URANIUM - BEARING QUARTZ - PEBBLE CONGLOMERATES: EXPLORATION MODEL AND UNITED STATES RESOURCE POTENTIAL

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INTRODUCTION

Uranium has been discovered in fluvial quartz-pebble conglomerates in most of the Precambrian shield areas of the world, including the Canadian, African, South American, Indian, Baltic, and Australian shields. Occurrences in these and other areas are shown in Figure 1.1. Two of these occurrences, the Huronian supergroup of Canada and the Witwatersrand deposit of South Africa contain 20-30 percent of the planet's known uranium reserves (Nininger, 1974). Thus it is critical that we understand the origin of these deposits and develop exploration models that can aid in finding new deposits. Inasmuch as these uranium-bearing conglomerates are confined almost entirely to rocks of Precambrian age, Part I of this review begins with a discussion of Precambrian geology as it applies to the conglomerates. This is followed by a discussion of genetic concepts, a discussion of unresolved problems, and finally a suggested exploration model. Part II summarizes known and potential occurrences of Precambrian fossil placers in the world and evaluates them in terms of the suggested exploration model. Part III discusses the potential for important Precambrian fossil-placer uranium deposits in the United States and includes suggestions that may be helpful in establishing an exploration program in this country. Part III also brings together new (1975-1978) data on uranium occurrences in the Precambrian of the Wyoming Province. Part IV is a complete bibliography of Precambrian fossil placers, divided according to geographical areas. In total, this paper is designed to be a comprehensive review of Precambrian uranium-bearing fossil placers which will be of use to uranium explorationists and to students of Precambrian geology.

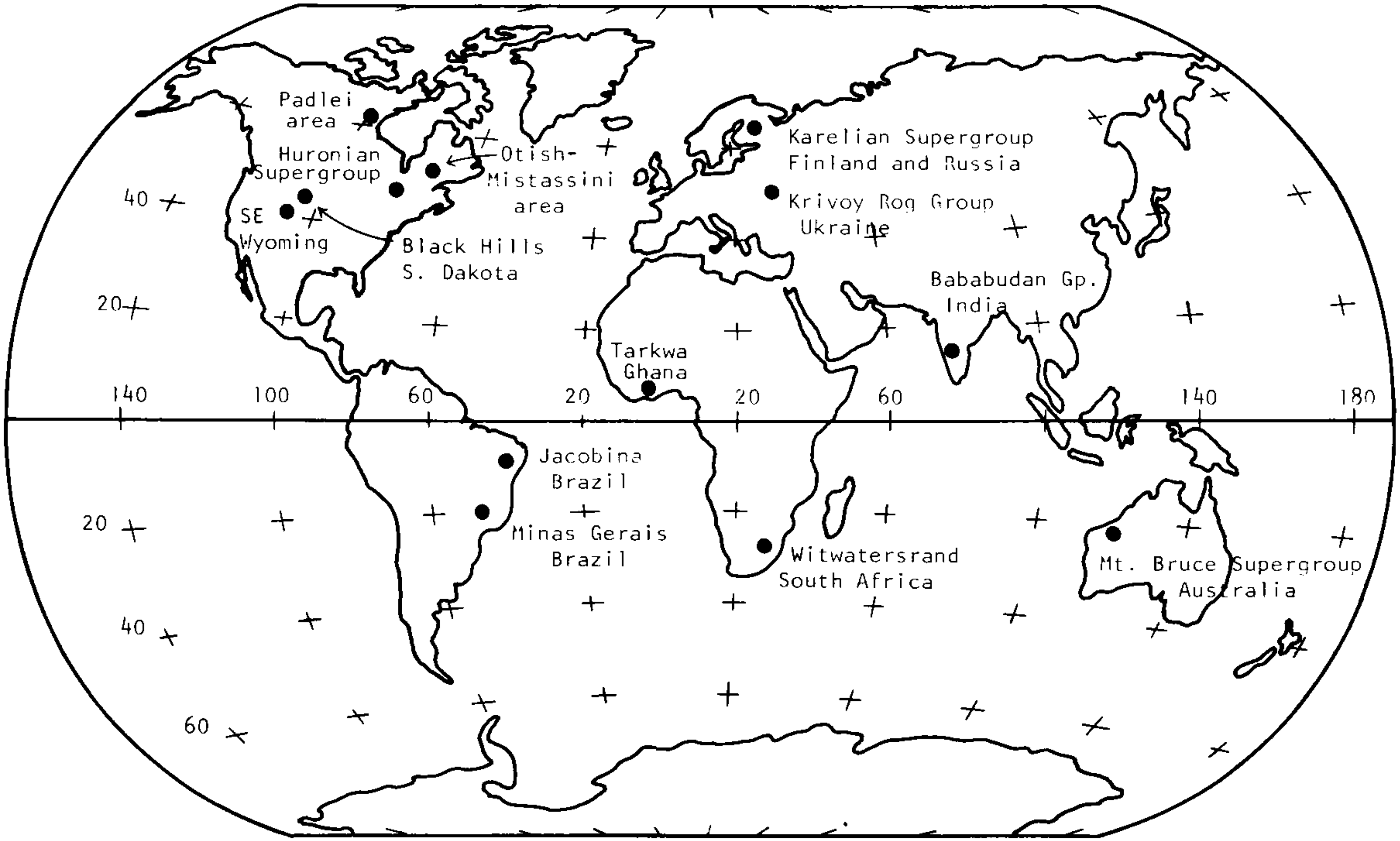


Figure 1.1. World-wide distribution of Early Proterozoic-type fossil placer uranium and gold occurrences.

PART I. EXPLORATION MODEL FOR PRECAMBRIAN URANIUM-BEARING FOSSIL PLACERS

PRECAMBRIAN GEOLOGIC SETTING

The Precambrian encompasses a time span from the beginning of the earth's history, more than 4500 m.y. ago, to 570 m.y. -- the great bulk of the geologic history of the earth. Obviously the most ancient rocks have complex geologic histories and, in many areas, the earlier history is obscured or obliterated by more recent geologic events. The complexity of Precambrian rocks, compared to generally less deformed Phanerozoic rocks, plus the fact that for five decades or more much of the economic impetus for geologic studies has come from the oil and gas industry which is primarily interested in Phanerozoic rocks, has meant that far less is known about the first 4000 m.y. or more of the earth's history than the last 600 m.y.

A beginning synthesis of Precambrian geologic history was made by Rankama (1963, 1965, 1967, 1970) who edited a four volume set of review articles by various authors designed to bring geologists up-to-date on Precambrian geology. As Rankama pointed out, the key to the understanding of Precambrian geologic history is the development of isotope dating methods that establish a time framework for rocks without guide fossils. The review articles in Rankama's volumes are a good first-look at Precambrian geology, but substantial progress in the understanding of Precambrian geology has been made in the last decade and the following recent symposia and review papers are especially commendable: McCall (1977); Windley (1976); Windley (1977); and Sutton and Windley (1973).

TERMINOLOGY

One of the major problems in Precambrian terminology involves the subdivision of Precambrian time. In this report, we will follow the recommendations of the I.U.G.S. working group on the Precambrian (1978), and James (1978) by using the following subdivisions:

Late Proterozoic	900-570 m.y.
Middle Proterozoic	1600-900 m.y.
Early Proterozoic	2500-1600 m.y.
Late Archean	2900-2500 m.y.
Middle Archean	3300-2900 m.y.
Early Archean	> 3300 m.y.

However, important breaks in the geologic record, which have often been used in the literature to subdivide the Precambrian, do not always conform to this scheme. For example, the Pongola and Witwatersrand Sequences of South Africa are Archean in age according to the subdivision given above, yet they are similar in character to Early Proterozoic sedimentary successions found elsewhere in the world. We will refer to these as "Early Proterozoic-type" successions and, to avoid confusion, absolute dates are included whenever such information is available -- e.g. Pongola Sequence (3100-2900 m.y.).

Our use of the terms shield, craton, greenstone belt, and mobile belt conforms to the definitions given by Anhaeusser and others (1969). A shield is a continental or sub-continental area of exposed crystalline rocks and a craton is a stable Archean nucleus within a shield area. In addition, we use the term platform as the next larger continental element; that part of a continent which is underlain by Precambrian crystalline rocks. The hierarchy, from largest to smallest, is thus platform-shield-craton.

ARCHEAN CRUST

This section considers only those aspects of Archean geology that we believe have a bearing on the origin of uranium-bearing quartz-pebble conglomerates. Of first concern is the nature of the source of the uranium now found in fossil placers. Which Archean rocks are the most likely source of the uranium minerals and what period of Archean crustal evolution was most favorable for the development of these sources?

One approach to this question is to examine the distribution of the major radioactive elements -- K, U, and Th, in the earth through geophysical studies of heat generation. Heat flow studies supply two important pieces of information. First, the radioactive elements appear to have been strongly concentrated in the crust and upper mantle very early in the earth's history, probably by fractional melting of the mantle during large scale differentiation of the earth prior to 4000 m.y. (Birch, 1965; Hanks and Anderson, 1969). Second, rates of heat generation due to decay of K and U were on the order of two times higher than present at 2600 m.y. and three times higher at 3600 m.y. (Lambert, 1976). This indicates that Early Archean crustal rocks were appreciably richer in uranium and geothermal gradients were steeper then than now (Lambert, 1976). Both of these suggest that low-grade sources of uranium may have existed very early in the Archean.

A more useful approach in understanding Archean source areas is through petrologic and geochemical studies. Many workers have shown that uranium is preferentially distributed in felsic rocks and is nearly absent in mafic and ultramafic rocks. As a consequence, the appearance of uranium-bearing quartz-pebble conglomerates on earth must have post-dated the development of extensive areas of uranium-bearing Archean felsic rocks and it is necessary to ask when such felsic rocks first appeared on earth and which felsic rocks are richest in uranium.

Unfortunately there is no consensus among Precambrian geologists on either the time of development or composition of the earliest crustal rocks. Archean rocks can be subdivided into three major types: granulite terrains which consist of metamorphic and igneous rocks of highest metamorphic rank and are dominated by felsic gneisses and migmatites; gneissic terrains of intermediate metamorphic rank that may or may not be a retrograde or reconstituted granulite terrains; and greenstone belts that are low rank metamorphic rocks dominated by volcanic rock suites. A critical debate among Precambrian geologists is the relative age of the gneissic terrains versus that of the greenstone belts. One school of thought is that, at least, some greenstone belts represent primordial crust and that high-grade gneiss terrains are younger than these greenstone belts. Proponents of this idea include: Anhaeusser (1973), Anhaeusser and others (1969), Glikson (1976), and Glikson and Lambert (1976). Another school of thought is that the gneissic terrains are older and formed the basement for the greenstone belts. This idea is supported by Windley and Bridgwater (1971), McGregor (1973), Hunter (1974), Moorbath (1975), and Windley (1977). A student of Phanerozoic rocks might suggest that a careful study of contact relationships between the two rock sequences should solve the problem. Unfortunately most contacts between greenstone belt rocks and gneissic "basement" are invaded by younger igneous rocks or are so deformed and metamorphosed that it is difficult to establish age relationships. Even age determinations of rocks in the two terrains does not solve the problem because it is difficult to date the mafic and ultramafic volcanic rocks of the greenstone belts and periods of multi-deformation and metamorphism create ambiquity in establishing the age of the gneissic terrains.

We do not consider this problem to be solved, but we believe most of the world's greenstone belts were formed after a felsic gneissic basement had been developed because: the most ancient rocks dated are 3700-3900 m.y. old rocks in the gneissic terrains (Moorbath, 1975; Goldich and Hedge, 1974); granitic clasts are common in conglomerates in greenstone belts (although they are often in the younger rocks of the greenstone succession); and finally, where careful reviews of field evidence are made, the evidence generally favors an older gneissic terrain (Bliss and Stidolph, 1969; Hunter, 1970, 1974; McGregor, 1973, Collerson and others, 1976). Nevertheless, we agree with Moorbath (1975) who cautions that: "there is no clear evidence that any particular greenstone-granite association represents primordial crust, or that stratigraphical relationships between the two rock units should always be the same way round".

It seems safe to assume that felsic rocks existed very early in the earth's history which could have constituted a basement for supracrustal rocks. These granitic rocks may or may not be primordial crust but that point is not critical to this discussion. The important factor for the uranium geologist is whether or not these felsic gneisses were rich enough in uranium-bearing minerals to supply uranium to overlying supracrustal sedimentary rocks.

The chemical composition of gneissic terrains has been reviewed by Tarney (1976) and Heier (1973). Tarney (1976, p. 405-415) pointed out that early basement gneisses (>3000 m.y.), both granulite facies and amphibolite facies, are characteristically depleted in K, Rb, Th, and uranium. There is no agreement on the cause of this depletion of K, Rb, Th, and U (Tarney, 1976, p. 413-415), but some type of flushing mechanism, during deformation and metamorphism especially for granulite facies rocks seems most probable.

The fact that uranium is relatively depleted in Archean felsic rocks older than 3000 m.y. is significant for prospecting because it implies that these rocks are an unpromising source of uranium minerals for subsequent deposition in younger sedimentary rocks and also that they are a poor place to prospect for uranium, in general. This geochemical argument seems to be supported by geologic data because geologists note (Smith, 1974) that few if any vein deposits of uranium have been recognized in rocks of Archean age. Admittedly, it remains possible that even a very low-grade source of uranium such as this could have been the source of placer deposits if the concentration process was extremely efficient. However, a better solution to the source area problem is to examine other Archean felsic rocks. Are there Archean granites which are richer in uranium than the Archean gneissic terrains?

Geochemical data indicate that some Late Archean granites are enriched in uranium (Roscoe, 1969; Stuckless, 1978; Rye and Roy, 1978). For example, concentrations up to 100 ppm Th and 30 ppm U were reported by Roscoe and Steacy (1958) for granites of the southern Superior Province. These Late Archean granites are probably a more suitable source for uranium minerals than the pre-2900 m.y. Archean gneissic terrains. If these late Archean granites were derived, even partially, by melting of older gneissic rocks, radioactive elements lost during basement mobilization (i.e. U, K, Th, Rb) might be added to the melts which formed at near granite minimum temperatures and hence, might be concentrated in Late Archean intrusives. There are arguments for (Tarney, 1976; Rye and Roy, 1978) and against (Moorbath and Pankhurst, 1976) this hypothesis, with the final resolution of the argument probably dependent on the interpretation of strontium isotope studies which often show low initial Sr^{87}/Sr^{86} ratios for Late Archean granites indicating that the granites do not represent rejuvenated crustal material. Nevertheless, empirical evidence indicates that late Archean K-rich granites should be considered as more favorable source rocks for uranium-bearing fossil placers than Archean high-grade gneisses.

ARCHEAN METASEDIMENTARY ROCKS

Most uranium-bearing quartz-pebble conglomerates are in Early Proterozoic-type successions which range in age from about 2700 to 2000 m.y. To evaluate whether the search for such conglomerates should be confined only to rocks of this age, it is necessary to discuss the physical characteristics of the Precambrian sedimentary rocks and the evolution and composition of the hydrosphere and atmosphere.

This section discusses two end-member types of Archean metasedimentary successions; low-grade metasediments which are part of greenstone belts and highly metamorphosed and deformed metasedimentary rocks that occur as inliers in granulite and amphibolite facies gneissic terrains. A third type consists of Late Archean sequences containing mature clastic sediments. However, these are similar to Early Proterozoic-type metasedimentary platform successions and will be discussed in later sections.

According to Anhaeusser (1973), a typical greenstone belt is made up of a lower ultramafic-mafic group, a middle calc-alkaline volcanic group, and an upper sedimentary group. The lower ultramafic group is composed of mafic and ultramafic volcanic rocks that may retain pillow structures; some layered ultramafic complexes; and very minor quartzitic cherts, schists, and pelites. These rocks appear to be entirely of marine origin and the volcanic rocks are characterized by a high CaO/Al_2O_3 ratio, high MgO, and low K_2O. Rocks of the lower ultramafic group do not appear to be a promising source of uranium and, to our knowledge, no uranium deposits are known in this group. However, these rocks often contain gold deposits and can be an important source of fossil-placer gold deposits (Pretorius, 1976a).

The middle calc-alkaline volcanic group consists of cyclic successions of volcanic rocks that are composed chiefly of mafic volcanic rocks but also contain felsic assemblages in the upper part of each cycle. The presence of pillow structures in the mafic volcanics and the presence of chemical sedimentary rocks such as chert and iron-formation suggests a marine depositional environment. The calc-alkaline group rocks are similar to the ultramafic group in that they do not appear to be a promising source of uranium but may be a source of gold.

The upper sedimentary group is primarily graywacke, shale, and impure sandstone in the lower part and conglomerate, quartzite, limestone, and iron formation in the upper part. These sedimentary rocks show a general transition from deep to shallow water sedimentation. The sedimentary rocks deposited in shallow water include both paraconglomerates and polymictic conglomerates with clasts of a great variety of rock types. The source area for these clasts was certainly granitic in part because clasts of granite, gneissic granite, felsic gneiss, and vein quartz are common in the conglomerates and the quartzites contain abundant feldspar. We know of no fluvial uranium-bearing conglomerates from this shallow water succession but we are unwilling to rule out the possibility that the upper part of greenstone successions might contain uranium- or gold-bearing conglomerates.

The distinction between the shallow-water sediments of the upper part of greenstone successions and Early Proterozoic-type platform successions is far from clear-cut. Supracrustal successions seem to have diachronously evolved through time and there appears to have been a gradual change from the older greenstone successions, dominated by volcanic rocks, to platform successions dominated by quartzites. During this change, in the interval 3000-2500 m.y., successions containing nearly equal volumes of volcanics and quartzites were deposited and it is not always clear whether to call them Archean greenstone successions or Early Proterozoic-type platform successions. These successions may include fluvial and shallow water quartzites that contain detrital uranium and gold minerals. For example, the Pongola sequence of South Africa (3100-2900 m.y.) contains about 55% volcanics and 45% sediments (Anhaeusser, 1973, figure 4) and contains detrital gold in quartzites. This sequence is often referred to as an Early Proterozoic sequence in spite of its age.

The Bababudan Group of India (2600-2100 m.y.) could also be one of these transitional metasedimentary sequences. It contains abundant mafic volcanics as well as quartzites and uranium-bearing conglomerates and is referred to by most Indian geologists as a greenstone sequence (Radhakrishna and Vasudev, 1977; Ramakrishnan and others, 1976). However, descriptions of the lithologies of this unit suggest to us that it is probably an Early Proterozoic-type succession. Indeed, Radhakrishna (1975) describes this "greenstone sequence" as geosynclinal in character and Proterozoic in age and it seems odd that he insists on the term "greenstone sequence". At any rate, the main point of this discussion is that sedimentary successions referred to as greenstone belts in the literature should not be arbitrarily ignored by the uranium explorationist without a careful examination of the rocks themselves.

The second type of Archean metasedimentary rocks that may be of interest to the uranium geologist is the metasedimentary rocks that occur within granulite and amphibolite facies gneiss terrains. These metasedimentary rocks include mica schists -- some of which are graphitic; marbles; quartzite which is often fuchsitic; and iron formation. Amphibolite and hornblende gneiss are also common in the gneissic terrains but are probably mainly of volcanic origin. These metasediments are more intensely deformed than rocks of the greenstone belts and are usually preserved as small synformal bodies within the gneissic terrain. They may represent very ancient sedimentary rocks preserved only in this "basement" terrain, remnants of disrupted and granitized greenstone belts, or even platform-type sediments which have been highly deformed and granitized. Too little is known about these metasedimentary rocks to appraise their uranium potential, but it seems probable that deep burial, severe deformation and accompanying metamorphism and metasomatism promoted a loss of uranium, especially in the granulite gneiss terrains. On the other hand, if some Archean gneissic terrains contain remnants of platform-type sediments, they could contain uranium. Thus, the younger Archean gneissic terrains should also not be ignored by uranium geologists.

EARLY ATMOSPHERE AND HYDROSPHERE

The earth's atmosphere and hydrosphere are considered by most workers (including Rubey, 1951, 1955; Rutten, 1964; Holland, 1962; Berkner and Marshall, 1967) to be of secondary origin. Rubey (1951, 1955) and Brown (1952), among others, have pointed out that certain volatile materials (H_2O, CO_2, Cl, N, and S) are too abundant in today's atmosphere and hydrosphere to be explained as products of rock weathering. Also, the noble gases are not abundant enough if our planet had an original atmosphere and hydrosphere with elemental abundances comparable to that of the universe. This suggests that the escape velocities of the constituents that might have constituted an early hydrosphere and atmosphere were too high to allow these elements to be retained on the earth's surface during formation of the planet. The planet therefore did not have an atmosphere and hydrosphere in earliest Archean, and the atmosphere and hydrosphere probably evolved through time.

