the gold is Precambrian or Tertiary (Slaughter, 1968), the overall characteristics of the deposits do not rule out redistribution and concentration of Precambrian gold at the time of emplacement of Tertiary rhyolite plugs (Woodfill, this volume). Gold and other metals may have been deposited in the reducing Precambrian environment and may possibly have had an ultimate source in Precambrian volcanics. If there is any substance to this hypothesis, it might be worthwhile to examine similar facies in other areas such as the Early Proterozoic French Slate of the Medicine Bow Mountains of southern Wyoming (Houston and others, 1968).

While rifting in the Black Hills may have resulted in the development of an aulacogen of limited areal exent, Early Proterozoic rifting in southern Wyoming is believed to have resulted in the complete separation of a southern block of unknown character and proportions, resulting in the development of an Atlantic-type margin (Karlstrom and others, 1983). Ultimately, perhaps around 1,700 m.y. ago, mature island arcs that may have developed in what is now Colorado collided with the rifted southern margin of the Wyoming Province (Hills and Houston, 1979), bringing eugeoclinal environments against the Atlantic-type margin -- somewhat like the Appalachian models of Williams (1979) and Hatcher (1978). These events, if interpreted correctly, offer additional opportunities for mineral exploration. For example, tin- and tungsten-bearing granites may be found near collision sutures; stratiform sulfide deposits may occur in the island arc successions (Houston and Lane, 1984); and if deeper levels of the crust are brought up at such sutures as suggested by Hills and Houston (1979), mafic complexes may have sulfide (Theobald and Thompson, 1968) or oxide mineralization (Houston and Orback, 1976) of economic interest.

Summary

The Precambrian history of the Wyoming Province is interesting and perhaps even unique in some ways. There are enough critical anomalies in the Archean history of the Wyoming Province to suggest that the new time-bound evolutionary models do not fit the geology well enough to use them as a guide in exploration. I suspect that this has much to do with our incomplete knowledge of the Archean geology and geochronology of the Wyoming Province and perhaps also with the stage of development of Archean time-bound evolutionary models themselves. Certainly the better we understand the geologic history, the better our opportunities for using geologic models in mineral exploration. I believe that we are doing much better in the Early Proterozoic; here we are able to successfully apply plate tectonic models and thus better understand geology and should have greater success in mineral exploration.

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Gold from Wyoming greenstone belts -production and prognostications

Paul J. Graff and W. Dan Hausel

Introduction

The term "greenstone belt" has been applied to a variety of Precambrian volcanic-sedimentary supracrustal sequences, which vary significantly in age, in the lithology of the volcanic section, and in the style of mineralization (Anhaeusser, 1981; Meyer, 1981). These Archean greenstone terrains have long been favorite targets for exploration geologists because of the large quantities of gold mined from such belts in the past. Historic gold production from free-world greenstone terrains is approximately 480 million ounces. This estimate was derived by updating figures for belts in Africa, Australia, India, Brazil, Canada, and other shield areas cited by Anhaeusser (1976).

Worldwide gold production from greenstone belts

Production from southern Africa comes from two Archean domains, the Kaapvaal and Rhodesian cratons. At least 29 volcanic-sedimentary belts have been mapped on the Kaapvaal craton (Anhaeusser, 1976). The best studied and most productive of these is the Barberton belt, where companies have been mining gold for nearly 100 years. Total production from the Barberton belt is over seven million ounces. Another million ounces of gold have been produced from other areas of the Kaapvaal craton. Mines in the greenstone belts of the Rhodesian craton, located north of the Kaapvaal craton across the Limpopo mobile belt, have produced about 74 million ounces of gold.

Mining in the African belts has concentrated on occurrences in favorable structural and stratigraphic sites. These include carbonate-rich layers, arsenopyrite-rich strata, and gash veins formed late in the structural history. Ore-mineral assemblages for the auriferous veins suggest that temperatures were around 400°C to 450°C at the time of deposition. Quartz lode and stratiform-type deposits are the principal exploration targets in southern Africa and have been the sources for more than 82 million ounces of gold extracted from some 4,000 mines (Anhaeusser, 1976).

Archean greenstone belts of the Pilbara and Yilgarn blocks of western Australia have been the sources of more than 78 million ounces of mined gold. It appears that most gold in the Pilbara region originated in ultramafic rocks in the lowest section of the greenstone succession and was mobilized during regional metamorphism to favorable depositional sites. The gold in the Yilgarn block is found in a variety of settings, but seems more prevalent in mafic rocks (Hutchison, 1983).

Records from the Indian Shield are very poor, but estimates place production at about 24 million ounces. The bulk of that was won from the Campion Lode of the Kolar gold fields (Anhaeusser, 1976).

Brazilian production records are not at all clear, but the total production probably approaches 50 million ounces. Most of that came from Archean rocks in the Minas Gerais province.

Canada is the world's largest producer of gold from Archean rocks. Total Canadian gold production is approximately 235 million ounces. Hutchison (1983) estimated that three-fourths of Canada's annual production of 1.6 million ounces comes from quartz lodes such as those of the Porcupine district and the other one-fourth from massive sulfide deposits such as the Noranda and Kidd Creek deposits of the Abitibi greenstone belt. Canadian deposits appear to have exhalites and felsic volcanics as source rocks (Goodwin and Ridler, 1970; Ridler, 1976). Both massive sulfide and quartz-lode styles of mineralization are closely associated with greenstone belts. Although South African and Canadian deposits differ in genesis, both types appear to have formed when gold was remobilized into structurally prepared sites.

Gold in Wyoming greenstone belts

The greenstone belts of the Wyoming Province (Plate 1, back pocket) have not been major contributors to the total production of gold in the free world. Optimistic estimates put gold production at 327 thousand ounces from the South Pass-Atlantic City district in the South Pass greenstone belt and at no less than 500 ounces from the Seminoe Mountains greenstone belt (Hausel, 1980). Although the Lewiston district along the southeastern margin of the South Pass greenstone belt contributed some gold to the overall production from Wyoming's greenstone belts, the amount of gold mined from this district apparently went unreported (Hausel, 1986a). Mines in other known belts of the province have produced only minor amounts of gold. Based on very poor records, up to 20 thousand ounces may have been produced from quartz lodes in the 2.5 billion year (b.y.) old granites in the Corner Gulch mine in the Raft River Mountains of northern Utah. These granites intrude a very poorly exposed metavolcanic-metasedimentary belt. A brief visit to this area revealed what looked like highly metamorphosed pillow basalts with inclusions of de-carbonatized carbonates filling the space between the pillow-like structures. If these rocks represent a greenstone belt sequence, then the Corner Gulch gold mine may be analogous to mines in the Rhodesian craton that are in granites that intruded the flanks of greenstone belts.

Total production of gold from Wyoming greenstone belts is apparently more than 350 thousand ounces and may be as high as 500 thousand ounces (Hausel, 1986b). Significant gold production of 35 million ounces is recorded from the Homestake Mine in the Black Hills (Woodfill, this volume) at the edge of the Wyoming Province, or perhaps from a remobilized zone within the province.

Gold from the Homestake mine may be of several ages. Two possibilities for its origin are of interest: (1) remobilization of Archean greenstone deposits, and (2) emplacement during the formation of a Proterozoic volcanogenic belt. If either of these hypotheses is correct, then the Wyoming belts may be tantalizing targets for gold exploration.

Comparison of greenstone belts

We have noted differences in styles of mineralization in greenstone belts, and corresponding differences in the physiology of supracrustal belts themselves. Although the detailed studies of structure, lithology, and chemistry available for the Canadian and African belts are lacking in Wyoming, data from available reconnaissance studies indicates that the belts of the Wyoming Province are not direct correlatives of the Canadian and African belts.

None of the Wyoming belts are greenstone belts in the strict sense; most differ in lithology, sedimentary sequence, and structure from the classic greenstone belts of South Africa and Canada. Wyoming belts do not contain vast amounts of prograde greenschist facies rock. Instead, they exhibit metamorphic grades of at least middle amphibolite facies. The original textures of the sequences have been destroyed during metamorphism. For example, some ultramafic rocks in the Seminoe Mountains belt have been termed spinifex-textured (Figure 1). Although the crystals from these ultramafic rocks display the same textures as the komatiites described by Naldrett (1970), there is no pyroxene or olivine in the coarse-splayed crystals, but only amphiboles produced during metamorphism. It is not yet determined if the amphiboles are pseudomorphic after olivine and pyroxene. Chemical compositions may likewise have been altered during metamorphism.

The sedimentary sequences and cyclicity exhibited in the classic greenstone belts of the world are seldom preserved in the Wyoming belts. Unlike the exposed shield areas of South Africa and Canada, exposures in Wyoming are limited to blocks of Precambrian rocks uplifted during the Laramide orogeny. The belts themselves may occur near the toes of the uplifted blocks and be disrupted by later faulting. Such structural disruption may have destroyed evidence of the volcanic-sedimentary cycles and juxtaposed greatly different levels or units from within the belts. In the South Pass belt, the sequence is dominated by aluminous rocks (in particular graywackes), and the ultramafic and mafic to felsic volcanics are poorly exposed. In the Seminoe Mountains, fairly well-developed banded iron formation, which is often characteristic of the uppermost sediments of greenstone belts, lies directly on rocks of komatiitic affinities, characteristic of the bases of greenstone sequences.

However, disregarding metamorphic grade and considering only structural and stratigraphic data, one greenstone belt in the Wyoming Province, the Elmers Rock greenstone belt of the central Laramie Range, shows similarities to classic greenstone belts. Structurally, the Elmers Rock belt is a domed series of anastomosing synclinal keels which

Figure 1. Spinifex(?) texture in ultramafic rock from Seminoe Mountains belt, Wyoming. Large crystals are amphiboles with no relict pyroxene or olivine identified.

are tightly folded and upright in the center of the belt, but are overturned near enclosing or intruding granite-gneiss domes. Contacts with granitic bodies are highly strained; schistose elements have been rotated parallel to bedding and coaxial planar faults occur along the synclines. The belt is locally cut by post-metamorphic granites (Figure 2).

The Elmers Rock greenstone belt contains ultramafic to mafic flows and sills that form the volcanic group in the lower part of the sequence. The chemical equivalents of komatiites are found within this group as sinuous, lenticular bands (Graff and others, 1982). Pillow basalts and thin interflow metasedimentary lenses overlie mafics and form the upper portion of the volcanic

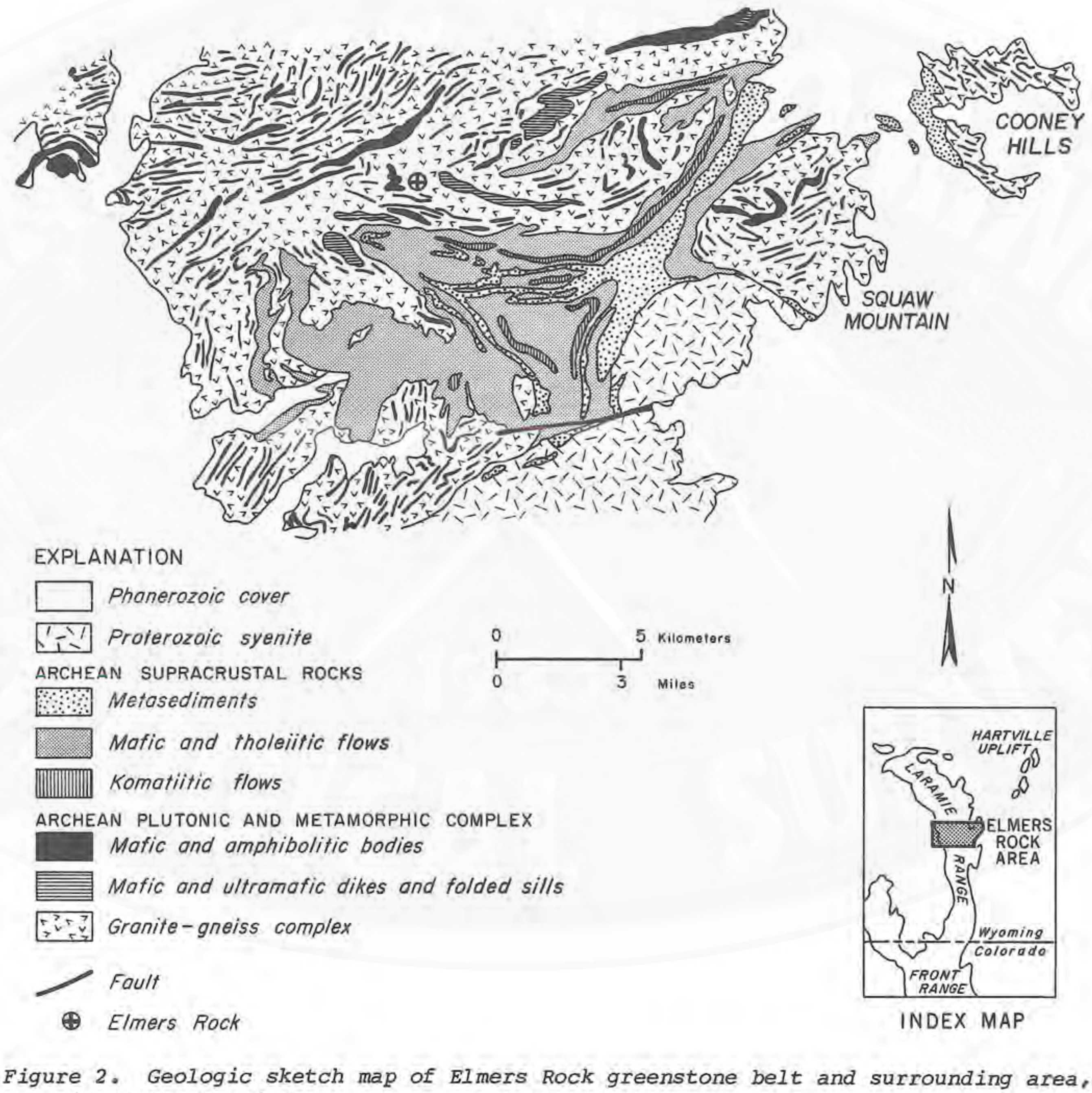

Figure 2. Geologic sketch map of Elmers Rock greenstone belt and surrounding area, Laramie Range, Wyoming.

group. The sedimentary portions of the Elmers Rock area are also like those of the shield areas. Coarse clastic facies, conglomerates, and graywackes are deposited on, and are occasionally interlayered with the flows of the upper part of the volcanic group. These immature sediments are overlain by thin quartzite and marble layers. Rarely, stromatolitic facies occur in the carbonate rocks. Pelites and poorly developed banded iron formation overlie the more mature clastic and carbonate rocks. Figure 3 provides comparative sketches of the Elmers Rock belt and the classic Barberton greenstone belt South Africa, and shows the striking similarities of these two belts.

There are some fundamental differences in lithology between Elmers Rock and the auriferous belts of Africa and Canada (Figure 4). The lithologic percentages for the Elmers Rock belt are estimated from preliminary mapping of the region. When volcanic constituents of each area are compared it appears that the Wyoming rocks are more like the South African than the Canadian sequence. The African and Wyoming belts are dominated by basaltic and ultramafic rocks while the Abitibi belt shows a well-developed mafic to felsic trend, which progresses up the volcanic stratigraphic sequence. While the Canadian belt contains negligible (less than 5 percent) ultramafic flows and a thick andesitic series, the Wyoming and African systems have almost no andesite to rhyolite series rocks.

As stated earlier, the styles of mineralization in the Canadian and African belts are also different. Ridler (1976) showed that gold mineralization in the Abitibi belt is often concentrated in the felsic rocks of the mafic to felsic suite -- the series containing andesites. The deposits are most likely related to exhalatives. In Africa, ultramafic magmas are believed to be the source of most of the gold. In establishing a link between ultramafic magmas and the occurrence of gold, Anhaeusser (1976) cites (1) field relations between ultramafics and gold de-

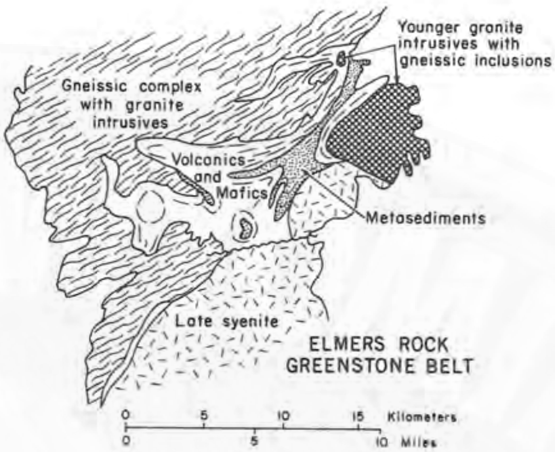

Figure 3. Comparative sketches of Elmers Rock greenstone belt, Wyoming, and Barberton greenstone belt, southern Africa. (After Anhaeusser, 1976 and Graff and others, 1982.) Fine broken lines indicate foliation trends.

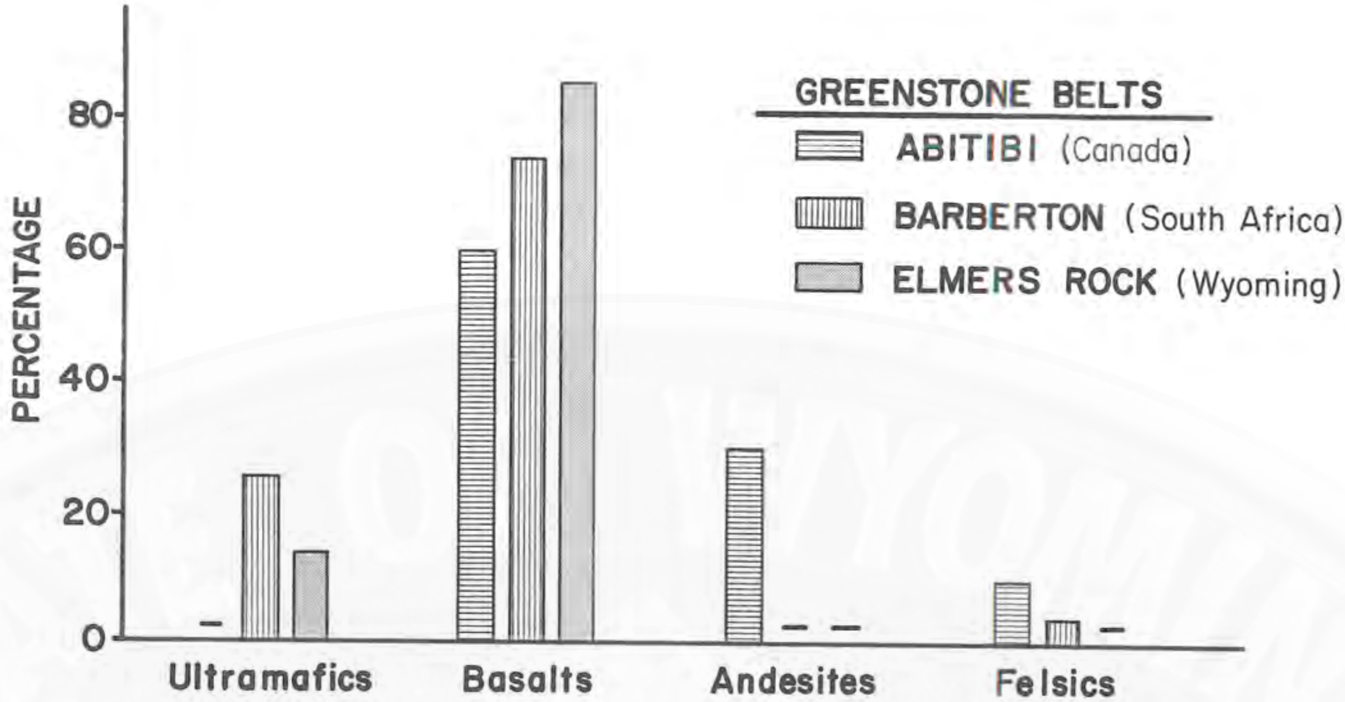

Figure 4. Lithologic comparison of volcanic rocks in the Abitibi (Goodwin, 1971), Barberton (Anhaeusser, 1976), and Elmers Rock greenstone belts.

posits, (2) the hydrated nature of the komatiites as evidence of alteration that may have moved gold, and (3) the presence of gold in fresh olivine crystals. Fresh and similarly altered ultramafic rocks are present in the Wyoming and African belts, but felsic rocks and associated exhalatives are not widely recognized in either of these belts.

Exploration in Wyoming supracrustal belts

The above comparisons are based on general features of belts in Wyoming and other Archean cratons and do not address the many details in which these belts compare favorably or unfavorably. Such fundamental comparisons may prove useful in constructing exploration programs and separating greenstone belts from other Archean supracrustal belts.

We are currently at a crossroads in the study of Wyoming belts. Reconnaissance studies are just beginning to provide detailed surface data on a few sequences, but there is virtually no subsurface data. Industry needs an easy to use, practical classification of supracrustal sequences for the Wyoming Province today, but the data necessary for the construction of such a classification is not available.

Although the exact number of metavolcanic-metasedimentary supracrustal belts in the Wyoming Province varies epending on how one chooses to group them, there are 14 areas which have been called greenstone belts (Houston and Karlstrom, 1980; Houston, this volume). These Archean metasedimentary belts show great variety, and defining and reclassifying these belts must await the detailed studies that are needed.

From reconnaissance studies and previous work, some suggestions can be made. First, nearly all of the metasedimentary sequences are probably fragmental belts, because, unlike belts located in the vast shield areas, all of the Wyoming belts are exposed in basement slabs that have been up-thrust as a result of brittle crustal failure. Because they have been transported for miles in both a vertical and horizontal sense, the Wyoming "greenstone belts" are likely to be fractured and structurally disrupted rather than preserved intact. This is especially true of belts of the central Wyoming Province. Before we can classify them, we must recognize the possibility that portions of the belts, necessary for exact classification may have been structurally removed or displaced.

The second suggestion is a twofold classification that can be effectively employed at our present level of understanding and that will remain usable as we learn more about these belts. We suggest the use of the terms "greenstone belts" and what Windley (1977) has described as "high-grade supracrustal belts." Such a classification of Archean supracrustals will be of value to the exploration geologist because, as a group, greenstone belts are related to world-class gold deposits while high-grade supracrustal belts contain only a limited number of economically valuable deposits. For the present, the Elmers Rock belt and two fragmental belts, the Seminoe Mountains and South Pass belts, probably qualify as greenstone belts (Plate 1, back pocket). High-grade supracrustal belts include belts of the Granite Mountains, Copper Mountain, Casper Mountain, and possibly equivalent rocks in the Wasatch and Raft River Ranges. Most of these sequences are also fragmental belts. These thin belts, which lack the structural characteristics, stratigraphic sequences, and cyclicity of true greenstone belts, probably represent depositional enclaves on the early Archean crust. Whether these enclaves lay between early sialic protoliths or within the more mature Archean craton, they were subject to tectonic interleaving, folding, and migmatization of styles not widely recognized in greenstone belts proper. Such belts are best typified by the occurrence at Copper Mountain. Condie (1969) and Granath (1975) ranked this sequence as a greenstone belt, but studies by Hausel and Graff (1983) and Hausel and others (1985) showed that metasediments of the Owl Creek Mountains are more akin to the high-grade supracrustal belts.

Not all Archean metasedimentary sequences should be ranked in this twofold classification. Falling outside are the quartz-rich, continentally derived clastic series in the Tobacco Root region and thick volcanogenic, clastic, and carbonate assemblages in the Black Hills and Hartville uplift. Archean volcanic-sedimentary sequences in the Medicine Bow Mountains and Sierra Madre are more characteristic of rocks along the leading edge of major plates, and since they are located on the margin of the craton, they may represent protoplate tectonic subduction systems. In some places, the structural setting and stratigraphic sequences of these groups are more like ophiolites. In other places, they are more similar to marginal volcanic or perhaps Archean island-arc terrains. Evidence from mapping suggests depositional environments that differ significantly from the environments forming either greenstone or high-grade supracrustal belts. Exploration programs in these areas must be tailored to meet the problems of those terrains; programs designed for greenstone belts may not be applicable.

Summary

In summary, it is evident that some of the world's richest ore deposits, those containing precious metals, are related to greenstone belts. It is also evident that supracrustal assemblages of the Wyoming Province, whether they are greenstone belts or other metasedimentary terrains, are not yet fully understood, and may well contain important economic concentrations of metals. Changes in the character of deposits from belt to belt elsewhere in the world suggest that each occurrence is different. Exploration in the Wyoming Province must be innovative, and we will probably have to modify models based on known belts and mineral occurrences.

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Base and precious metal mineralization associated with the Mullen Creek-Nash Fork shear zone and related fault systems

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The Mullen Creek-Nash Fork shear zone and associated fault systems in the Medicine Bow Mountains are characterized by locally significant copper mineralization, copper-gold and copper-gold-platiunum-paladium mineralization, and minor copper-silver-paladium mineralization. Except for platinoids, similar mineralization is associated with structures of this major tectonic zone in adjacent ranges. Geological relationships of mineralized zones and geochemistry of ore and host rock assemblages suggest that the majority of these deposits were formed by lateral secretion processes. Although the deposits are enclosed in a variety of felsic and mafic intrusive metavolcanic and metasedimentary rocks, all demonstrate a pronounced spatial affinity to mafic assemblages that apparently were the source of both base and precious metals. Mineralization generally is localized at intersections of faults and shear zones, in brecciated zones and fissures in shear zones and faults, and in minor tear faults and fissures associated with major faults.

No dates are available for the various ore deposits, but field relations suggest that they are essentially penecontemporaneous. The principal mineralizing event may have been related to a thermal-tectonic disturbance associated with intrusion of the Sherman batholith approximately 1,400 m.y. ago. This and associated intrusions may have served as the heat source needed for the mobilization of metals for comminuted mafic materials present in faults and shear zones. Some ore assemblages show the effects of post-crystallization deformation, and a few have been extensively weathered to supergene products.

Chemical and mineralogical studies of ore assemblages and altered wall rocks indicate that the base and precious metals were transported as chloride complexes. Sericitic alteration of wall rock adjacent to most deposits infers low pH conditions that do not favor transport of metals as sulfide complexes. Gold-rich deposits all contain carbonates, and associated mafic rocks have been variably carbonatized, characteristics that typify many lode gold deposits throughout the world.

Metavolcanic rocks and associated volcanogenic mineral deposits of the Fletcher Park and Green Mountain areas, Sierra Madre, Wyoming

Thomas Schmidt

Introduction

Stratabound volcanogenic sulfide deposits occur in a number of localities in the Precambrian metavolcanic-sedimentary rocks of the Sierra Madre, Wyoming. The deposits are contained within a 12,000-foot succession of Early Proterozoic (greater than 1,791 m.y. old, W. Premo, personal communication, 1983) submarine volcanic, volcaniclastic, and sedimentary rocks. Reconstruction of the depositional environment indicates alternating mafic and felsic volcanism occurred during the Early Proterozoic. Large accumulations of volcanic and clastic sediments were deposited above the mafic and felsic rocks. The volcanic-sedimentary strata have been tightly folded and later intruded by granite and granodiorite stocks.

Graff (1978) proposed that the metavolcanic rocks south of the Cheyenne belt (Mullen Creek-Nash Fork shear zone) were part of an island arc that collided with the Archean craton in southern Wyoming 1,700 m.y. ago. This paper supports his idea by showing that the chemistry of the metavolcanic rocks in the Sierra Madre is similar to the chemistry of rocks found at recent convergent plate boundaries.

The metallic deposits consist primarily of pyrite, chalcopyrite, and magnetite and are associated with calcium-rich and brecciated rock. Spencer (1904) described five types of mineralization in the Sierra Madre. The description of one of those types, the Hinton-type, is similar to the description of the stratabound Kuroko deposits in Japan. This study proposes that the Hinton-type deposits are stratabound volcanogenic sulfide deposits.

Regional geologic setting

The Sierra Madre is a northwest-trending uplift of Precambrian rocks that forms a portion of the continental divide in southern Wyoming. The Precambrian core of the Sierra Madre was uplifted during the Laramide orogeny (Graff, 1978). West-dipping Mesozoic sedimentary rocks occur along the west flank of the Sierra Madre. Pleistocene glacial and Holocene fluviatile deposits cover the valley bottoms between ridges of Precambrian rocks.

Graff (1978) divided the Precambrian rocks in the Sierra Madre into three portions -- the northern, central and southern domains (Figure 1). The northern domain is composed of Archean granitic gneisses, granite and metasediments. The central domain contains Proterozoic metasediments that lie unconformably on rocks of the northern domain and are in fault contact with rocks of the southern domain. The southern domain is composed of intrusive igneous rocks and metamorphic rocks. The southern domain is separated from the central domain by a shear zone thought to be an extension of the Mullen Creek-Nash Fork shear zone of the Medicine Bow Mountains and part of the Cheyenne belt of Houston and others (1979).

This study is concerned with the metavolcanic rocks that are in the north portion of Graff's (1978) southern

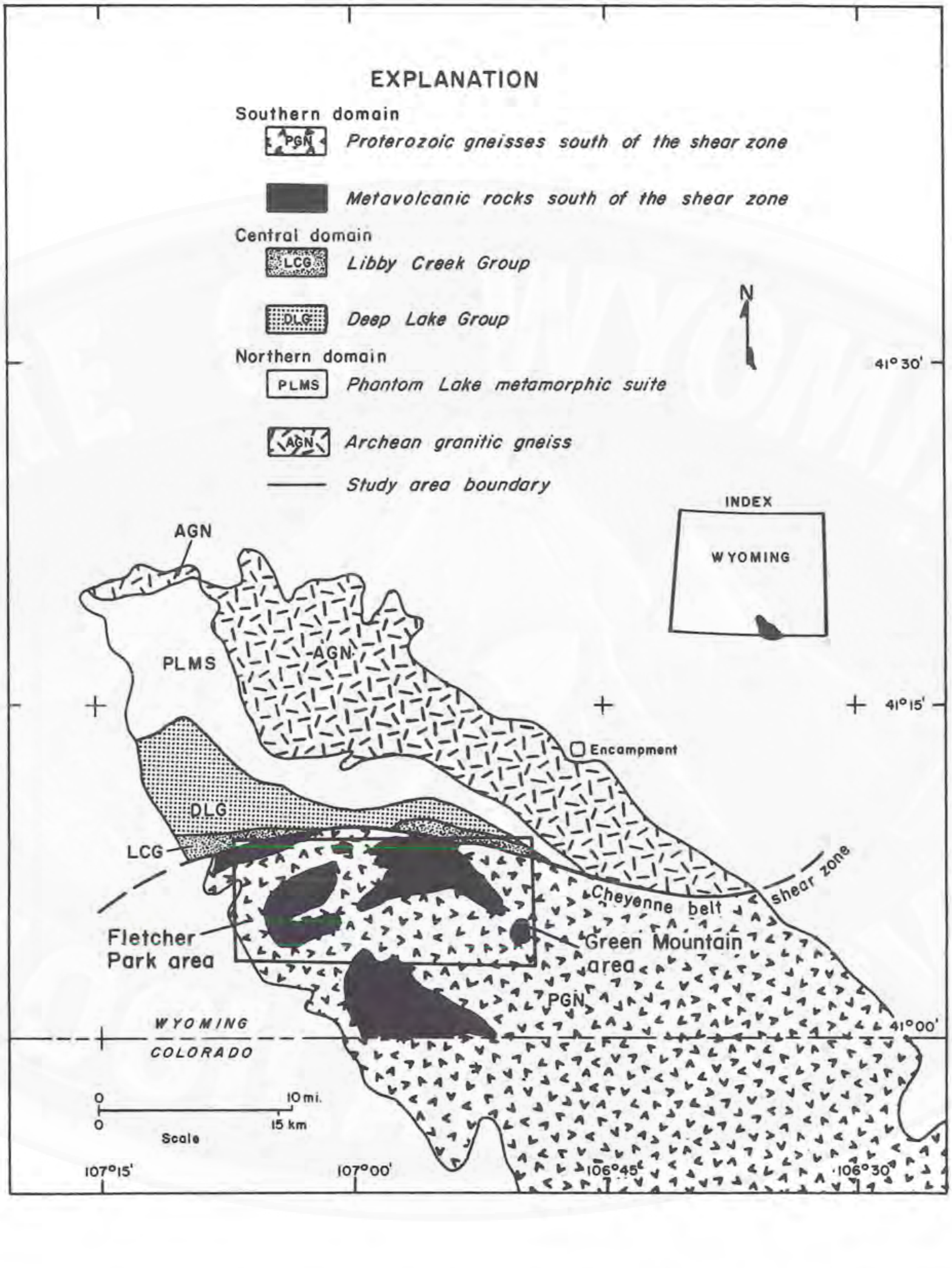

Figure 1. Generalized geologic map of the Sierra Madre. (Modified from Karlstorm and others, 1981; domain designations from Graff, 1978).

domain. These metavolcanic rocks are bounded on the north by granite, gneiss, sheared dolomite, and mylonite of the Cheyenne belt. On the south, the metavolcanic rocks are bounded by granite and quartz monzonite (Figure 1).

Green Mountain area

The Green Mountain area consists of a group of metamorphic rocks that are thought to be Early Proterozoic. These rocks were deposited as a seqnence of volcanic tuff, lapilli tuff, lava flows, and volcaniclastic sediments. The composition of the rocks ranges from basalt to rhyolite. Basalt and dacite are most abundant. Generally, the succession has a mafic volcanic base, a felsic volcanic top, and a metasedimentary central portion (Figure 2). These metavolcanic rocks were cut by gabbro dikes and later surrounded and cut by the 1,680 m.y. old Sierra Madre Granite (Divis, 1976).

Spencer (1904), was the first to recognize the relict volcanic textures. He noted that the rocks have been recrystallized, but their original character can be seen in the angular fragmental texture and relict amygdaloidal structure that are characteristic of volcanic lapilli tuff and lava flows.

The tuffs, lava flows, lapilli tuff and metasediments in the Green Mountain area have a variable lateral extent and pinch out or grade into one another (Figure 3). This type of chaotic deposition is common in volcanic rocks. The occurrence of graded sediments, cross bedding, and carbonate rock suggest that the volcanic rocks were at least in part deposited into a submarine environment.