This concept requires that the hydrosphere and atmosphere evolved from some juvenile source and Rubey (1951, 1955) suggested that this source may have been volcanic gases that escaped during the crystallization of magmas.

These gases may have been added to the atmosphere and hydrosphere in one catastrophic event in the Early Archean (Walker, 1976) or, more probably, in many minor periods of Archean volcanism (Fisher, 1976). If these volcanic emanations were similar to modern volcanic emanations, they contained mainly H_2O and CO_2 with a variety of less abundant gases but essentially no free oxygen (Schidlowski, 1976). Oxygen was probably contributed to the atmosphere in small quantities by dissociation of water and in large, quantities by organic photosynthesis, once life began to evolve.

Holland (1962) took Rubey's concept of a volcanic source of volatiles one step farther by considering the composition of gases that might be in equilibrium with magmas generated at various periods of the planet's development. Holland (1962) proposed a three-stage model for the evolution of the earth's atmosphere. Holland's earliest atmosphere developed prior to core formation at a time when native iron was present in the outer planet. Holland did not believe this stage was of long duration (∿0.5 b.y.), but gas in equilibrium with magmas generated at this time would be strongly reducing and have as major constituents CH_4 and H_2 with lesser amounts of H_2O, N_2, H_2S, NH_3, and Ar. This atmosphere is of little interest to the uranium geologists because of its short duration and the probability of very limited hydrosphere in equilibrium with it. Holland's second stage atmosphere (∿3 b.y. to ∿2 b.y.) which was in equilibrium with Hawaiian-type magma, was less reduced and was made up mainly of nitrogen, water, and carbon dioxide with Ar, Ne, He, and CH_4 in modest amounts ($>10^{-6}$ - $<10^{-4}$ atm). Holland also considers that NH_3, SO_2, and H_2S may have been present in trace amounts, and that O_2 was present as a nonequilibrium trace. Holland's calculation of oxygen pressure, however, is based on the stability of uraninite in surface waters, and may be subject to revision. Holland's third stage atmosphere (2 b.y.-present) became increasingly like our present atmosphere.

The second stage of Holland's model of atmospheric evolution is of great interest to the uranium geologist because the scarcity of free oxygen in the Archean and Early Proterozoic atmosphere and hydrosphere may be one key to understanding the formation of uranium-bearing fossil placers (Roscoe, 1969; Pretorius, 1976a). Uraninite and pyrite are unstable minerals in oxidizing environments but appear to have been mechanically weathered and transported as detrital minerals in the low-oxygen atmosphere and hydrosphere which existed in the Archean and Early Proterozoic. The main lines of evidence supporting an anoxygenic atmosphere and hydrosphere in the Archean and Early Proterozoic are given below.

(1) Geochemistry of iron. There are several lines of evidence which indicate that Fe^{+2} was transported in solution in large quantities in early Precambrian waters, that hematite was unstable in the atmosphere, and that pyrite was stable. Collins (1925) and later workers have identified a regolith between the Huronian Supergroup and underlying Archean granites. Roscoe (1969) reported that this regolith was depleted in total iron and that the Fe^{+2}/Fe^{+3} ratio was increased relative to nearby unweathered Archean granites. This is opposite to modern soils where iron is enriched because of its low solubility in oxidized groundwaters and it implies a low-oxygen atmosphere in the early Precambrian. Second, the transport of abundant Fe^{+2} in solution is also indicated by the Early Proterozoic banded iron formations. Most models for genesis of these deposits require a low-oxygen atmosphere and hydrosphere. Third, Archean and Early Proterozoic clastic sediments are mainly a drab green color and red-beds are significantly absent before about 2300 m.y. This suggests that hematite was not stable during deposition and diagenesis of the sediments and that the atmosphere was low in oxygen. The presence of detrital pyrite in Early Proterozoic sediments is additional evidence for low oxygen because this mineral is only

stable under low Eh conditions (at normal pH). See Figure 1.2 for the stability fields of various iron minerals under different Eh and pH conditions.

(2) Geochemistry of uranium. In the basal Huronian regolith, uranium was not depleted during weathering (Roscoe, 1969). This suggests that uraninite was a resistate mineral in the early Precambrian soils. Also, the presence of detrital uraninite in fossil placer deposits indicates that uraninite was stable in the hydrosphere during fluvial transport. This is opposite to modern environments where uraninite is highly soluble in the hydrosphere and it suggests low-oxygen in the early Precambrian environment.

(3) Organic evolution models. If one accepts the hypothesis that the atmosphere is secondary and originated by volcanic emanations, it is necessary to consider the source of atmospheric oxygen. Dissociation of water is one possibility (Berkner and Marshall, 1965) but Schidlowski (1971) suggested that it was far less important than organic photosynthesis as a source of O_2. Photosynthesis may have begun as much as 3700 m.y. ago (Junge and others, 1975) but most workers believe that life, as blue-green algae, and appreciable quantities of free oxygen did not become widespread until the Early Proterozoic (Cloud, 1968, 1972; Schopf, 1975).

From the viewpont of the explorationist, the important question is the time when the atmosphere became sufficiently oxidizing to oxidize uranium during weathering of source rocks and prevent transportation of uraninite and pyrite as detrital minerals in streams and rivers. This appears to have occurred about 2300-2000 m.y. ago. The main evidence is the change in heavy mineral assemblages found in Early Proterozoic clastic rocks. Roscoe (1973) suggested that Early Proterozoic successions in Canada, South Africa, Brazil, and Australia record a dramatic change in the oxygen content of the atmosphere. In each case, the basal units contain pyrite as the main iron-bearing phase and uraninite in the heavy mineral

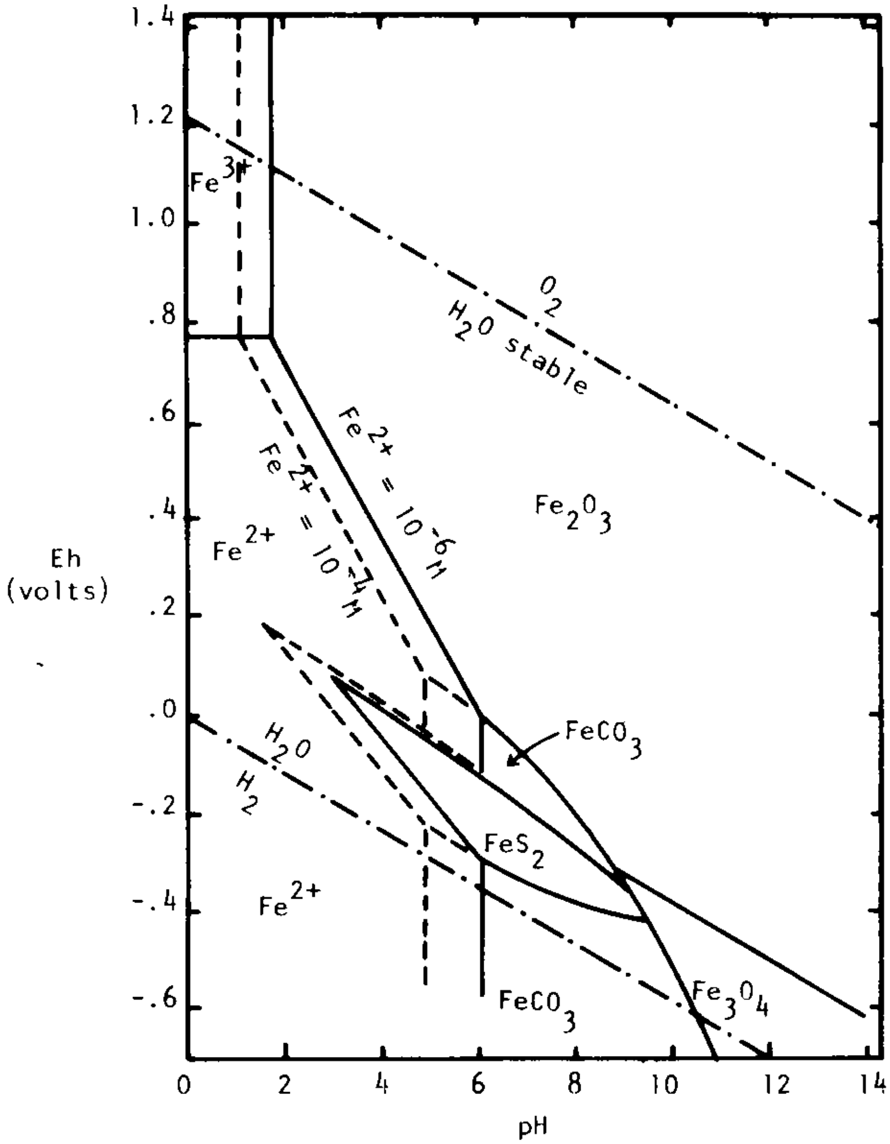


Figure 1.2. Eh-pH diagram showing stability fields of common iron minerals. Total activity of dissolved carbonate, 1M, of dissolved sulfur, 10^{-6}M. Solid field boundaries on left side of diagram are for total dissolved iron = 10^{-6}M, dashed lines for 10^{-4}M. Taken from Krauskopf (1967, p. 252). Note that the stable iron phase under low Eh conditions may be Fe^{+2}, pyrite, siderite, or magnetite, depending on pH and on concentrations of sulfur and carbonate but that hematite is not stable. If sulfur is abundant, the pyrite field expands to cover most of the lower part of the diagram. From: Introduction to Geochemistry by K. B. Krauskopf. Copyright (c) 1967) by McGraw-Hill Inc. Used with permission of McGraw-Hill Book Company.

suites -- and there are no red beds. In contrast, the upper units contain iron oxides instead of pyrite and thorium phases such as monazite instead of uraninite -- and red beds began to appear. It must be admitted that existing geochronologic data on some of these successions is still imprecise. However, where good bracketing dates exist, such as in Canada and South Africa, the transition from uranium and pyrite bearing fossil placers to oxide and monazite bearing placers took place about 2300-2200 m.y. ago. Roscoe (1973) suggested that this was a worldwide event related to the appearance of appreciable free oxygen in the atmosphere about this time.

Other evidence which supports the increase in free oxygen in the atmosphere about 2300-2000 m.y. is the appearance and disappearance of the largest Proterozoic banded iron formations during the interval 2200-1800 m.y. (see page 24) and the appearance of more widespread biogenic carbonates after about 2000 m.y. It seems clear that the Early Proterozoic was a time of rapid changes in the atmosphere and biosphere and that atmospheric conditions most favorable for deposition of uranium-bearing conglomerates occurred prior to about 2300-2000 m.y. ago.

The evolution of the hydrosphere is also of interest to the uranium geologist. It is quite clear from the sedimentary record that a hydrosphere existed at an early stage in the planet's evolution. Evidence includes the presence of rounded granite boulder conglomerates and marbles in the 3750 m.y. old Isua supracrustals of West Greenland (McGregor, 1973); the presence of pillow structures in 3400 m.y. volcanics of the Onverwacht Group of South Africa (Viljoen and Viljoen, 1969); and the presence of stromatolites in the 3000 m.y. old limestones of Rhodesia (Schopf and others, 1971). The key question, however, is when did the hydrologic cycle become well enough developed and land masses stable enough to support large fluvial systems and large sedimentary basins. As Rubey (1951) pointed out the volume of the ocean may have grown through time, a concept that fits well with that of a gradual growth of land masses and deepening of ocean basins until a point was attained during the Late Archean and Early Proterozoic when continents became stable and major transgression and regressions might be possible. Evidence from the sedimentary record suggests that this point was reached in South Africa prior to 3000 m.y. ago and elsewhere on earth by 2500 m.y. This is discussed more fully in the next sections.

ARCHEAN-PROTEROZOIC TRANSITION

Geologists working in Phanerozoic rocks are familiar with the problems that have existed in setting the exact date for the beginning of the Phanerozoic Eòn, even in highly fossiliferous rocks of Cambrian age. Settina the date for the beginning of the Proterozoic Eon some 2500 m.y. ago, in rocks without guide fossils, it is even more difficult.

The Archean-Proterozoic boundary represents the most profound change in tectonic style in the earth's history. The Archean was characterized by high mobility of crustal material, steep geothermal gradients, and deposition of greenstone belts. This period ended about 2500 m.y. ago and the end is marked by three occurrences: 1) concentration of high heat flow along linear mobile belts adjacent and surrounding the (newly) inactive cratonic blocks, 2) extensional fracturing of the rigid crust and emplacement of basic dike swarms, and 3) deposition of mature quartz-rich clastic platform sediments on the stable margins of the cratonic blocks and in intracratonic basins. These occurrences are generally considered to represent a change in tectonic processes which resulted from an increase in the thickness and rigidity of the sialic cratons (Anhaeusser and others, 1969; Windley, 1973; Sutton, 1973; Burke and Dewey, 1973). Viewed as such, the Archean-Proterozoic boundary is not a sharp time line but instead is a transitional phase in an evolving planet.

Early Proterozoic mobile belts, dike swarms, and platform sediments are identified on all the major continents but they appeared diachronously. For example, the earliest platform sediments in South Africa, the Pongola Sequence, were deposited between 3100 and 2900 m.y. ago whereas the earliest platform sediments in the Canadian shield, the Huronian Supergroup, were not deposited until after 2500 m.y. ago. Thus in South Africa the Archean-Proterozoic boundary is often placed at around 3000 m.y. whereas in Canada it is placed at 2500 m.y. Also, Early Proterozoic-type events overlapped temporally with Archean-type events. For example, initiation of the Proterozoic-type Limpopo mobile belt of South Africa began prior to 3000 m.y. (Mason, 1973), at about the same time that Archean greenstones of the upper Swaziland Sequence were being deposited elsewhere on the same craton. Similarly, platform sediments of the Pongola Sequence were deposited at about the same time that Archean granites were diapirically intruding the Swaziland Sequence (Anhaeusser, 1973).

Because of the transitional nature of the Archean-Proterozoic boundary, Precambrian geologists have not yet accepted a specific date for the end of the Archean although 2500-2600 m.y. is seen most often in the literature. So long as the uranium geologist recognizes that this problem exists, he can search for uranium in Proterozoic-type metasedimentary rocks keeping in mind that it is the environment of deposition not the precise age that is critical to the development of fossil placers.

EARLY PROTEROZOIC-TYPE SEDIMENTARY ROCKS

Proterozoic sedimentary successions include thick piles of clean, well-sorted quartzites, conglomerates, stromatolitic limestones and shales that are quite similar to miogeosynclinal rocks of the Phanerozoic Eon. The major change from the Archean style of sedimentation is that Early Proterozoic type successions are not dominated by volcanic rocks as they were in the Archean and that sedimentation took place in environments of relative tectonic quiescence. This striking change in the geologic record is recognized worldwide and took place between 3000 and 2500 m.y. ago.

The beginning of deposition of Proterozoic-type sedimentary successions is commonly said to be the result of a less active tectonic regime on earth which allowed the development of extensive areas of stable continental crust. The existence of these stable landmasses, in turn, allowed the development of a mature erosional cycle which supplied detritus to major river systems which flowed into large intracratonic basins and continental shelf environments. This clastic material consisted of mechanically stable minerals such as quartz, chert, and various heavy minerals, including uraninite, pyrite, and gold which were all stable in the low-oxygen hydrosphere and atmosphere of the Early Proterozoic.

One of the most fascinating aspects of Early Proterozoic-type sedimentation is its cyclical nature. For example, Pretorius (1966, 1976a) described the sedimentary successions of South Africa in terms of tectonically controlled cycles of various frequency. Increased tectonic unrest produced periods of volcanism, erosion, or deposition of coarse-clastics whereas tectonic quiescence produced non-clastics. Similarly, Roscoe (1969, 1973) and Frarey and Roscoe (1970) viewed sedimentation of the Huronian Supergroup in terms of major cycles which, in this case, appear to have been controlled by advance and retreat of continental glaciers in response to climatic changes. Sedimentary cycles have also been proposed for Early Proterozoic sediments in the Medicine Bow Mountains of Wyoming (Karlstrom, 1977). Aside from implications of crustal evolution, cyclical sedimentation in the Early Proterozoic was important in terms of concentrating and preserving heavy minerals in placers during periods of marine regression and transgression. Regression in fluvial environments causes continued reworking of sediments and concentration of heavy minerals (Pretorius, 1976a) and transgression was effective in preserving previously deposited sediments.

Early Proterozoic-type sedimentary successions are the most favorable hosts for fossil-placer uranium deposits because of the interaction of a variety of environmental factors which acted to create favorable sedimentary conditions. These sedimentary successions developed at a time when continental landmasses were newly stabilized. This created, for the first time on earth, a situation where extensive river systems existed which could transport and concentrate resistate heavy minerals as placers. Cyclical tectonic and/or climatic changes caused episodic marine regressions and transgressions which were effective in preserving the heavy minerals in placers.

EVOLUTION OF PRECAMBRIAN SEDIMENTARY ROCKS

Uranium-bearing conglomerates are mainly restricted to sedimentary successions which are between 2700 and 2000 m.y. old. To understand why this is true, it is necessary to examine the changes in the earth's sedimentary record through time and to examine the factors that have controlled these changes. Such factors include: changes in tectonic processes, evolution of the hydrosphere, atmosphere, and biosphere, changes in the types of source areas for sediments, and sedimentary recycling (Windley, 1977, p. 304). It is critical that the uranium explorationist understand the appearance and disappearance of uraniferous conglomerates in the geologic record in terms of the larger trends in the evolution of Precambrian sedimentation. As an illustration, Figure 1.3 shows that red beds and carbonates became well developed on earth sometime after uraniferous conglomerates. Thus, sequences which contain thick red bed or carbonate sections can be regarded as relatively poor targets for uraniferous conglomerate and the explorationist can assume that stratigraphically lower rocks are likely to be more favorable.

As discussed earlier, the change from the Archean to the Proterozoic Eon is marked by a change in sedimentary style represented by a decrease in deposition of volcanogenic rocks and an increase in deposition of clastic sediments such as arkoses, quartz-arenites, and quartz-pebble conglomerates (Figure 1.3). Another major sedimentary change was in the deposition of iron formation. Iron formation is an important Precambrian rock-type which is present in every Precambrian shield area and may account for about 15 percent of the total thickness of sedimentary successions (Ronov, 1964). Archean iron formations are typically thin lenses intercalated with volcanic and volcaniclastic rocks and the iron minerals are usually oxides. In contrast, Early Proterozoic iron formations tend to be much thicker and they are associated with stromatolitic dolomite, quartzite, slate, and chert. Iron minerals are distributed into oxide, silicate, carbonate, and sulphide assemblages which are interpreted to represent different facies of sedimentation (James, 1966). This change in the character of iron formations can best be understood in terms of the changing tectonic regime which allowed large stable land masses to develop.