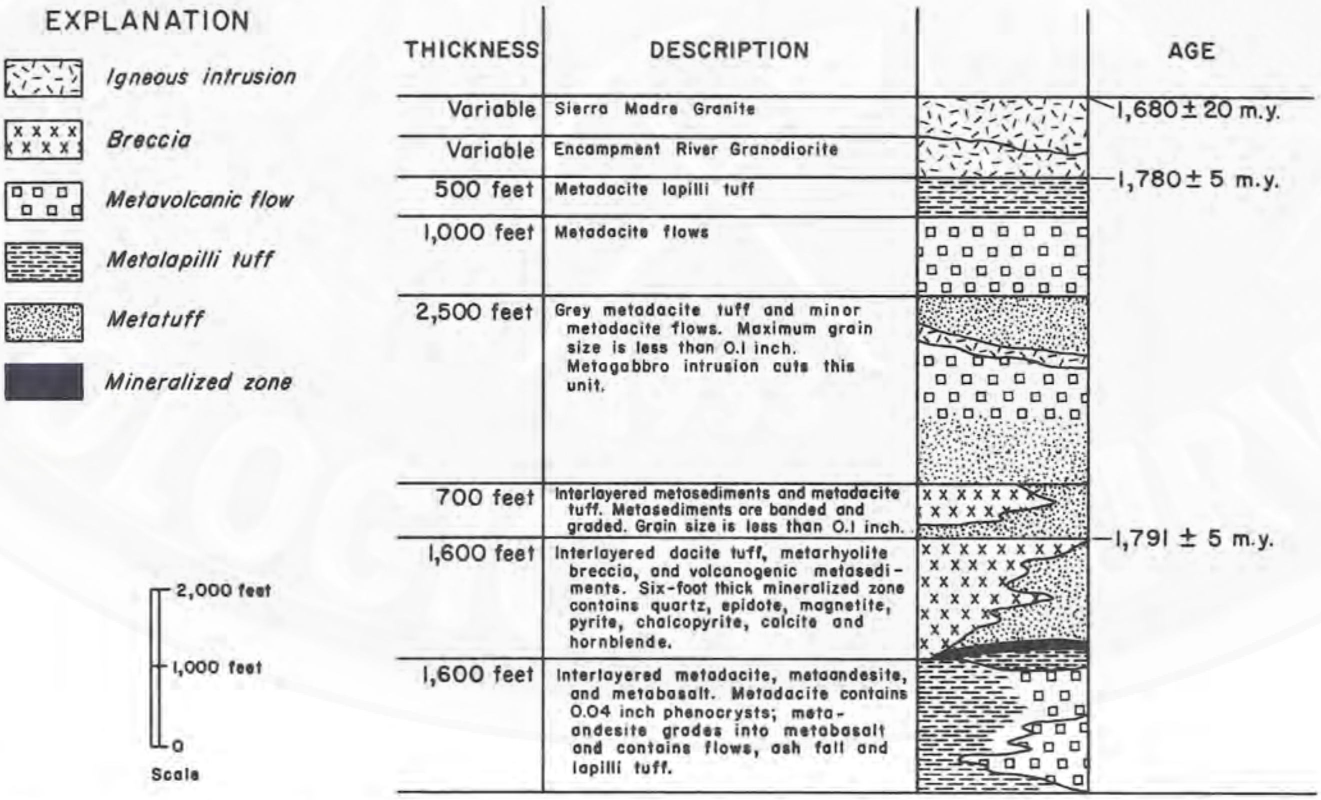

Figure 2. Generalized stratigraphic column of the rocks in the Green Mountain area. Age relations are from Premo and Van Schmus (1982), W. Premo (personal communication, 1983), and Divis (1976).

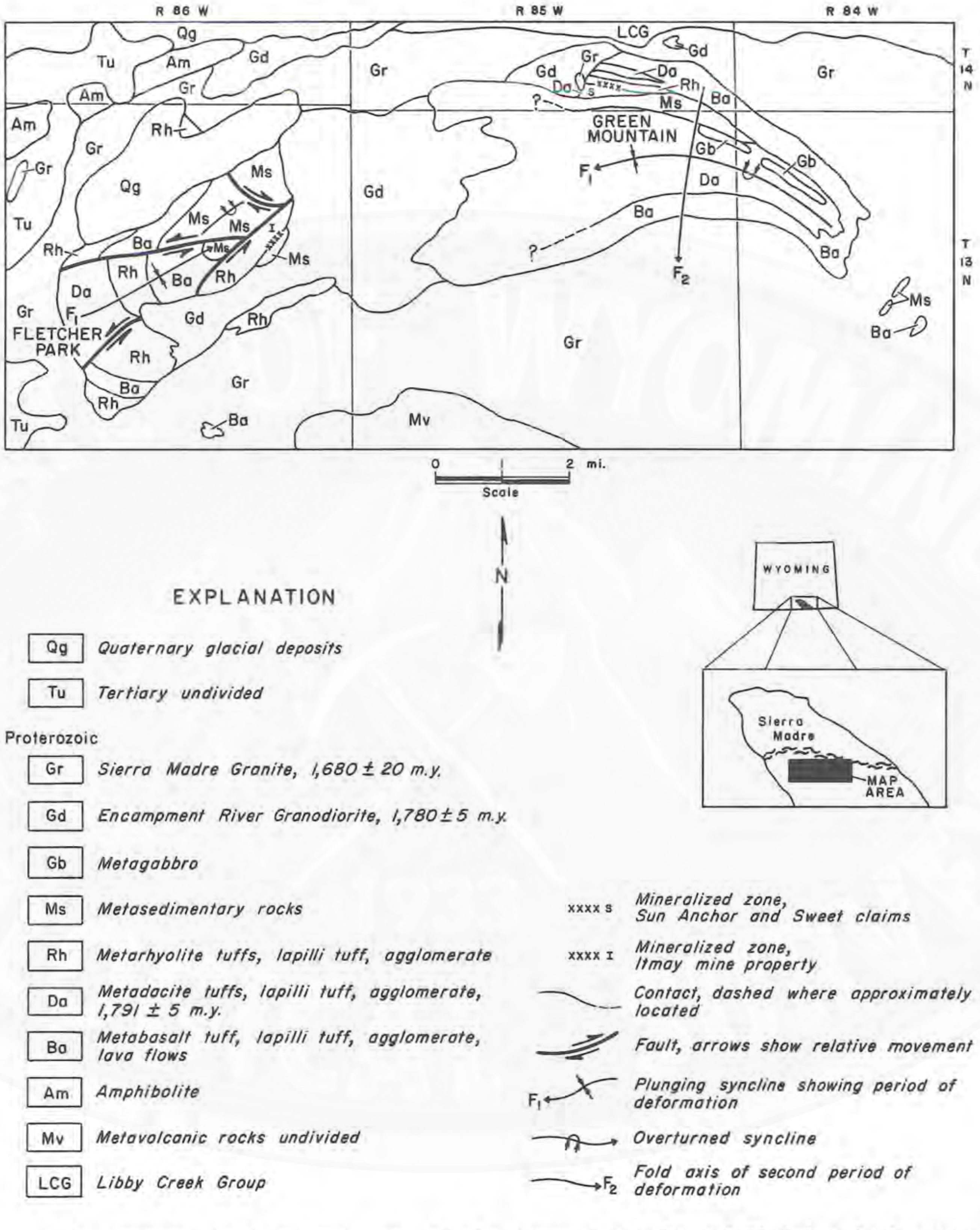

Figure 3. Generalized geologic map of the Green Mountain and Fletcher Park areas. (Modified from Schmidt, 1983; Spencer, 1904.) Radiometric dates are from Premo and Van Schmus (1982), W. Premo (personal communication, 1983), and Divis (1976).

Fletcher Park area

The Fletcher Park area consists of a group of metamorphic rocks that are thought to be Early Proterozoic (Divis, 1976). Exposed within the region are metamorphosed volcanic tuff, lapilli tuff, agglomerate, lava flows and volcaniclastic sediments (Figure 4).

The composition of the volcanic rocks ranges from basalt to rhyolite. The volcanic succession has a mafic base, a felsic central portion, and a mafic upper portion that represents repeated cycles of mafic to felsic volcanism. Sediments occur above the volcanic rocks.

Numerous thin layers of tuffs, agglomerate, and flows indicate many periods of explosive volcanism. As in the Green Mountain area, the occurrence of graded sediments, cross bedding, and calc-silicate rock suggest that the rocks were at least in part deposited in an aqueous environment. Field relations show that the Encampment River Granodiorite intruded the volcanic rocks and was subsequently intruded by the Sierra Madre Granite (Figure 3).

Geochemistry

Rocks of the Green Mountain and Fletcher Park areas were deposited about 1,780 m.y. ago (W. Premo, personal communication, 1983), and have been affected by at least two periods of deformation and metamorphism. The extent of compositional changes that these rocks have undergone cannot be absolutely determined. Nevertheless, the geochemical evidence presented here will show that the volcanic rocks in these areas were most likely formed at a convergent plate boundary and are arc related.

Variation diagrams can be used to determine the tectonic setting of a suite of rocks. Chemical analyses of samples from the Green Mountain and Fletcher Park areas are plotted on Figure 5 in order to compare them with oceanic ridge tholeiites (ORT), Aleutian andesites (island arc) (AA), and Medicine Lake volcanics (continental arc) (ML). These samples plot along roughly the same trend as the Aleutian arc and Medicine Lake rocks. The TiO_2 and MgO diagrams show this particularly well. Note that oceanic ridge tholeiites contain much higher amounts of TiO_2 and MgO than the arc volcanics. In general, TiO_2 values of less than 1.3 percent are typical of subduction zones (Fiess, 1982).

AMF diagrams are commonly used to separate tholeiitic from calc-alkaline rocks. On this type of diagram tholeiitic and alkaline rocks show an initial trend of iron enrichment that changes to alkali enrichment. Calc-alkaline suites on the other hand have a more direct trend of alkali enrichment (Irvine and Baragar, 1971). Samples from the Green Mountain and Fletcher Park areas plot in both calc-alkaline and tholeiitic fields (Figure 6).

Given the degree of alteration of the rocks in the Green Mountain and Fletcher Park areas, it appears unlikely that the chemistry can give unequivocal answers to the original rock composition and tectonic setting. However, the trend of the samples plotted on variation diagrams show that the volcanic rocks were deposited in an arc-related environment. The low TiO_2 values of these samples suggest a subduction-related setting.

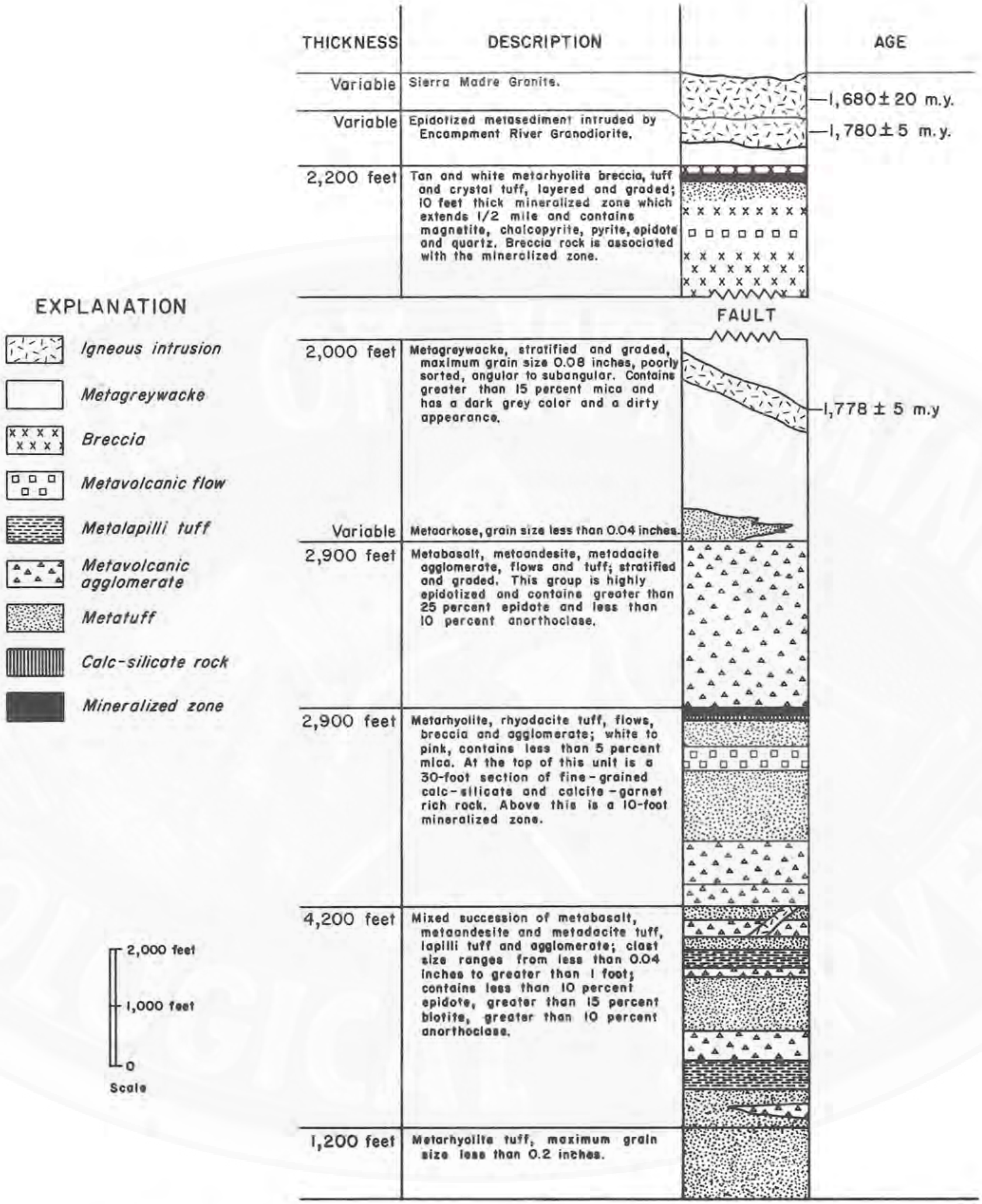

Figure 4. Generalized stratigraphic column of the rocks in the Fletcher Park area. Age relations are from Premo and Van Schmus (1982), W. Premo (personal communication, 1983), and Divis (1976).

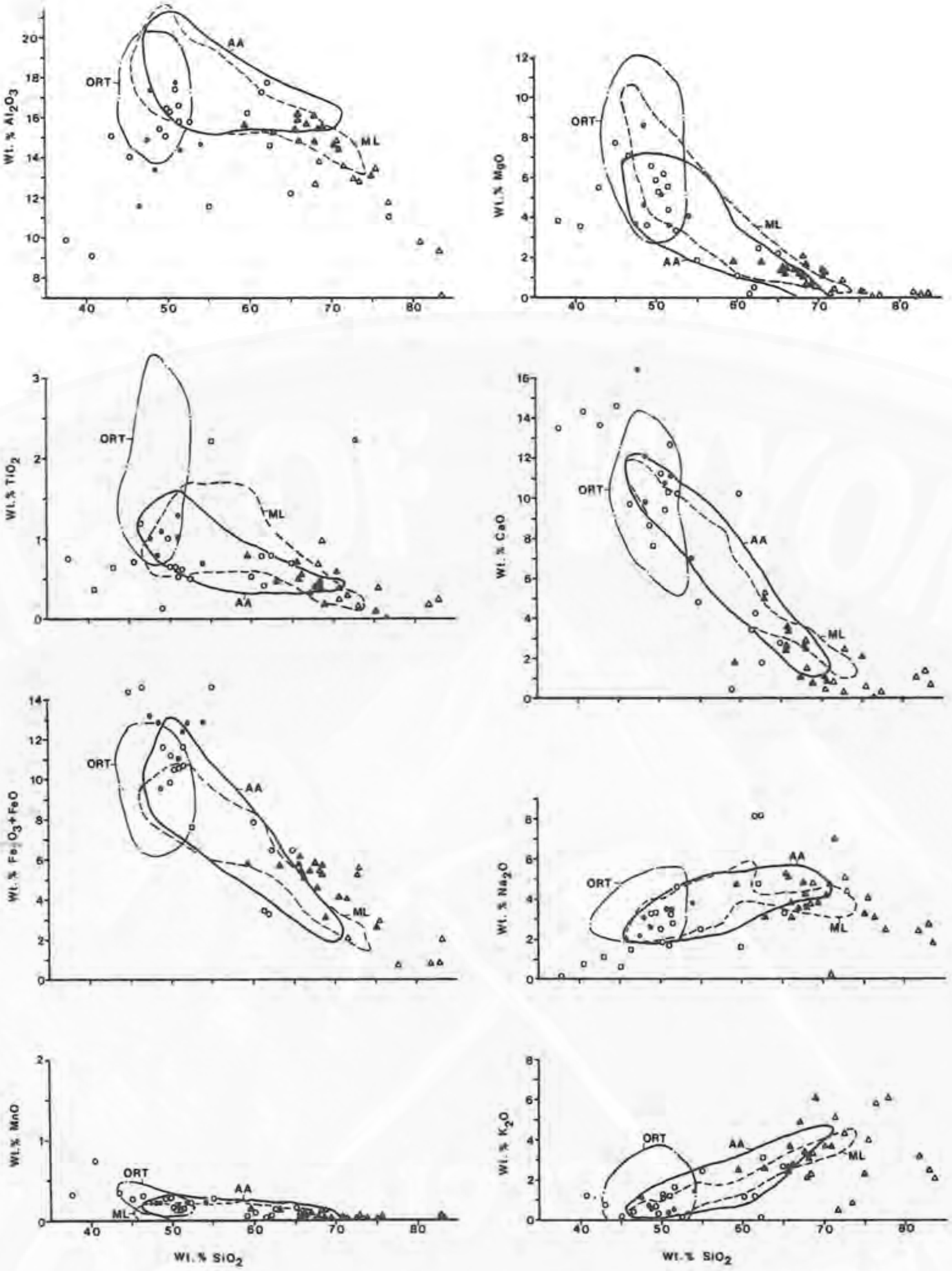

Figure 5. Variation diagrams (Schmidt, 1983). Samples from the Green Mountain area are shown as closed symbols, samples from the Fletcher Park area are shown as open symbols. Outlined areas enclose oceanic ridge tholeiites (ORT) (Engel and Engel, 1963; 1964a; 1964b; Engel and others, 1965; Muir and Tilley, 1964; Aumeto, 1968); Aleutian andesites (AA); (Kay, 1977; Coats, 1952; Marsh, 1976; Arculas and others, 1977; Barth, 1956; Byers, 1961; Coats, 1959; Wilcox, 1959; Marsh, 1976; Drewes and others, 1961; Simons and Mathewson, 1947; 1955; Byers, 1959; Delong and others, 1975; Coats, 1956a; 1956b; Fraser and Barnett, 1959; Forbes and others 1969; Waldron, 1961; Perfit and others, 1980; Fraser and Snyder, 1959); and Medicine Lake volcanics (ML) (Hayslip, 1973; Mertzman, 1977; 1979; Powers, 1932; Anderson, 1933; 1941).

Structure

This region of southern Wyoming is thought to be an ancient plate boundary. The granitic gneiss of the northern domain (Figure 1) represents the Archean craton. The Proterozoic metasediments of the central domain are thought to represent continental shelf-type sediments. They have been folded into an east-west trending system of nappes. The southern domain is composed of intrusive igneous rocks and metamorphic rocks and is separated from the central domain by an east-west trending shear zone of mylonite and ultramylonite.

Green Mountain area. The metavolcanic rocks in the Green Mountain area have undergone at least two periods of deformation. The first period (F_1) can be seen in the tightly folded syncline of Green Mountain (Figure 3). The axis of this syncline trends east-west and plunges to the west. The faulted axial surface dips steeply to the south. Gabbro intruded along the fault and was later folded during the second period of deformation.

The second period of deformation (F_2) refolded the metavolcanic rocks into an open fold about an axis trending S32°W, 70°S. This axis can be seen in the lineated hornblende needles that plunge steeply to the southwest.

Fletcher Park area. The metavolcanic rocks in the Fletcher Park area form a plunging syncline (Figure 3). This tightly folded F_1 syncline plunges N70°E, 40°E. The axial surface of this fold is oriented N70°E, 70°S. Two sets of faults displace the metavolcanic rocks in the Fletcher Park area. The first set of faults are left lateral and trend N50°E, subparallel to the axial surface; the second set of faults trend N45°W and are right lateral.

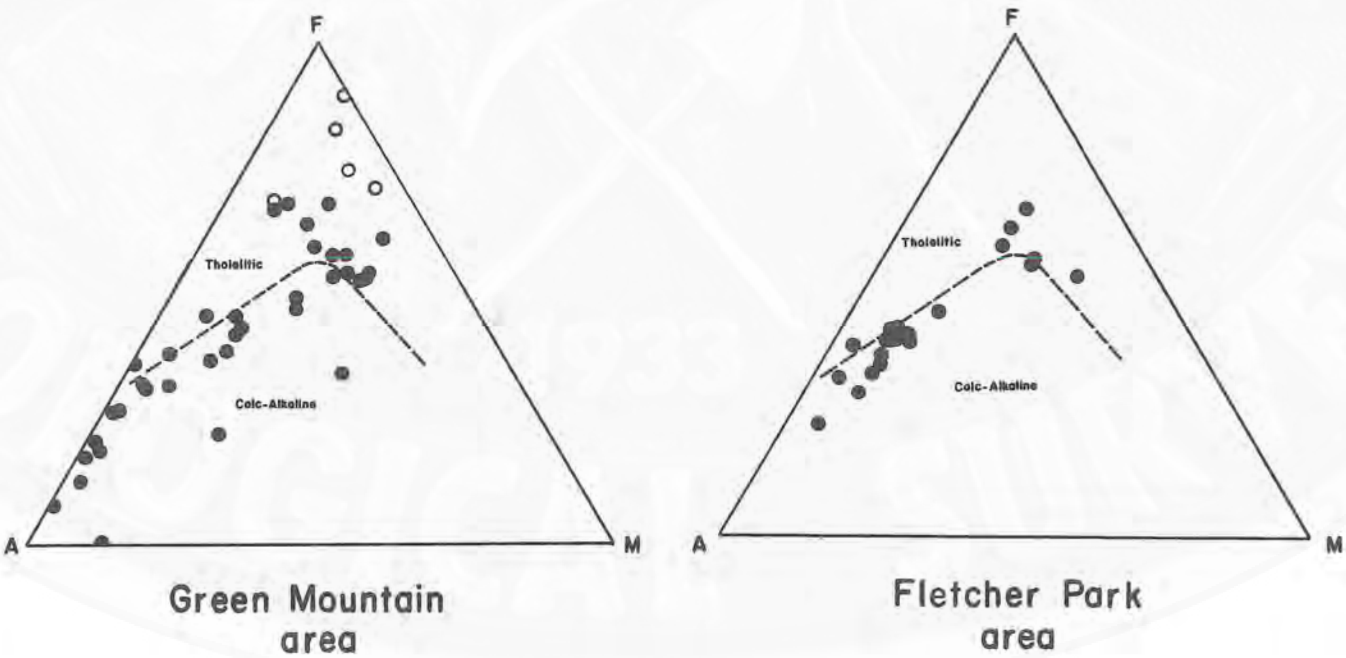

Figure 6. AMF plots of samples from the Green Mountain and Fletcher Park areas. A=NA_2O + K_2O; M=MgO; F=Fe_2O_3 + FeO. Open circles are samples taken near iron formation and have higher than normal iron content. (Dividing line from Irvine and Baragar, 1971.)

Economic geology

Spencer (1904) was the first to describe the economic geology of the Encampment district. Copper-bearing quartz veins were discovered in 1868 by J.W. Southwick and systematic mining of the area began in 1881 at the Doane-Rambler mine. Spencer describes five types of copper occurrences; the Hinton type is the occurrence most important to this report. Spencer noted that the Hinton ore was confined to distinct bands but that there was an absence of fractures by which the metals could have been introduced. He assumed that the ores were produced during metamorphism. However, these rocks have characteristics that are similar to Kuroko-type deposits and are more likely to have been formed by submarine volcanic processes.

The composition and nature of the associated rocks in these deposits show them to be submarine volcanogenic. The rocks in the Green Mountain and Fletcher Park areas show repeated cycles of mafic to felsic volcanism and the rock composition ranges from basalt and andesite at the base of the section to dacite at the top (Figures 2 and 4). Mineralized zones occur above the felsic units and are associated with volcanic breccia and felsic metasediments.

Green Mountain area

A stratabound mineralized zone can be found along the north face of Green Mountain that extends in an arc east from section 34, T.14N., R.85W. (Figure 3). Numerous prospect pits and small mines mark this magnetite deposit. The Sun Anchor and Sweet claims were located along this stratabound deposit but were still undeveloped when Spencer visited them in 1902. There is no report available of the ore grade at these claims. When visited in 1982, all the mine shafts were caved a few feet below ground level.

The stratabound mineralized zone consists of a magnetic layer eight feet thick that extends 100 feet before it is covered by glacial debris. This layer can be traced east from section 34 for about one mile. The magnetite is interlayered with epidote and quartz. In this section, the magnetite layer consists of 25 percent magnetite, 20 percent quartz, 20 percent hornblende, 15 percent carbonate, 10 percent almandite garnet and minor biotite. The magnetite occurs as equant particles and has a maximum grain size of 0.002 inches.

Fletcher Park area

There are two types of copper deposits in the Fletcher Park area, a contact metamorphic deposit and a stratabound deposit. Spencer (1904) described the contact metamorphic deposit at the Itmay mine (Figure 3). This copper deposit occurs along the walls of a mafic dike that cuts granodiorite. The dike has been altered to chlorite. Spencer reports a band of schistose gouge on the north side of the dike that contains streak copper ore and varies from a few inches to two feet wide. He also reports that there are practically no other copper minerals besides chalcopyrite and that one sample contains 0.05 ounces of gold per ton and 17.92 percent copper (Spencer, 1904, p. 51). When visited in 1982, the shaft was caved ten feet below the surface; only small amounts of chalcopyrite were found on the mine dump and no gold was detected.

More important to this report is the second type of copper deposit that can be found 160 feet north of the Itmay mine. This deposit is similar to the Hinton type (Spencer, 1904) and to the stratabound volcanogenic deposit found at Green Mountain. The mineralized zone occurs at the top of a metarhyolite unit and is closely associated with a felsic volcanic breccia (Figure 7). This zone

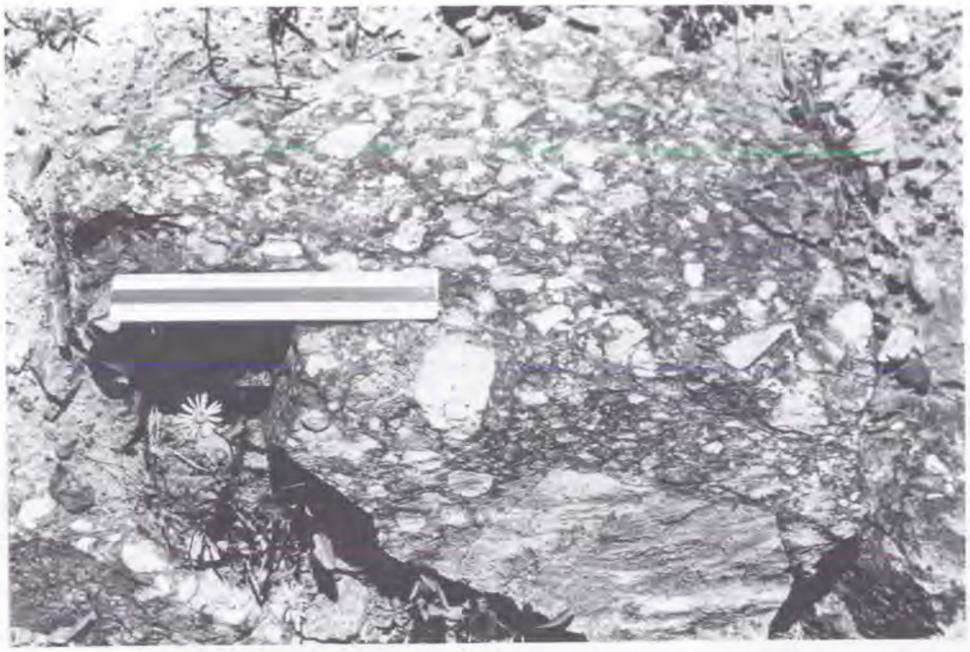

Figure 7. Mill rock breccia near the mineralized zone. Scale is in inches.

is 10 to 30 feet thick and can be traced for over 2,000 feet. The outcrop shows thin layers of magnetite and epidote that are arranged into disharmonic layers and there is no vein or fracture that would allow for the introduction of the metal into the surrounding rock. A fresh sample from the mine dump contains magnetite, pyrite, chalcopyrite, quartz, chlorite, epidote, and apatite. Chemical analysis of this sample (Table 1, no. TGS 30-81) shows 15,000 parts per million copper and 200 parts per million nickel. Four samples from the Fletcher Park area show mildly anomalous zinc values. One sample (TGS 73-81) contains a small amount of tungsten.

Table 1. Chemical analyses of samples from Green Mountain and Fletcher Park areas (Schmidt, 1983).

Sample	FE	Ti	Ag	Au	Ba	Bi	Cu	Ni	Pb	Zn	W	
	percent					parts per million						
						Green Mountain Area						
TGS 5-82	10	0.3	500	N	G(5,000)	500	G(20,000)	30	200	200	N	Granite intrusion
TGS 9B-82	15	0.5	L	N	5,000	N	200	20	500	300	N	Quartz vein
TGS 33-82	G(20)	0.1	N	N	100	300	20	L	10	N	N	Sun Anchor claim
TGS 80-81	10	0.5	0.7	N	200	50	15,000	100	10	N	N	Chlorite schist
						Fletcher Park Area						
TGS 30-81	20	0.1	3	N	700	N	15,000	200	10	N	N	Itmay mine
TGS 36-81	10	0.5	N	N	500	N	50	50	15	500	N	Basalt lapilli tuff
TGS 37081	10	0.3	N	N	500	N	20	70	30	300	N	Andesite lithic tuff
TGS 40-81	20	0.3	N	N	300	N	20	15	15	700	N	Basalt flow
TGS 45-81	10	0.5	3	N	1,000	N	15	7	300	500	N	Dacite agglomerate
TGS 66-81	20	0.02	N	N	20	N	200	30	15	N	N	Gossan
TGS 73-81	20	0.05	N	N	300	N	1,500	7	15	N	100	Gossan

N = none detected; L = detected but in amount below detection limit; G = greater than value shown.

Comparison to other base-metal massive sulfide deposits

Hutchinson (1980) described four types of massive base-metal sulfide deposits under his exhalative volcanogenic category. The oldest and most primitive is the copper-zinc type, which occurs in subsiding basins. This group forms in deep to shallow water, and includes well-differentiated tholeiitic to calc-alkaline, mafic and felsic volcanic rocks. Clastic sedimentary rocks are common. The ore bodies are concordant to bedding and occur above felsic breccia. Examples of this type include Noranda, Quebec (Spence and DeRosen-Spence, 1975) and Jerome, Arizona (Anderson and Nash, 1972).

Hutchinson's polymetallic or Kuroko zinc-lead-copper-type massive sulfide deposit occurs in shallower basins, along with increasing amounts of craton-derived and carbonate-sulfate rich sedimentary rocks. These deposits are associated with bimodal rock suites including tholeiitic basalts and calc-alkaline lavas. These rocks were deposited in very shallow subaqueous or possibly subaerial environments. This group is commonly rich in lead and zinc. Examples of this type of deposit include Buchans, Newfoundland (Thurlow and others, 1975) and the famous Kuroko deposit (Lambert and Sato, 1974).

The third type of massive sulfide deposit is the cuperous pyrite group, which is very different from the previous two in that it is associated with ultramafic and tholeiitic rocks. The ores occur in pillow lavas and there is only minor sedimentary cover. The fourth group is the Besshi type. This type consists of thick sedimentary sequences and tholeiitic volcanic rocks and there is a lack of calc-alkaline or felsic rocks.

The stratabound deposits in the Green Mountain and Fletcher Park areas have characteristics similar to both the primitive type and the polymetallic group, but seem to be more closely related to the polymetallic group. Aspects of these deposits that resemble the polymetallic group include composition and nature of the associated rocks. On variation diagrams, the rocks demonstrate bimodal basalt-rhyolite trends with relatively few samples of intermediate composition. AMF diagrams show the rocks to have both calc-alkaline and tholeiitic composition, which is characteristic of the polymetallic group. The carbonate and calc-silicate rocks that are found in and near the mineralized zone represent submarine limestone deposition during times of volcanic quiescence. The lack of pillow basalts in the Fletcher Park and Green Mountain areas suggest the rocks were deposited in very shallow water, or at times, subaerially. Metasediments in the Green Mountain area are commonly mature and possibly craton derived. The most striking similarity between the deposits of this study and polymetallic deposits is their occurrence at the top of felsic cycles and their association with felsic breccia.

Proposed depositional environment

Massive base-metal sulfide deposits are formed at discharge vents of submarine hydrothermal systems (Sangster, 1980). Hot metal-bearing fluids flowing from these vents mix with sea water, and the precipitation of sulfides is caused by changes in temperature, pH, and pressure that result from either mixing or boiling (Franklin and others, 1981). The presence of polymict agglomerate and stringer sulfides suggest boiling of hydrothermal fluids and steam explosions were common during formation of base-metal massive sulfide deposits. Sangster (1980) described two types of sulfide deposits: proximal and distal. Proximal deposits are precipitated around the fumerolic vent and are associated with boiling, explosion breccia and agglomerate. Distal deposits lack signs of explosive volcanism and occur in layers over a wide area. Density of the metal-bearing fluid controls whether the deposit develops into a proximal or distal type. Fluids that are more dense than sea water would flow downhill and collect in depressions near the vent. Fluids that are less dense than the sea water would gradually increase density during mixing and the deposition of sulfide would occur as a thin layer over a large area away from the vent. In the Green Mountain and Fletcher Park areas the deposits generally occur as uniformly thick layers that extend over 0.5 miles and are associated with brecciated and calc-silicate rocks. This suggests both proximal and distal deposits occur in the Green Mountain and Fletcher Park areas.