Another important change between the Archean and the Early Proterozoic is in the character of paraconglomerates (also called tilloids or diamictites) which are open framework, matrix-dominated conglomerates. These rocks occur in Archean successions and are quite common in the sedimentary unit of the greenstone belt where they are usually interpreted as mudflows or proximal turbitites. Paraconglomerates that appear in the Early Proterozoic,

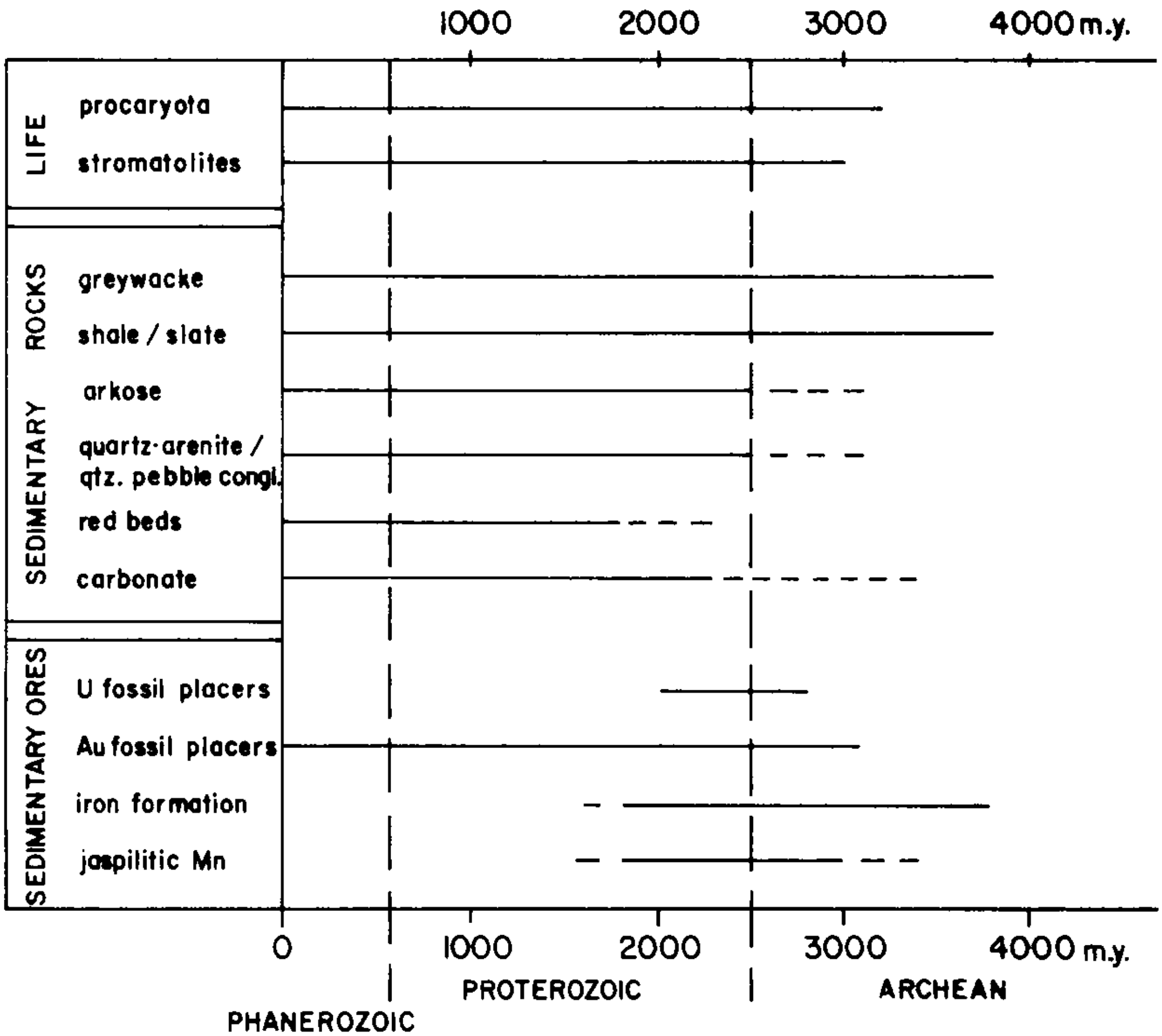


Figure 1.3. Summary of the temporal distribution of sedimentary rocks and sedimentary ore deposits in geologic history. Adapted from Veizer (1976, p. 25). Reproduced with permission from Elsevier Scientific Publishing Company and the author.

however, are more widespread and can perhaps be correlated over vast areas (Young, 1970). These paraconglomerates have associated laminated phyllites with dropstones, large clasts with possible striations, and even local striated pavements at their base. They have been interpreted as tillites (Lindsey, 1966; Cassyhap, 1969, Young, 1973) and may represent the earliest examples of continental glaciation. Certainly these deposits are further evidence of the development of large stable landmasses by the Early Proterozoic. They may also represent a time of change in the temperature regime of the planet sufficient to allow glaciation, or a time when the hydrosphere was large enough volumetrically so that major ice sheets could form. Various suggestions have been made by geologists (Roscoe, 1969; Wiebols, 1955) to relate the formation of these tillites to the origin of uranium-bearing quartz-pebble conglomerate. The writers cannot make a direct connection between these two Early Proterozoic lithologies, except that they are often found together, especially in North America. This topic will be discussed in greater detail in a later section.

In addition to changes in sedimentary style between Archean and Early Proterozoic times, there were important changes within the Proterozoic Eon. The most intriguing of these changes is the sudden disappearance of banded iron formation about 1800 m.y. ago and the appearance of widespread red beds at about the same time (Figure 1.3). In recent years, this curious disappearance of iron formation has been explained most commonly in terms of the oxygen fugacity of the Precambrian atmosphere. Oxygen is believed to be a volumetrically unimportant constituent of the early (Archean) atmosphere (Schidlowski, 1976), but is believed to have evolved through time as plants developed that could produce oxygen through photosynthesis. Iron can readily dissolve in surface waters deficient in oxygen and should therefore have been an important constituent of the Archean hydrosphere. In the Archean, this iron was chemically precipitated along with silica to form the typical laminated beds of alternating silica-rich and iron rich layers. The characteristic association with volcanic rocks in the Archean suggests that during volcanism both iron and silica may have become locally saturated in the sea water and chemically precipitated to form iron formation. In the Early Proterozoic, iron formations are not typically associated with volcanic rocks and a different solution must be found to explain their origin. As oxygen producing plants evolved, the oxygen fugacity undoubtedly slowly increased in the Precambrian atmosphere and hydrosphere. Although there is evidence of life in Early Archean rocks as stated above, stromatolites became really abundant for the first time in the Early Proterozoic. With the increase in these oxygen-producing plants, a critical point was probably attained where iron was less soluble in surface waters and was more easily extracted from surface waters by various mechanisms. Under such circumstances a general period of widespread deposition of iron in lakes, bays, rift zones, or shelves bordering the open ocean might be expected. Numerous mechanisms have been proposed to account for deposition of iron and silica in various environments, (Trendall, 1973, Eugster, 1969, Lepp and Goldich, 1964; Govett, 1966; James, 1966 and 1969, Drever, 1974), but there are two main points that are of interest in terms of trends in Precambrian sedimentation. First, most of the above models suggest that oxygen fugacity played a role in the transportation and precipitation of iron. Second, most Early Proterozoic iron formation is of vast extent and individual beds can be correlated for hundreds of miles. This is additional evidence that large continental platforms of some type must have existed at this time.

Therefore, evolution of iron formation deposits provide independent evidence of changing oxygen fugacity through time and of the development of large stable landmasses. Iron formations also provide a clue as to a time of the development of an oxidizing atmosphere similar to that of today. As stated above, true iron formation is uncommon in rocks younger than 1800 m.y. In addition, true red beds began to appear in Proterozoic sedimentary successions at about the same time that iron formation disappeared (Cloud, 1968; Veizer, 1976). There are exceptions and overlaps (Dimroth, 1975) in iron formation and red beds, but the disposition through time of these two rock types along with the increasing number of stromatolitic limestones is most simply explained by a gradual increase in the oxygen fugacity of the atmosphere.

So far, we have presented several lines of evidence that the Early Proterozoic was a time of major changes in tectonic regimes, sedimentary styles, and atmosphere and hydrosphere composition. The way that these changes affected the types of fossil-placer deposits which formed is summarized in Figure 1.3 which shows that uranium-bearing placers are restricted to the time interval between 2800 and 2000 m.y. but that gold-bearing placers are found as early as 3100 m.y. and are still found today. The older limit of about 3100 m.y. for gold placers can best be explained in terms of the evolving hydrosphere and changes in sedimentary style. A source of gold was available as early as 3400 m.y. in greenstone belts of South Africa but the deposition of gold-placers had to await the development of a sedimentary environment capable of depositing relatively mature quartzites and conglomerates. The fact that the first uranium placers were not deposited until somewhat later (2700 m.y.) probably reflects the fact that the main source of uranium was Late Archean K-rich granites (2900-2500 m.y.). Nevertheless, there is no reason that uranium placers might not have developed earlier if suitable sources and depositional environments could be found. The younger limit of about 2000 m.y. for uranium placers is best explained in terms of the evolution of the atmosphere. By about 2000 m.y. the amount of free oxygen appears to have been sufficient to make detrital uraninite and pyrite unstable in the existing hydrosphere. In contrast, detrital gold is stable in oxidizing waters so that gold placers persist in modern environments.

SUMMARY AND IMPLICATIONS FOR EXPLORATION

In the discussion of Precambrian geology certain facts have been developed that may assist the explorationist. They are listed below:

1. There is evidence of a granitic crust as early as 3800-3900 m.y., but geochemical studies suggest that this early crust is uranium poor and is a poor source for uranium minerals. In fact, very few uranium deposits have been found in rocks older than 2800 m.y.

2. Archean gneissic terrains contain abundant younger granites (3000-2500 m.y.) that differ chemically from earlier rocks. Some of these granites contain significant uranium, with concentrations of up to 10-20 ppm. They may be the source of uranium in younger quartz-pebble conglomerate and therefore their location may be significant in the search for uranium.

3. No significant uranium deposits have been found in greenstone belt successions, but we cannot rule out the possibility that the sedimentary rocks of the greenstone belts, especially of the younger greenstone belts (2800-2500 m.y.), may contain uranium-bearing quartz-pebble conglomerates. Furthermore, some Late Archean (2900-2500 m.y.) supracrustal successions are very similar in overall characteristics to Early Proterozoic-type sedimentary basins that typically contain uranium-bearing quartz pebble conglomerates.

The explorationist should therefore look carefully at sedimentary-volcanic successions that are referred to as greenstone belts -- they may be a different sequence of rocks entirely.

4. The Early Proterozoic represents a striking change from the Archean greenstone belt deposition to a type of sedimentation more like the Phanerozoic. This change did not take place at the same time everywhere on the planet, and the beginning of Early Proterozoic-type sedimentation ranges from 3100 m.y. in some areas to 2500 m.y. in others. Thus the age of the rocks cannot be used exclusively to identify favorable metasedimentary successions.

5. By 2500 m.y. ago, large stable landmasses had developed everywhere on the planet. As they evolved, perhaps beginning at 3000 m.y., major drainage basins could form and extensive, relatively clean detrital sediments could be deposited in fluvial systems. These stable landmasses allowed transgression and regression over large areas and ultimate burial and preservation of thick fluvial successions.

6. There is evidence in Early Proterozoic sedimentary rocks, independent of the distribution through time of uranium-bearing conglomerate, to substantiate a change in the composition of the hydrosphere and atmosphere to a point at approximately 2000-1800 m.y. where there was adequate oxygen in the atmosphere to oxidize uranium minerals in source areas, during transportation or after deposition and thus prevent the formation of uranium-bearing fluvial placers.

GENETIC CONCEPTS

HISTORY OF EXPLORATION

The discovery that led to the development of the gold- and uranium-bearing quartz-pebble conglomerate of the Witwatersrand of South Africa was made by George Walker and George Harrison in 1886 (Whiteside and others, 1976). This discovery ultimately led to the development of mines over most of the northern Witwatersrand basin. The mines were developed for their gold values, but uraninite was determined to be an additional constituent of the conglomerate (Cooper, 1924) and now the conglomerates are mined for both their gold and uranium values. These conglomerates contain the world's largest reserves of gold and major resources of uranium. One of the most atonishing facts concerning this deposit is that geologists were extremely slow in utilizing the information gained concerning the characteristics of the Witwatersrand deposits to discover, or at least, explore for additional reserves of the same type. There are probably two factors that played a role in this: one was the economic conditions which controlled the price of gold and uranium and the other was the long standing debate over the origin of the deposits.

Gold mining was stimulated by South Africa's decision to go off the gold standard in 1932 and more recently by greatly increased market prices for gold. After the second World War the price of uranium was artificially pegged at high prices to stimulate prospecting and recently, free market uranium prices have increased more than four-fold. Interestingly, it appears that the post World War II uranium market did more to stimulate prospecting for uranium and gold in quartz-pebble conglomerates than any other factor. For example, uranium-bearing quartz-pebble conglomerates were discovered in the Blind River area of Ontario, Canada in 1948 (Robertson and Card, 1972) and gold-and uranium-bearing conglomerates were discovered in the Serra de

Jacobina in the state of Bahia, Brazil in 1953 (Bowie, 1970, p. 27). The discovery at Blind River in Canada led to additional finds at Elliot Lake; at Agnew Lake, in the area between Sudbury and Saulte St. Marie, Ontario; in the Montgomery Lake-Padlei area in the Northwest Territories of Canada; and in the Sakami Lake area in Quebec. In addition, more recent finds have been reported in Australia (Richards, 1972), Brazil (D.S. Robertson, 1974), India (Viswanatha and others, 1977), South Dakota (Hills, 1977) and Wyoming (Houston and others, 1977). All of these deposits are in Early Proterozoic metasedimentary rocks and are very similar in their overall geologic setting to the deposits discovered in South Africa in 1886.

HISTORY OF GENETIC CONCEPTS

Witwatersrand Sequence, South Africa

Uranium and gold-bearing quartz-pebble conglomerates in South Africa have been the subject of one of the most remarkable debates in the history of economic geology. The controversy began more than 90 years ago and has not been entirely put to rest even today. As stated above, this controversy prevented the acceptance of a genetic model for the deposits and was probably a major contributor to delayed exploration for additional deposits of the same type.

The controversy has had two facets. The first involved a debate on the origin of the uranium and gold minerals in the conglomerates and the second involved a debate on the depositional environment(s) of the conglomerates.

From the beginning (1886), most South African geologists subscribed to the syngenetic, placer model for the origin of the gold. However, they were opposed by a vociferous group of geologists from outside South Africa who insisted that the gold was derived from a magmatic source and transported into the Witwatersrand basin by epigenetic hydrothermal solutions. This debate over a syngenetic versus epigenetic origin for the gold mirrored the debate between the neptunists school of economic geologists, who believed that many metals in ore deposits came from non-magmatic sources, and the magmatists, who maintained a rather rigid concept that virtually all metals were derived from magma.

A fascinating aspect of the debate on the origin of the gold is that the magmatists (Maclaren, 1908; Horwood, 1917; Graton, 1930; Davidson, 1953, 1957, 1964; Fisher, 1939; Hatch and Corstorphine, 1904; H. Louis, J.H. Hammond, and R. Beck, in Lindgren, 1933, p. 244) were largely from academic and research institutions located outside of South Africa and they had very limited experience in the Witwatersrand mines. In contrast, most of the placerists (Ballot, 1888; De Launay, 1896; Young, 1911, 1917; Mellor, 1916; Rogers, 1922; Du Toit, 1926; Nel, 1927; Sharpe, 1949; Antrobus, 1956; Viljoen, 1963; Brock and Pretorius, 1964a; Pretorius, 1966, 1974; Minter, 1972, 1976) were South African geologists and mine workers. This situation led Davidson (1953), who was the principal proponent of the magmatic theory, to state that the placer theory was "a national article of faith" in South Africa. However, it seems more likely that this adherence to the placer model, by South Africans, came from practical experience which showed that the placer model could be successfully applied to exploration and development of the deposits (Reinecke, 1927).

The evidence presented by the magmatists was based on mineralogical, textural, and structural characteristics of the ore deposits which are not typical of modern fluvial placers. This evidence is summarized below:

1) in many samples, the gold is not rounded but, instead, occurs as irregular masses which are sometimes well crystallized,
2) the pyrite in the matrix of the conglomerates is often crystalline or anhedral,

3) there is evidence of replacement of quartz pebbles by pyrite,

4) veinlets of gold cut quartz pebbles, and gold is also found in veinlets cutting pyrite grains,

5) some conglomerates contain abundant sericite and chlorite which were said to be of hydrothermal origin,

6) sulfur isotope studies show that values of S^{34}/S^{32} are very close to that of meteoritic sulfur and there is little deviation from this value--this suggests the sulfur had a magmatic source.

There are two additional major points made by magmatists that require discussion: one is the significance of uraninite and magnetite and the other is the role played by the organic material referred to as thucholite. Uraninite is a rare constituent of modern placers, but an important one in the Witwatersrand, whereas magnetite is a typical constituent of modern placers and is not present in the Witwatersrand. Magmatists, most notably Davidson (1953, 1955, 1957, 1961, 1962, 1964), have maintained that uraninite is not a constituent of modern placers because it cannot survive in an oxidizing atmosphere and thus there is no reason to expect it in fossil placers. In addition, the virtual absence of magnetite in the Witwatersrand, and for that matter in most other Proterozoic quartz-pebble conglomerates, is remarkable considering that magnetite is one of the most common constituents of modern placers and its presence has even been used as an indicator of the possible presence of gold. Thus, the presence of uraninite and the absence of magnetite in the Witwatersrand conglomerates was cited as evidence that the deposits are not normal gold placers.

The organic material thucholite is clearly associated with both gold and uraninite in the Witwatersrand. It is really a hydrocarbon containing variable amounts of uranium and thorium with a variety of elements in minor amounts such as Pb, Mg, Zr, K, Ce, La, Yt, Er, Ti, Fe, Al, V, Ca, Na, P, and

Si. The name thucholite was given by Ellsworth (1928) to an organic material from the elements Th, U, C, H, and O present in the substance. Subsequently this term has been used for any organic material containing uranium and/or thorium and the name has clearly been applied to organic material of quite different origin, composition, and structure.

Thucholite in the Witwatersrand Sequence occurs as small granules about 2 mm or less in diameter, in seams up to 15 mm thick that parallel bedding planes in quartzite, and as veins that transect primary structure. It contains inclusions of uraninite, pyrite, and gold. Magmatists maintained that the thucholite was formed from hydrothermal emanations, probably gaseous hydrocarbons, and they cited pegmatite and vein thucholite as analogues (Davidson and Bowie, 1951; Liebenberg, 1955). They suggested that there was a definite paragenesis with deposition of pyrite, carbon, pyrite, and finally gold (Fisher, 1939).

The second aspect of the controversy concerned the depositional environment(s) of the conglomerates. This aspect of the Witwatersrand controversy was summarized by Pretorius (1975), who divided the history of work on the Witwatersrand into several phases. The early history of the Witwatersrand (up to 1928) was characterized by a wide variety of speculations on the depositional environment of the conglomerates. Most geologists of this period favored deposition in a neritic, littoral or shoreline delta environment and conglomerates were usually considered to be beach gravels. After 1928, as a result of increased use of quantitative sedimentological techniques, geologists began to give preference to various fluvial models and to combinations of fluvial and shallow marine environments. At present (1979), the most widely accepted model combines braided fluvial and lacustrine depositional environments. This model is summarized by Pretorius (1976a, 1976b) and will be discussed in a later section.

Blind River - Elliot Lake Area, Canada

The discovery of uranium-bearing conglomerates of the Huronian Supergroup in 1948 at Blind River, Ontario came at a time when the magmatists were strongly supporting an epigenetic, hydrothermal origin for the gold and uranium of the Witwatersrand Sequence. As a consequence, many early workers considered the Blind River uranium deposits to also be epigenetic and hydrothermal in origin (Bateman, 1955; Joubin and James, 1957; Davidson, 1957; Heinrich, 1958; Patchett, 1960). As in South Africa, however, the majority of the field geologists working in the Blind River - Elliot Lake area favored a placer origin for the uranium and a fluvial origin for the conglomerates (Abraham, 1953; McDowell, 1957, 1963; Pienaar, 1958, 1963, Roscoe, 1957, 1959, 1969; D. S. Robertson, 1962a, 1962b, J. A. Robertson, 1960, 1968, 1969, 1976).

A variation of the placerist view was expounded by Holmes (1957), among others, who suggested that the ores were deposited as placers but were later modified by diagenetic, metamorphic, and deformational events which created some new secondary minerals and textures. This has been called the "modified placer hpothesis" and it is now (1979) the favored hypothesis for formation of the Blind River - Elliot Lake deposits (Robertson, 1976). From the explorationists viewpoint, this modified placer hypothesis is not much different from the placer hypothesis--the distribution of heavy minerals is still controlled by primary sedimentary features. Nevertheless, it is important to be aware of the nature of the post-depositional changes, particularly in exploring for fossil placers in amphibolite and higher facies terrains These changes are discussed in a later section.

Jacobina, Brazil

Gold- and uranium-bearing conglomerates were discovered at Jacobina in 1953, again during the of the magmatists. These deposits are more intensely metamorphosed than the Blind River deposits and they contain abundant evidence of gold and pitchblende in fractures and veins cross-cutting the conglomerates and along the contacts with ultramafic intrusives (Gross, 1968). This evidence was cited by White (1956; 1957; 1961; 1964), Bateman (1958), and Ridge (1972) to support a hydrothermal origin of the gold and uranium. However, Cox (1967) and Gross (1968) presented sedimentological evidence that the gold was syngenetic and they favored the modified placer hypothesis. Gross (1968) said that gold was found in the matrix of nearly all the conglomerates and the highest concentrations were along foreset beds. He suggested that remobilization of gold took place during folding, metamorphism, and intrusion of mafic dikes.