In summary, the stratabound sulfide deposits in the Green Mountain and Fletcher Park areas were formed in a complex volcano-sedimentary pile and are classified as the polymetallic type on the basis of bimodal affinity; shallow water deposited, craton derived sediments; and occurrence of sulfides and brecciated rock at the top of mafic and felsic cycles.

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On the reactions leading to the formation of magnetite in ferruginous rock and their implications for the classification of iron formations

B. Ronald Frost

Recent studies of phase relations in metamorphosed iron formations indicate that the protoliths of most of these rocks probably contained hematite as the major iron oxide rather than magnetite. Petrographic studies of weakly metamorphosed iron formations also show ample evidence for the reaction of greenalite and siderite to magnetite. Due to the extremely low abundance of free oxygen in metamorphic fluids, any conversion of silicates or carbonates to magnetite must have occurred as the result of oxygen-conserved reactions. Two of these reactions may have occurred as internal buffering of the fluid drove the oxygen fugacity down to the hematite-magnetite buffer. As the fluid crossed the buffer these reactions would have ensued:

(1) greenalite + 3 hematite = 3 magnetite + 2 quartz + $2H_2O$

(2) siderite + hematite = magnetite + CH_4

If siderite and greenalite were still present in the rock after reactions (1) and (2) had gone to completion, additional magnetite may have formed by the reaction:

(3) greenalite + 9 siderite = 2 quartz + 4 magnetite + $8CO_2$ + CH_4

Reaction (3) will occur only as long as conditions are sufficiently reducing to allow CH_4 to form. As the reaction proceeds, however, the conversion of the iron component to magnetite will progressively enrich greenalite in magnesium. This will cause the buffering surface for the assemblage greenalite-magnetite-quartz to rise to higher oxygen fugacity and eventually this upward shift will cause reaction (3) to cease. Once this happens no further magnetite will form, and the remaining reactions occurring with prograde metamorphism of iron formation will involve only dehydration and decarbonation. The primary effect of these higher-grade reactions is to convert siderite to silicate, a process which for most bulk compositions is complete by lower amphibolite facies.

Recognition that these processes occur in low-grade metamorphism of iron formation leads to the observation that the traditional classification of iron formations into oxide, silicate, carbonate, and sulfide facies can lead to gross misinterpretation of the sedimentary environment of some rocks. For most iron formations the primary sediment either consisted of interlayered hematitic chert and siderite (i.e. the Algoman type iron formation) or of interlayered hematitic chert-siderite and greenalite (i.e. the Lake Superior type iron formation). Whether an oxide, silicate, or carbonate iron formation formed during diagenesis of these rocks would have been as much a function of the original abundance of the primary phases and degree of metamorphism as of the sedimentary environment. For example, an "oxide-facies iron formation" could have formed from a sediment that was originally poor in siderite relative to hematite while a sediment from the same environment that was richer in siderite could have formed a "carbonate-facies iron formation" during diagenesis. Primary magnetite and iron sulfides may have occurred in some iron formations, but they are likely to have been restricted to exhalative environments where they could have formed either as a primary deposit on the distal portions of a fumarolic center, or as the result of incomplete suboceanic weathering products of massive sulfides.

The geology of the Homestake mine, Lead, South Dakota

Robert D. Woodfill

The Homestake mine at Lead, South Dakota is a world-class gold mine. Since it began production in 1876, the mine has produced over 35 million troy ounces of gold, worth $16 billion at present prices.

The ore at Lead is contained within Precambrian metasedimentary rocks that total 17,000 feet thick. The Poorman Formation, the oldest, is an ankerite-bearing phyllite. The overlying Homestake Formation is a quartz-sideroplesite schist altered to cummingtonite schist at higher metamorphic grades. The overlying Elliston Formation is comprised of quartzites and interbedded quartz-mica phyllites. Younger formations are argillaceous with some quartzites. The Cambrian Deadwood Formation unconformably overlies the Precambrian.

The dominant structure at Lead is an elongated dome about 10 by 12 miles long whose center is cut by an Eocene stock. Isoclinal folds in the Precambrian rocks plunge 10° to 45° SE, with axial planes dipping slightly to the east. At least eleven deformational events have been recognized at the Homestake mine.

Progressive metamorphism of the Precambrian rocks at Lead increases from the southwest to the northeast from the biotite zone, through the garnet zone, to the staurolite zone. Ore deposits are found only in the biotite and garnet zones.

Two distinct types of mineralization are recognized in the northern Black Hills: Homestake-type ores that are restricted to Precambrian host rocks, and Tertiary replacement ores that are restricted mostly to Paleozoic host rocks. At the Homestake mine, major ore bodies, termed ledges, are a group of ore shoots, each pencil-like in shape and parallel to the plunging folds to which they are confined as replacements of the Homestake Formation. These ore bodies are in highly chloritized parts of the Homestake Formation and are associated with abundant veins and masses of quartz with pyrrhotite, pyrite, and arsenopyrite. Gold is generally associated with arsenopyrite.

Rye and Rye (1974) showed that stable isotope data are consistent with a sedimentary Precambrian origin for the major ore constituents with later migration of syngenetic components during metamorphism.

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Review of kimberlite exploration and evaluation methods

M.E. McCallum and James B. Lincoln

Kimberlite is the principal primary source of naturally-occurring diamonds, and as such, it is the target of extensive exploration activity. Exploration for kimberlite is an integrated effort that involves a variety of techniques. Effectiveness of any given method varies as a function of the size; geologic, geographic and climatic nature of the target area; and whether kimberlite is known to be present. Regional studies are designed primarily to establish districts favorable for more detailed evaluation. Airborne surveys are most applicable regionally, and comparison of similarly scaled LANDSAT (infrared reflectance) and radar images and high-altitude aerial photographs with aeromagnetic, gravity, geologic, topographic, and drainage maps, along with any other available comparable data that may provide information on the surface and subsurface character of an area, may reveal the presence of intersecting or interrupted regional trends. This textural interference may reflect deep-seated structures that could have provided the zones of weakness necessary for the emplacement of deeply derived kimberlite melt.

Airborne techniques are effective locally in identifying individual kimberlite occurrences. Surface and near surface structural trends that may be related to pipe and (or) dike emplacement are readily recognized on a variety of images, and both aeromagnetic and gravity surveys can reveal the location of kimberlite if host rocks are sufficiently different in physical character. Ground geophysical methods appear to be effective only in the delineation of individual kimberlite bodies. Magnetic, gravity, electromagnetic, electrical resistivity, refraction seismic, radioactivity and radiometric surveys have been conducted in a number of kimberlite districts, but generally only the magnetic, electromagnetic, and resistivity methods are useful in defining kimberlite-host rock contacts.

Most effective in the search for kimberlite are heavy mineral surveys that rely on recognition of one or more of the kimberlite indicator minerals (pyrope garnet, magnesian ilmenite, chromian diopside, and chromian spinel) in the heavy mineral fraction of alluvial, colluvial, and eluvial material. These minerals are common only in ultramafic rocks and all but diopside are relatively resistant to abrasion and weathering, thus they persist for considerable lengths of time and over considerable transport distances (miles to tens of miles) except in tropical regions. Definitive identification of the indicator minerals is dependent upon chemical, optical, and (or) x-ray determinations; however, they generally can be recognized by color and magnetic properties. The opaque oxides Mg-ilmenite (picroilmenite) and chromian spinel commonly can be distinguished from their more abundant counterparts ferro-ilmenite and magnetite by their lower magnetic potentials, although more iron-rich varieties of both indicators may be appreciably magnetic. Chromian diopside is readily recognized by its emerald green color, and its presence indicates relatively close proximity to the source. However, this indicator mineral also is characteristic of a variety of locally abundant non-kimberlitic rocks (e.g., alpine peridotites). Pyrope garnets generally can be recognized by their red-purple, burgundy, to deep greenish purple colors, although some may be characterized by lighter shades of red and purple. Other garnets derived from deep-seated xenoliths transported to the surface in kimberlite magma also may be useful locally in exploration if no other deep-source intrusive rocks are present. For example, many pink, reddish orange to reddish brown, and brown garnets are characteristic of websterites, pyroxenites, eclogites, and granulites that may represent xenolithic assemblages. A detailed color classification of garnet grains using a Rock Color Chart is a rapid and useful tool in defining garnet populations that although not "kimberlitic", might be related to associated xenolith suites.

Geochemical surveys are useful in some districts, but generally only as part of detailed evaluations of target areas established by other means. Kimberlites are usually enriched in such elements as Mg, Ca, Ni, Co, Cr, Ti, Nb, and rare earths, but use of these elements as pathfinders is dependent upon their dispersion into overlying and adjacent materials and their insignificance or absence in host rocks.

Final ground surveys in target areas utilize a variety of surface characteristics to establish the presence of kimberlite. Of particular value are linear structural controls and the intersections of cross-structures; the presence of contrasting soils (e.g., green to gray montmorillonite-rich vs. brown to orange kaolinite-rich); sharp vegetative changes; nodules or exotic rocks (e.g., in Colorado-Wyoming kimberlite district, rounded xenoliths of garnet peridotite, eclogite, websterite, granulite, and lower Paleozoic sedimentary units occur in weathered granitic material); nodules and xenocrysts of kimberlite minerals (e.g., pyrope, magnesium ilmenite, and diopside); and localized caliche deposits in non-carbonate terrains.

Many models have been proposed to relate various parameters of a kimberlite to its diamond content. Most of these models suggest a correlation between the Cr, Ni, Ti, and Fe content of the kimberlite and the presence or absence of diamond. Although the indicated trends may apply locally, it does not appear that such models have regional applicability. More recent studies on the chemistry of pyrope, magnesian ilmenite, and spinel demonstrate considerable promise. The Cr_2O_3-CaO content of pyrope and the Cr_2O_3-MgO contents of magnesian ilmenite and spinel in kimberlite appear to correlate reasonably well with the diamond content of individual kimberlite occurrences.

Consideration of the morphological features of diamonds present in a kimberlite may also be of value in predicting the economic potential of a given pipe. Resorption of diamond crystals in kimberlite is an extremely important process, and it is responsible for converting primary octahedron and cube forms to the tetrahexahedroid form (rounded dodecahedron). Robinson (1980) demonstrated that this conversion process involves a mass loss of at least 45 percent and that there is a correlation between the proportion of diamonds with octahedral and (or) cubic surfaces and the diamond content of the host kimberlite.

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Remote sensing techniques applied to kimberlite exploration in southeast Wyoming and north-central Colorado

Janet E. Marks and R.W. Marrs

Introduction

M.E. McCallum first recognized kimberlite in the Colorado-Wyoming region in 1964 (Chronic and others, 1965). Since that time, approximately 100 kimberlite occurrences have been identified in the Laramie Range of southeastern Wyoming and the Front Range of north-central Colorado. At least 15 of these are now known to host diamonds (Hausel, 1985, p. 10).

Kimberlite diatremes in the Colorado-Wyoming region are difficult to locate using conventional geological methods because the terrain is mountainous, the field season is limited, and exposures of kimberlite are poor. Many common exploration techniques -- geochemical and mineralogical alluvial sampling, field mapping, and geophysical methods (gravity, conductivity, magnetics and seismic reflection)--are time-consuming and costly. Multispectral remote sensing techniques, the subject of this report, are relatively fast and economical. The major objective of the research described here was to develop techniques for detecting poorly exposed kimberlite diatremes using an eight-channel multispectral scanner (NS-001) as a remote-sensing exploration tool. Data from the scanner were digitally enhanced and then interpreted to define target areas to be field checked. Field checks include sampling and geophysical surveys.

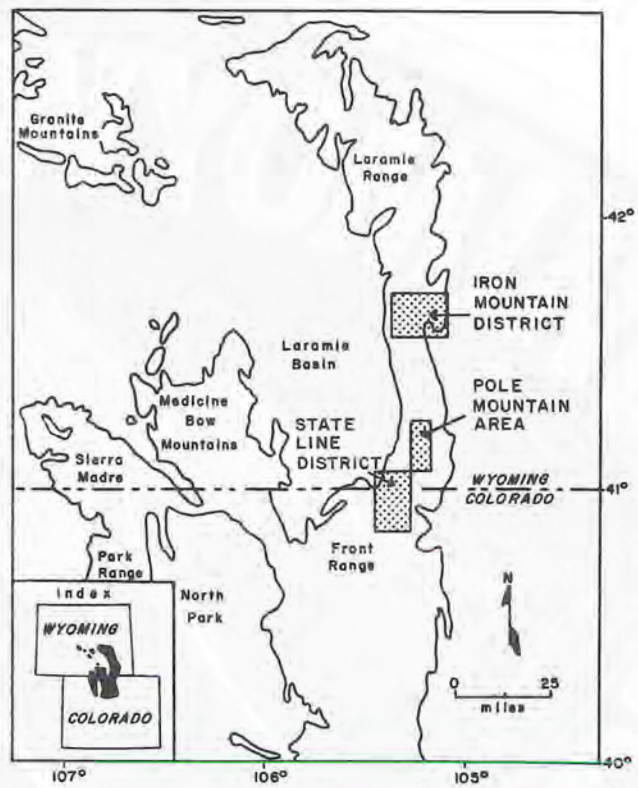


Figure 1. Location map of kimberlite diatremes and test areas.

Three test areas (Figure 1) in southeastern Wyoming and north-central Colorado were selected on the basis of their proximity to known kimberlite occurrences (Hausel, 1985). Scanner coverage and false-color infrared photographs for these areas were obtained from the National Aeronautics and Space Administration (NASA) through the NASA/Ames medium-altitude aircraft program.

Control sites were chosen in areas where known kimberlite diatremes are exposed. With the infrared photographs as location guides, multispectral image data for the control sites were selected and corresponding images were processed to enhance the spectral contrast of kimberlitic material. Using interpretation techniques that best expressed the kimberlites in the control sites, potential target areas were identified. These new target areas were checked by field examination and electrical and magnetic surveys.

Our approach was to (1) analyse the unique spectral and spatial surface characteristics of kimberlite diatremes; (2) define a sequence of digital image processing techniques that would best discriminate kimberlite diatremes from the surrounding granitic country rock; (3) identify several target areas where hidden kimberlite pipes might be found; and (4) confirm and map -- or disprove -- the occurrence of kimberlite in these target areas on the basis of field observation, geophysical surveys, and mineralogical and geochemical analyses.

We hoped to discriminate kimberlite from the surrounding country rock on the basis of geomorphology, structural relationships, and spectral characteristics. Kimberlite shows such distinguishing characteristics as elliptical plan, association with faults or fractures, bluish soil color, lack of tree cover, lush grassy vegetation, and clay-rich soil that holds abundant moisture. To the degree that these unique characteristics are discernable, kimberlite may be detected by interpretation of the digitally processed multispectral scanner data, even if the kimberlite is very difficult to detect on the ground.

The area covered in one digitally reconstructed scene is a little over one square mile. The spatial resolution is approximately 12.5 feet on the ground. The NS-001 scanner provides spectral images in the range of 400 to 2,350 nanometers (nm) and a thermal infrared band (9,000 to 11,000 nm).

Digital processes were used to emphasize spectral and spatial contrasts between kimberlite and its host rock. Spectral band ratios and principal-component color composites were found to be the most valuable process products for discriminating kimberlite. When the spectral signatures of the known kimberlite diatremes (control sites) had been determined on the processed images, other target areas could be identified on the basis of similar spectral and spatial expression.

Regional Precambrian geologic setting

The southern Laramie Range is a north-trending, broad, asymmetrical anticline with highly jointed Precambrian crystalline terrain exposed in the core. The Precambrian crystalline core consists mainly of older, complexly folded granite gneiss, intrusive granite, anorthosite, and syenite, and minor volcanogenic metamorphic rocks.

A major age discontinuity divides the Precambrian rocks of the Laramie Range and continues to the west, through the Medicine Bow Mountains and Sierra Madre, where it is marked by the northeast-trending Mullen Creek-Nash Fork shear zone (Houston and others, 1968; Hills and others, 1968). The Laramie Range anorthosite-syenite complex emplaced around 1,510 $\pm$ 30 m.y. ago (Hills and Armstrong, 1974), masks evidence of the shear zone in the Laramie Range. Its emplacement may have been controlled by the shear zone.

The Sherman Granite batholith intruded the Laramie anorthosite-syenite complex on its southern and eastern border approximately 1,335 $\pm$ 30 m.y. ago (Hills and others, 1968) or 1,430 $\pm$ 60 m.y. ago (Hills and Houston, 1979). Within the Precambrian Sherman Granite terrain are small exposures of gneissic rock dated at 1,750 m.y. ago (Peterman and others, 1968).

Diatreme geology

The known kimberlite diatremes penetrate the Precambrian crystalline terrain and are scattered in a north-trending region coincident with the trend of the Laramie Range. They extend north to Sybille Canyon, approximately 45 miles north of the Colorado-Wyoming State Line, and south to Boulder, Colorado, approximately 70 miles south of the State Line. The kimberlite pipes are dated between latest Silurian and Early Devonian on the basis of Ordovician and Silurian sedimentary rocks incorporated in the diatremes. Naeser and McCallum (1977) report an age of 377 $\pm$ 9 m.y., based on fission-track dating of the kimberlites. This date is consistent with the Devonian age implied by the inclusions.

The State Line district in the Laramie Range (Front Range) (Figure 1) is a north-trending zone, 17 miles long by 5 miles wide. This district contains approximately 40 kimberlite diatremes that penetrate the Precambrian Sherman Granite and Log Cabin Granite batholiths. The Sherman Granite includes the Trail Creek Granite and the Inner Cap Rock quartz monzonite. Much of the State Line district (including the Schaffer pipe area discussion below) lies within the Virginia Dale ring-dike complex, a circular structure 8 miles in diameter within the Sherman Granite batholith (Eggler, 1968). The Sherman Granite terrain in this vicinity shows a dominant north-northwest structural trend of joints and diabasic dikes (Hausel and others, 1979a; 1979b; 1981).

The Schaffer cluster of kimberlites is located in the State Line district in Wyoming. The kimberlite diatremes are emplaced on a N30°W - N40°W trend (Hausel and others, 1979a; 1979b; 1981). Because of this linear pattern, their emplacement is suspected to be controlled by northwest-trending shear zones. Approximately 20 pipes occur in this cluster; they intrude the Inner Cap Rock quartz monzonite (Hausel and others, 1981).

The Schaffer pipes are relatively small, averaging 85 to 120 feet wide by 130 to 280 feet long (Puckett, 1971). The largest, the Schaffer 13, is 1,800 feet in the maximum horizontal dimension. These diatremes were poorly exposed, but most have been extensively prospected since 1980 by Cominco American, Incorporated and are now partially revegetated.

Because of its large size, relatively easy access, and good exposure, the Schaffer 13 pipe was used as an initial control site for determining the best techniques for kimberlite detection using digital image processing. Ground measurements were taken at this control site both before and during the overflight.

Description of kimberlite and its surface expression

Dawson, (1971, p. 188) defined kimberlite:

Kimberlite is a very rare, potassic, ultrabasic, hybrid igneous rock that occurs in small diatremes, or in dykes or sills of limited extent. It has an inequigranular texture, the porphyritic aspect being due to megacrysts of olivine, enstatite, chrome diopside, pyrope, picroilmenite and phlogopite, set in a finer-grained matrix of which serpentine, carbonates, phlogopite, magnetite and perovskite form the major part. Many of the megacrysts are derived from fragmentation of mantle-derived garnet lherzolite (blocks of which are embedded in the kimberlite) and are in various stages of reaction with the kimberlite matrix. The matrix may, or may not, contain diamond; even in the most diamondiferous kimberlites, diamond is a very rare and widely dispersed mineral.

Nodules derived from the upper mantle and lower crust are often present in the kimberlitic matrix. These include garnet and spinel peridotite, eclogite, granulite and monomineralic xenocrysts. The olivine and orthopyroxene are usually completely serpentinized, leaving only relict outlines.

Kimberlite is always altered, although the degree of alteration varies depending upon depth and on whether the pipe records repeated eruptions or a single intrusive event. The alteration processes typical of kimberlite are hydration (serpentinization and clay formation), carbonatization, and fenitization.

Kimberlitic soil (soil derived from kimberlite) in the study area is generally devoid of tree growth, while the host granitic rock is commonly forested with Ponderosa pine. The kimberlite is typically covered by relatively thick, lush, grassy vegetation and scrub brush. However, many grassy, non-forested areas are not kimberlitic. Kimberlitic soil tends to be moister than granitic soil because the clays, especially smectite (montmorillonite), in kimberlite absorb water. Granitic soil tends to dry more quickly and to support xeric vegetation. The contact between well-exposed kimberlite and the host rock can sometimes be defined by vegetation differences. For detection of vegetation contrasts by remote sensing, data should be obtained when the moisture and vegetation differences are most pronounced, in late spring or early summer.

Kimberlite is unstable at surface temperatures and pressures, and weathers more easily and deeper (10 to 100 feet), than the surrounding granite. Granite typically forms massive outcrops, while kimberlite does not crop out prominently. Kimberlite decomposes and erodes readily to produce topographically low areas, except where it has undergone silicic alteration (Smith, 1977).

Kimberlite dikes and diatremes in the study area are quite small, varying from a few feet to nearly 1,800 feet wide. They are typically elliptical to elongate in plan, with their long axes parallel to prominent fractures (Mannard, 1968). The close association of fractures and kimberlite suggests that their emplacement is often structurally controlled by preexisting faults and fractures.

The saprolite developed on kimberlite is generally greenish gray or bluish gray due to the weathering products of serpentine, smectite, chlorite, calcite, and talc. Although kimberlites are commonly covered by soil derived from the surrounding host-rock terrain, badgers and other burrowing animals often expose this "blue ground" in their burrowing. Some kimberlite bodies in the study area that contain calcareous lower Paleozoic sedimentary xenoliths show a white saprolite coloration. Nodules of garnet and spinel peridotite, eclogite, granulite, and monomineralic xenocrysts derived from the upper mantle and lower crust may also be present. These nodules are generally ellipsoidal and 2 to 10 centimeters in size. They are usually altered by hydration to serpentine and smectite.

All of these characteristics of weathered kimberlites are readily detectable on remote sensing images of good kimberlite outcrops, but are not detectable on images of kimberlitic bodies that are poorly exposed or covered by a layer of soil. Consequently, it is critical to distinguish minor amounts of kimberlitic soil mixed with soil derived from granite, or to distinguish subtle vegetation contrasts. Multispectral data, enhanced as described below, can provide discrimination of these very subtle soil and vegetation contrasts and reveal covered kimberlite locations.

Description of imagery

The 8-channel NS-001 multispectral scanner is an airborne, scanning radiometer designed to record reflected radiation in spectral bands that respond to subtle variations in the character and condition of rocks and soils. It has an instantaneous field of view (IFOV) of 2.5 mrad. At an altitude of 5,000 feet above mean terrain it provides spatial resolution approaching 12.5 feet and a total swath width of approximately 8,740 feet. The total, usable scan-angle is 100 degrees, and there are 699 pixels per scanline oriented perpendicular to the flight direction.

The scanner images in the seven Landsat-D Thematic Mapper bands plus a band from 1.00 to 1.30 micrometers, band 6 (Table 1). These bands can be combined in various ways to enhance subtle spectral contrasts.

Table 1. Description of the NS-011 bands. (After Freden and Gordon, 1983, p. 548.)

Band	Type of radiation	Wavelength (in microns)	Principal applications
1	Visible blue	0.45-0.52	Differentiates water, soil and vegetation.
2	Visible green	0.52-0.60	Shows high reflectance of healthy vegetation.
3	Visible red	0.63-0.69	Absorbed by chlorophyll.
4	Near-infrared	0.76-0.90	Delineates water bodies that absorb electromagnetic radiation in this band. Ferrous iron also shows an absorption in this band due to electronic transitions.
*5	Near-infrared	1.00-1.30	Chlorophyll in healthy vegetation shows a high reflectance in this spectral region.
6	Near-infrared	1.55-1.75	The peak rock reflectance band; vegetation moisture detection.
7	Near-infrared	2.08-2.35	The clay band, discriminates hydrothermally altered rocks from unaltered rocks, hydroxyl ions in clays and other minerals show a strong absorption near 2.20 due to Al-hydroxyl vibration or 2.30 due to Mg-hydroxyl vibration.
8	Thermal infrared emission	10.4 -12.5	Sensitive to moisture and vegetation changes.

* Not used on Landsat-D Thematic Mapper

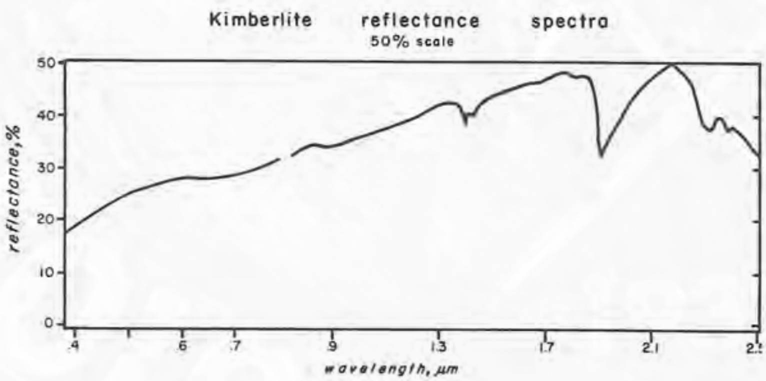
Color infrared photographs were also acquired over the study area with a Zeiss camera while the eight-channel Thematic Mapper Simulator (multispectral scanner) gathered multispectral information. Flights were flown in late June and early July, 1982. The aircraft was flown at an average altitude of 5,000 feet above mean ground elevation to yield adequate resolution with the scanner. Ground data were acquired before and during the overflight.

The color infrared photographs supplement the digital scanner data by providing location and spatial reference. Earlier work indicated that color infrared photographs are very useful for general mapping, although they are not generally amenable to the detection of poorly exposed kimberlite in this region (McCallum, 1974).

Interactive digital processing and enhancement techniques

Processing procedures were selected on the basis of trial application in areas of known kimberlite. The selection procedure is as follows: (1) a control site is first defined that includes a known kimberlite diatreme; (2) the unique surface expression and spectral characteristics of kimberlite are then determined in an examination of the photographs and unenhanced images; and (3) digital image processing techniques are found that enhance subtle spatial and spectral characteristics of the kimberlitic soils (Short 1982, p. 145-178 and 433-444).

The moderately well-exposed Schaffer 13 reference area, located on the Wyoming-Colorado border, served as an initial control (calibration) site from which to define distinct spectral contrasts between granitic rocks and kimberlitic soils (Figure 2). This kimberlite has been disturbed by mining activity but is now mostly revegetated.



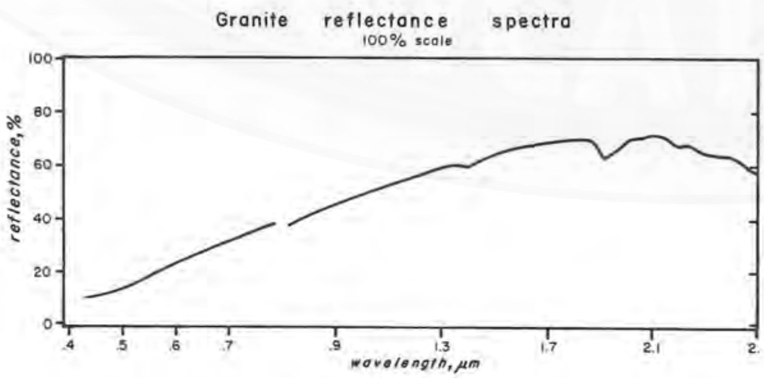


Figure 2. Reflectance spectra of kimberlitic and granitic soils and rocks. (From Marks and Marrs, 1983).

Analysis of single bands of imagery

Analysis was applied to single spectral bands selected from the 8-band image set. Because many spectral reflectance differences are subtle, interpreting low-contrast differences is important. Interpreting the images is much easier when they are contrast-stretched. To accomplish a useful contrast-stretch, a histogram is produced representing the frequency of occurrence of various pixel values. The density-level distribution is then expanded to take advantage of the full dynamic range of the image processor (0-255 digital grey-levels).

The clay band (band 7) is the most diagnostic single band, because soil derived from kimberlite contains smectite. Smectite, with its unique crystal structure, contains abundant hydroxyl ions. When a hydroxyl ion is bonded to a metal cation, vibrational processes absorb electromagnetic radiation in this wavelength range, and produce an absorption anomaly over the kimberlite when the clay band is displayed. Therefore, kimberlite appears dark in the contrast-stretched band-7 image.

Analysis of combinations of two bands was applied using arithmetic operations. Ratioing is the most common and useful combination technique because it diminishes strong contrasts while enhancing subtle spectral detail. These, however, do not appear to be as useful for lithologic discrimination as ratioing. Other divariant analysis techniques are summing, subtracting, and multiplying. However, these do not appear to be as useful for lithologic discrimination as ratioing. Ratioing is performed by dividing the intensity level in one band by the intensity level in a second band, pixel by pixel. The resulting quotients fall over a very narrow range so contrast-stretching is routinely performed to enhance the ratio combinations. Higher values are assigned to lighter tones.

Ratio images have the unique characteristic of enhancing subtle spectral reflectance differences relative to first-order radiance differences. Instrument noise is muted. Therefore, materials that are spectrally similar in the ratioed bands will appear the same regardless of albedo (intensity) variations. Areas that are spectrally unique show up as stronger contrasts against the normalized background. Because rocks are often distinguishable on the basis of subtle spectral contrasts, ratioing is an extremely useful technique for discriminating among rock types.

The choice of the multispectral bands to be used depends on the spectral properties of the rock types to be distinguished. For kimberlite detection, ratios of bands 4/3 (near-infrared/red) and 6/7) (peak rock/clay) appear to be the most effective. Kimberlite has moisture-retaining clays that absorb much of band 7 radiation, so kimberlite appears dark. Vegetation appears light, and granite yields an intermediate response on the 6/7 ratio. Kimberlite sometimes shows high reflectivity in the near-infrared spectral bands due to the presence of lush, grassy vegetation cover. Therefore, kimberlite should appear light and granite dark on a 4/3 band ratio.

Some enchancement techniques involve manipulation of three or more spectral bands. Color composites, stretched-ratio color composites, and principal component analysis fall into this category. Any of the previously discussed techniques may also be performed on the resulting multiband composite images.

Color composites are constructed by assigning three different images to the red, green, and blue hues of the video display. Ratio color composites are produced by assigning each ratio of two bands a different color on the video display.

In principal component analysis (factor analysis), the computer selects the best combinations of bands with which to display the brightness differences in all bands of the multiband scene. Principal component computations effectively reduce the number of spectral bands needed to represent variations in the data. The number of images that must be evaluated simultaneously is reduced without losing any essential information.

Principal component analysis determines an array of correlation coefficients between eight sets of data. It then constructs a smaller set of variables (components or factors) by mathematical transformations of the original eight-band data. The components are selected to be mathematically orthogonal, or independent of one another. The process then rearranges the data by rotation and translation of the spectral bands to maximize the information content of the first factors constructed. The last factors can be discarded. With eight-channel data, the essential information is condensed to four or five principal components. Principal component images may be displayed on the image processor monitor either individually or color-combined to produce a principal-component color composite.

The fourth principal component (factor) seems to be best for discriminating kimberlitic soils from granitic soils (Figure 3). This factor usually reveals the very subtle differences between rock and soil types. Principal component color composites that incorporate the fourth principal component were selected as the most interpretable of the enhanced image products.

Results

All of the standard digital image-enhancement techniques discussed above were examined to see which produced the optimum image product for discriminating kimberlitic soils from the surrounding host rock. The Schaffer 13 kimberlite was used as the initial control site. Principal component analysis was selected as the most effective standard image enhancement technique. The fourth principal component offers the most distinctive image display of kimberlites. The fourth principal component (Figure 3), and principal-component color composites that include the fourth principal component, were examined for zones that appeared anomalous and similar to spectral features associated with the known kimberlite diatremes in the Schaffer control site.

Twenty possible anomalous zones were defined by interpretation of several enhanced-image products (Figure 4). Eight of these interpreted anomalies were correlated with known kimberlites in the area, and twelve were not. Upon review of the interpretation with W.D. Hausel of the Geological Survey of Wyoming, three of the twelve anomalies (8, 9, and 12) (Figure 4) were found to be coincident with field-reported occurrences of kimberlite in areas that have not been mapped because access to the property is restricted.

The remote sensing anomalies 1 and 2 (Figure 4) were examined in the field with electromagnetic and magnetic equipment to confirm the occurrence of kimberlite and map the subsurface distribution of each diatreme or disprove its presence. Both of these anomalies yielded negative results.

Field evidence indicates a success rate of 60 percent for detection of kimberlitic bodies by this method. At this time we have field information for only five of the twelve identified target areas. Features 3 through 7, 10, and 11 have not yet been verified as kimberlite, nor has the occurrence of kimberlite been disproved.

The results of this project were gen-

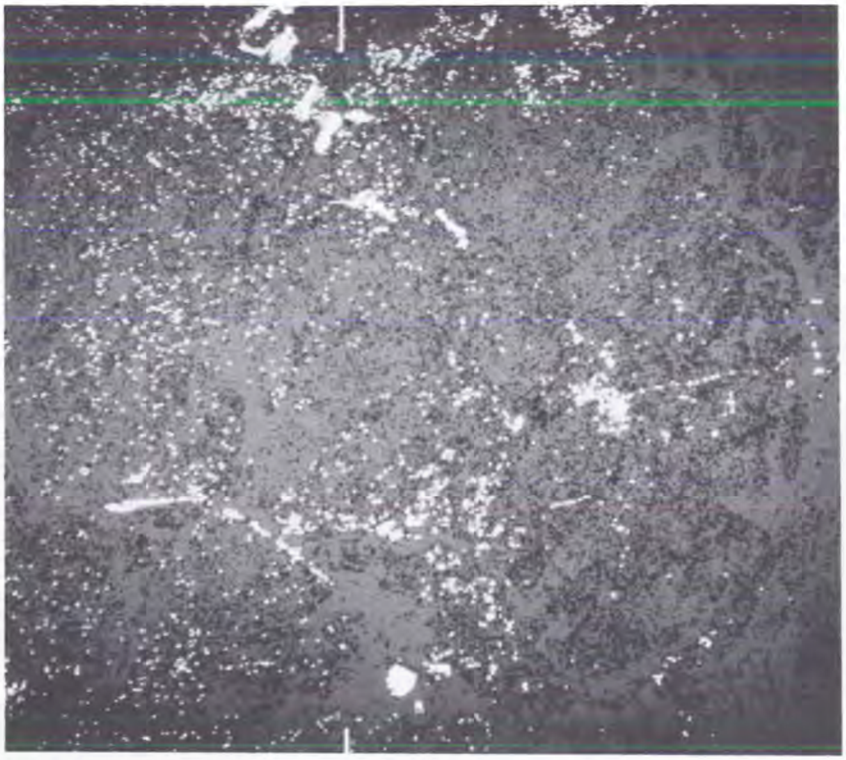


Figure 3A. Reconstructed image of the fourth principal component for the State Line area TMS data. Density slice of fourth principal component. Kimberlites, road, lakes, and barren (disturbed) areas appear light toned.