CURRENT VIEWS ON GENESIS

Evidence for Placer Origin of Uranium and Gold

In the last several decades a wealth of information has become available on uranium and gold-bearing quartz-pebble conglomerates not only in South Africa, but also in other localities. Most of this information supports the placerist viewpoint. In South Africa notable progress has been made in understanding sedimentology, mineralogy, and the nature of the organic material, and a major part of the published work is by D.A. Pretorius of the University of the Witwatersrand and his students. In Canada, papers by S.M. Roscoe and J.A. Robertson, among others, have strongly supported the placer viewpoint. Evidence which supports the placer origin of the uranium and gold is summarized below.

1. Sedimentological studies have established that the distribution of uranium, gold, pyrite, and other heavy minerals in the metasediments is strongly controlled by primary sedimentary features. Heavy minerals are commonly concentrated in the matrix of conglomerates (Liebenberg, 1973) and tend

to be best concentrated near the base of individual conglomerate layers. In addition, heavy minerals are found in economic concentrations along the foresets of cross-bedded sandstones that fill erosion channels and in sandstone, shale, and carbonaceous seams which directly overlie local unconformities (Pretorius, 1976a). In Canada, heavy minerals are concentrated in conglomerates which were deposited in topographic depressions that developed on the Archean erosion surface (Roscoe, 1969). Another convincing piece of evidence which favors a sedimentological control of heavies is that paystreaks in the Witwatersrand conglomerates are elongate in the directions of the paleocurrents and form fan-shaped patterns reminiscent of braided river channels (Reinecke, 1927; Minter, 1976).

2. Textural studies in both Canada and South Africa show that many of the heavy minerals occur as rounded grains; even the sulphides--pyrite, cobaltite, linnaeite, and arseno-pyrite. The pyrite, which constitutes 90% of the sulphide fraction (Saager and Esslaar, 1969), is typically rounded and is referred to as buckshot pyrite. Uraninite is also commonly rounded and even rounded grains of gold have been found (this mineral is most typically irregular in form and occurs in minute particles 0.005 to 0.5 mm in width). These rounded grains are viewed as detrital and stream transported.

3. Where the original detrital form of heavy minerals can be determined, it has been shown (Koen, 1961; Viljoen, 1963; Simpson and Bowles, 1977) that these minerals are generally in hydraulic equivalence with associated light minerals in the matrix of the conglomerates and with each other. This suggests that the heavy minerals were transported in suspension, probably in a fluvial system. As a corollary to this argument, Minter (1972) showed that there is a relationship between the specific gravity of heavy minerals and their proximity to the source area. Uraninite, for example, is lighter than gold and is generally more abundant farther down the paleoslope whereas gold

tends to be concentrated closer to the source. Also, Viljoen (1963) showed that there is a systematic decrease in grain-size of heavy minerals radially away from the points where the sediment entered the basin. All of these facts suggest a detrital origin for the heavy minerals. Admittedly, from the sedimentologist's viewpoint, there are major problems with the concept of hydraulic equivalence which is based on particle settling velocities (Rittenhouse, 1943). Nevertheless, it is difficult to imagine that hydrothermal processes would form a suite of heavy minerals which are in hydraulic equivalence and it seems reasonable to invoke sedimentary processes to explain this relationship.

4. Recent detailed studies of the organic material thucholite (Hallbauer, 1974, 1975) show that this material is some form of primitive plant, perhaps a primitive conidia producing fungus or some form of blue-green algae. Its principal mode of occurrence is in mats that range in thickness from one to twelve millimeters, but it is also common as granules in conglomerate layers (Hallbauer, 1975). Placerists interpret this carbonaceous material to represent algal colonies which flourished near the mouths of the rivers and which collected gold and uranium by physically filtering the river water and, possibly, by chemically precipitating gold and uranium from solution.

5. S^{34}/S^{32} values are near meteoritic values and very consistent. Placerists interpret this to indicate an ultimate magmatic source for pyrite sulphur which was then deposited in an igneous environment of the Archean source area. This pyrite was subsequently eroded, transported, and deposited in a reducing environment so that no oxidation-reduction reactions occurred that might have resulted in sulphur isotope fractionation. Furthermore, the pyrite was largely protected from alteration after it was buried.

6. Unquestionably one of the most significant points made by supporters of the theory of placer origin for the uranium in the Witwatersrand is from

age determinations. The basement for the Witwatersrand metasedimentary rocks has been dated by determining the age of Archean granites. These granites yield ages ranging from 3100 to 2750 m.y. (Hales, 1960; Allsopp, 1964). In addition, direct dates of volcanic rocks in the lowermost Witwatersrand Sequence yield ages of 2820 ± 55 m.y. by Rb/Sr and 2800 ± 50 m.y. by U/Pb (Van Niekerk and Burger, 1964). These data indicate that the Witwatersrand Sequence is younger than about 2800-2750 m.y. A minimum age for the Witwatersrand Sequence is given by a 2300 ± 100 m.y. date on the Ventersdorp Sequence, which stratigraphically overlies the Witwatersrand Sequence (Van Niekerk and Burger, 1964) so that uraniferous conglomerates are bracketed between 2750 and 2300 m.y. old. However, uraninite grains from these conglomerates have been dated as 3050 ± 50 m.y. (Rundle and Snelling, 1977) and monazite as 3160 ± 100 m.y. (Nicolaysen and others, 1962). Inasmuch as these grains are older than the deposits in which they occur, it is clear that they cannot be of hydrothermal, epigenetic origin. Thus, the placer theory is strongly supported by geochronological data.

7. Thorium is a typical constituent of most Precambrian fossil placers. This association is in strong contrast to uranium deposits that are clearly of chemical origin; in fact, thorium is not even mentioned as a constituent of epigenetic uranium deposits of the United States in Finch's 1967 detailed summary of their characteristics. Thorium has an extremely low solubility in modern natural waters which are generally alkaline. It is typically concentrated in resistate minerals or is absorbed by clay minerals in weathered rocks of today. We cannot be certain that thorium followed this same pattern during the Early Proterozoic because thorium is more soluble in acid solutions. But, if Roscoe and Steacy (1958) are correct, thorium was weathered during the Early Proterozoic much as it is today. Therefore, it seems unlikely that thorium was transported in solution during the Early Proterozoic

and thus the uranothorite and uranothorianite of these deposits is probably detrital, and the presence of significant amounts of thorium in these deposits can be taken as additional evidence of detrital origin for most of the constituents.

Evidence for Fluvial Origin of Conglomerates

Sedimentological studies, in addition to establishing the distribution of heavy minerals, have defined the environments of deposition of the host rocks. Most geologists now agree that the conglomerates, which host much of the uranium and gold mineralization, are fluvial in origin. The most convincing evidence comes from the Witwatersrand Sequence, which has been the most thoroughly studied of known deposits. However, there is also convincing evidence from the Canadian rocks. The evidence falls into several categories which are discussed below.

1. Distribution of conglomerates--In the Witwatersrand, conglomerates occur in several bi-lobate fanshaped bodies on the margin of the basin, nestled between granite domes (Steyn, 1963; Pretorius, 1976a). These have been interpreted as fluvial fan-deltas. In Canada, conglomerates are thickest in valleys and basins in the underlying Archean erosion surface (Roscoe, 1969). In both areas, the large-scale distribution of conglomerates appears to have been controlled by local topography. Within conglomeratic zones, in both areas, individual conglomerate layers are lenticular and elongated parallel to the paleoslope (Armstrong, 1968; Reinecke, 1927; Pienaar, 1963; Roscoe, 1969). This distribution suggests that the conglomerates represent gravels which accumulated in channels and bars of a braided river system (Pretorius, 1976b).

2. Paleocurrents--Paleocurrent distributions are unimodal and have a low variance, suggesting fluvial transport (Minter, 1976; Pienaar, 1963,

McDowell, 1957). In the Witwatersrand, paleocurrent directions radiate out from the apex of the fluvial fan-deltas, suggesting fluvial transport away from the point where the river emptied into the basin (Pretorius, 1975).

3. Down-paleoslope variations--In the Witwatersrand, pebbles in the conglomerates become finer-grained and less rounded down the paleoslope and the quartzite/conglomerate ratio increases (Viljoen, 1963). This change is accompanied by an increase in indicators of wave and tidal action such as oscillation ripples. These variations indicate a decrease in energy of transport down the paleoslope--a typical characteristic of fluvial systems.

4. Stratification sequences--Pretorius (1976b) describes fining-upwards sedimentary cycles in the Witwatersrand sediments. Each cycle began with deposition of gravels on a local unconformity, followed by deposition of sands, and sometimes silts and clays. The abrupt change across the unconformities, from the silts and clays at the top of one cycle to gravels at the base of the next, was cited by Mellor (1915) and Roberts and Kronsdorff (1938) as evidence for fluvial deposition.

5. Sedimentary structures--Abundant channels, small scours, and trough cross-bedding in the conglomerates have been cited as evidence for fluvial deposition (Winter, 1965; Pienaar, 1963). Also, fluvial-type pebble imbrication was reported by Armstrong (1968), Knowles (1966), and Pienaar (1963).

6. Lithology--The conglomerates are bimodal sediments which consist of a pebble fraction ($\sim$ 80 percent by weight) consisting of quartz and chert pebbles and a matrix fraction consisting of poorly sorted, mature to submature sand with relatively large amounts of clay or mica, and heavy minerals (Viljoen, 1963; Pretorius, 1976b; Pienaar, 1963; Roscoe, 1969). The conglomerates are interlayered with poorly sorted, submature subarkoses which are less mature than the matrix sands of the conglomerates (Minter, 1976; Roscoe, 1969). The poor sorting and submature nature of the sands suggests rapid fluvial deposition.

Sedimentary Model for the Witwatersrand Goldfields

Fluvial deposition of heavy minerals in conglomerates is only part of the story of the formation of uranium- and gold-bearing conglomerates. In addition to fluvial deposits, the Witwatersrand Sequence contains a variety of shallow water clastic sediments, mafic and felsic volcanic rocks, and iron formation. To understand the conglomerates in terms of the total stratigraphic succession, it is necessary to formulate models of the paleogeographic setting and sedimentary history of the entire sedimentary basin. This was done for the Witwatersrand basin in recent papers by D.A. Pretorius (1975, 1976a, 1976b), which are very briefly summarized here. A more detailed discussion of these papers is presented in a later section.

Pretorius (1976a, 1976b) viewed the Witwatersrand basin as a large, intermontane lake or inland sea bordered on the north and west by tectonically active highlands. Clastic material was carried into the basin by major river systems which flowed through deep canyons in the Archean source terrain, across a relatively short piedmont plain, and then into the basin. Large, fluvial fan-deltas developed at each of the six points where major rivers entered the basin. Uranium- and gold-bearing sands and gravels were deposited on these fan-deltas in braided channels and on bars. Finer-grained sediments were carried farther into the basin and were reworked and redeposited by clockwise longshore currents. Deeper in the basin, chemical sediments were occasionally deposited.

Concentration of most of the gold and uranium in the sediments was related to cyclic episodes of regression and transgression which, presumably, were controlled by tectonism. Regressions were initiated by uplift in the source area. This caused an increase in the supply of clastic material entering the basin and progradation of the fan-deltas. Progradation took place in pulses; first an openwork gravel was deposited, then sands were deposited which filtered into the openwork gravel. Gold and uranium were brought in with the sand phase. Continued reworking of previously deposited sediments during progradation allowed heavy minerals to be very efficiently concentrated. Many of the economically important reefs in the Witwatersrand Sequence formed by this process of continued sediment reworking during regressional episodes.

As tectonic activity waned, the supply of clastic material decreased, stream gradients became lower, and the strandline transgressed. Finer-grained material was deposited and, locally, algal colonies flourished in quiescent areas. These algal colonies collected very fine-grained gold and uranium by both mechanical filtration and biochemical precipitation. During periods of non-deposition, longshore currents caused winnowing of the light mineral fraction and concentration of the residual heavy minerals on the erosion surface. Many economic concentrations of gold and uranium occur directly above such erosion surfaces in carbonaceous mats, sands, or shales. Renewed tectonism and uplift of the source area, then, would start the cycle over again.

CURRENT PROBLEMS

In spite of the general consensus among geologists, that the Precambrian uranium- and gold-bearing conglomerates are fossil placers which were deposited by braided rivers in a low-oxygen atmosphere, there are important aspects of the deposits which are not explained by the current model. First, there is evidence that not all gold, uranium, and pyrite grains in the deposits are detrital in origin--but instead some have formed by precipitation from solution. This fact has given rise to various modified placer hypotheses. The usual modified placer hypothesis views the ore metals to have been deposited first as placers and then to have been redistributed and recrystalized during diagenesis and/or metamorphism (Stanton, 1972). More recent modified placer hypotheses suggest that gold and uranium were transported in solution as well as in detrital minerals during deposition of the conglomerates.

Second, there are still unanswered questions regarding the geochemistry of the Precambrian environment. How much oxygen was in the presumed low-oxygen atmosphere? What role did organic material play in transporting and concentrating ore minerals? And why are the placers essentially devoid of magnetite but rich in pyrite?

Third, the paleoclimatic and tectonic setting in which these rocks were deposited is not well enough understood. Why are they often stratigraphically associated with presumed glacial deposits and what was the paleoclimate and paleolatitude during deposition? Also, what is the significance of the fact that the world-wide distribution of Early Proterozoic conglomerates is very similar to the distribution of Proterozoic banded iron formations? These controversial aspects of Precambrian fossil-placer deposits are discussed in the following sections.

Modified Placer Hypothesis

There is good evidence in most Precambrian fossil-placer deposits for some movement of gold, uranium, and other constituents in solution. In the Witwatersrand, the association of gold and uranium with carbonaceous seams suggests biochemical precipitation of mineralized waters (Pretorius, 1976b; Hallbauer, 1975) and textural studies of ore minerals suggest that some of the gold, uranium minerals, and pyrite are not of detrital origin (Saager, 1970; Feather and Koen, 1975; Simpson and Bowles, 1977). Similarly, in Canada, mineralogical and textural studies of the primary uranium mineral--brannerite--suggest that it may have formed by alteration of titanium phases in the presence of dissolved uranium (Ramdohr, 1957; Ferris and Rudd, 1971). In Brazil, textures also indicate considerable mobilization of gold and uranium (Bateman, 1958; Gross, 1968). The evidence for mobility of gold and uranium is overwhelming but the timing of migration of these elements is difficult to establish. Migration may have occurred syngenetically, diagenetically, during metamorphism or all three. Various workers have suggested each of these possibilities. In this section, we discuss metamorphic and diagenetic movement of metals; syngenetic transport will be discussed in the next section.

There are several lines of evidence which demonstrate that uranium and gold in conglomerates can be remobilized during metamorphism. Even in the Witwatersrand Sequence, where metamorphic temperatures were only 200-250^{o}C (Schidlowski, 1968), there are textures which indicate metamorphic recrystallization. For example, Schidlowski (1968) reported intergrowths between gold particles, gold in the pressure shadows of quartz grains, and gold overgrowths on clearly authigenic pyrite nodules. He interpreted these as evidence for small scale transport (several mm) of gold as AuS^- during low-grade regional metamorphism and local thermal metamorphism associated with mafic dikes. Also, Pretorius (1964) reported migration of gold some tens of meters due to intrusion of mafic dikes. Geochronological studies of uraninite (Burger and others, 1962; Rundle and Snelling, 1977) demonstrate an original age of 3050 $\pm$ 50 m.y. for detrital uraninite, but suggest a major reworking at 2040 $\pm$ 40 m.y. which reset uranium ages and released radiogenic lead which was deposited in veins and fissures as galena. There is also geochronological evidence of uranium loss, but the timing is uncertain.

Obviously, metamorphic effects become more pronounced in higher grade deposits, of which the Jacobina Series in Brazil is a good example. These rocks were metamorphosed to amphibolite facies (Cox, 1967) and both gold and uranium have been remobilized. Gross (1968) reported gold and uraninite in fractures cutting quartz pebbles and pyrite grains and these metals were enriched in contact zones adjacent to mafic dikes. He suggested that the metals were remobilized during regional metamorphism and subsequently reconcentrated during dike intrusion, in areas where dikes cut conglomerates. Pyrite was also recrystallized, at times surrounding gold grains and at times filling fractures in the matrix of the conglomerates. This type of textural evidence was originally used by Bateman (1958) and White (1961) as evidence for an epigenetic origin of the mineralization but was interpreted by Cox (1967) and Gross (1968) as evidence for a modified-placer origin.

In general, there appear to be two types of metamorphic changes in fossil placers. The first is recrystallization and very minor redistribution (on the scale of mm or perhaps cm) of metals during regional metamorphism. The second is reconcentration of metals due to thermal metamorphism associated with mafic intrusives. The first is not of great concern to the explorationist because, even at amphibolite facies, the distribution of metals is still directly related to sedimentary features. The second, however, is of potential economic significance because this mechanism appears capable of creating economic concentrations of uranium and gold adjacent to intrusive contacts in conglomerates which would otherwise be of sub-economic interest. Jacobina is one example and the Paukkajanvaara Mine of Finland is another (see page 263).

Diagenetic remobilization of metals is difficult to distinguish on a textural basis from metamorphic remobilization (or from syngenetic movement for that matter), and the distinction is certainly not clear for the low-grade deposits of Canada and South Africa. Nevertheless, there are several mineralogical and textural changes which have been attributed to diagenetic processes--to the extreme position advocated by Davidson in one of his later papers (1965) that all uranium mineralization represents diagenetic precipitation of uranium from circulating ground waters. This idea is not very convincing in view of the abundant mineralogical evidence for detrital uraninite grains (Liebenberg, 1955; Roscoe, 1969; Feather and Koen, 1975; Simpson and Bowles, 1977) but diagenetic changes do warrant consideration for minor redistribution of uranium and for formation of brannerite--a uranium-titanium mineral commonly found in the Canadian and South African conglomerates.

Brannerite typically contains a delicate internal structure which would be unlikely to survive detrital deposition and so indicates epigenetic processes (Ramdohr, 1957; Roscoe, 1969; Ferris and Rudd, 1971). Ramdohr (1957) believed that brannerite is a metamorphic mineral formed by the "Pronto reaction":

$$UO_2 + 2\text{-}3TiO_2 \rightarrow UTi_{2-3}O_{6-8}$$

in which detrital grains of uraninite or ilmenite were converted to brannerite. Roscoe (1969) and Ferris and Rudd (1971), on the other hand, suggested that brannerite represented ilmenite which was altered diagenetically in the presence of uranium in solution. This view may be difficult to reconcile with geochemical evidence that uranium is not appreciably soluble in reducing waters--a point which is discussed in the next section. Nevertheless, diagenetic movement of uranium remains a possibility.

Syngenetic Transport of Gold and Uranium

Syngenetic transport of gold has been proposed by Pretorius (1975, 1976a, 1976b) to help explain the gold found as coatings and replacements of carbonaceous filaments in carbon reefs of the Witwatersrand Sequence and to explain the very fine-grained gold found in sediments of the Transvaal Sequence (2300-2100 m.y.) which overlies the Witwatersrand Sequence. The Transvaal Sequence differs from the Witwatersrand Sequence by containing a significantly higher proportion of fine clastic sediments and dolomites. These were considered by Pretorius (1976a) to be mixtures of deltaic muds and silts with shallow-water carbonates. Stromatolites are abundant in dolomites of the Transvaal Sequence suggesting a significant increase in algal activity. Pretorius (1976a, p. 25) stated that the gold in these fine-grained sedimentary rocks was probably transported in solution and precipitated either chemically or biologically. He considered the precipitants to have been carbonate material, colloidal particles of clay, and algae, either separately or in combination. He further suggested that chloride- and cyanide-complexes, which may have existed in the anoxygenic hydrosphere of the times, promoted the solubility of gold in the Precambrian hydrosphere. Obviously, if gold was transported in solution during Transvaal time it could also be transported during Witwatersrand time and many textures and structures that were interpreted as evidence for hydrothermal origin of gold might simply be the result of syngenetic chemical precipitation from surface waters.