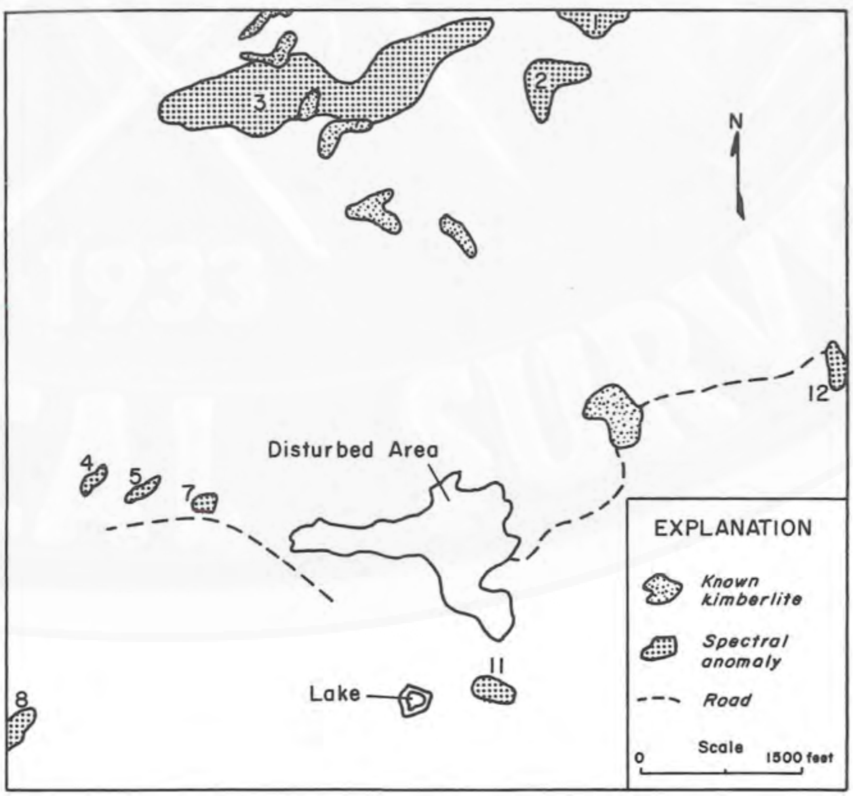


Figure 3B. Interpretation of fourth principal component. Both known kimberlites and areas interpreted as possible kimberlite occurrence are shown. Shape of the imaged area differs slightly from Figure 4 due to scan-angle distortion.

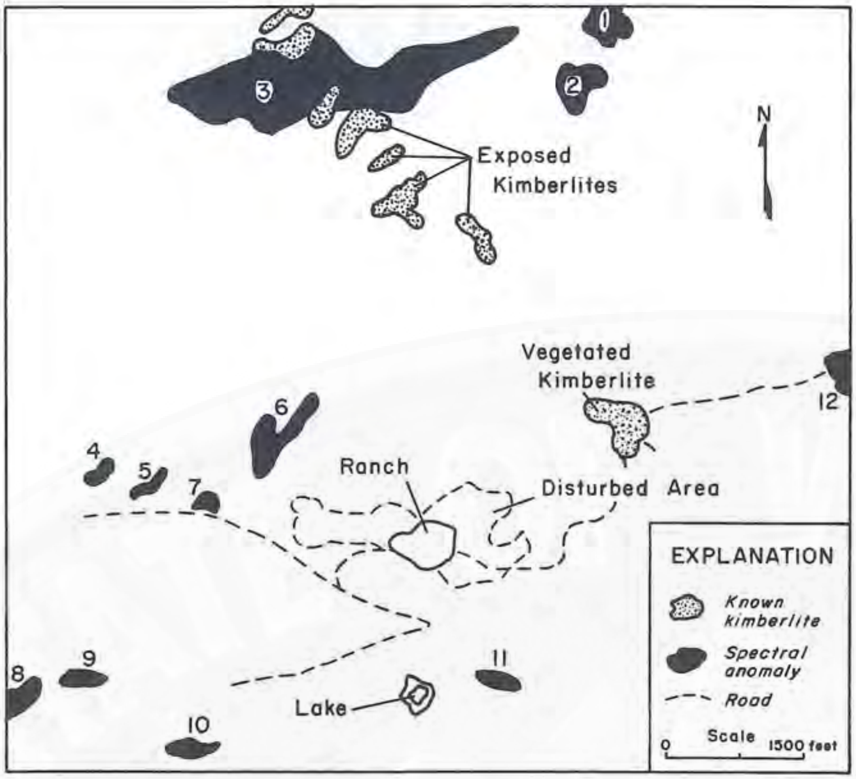


Figure 4. Interpretation of the color composites of principal components 2, 3, and 4 and 3, 4, and 5. Both known kimberlites and areas interpreted as possible kimberlite occurrences from spectral anomalies (numbers 1-12) are shown. (From Marrs and others, 1983.)

erally encouraging. Computer-processed multispectral imagery has excellent potential as a tool for regional reconnaissance for kimberlite intrusives. Both ratioing and principal component analysis of the eight channel multispectral data effectively enhance the subtle contrasts in soils and vegetation.

Acknowledgments

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We especially wish to thank W.D. Hausel for numerous discussions on kimberlite emplacement and generation, and Reid Fletcher for help with image analysis.

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Lake Alice Copper district, Lincoln County, Wyoming

W.W. Boberg

The Lake Alice district is within the Overthrust Belt of western Wyoming about 20 miles north of Cokeville, Wyoming. The principal prospect of the district is the Griggs mine, which was discovered in 1895. The Griggs mine consists of several adits into sediments of the Triassic/Jurassic Nugget Sandstone and the Gypsum Spring Member of the Jurassic Twin Creek Limestone. It had minor production between 1914 and 1920. Bleached-white sandstones of the normally reddish-colored Nugget Sandstone; and dirty, gray, silty sandstones and petroliferous, cherty dolomites of the Gypsum Spring Member of the Twin Creek Limestone contain extremely variable concentrations of copper, lead, and zinc sulfides with some associated silver.

The Griggs mine prospect contains the greatest observed thicknesses and highest grades of sulfide mineralization in the district. Sulfide mineralization was apparently concentrated in a structurally-prepared zone on the flank of an anticline. This is the only area where sulfides can be observed in bleached-white Nugget Sandstone. Outcrops of bleached-white Nugget Sandstone occasionally contain numerous cavities that probably represent the complete leaching of sulfide minerals at the surface.

Sulfides are nearly ubiquitous in low but observable concentrations within specific units of the Gypsum Spring Member of the Twin Creek Limestone. A locally continuous, two- to three-foot thick, gray to yellow-gray, silty sandstone within the Gypsum Spring Member contains low-grade copper (0.2 percent Cu) and silver (0.1 ounce/ton Ag) wherever it is found in an area of more than two square miles. Another thin (two to five feet thick), discontinuous, cherty, petroliferous dolomite and limestone contains variable amounts of lead and zinc as well as some silver.

It is believed that the sulfide occurrences at Lake Alice formed solely by sedimentary and hydrodynamic processes in much the same manner as the major stratiform sulfide deposits of central Europe. The metals probably precipitated from a briny formation fluid developed within the red-bed sequence of the Nugget Sandstone and underlying Ankareh Formation. The presence of a petroliferous shale, limestone, or dolomite above the red-bed sandstones was important in limiting fluid flow as well as in providing varying chemical environments that affected the composition of briny formation fluid. The type and amount of metals available in the red-bed sequence determined the availability of metals in the briny formation fluid. During diagenesis and burial, deep-basin hydrodynamic flow moved the variously metal-enriched briny formation fluid upward into and through the less permeable, reducing, reactive petroliferous and limy sediments. Precipitation of metals carried by the formation fluid occurred at or near the formation contact. Specific facies of the overlying sediments (Gypsum Spring Member) tended to localize greater amounts of metals or were more consistent in localizing metals.

Another structurally-controlled phase of mineralization occurred during Laramide time when overthrusting caused intense variations in hydrodynamic flow in the Nugget Sandstone, the formation of folds, and the opening of faults near the crest of anticlines. It appears that the metal content of the initial ore-forming briny formation fluid was not sufficient to develop large, blanket-like, sulfide deposits, and it took the second stage of structurally-controlled fluid flow to develop small deposits of higher grade, such as the Griggs mine prospect. Hydrocarbons within the Nugget Sandstone were subject to the same pressures and hydrodynamic flow during overthrusting and may have been involved in the movement of metals into the later, structurally-controlled sites of deposition.

Further thoughts on the genesis of platinum reefs of the Stillwater/Bushveld type

Craig S. Bow

Platinum group metal (PGM) mineralization within the Stillwater Complex occurs in a distinctive portion of banded zone stratigraphy, coincident with the reappearance of olivine in the crystallization of the complex. The immediate host rocks to the mineralization are distinguished by variable proportions of olivine, orthopyroxene, plagioclase, and augite. Hornblende, phlogopite, apatite, sulfides, and chrome-spinel are accessory.

Significant concentrations of PGM within the reef are restricted to sulfide minerals. The sulfides show clear evidence for an origin as accumulated, immiscible sulfide liquid droplets. The high concentrations of PGM in a subordinate weight fraction of sulfide melt, confined to an extremely narrow stratigraphic interval, constitutes the most fundamental constraint on models of genesis.

The presence of olivine- and sulfide-bearing rocks within reef stratigraphy is attributed to the influx of a pulse of primitive magma during a unique interval in the evolution of the chamber. The prolonged interval of plagioclase crystallization resulted in a significant decrease in the residual melt density, to the point where the new pulse of mafic/ultramafic magma was actually buoyant. As a consequence of this density inversion, a diapiric upwelling of hot, but less dense, magma ascended towards the roof of the chamber, rather than flowing across the floor. The critical PGM collector was provided by subsequent saturation and formation of an immiscible sulfide melt. The ascent, ponding, and subsequent decay of the plume, with attendant magma mixing and crystallization, accounts for the high volume of silicate magma to sulfide melt required to explain the high PGM concentrations in the reef. This mechanism also explains, at least in an intuitive way, the complex mineral assemblage and sulfide-silicate textures of the reef.

The fundamental lithologic makeup and distribution of economic mineralization within the Stillwater and Bushveld igneous complexes are strikingly similar. Certainly the most singular and enigmatic similarity between the two intrusions -- from an economic geologist's point of view -- is the positioning of the PGM reefs with respect to the major cumulus phases. This is compelling evidence that very similar mechanisms controlled the evolution of these deposits.

One of the many shared characteristics between the Bushveld and Stillwater intrusions is the presence of thick cycles of ultramafic cumulates in the lower portions of the stratigraphy. Recent research has provided new insight into the origins of cyclic layering in layered intrusions that may affect our understanding of PGM mineralization. The fluid dynamic behavior of successive pulses of mafic magma, introduced into a partly crystallized chamber, has been modeled theoretically, and with low-temperature solution analogs (Huppert and Sparks, 1980). These studies, interpreted in the light of fractional crystallization processes in open-system magma chambers, provide a plausible mechanism for enriching residual melts in PGM prior to the critical pluming event.

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Biogeochemical prospecting in the Stillwater (Pt) Complex, Montana

Walter C. Riese and Gerald K. Arp

Introduction and background

The Stillwater Complex is a large, stratiform, tholeiitic igneous body that was intruded into a stable Archean craton approximately 2,700 m.y. ago. Platinum-group elements are concentrated in the lower part of the banded zone of the complex in a unit referred to as the Howland reef. Significant concentrations of platinum-group elements within the reef are restricted to sulfide minerals; the dominant minerals belong to the braggite-vysotskite [(Pt, Pd, Ni)S] solid solution series.

The Howland reef is continuous over the exposed length of the Stillwater Complex. It strikes approximately N70°W and dips steeply to the north at between 65° to 70° in the area covered by this study. High-angle reverse faults cause the zone of platinum group-bearing sulfides to be repeated in places.

Geochemistry

The geochemical orientation survey performed over the Howland reef portion of the Stillwater Complex in 1982 was designed to test the utility of lithogeochemical, pedogeochemical, and biogeochemical samples in mapping the outcrop/subcrop of this zone. We also hoped that the more platinum-rich zones could be mapped or projected using these media.

Lithogeochemical samples were quickly ruled out as a viable sampling medium because of the paucity of outcrop. Rock chip samples from C-horizon float were also ineffective: this region has been the site of alpine glaciation, and much of the area is covered by kame-terrace deposits which have moved downslope due to solifluction. In short, there is no way to relate the geochemistry of these rock chip samples to the underlying lithologies.

For the same reasons, the pedogeochemical samples were not anticipated to be useful. Our hypothesis, however, was that because the platinum- and palladium-bearing phases were sulfides, these elements should be available to oxidizing ground-water solutions and might migrate sufficient distances to produce break-in-slope anomalies downslope from the more platinum-rich zones. This process of anomaly formation would be further augmented if organic acids were available to the ground-water system; the amount of vegetation present in the area suggests that such acids are available (Huang and Keller, 1970a, 1970b, 1971; Riese and others, 1978). Pedogeochemical samples were therefore collected.

With regard to the biogeochemical samples, several conifer species indigenous to the area offered possible sampling media: Douglas fir *(Pseudotsuga menziessi)*, Ponderosa pine *(Pinus ponderosa)*, Limber pine *(Pinus flexillis)*, and Lodgepole pine *(Pinus contorta)*. Only Douglas fir was sufficiently well distributed to be viable as a sampling medium. Because this was an orientation survey, we did not choose a multiple-species program: it was felt that interspecies normalization might be too difficult for so small a survey.

The survey was conducted in early August of 1982. In all, 65 samples of each medium were collected at 8-foot (25 meter) intervals along lines that were approximately 100 feet (300 meters) apart (Figure 1).

The soil samples were prepared for analysis in Anaconda's Monte Vista, Colorado, facility. This involved drying and sieving to -80 mesh. The prepared pulps were then sent, with standards as well as field and laboratory replicates, to Bondar-Clegg, Lakewood, Colorado. Analyses for copper, zinc, nickel, and cobalt were performed

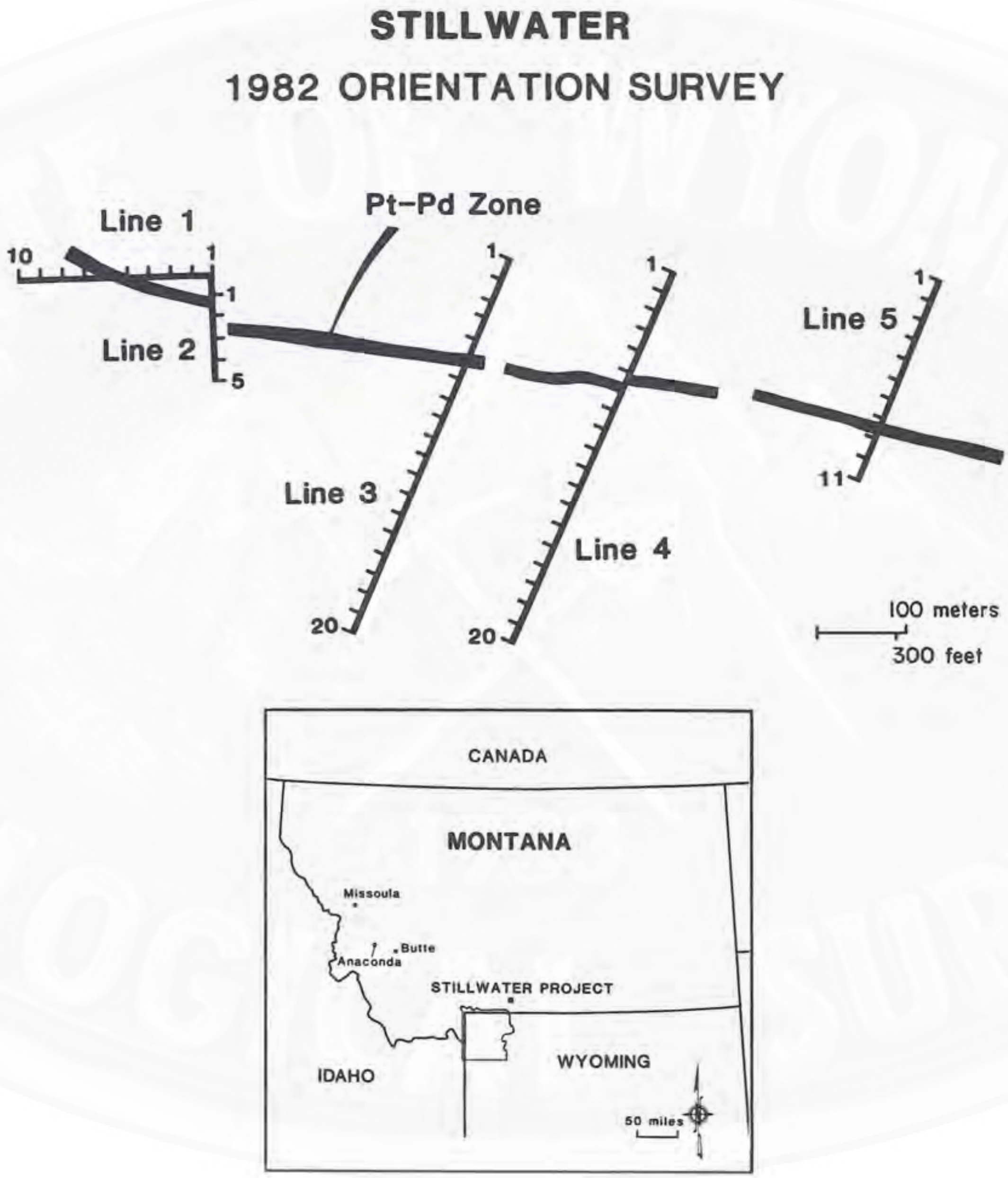

Figure 1. Location of Stillwater project and orientation survey lines. Numbers and ticks on lines designate samples.

using an aqua regia digestion and AA finish; analyses for chromium, platinum, and palladium were performed using either fusion or hydrofluoric digestion and an AA finish.

The biogeochemical samples were all three-tree composite samples of Douglas fir. These were prepared and analyzed at the Arco Exploration and Production Research Center in Plano, Texas. All needles were first separated from the woody material that was to be analyzed; initial analyses demonstrated that the stems were preferential accumulators. Approximately five to seven years of woody material growth was then dried at 110°C, ashed at 550°C, and digested with aqua regia. Palladium, nickel, and chromium were analyzed by ICP.

Discussion

Line plots and topographic profiles of the data are presented as Figures 2 through 4.

Pedogeochemistry

Of the elements analyzed in soils, neither chromium nor nickel was found to be of use in mapping the subcrop of the Howland reef. Palladium was consistently mapped as anomalous over the main zone and offered suggestions that other imbricate zones might be present (Figure 2). This is most strongly suggested in lines 3 and 4, where strong, additional peaks were found across the creek from the main Howland reef zone in areas where the topography precludes the possibility that these are the product of hydromorphic dispersion.

The soil platinum analyses (Figure 3) revealed the location of the main platinum-palladium zone in all lines except line 5. There, there is a weak platinum anomaly several hundred feet downslope from the projected subcrop of the main zone: this may mark the development of a hydromorphically displaced, break-in-slope anomaly.

Biogeochemistry

In general, neither nickel nor chromium was found to be an acceptable or reliable pathfinder for the main platinum-palladium zone of the Howland reef (Figure 4); this is probably due to the overall high background in the ultramafics. The two exceptions to this are the nickel profiles in lines 3 and 4 (Figure 4). In line 3, the nickel in the biogeochemical samples does appear to be anomalous over the main zone. In line 4, the nickel maps quite well define the subcrop of a fault sliver(?) of this zone (compare nickel in Figure 4, line 4 with the biogeochemical platinum in Figure 3, line 4).

The palladium in biogeochemical samples was very difficult to analyze, so the quality of these data may not be as good as it should be. Conversely, lines 1, 3, 4, and 5, all show anomalous palladium concentrations in biogeochemical samples collected downslope of the main Howland reef zone. The only interpretation that can be applied to this distribution is that the palladium, being more readily oxidized and complexed in solution (Pourbaix, 1966), is more mobile and is migrating down the hydrologic gradient. Plants are, therefore, picking up the palladium from ground water and not directly from the rock. This, of course, represents a false anomaly.

The platinum in biogeochemical samples (Figures 3) consistently showed the position of the main Howland reef platinum-palladium zone. The anomalies of this element in lines 3 and 4 suggest that the zone of interest may be wider here than mapped.

The primary criteria in selecting

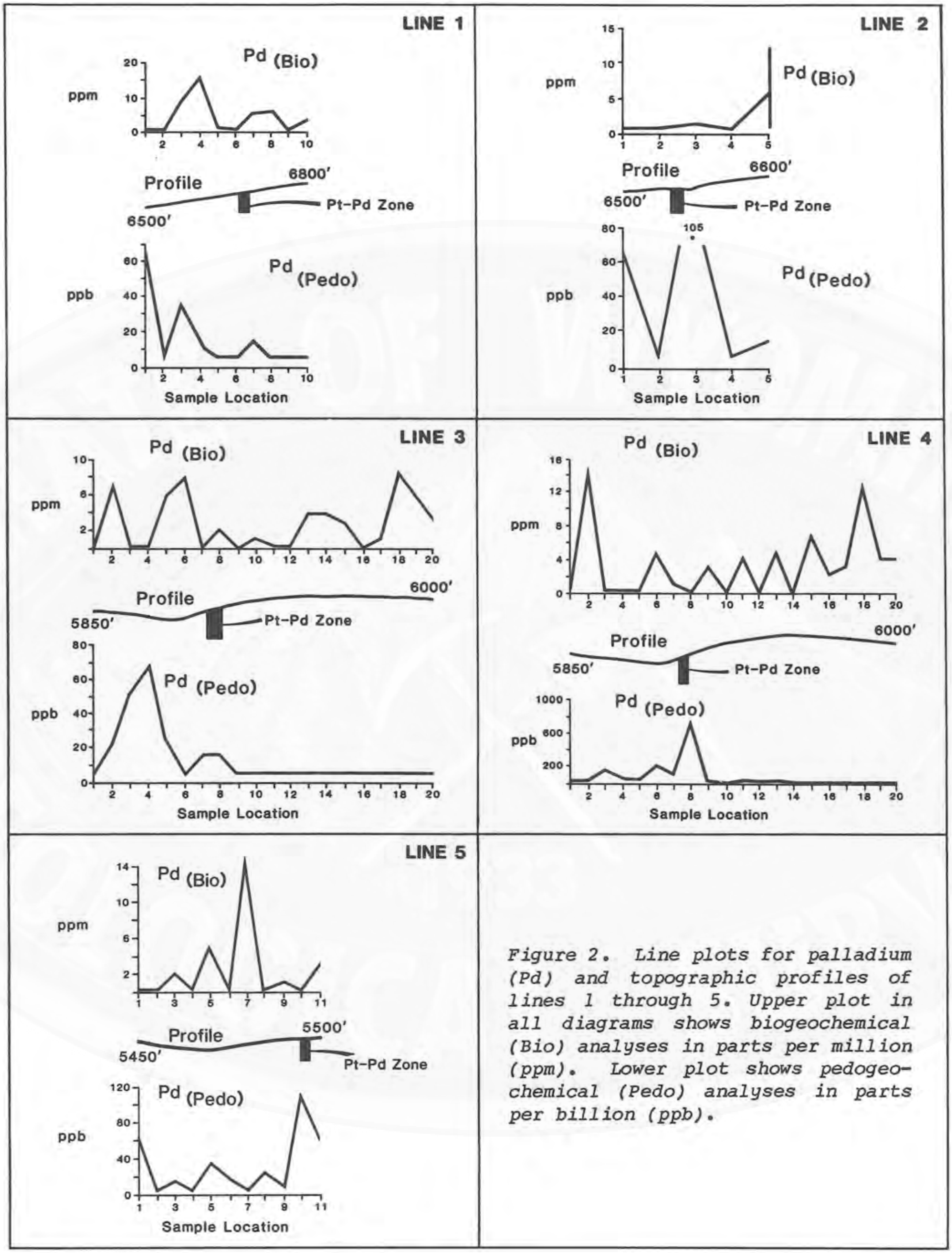

Figure 2. Line plots for palladium (Pd) and topographic profiles of lines 1 through 5. Upper plot in all diagrams shows biogeochemical (Bio) analyses in parts per million (ppm). Lower plot shows pedogeochemical (Pedo) analyses in parts per billion (ppb).

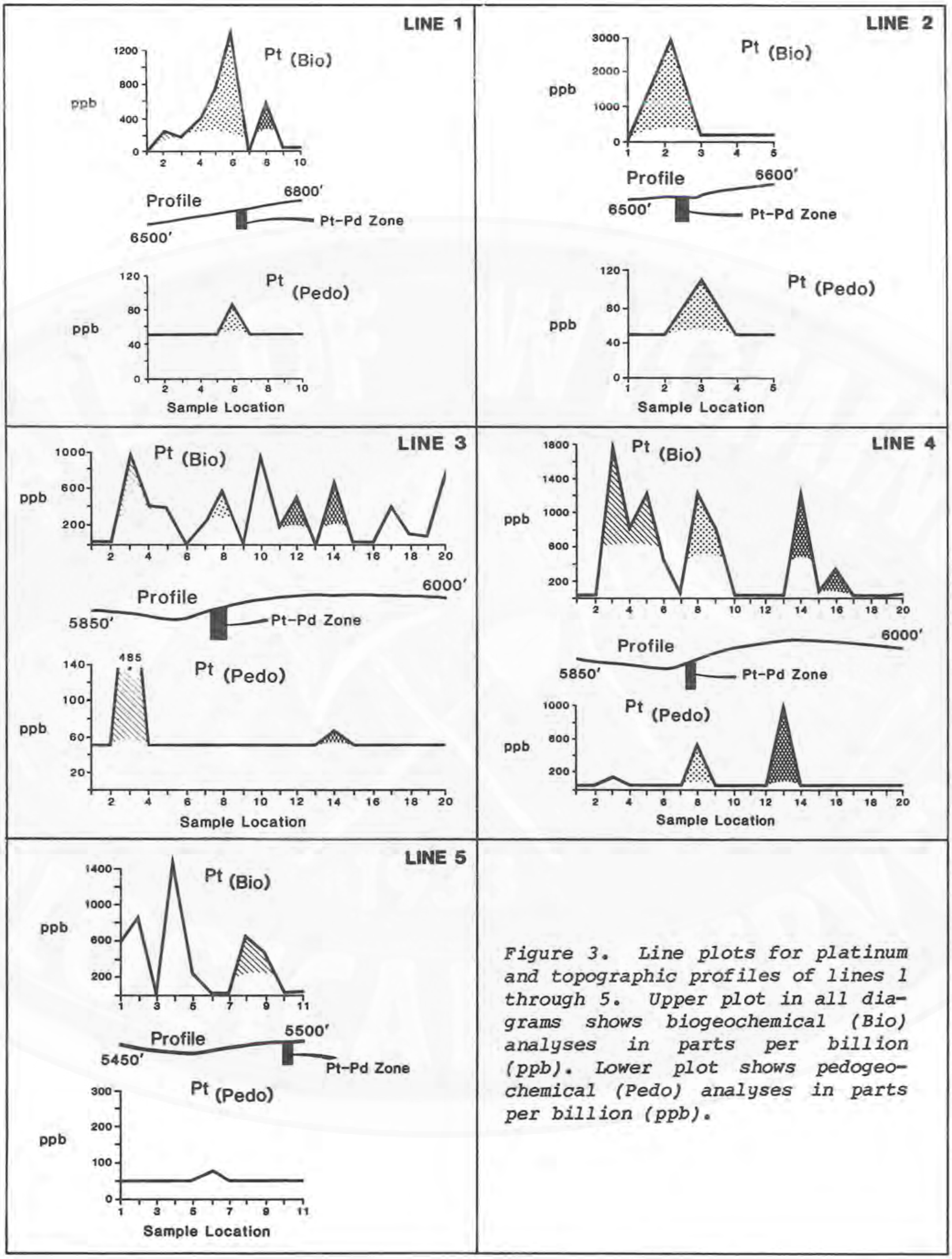

Figure 3. Line plots for platinum and topographic profiles of lines 1 through 5. Upper plot in all diagrams shows biogeochemical (Bio) analyses in parts per billion (ppb). Lower plot shows pedogeochemical (Pedo) analyses in parts per billion (ppb).

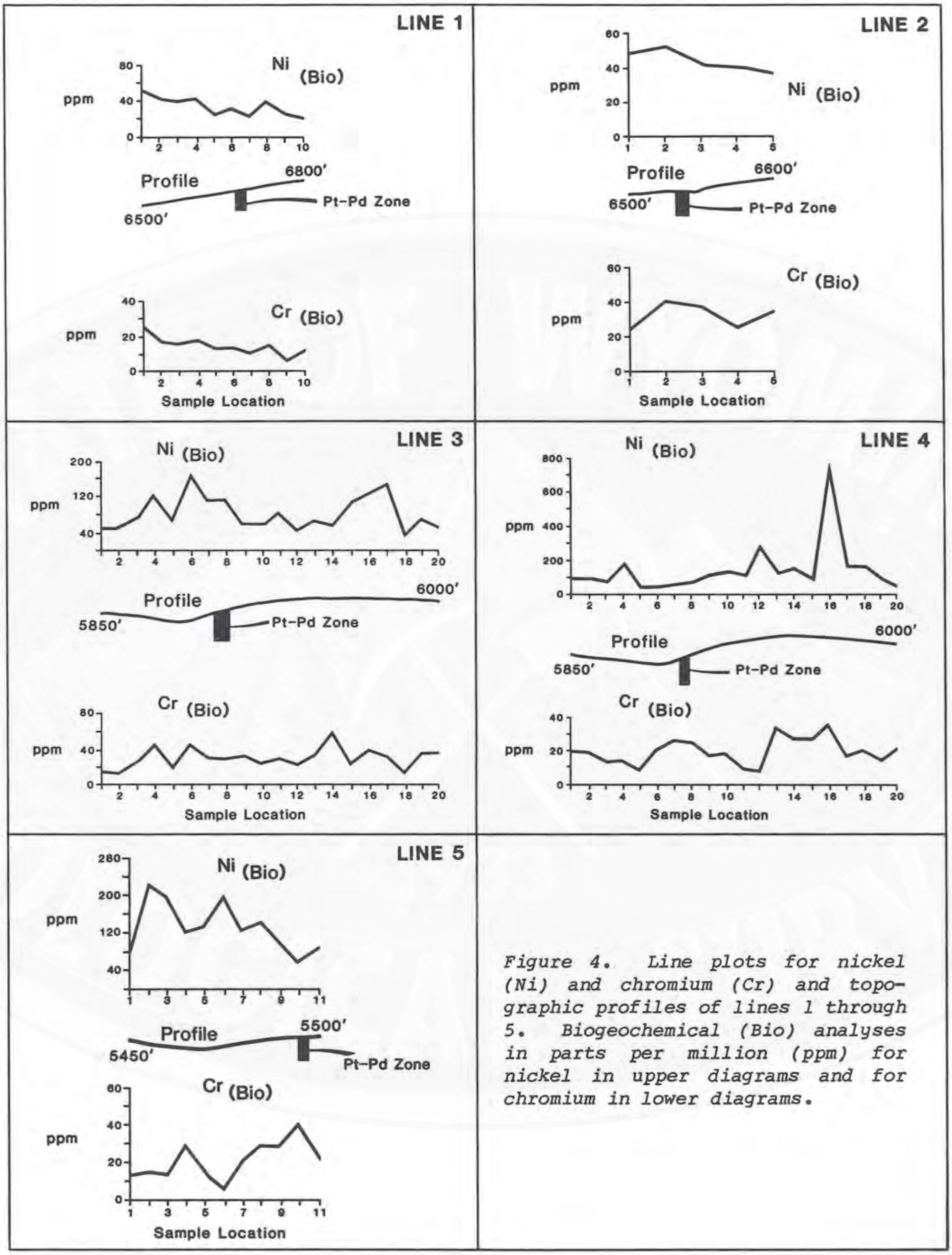

Figure 4. Line plots for nickel (Ni) and chromium (Cr) and topographic profiles of lines 1 through 5. Biogeochemical (Bio) analyses in parts per million (ppm) for nickel in upper diagrams and for chromium in lower diagrams.

an element and sampling medium for a geochemical survey is that the two provide an interpretable product. A low noise-to-signal ratio, high anomaly-to-background contrast, a range of concentration at analytically sound levels, and high absolute contrast between background and anomaly peak are all desirable factors. These factors were best met by the platinum in biogeochemical samples collected during this survey. In general, the noise-to-signal ratio is extremely low, certainly lower than in the pedogeochemical data. The concentration range is approximately one order of magnitude higher in the biogeochemical data than in the pedogeochemical data and is at levels sufficiently high to eliminate any question about analytical relics. Finally, although the anomaly-to-background contrast is about the same for both the biogeochemical and pedogeochemical data, the absolute variance in this measurement is much greater and therefore more desirable in the biogeochemical data.

Conclusions

This orientation survey was designed to test the utility of pedogeochemical and biogeochemical samples to map the Howland reef platinum-palladium zone of the Stillwater Complex. Palladium and platinum appear to be useful for this purpose in both media although the palladium in biogeochemical samples is admittedly questionable.

Our geochemical orientation survey over this zone of mineralizatiom demonstrated that *Psedotsuga menziessi* (Douglas fir) is an effective accumulator of platinum and that this element in this medium is a useful exploration tool. The survey also suggests that neither nickel nor chromium is selectively enriched by this species and that neither is a useful indicator of mineralization.