The concept of syngenetic transport of uranium in solution has also had its supporters (Derry, 1960; Joubin, 1960; Simpson and Bowles, 1977). Simpson and Bowles (1977) have made a detailed study of ores from the Witwatersrand Sequence using a variety of mineralogical techniques, They propose that there are two types of uranium ore; one the result of transportation and deposition of clastic grains of uraninite and the other the result of transportation and deposition of uranium from solution. Uranium thought to be deposited from solution occurs in various ways: in secondary clay minerals forming marginal overgrowths on pyrite; in interstitial clay phases of pyrite concretions; disseminated through rounded granules that consist of quartz, kaolinite, and pyrophyllite; as a dendritic vein system containing uraninite, gold, pyrite, and galena; in rounded grains of thucholite; in thucholite seams as dendritic growths; in sutures that parallel the columnar structure of the thucholite; and in internal pore spaces of the thucholite. All of these textures are believed to demonstrate that uranium (and gold) were precipitated from solution and that precipitation was controlled largely by the presence of the organic material thucholite and by clay minerals.

An interesting point made by Simpson and Bowles is that there is a significant difference in thorium content between the detrital uraninite and the uraninite believed to be deposited from solution. The Witwatersrand detrital uraninite is relatively high in thorium and is similar to detrital uraninite in modern alluvial deposits of the Indus River in Kashmir, India and Hazro, Pakistan which is also high in thorium. According to Simpson and Bowles (1977, p. 312), a thorium uraninite containing one percent or more thorium oxide can survive in river detritus under oxidizing atmospheric conditions and, inasmuch as the detrital uraninite of the Witwatersrand is high thorium, these authors believe that it is not necessary to have had a low-oxygen atmosphere for transport and deposition of uranium. In fact, they suggest that uranyl ions in solution might require the existence of an oxidizing atmosphere.

This point concerning the Precambrian atmosphere is an interesting one because both gold and uranium are believed to have been syngenetically transported in solution in Precambrian waters. According to Pretorius, chloride- or cyanide-complexes may have been the medium for gold transportation but he suggests that this would require an anoxygenic atmosphere, whereas Simpson and Bowles suggest the reverse situation (an oxidizing atmosphere) as best for transportation of uranium. If we consider all of the geological and geochemical evidence presented earlier in this report, it is difficult to believe that free oxygen was present in the Precambrian atmosphere in significant amounts prior to about 2000 m.y. ago. As a consequence, we are forced to try to reconcile the geologic evidence of transport of both uranium and gold in solution with the geologic and geochemical evidence of an atmosphere and hydrosphere low in oxygen.

Gold cannot be transported in its ionic forms, Au^{+} or Au^{3+}, because these species are stronger oxidizing agents than water and, therefore, water would decompose before Au^{+} or Au^{3+} would form (Krauskopf, 1967; Ling Ong and

Swanson, 1969). Pretorius (1976a) suggested that gold was transported as chloride or cyanide complexes. However, as shown in Figure 1.4a, transportation of significant quantities of gold chlorites in solution requires strongly acid waters (pH < 3) plus the presence of a strong oxidizing agent such as MnO_2, Fe^{3+}, or Cu^{2+} (Krauskopf, 1967; Ling Ong and Swanson, 1969). These specialized conditions have been observed in some mine waters, volcanic waters, and hydrothermal waters (Ling Ong and Swanson, 1969) but it seems unlikely that such conditions could have existed in the rivers which fed the Witwatersrand basin. The possibility of the transport of gold as cyanide complexes also has difficulties, as Hallbauer and Utter (1976) suggested , in that the presence of significant amounts of HCN, $(CN)_2$, and NH_3 in the Precambrian atmosphere is unsubstantiated. Perhaps the most likely possibility is transport of gold as AuS^-. As shown in Figure 1.4a, this can take place in strongly reducing solutions containing $HS^- > 10^{-6}$ moles/liter and such conditions seem possible for the Witwatersrand basin.

An alternative interpretation was presented by Hallbauer and Utter (1976) who suggested that gold was transported as organic supported colloids during Witwatersrand and Transvaal deposition and that no gold was transported in solution. Hallbauer (1975) maintained that such an interpretation can explain the observed occurrence of fine-grained gold in carbonaceous mats and calcareous muds and he cited work by Ling Ong and Swanson (1969) which demonstrated that organic acids stabilize colloidal gold by giving them protective coatings which inhibit precipitation. According to Ling Ong and Swanson (1969), this mechanism for transport of gold requires concentrations of organic acids greater than 3 ppm and a pH between 4 and 9. These conditions also seem plausible for the Witwatersrand basin and it appears to us that the transport of gold as colloidal particles is at least as probable as the transport of gold as AuS^- in solution -- and both are capable of accounting

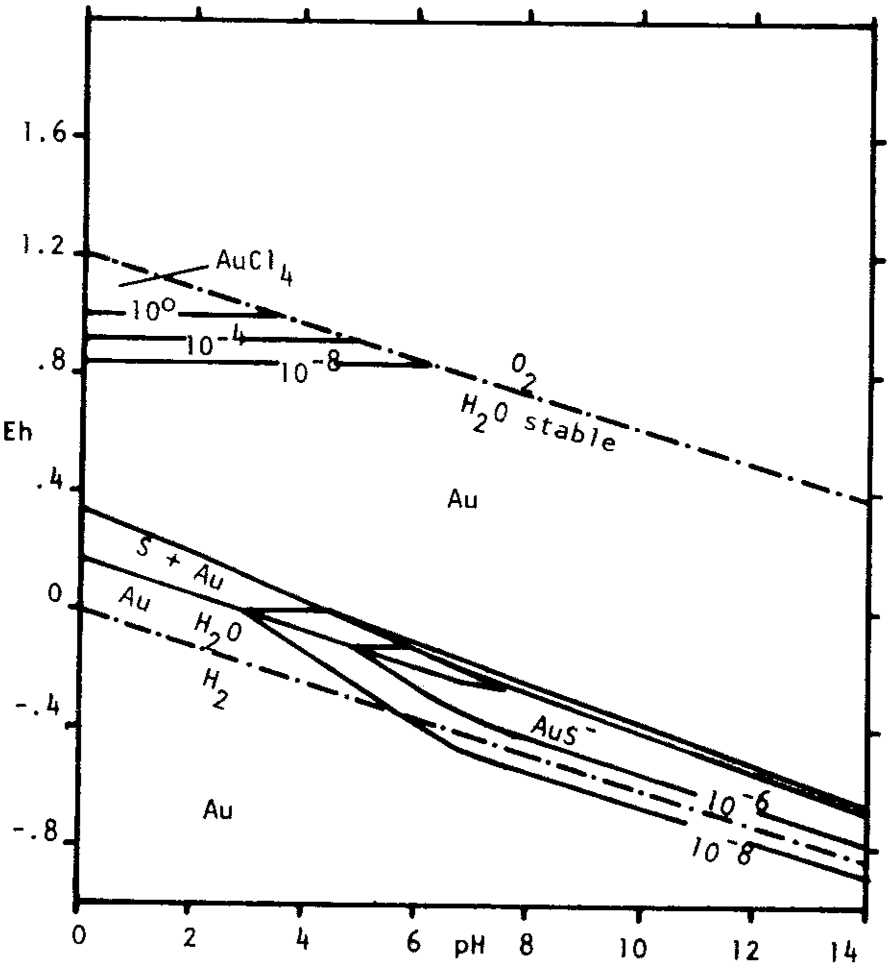


Figure 1.4a. Stability relations among some gold compounds in water at 1 atm. total pressure. Total dissolved chloride species = 1M; total dissolved sulphur species = 10^{-1}M. Shows stability fields at low and high chloride and sulfur contents. From Garrels and Christ (1965, p. 258): Solutions, Minerals and Equilibria. Reproduced with permission from Freeman, Cooper and Company.

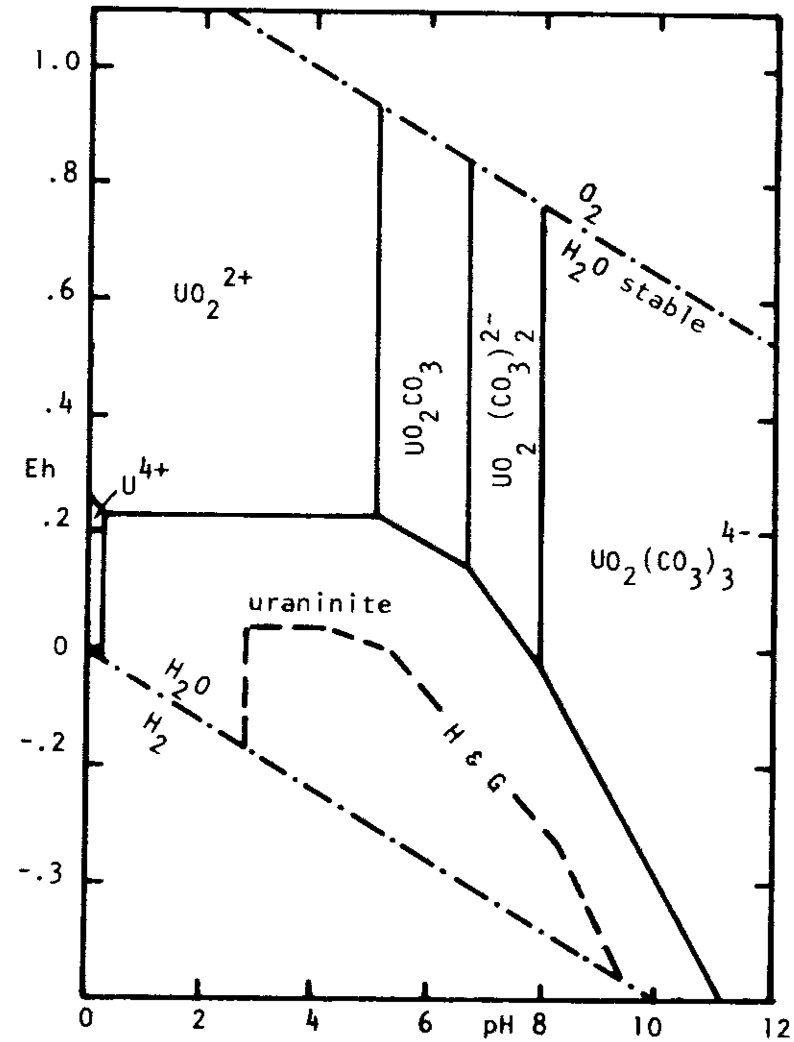


Figure 1.4b. Eh-pH diagram in the $U\text{-}O_2\text{-}CO_2\text{-}H_2O$ system at 25° C for $P_{CO_2} = 10^{-2}$ atm. Uraninite, UO_2(C), solution boundaries are drawn at 10^{-6}M (.24 ppm) dissolved uranium species. H & G denotes the boundary of the uraninite stability field according to Hostetler and Garrels (1962). Taken from Langmuir (1976, p. 560). Reproduced with permission from Pergamon Press Ltd. and the author.

for the geologic observations that gold may have moved in solutions as well as in detrital minerals during deposition of the Witwatersrand and Transvaal Sequences.

Transport of uranium in solution in a low-oxygen atmosphere and hydrosphere is at least as problematical as transport of gold. According to Langmuir (1978), uranium is soluble under anoxygenic conditions (in waters containing typical amounts of CO_3, Cl, F, PO_3, and SO_3), as uranous (U^{+4}) flouride complexes only if the pH is less than 4 or as uranous hydroxy complexes at higher pH (Figure 1.4b). However, transport of uranium as hydroxy complexes is severely limited by the fact that uraninite and coffinite are stable and highly insoluble in anoxygenic waters so there is no strong tendency for uranium to go into solution.

To answer this problem, Simpson and Bowles (1977) proposed that the atmosphere may have been oxidizing during deposition of the Witwatersrand Sequence--as early as 3000 m.y. ago. We are reluctant to agree with them because most other evidence is against this view. Nevertheless, if they are correct in believing that uranium was syngenetically transported in solution, it is either necessary to propose an oxidizing environment or to propose other geochemical explanations. For example, one might invoke a very low pH (less than 4) for the river waters. If this were the case, appreciable gold could be transported as chloride complexes and uranium as flouride complexes. Unfortunately, a low pH may be incompatible with the abundance of organic carbon in the Witwatersrand Sequence and the abundance of carbonate in the Transvaal Sequence. This problem of simultaneous transport of uranium and gold in solution and as detrital minerals needs further attention from the geochemists.

Magnetite - Pyrite

The problem of the abundance of pyrite and the relative absence of magnetite in fossil placers compared to modern placers was used extensively by hydrothermalists (Davidson, 1957; 1960; 1962) as evidence against a syngenetic origin of the mineralization. Explanations advanced by placerists, to this date, are only partially able to answer the problem.

The placerists argue that the pyrite is mainly detrital and was supplied from the Archean greenstone belts. This position was supported by Köppel and Saager (1974) who showed that the isotopic composition of Witwatersrand pyrite was similar to that of the pyrite in the Swaziland Sequence to the northeast. Placerists further maintain that the magnetite and ilmenite seen in the Archean source area were unstable in the Precambrian atmosphere and hydrosphere and were altered to pyrite during weathering, transport, deposition, and diagenesis.

The key to this argument is the mineralogic work of Ramdohr (1958), Roscoe (1969), Saager (1970), and Köppel and Saager (1974) which demonstrated that there are several varieties of pyrite in the conglomerates. In addition to abundant detrital pyrite there are concretionary grains which are rounded and porous. These may have been deposited detritally (Simpson and Bowles, 1977) or they may have formed diagenetically (Saager, 1970). According to Feather and Koen (1975) many of these grains may represent altered iron-oxides. A third type of pyrite is recrystallized authigenic pyrite which includes euhedral grains and overgrowths. These probably formed or were modified during metamorphism and dike intrusion (Ramdohr, 1958; Saager, 1970; Köppel and Saager, 1974).

The presence of clearly non-detrital pyrite led Ramdohr (1958) to suggest that the absence of appreciable magnetite and ilmenite in the placers is the result of pyritization--a process leading to the conversion of iron-titanium oxides to rutile plus new pyrite. This process was also advocated by Ferris and Rudd (1971) for the Canadian deposits. They proposed the following reactions:

$$FeTiO_2 + H_2S \rightarrow TiO_2 + FeS + H_2O$$

$$FeS + H_2S \rightarrow FeS_2 + H_2$$

The rutile that formed in the first reaction was then capable of taking up uranium from solution to form brannerite via the "Pronto reaction":

$$2\text{-}3TiO_2 + UO_2 \rightarrow UTi_{2-3}\,O_{6-8}$$

Ferris and Rudd (1971) believed that this type of reaction exchange could explain the high pyrite-low ilmenite content of the ores, the texture and composition of the brannerite with its complex interpenetrating textures of rutile and brannerite crystals and its variable uranium content, and the overgrowths of new pyrite on old pyrite and brannerite crystals.

We agree that this process of sulfidization of iron-titanium oxides is the most likely explanation for the presence of some of the authigenic pyrite and for the lack of iron-oxides. Indeed, Figure 1.2 shows that pyrite should be the stable iron phase in low oxygen waters (see also Krauskopf, 1967). However, there are several problems with the model. First, it is not clear whether sulfidization took place syngenetically or epigenetically and second, it is not clear where the sulfur came from.

The timing is difficult to establish on textural grounds. Ramdohr (1958) suggested that much of the sulfidization took place during deposition of the sediments whereas Ferris and Rudd (1971) and Saager (1970) suggested it took place diagenetically. Both possibilities seem plausible to us.

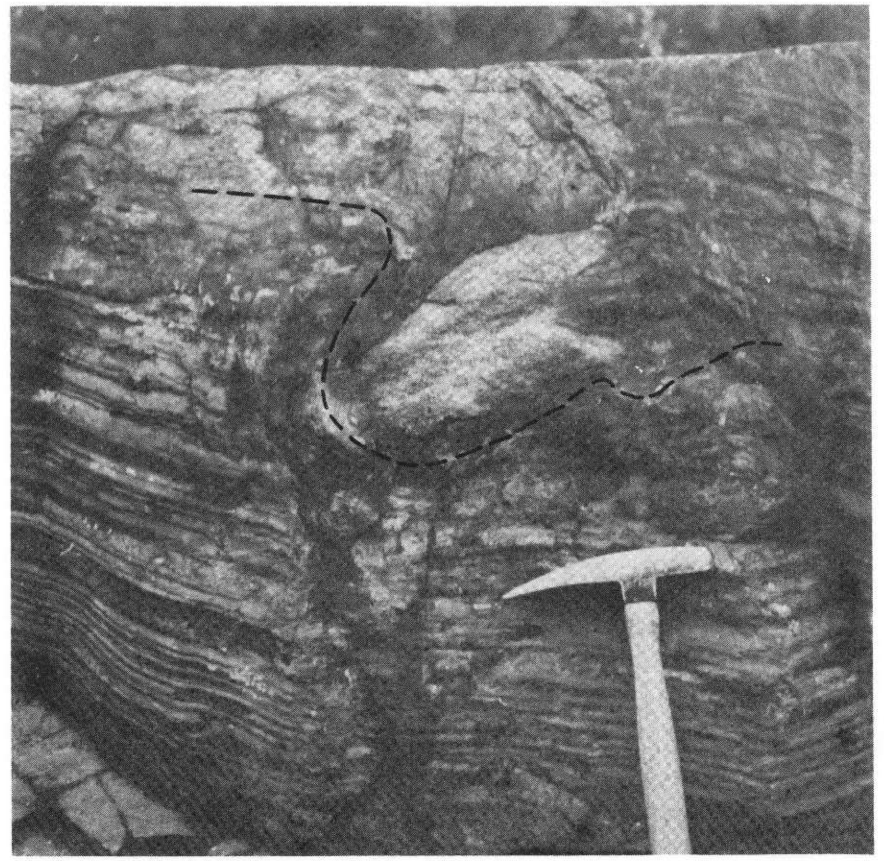
Ferris and Rudd (1971) suggested a volcanic source for the sulfur and Saager (1970) suggested a biogenic origin. The consistent values of S^{34}/S^{32} in both Canada and South Africa, however, seem to favor a volcanic origin and suggest that biogenic fractionation of sulfur isotopes was minimal and that the sulfur was probably never in the form of sulfate. However, there is no direct evidence for volcanism contemporaneous with deposition of the Witwatersrand or Blind River conglomerates which could supply the sulfur. The explanation is still missing but we believe that sulfur, as HS^-, was present and was in some way characteristic of the Early Proterozoic atmosphere and hydrosphere because every fossil placer (except the Ghana deposit) is rich in pyrite and poor in iron-titanium oxides. We doubt that each deposit had the same metamorphic history, but they were most likely deposited under similar atmospheric conditions.

Paleoclimate - The Glacial Association

An interesting feature of some Precambrian uranium-bearing conglomerates is their association with paraconglomerates of possible glacial origin--i.e. tillites. The best example is the Blind River - Elliot Lake deposit of Canada but there are several others, especially in the Northern Hemisphere. These include: the Hurwitz Group, Northwest Territories (Bell, 1970; Young, 1973); the Mistassini Lake area of Quebec (Roscoe, 1973); southeastern Wyoming (Blackwelder, 1926); and the Baltic and Ukranian Shields (Salop, 1977, Plate 3). These locations, shown in Figure 1.1, are all areas in which uranium-bearing conglomerates have also been reported. In the Southern Hemisphere, presumed glacial conglomerates have been reported in the Lower Division of the Witwatersrand Sequence (Wiebols, 1955) but, to our knowledge, uranium-bearing conglomerates in Brazil, India, and Australia are not associated with glacial deposits. The important question to the explorationist is the possibility of a genetic relationship between glacial processes and the deposition of fossil placers.