In selecting a sampling medium and elemental suite for exploration, particular care must be taken to assure that the resulting data will provide an interpretable background-to-anomaly contrast. If possible, these thresholds should be at absolute levels sufficiently high to eliminate any questions that they may be analytical relics. The platinum in Douglas fir fulfills these requirements and provides a reliable sampling medium for subsequent work in this area.

Acknowledgments

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The genesis of uranium deposits in Athabasca, Canada and northern Australia -- Wyoming exploration significance

Ray E. Harris

Introduction

The world's largest uranium reserves are found in the Northern Territory, Australia, and the Athabasca Basin of northern Saskatchewan, Canada. These reserves occur in similar geologic environments, and may have formed in similar ways. Reported reserves for these two regions are 436,360,000 tons of U_3O_8, or one quarter to one third of the noncommunist world's proven uranium reserves (Dahlkamp and Adams, 1981). In comparison, total 1982 United States uranium \$30 reserves (ore minable at a price of \$30.00 per pound of yellowcake, U_3O_8) were 205,000 tons of U_3O_8, and total 1982 Wyoming uranium \$30 reserves were 52,700 tons of U_3O_8 (Hansen, 1982).

Wyoming has some uranium occurrences in geologic environments similar to those of Australia and the Athabasca Basin, and appears to have the potential for a uranium deposit similar in magnitude to those depoists.

The purpose of this report is to summarize the geology and theories of origin of the Canadian and Australian deposits, to discuss and compare similar uranium occurrences in Wyoming and other areas in the United States, and to indicate promising areas for uranium exploration in Wyoming.

Since 1945, uranium occurrences have been known to exist in the Athabasca region of Northern Saskatchewan, Canada. Uranium exploration in that area was, at first, unsuccessfully directed towards the discovery of deposits of the Beaverlodge type, (Tremblay, 1982) or the Shinkolobwe type. Beaverlodge-type deposits are controversial high-temperature or low-temperature deposits found in high-grade metamorphic rocks. Shinkolobwe-type deposits are hydrothermal deposits in metamorphic rock, and contain many unusual uranium minerals. These two deposits were the primary supplies of uranium to atomic weapons and energy research before 1945. Prior to the mid-1960s, uranium exploration companies largely ignored the Athabasca area, for it was then the the fashion to search for Beaverlodge-type and Shinkolobwe-type deposits within metamorphic rocks or for sandstone-type occurrences similar to those in Wyoming and the Colorado Plateau in the United States.

From 1965 to 1968, a few pioneering companies, guided by models of supergene transport of uranium in sandstones and igneous and metamorphic rocks -- as developed by Garrels and Christ (1965), Finch (1967), and especially Moreau and others (1966) --began conducting geochemical and radiometric surveys of the Athabasca Basin. In 1968, the discovery of the Rabbit Lake orebody was announced. In 1969, the discovery of the first orebody at Cluff Lake was announced. Since that time, as a result of the uranium exploration rush generated by these announcements, about fifteen deposits of substantial size have been discovered in Saskatchewan (Figure 1). In November, 1984, the discovery of an orebody at Cigar Lake, 40 miles southwest of Rabbit Lake, was announced. This orebody is remarkable for its 8.49 percent average grade of uranium (Mining Magazine, 1984).

Several surface occurrences of uranium were mined in the Northern Terri-

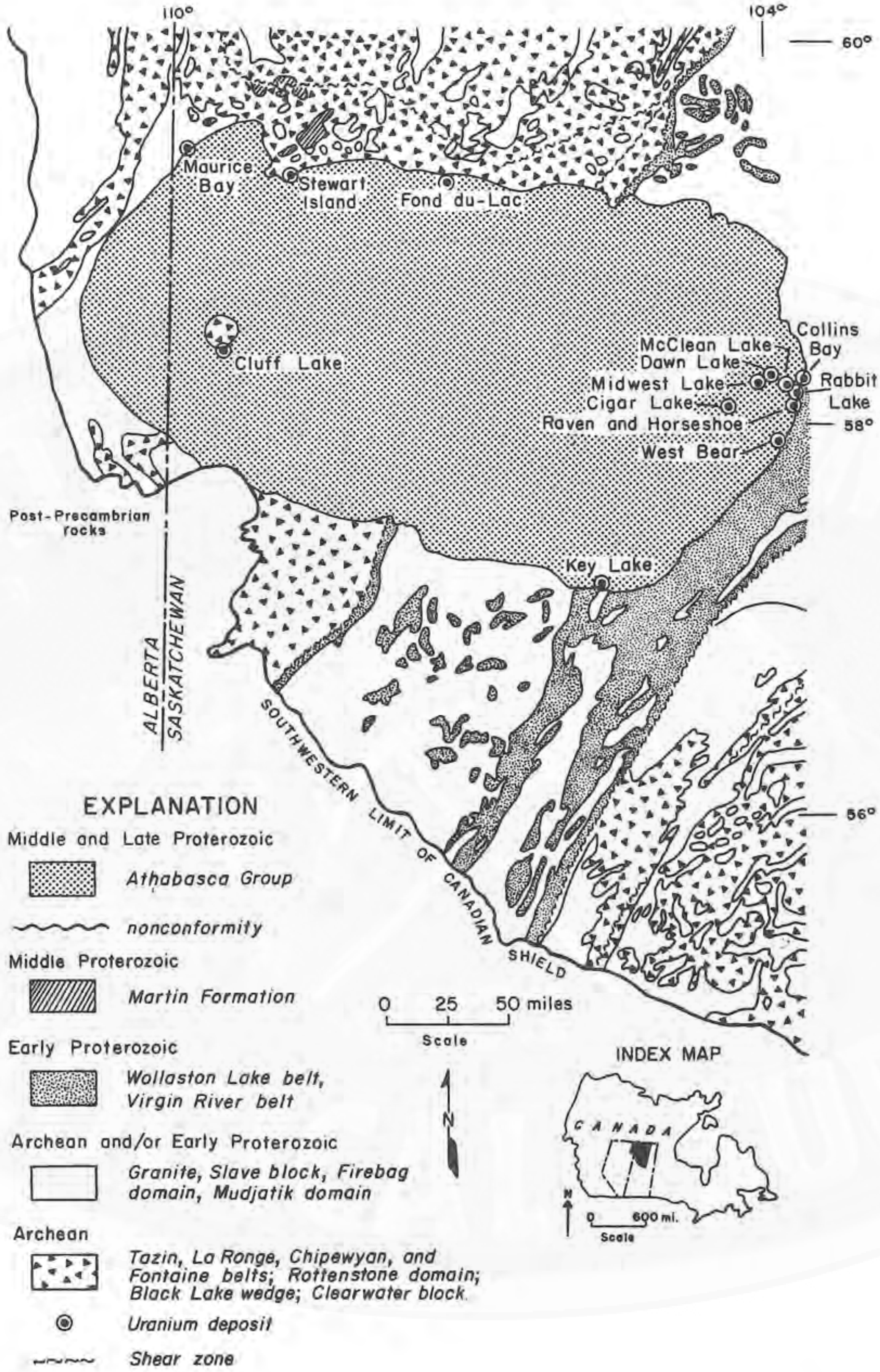

Figure 1. General geology and uranium deposits of the Athabasca region, northern Saskatchewan. (After Tremblay, 1982.)

tory, Australia (Alligator River and Rum Jungle areas) in the early 1950's (Kalliokoski and others, 1978). Following the Athabasca discoveries the area received intensive airborne radiometric and surface geochemical exploration during 1970-1973, and several large discoveries were announced, beginning with Nabarlek in 1970. The Jabiluka II ore body, discovered in 1973, is the second largest single uranium deposit in the world mineable under current economic conditions, with announced reserves of 203,800,000 tons of uranium oxide at a grade of 0.39 percent (Figure 2).

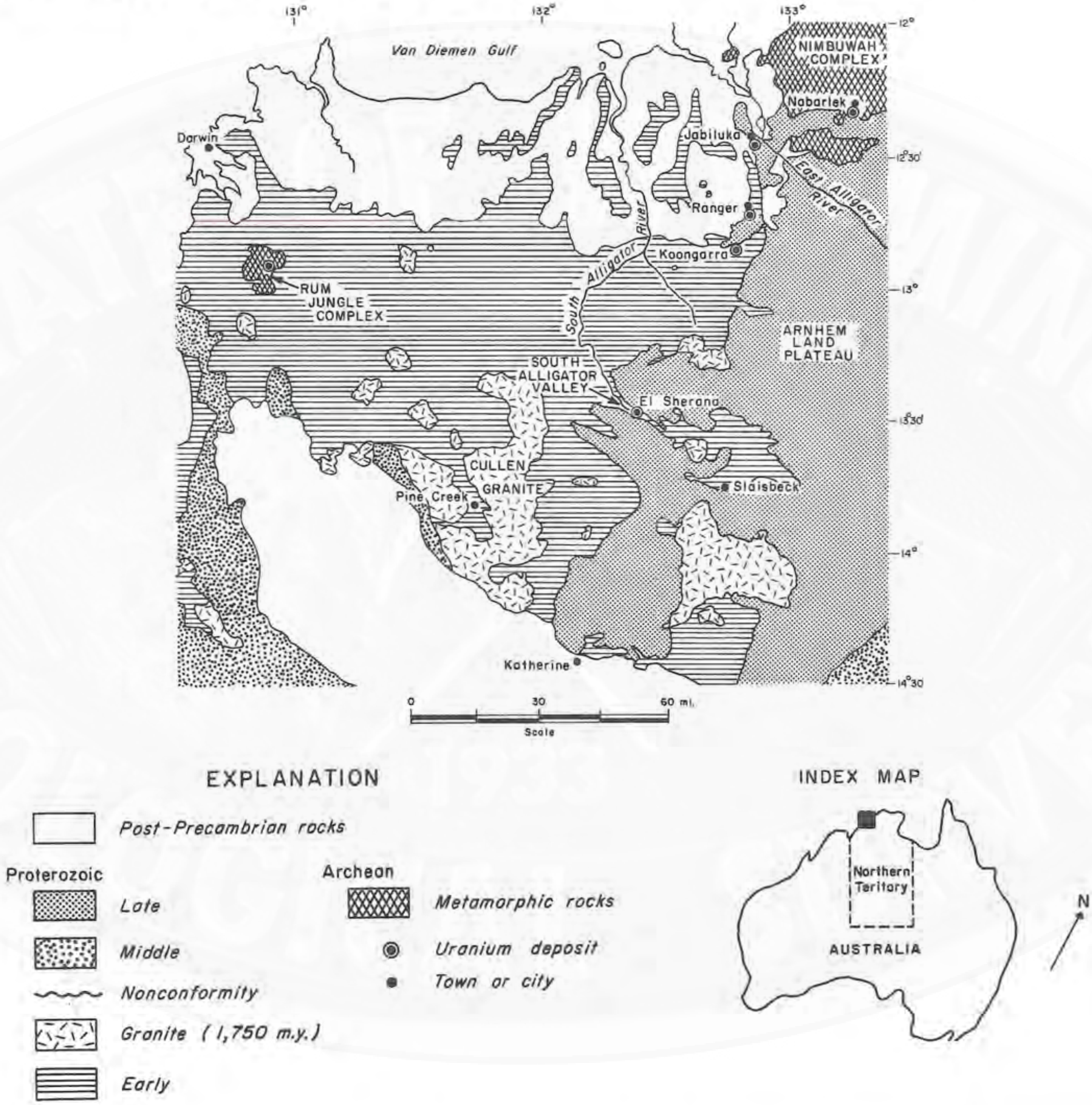

Figure 2. Geology of the Northern Australian uranium producing area and major uranium deposits. (After Kalliokoski and others, 1978.)

There are no producing orebodies in the United States similar to those of the Athabasca Basin and Northern Australia, but two deposits, not currently being mined, may be of similar genesis. These are the deposits near Chatham, Pittsylvania County, Virginia, and at Copper Mountain, Fremont County, Wyoming. These and other, similar deposits in Wyoming are discussed below.

The geology of nonconformity-related uranium deposits

It is beyond the scope of this paper to present detailed geologic summaries of the individual Athabasca and Northern Australia deposits, but a general discussion of those regions is in order.

In Saskatchewan (Figure 1), uranium occurs in large concentrations in metamorphic rocks of the Wollaston Lake belt (Early Proterozoic) and the Tazin and Chipewyan belts (Archean) near a nonconformity overlain by slightly metamorphosed, red, quartz-rich sandstones of the Athabasca Group (Middle and Late Proterozoic). Uranium mineralization within the sub-nonconformity metamorphics is structurally and lithologically controlled (Figure 3). The mineralization is concentrated along faults, fractures, and shear zones in chlorite-rich and graphite-rich feldspathic rocks and calc-silicate units. The deposits appear to extend to only 320 feet below the nonconformity and 130 feet into the Athabasca Group above the nonconformity. Host rocks have been altered (1) by argillization to sericite and chlorite, possibly by low-temperature hydrothermal processes, or (2) by surface weathering prior to the deposition of the Athabasca Group, which produced additional sericite clay, and chlorite as a regolith. Locally, the overlying Athabasca Sandstone has been converted to quartzite by silicification. Other elements present in some

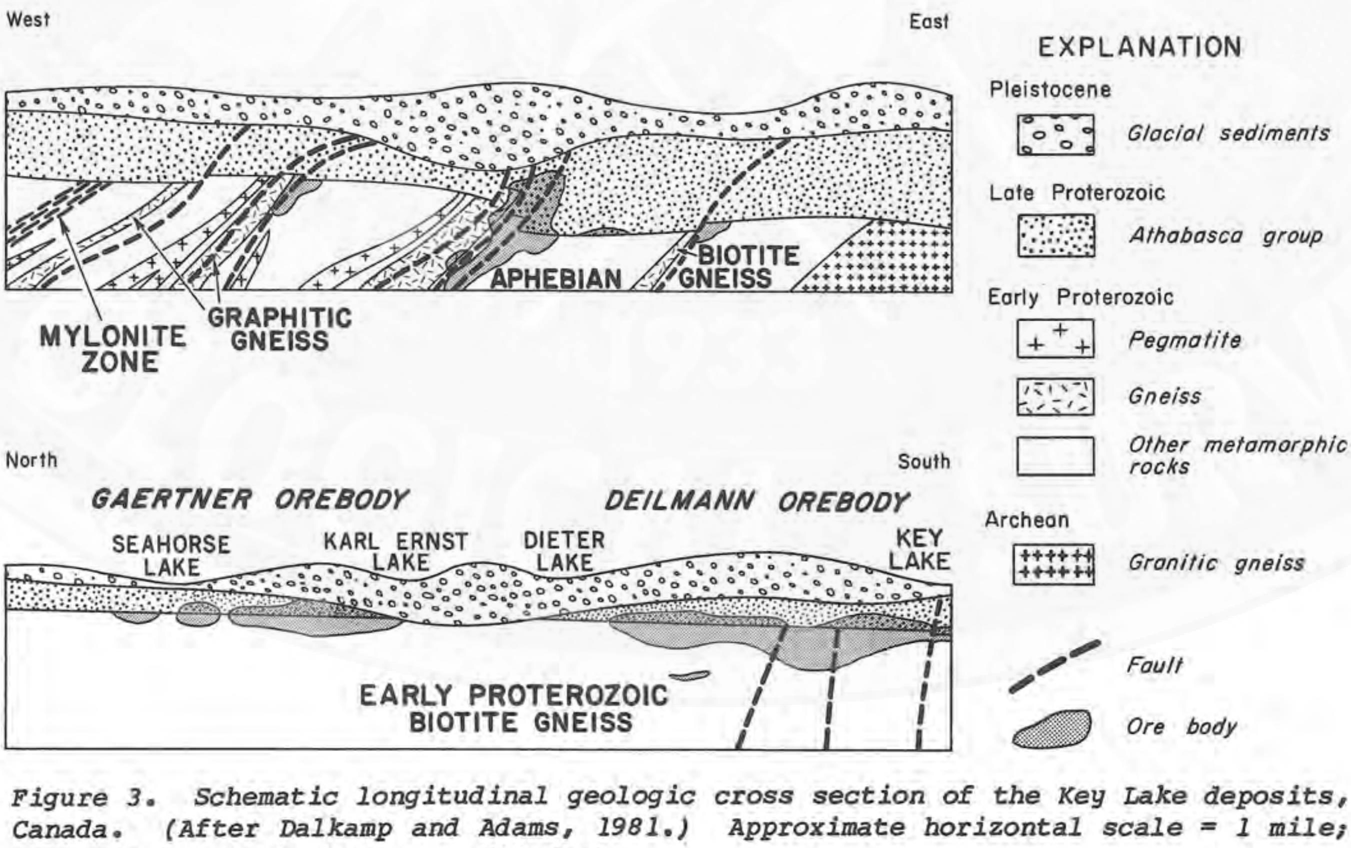

Figure 3. Schematic longitudinal geologic cross section of the Key Lake deposits, Canada. (After Dalkamp and Adams, 1981.) Approximate horizontal scale = 1 mile; approximate vertical scale = 200 feet.

deposits are cobalt, nickel, arsenic, selenium, tellurium, gold, vanadium, molybdenium, copper, lead, and silver. Some deposits only contain high concentrations of uranium, while the Gaertner deposit contains significant nickel concentrations, and Cluff Lake D deposit contains significant gold. Uranium minerals are generally low-temperature pitchblende and coffinite, but where selenium and arsenic are abundant, minerals may include complex selenides and arsenides (Tremblay, 1982). The age of the deposits is 1,281$\pm$ 11 m.y. (Cumming and Rimsaite, 1979). The temperature of mineralization is low, below 300°C (Dahlkamp, 1978).

In Northern Australia (Figure 2), uranium occurrences in lower Proterozoic and Archean metamorphic rocks at a nonconformity overlain by middle and upper Proterozoic, quartz-rich, red sandstones provide much the same setting as the Athabasca Region. Sub-nonconformity host rocks are feldspathic, pyritic, and graphitic gneisses and schists. Alteration consists of argillization, sericitization, and chloritization in the vicinity of the ore bodies (Kalliokoski and others, 1978). Ore bodies occur roughly parallel to foliation (Figure 4) and trend along faults, shear zones, fold hinges, and joints (Dodson and others, 1974). The overlying unit is the Kombolgie Formation of the Katherine River Group (Figure 5), a red, quartz-rich sandstone containing a few interbedded volcanics (Needham, 1982), deposited in a braided stream system in an alluvial plain environment (Ojakangas, 1979).

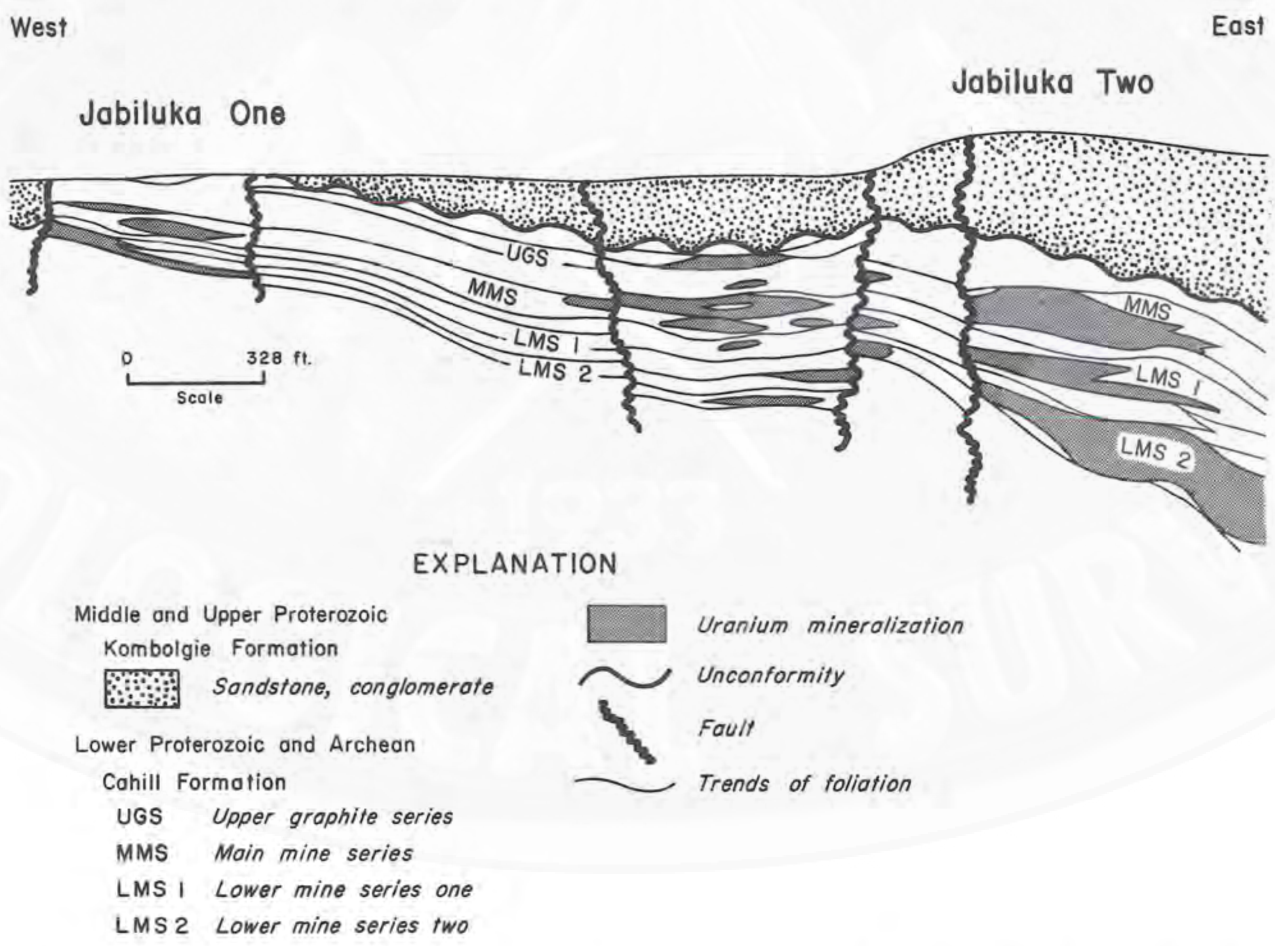

Figure 4. Cross section of the Jabiluka uranium deposit, Northern Territory, Australia. (After Dahlkamp and Adams, 1981.)

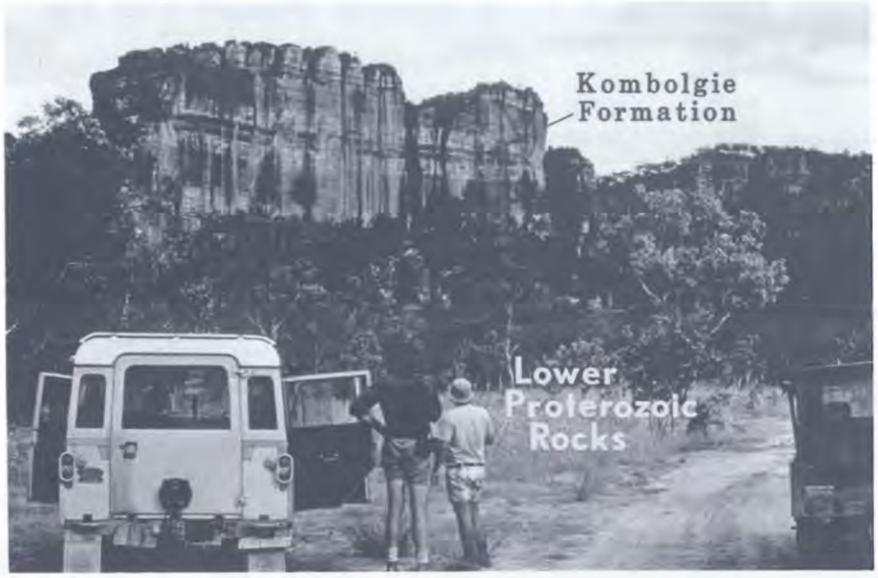

Figure 5a. Kombolgie Formation escarpment near Jabiluka, Northern Territory, Australia. (Photo by Dave Mickle.)

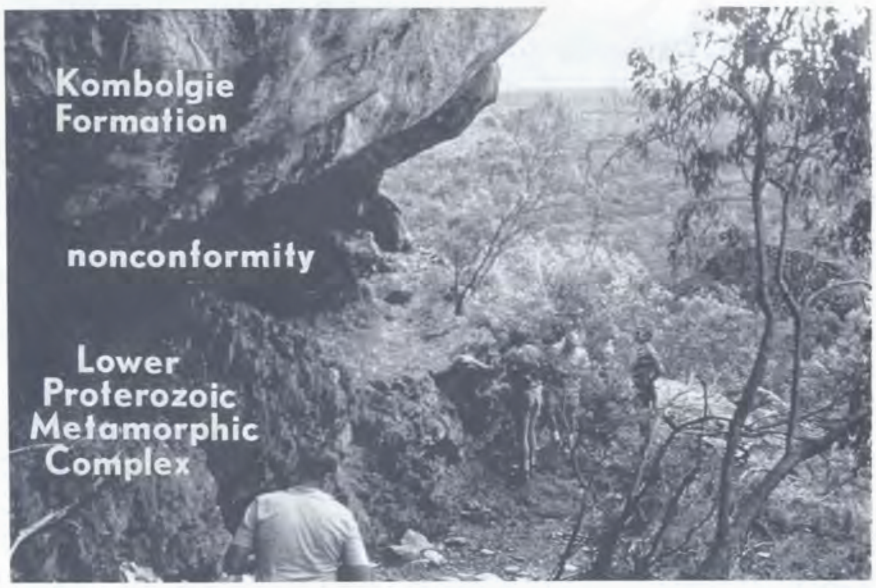

Figure 5b. Unconformity at the base of the Kombolgie Formation, Northern Territory, Australia. (Photo by Dave Mickle.)

Mineralization consists primarily of low-temperature pitchblende and minor coffinite and brannerite. Gold is associated with uranium at Jabiluka, and traces of nickel, copper, zinc, lead and tellurium are present (Kalliokoski and others, 1978). The age of the Australian deposits is not known with certainty, but appears to be near 1,700 m.y. with remobilization at 900 and 600 m.y. (Kalliokoski and others, 1978).

Common characteristics of uranium ore bodies of the type found in the Athabasca region and in Northern Australia are:

(1) They are spatially associated with a nonconformity between older metasedimentary or igenous rocks and an overlying, relatively unmetamorphosed, sandstone unit.

(2) The overlying sandstone has been oxidized, and is red.

(3) The overlying unit is medium- to coarse-grained, well-sorted, fluvial sandstone.

(4) The sub-nonconformity units contain reductants, usually graphite, but sometimes pyrite and other sulfides.

(5) The uranium ore bodies are controlled by structures in the sub-nonconformity rocks.

(6) The uranium minerals formed at low temperature (less than 350°C).

(7) Alteration of the host rock is characteristically argillization and associated chloritization and sericitization.

(8) Primary thorium is never present.

(9) The deposits formed 1,200 to 1,800 m.y. ago in the Middle to Late Proterozic (the Canadian deposits seem to be somewhat younger than the Australian).

Characteristics of some, but not all, of the deposits include the following:

(1) A regolith is developed at the nonconformity prior to deposition of the overlying sandstone.

(2) Cobalt, nickel, gold, copper, lead, zinc, silver, molybdenum, selenium, or tellurium are present.

(3) The overlying sandstone may contain some mineralization, but never in amounts greater than the sub-nonconformity metamorphics.

Genesis of the Canadian and Australian deposits

Three primary theories have been proposed for the origin of the Athabascan and northern Australian deposits:

(1) A hydrothermal theory: Uraniferous hydrothermal fluids originated at depth, migrated upward through the sub-nonconformity rocks, and deposited uranium at the unconformity.

(2) A diagenetic theory: The source of the uranium was weathering of the sub-nonconformity rocks. The uranium was then transported by surface or near-surface water to its present location. Some proponents of this theory believe that a regolith formed over the deposits, protecting them from later oxidation and weathering, particularly during deposition of the overlying units.

(3) A modified diagenetic theory: Uranium was carried through the overlying sandstones in solution and deposited in the host rocks by reduction. This theory does not give a source for the uranium.

The writer favors the modified diagenetic theory for the origin of the deposits. There is little evidence for characteristic hydrothermal uranium associations, particularly thorium and rare-earth elements. The absence of large quantities of hydrothermal sulfides in these deposits also seems to negate this theory. Diagenetic formation before post-unconformity sandstone deposition is chemically possible. However, the location of all these deposits beneath red sandstones suggests that the overlying unit was involved in the formation of the uranium deposits. No deposit of this type has been reported beneath nonconformities overlain by limestones or other marine units, or beneath clay-rich or fine-grained sedimentary units. The essential difference is that the sand would have been originally quite permeable, thereby providing a conduit for uranium-bearing fluids. The writer proposes the following theory of origin for these deposits (Figure 6):

(1) Formation of pre-nonconformity units through sedimentation, metamorphism and intrusion. Favorable units contain graphite or pyrite.

(2) Structural preparation of pre-nonconformity rocks, which may have been contemporaneous with the last stages of step 1, forming fault zones, shear zones, breccia zones, and joint systems into which uranium-

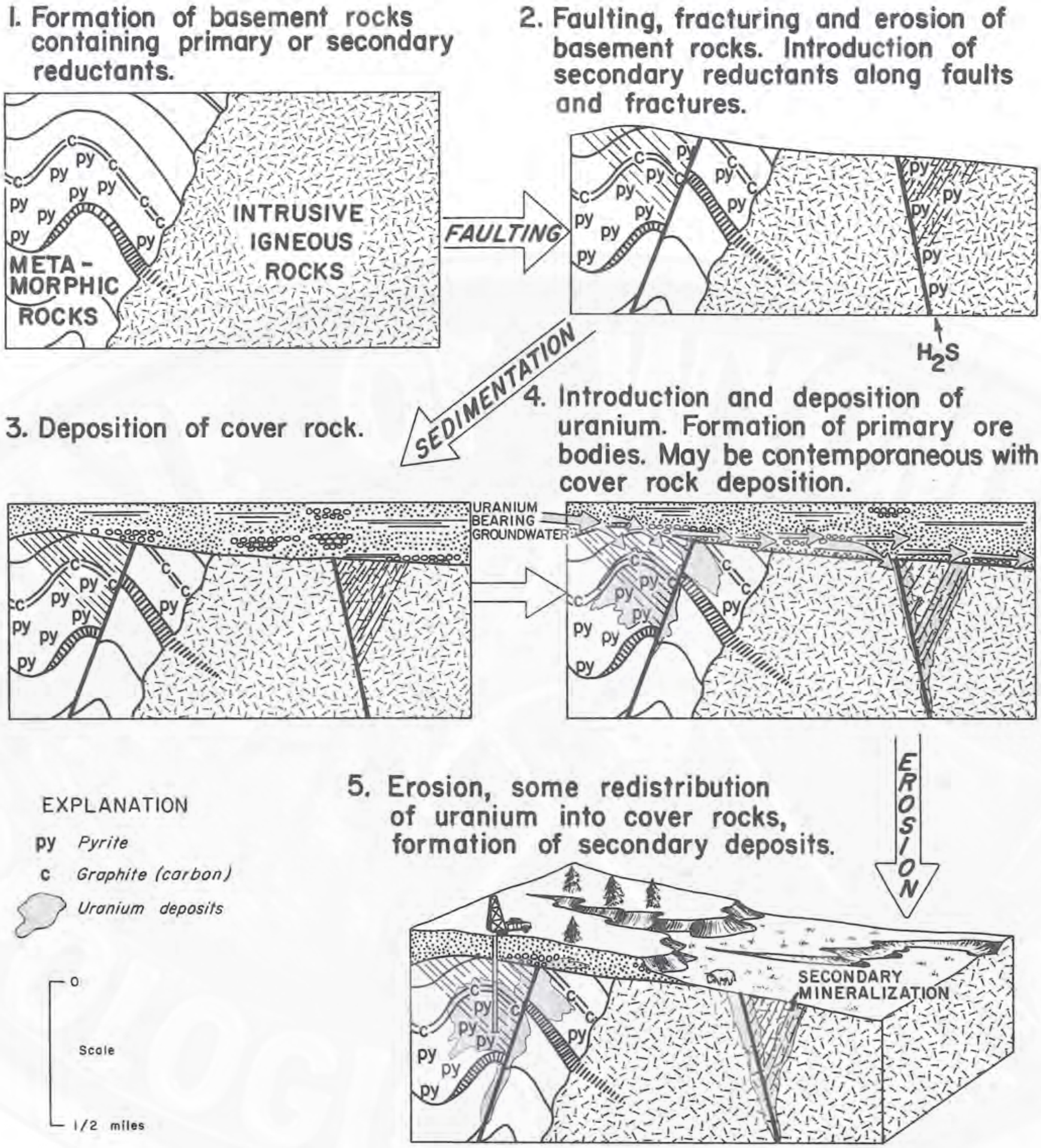

Figure 6. Formation of nonconformity-related uranium deposits. Blocks depict about 2,000 foot horizontal scale.

bearing solutions could penetrate. The deposition of pyrite may have accompanied this step, providing reductants. Folding episodes may also have produced permeable zones, particularly in the hinge areas of the folds. This was followed by erosion of metamorphic and igneous rocks to a surface of nonconformity. In Canada, it appears to have ended with a period of lateritic weathering and formation of the regolith.

(3) Deposition of the overlying sandstone in either an oxidizing or a reducing environment.

(4) Diagenesis of the overlying sandstone unit with transportation of uranium in oxidized form. If the overlying sandstone was deposited in a reducing environment, oxidizing water, containing uranium, could have oxidized the sandstone in a manner similar to that which produces redox type (roll-front and peneconcordant) deposits in sandstone (Harris, 1982; 1984). The solutions entered the units beneath and along the nonconformity, especially where the rocks were structurally prepared. Upon encountering reductants, uranium was reduced and precipitated, along with other trace elements. Low-temperature alteration produced clay from feldspar and removed biotite and hornblende, from which iron was freed for the production of chlorite. This alteration is charcteristic of low-temperature roll-front deposits (Harris, 1982; 1984). Continued diagenesis and deeper burial altered the hydrology of the system and the formation of ore-deposits ceased. Silicification cemented the overlying sandstone into quartzite. Some uranium may have been remobilized into the overlying sandstone, as the sands were locally reduced. In some areas thermal events altered the ore bodies and reset the uranium-lead clocks, resulting in relatively young age determinations. If this diagenesis was too severe, uranium deposits formed earlier could have been destroyed, with uranium mobilized into high-temperature solutions and transported elsewhere. No highly metamorphosed nonconformity has been found to contain uranium of this mode of occurrence.

(5) Uplift and erosion of the present cycle brought the uranium deposits to near the surface, permitting discovery. Probably, uranium deposits occur beneath the Kombolgie (Ojakangas, 1979) and Athabásca sandstones away from their erosional edges.

This theory does not designate a source of uranium for the water moving through the overlying sandstone. As in the case of redox deposits in sandstone, it could be from the weathering of granite, volcanic ash, or the sandstone unit itself. Ojakangas (1979) proposes a theory in which the sandstone overlying the nonconformity, can be a source of the uranium. This case seems likely, particularly since acidic volcanogenic units are scarce or absent in the Athabasca region. Volcanic ash can be a potential source in other areas, however. The relatively unique occurrence of these deposits near nonconformities of Middle Proterozoic age suggests that ground water was transporting large quantities of uranium at this time. The earth's atmosphere is considered to have been reducing prior to 2,300 m.y. ago, and uranium minerals were stable at the surface. After 2,300 m.y. ago, the atmosphere became oxidizing (Roscoe, 1973); these minerals were no longer stable and were easily weathered, liberating uranium into surface water and ground water. It is likely that most surface water and ground water at this time were greatly enriched in uranium, and that more uranium was available to form deposits in favorable areas.

This theory does not, however, restrict formation of uranium deposits to the Middle and Late Proterozoic, as suggested by Dahlkamp and Adams (1981). If the geologic conditions are favorable, and if there is sufficient uranium in the system, nonconformity-related uranium deposits may be formed in any age. In Wyoming, several uranium occurrences near unconformities of other ages may have formed similarly and are classed as nonconformity-type deposits. In fact, the bounding surface between the favorable host rocks and oxidized, permeable sandstone need not be a nonconformity, but can also be a fault plane surface.

Derry (1973) refers to "contemporaneous surfaces" in his observations on the origin of the Beaverlodge occurrences and other low-temperature deposits. The presence of a surface at which reactions can occur is the critical relationship.

The proposed theory of formation for these deposits, then, can be related to other surfaces between oxidized, permeable rocks capable of supplying uranium-bearing solutions and adjoining rocks containing reductants. Unconformities of all types are possible targets, as well as fault surfaces. Roll fronts in sediments, as boundaries between oxidizing and reducing zones, can possibly be classified contemporaneous surfaces as the term is used by Derry (1973).

Examples of nonconformity-related uranium occurrences in the United States

Exploration for nonconformity-related uranium deposits in the United States has generally been restricted to nonconformities of the same age as the pre-Athabasca nonconformity in Canada and the pre-Kombolgie nonconformity in Australia. Although nonconformity-related uranium occurrences have been found in a few areas to date (Figure 7), only one uranium discovery has been announced.

The largest and one of the most intensively explored area of this type in the United States is the Fond du Lac region of Minnesota (Figure 7). There, the Keweenawan Fond du Lac Formation, a red, arkosic, slightly metamorphosed sandstone, rests upon older units. The nonconformity is not exposed, except for a small area in the suburbs of Duluth that is exposed from time to time near an intermittently active landslide. Due to the lack of exposure and the thick glacial cover, prospecting in the area is expensive and relies heavily on geophysics and other indirect methods. To date, some drilling and geophysical

Figure 7. Areas of nonconformity-related uranium occurrences and exploration in the United States.

studies have been completed, but no uranium discovery has been announced. A related area is the small outcrop of the Keweenawan Puckwunge Formation in Minnesota, northeast of Duluth. Uranium occurrences have been located near Keweenawan units in the upper peninsula of Michigan (Figure 7). Other areas of Precambrian nonconformities have not been extensively explored.

A recent uranium discovery with 30 million pounds of uranium oxide at a grade more than four pounds per ton of ore near Chatham, Pittsylvania County, Virginia, is of interest (Figure 7) (The Mining Record, 1982). This deposit occurs along a fault in which a red, arkosic sandstone of Triassic age is in contact with high-grade metamorphic units of the Piedmont Belt. The ore occurs mainly in the Piedmont metamorphic units. The nature of the mineralization has not been made public, but the setting is similar to nonconformity-uranium deposits, with two exceptions: the contemporaneous surface is a fault, and the age of the deposit may be younger than Triassic. Nevertheless, on first glance, it seems to have formed similarly to the Athabasca or Northern Australia deposits.

Similar geologic settings are found in two Triassic basins south of the Chatham area (the Wadesboro and the Raleigh-Durham Basins in North and South Carolina) and elsewhere in the eastern United States. There are uranium geochemical anomalies in the Raleigh-Durhan and Wadesboro Basins. Uranium minerals are present in Triassic arkose near the fault boundary of the Connecticut Basin.

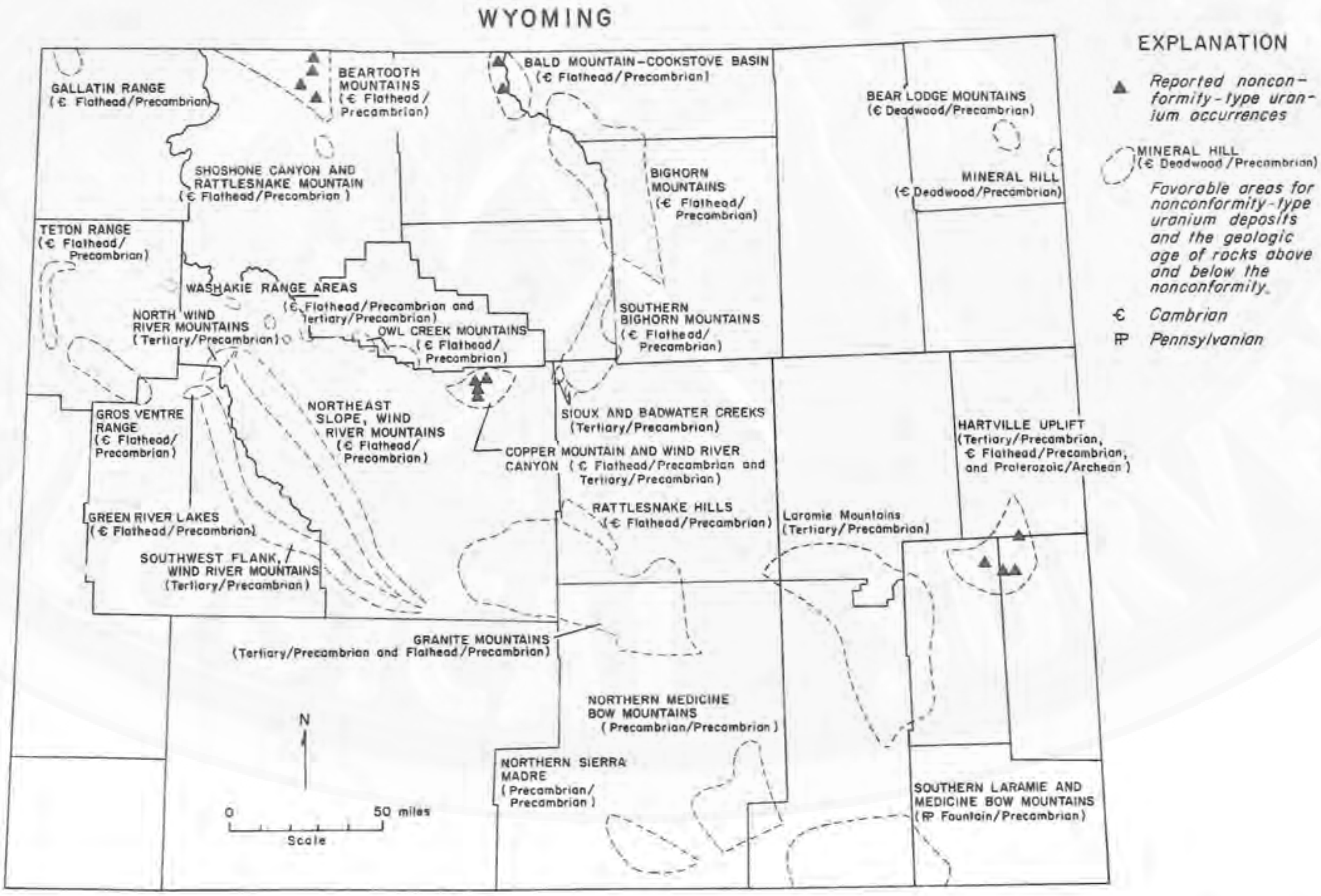

Figure 8. Nonconformity-related uranium occurrences and favorable areas for exploration in Wyoming.

Nonconformity-related uranium occurrences in Wyoming

In Wyoming, radioactivity anomalies are found along the nonconformable contacts between the Cambrian Flathead Sandstone and underlying Archean igneous and metamorphic rocks around the margin of the Bighorn Basin and in surrounding uplifts (Figure 8). (See also Harris, 1983.)

In Montana, along the Beartooth Mountain front, about six miles north of the Wyoming State Line, sooty pitchblende is present in fractures in granitic and gneissic Precambrian rocks and the Cambrian Flathead Sandstone. In Wyoming, radioactivity occurrences are common along the nonconformity in the Shoshone River, Clarks Fork, and Littlerock Creek canyons, and on the divides between these drainages. Along Sunlight Creek, significant radioactivity extends into the Precambrian along joints in the gneissic units.

The Flathead Formation contains concentrations of radioactive heavy minerals, particularly at Bald Mountain and Cookstove Basin on the east side of the Bighorn Basin (Figure 8). Those occurrences along the nonconformity that are primarily located in the Precambrian units are nonconformity related. At the Clarks Fork locality, anomalies occur sporadically along the nonconformity from the river level to the crest of the ridge north of the river (Figure 9).

In other areas in the state, notably the Hartville uplift, radioactivity anomalies, as well as a general increase in background radiation, are associated with this nonconformity (Figure 8). On South Twin Hill, sec. 17, T.28N., R.64W., above-normal radioactivity (from uranium) occurs with visible copper mineralization in gneissic units beneath the Flathead Formation (Figure 10). At this locality, the Flathead is near its wedge-edge and is only about five feet thick. The radioactivity and copper mineralization appear to be selectively concentrated in mafic units and along fractures in the Precambrian rocks beneath the nonconformity. The Flathead is also anomalously radioactive in this area. The overlying Guernsey Limestone (Mississippian-Devonian) contains no anomalous radioactivity.

Fluvial Tertiary rocks, which host all of the currently producing and most of the past-producing uranium deposits in Wyoming, rarely directly overlie Precambrian igneous and metamorphic rocks. At the Copper Mountain area (Figure 8), uranium occurs in fractured and faulted

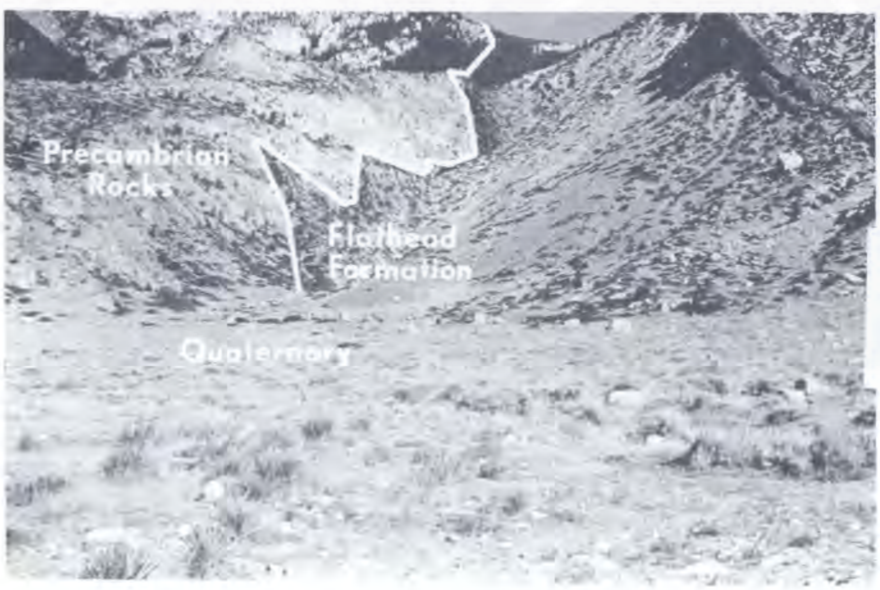

Figure 9. Basal Flathead Formation (Cambrian) nonconformity,Clarks Fork Canyon, Park County, Wyoming.

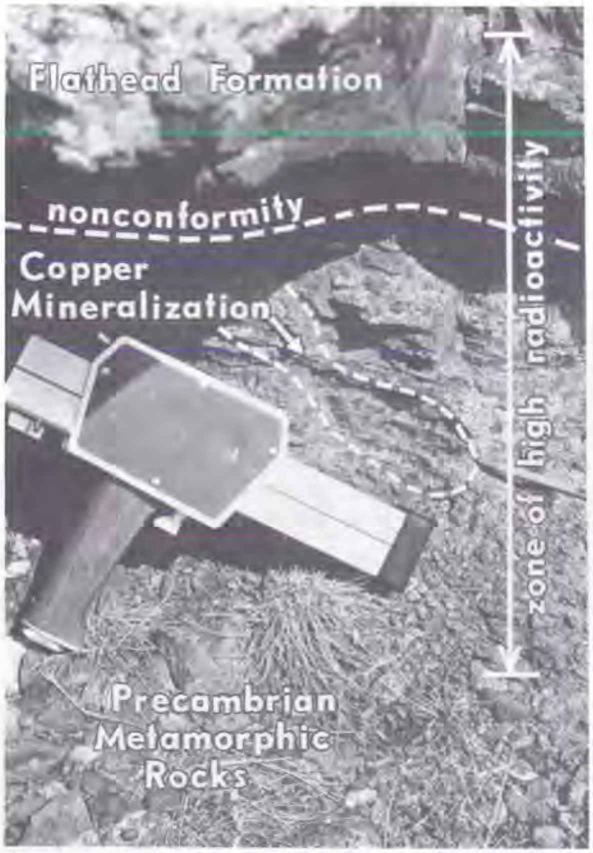

Figure 10. Basal Flathead Formation (Cambrian) nonconformity with visible copper mineralization, South Twin Hills, Goshen County, Wyoming.

Precambrian rocks and in the nonconformably overlying Eocene Tepee Trail Formation. The uranium occurrence is subeconomic but of promising grade and size. Rocky Mountain Energy Company has conducted detailed drilling on the North Canning deposit in the Copper Mountain area. According to Yellich and others (1978), uranium is spatially related to fractures and subsidiary faults associated with the Laramide North Canning fault. Mineralization occurs in Precambrian granitic rocks and enclosed metasediments. Figure 11 illustrates the spatial relationship of ore to the host rocks there. Alteration includes propylitization and argillitization. Mineralization is primarily low-temperature pitchblende and coffinite, with secondary minerals developed in zones of oxidation.

In mineralization and alteration, the North Canning deposit is similar to nonconformity-related uranium deposits. It is likely that the deposit formed by processes similar to those that operated in the Athabasca and Northern Australia regions. Yellich and others (1978) suggest a modified nonconformity-related theory involving redox changes during the structural evolution of the area. This theory, although more complex than the modified diagenetic theory, involves similar processes, with uranium-bearing solutions migrating through overlying sediments, depositing the uranium in structurally controlled zones in multiple cycles of redox changes. Multiple cycles may also have been factors in the Athabascan and Northern Australian areas, which exhibit much less structural complexity than Copper Mountain.

Another nonconformity in Wyoming that meets the general criteria of nonconformity-related uranium producing areas is between Precambrian granites and metamorphic rocks and the overlying Fountain Formation (Pennsylvanian) in the southern Laramie Mountains and southern Medicine Bow Mountains in southeastern Wyoming (Figure 8). Uranium radiometric anomalies occur locally in the Fountain Formation, a red, coarse-grained arkose (Harris and Hausel, 1984). Uranium-bearing veins occur in Precambrian rocks immediately below the nonconformity southwest of Fort Collins, Colorado, on Horsetooth Mountain, 40 miles south of the Wyoming State Line. Favorable metamorphic rocks, containing reductants, are present beneath the Fountain Formation in the area of Jelm and Ring Mountains and the southern Medicine Bow Mountains in Albany County, Wyoming.

Several unconformities within a Proterozoic and Archean metasedimentary sequence occur in the Sierra Madre and northern Medicine Bow Mountains in south-central Wyoming (Figure 8). In the Sierra Madre, rocks of the Deep Lake Group overlie Archean units (Figure 12). The Magnolia Formation, the lowest unit

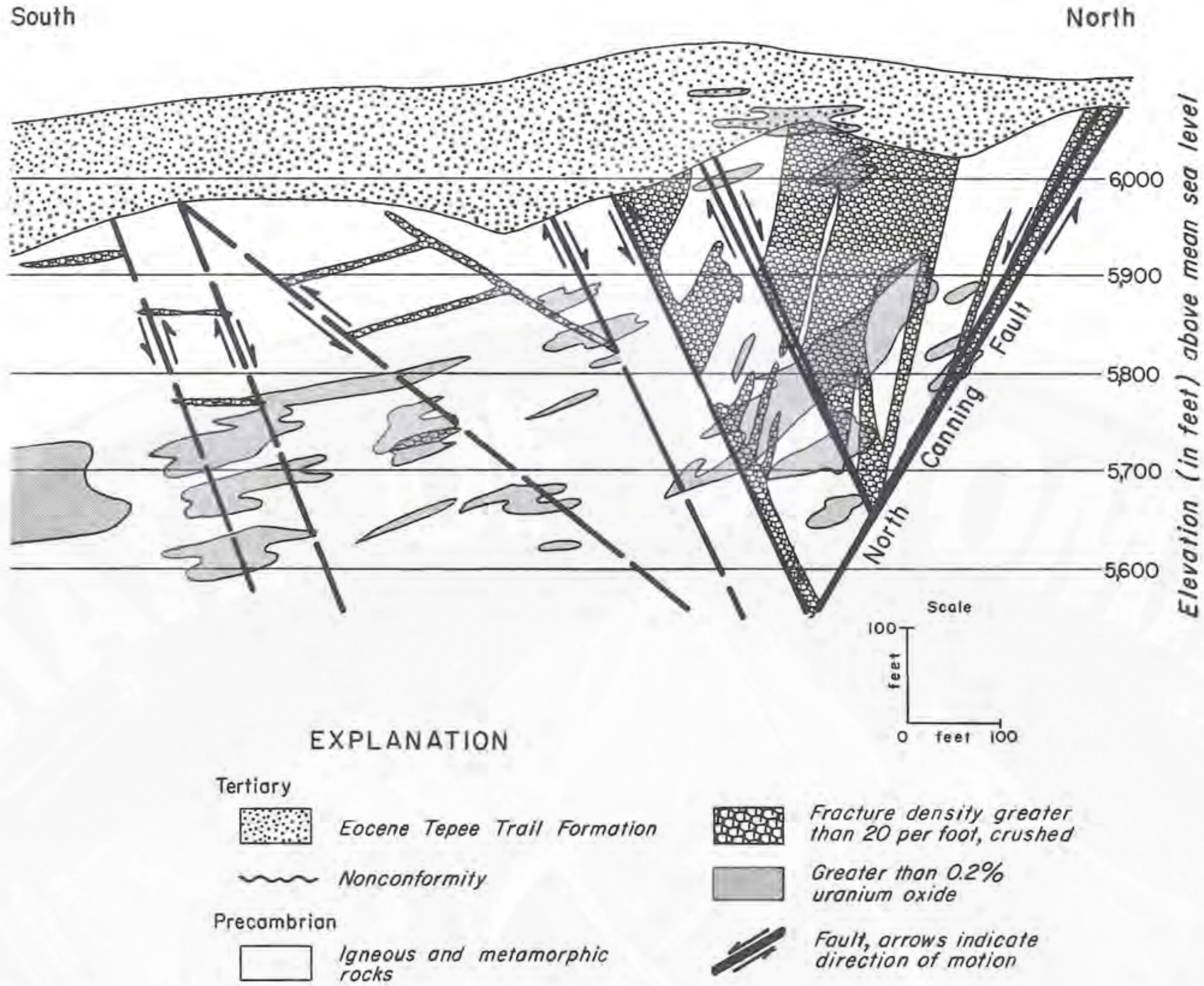

Figure 11. Schematic cross section, North Canning area, Copper Mountain, Fremont County, Wyoming. (After Yellich and others, 1978.) Compare with Figures 2 and 4.

of the Deep Lake Group, contains some uraniferous quartz-pebble conglomerates (Karlstrom and others, 1981), indicating that this nonconformity may have formed before the post-2.3 billion year atmospheric change, and oxidizing solutions carrying uranium would not have been present. Younger unconformities, at the base of the Vagner Formation and Cascade Quartzite (Figure 12), may be more favorable. Similarly, in the Medicine Bow Mountains, rocks of the Deep Lake Group (Magnolia Formation) unconformably overlie older units of the Phantom Lake Metamorphic Suite (Karlstrom and others, 1981). Younger unconformities at the base of the Cascade Quartzite and Vagner Formation (Figure 12) may be more favorable. Uranium occurrences are present in the quartz pebble conglomerates (Karlstrom and others, 1981), but there are no reported uranium anomalies associated with unconformities. Areas in the northern Medicine Bow Mountains, in which rocks of the metasedimentary sequence overlie Archean gneisses (Houston and others, 1978), may also be favorable. Since these unconformities are not nonconformities, in that they involve relatively unmetamorphosed stratified sedimentary rocks both above and below the erosional surface, they do not fit into the nonconformity category. They may be related to a contemporaneous surface, as used by Derry (1973), wherein other unconformities, depositional

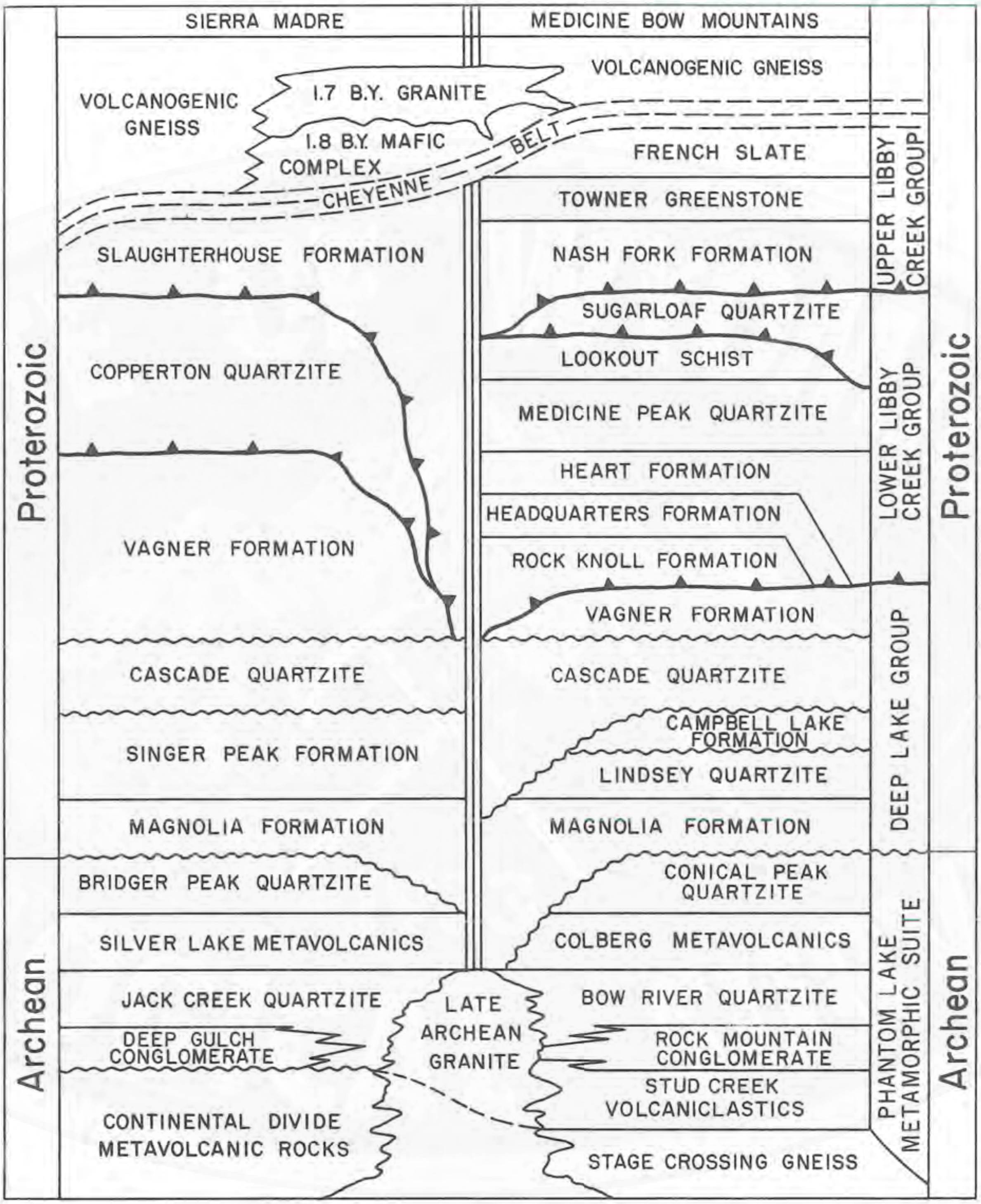

Figure 12. Stratigraphy of Precambrian rocks of the Sierra Madre and Medicine Bow Mountains. (Modified from Karlstrom, and others, 1981.)

boundaries, or faults, may be boundaries along which similar deposits may form.

In the Hartville uplift in east-central Wyoming (Figure 8), a nonconformity exists between overlying Proterozoic metasedimentary rocks (including a basal quartzite named the Whalen Group by Snyder, 1980) and Archean gneiss and granite. Near Hell Gap (sec. 15, T.28N., R.65W.), radiometric anomalies at the base of the Whalen Group may indicate nonconformity-type uranium occurrences. Elsewhere along the outcrop belt of the Precambrian of the Hartville uplift, uranium radioactivity anomalies are common. The Hartville uplift may be a target worth exploring with efforts concentrated along both the basal Cambrian and basal Proterozoic nonconformities.

Summary of areas in Wyoming favorable for nonconformity-related uranium deposits

Several areas in Wyoming are favorable for nonconformity-related uranium deposits (Figure 8). Criteria are: a nonconformity overlain by red, oxidized, permeable (or formerly permeable) sediments of fluvial or near-shore origin overlying metamorphic or igneous rocks that may contain reductants and be structurally favorable. Three nonconformities that meet these criteria are the basal Cambrian (Flathead Sandstone or Deadwood Formation) nonconformity, the basal Tertiary (Tepee Trail Formation and others) nonconformity, and the basal Pennsylvanian (Fountain Formation) nonconformity. In addition, unconformities are present in Precambrian rocks in the Sierra Madre, Medicine Bow Mountains, and Hartville uplift where most of the favorability criterion are met. These last unconformities are of the same age as the Athabascan and Northern Australian producing regions.

Other areas worth exploring are those in which Precambrian igneous rocks are faulted against red oxidized sandstones. Although of limited extent, these areas provide geologic conditions similar to those of the Chatham, Virginia, uranium deposit.

Wyoming is a uranium province (Stuckless, 1978), with uranium occurring in rocks of nearly all major time divisions in the state. Uranium was available for mobilization during every major weathering period related to the nonconformities described. The key to exploration is detailed examination of the basal igneous and metamorphic units. These should be studied for structural trends and lithology. Areas of favorability along the nonconformities can be determined where the basal units are faulted and fractured and sometimes folded, and contain reductants, particularly graphite but also pyrite or other sulfides. Copper Mountain is a good example of an area of basal Precambrian rocks that are structurally prepared and contain reductants. This and other areas need to be explored by detailed geophysical methods to locate structurally and lithologically favorable zones beneath the nonconformity. Electromagnetic and conductivity or resistivity methods are most useful, and should be supplemented by radiometric (gamma or alpha particle) and geochemical (ground-water) studies. Favorable areas should then be drilled.

Given the impressive length of exposure, the relatively shallow subcrop depths of favorable nonconformities in Wyoming, and the great amounts of uranium available for mobilization, a nonconformity-related uranium deposit should exist somewhere in Wyoming. The Geological Survey of Wyoming is conducting field studies to define favorable nonconformities and locate areas along them where lithology and pre-nonconformity structures could have contributed to the formation of uranium deposits.

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Origin and characteristics of Wyoming bentonite deposits

David L. Rath

Introduction

Wyoming bentonite deposits are the result of the in situ alteration of volcanic ash deposits in shallow marine environments. The alteration products are primarily clay minerals of the smectite group, predominantly montmorillonite and sometimes beidellite. Most Wyoming bentonite is composed of a sodium-rich montmorillonite that gives Wyoming bentonites special qualities not associated with the more common calcium-montmorillonite bentonite. Wyoming bentonite swells many times its volume in water and forms viscous suspensions and thixotropic gels in water containing relatively low concentrations of bentonite; it has high dry-bonding strength, high adsorption capacity, and the ability to form an impervious seal. Such properties make Wyoming bentonite useful for many industrial, construction engineering, and agricultural applications.

Bentonite was first commercially produced in Wyoming in 1888 at the Taylor Ranch north of Rock River. The product was initially called taylorite but because the name taylorite was already in use, Knight (1898) changed the name to bentonite after the Fort Benton Formation from which it was mined. Since then, numerous uses for bentonite have been discovered, and major markets have developed. Despite many fluctuations caused by changing economic conditions and market adjustments, demand for bentonite has increased over the years.

In 1984 bentonite companies produced 3,083,324 tons of Wyoming bentonite (Ad Valorem Tax Division, 1985). This figure is up by approximately one million tons from the 1983 production total (Ad Valorem Tax Division, 1984). It is estimated that about half of the 1984 production was sold to the drilling mud industry. Pelletizing and foundry sand bonding were the next largest uses. The balance was consumed by a variety of other industries. The price for all grades of bentonite as of December, 1985 ranged from \$28.60 to \$45.50 per ton (Anonymous, 1985).

Origin

The source of the volcanic ash that formed Wyoming's bentonite deposits was explosive volcanoes active in western Wyoming and Idaho during the Cretaceous. In their report on the Mowry Formation bentonites in the Bighorn Basin, Slaughter and Earley (1965) theorized that these volcanic eruptions were associated with regional tectonics and the emplacement of the Idaho batholith. The large ash clouds were carried eastward by the prevailing westerly winds. The source direction is evidenced by the thick and coarse-grained deposits in central Wyoming that thin and fine eastward toward the Black Hills. Models for the transport and depositional distribution of such ash falls have been developed by Slaughter and Hamil (1970). The process was very similar (but on a much larger scale) to the 1980 Mount St. Helens eruption.

During the Cretaceous much of Wyoming was covered by a shallow sea. Alteration of the ash probably began immediately upon contact with the sea water and continued as the ash particles filtered down to settle on the bottom. Water-soluble salts coating the ash particles would have dissolved first, altering the chemistry of the sea water. Recent studies of leaching characteristics of Mount St. Helens' ash were conducted by the U.S. Geological Survey (Smith and others, 1982). Sea water chemistry as well as the degree of circulation and off-shore currents probably played an important role in the alteration process. In the ocean environment, the volcanic glass began to dissolve quickly, freeing the elements necessary to form montmorillonite. The heavy insoluble constituents of the ash settled faster and are concentrated toward the bottom of the bentonite beds.

Characteristics of deposits

Outcrops of bentonite deposits occur in many parts of the state. Those of commercial interest are usually found where Cretaceous sediments are exposed. Locations of major bentonite deposits are shown in Figure 1. The most persistent bed of bentonite is the Clay Spur bed which occurs at the top of the Cretaceous Mowry Formation. Historically, most of the high-yield bentonite mined has come from the Clay Spur bed. Other formations hosting bentonite of commercial interest include the Early Cretaceous Thermopolis Shale (central Wyoming) and Newcastle Formation (Black Hills district); and the Late Cretaceous Frontier Formation (central Wyoming) Belle Fourche Shale, and Greenhorn Formation (Black Hills district).

Because of varied surface topography and enclosing rock competency, the bentonite outcrops may be covered by surficial material. Generally, however, an abrupt change in vegetation or a profusion of selenite crystals on the surface will indicate bentonite. Outcrops exposed to weathering will often exhibit a frothy or popcorn-like texture caused by swelling and shrinking of the bentonite during wet and dry conditions. Calcium bentonites or bentonites of low swelling capacity will exhibit an alligator-type cracked surface. The average moisture content of bentonite is approximately 30 percent. The material does not appear to be that wet because much of the water is adsorbed between the layers of the clay. Unweathered bentonite has a waxy texture.

The contact with the underlying rock is sharp. The rock immediately below the bentonite is usually very siliceous and is capped by a thin layer of glassy to cherty material that was formed by silica leached from the ash during alteration. The bottom inch or so of bentonite is very gritty and is usually of poor quality. This layer is usually lighter colored than the overlying material. However, in some beds in certain areas this layer is darker. Bentonite overlying the gritty zone is generally massive or blocky, grading to coarsely laminated and then to finely laminated toward the top of the bed. Color varies with increasing overburden from light yellow, to cream, to yellow green, to olive, to bluish gray. Variations in color reflect the degree of oxidation, which is a surface-weathering related process caused by the infiltration of oxygenated surface water. Generally, the color of all bentonite beds becomes bluish gray to gray with depth. In the transition zone between the oxidized and unoxidized bentonite the bentonite has a blocky character. Joint surfaces are stained orange red with iron oxides and sometimes filled with selenite or calcite. Depending on the amount of weathering, the blocks have a bluish gray core surrounded by a yellow rim of oxidation.