In the Blind River - Elliot Lake area of Canada, paraconglomerates occur as thick and extensive beds at the base of three of the four groups within the Huronian Supergroup: the Ramsey Lake Formation at the base of the Hough Lake Group, the Bruce Formation at the base of the Quirke Lake Group, and the Gowganda Formation at the base of the Cobalt Group (Table 2.1, page 106). The best documented glacial deposits are in the Gowganda Formation, shown in Figure 1.5, where a striated pavement, striated boulders, dropstones, varve-like beds and other evidence indicates a glacial origin. The Ramsey Lake Formation and Bruce Formation are probably also glacial but the evidence for this is not so convincing (Young, 1973; Roscoe, 1969). Roscoe (1969, p. 79) pointed out that quartz-pebble conglomerates occur beneath the Ramsey Lake paraconglomerate, the Bruce paraconglomerate, and the Gowganda paraconglomerate, and he suggested that this association might be related to glaciation. If the paraconglomerates represent tills deposited by continental glaciers, the quartz-pebble conglomerates may represent the better-sorted material of the glacial outwash plain or sandur. Sandurs are outwash plains formed by rivers carrying meltwater away from fronts of glaciers or ice caps. They are aggradational sand plains that are deposited by braided streams that continually shift position. Inasmuch as sedimentological studies suggest that uranium-bearing quartz-pebble conglomerates were deposited by braided streams and the conglomerates are interbedded with poorly-sorted, coarse-grained sandstone that has a substantial mica fraction (originally clay mineral?); the sandur model is certainly worth considering. According to Flint (1971, p. 190) the bulk of the sandur deposits seen in modern sedimentary successions was deposited during shrinkage of glaciers of the last glacial age, and Flint believes that sandur deposits are more likely to have developed during glacial retreat than during advance. As an ice-sheet retreats the sandur deposits should be laid down on top of till (paraconglomerate) and therefore the succession should be quartz-pebble conglomerate overlying

A



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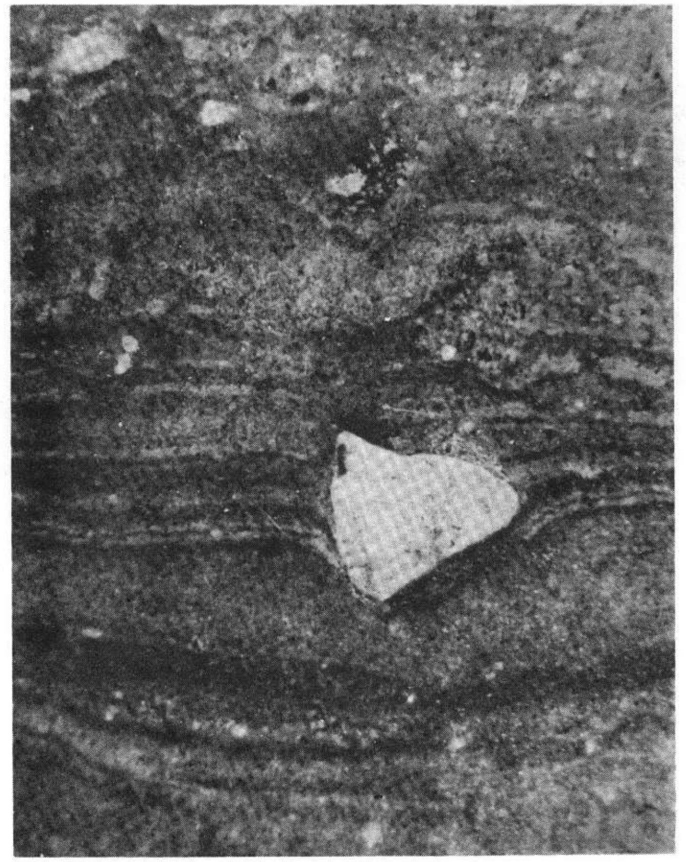


D

B

Figure 1.5. Photographs of glacial (?) diamictites from the Huronian Supergroup, Ontario and the Deep Lake Group, Wyoming.

A. Lower Gowganda Formation from the core of the Quirke Syncline, north of Elliot Lake, Ontario. Sub-angular to rounded clasts of pink granite are irregularly scattered in a dark-gray, fine-grained, chloritic matrix.

B. Outcrop of polymictic paraconglomerate of the Vagner Formation, Medicine Bow Mountains, Wyoming. Clasts of white granite, quartzite, and mafic schist are scattered in a medium-grained lithic arkosic wacke matrix.

C. Transition zone between the Ramsey Lake and Pecors Formations near Quirke Lake, Ontario. Large granite boulder was apparently dropped from floating icebergs into unconsolidated silt and clay, causing bedding to be depressed (Robertson and Card, 1972).

D. Polished slab of a dropstone structure in the Headquarters Formation, Medicine Bow Mountains, Wyoming. Bedding below the granite clast is distorted; bedding above goes across undisturbed. This suggests that the clast was ice-rafted and dropped from above into laminated subaqueous sediments. Photograph is actual size. Taken from Sylvester (1973, p. 57).

paraconglomerate -- the opposite of the Blind River-Elliot Lake succession. However, sandur deposits can develop in front of advancing ice sheets and Flint (1971, p. 154, fig. 7-1) shows an illustration of a Pleistocene example. So the succession of quartz-pebble conglomerate overlain by paraconglomerate seen at Blind River-Elliot Lake could represent this type of sequence.

Canadian geologists most familiar with the Huronian succession do not make a direct connection between possible glacial deposits and the deposition of the uranium-bearing quartz-pebble conglomerate -- that is, they do not consider the uranium-bearing conglomerate as part of sandur deposits laid down in front of advancing ice sheets or behind retreating ones. Instead indirect connections are suggested such as: deposition of fluviatile sedimentary rocks during a period of high precipitation preceding continental glaciation (Roscoe, 1969, p. 79); or deposition of fluviatile sediments following glacial rebound (Young, 1973); or deposition of uraniferous conglomerates in rapidly moving rivers "under cold, possibly frigid reducing conditions" (Robertson, 1976, p. 35). The question of a direct versus indirect connection between uraniferous conglomerates and glacial (?) paraconglomerates is really a question of timing of events and rates of sedimentation during Huronian deposition. If these deposits were laid down during a short time interval, the direct glacial connection might be established. On the other hand, if there was a long time interval between deposition of uraniferous conglomerates and glacial (?) paraconglomerates, the direct glacial connection seems unlikely. Regional stratigraphic relationships indicate that the Ramsey Lake paraconglomerate is separated from the overlying uraniferous conglomerate of the Matinenda Formation in places by up to 2150 meters of argillite of the McKim Formation and an erosional hiatus. This suggests that there was a considerable length of time between deposition of the uraniferous congloemrate and deposition of the overlying glacial units -- hence the sandur model for deposition of the Canadian uranium-bearing conglomerates seems unlikely.

A sandur-type model was also suggested by Wiebols (1955) to explain the gold- and uranium-bearing conglomerates in the Witwatersrand basin. Wiebols suggested that the Witwatersrand was an enclosed basin or inland sea situated on the edge of an ice sheet. Glacial-fluvial sedimentation was believed to have taken place on a peneplain bordering the sea where sandurs developed on the margin of the ice sheet. Wiebols thought that periodically the ice sheet advanced and deposited tills on top of the glacio-fluviatile sediments. When the ice-sheet retreated, the tills were reworked by the waters of the transgressive sea -- washing out clay and concentrating the heavy minearls. Wiebols' concept was criticized by Broch and Pretorius (1964b, p. 571-572) because it ignored the tectonic setting of the basin (there is strong evidence for marginal faulting controlling pulses of sedimentation) and because "tills" are not believed to be a really significant part of the Witwatersrand succession. Nonetheless, Weibols' concept does explain the cycles (also proposed by Pretorius, 1976a) that begin with active deposition of coarse fluviatile sediments followed by trangression and winnowing of previously deposited and sedimentary rocks.

Ahtough the direct glacial connection--fluvial deposition of uraniferous conglomerates in glacial outwash plains -- doesn't appear to withstand close scrutiny, the indirect connections cannot be dismissed. Deposition of both extensive paraconglomerates and uraniferous conglomerates appears to havebeen one of the characteristic features of Early Proterozoic successions, especially in North America. It is possible that glacial and periglacial conditions may have contributed to the large amount of coarse clastic material found in Precambrian fluviatile deposits and that the cyclic aspects of Early Proterozoic sedimentation could be explained in terms of world-wide climatic cycles.

It is also possible that the cold conditions of glacial or periglacial deposition may have been conducive to the transport of uraninite as a detrital mineral. Along this line, Trow (1977, p. 1205-1206) has proposed a model to demonstrate that uranium-bearing quartz-pebble conglomerates might form in CO_2-deficient atmospheres that are synchronous with glaciation regardless of the age of the deposit. Trow (1977, p. 1206) suggested that clastic uranyl oxide pseudomorphs after uraninite might have formed in Early Proterozoic oxidizing atmospheres if the solubility of uranium as uranyl carbonate complexes was decreased in a CO_2-deficient atmosphere. This clastic uranyl oxide then could be transported by streams to a sharp interface between oxidizing and reducing environments, where uranyl oxide would be reduced to uraninite; magnetite reduced to pyrite; ilmenite to pyrrhotite; TiO_2 would combine with UO_2 to form brannerite; and placer gold would dissolve temporarily as AuS^- to mobilize and lose its placer morphology (Trow, 1977, p. 1206). Trow further suggested that ore metals were preserved by being sealed from renewed oxidation during post-glacial conditions. This is an interesting theory but it is counter to many concepts previously proposed. It requires an oxidizing atmosphere in the Early Precambrian, a direct tie to glaciation, a rather spectacular change from oxidizing to reducing environments in the fluviatile system, and the ability of a uranyl oxide of questionable hardness to resist abrasion. Again we return to the O_2 pressure of the Precambrian atmosphere as the key issue, and again the majority of the evidence seems to favor a low P_{O_2}--in which case, a low-CO_2 atmosphere would not be required to permit the transport of clastic uraninite.

It might be suggested that paleomagnetic evidence of the paleolatitude of deposition of the paraconglomerates could resolve the issue--if glacial deposition was restricted to high latitudes in the Early Proterozoic. The Huronian basin was at latitudes of 35° during deposition of the Thessalon Volcanics (2400 m.y.), which directly underlie the uranium-bearing Matinenda Formation (Symons and O'Leary, 1978). And the Witwatersrand basin was at latitudes of about 65° during deposition of the Lower Ventersdorp lavas (2400-2500 m.y.) which overlie the radioactive units of the Witwatersrand Sequence (Piper, 1976b). Thus, the world's two biggest deposits of Precambrian fossil placers appear to have been deposited at intermediate latitudes.

However, the importance of this information is difficult to assess largely because of the scarcity of both paleomagnetic and geochronologic data from other Early Proterozoic-type successions and the inexact nature of the data which do exist. What is worse, there is no agreement among geologists that ancient glacial deposits need have been deposited at high latitudes (Harland, 1964; Piper, 1973) so, at present, paleolatitude data neither support nor undermine the glacial hypothesis.

The most frustrating aspect of this whole discussion of the possible glacial association is that there is but scanty evidence that the Proterozoic paraconglomerates are indeed of glacial origin. The best case for glacial deposition can be made for the Gowganda Formation of Canada but even here there is some doubt (Pienaar, 1963). For other paraconglomerates both in the Huronian Supergroup and elsewhere, tectonically sponsored mudflow or turbidite modes of deposition of paraconglomerates may be an equally valid alternative. Certainly tectonic cycles, such as were proposed by Pretorius (1976a) for the Witwatersrand basin, are capable of producing the observed cyclic stratigraphic successions which are common in Early Proterozoic metasediments.

World-Wide Distribution of Precambrian Fossil Placers

Paleomagnetic evidence, scanty though it is, has been interpreted by Piper (1976a; 1976b) to demonstrate that all the continents were grouped together into one supercontinent from at least 2200 m.y. (and maybe 2700 m.y.) until about 1000 m.y. This supercontinent is shown in Figure 1.6, as are the distributions of both Precambrian fossil placers and Proterozoic banded iron formations (after Goodwin, 1973). The two rock types appear to be closely associated in terms of their world-wide distribution -- both occur mainly in a narrow zone which traverses the entire supercontinent.

The implications of this are striking. First, the zone is mainly intracratonic on the pre-drift reconstruction which suggests that both fossil-placer conglomerates and banded iron formations were deposited in intracratonic basins and troughs which formed this linear zone. Second, the zone is subparallel to modern continental margins, especially between South America and Africa and in Asia (Goodwin, 1973). This suggests that this zone of sedimentation later became a zone of Phanerozoic rifting. Goodwin (1973) suggested that some unique feature of Precambrian crustal evolution caused this alignment of Proterozoic iron formations, and it is evident from Figure 1.6 that the sedimentary basins of major iron formation deposition were already in existence--and already aligned in a zone--several hundred million years earlier, during deposition of the first Early Proterozoic-type clastic

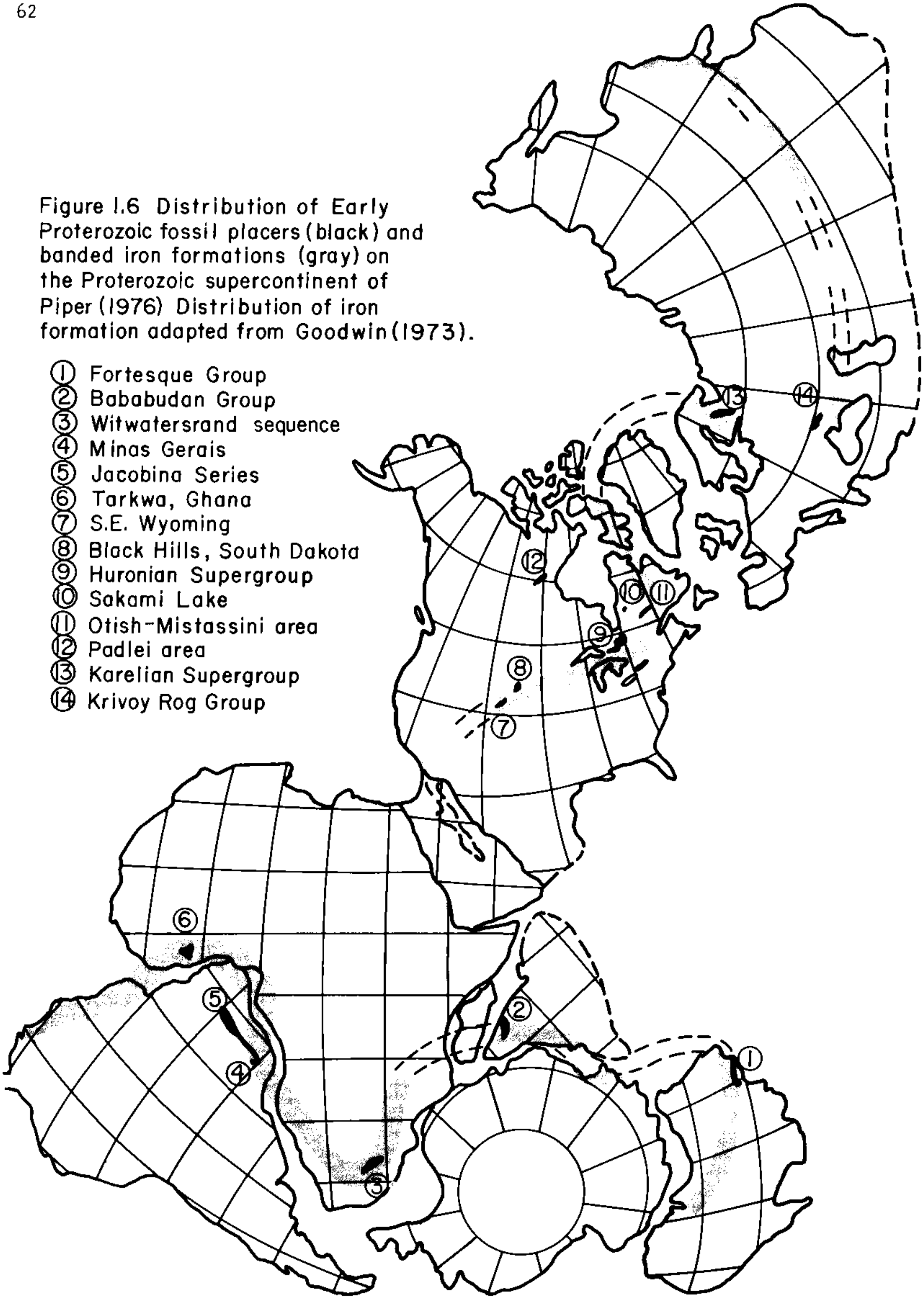


Figure 1.6 Distribution of Early Proterozoic fossil placers (black) and banded iron formations (gray) on the Proterozoic supercontinent of Piper (1976) Distribution of iron formation adapted from Goodwin (1973).

sedimentary successions. This suggests to us that the same crust-mantle tectonic regime which resulted in stabilization of continental masses at the end of the Archean and in deposition of the first clastic continental platform sequences was also responsible for the linear distribution of sedimentary basins. Perhaps incipient rifting along this zone, during the Archean-Proterozoic transition, created the sedimentary basins and provided a suitable tectonic environment for clastic deposition--and then iron formation deposition.

The most important implication to the explorationist is that, on the scale of a continent, fossil placer deposits and Proterozoic banded iron formations are spatially associated. Examples are numerous. In North America, the Huronian Supergroup is about 200-300 km east of the Lake Superior iron deposits. In South Africa, the main Witwatersrand fossil placers are just south of the iron formation in the Transvaal Sequence. In Australia, low-grade fossil placers of the Fortesque Group were deposited in the same sedimentary basin but just north of the Hamersley iron formation. In India, iron formation occurs in the Chitradurga Group which overlies the uranium-bearing Bababudan Group, and iron formation occurs in the Iron-ore Series in the Singhbhum craton to the east. In Russia, the Krivoy Rog iron formation of the Ukraine overlies rocks containing fossil-placers. And possibly the best example is the Minas Gerais area of Brazil where uraniferous conglomerates of the Minas Series pass up-section into fine-grained clastics and then to iron formation without significant hiatus. In most of these areas, the fossil placers and the iron formations occupy approximately the same sedimentary basin and, in every case, the Proterozoic banded iron formations are younger than the fossil placer deposits. Both of these facts should be useful to the explorationist.

Differences Between Major Deposits

There are important differences between fossil-placer occurrences which, ultimately, might make it possible to classify these deposits into specific types based on their geologic setting and sedimentological characteristics. In this section we attempt such a classification. However, we also recognize that there is simply too little detailed information on some fossil-placer occurrences to catagorize them properly and we consider the following classification to be provisional and subject to revision.

Perhaps the most important difference between major Precambrian fossil-placer deposits is in the distribution of reefs (ore zones) within the stratigraphic section. In some deposits there is a single ore-bearing formation which occurs at the base of the clastic succession and directly overlies the Archean basement. The best example is the Huronian Supergroup of Canada and we therefore weill refer to this type of deposit as Huronian-type. In other deposits, there are several or numrous reefs which occur higher in the stratigraphic succession and are associated with local unconformities developed on quartzites, shales or volcanics. The best example of this type of fossil-placer is the Witwatersrand Sequence and we refer to it as the Witwatersrand-type.

In Huronian-type deposits there is likely to be only one major placer zone and it occurs directly above a pronounced Archean erosion surface. This erosion surface is the key to the deposit. It represents a prolonged period of erosion in which detrital ehavy minerals were mechanically liberated from the source terrain and accumulated in a weathering residuum. This detrital residue was then swept into localized topographic depressions in the erosion surface by streams and these depressions became the sites of placer deposition. The erosion surface is often represented by a regolith which gradationally separates arkosic conglomerate from unweathered granite. Ore may occur directly above the regolith but, more commonly, the ore occurs in one or several discontinuous conglomerate lenses within a basal quartzite sequence several hundred meters thick. Exploration for this type of deposit is relatively simple; it focuses on identifying the basal Proterozoic nonconformity and sampling the basal clastics.

Witwatersrand-type deposits are more complex. Here, heavy minerals accumulated at various levels in the stratigraphic sequence, in specific parts of tectonically controlled (?) sedimentary cycles. Most commonly, reefs occur at the conglomeratic base of regressional, fining-upward sequences. This type of deposit is recognized by the local unconformity at the base of the regressional cycle--there are as many potential ore horizons as there are local unconformities. Exploration for this type of deposit requires detailed sedimentological study to outline the facies patterns, stratification sequences, and depositional environments of the sediments. The explorationist needs to identify both large and small sedimentary cycles, with emphasis on conglomerates overlying local unconformities within large-scale regressional sequences.

Table 1.1 is an attempt to classify Precambrian fossil placers according to two end members: Huronian versus Witwatersrand types. We recognize a continuum between the two types so that classification of some deposits was rather arbitrary and depended upon which type of exploration approach we thought would be most useful in outlining the deposit. If the uranium- or gold-bearing units in a sequence are closely associated with a profound basal Proterozoic nonconformity we called it Huronian-type. If, however, the mineralized horizons are associated with localized unconformities within the stratigraphic succession so that more detailed mapping and sedimentology are required to outline the deposit--we called it Witwatersrand type. For some deposits, the distinction cannot be readily made based on available information so Table 1.1 must be viewed as a provisional classification. The main facts used to classify each occurrence are summarized below. More detailed information on individual occurrences is presented in Part II.

Table 1.1 Provisional Classification of Precambrian Fossil Placers

Huronian - type	Witwatersrand - type
Huronian Supergroup, Ontario	Witwatersrand, South Africa
Fortesque Group, Australia	Tarkwa Group, Ghana
Krivoy Rog Group, Ukraine	Deep Lake Group, Wyoming
Bababudan Group, India	Estes Group, South Dakota
Jacobina Series, Brazil---→	←---Minas Gerais, Brazil
Karelian Supergroup, Finland---→	←---Karelian Supergroup, Russia
Montgomery Lake Sediments, N.W. Terr.	
Sakami Formation, Quebec---→	

Arrows indicate that deposits may be intermediate in character between two end member-types.

Huronian-type Deposits

Fortesque Group, Australia. Uranium-bearing conglomerates occur in the basal clastic unit of the Fortesque Group which crops out along the south margin of the Pilbara Archean craton and possibly along the northeast margin of the Yilgarn craton. The basal clastic unit of the Fortesque Group occupies topographic depressions in the Archean erosion surface and the best exploration technique would appear to be a detailed sampling program along this erosion surface. Hence, we classify this occurrence as Huronian-type.

Krivoy Rog Group, Ukraine. Conglomerates containing gold, pyrite, and presumably uranium (though not reported in the Russian literature) occur in the basal clastic units of the Krivoy Rog Group which directly overlie Archean basement.

Bababudan Group, India. Uranium and gold occur in the basal Chickmagalur conglomerate of the Bababudan Group of Karnataka. This conglomerate unconformably overlies Archean basement so we consider it to be Huronian-type.

Jacobina Series, Brazil. Gold-bearing reefs occur within the lower 500 m of the Jacobina Series. According to Gross (1968), these rocks nonconformably overlie Archean basement. However, Cox (1967) reported a basal fault contact and, in places, the Jacobina Series is underlain by older metasediments. Thus this deposit may be intermediate in character. However, we classify this occurrence as Huronian-type because the ores are distributed along the Archean-Proterozoic contact.

Karelian Supergroup, Finland. Uranium occurs mainly in quartzites and conglomerates which closely overlie a regolith developed on the Archean erosion surface. Hence we classify this occurrence as Huronian-type.

Montgomery Lake Sediments, Northwest Territories. Mineralized conglomerates occur in a quartzite sequence which is preserved only in isolated synclinal inliers in Archean basement. These may, in part, represent original topographic depressions in the Archean erosion surface in which gravels accumulated.

Sakami Formation, Quebec. Uraniferous conglomerates overlie a tightly folded Archean greenstone sequence and appear to be basal Proterozoic conglomerates. However, the details on conglomerate distribution are not available to us and this occurrence could be intermediate in character.

Witwatersrand-type Deposits

Tarkwa Sequence, Ghana. Gold-bearing conglomerates occur in several reefs located above local unconformities in the Tarkwa Sequence, which overlies an older metasedimentary succession. This occurrence is very similar to the Witwatersrand deposit in that sedimentological studies play a key role in outlining the mineralized zones. They are different, however, in that the Tarkwa deposit contains gold and hematite--and no uranium, whereas the Witwatersrand contains gold, pyrite, and uranium.

Deep Lake Group, Wyoming. Uraniferous conglomerates occur in the lower part of the Deep Lake Group which unconformably overlies both older metasediments and Archean basement. The main mineralized horizon occurs above this major unconformity and, in that respect the Deep Lake Group resembles a Huronian-type deposit. However, uraniferous rocks also occur in the older metasedimentary sequence and in quartzites well up in the Deep Lake Group. Also, sedimentological studies and detailed mapping of various unconformities may be critical in outlining the potential of this occurrence. We consider this occurrence to be perhaps more similar to the Witwatersrand than the Huronian deposit.

Estes Group, South Dakota. The situation here is similar to that of the Deep Lake Group because uraniferous conglomerates overlie both an older metasedimentary succession and Archean basement. We consider it to be transitional in character based on this analogy with the Deep Lake Group.

Minas Gerais, Brazil. Uraniferous conglomerates are located within a 300 m thick quartzite and conglomerate sequence at the base of the Minas Series, which unconformably overlies an older metasedimentary sequence. This occurrence is intermediate in character--the reefs are related to one major unconformity but this unconformity is not a basal Proterozoic nonconformity.

Karelian Supergroup, Russia. Gold-, pyrite-, and presumably uranium-bearing conglomerates occur in the Karelian Supergroup of Soviet Karelia as well as Finland. Here, mineralized conglomerates overlie both Archean basement and older quartzites, shales, and felsic volcanics and they appear to occur in various horizons--all related to unconformities, but not necessarily to a basal Proterozoic nonconformity.

EXPLORATION MODEL

A number of very good exploration models have been proposed for Precambrian uranium-bearing, quartz-pebble conglomerates including two exceptional ones by Pretorius (1976a) and Roscoe (1969, p. 166-167). These two models, as well as those of Whiteside (1970) and D.S. Robertson (1974), are based on the concept that the uranium minerals were detrital and were deposited in fluviatile sedimentary rocks. The writers are in general agreement with this concept based on the evidence cited above, but we recognize that many questions concerning these deposits are unanswered, and we will therefore present a preferred model followed by a brief discussion of some alternate concepts.

PREFERRED MODEL

1. Age Constraints

Uranium-bearing quartz-pebble conglomerates are found almost entirely in Early Proterozoic-type metasedimentary successions which range in age from 2800 m.y. to about 2000 m.y.. Gold-bearing placers range in age from about 3100 m.y. to the present.

(a) The maximum age for uranium and gold placers is related to crustal evolution. Early Proterozoic-type sedimentary rocks are considered to have developed after a period of Late Archean (3000-2500 m.y.) cratonization when a substantial continental mass or masses developed allowing widespread sedimentation on low-lying stable platforms. This cratonization did not take place at the same time everywhere on the planet but was essentially completed everywhere by 2500 m.y..

The diachronous nature of the Archean-Proterozoic transition means that rocks classed as Archean by some observers may have Early Proterozoic-type sedimentary successions.

(b) The minimum age for uranium-bearing fossil placers is related to atmospheric evolution. Exploration for uranium-bearing quartz-pebble conglomerates should be confined to sedimentary successions older than about 2000 m.y. because most geologic, geochemical and paleontologic evidence suggests that the Precambrian atmosphere contained sufficient oxygen by this time to allow oxidation and dissolution of uranium minerals in the zone of weathering and thus make them unstable for transportation and deposition in fluviatile deposits. In contrast, there is no apparent minimum age for gold placers.

2. Source Area Constraints

The best source area contains Late Archean, K-rich granites and Archean greenstone belts.

(a) The source of uranium in Early Proterozoic metasedimentary rocks was probably Late Archean (3000-2500 m.y.) granites. There is substantial evidence that Early Archean granites and gneisses are uranium poor, but some Late Archean granites contain appreciable uranium. A geochemical study of these granites might be fruitful in locating provinces where there is a uranium potential in younger rocks.

(b) The source of gold and much of the detrital pyrite in the placers was probably Archean greenstone belts. Thus, the favorability of a fossil placer for gold can, in part, be assessed by determining whether gold-bearing greenstone belts are present in the Archean source terrain.

3. Stratigraphic Constraints

The most favorable rocks are Early Proterozoic sedimentary successions with a preponderance of sandstone, conglomerate, shale, and carbonate and a relatively minor proportion of volcanic rocks. They can be recognized by their strong resemblance to Phanerozoic sedimentary successions and by certain characteristics that may be unique:

(a) Most Early Proterozoic-type sedimentary successions lie nonconformably on a more highly metamorphosed Archean granite gneiss terrain which may contain infolded greenstone belts. This unconfomity represents considerable hiatus and the Archean rocks may show evidence of deep residual weathering prior to deposition of Early Proterozoic rocks.

(b) Other Early Proterozoic-type successions unconformably overlie earlier metasedimentary successions which contain a higher proportion (> 50 percent) of volcanic rocks interbedded with clastic sediments.

(c) The thickest and most extensive beds of laminated iron formation are in Early Proterozoic sedimentary successions, but these rocks are usually younger than uranium-bearing quartz-pebble conglomerate and are only rarely in direct contact with them.

(d) Many Early Proterozoic successions are characterized by stromatolitic dolomites, but again these rocks are not normally in close association with uranium-bearing, quartz-pebble conglomerate. They usually lie above them.

(e) Early Proterozoic sedimentary rock-types in direct association with uranium-bearing quartz-pebble conglomerate include paraconglomerate, quartzite, and lesser amounts of slate and limestone. These metasedimentary rocks may be arranged in sequences with paraconglomerates at the base, grading upward into shales and limestones, in turn overlain by quartzite that may be in part, fluviatile. This sequence may be repeated and may be in some manner related to Precambrian glacial episodes, but the cyclic successions may or may not control the deposition of uranium-bearing, quartz-pebble conglomerate. Although we do not understand the manner of formation of these sequences they are still quite useful in prospecting simply as a rock succession with which uranium-bearing, quartz-pebble conglomerate may be associated.

(f) Some Late Archean supracrustal successions contain sedimentary sequences that are transitional between clastic rocks of true greenstone belts and clastic rocks of the Early Proterozoic-type successions. No uranium-bearing quartz-pebble conglomerate has been reported in these sedimentary successions, but we consider them to be legitimate exploration targets, albeit of secondary interest to the true Early Proterozoic-type sedimentary succession.

4. Sedimentological Constraints

The sedimentary environment most favorable for deposition of uranium-bearing quartz-pebble conglomerates is considered to be a fluvial-deltaic environment that has the overall characteristics of braided stream or river systems.

(a) The sedimentary succession is marked by fining-upward stratification sequences, with clast-supported conglomerate at the base grading upward into cross-bedded, coarse-grained sandstone (cross-beds may be low-angle, or inclined about 30°, or trough), and having climbing ripple sets near the top of the sequence.

Unlike meandering river systems the braided stream fining-upward sets do not usually contain shales or siltstones near the top and the in-channel deposits (coarse-grained sandstones) may be quite thick. A diagrammatic illustration of such a stratification sequence is shown in Figure 1.7. However, the exact characteristics of the stratification sequence will vary depending on proximity to source, tectonism, or climate so that the development of the various units within the fining-upward sets may vary significantly. Usually these fining-upward sets are repeated in ancient braided river deposits as a result of channel migration.

(b) Sedimentary structures include abundant channels, small scours, and trough cross-bedding. The conglomerates may show upstream pebble imbrication. Paleocurrent distributions tend to be unimodal and low-variance in individual areas but may vary considerably over the depositional basin reflecting different points where rivers entered the basin.

(c) The sedimentary succession may show evidence of cyclic sedimentation. This may take the form of repetition of fining-upwards stratification sequences or repeated paraconglomerate-argillite-quartzite sequences. This is an important aspect of Early Proterozoic sedimentation because it indicates the occurrence of regressional and transgressional episodes which were important in continually reworking and reconcentrating the placers. The control of these cycles may have been tectonic movements in the source area and basin margins and/or fluctuations in sea level due to climate changes.

(d) As a result of cyclic sedimentation, uranium-bearing conglomerates often occur at more than one horizon within the quartzite succession. These layers may range in thickness from that of a single

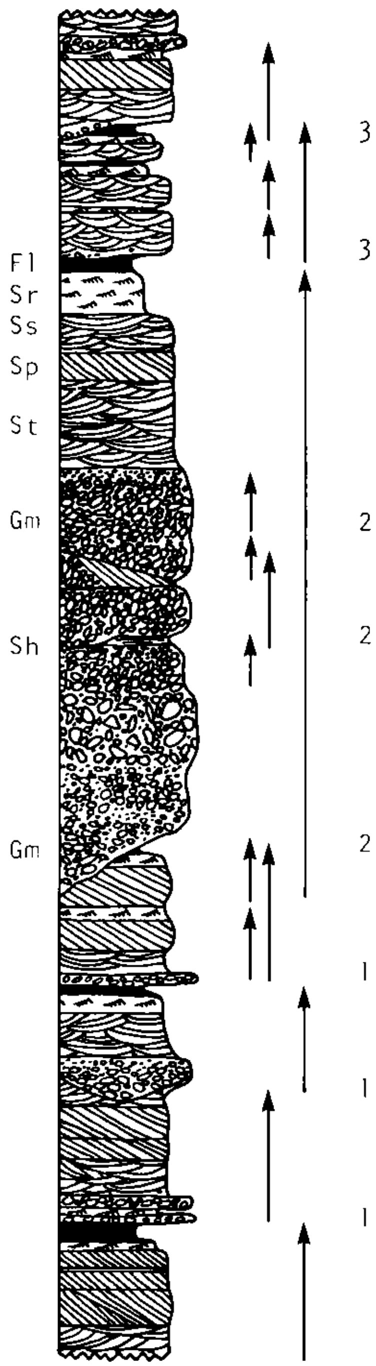


Figure 1.7. Diagrammatic stratigraphic column showing idealized distribution of sand and gravel lithofacies in a braided river deposit (facies labels explained below). Taken from Miall (1977). Arrows at right of column show major and minor cycles which represent main and secondary channel systems. The scale of the entire column can vary from meters to tens of meters as can the lateral extent of individual facies units.

Placer uranium and gold are found in several places in such a sequence:

1. in isolated conglomerate beds at the base of fining-upwards cycles. Such conglomerates represent deposition on isolated longitudinal bars and may or may not be of economic importance.

2. in thick conglomerate sequences containing many gravel-sand cycles which represent deposition of superimposed longitudinal bars in response to fluctuating flood or valley-fill conditions. Examples are the Main-Bird series of the Witwatersrand and the Matinenda ores of Blind River, Canada.

3. in coarse sands overlying local unconformities. Erosion of underlying units causes accumulation of residual heavy minerals on the erosion surface so this type of horizon can be economically important if the unconformity cuts placer-bearing gravels or if carbonaceous material developed on the unconformity.

Facies labels: Gm - massive gravel; Sh - horizontally-bedded sand; St - trough crossbedded sand; Sp - planar cross-bedded sand; Ss - scour-fill sand; Sr - ripple cross-laminated sand; Fl - laminated sand, silt, and mud.

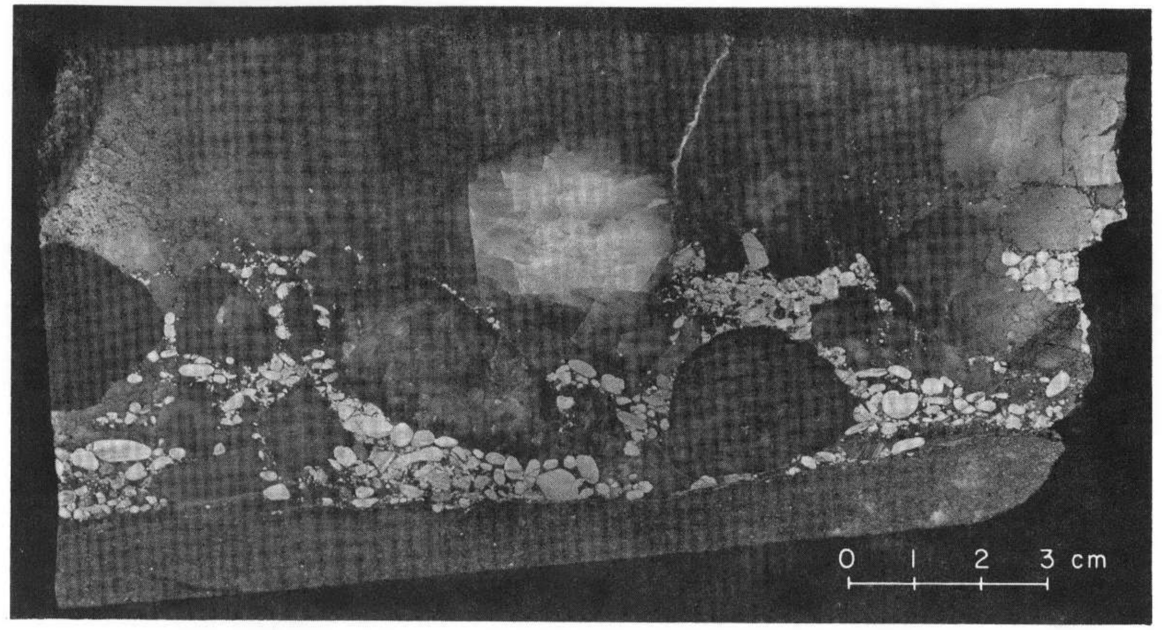
pebble to compound zones of conglomerate ten meters or so in thickness.

(e) The most important sedimentologic feature in locating mineralized zones within the sedimentary succession is the unconformity. All mineralized zones in fossil placer deposits are related to regional or local unconformities. A simplistic model (which is often the best) might be expressed by stating that the best opportunity for development of uranium-bearing, quartz-pebble conglomerate is at the unconformity that marks the base of the first true Proterozoic-type sedimentary rocks. Major Canadian, Brazilian, Indian, Australian, and Finnish deposits lie at or near this Proterozoic-Archean nonconformity. Thus, if the Archean-Proterozoic unconformity can be located in a given area it is clearly a good exploration target. However, there are important fossil placer deposits that are not directly associated with a basal Proterozoic unconformity such as the Witwatersrand, Ghana, Wyoming, and Minas Gerais, Brazil deposits. For these, the key to exploration is to identify the local unconformities at the base of large, regressional fining-upward stratification sequences.

5. Lithologic characteristics

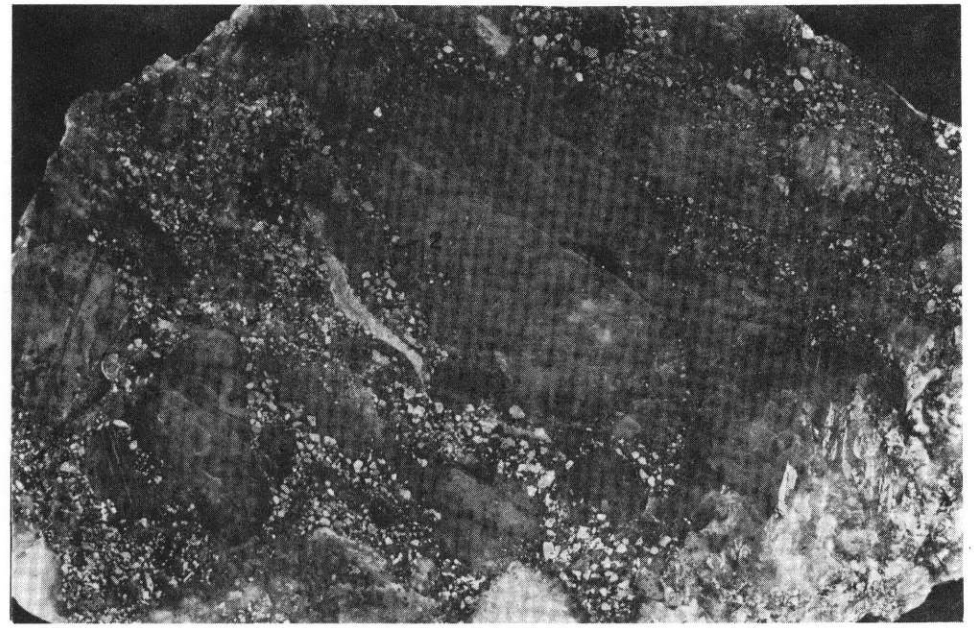
Certain lithologic characteristics seem to typify rock types that contain Precambrain uranium-bearing quartz-pebble conglomerates. These characteristics can be seen in Figure 1.8.