The contact with the overlying shale is gradual. Fine laminae of organic-rich bentonitic shale alternate with bentonite, and the proportion of shale laminae increases upward. The thickness of the

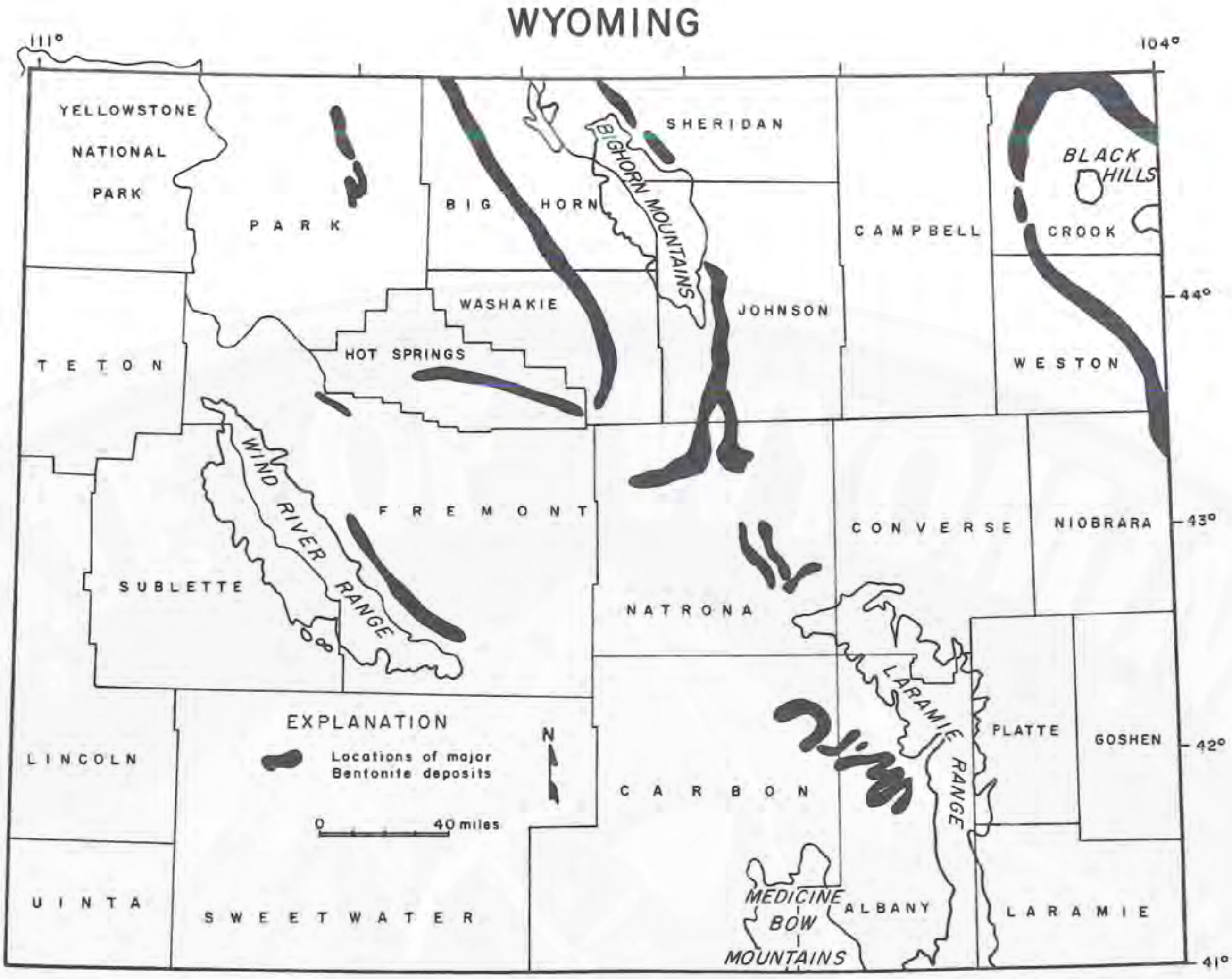

Figure 1. Location of major bentonite deposits in Wyoming.

zone varies from a few inches to as much as half the thickness of the main bentonite bed. Locally, this zone is called "bonded clay". Often this bentonite is of poor quality and is discarded as overburden during the stripping operation. However, this layer should be spot checked because in a few cases it has been of good quality and sometimes even of higher quality than the underlying material.

Local and regional structural features resulting from tectonic events that occurred since the deposition of the bentonite have had a significant influence on the character of the bentonite deposits. Regionally, dips range from nearly level to nearly vertical. In the northern Black Hills, folding and erosion associated with the Black Hills uplift has repeatedly brought the bentonite beds to the surface (Knechtel and Patterson, 1962). Locally, bentonite beds exhibit rather abrupt thickening and thinning, which may be due to depositional or tectonic events. Thickness of beds may vary from a few inches to more than 10 feet and average between 2.5 and 5 feet. Swales and rolls are common features. The beds in many places are crisscrossed with faults, most with displacements of less than 10 feet. A number of deposits are controlled by small combination structures (Davis, 1965). In addition, local structure can significantly affect the quality of the bentonite because faulting and joints in the overlying shale act as conduits for surface water to penetrate the beds.

Mineralogy

Unaltered mineral assemblage studies and estimates of the composition of the unaltered glass indicate the composition of the original ash for Wyoming bentonites varied from rhyolite to dacite or trachyte to latite (Slaughter and Earley, 1965).

The clay mineral montmorillonite of the smectite group is the major constituent in bentonite (65 to 90 percent). It is this clay, when sodium is the dominant exchangeable cation, that is responsible for Wyoming bentonite's unique properties. Montmorillonite occurs in the bentonite as extremely small flake-like particles composed of a number of stacked montmorillonite layers. The structure of montmorillonite has been described by Hoffman and others (1933), Marshall (1935), and Hendricks (1942). In addition, literature relating to the chemical and physical properties of montmorillonite has been summarized by Ross and Hendricks (1945), Grim (1953) and Bleifus (1972). Essentially, montmorillonite is composed of three structural sheets: an octahedral sheet of aluminum coordinated with oxygen and hydroxyls sandwiched between two silica tetrahedral sheets. The sheets form a structural layer by sharing oxygen ions. Within this arrangement a considerable amount of substitution occurs. In the tetrahedral sheet, some Al^{+3} may substitute for Si^{+4}, and in the octahedral sheet Mg^{+2}, Fe^{+2}, and Fe^{+3} will readily substitute for Al^{+3}. A typical chemical analysis of Wyoming bentonite shows the percent of major elements present as oxides:

SiO_2	54.48%
Al_2O_3	20.23
Fe_2O_3	1.19
FeO	2.00
Na_2O	2.40
CaO	0.61
MgO	2.01
K_2O	0.49

Such substitutions cause a positive charge deficiency on the structural unit layers that attracts available alkaline cations. These cations, which are usually Na^+, Ca^{+2}, and Mg^{+2} and sometimes K^+ and H^+, are weakly attracted to positions between the layers and, therefore, are readily exchangeable for other cations. When Na^+ and Ca^{+2} are the cations filling the exchange sites, water is adsorbed between the montmorillonite layers. Ca^{+2} provides a stronger bond between the layers; however, when Na^+ is the dominant cation, a greater number of water layers may be adsorbed, thereby expanding the crystal structural layers in the c-axis direction. This accounts for Wyoming bentonite's ability to swell many times its volume when wet. A typical distribution of exchangeable cations is shown below:

	MEQ/100 grams
Sodium - Na^+	62
Calcium - Ca^{+2}	23
Magnesium - Mg^{+2}	9
Potassium - K^+	1
	95

MEQ/100 grams = milliequivalents per 100 grams

Other clay minerals sometimes present in minor amounts are kaolinite and less often, illite. Non-clay minerals found in bentonite are feldspars, quartz, cristobalite, biotite, zeolite, mica, and minor amounts of minerals of igneous origin. Selenite, iron oxides, and some carbonates and sulfates may also be present.

Quality

The desired quality of bentonite depends primarily on its intended use. Properties desirable for one use may be undesirable for another use. Historically, natural bentonite that exceeded the American Petroleum Institute's (API) specifications, in which one ton of bentonite yields more than 92 barrels of drilling mud having a viscosity of 15 centipoises, was considered high grade. However, bentonite has numerous other uses and nearly as many specifications. Information on published specifications may be found in the references listed. In many cases specifications are established by the individual customers, some of whom consider such information as proprietary process information. Because most published work in this area has dealt with correlating physical and chemical properties of bentonite to standard API specifications, this discussion will be limited to rheologic properties desirable for drilling mud.

Attempts to correlate physical and chemical properties of bentonite with rheologic properties have been mostly unsuccessful. A recent attempt at correlation was conducted by the U.S. Geological Survey (Wolfbauer, 1977). The immediate problem with bentonite is that some of the rheologic properties do not correlate with one another. Most bentonite specifications include more than one property. For example, API standards include specifications for a minimum viscosity and a maximum water loss or filtrate loss. Bentonites characterized by low water loss exhibit a wide range of viscosities (Wolfbauer, 1977). In the Wolfbauer study, exchangeable cation composition and cation exchange capacity were selected because of their strong influence on the quality of bentonite. In addition to cation exchange capacity and the amount of exchangeable cations present, the ratio of cations present is also important. A good quality bentonite will usually have a ratio of $Na^{+}/Ca^{+2} + Mg^{+2}$ of 2 or more; however, because of other variables, this ratio is no guarantee of good quality. Other factors that may influence the quality of bentonite under certain conditions are the ratio of Fe^{+2} to Fe^{+3} (Foster, 1953); net charge imbalances and selective adsorption of cations (Solmonson, 1961; Slovinski, 1958); presence of other minerals; the degree of crystallinity; and particle size and shape.

Most of the factors mentioned above are interrelated. Some may vary widely without having much effect on the quality of the bentonite, while others, such as the $Na^{+}/Ca^{+2} + Mg^{+2}$ ratio, will induce large changes in quality with only slight shifts in value (Williams and others, 1954). In areas where bentonite exists under light overburden, weathering has altered the chemical characteristics of the bentonite, primarily by the introduction of alkaline cations leached from the overlying formations, and by oxidation. In many cases, such weathering significantly improves the rheologic properties of the bentonite. However, excessive weathering is generally detrimental.

Reserves

A conservative estimate of remaining reserves is around 90 to 100 million tons. Most companies do not willingly give out reserve figures. As with all reserve estimates, figures change significantly with changing economic conditions. Geologically, there is considerably more bentonite available than is presently commercially mineable.

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Mineral deposits associated with alkaline intrusive complexes: examples from Wyoming, Colorado, Montana, and California

Theodore J. Armbrustmacher

Ore-forming processes have operated in and around certain alkaline intrusive complexes, resulting in economic and near-economic deposits of rare-earth elements, niobium, titanium, copper, iron, thorium, vermiculite, phosphate, and other commodities. Mineral deposits have been found in nearly every part of a typical alkaline intrusive complex: early stage mafic-ultramafic rocks are known to contain iron and titanium minerals concentrated in magmatic segregations (Powderhorn, Colorado), and vermiculite formed by hydrothermal alteration and weathering (Rainy Creek, Montana); more leucocratic rocks, such as nepheline syenite, ijolite, and others, can contain disseminated pyrochlore (Miask, U.S.S.R.) or eudialyte (Ilimaussaq, Greenland), and residual concentrations of bauxite (Arkansas); central-core carbonatite can contain disseminated rare-earth mineral (Mountain Pass, California) or niobium (Oka, Quebec), and residual concentrations of phosphate (Cargill, Ontario); fenite zones surrounding the complex can be enriched in uranium and thorium (Ilimaussaq, Greenland); later stage carbonatite dikes can contain anomalous amounts of rare-earth elements and thorium (Powderhorn, Colorado); and latest stage hydrothermal veins can contain abundant thorium and barite (Wet Mountains, Colorado) or copper (Bear Lodge Mountains, Wyoming).

Most of these kinds of mineral deposits tend to be found in *carbonatitic* alkaline complexes, such as the Iron Hill complex at Powderhorn, Colorado. *Gabbroic* alkaline complexes are found in the western United States, for example at Iron Mountain in the Wet Mountains area, Colorado, and at Rainy Creek, Montana, but few mineral deposits, other than vermiculite, appear to be genetically related to these kinds of complexes.

Volcanic geology and mineralization in the upper Wood River-Kirwin area, Absaroka Mountains, Park County, Wyoming

William H. Wilson

Rock exposed in the Wood River-Kirwin area (located 20 to 30 miles southwest of Meeteetse and 45 miles south of Cody, Wyoming), consist of the lower Eocene Willwood Formation, the middle Eocene Pitchfork Formation, and the upper Eocene Wiggins Formation. The contact between the base of the Pitchfork Formation and the top of the Willwood Formation has been mapped as the toe of a detachment fault. Breakaway faults and other vertical faults and shear zones cut and offset both the Pitchfork and Wiggins Formations.

Most of the area is underlain by volcanic rocks of the Wiggins Formation. These andesite flows and flow breccias, which probably represent a shield volcano or large complex vent, intertongue with and grade into a volcaniclastic facies in all directions away from the central Kirwin vent area. Although the vent and alluvial facies are essentially andesitic in composition; the former is more felsic. Total thickness of the volcanic rocks is 6,500 feet. Intrusive rocks consist of granodiorite, andesite, and dacite plugs, as well as many andesite and dacite dikes north-northwest trending to sub-radial around the granodiorites.

Mineralization at Kirwin consists of narrow, steeply dipping, lead-silver, and minor gold and copper veins, essentially clustered radially around the central disseminated copper-molybdenum zone in the Bald Mountain area. Drill hole data indicate that the rocks on Bald Mountain are composed of an outer doughnut-shaped mineralized stockworks; a central volcanic vent zone of intrusive breccia of older rocks and mineralized rock fragments-partially intruded into the stockworks; and fine-grained quartz porphyry (or highly altered and silicified andesite ?) that occupies a major area of the central vent zone. A zone of welded(?) tuff, which occurs to the north-northwest, is of obscure relationship, but may intrude the rocks described above.

The stockworks shows a crude hypogene zoning in which low-grade pyrite and chalcopyrite and molybdenite occur in the center, and high-grade pyrite and chalcopyrite and low-grade molybdenite occur in the inner zone with high-grade molybdenum around the vent rocks contact described above. A secondary enriched zone, containing chalcocite, digenite, and covellite overlies part of the stockworks, which, in turn, is overlain by a barren leached cap.

Copper mineralization in Meadow Creek (approximately 4 miles north of Kirwin) occurs in a multiple pluton composed of a westernmost, older, dark gray, fine-grained granodiorite in contact with an easternmost, younger, light to medium gray, coarse-grained to porphyritic granodiorite. An ill-defined zone trending north-northwest across the western half of the stock is more granitic in composition and is believed to represent the chilled contact facies of the younger granodiorite. The location of the intrusion, as a whole, is believed to have been partially controlled by the intersection of the main east-west Wood River fault and a north-trending fault which offsets the volcanics to the north of the plug.

An area of intense quartz-sericite-pyrite alteration and disseminated copper mineralization occurs in an irregular north-striking zone within the younger granodiorite along the north fork of Meadow Creek. The zone, which is about 3,000 feet long and averages 300 feet wide, contains disseminated chalcopyrite and azurite- and malachite-coated fracture surfaces. Traces of molybdenite(?) associated with quartz veining also occur here. Copper concentrations in this zone range from 200 to 700 parts per million.

Rhyolite and granodiorite plugs, crosscut by andesite dikes, crop out in the Yellow Ridge area at the head of the Greybull River. A northeasterly trending mineralized zone, 200 to 300 feet wide by 2,500 feet long, associated with an andesite (dike ?), is exposed in the southeasterly part of the composite granodiorite (andesite ?) plug. Data are sparse, but malachite, pyrite, chalcopyrite, bornite (?), and molybdenite (?), in order of decreasing abundance, occur as fracture coatings and disseminations in the main mineralized zone and sporadically elsewhere in the intrusive mass.

In summary, mineralization is confined to areas of Wiggins vent facies rocks intruded by granodiorite plugs, which are in turn cut by andesite dikes that are often weakly mineralized. All mineralized areas show varying degrees of bleaching, silicification, and argillic and potassic alteration. Propylitization of andesite flows and breccias is common and generally increases in intensity (along with secondary silicification) towards the centers of mineralization.

Genesis and distribution of trona deposits in southwest Wyoming

William C. Culbertson

As many as 42 beds of trona ($Na_2CO_3 \cdot NaHCO_3 \cdot 2H_2O$), a primary source of the industrial chemical soda ash, underlie an area of about 1,300 square miles in the Green River Basin west of Rock Springs, Wyoming. The 25 thickest and most extensive beds range from 3 to 37 feet thick, from 105 to 870 square miles in areal extent, and up to 3,500 feet below the surface. They contain an estimated 81.7 billion tons of trona and 52.7 billion tons of mixed trona and halite.

The trona beds were formed about 50 m.y. ago by the recurrent evaporation of an inland lake. They occur within a sequence of persistent thin beds of oil shale, marlstone, and tuff, and blanket-like beds of sandstone and mudstone, which constitutes the Wilkins Peak Member of the Eocene Green River Formation.

Three episodes of trona deposition are recognized. Most beds of the first episode (beds 1-18) occupy areas of 300 to 870 square miles in the southern and central parts of the trona area, and locally contain intermixed or interlayered halite. Beds of the second episode (beds 19-23) occupy areas of 200 to 300 square miles in the northwest, and are halite-free. Beds of the last episode (beds 24 and 25) occupy areas of 105 to 150 square miles in the northeast, and also are halite-free.

Relationships between modern wetlands and ancient environments of peat deposition in Wyoming

Fredrick J. Rich

Introduction

Wetlands are a common and geologically significant part of the earth's diverse biosphere. Whether called swamps, marshes, fens, or bogs, these typically lowland areas of impounded or slowly moving water are the home of a great variety of plant species. Usually the wetlands have permanently saturated soils or, in many cases, the plants grow in water which may be many inches, or even several feet deep. These wet conditions can lead to preservation of some of the remains of the plants, and a sediment rich in organic materials accumulates around the plants' roots. The deposition of this organic sediment, which may range from organic clay to sandy peat or peat, makes wetlands interesting to geologists. Given proper burial conditions and a suitable geologic history, these deposits may become coal beds, carbonaceous shales, etc. Wyoming has a vast amount of coal of various ages, all of which was derived from ancient wetlands that may be likened to modern swamps and marshes. This paper presents a comparison of modern wetlands with the ancient swamps, marshes, and bogs of Wyoming.

General characteristics of wetlands

Size

The size of wetland areas is one of the chief variables characterizing them. The smallest may be sinkholes or kettles a few dozen square feet in diameter. Wetlands may appear as singular features, or, more commonly, as aggregates of individuals. Thus, dozens or hundreds of individual lakes and swales comprise the wetland morass of the Coteau des Prairies in northeastern South Dakota, the Kettle Moraine of eastern central Wisconsin, and the karst region of northern Florida. The environmental effect of these lakes, ponds, and marshes may not be significant over a very large area, but they are critical in holding ground water. Other, more extensive wetlands are known, however, which make the marshes of the Coteau vanish in importance. For example, the arctic muskeg lies in a circumboreal fashion from Scandinavia, across northern Asia to Alaska and across Canada. Radforth (1969) estimated that nearly 500,000 square miles of muskeg may exist in Canada alone. The aggregate extent of muskeg worldwide must be tremendous.

Wetlands in the central and southern United States were once also of great extent. Blair (1981), in an appeal to preserve what is left of the North American bottomland hardwood forest, estimated that 50 million acres of the Mississippi Valley and southeastern United States were once covered by swamp. A slightly smaller area of northern Africa is still the habitat of giant marsh grasses and papyrus. There, in the Sudan, the Bahr el Jebel, Bahr el Zeraf, and Bahr el Ghazal cover nearly 11,500 square miles of land surface (Rzoska, 1974).

Bogs, swamps, and marshes of such an extent certainly have a pervasive influence on regional water cycling, nutrient supplies, and sediment deposition patterns. Though the impact may be difficult to access, the tremendous extent of muskeg probably even influences global weather patterns. The actual effect that most wetlands have on the external environment apparently has not been measured in most cases. One swamp, however, has been scrutinized so thoroughly from so many different scientific perspectives that its influence is well known. The Okefenokee Swamp, in southeastern Georgia and an adjacent part of Florida, covers nearly 600 square miles. As headwaters area for both the St. Marys and Suwannee Rivers, the Okefenokee occupies a position of great importance hydrologically. The fact that the swamp occupies a confined basin with a perched water table seems to dictate eventual filling of the basin with peat, and ultimate self destruction of the swamp. However, Reuter and Beck (1974) found that, through breakdown of the peat mass by normal biological processes, and subsequent flushing by the Suwannee River, upwards of 110,000 tons of humic (organic) material is removed from the swamp each year. The humic materials are carried seaward by the tea-colored St. Marys and Suwannee Rivers and probably influence nutrient cycling and sediment deposition at the mouths of the rivers.

Species composition and chemistry

The species composition of modern wetlands is tremendously variable. In fact, it even determines how we refer to them. The term swamp, for example, may be used by the Platte Valley rancher for his patch of cattails, or it may be employed by a Georgia forester as he surveys 100 square miles of cypress forest. Even the professionals confuse the issue at this point, and Bellamy (1967), for example, refers to ephemeral swamp ecosystems when he is, in fact, writing about shallow depressions that fill with rainwater during the monsoons. These swamps of floating and rooted aquatic herbs, which last only a few months, are not considered swamps in this study.

The definition of swamp used here is: a wetland area with saturated soil, which may be partially or completely inundated with water most of the year, and which supports a cover predominantly of trees and shrubs. Swamp is, thus, distinguished from herbaceous communities (marsh, bog, moor, muskeg, fen, etc.). The distinction is important simply because it may take many years, perhaps even centuries for a mature swamp to develop, whereas a grassy marsh may form in just a few years or less. Marshes, bogs, and other wetland communities are dominated by herbaceous species and this also distinguishes them from swamps, in terms of the type of organic sediment they produce. There are sharp differences in the appearance and composition of herbaceous peats and woody peats (e.g. Cohen and Spackman, 1977), and the differences are carried through the coalification process.

The chemistry of wetlands also varies and may be as important a distinguishing factor as vegetation type. European scientists have differentiated many herb-dominated wetlands, depending upon whether the soil is always, sometimes, or only partially saturated. Alkalinity is also an important factor and, in spite of the usual reference to acid swamp conditions, alkaline wetlands are also significant. Figure 1 illustrates some of these chemical environmental parameters and identifies the types of wetlands associated with particular combinations of them.

Hundreds of plant species found in various parts of the world have adapted to the peculiar physical and chemical properties of the wetland environment. To treat the many plant communities here would take too much time and space, and, thus only a few kinds of modern wetlands will be discussed. Only a few swamp/ marsh types appear to have been impor-

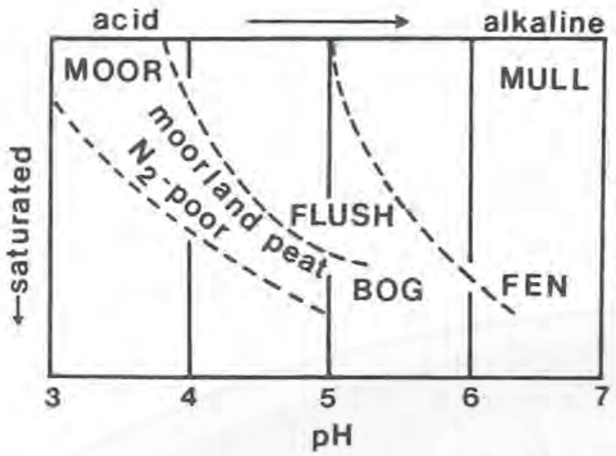


Figure 1. Some chemical and physical characteristics of European wetlands (After Pearsall, 1965.)

tant in Wyoming's distant past, and those plants must certainly have had physiological characteristics similar to modern species. Plants of all wetland habitats, for example, share certain capabilities that make them successful where upland species fail. Foremost is their ability to withstand drowning in a habitat that is often submerged. Meanley (1972) notes that parts of the White River swamp in Arkansas may be submerged under 34 feet of water for days at a time, with no apparent ill effects on the plants. However, it is also true that even cypress trees *(Taxodium distichum)* can be drowned if the water stays too deep for too long. Just how swamp/marsh species maintain themselves in a flooded environment is not well understood, and problems associated with gas exchange, in particular, are unresolved.

Nutrient absorption presents another difficulty for wetland species, especially those inhabiting bogs and swamps of high acidity. The low pH conditions in which many species live (pH 3.5-5.0) have led to a low plasma permeability, or salt absorption ability. At a pH of 4, even relatively high concentrations of mineral salts in the water will not be absorbed. Decreased acidity, on the other hand, results in a disruption of the plasma permeability and cells become flooded with salts (Ruttner, 1953). The low plasma permeability, coupled with the naturally low nutrient content of swamp or bog waters results in such places being fairly sterile environments, habitable only by certain well-adapted species. Exceptional members of the acid-water community have developed unconventional means of compensating for the meager nutrient conditions in their environment. Thus, many modern species have adopted the carnivorous habit and trap and digest insects and small crustaceans to supplement their intake of nitrogen and phosphorous (Darwin, 1893; Erickson, 1968).

History of ancient wetland development in Wyoming

Wetlands have evidently been important world wide since mid-Paleozoic times. Flourishing Devonian swamps produced coal beds of variable thickness and extent in such places as Virginia and Norway. The area now occupied by Wyoming did not lie within the limits of the Paleozoic coal swamps, however, and appreciable amounts of peat (and, hence, coal) did not accumulate in Wyoming until the earliest Cretaceous. The Cambria coal, found near Newcastle, Wyoming, in Weston County is earliest Cretaceous and is associated with fluvial sediments of the Inyan Kara Group. The peculiar nature of the coal and the nearly complete lack of macrofossil remains from the coal-bearing units leave the Cambria coal a genuine enigma. Knell (1985) showed that the coal may have formed as transported plant debris accumulated in a large enclosed flood basin, perhaps similar to a lake basin.

The majority of Wyoming coals are Late Cretaceous and early Tertiary. Among the 11 types of modern wetlands that Rich (1982) discussed as analogues to ancient environments of peat deposition, three appear to have been most important in Wyoming during the Late Cretaceous and (or) Tertiary. The three types include, (1) deltaic swamps, (2) non-deltaic coastal plain swamps, and (3) inland river swamps. The significance and characteristics of these swamp types are discussed below.

Deltaic swamps

Coal-bearing units of undisputed deltaic origin are found principally within the Cretaceous System. Tewalt and others (1983) and Ayers and Kaiser (1984) describe what they consider to be deltaic deposits within the Paleocene Fort Union Formation. These deltas are presumably entered Lake Lebo from the margins of the Powder River Basin. Though substantiation of the lacustrine origin of the Lebo Shale may prove the existence of such lacustrine deltas, it seems much more likely that the sand bodies that Tewalt and others (1983) and Ayers and Kaiser (1984) identified were actually deposited in humid alluvial fans.

The best-documented deltaic sequences occur in the Upper Cretaceous Rock Springs and Blair Formations (Horne and others, 1980) and the Adaville Formation (Lawrence, 1982). Coal beds in those formations are associated with fluvial sandstones, interdistributary siltstones and clays, and occasional oyster beds that represent brief marine incursions. The thickness of the 32 reported beds of the Adaville Formation vary considerably, with the thickest one being the Adaville no. 1 (at 74 to 84 feet, Glass, 1978). Lawrence (1982) attributes the variable thickness and lateral extent of the beds to the dynamics of the delta. Rapid deltaic progradation, for example, would produce thin continuous peat beds, while strand-line stabilization would produce extraordinarily thick peats, and delta abandonment would result in destruction of the swamp flora and its peat deposit.

A deltaic system as active as the one that produced the Adaville sequence is likely to have been very dynamic. Overall progradation and recession of the deltas, and local stream channel shifting or delta lobe abandonment would have created a variety of habitats suitable for many different plant communities. Opportunistic herbaceous species could have taken residence in ephemeral habitats, such as sandy distributary mouth bars or point bars in streams, while more slowly growing species probably dominated the flood-plain swamps. In all likelihood there are distinct plant communities, ranging from marshes to swamps or, perhaps, floating bogs. These types of communities are common on the Mississippi and Fraser River Deltas (Montz and Cherubini, 1973; Styan and Bustin, 1983, respectively), and have resulted in deposition of variable quantities of distinctly different types of peat. Unfortunately, there are not many published accounts of the paleobotany and palynology of the Cretaceous deltaic coals. We can only guess which plants occupied the marshes. Grasses and sedges had not yet evolved, and considering the great diversity of ferns and their allies that were common in Mesozoic times, it may have been ferns that comprised the herbaceous vegetation. This contention is based in part upon work by Parker (1976) on the Upper Cretaceous Blackhawk Formation in Utah. The concept is also supported by analyses of a megaflora from the Frontier Formation of Western Wyoming (Tidwell and others, 1976). The Fronter is slightly older than the Adaville Formation, but the plant communities were probably similar during deposition of the two formations. Andrews and Pearsall (1941) reported seven species of ferns among their 25 described Frontier plant species. Trees may have been represented by a variety of conifers and hardwoods. Tidwell and others (1976) mention numerous authors who have reported *Sequoia, Ficus, Myrica, Magnolia, Salix, Populus, Platanas, and Sabalites* from sediments in Utah which are stratigraphically equivalent to the Adaville. Reconstruction of the ancient Adaville delta swamp/marsh communities must await further detailed paleontological research. Palynostratigraphic work of the type performed by Nichols and Jacobsen (1982) would be most welcome, though emphasis should be placed on paleoecological reconstruction rather than biostratigraphy.

Non-deltaic coastal plain swamps

Nondeltaic swamps are very common features along the coastal plain in the eastern United States. Tens of thousands of square kilometers of southern New Jersey, eastern Virginia, the Carolinas, Georgia, and Florida are covered with mixtures of conifer and hardwood swamps and extensive marshes. White cedar swamp is common in New Jersey and Virginia, whereas cypress swamp is prevalent further south where Carolina bays are common features. These bays are small lakes or ponds whose origin is problematical, but which are associated with extensive fresh-water swamps. The floral diversity of the coastal plain swamp/marsh complex culminates in the Everglades, which is the single largest tract of this type of landscape remaining in America. Carr (1973) estimated that the Everglades covers 13,000 square miles of southern Florida. Folkerts (1982) describes a less extensive, and rather bizarre element of the Gulf Coastal Plain swamps, pitcher plant bogs. Though these communities of carnivorous plants may have once covered nearly 1,100 square miles of Alabama, Mississippi, and Louisiana, they have been reduced to a fraction of that extent.

The white cedar and cypress swamps or pitcher plant bogs all exist where they do because drainage is impeded on the coastal plain by the very low relief. The water tables are also very high, being near or actually above the ground surface in many places, and the climate is mild. These are excellent conditions for plant growth and peat deposition, so the coastal plain of the eastern and southeastern United States is almost literally covered with peat.

The non-deltaic coastal plain swamp/marsh/bog complex has some ancient counterparts, although this type of wetland does not appear to have been as common in the past as it is now. Livingston and Wilson (1982) describe lower coastal plain fluvial/paludal deposits from Carbon County, Wyoming. There, within a lower coal-bearing unit of the Almond Formation, sediments consist of a fluvial, fresh-water dominated system of channel, overbank, splay, and swamp deposits typical of lower coastal plains (Livingston and Wilson, 1982). Massoth (1982) described a similar interdeltaic coastal plain sequence near Craig, Colorado. The tabular coal seams there are laterally continuous (30 square miles) and of moderate thickness (1 to 15 feet). While Massoth (1982) illustrated swamps, marshes, and lakes or ponds in his paleoenvironmental reconstruction, the actual paleobotanical composition of those coal-forming communities remains to be described. In view of the great variety of modern plant communities on the coastal plain, one might expect the Cretaceous coals to have been derived from diverse plant types.

Inland river swamps

Inland river swamps comprise one of the most visible of modern wetland types. Nearly every permanent stream of any size has some swamp/marsh land within its flood plain. The wetlands may be as apparently insignificant as the cottonwood groves and marshes along the Platte River in Wyoming, or as extensive and impressive as the tupelo forest lining northern Florida's Chipola River. Modern inland river swamps vary in composition from the tupelo/cypress forests of the Deep South, to canebreak marshes of western Tennessee, and silver maple/rover birch forests of Wisconsin and Illinois. Maximum peat development probably occurs in the South, where the growing season is long and the climate is mild. The best examples of inland river swamps in North America are those of the Deep South. They line the banks of rivers with names like the Suwannee, Appalachicola, Choctawhatchee, and Pascagoula. There, and inland along the

Mississippi River, some of the most extensive wetlands in the United States once existed.

Ancient river swamp deposits are usually associated with prominent channel sandstones. Thus, the fluvial Fort Union and Wasatch Formations of the Power River Basin contain not only numerous, often thick coals, but abundant channel sandstones as well. The work of Flores and others, and Obernyer (*in* Glass, 1980) and many other workers has shown very clearly how the early Tertiary swamps grew along what were probably large, slowly moving streams. Abundant natural exposures of fossiliferous shales and naturally fired clinker show further that the swamp flora was domianted by conifers (e.g. *Metasequoia*) and cottonwood-like hardwoods, while numerous herbaceous species (e.g. the tropical fern *Lygodium*) lived among the trees. Preliminary petrographic work (Dorsett, 1984) has shown that the vertical composition of the Wyodak bed near Gillette is variable and reflects peat deposition under alternating wet and dry conditions. Petrographic and palynological investigations of the coals from the Power River Basin have only begun however, and it may be several years before we understand the genesis of the coals themselves.

Coal beds within the Powder River Basin are unusually thick. For example, the Wyodak bed at the Wyodak mine is 100 feet thick. Other currently mined coals such as the Badger coal at the Dave Johnston mine, and the Monarch coal at the Big Horn No. 1 mine are also quite thick (16.8 to 26.1 feet respectively, as reported by Glass, 1975). The unusual thickness of Powder River Basin coal beds has been attributed to their youth and low rank and had not been considered to be much more than an oddity until about three years ago when the U.S. Geological Survey reported the discovery of Big George. Pierce and others (1982) describe this super-thick coal bed, which is 182 feet thick at the discovery site. Big George lies 1,100 feet below the surface, and extends over 950 square miles. Current exploration and research will provide us with more information about Big George. Until then it is tempting to speculate about Big George, the Wyodak bed, and other unusually thick coals, and relate their formation to some reasonable depositional/tectonic framework.

Several fundamental ideas must be considered first:

(1) Pierce and others (1982) note that throughout the basin, coal resources are commonly concentrated in definable subareas where several coal beds merge locally to form thick deposits. B.H. Kent (U.S. Geological Survey, personal communication, 1982) has shown that these subareas usually lie in a north-south orientation.

(2) The axis of the Powder River Basin strikes north-south.

(3) The Powder River Basin is strongly asymmetrical, with the axis adjacent to the base of the Bighorn Mountains.

(4) Regional paleochannel flow directions were south to north (Etheridge and Jackson, *in* Glass, 1980; Flores, 1982).

(5) Sand percent maps and distribution of lithologies in the subsurface suggest that during deposition of the Tongue River Member of the Fort Union Formation the basin was filled by two deltaic lobes that were supplied sediment from an eastern source in the ancestral Black Hills (Tewalt and others, 1983).

(6) The Powder River Basin was subsiding while the Black Hills and Bighorn Mountains rose.

Certain assumptions may be added:

(1) The basin was not filled with water before deposition of the deltaic lobes, therefore they were not deltas. More likely they were akin to alluvial fans.

(2) Maximum peat thickness probably developed during periods of greatest, or

most rapid subsidence.

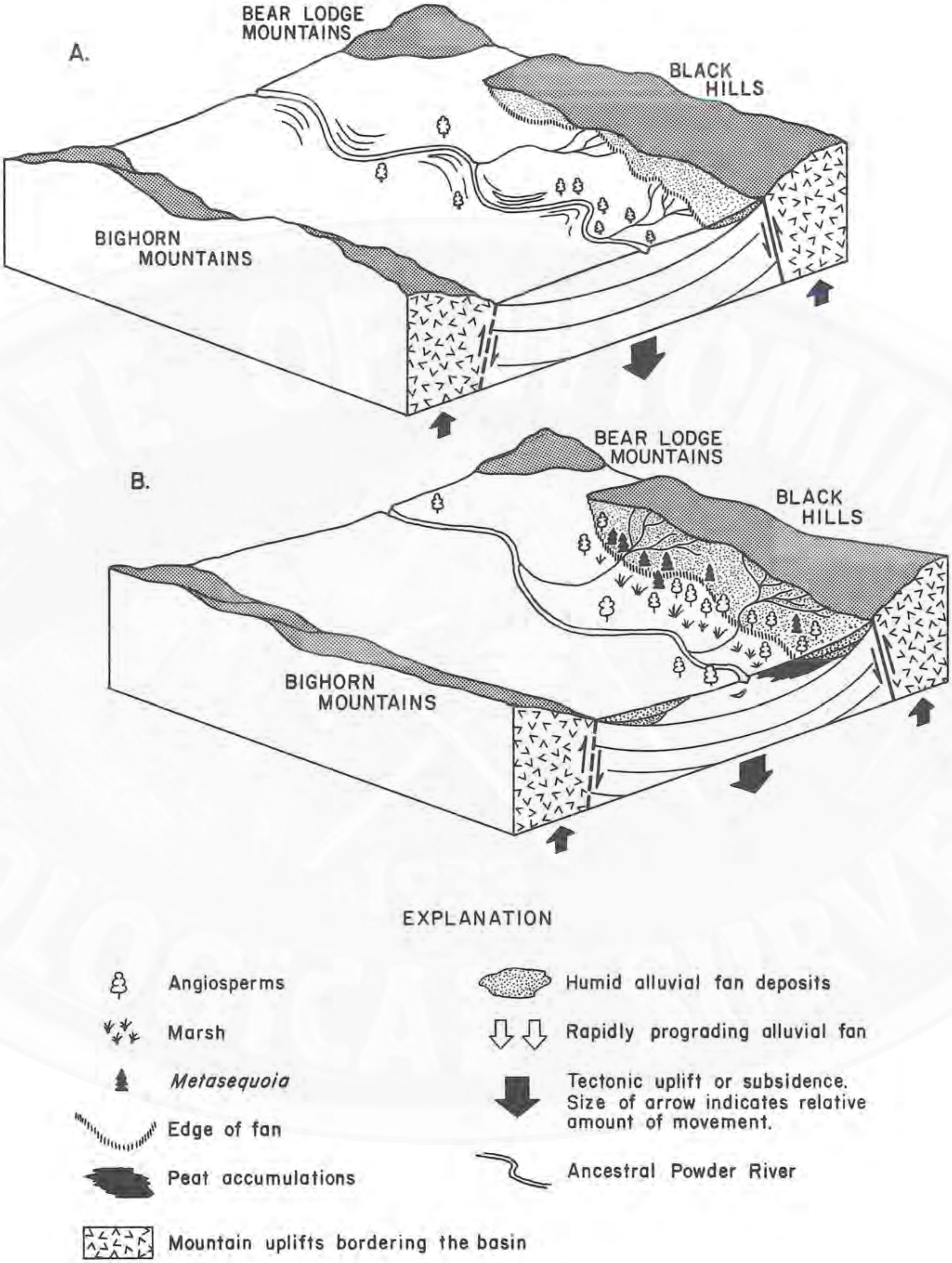
(3) The swamp/marsh vegetation probably grew in the most favorable location topographically, not in the major stream channels, which lay always to the west, and not on the edaphically dry alluvial plain to the east, but at the interface of the alluvial deposits and the flood plain of the rivers.

The tectonic scenario proposed here is that the basin and adjacent areas of uplift were intermittently active, with periods of basin subsidence and (or) mountain block uplift alternating with periods of tectonic quiescence. The asymmetry of the basin developed gradually, with the axis starting and remaining in the west, or migrating westward to its present position.

Coarse clastics were stripped from uplifted areas and deposited in the basin either as alluvial deposits that built out into the basin from the slopes of the Black Hills and Bighorn Mountains (Figure 2A), or as finer-grained clastics which, upon reaching the major northward-flowing stream systems, were diverted toward the Williston Basin. Sediment production and transport would have been of intermittent intensity depending upon tectonic activity. If the basin was subsiding comparatively rapidly, the rivers presumably stayed within narrowly defined flood plains. With much slower rates of subsidence, the rivers meandered widely and reworked their own flood-plain sediments, and, perhaps, the distal margins of the alluvial deposits.

The peat-producing environments gradually moved laterally, from east to west, as the alluvial deposits encroached from the east, and the basin subsided progressively more in the west. The zone of optimal topographic relief shifted laterally, in effect taking the swamps and marshes with it. Thus, we can imagine long corridors of swamp vegetation extending north-south, between the major streams and the alluvial slopes. During periods of subsidence, peat produced by the swamps and marshes subsided rapidly into the swampy basin (Figure 2B). The warm, humid climate would have encouraged rapid, abundant plant growth. As very thick layers of plant debris periodically accumulated beneath the swamps and marshes, tremendous quantities of peat must have been produced. The very thick peats later became the unusually thick coal beds exemplified by the Wyodak coal. When subsidence slowed, river meandering disrupted peat accumulation by disturbing the swamps and reworking the peat deposits (Figure 2C). Layers of clastic sediment eventually covered the peat beds, perhaps destroying the swamp vegetation. Slow subsidence would also have kept the peat beds at the ground surface for long periods of time, where it was destroyed by oxidation and biological activity. This could account for the many charcoal-rich zones that Dorsett (1984) identified in the Wyodak coal. Such periods of slow basin subsidence might also have caused thin beds of peat to accumulate, resulting in thin coals. The thin coals that overlie some of the thicker beds were probably produced under these environmental and tectonic conditions. Large influxes of stream sediment periodically terminated peat accumulation altogether, and produced thick sequences of sandstones, shales, and siltstones such as those along Interstate-90 between Gillette and Sheridan. When rapid basin subsidence resumed, thick peats began to accumulate again at the distal margins of the alluvial slopes (Figure 2D), and another cycle of river stabilization and eventual meandering began.

The test of this theory will be to prove the lateral relationships of fluvial, paludal, and alluvial facies that have been described here. It will also be essential to describe the tectonic activity of the Powder River Basin using as many sources of information as possible in order to determine if the episodic nature described here is accurate. The periodic nature of basin subsidence, which Winczewski and Groenewold (1982) described for the Williston Basin, is quite plausible, and their work adds credibility to the hypothesis offered here.


A.
BEAR LODGE MOUNTAINS
BLACK HILLS
BIGHORN MOUNTAINS
B.
BEAR LODGE MOUNTAINS
BLACK HILLS
BIGHORN MOUNTAINS
EXPLANATION
Angiosperms
Marsh
Metasequoia
Edge of fan
Peat accumulations
Mountain uplifts bordering the basin
Humid alluvial fan deposits
Rapidly prograding alluvial fan
Tectonic uplift or subsidence. Size of arrow indicates relative amount of movement.
Ancestral Powder River

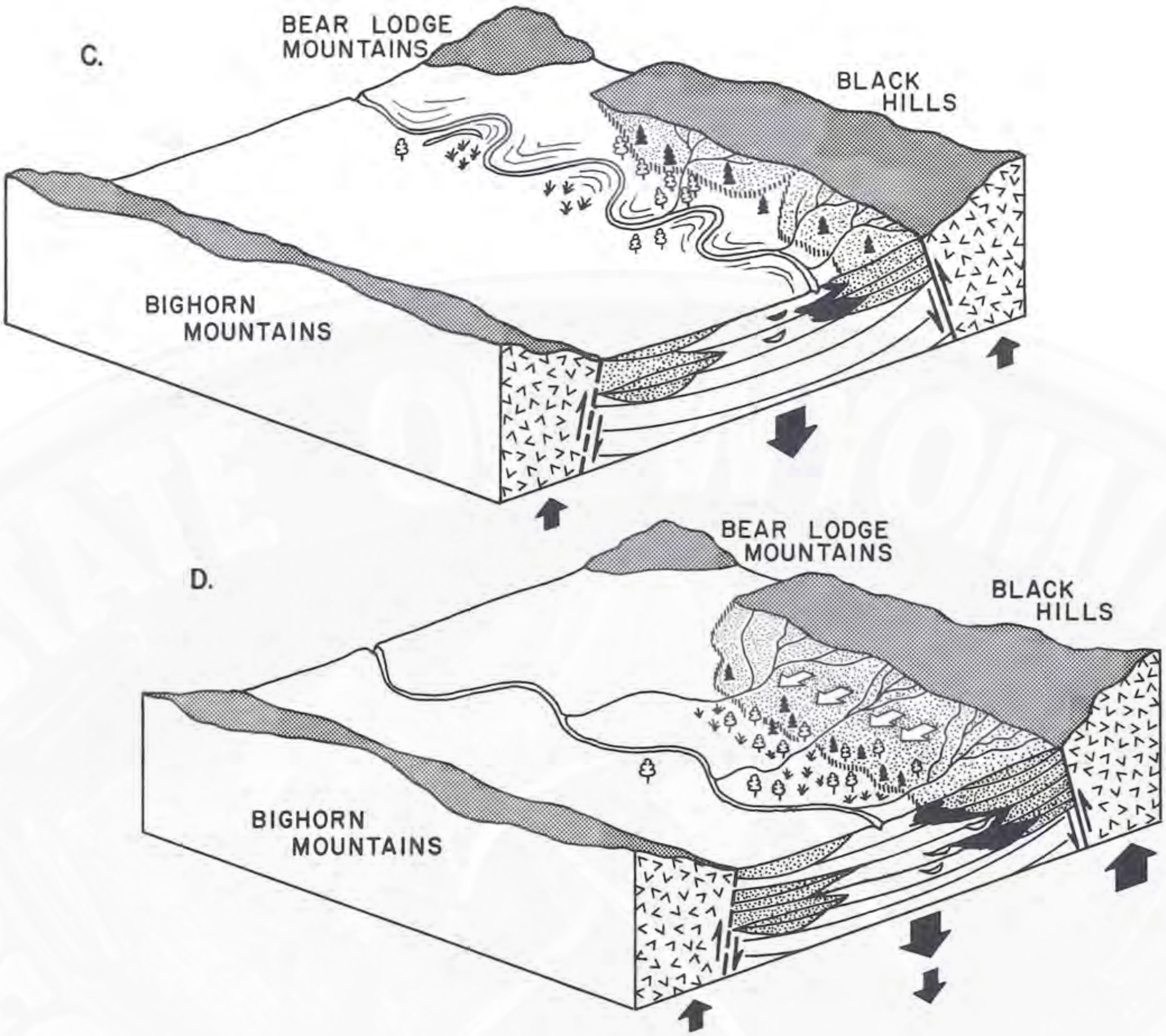


Figure 2. Generalized early Tertiary tectonic and depositional history of the Powder River Basin: (A) Initial uplift of Bighorn Mountains and Black Hills/Bear Lodge Mountains. Symmetrical subsidence of basin, with river lying in the east and alluvial fans developing on flanks of uplifts. (B) Moderate uplift and erosion of Black Hills, but increased basin subsidence. Fan margin, swamp, and river all occupy stable positions for many years, with subsequent accumulation of very thick peat. (C) Subsidence and uplift are both reduced, river meanders, truncating peat deposits and eroding toes of alluvial fans. (D) Basin subsidence and uplift are both renewed. Initial slow subsidence allows alluvial fans to migrate rapidly westward, causing westward shift of river. Subsequent increased subsidence once again causes stabilization of river, fan margins and swamp, with resulting thick accumulation of peat.

Conclusions

A wide variety of wetland plant communities exist. They range from small patches of cattails or grassy marsh, to cypress forest and the circumboreal muskeg. Each different community thrives according to the chemistry and climatic conditions of the environment where the plants grow. Under optimal conditions, each plant community will produce a peat deposit that may be preserved and altered to become coal.

Swamps and marshes of many different ages have produced coal beds in North America and elsewhere. Cretaceous and Tertiary wetlands were the primary source of Wyoming coal beds. Sedimentary and paleontological evidence show that three wetland types were important in producing Wyoming coals. These were: deltaic swamps, non-deltaic coastal plain swamps, and inland river swamps. These three swamp types are common today and suitable modern analogues to the ancient swamp/marsh environments can be identified. The Tertiary inland river swamp deposits of the Powder River Basin have unusually thick coal beds. It is suggested here that the thickness is the result of periodically rapid subsidence of the Powder River Basin and the existence of extensive corridors of swamp that grew on the distal slopes of large alluvial fans. Much more work remains to be done to prove or disprove this theory of thick coal bed formation.

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Zeolites in Washakie Basin

H.D. Curry and K. Santini

The Washakie Basin is a large, subrectangular, equidimensional structural basin located southeast of the Rock Springs uplift and immediately north of the Colorado-Wyoming border. An area of about 2,500 square miles is covered by rocks of the Wasatch, Green River, and Bridger (Washakie) Formations; the ages of these rocks span nearly the entire Eocene epoch. The upper Eocene in particular has long been known to contain considerable volcanic ash. Bradley (1964) and later workers have reported disseminated zeolites and essentially monomineralic units of analcite, mordenite, and clinoptilolite, but only the latter is here considered to have potential commercial importance.

Roehler (1973) noted that a variegated tuff occurring just above the thick, widespread, brown, cliff-forming sandstone that marks the base of the Adobe Town Member of the Washakie Formation, was clinoptilolite-bearing. He traced this unit (his marker bed 579) intermittently almost completely around the basin. The Adobe Town Member is best developed on the north. It is locally missing or obscured by structural complications, or by a large influx of sands in the southeast. The Miocene Browns Park Formation overlies it unconformably on the southwest. Although it is not one continuous bed (it pinches out laterally) and varies in color and sedimentary features, the Adobe Town Member is remarkably persistent. In a few places clinoptilolite lenses are also found just below the basal brown sandstone, but these lenses are small and of minor importance. Similarly, thin zeolitized tuffs occur in places several hundred feet stratigraphically higher in the section.

The color of the main clinoptilolite-rich tuff unit is variegated with pastel shades ranging from off-white through lemon yellow, buff, orange, light olive green, and a distinctive and striking robin's-egg blue. It is dense, fine grained, hard, brittle, splintery, and moderately resistant. The main unit varies from a few inches to as much as twelve feet thick and intertongues laterally and vertically with tuffaceous mudstones and siltstones. The clinoptilolite content may exceed 90 percent, with an ammonia adsorption capacity as high as 2.0 milliequivalents per gram. It is nearly completely altered, with common relict glass shards and scattered biotite flakes.

The volcanic ash apparently had a distant source, probably the Absaroka area of northwestern Wyoming which was volcanically active during the late Eocene. It is a largely reworked ashfall tuff deposited in a relatively high-energy, fluvial-lacustrine environment during one or more short episodes, rather than in a true deep-water lake such as is believed to characterize the underlying Laney Shale Member of the Green River Formation.

Rocky Mountain Energy Corporation staked claims on the clinoptilolite deposits several years ago. This claim block, plus contiguous acreage held in fee, is estimated to contain several million tons of relatively high-grade material. Other areas of zeolites are characterized by thin beds, low grade, thick overburden, and inaccessibility which make mining less attractive.

Although there are many clinoptilolite occurrences in the west, few have such desirable properties, mining configuration, and commercial promise. Eventually, these deposits should prove of real commercial value.

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Minerals exploration in Wyoming

W. Dan Hausel

Introduction

About 10 percent of the State's surface area lies within Laramide uplifts that are cored by Precambrian crystalline rock, and less than three percent (excluding Yellowstone National Park) is covered by Tertiary volcanics. Much of the remaining surface area occurs within broad Tertiary basins (with Cretaceous sediments exposed on their flanks) that are favorable environments for energy and industrial minerals, but generally unfavorable for base, ferrous and ferroalloy, and precious metals (other than placers).

The 10 percent of Wyoming's surface area that is exposed as mountainous terrain contains large regions of relatively unmineralized Archean granite-gneiss. These basement rocks are primarily gneisses, migmatites, and granitic plutons and batholiths. This considerably restricts the potential areas of exploration for metals; however Devonian diamond-bearing kimberlites could occur throughout the exposed Precambrian basement.

Exploration targets

Energy and industrial minerals have historically been more important to Wyoming's economy than metals. However, Wyoming has several regions in which the geology is favorable for metal and possibly diamond deposits. Supracrustal terrains such as South Pass have been the source for nearly all of the iron ore and gold that have been mined in Wyoming. Terrains similar to South Pass are scattered throughout the granite-gneiss basement complex (Plate 1, back pocket). Diamonds have been found in the Precambrian terrain of the Medicine Bow Mountains and southern Laramie Range. Large areas containing Cretaceous to Recent auriferous placers have been reported in northwestern Wyoming, and Tertiary to Recent gold placers occur in southwestern Wyoming.

Wyoming is also relatively unexplored compared to its neighbors. But within the last decade a number of important discoveries, rediscoveries, and interesting finds have been enticing to the exploration geologist. The following list reminds us of Wyoming's potential:

(1) In 1981 significant amounts of gold were detected in analyses of selected samples collected in the Bradley Peak area (Figure 1). These specimens were collected from an historic mining camp (Penn Camp) which lies within metagabbros and ultramafic rocks that form part of the Seminoe Mountains "greenstone belt" (Hausel, 1982a; Hausel and Harris, 1983). The metals are believed to have been mobilized from adjacent country rock and transferred to dilational zones (Klein, 1981). There appear to be similarities between the Bradley Peak gold deposits and the Yellowknife region of Canada (e.g. Boyle, 1959).

(2) Between 1975 and 1978, Precambrian quartz-pebble conglomerates in

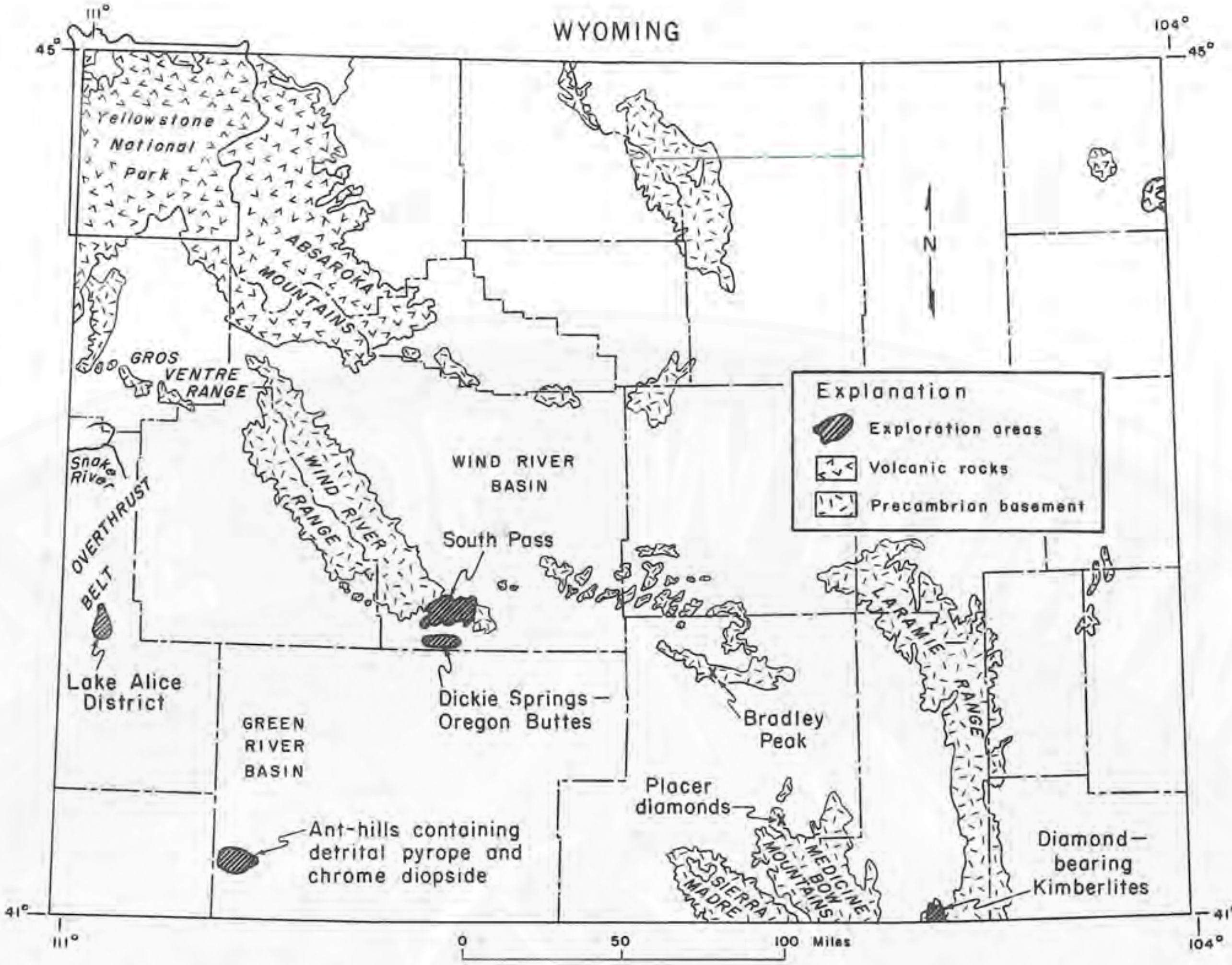

Figure 1. Location of exploration areas and discoveries described in the text.

the Sierra Madre and Medicine Bow Mountains (Figure 1) were discovered to be radioactive (Houston and others 1977; Houston and Karlstrom, 1979). Large regions of these metaconglomerates remain untested for uranium. They are also untested for gold, although it was suggested as early as 1968, that Precambrian conglomerates could contain fossil placer deposits (Houston and others, 1968).

(3) During 1979 and 1980, Conoco Minerals discovered a stratiform, volcanogenic, zinc-copper-silver sulfide deposit in the southern Sierra Madre (Figure 1). This deposit occurs in a eugeoclinal setting with thick successions of Proterozoic calc-alkaline metavolcanic and metavolcaniclastic rock (Divis, 1976; 1977). Although many copper mines and prospects in the area have not been evaluated as volcanogenic deposits, it has been suggested that two nearby mines, the Itmay and the Hinton (Verde), are possibly volcanogenic sulfide bodies (Hausel, 1982a; Hausel and Harris, 1983). The geologic settings of these historic mines have recently been described by Schmidt (1983) and Swift (1982), respectively.

(4) The 1975 accidental discovery of diamond in a mantle nodule collected from a Wyoming kimberlite pipe (McCallum and Mabarak, 1976) led to

the discovery of numerous kimberlite intrusives as well as additional diamond-bearing pipes (McCallum and others, 1977; 1979; Hausel and others, 1979; 1981; 1985). Large regions remain unexplored, but reconnaissance stream sampling in some areas has identified anomalies that suggest the presence of several undiscovered kimberlites. No potential placer deposits have been tested, even though the known kimberlites are considered to be deeply eroded (e.g. McCallum and Mabarak, 1976, Figure 6).

In addition, there are (1) the recent discovery of placer diamonds in the northern Medicine Bow Mountains; (2) the reported occurrence of pyrope garnets and chrome diopside in ant hills over a six to eight square mile area in the Green River Basin (Hausel and others, 1979;,McCandless, 1984); and (3) unverified reports of diamonds in the Wind River Range and in the Gros Ventre Range, (J.D. Love, personal communication, 1981) (Figure 1).

(5) The rediscovery of copper-silver-zinc mineralization in Nugget and Twin Creek Formation red beds of the Lake Alice District (Figure 1) (Love and Antweiler, 1973; Boberg, 1983) led to the discovery of several more occurrences in red beds of the Overthrust Belt (Love and Antweiler, 1973; Hausel and Harris, 1983).

(6) Love and others (1978) reported significant amounts of gold disseminated throughout Wasatch Formation (Tertiary) boulder conglomerates in the Dickie Springs region several miles south of South Pass (Figure 1). More than 28,500,000 ounces of gold may occur within these conglomerates (Love and others, 1978),

(7) Antweiler and others (1977) reported significant amounts of gold in Cretaceous and Tertiary conglomerates and alluvial gravels in the Snake River region (Figure 1).

(8) Several porphyry copper-molybdenum deposits have been discovered in the wilderness areas of the Absaroka Mountains (Figure 1) in recent years (Wilson, 1971; 1975; Fisher and Antweiler, 1980; Fisher and others, 1977; Fisher, 1981; Hausel, 1982b).

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Stream sediment exploration sampling for kimberlite in Colorado-Wyoming and techniques of diamond extraction

Carl Brink, Karl Albert, and W. Dan Hausel

Exploration for kimberlites in Wyoming and Colorado is based on methods that have been effectively used for many years in Africa, Australia, and the U.S.S.R. The most efficient of these is reconnaissance stream sediment sampling for the diagnostic kimberlite indicator minerals (pyrope garnet, chrome diopside, and magnesian ilmenite) that have been released into the surrounding soil and drainages by decomposition of kimberlite.

Stream sediment sampling. Once a promising area has been selected, sample sites are plotted on U.S. Geological Survey 7½- or 15-minute topographic maps. These sites are spaced 0.5 to 1.0 mile apart on all major drainages. This spacing has been chosen because studies done in Colorado (Leighton and McCallum, 1979) have shown that abrasion of grains during transport effectively destroys indicator minerals at relatively short distances from their parent rock. Magnesian ilmenite grains survive up to 2.5 miles, garnet 1.5 miles and chrome diopside only 0.25 miles.

Sample locations in the field are chosen to take advantage of the best possible heavy mineral collection site in the drainage. Samples are collected from one to three feet below the surface. The sample is shoveled into a grizzly pan, and material greater than 1/4 inch (coarse fraction) is visually examined before being rejected. Material smaller than 1/4 inch passes through the mesh and is collected in a gold pan. This fine fraction is repeatedly panned in gently running water until all lighter minerals have been washed away, leaving only a black heavy mineral concentrate. This pan concentrate is briefly examined for distinctive purple to lavender pyrope garnet and emerald green chrome diopside before being placed in a separate bag. Between three and five pounds of concentrate are collected at each site.

At the laboratory, the samples are dried and run through a magnetic separator to remove magnetic minerals. The coarse nonmagnetic fraction is further concentrated in a jig while the fines go to a Wilfley table where material of less than 3.2 specific gravity is removed. After the final concentrates are dried, they are ready for optical examination for indicator minerals.

The most important indicator minerals are pyrope garnet, chrome diopside, and magnesian ilmenite. Pyrope garnet is well-rounded and recognizable by its deep red-purple color and refractive index; chrome diopside is emerald green and has excellent cleavage and parting; and magnesian ilmenite is glossy black and nonmagnetic.

Diamond extraction techniques. Kimberlite is crushed and sieved into two size fractions. After passing through a magnetic separator, the coarse fraction (+16 mesh; >1.19 mm) is passed over a grease table, the grease consisting of a mixture of vaseline and paraffin up to 1/4-inch thick. Diamonds, being nonwettable by water, adhere to the grease while other minerals are washed off. Tests have shown that the grease table is 100 percent efficient on this size sample. The fine fraction (-16 mesh; <1.19 mm) is processed in a skin flotation unit (as well as on the grease table). In the skin flotation unit, the sample falls from a vibrating container onto a smooth tensional water surface created by a constant flow of rising water. Being non-wettable, diamonds float on this surface and are recovered as they pass over the edge of the container. Other minerals break through the surface tension layer and are effectively separated from diamonds. This unit is 80 percent effective.

The concentrates from the grease table and skin flotation unit are placed in a hydrofluoric acid bath to digest silicate minerals. The final sample is examined with a microscope for diamonds. The diamonds are recognizable by their characteristic crystal shapes (octahedrons and dodecahedrons) and surface features (trigons).

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A model for low temperature copper transport and deposition

Robert J. Starkey and B. Ronald Frost

The zonal sequence pyrite-chalcopyrite-bornite-chalcocite $\pm$ native copper has been described in sediment-hosted, syngenetic and diagenetic copper deposits of Zambia and White Pine, Michigan as well as in higher temperature hydrothermal deposits (the Sustut deposit). The variety of geologic settings and rock types in which these deposits occur suggest a similar, geologically simple mode of formation.

Copper mineralization typically is associated with a permeable hematitic oxidized layer and an impermeable reduced layer, leading many workers to deduce that the zonation is a result of progressive reduction of an oxidized, copper-bearing fluid, possibly in the presence of hydrogen sulfide. Simple calculations of copper mineral stabilities from 25°C to 300°C in the presence of aqueous sulfur species in log fo_2 versus pH space show that over geologically reasonable pH ranges at constant aqueous sulfur concentrations, the progression of mineral stabilities with increasing oxygen fugacity is pyrite-chalcopyrite-bornite-chalcocite-native copper. In addition, these calculations reveal that bornite, chalcocite, and native copper are stable with respect to hematite and aqueous sulfate. This suggests that copper and sulfur can be transported in the same fluid, and upon reduction of this fluid, copper sulfides will be precipitated. Calculations also show that native silver is stable with chalcocite and native copper and that lead and zinc are very soluble under the oxidizing conditions associated with native copper. This explains the common occurrence of native silver in low-temperature copper deposits and the absence of galena and sphalerite within the zonal sequence in these deposits.

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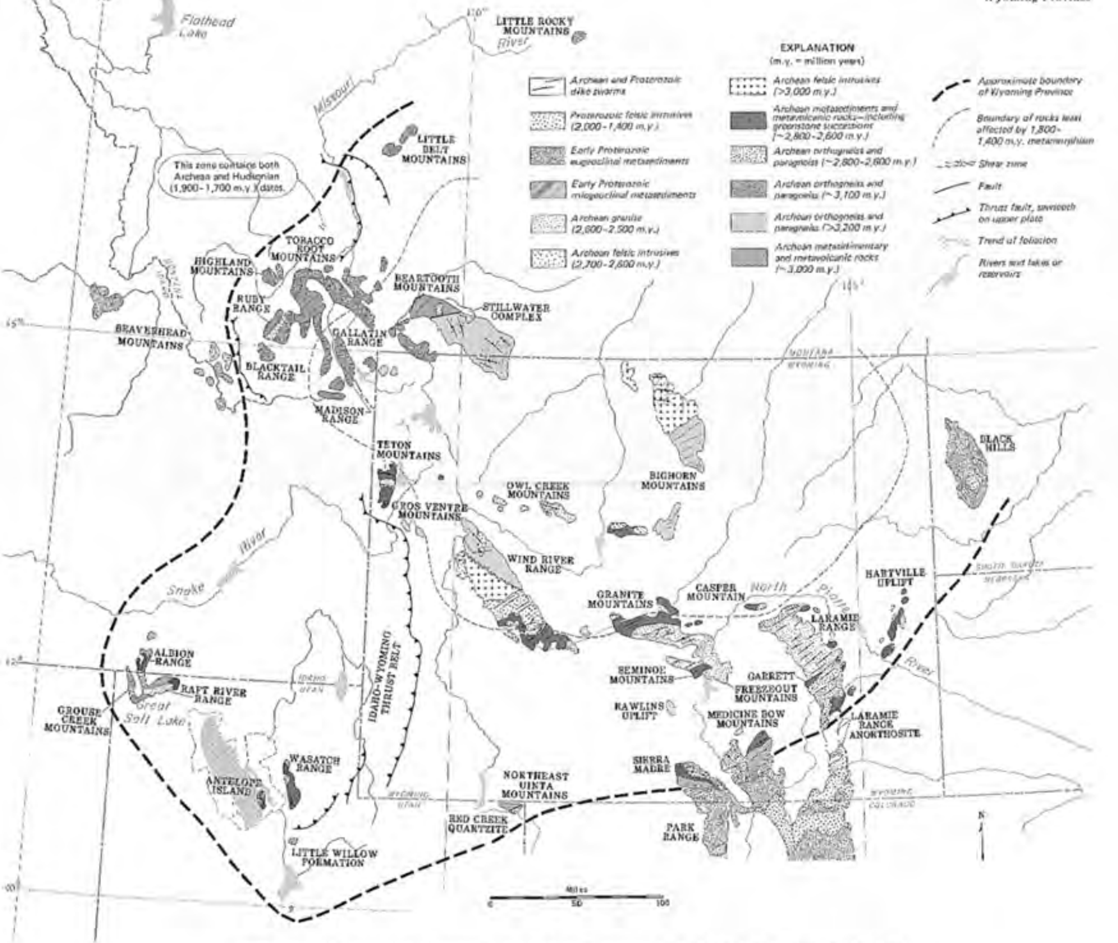

PRECAMBRIAN INDEX MAP OF THE WYOMING PROVINCE

by

Robert S. Houston