(a) Color--Fresh samples of sedimentary rocks deposited prior to 2000 m.y. ago tend to be characterized by drab colors--gray, green, white, or buff-color and they generally lack pink, red, orange and

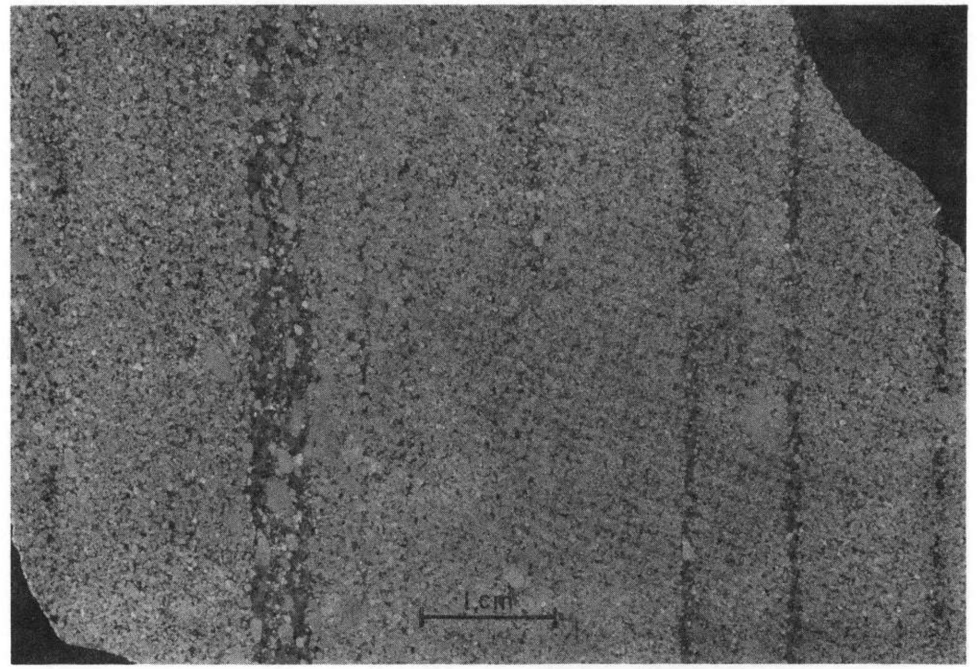


A. Vaal Reef placer from the Witwatersrand, South Africa; gold, uraninite, and rounded pyrite grains occur in the matrix of quartz- and chert-pebble conglomerate, which rests unconformably on underlying barren quartzite; note irregularity of basal contact. Average uranium content of the Vaal Reef is 500 ppm; average gold content is 15 ppm (Minter, 1976).

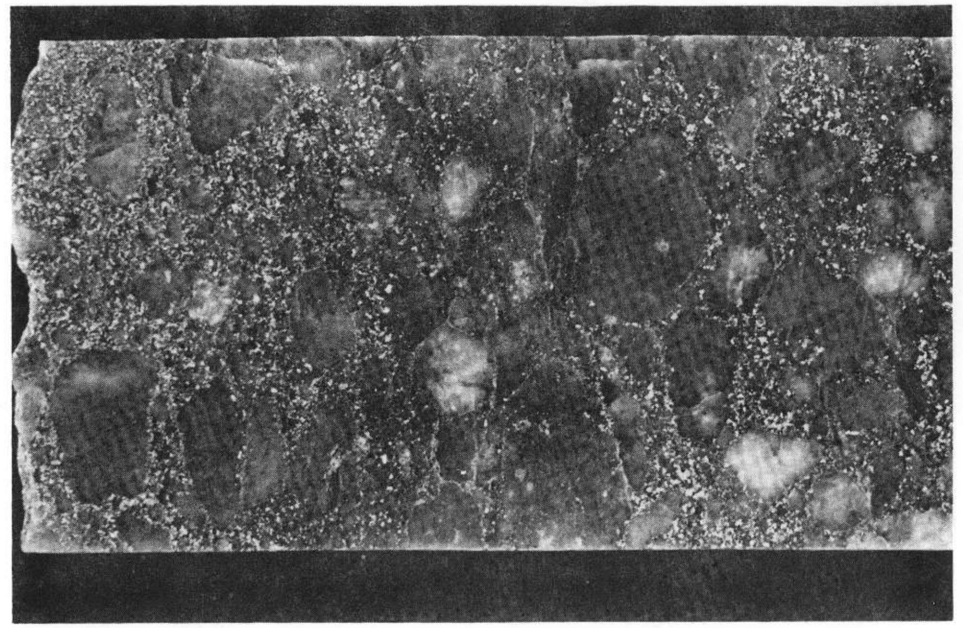
B. Detail of basal contact of Vaal Reef (from A) showing well-rounded granules of nodular pyrite along erosional unconformity. Fine-grained gold is concentrated within one centimeter of the basal contact and is interpreted by Minter (1976) to be a basal concentrate which lay in depressions on the unconformity surface prior to deposition of the gravels. Gold concentrations often reach 1500 ppm in the Vaal Reef (Minter, 1976).



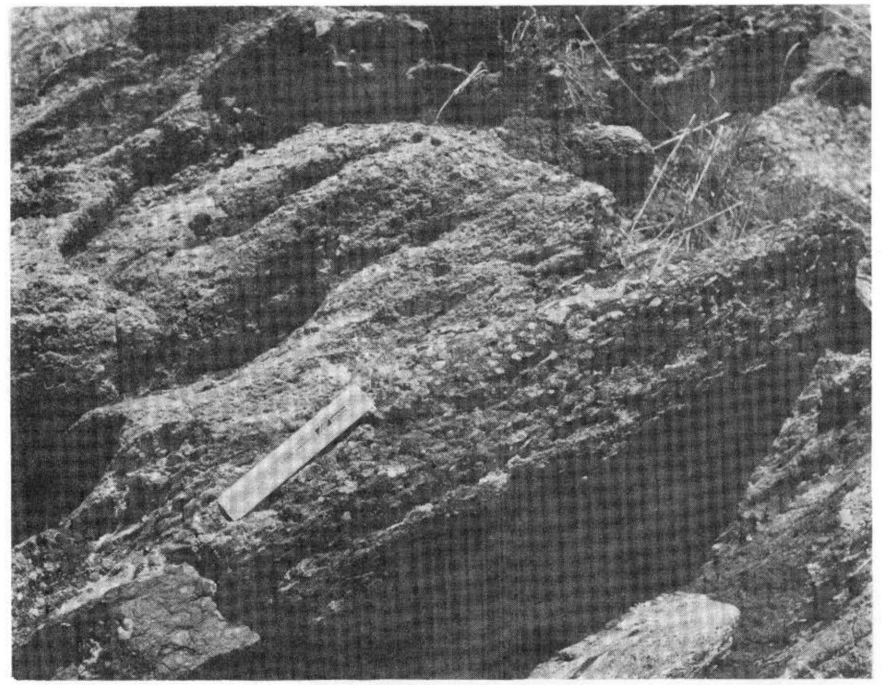
C. Quartz- and chert-pebble conglomerate of the Matinenda Formation from the Denison mine near Elliot Lake, Ontario. Pyrite is irregularly distributed through the matrix and occurs both as well-rounded grains (1) and euhedral grains (2). Pyrite euhedra were probably formed during post-depositional recrystallization of originally detrital grains. Uranium occurs in the matrix as brannerite and uraninite; uranium ore averages 14 percent from the Denison mine. Photograph is actual size.



D. Well-rounded pyrite sands from a lower-energy, distal portion of the Klerksdorp fan-delta, Witwatersrand, South Africa. Distal facies rocks such as this may contain appreciable uranium but are not as rich in gold as proximal facies rocks such as in A. Black layers show bedding orientation and are composed of quartz-rich sands; the light-gray grains are pyrite which makes up more than 85 percent of the rock.



E. Pyritic quartz-pebble conglomerate from drill core near the Carrico Ranch, northwestern Sierra Madre, Wyoming. The pyrite in the matrix of the conglomerate is relatively fine-grained (average diameter is about 1 mm) and grains are mainly euhedral. Assays from this core are not available yet but similar drill core from the northeastern Medicine Bows yielded 885 ppm U and 1870 ppm Th. Core diameter is 6.25 cm (2.5 inches).



F. Outcrop of the quartz-pebble conglomerate shown in E (above), from the Carrico Ranch. Pyrite is nearly completely oxidized in outcrop and only boxworks of square pyrite casts are visible. Oxidation of pyrite has stained conglomerates dark red. Uranium has also been strongly leached from outcrops and Th: U values are high (about 6:1). Maximum uranium content found so far is 131 ppm U. Scale is 17.5 cm (7 inches) long.

brown coloration. This is thought to reflect the oxidation state of iron, as Fe^{+2}, in the low-oxygen atmosphere. This is a useful exploration tool because uranium-bearing rocks are generally stratigraphically below the oldest red colored rocks.

(b) Composition--The most important lithologic types are quartz-pebble conglomerate and quartzite. Quartz-pebble conglomerates contain well-rounded quartz pebbles several centimeters in diameter in a quartz-sericite matrix. The pebbles commonly constitute 60-80 percent of the rock and are mainly quartz and chert, with minor rock fragments such as granite, quartzite, and schist. The pebbles are often, but not always, in contact, forming a pebble supported conglomerate. The matrix of the conglomerates contains quartz, sericite--which may constitute up to 25 percent of the matrix, and heavy minerals--usually less than 5-20 percent. Surrounding quartzites range from orthoquartzite to greywacke in composition but the predominant composition is subarkose.

(c) Textural maturity--Although uranium-bearing, quartz-pebble conglomerate has been referred to as mature orthoquartzitic conglomerate, implying that they are composed of quartz clasts, are well-sorted, and have a mineral cement, this is not generally true. Both the conglomerate and interbedded quartzite are only moderately well-sorted (bimodal or multi-modal), sericite and other phyllosilicate minerals are commonly abundant in the matrix, and detrital grains are often poorly rounded. In fact, in Canadian deposits, it is the poorly sorted sericitic grits and associated conglomerate units that are typically radioactive and interbedded orthoquartzites and orthoquartzite conglomerates are barren. However, it is true that the matrix of mineralized conglomerates tends to be compositionally and texturally more mature than adjacent quartzites and that clasts are well rounded and predominantly quartz and chert. In general, the sub-mature nature of mineralized conglomerates and quartzites seems to fit the fluvial model very well.

(d) Sericite--The association of sericite with mineralized conglomerates and quartzites is a useful exploration tool. Its presence might indicate that these sedimentary rocks contained a significant clay mineral fraction prior to diagenesis and metamorphism. Another possibility is that sericite is a primary mineral derived from a weathered granitic or granite gneiss source rock. We prefer the latter interpretation because the majority of the heavy minerals (uraninite, zircon, monazite, urano-thorite, ilmenorutile, rutile, allanite, cassiterite, magnetite (original), and ilmenite (original) were probably derived from a granitic source.

6. Mineralogical Characteristics

(a) Pyrite is the principal heavy mineral in Precambrian conglomerates that contain uranium minerals, and it is not only present in the conglomerates but is a common detrital mineral in associated quartzites. In conglomerates it commonly comprises 5-20 percent of the matrix and in banded quartzites it can comprise up to 25 percent of the entire rock. Its presence or absence is a good guide to the presence or absence of uranium minerals because both required a low-oxygen atmosphere for detrital transport.

(b) Uraninite is the most common uranium mineral. It occurs predominantly in grains about .1 mm in diameter in the matrix of conglomerates. It also occurs as minute crystals in thucholite. Other uranium minerals include brannerite, uranothorite, and coffinite.

The total uranium content of ore-grade conglomerates averages between about 1000 and 1500 ppm but uranium concentrations up to one or more percent have been reported. Also, uraninite veinlets associated with thucholite have yielded values higher than ten percent uranium.

(c) Thorium is an important constituent of Precambrian uranium-bearing quartz-pebble conglomerates; perhaps as typical of these conglomerates as uranium. Thorium minerals include monazite, uranothorite, uranothorienite, thorite, xenotime, and others. Thorium is also present in much of the uraninite. The presence of thorium contrasts with uranium deposits of chemical origin, where thorium is largely absent, and is additional evidence for a detrital origin of the mineralization.

(d) Gold is present only in some fossil placers. Where present, it is found in the basal part of conglomerate reefs, in banded quartzites, and in thin films along unconformity surfaces and carbon seams. The gold may be extremely fine-grained and not visible to the naked eye. In sampling for gold assays, it is extremely important to identify the unconformity surface and sample directly on or above it.

(e) The carbonaceous material commonly referred to as thucholite is probably some form of primitive plant that occurs in mat-like beds and as clastic particles in quartzite and conglomerate. The plants that produced this organic material seem to have flourished at the same time uranium-bearing, quartz-pebble conglomerates were deposited. This organic material is difficult to identify in the field, but its presence would certainly be a good indicator of the possible presence of detrital uranium minerals.

(f) Leaching has often altered the mineralogical composition of conglomerate outcrops because uraninite and pyrite are metastable minerals in modern oxidizing environments. Depending on the duration of exposure and type of climate, these minerals may be wholly or partially leached from surface outcrops. The outcrop, however, may show evidence of the former presence of these minerals in gossan-type residues containing reddish, hydrated iron oxides and voids with distinctive crystal form. Uranium may be wholly or partially leached from such outcrops, but inasmuch as thorium is usually present in uranium-thorium minerals or as detrital thorium minerals, the outcrop will be radioactive. These oxidized outcrops usually show a very high Th/U ratio (5-15), but samples acquired from below the zone of weathering usually exhibit a much lower Th/U ratio (1-2).

7. Preservation and Metamorphism

The biggest and best fossil placer deposits are in greenschist or lower grade metasedimentary rocks which have never been involved in orogenic deformation. This type of sequence is obviously the best exploration target. However, uranium and gold have not been appreciably remobilized in some deposits of amphibolite grade (e.g. Brazil and possibly Wyoming). Thus, it is reasonable to explore for fossil placer uranium deposits in any Early Proterozoic-type quartzite successions which are amphibolite facies or lower. Granulite facies sequences and rocks which have been intensely deformed are unlikely to contain fossil placer deposits because uranium is likely to have been remobilized. Even here, however, if the quartzites originally contained uranium, they might have been a good source for nearby epigenetic vein and unconformity uranium deposits.

ALTERNATIVE CONCEPTS

The previous model is, as all models are, an attempt to bring together a vast body of observational data and theoretical concepts into a coherent explanation of the genesis of Precambrian fossil placers. Formulation of such a model requires a process of evaluating different, and sometimes conflicting, ideas and observations--some to be believed and others to be ignored. The result is a generalization which hopefully will be of use in uranium exploration and hopefully will not assume the role of a restrictive dogma. The basis for good science is the use of the method of multiple working hypotheses (Chamberlin, 1897) and it is particularly important for the Precambrian geologist to remain open to new interpretations and new information. With this in mind, we briefly outline some alternative genetic concepts regarding fossil placers which should be used to salt the "preferred model".

There is one important aspect of the model which could be drastically wrong--that is our generalization that uranium-bearing placers are restricted to rocks older than 2000 m.y. This generalization comes from the fact that the few well-dated uranium placers are older than 2000 m.y. and from the concepts that the Early Proterozoic atmosphere was low in oxygen and that uraninite cannot be transported in today's atmosphere but could be transported in a low-oxygen atmosphere.

In all objectivity, however, it must be admitted that the only uranium-bearing fossil placers that we know absolutely to be older than 2000 m.y. are the two big ones--Blind River-Elliot Lake and the Witwatersrand. Also, the evidence for a low oxygen atmosphere in the Early Proterozoic is based on various geochemical models and has vague similarities to a house of cards. In addition, detrital uranium minerals are not entirely confined to fluviatile sedimentary rocks older than 2000 m.y.--but are also reported from recent alluvial deposits of the Indus River, along with gold and pyrite (Darnley, 1962; Miller, 1963; Simpson and Bowles, 1977). Thus, it behooves us to ask--what situations might promote deposition and preservation of uranium minerals in sedimentary rocks younger than 2000 m.y.?

The writers are not of the opinion that glaciation was a direct control on the formation of Precambrian uranium-bearing, quartz-pebble conglomerate for reasons cited above, but chemical weathering is inhibited during glacial episodes, and a large volume of clastic material is transported, deposited, and reworked in glacial outwash or sandur deposits. These sandur deposits should be investigated in areas where uranium minerals are known to be present in a source area. Trow (1977, p. 1205) states that uranium deposits have been found in both Late Precambrian and Eocambrian deposits using a glacio-chemical model that he proposed. We cannot fully evaluate Trow's model from information presented in his abstract, but we believe that the glacial environment deserves more consideration as a favorable one for uranium deposition than it has so far received.

Another geologic situation where weathering and deposition might be rapid enough to preserve detrital uranium minerals even in an oxidizing environment is in areas where there is a production of a substantial volume of felsic and intermediate composition volcanic detritus. Reworking of volcanic debris in streams, rivers, and finally on beaches has resulted in accumulation of thick black sands that consist primarily of heavy minerals of volcanic origin, in New Zealand, for example. These black sands are composed of common heavy accessory minerals such as magnetite and ilmenite. Nevertheless, some consideration might be given to volcanic source rocks that might provide detrital uranium and thorium minerals, either in modern or fossil volcanic sedimentary successions.

Inasmuch as detrital uranium minerals appear to have been deposited in

Precambrian uranium-bearing, quartz-pebble conglomerates in braided streams or river systems and inasmuch as the only significant occurrences of uraninite in a modern river (the Indus and Hunza Rivers) are in a braided river system, consideration might be given to the origin of such fluviatile systems present and past. These braided stream systems clearly develop where there is aggradation due to an overload of clastic material and they are also more common in arid climates where the water volume cannot overcome the supply of detritus. It is conceivable that the sedimentary environment is more important than it has been given credit for, and that, if available in a source area, detrital uranium minerals might be found in any fluviatile succession deposited by a braided stream or river system. In the western United States most of the chemically deposited uranium is in arkosic sandstones deposited by braided streams or rivers. Most observers have concluded that the uranium was leached from granites in adjacent highlands and/or from tuffs overlying the arkosic sandstone, but little attention was given to the possiblity that this uranium came from heavy minerals in the arkosic sandstone. In this connection it is interesting to review again the hypothesis of Simpson and Bowles (1977) who considered the uranium of the Witwatersrand to be deposited both as clastic uraninite (high thorium) and as chemical uraninite (low thorium) in an atmosphere not too different from today's. Obviously, if these gentlemen are correct, neither clastic uraninite nor chemically deposited uraninite need be confined to a specific geologic time span. Perhaps all of this is relative, and we are indeed correct in stating that clastic uraninite is more likely to be found in rocks older than 2000 m.y., but some may have been deposited in braided streams after this date. Thus we may find either clastic uraninite or chemical deposits derived from such uraninite if we have a thorough understanding of the geologic history of a given area.

There are two other aspects of our model which could be too restrictive. First, we have emphasized Early Proterozoic metasedimentary successions as the best hosts for uranium-bearing fossil placers because these are the hosts for known deposits. However, we see no reason that uranium-bearing sediments might not also be found in clastic rocks in the upper part of greenstone sequences or in other Archean supracrustal successions. Again the key may be the depositional environment and not age.

Second, we have emphasized detrital transport and placer deposition of uranium and gold so we have focused on conglomerates (and quartzites) as the most favorable host rocks. At the same time, we reviewed evidence that both uranium and gold may have been transported in solution as well. If this were true on a large scale, the explorationist should consider Early Proterozoic black shales and slates -- such as in the Wollaston Group of the Churchill Province of Canada, for low grade uranium deposits, and siltstones, mudstones, and carbonates -- such as the Transvaal Sequence of South Africa, for deposits of gold.

In conclusion, the uranium explorationist should remain open to a large range of possibilities. Uranium is commonly found in certain mineral districts, but, within these districts, it may occur in many different lithologies and in rocks of several different ages because of its high mobility. Thus, in areas of known placer deposits, the explorationist should consider the possibility that uranium may have been also chemically deposited (or redeposited) in a variety of environments and, in areas of known chemical uranium deposits, the explorationist should examine any nearby conglomerates and quartzites with the idea that low-grade uranium placers may have been one source for the uranium. Also, the explorationist should not ignore other minerals of potential economic interest when he examines fossil placer deposits.

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PART II. SUMMARIES OF KNOWN AND POTENTIAL OCCURRENCES OF URANIFEROUS AND AURIFEROUS QUARTZ-PEBBLE CONGLOMERATES

CANADIAN SHIELD

HURONIAN SUPERGROUP

Geologic Setting

The Huronian Supergroup is a sequence of low-grade metasedimentary rocks which crops out in the Southern Province of the Canadian Shield (Card and others, 1972), in a belt which extends from Sault Ste. Marie, Ontario to Noranda, Quebec. These rocks are bounded on the north by the Archean Superior Province, on the east by higher-grade metamorphic rocks of the Grenville Province, and on the south by onlapping, undeformed Paleozoic sedimentary rocks (Figure 2.1).

The Superior Province formed the basement on which the Huronian Supergroup was deposited and is composed of Keewatin-type greenstone belts surrounded by intrusive granites. Greenstone belts occur in the east-west trending arcuate belts and contain basic volcanics, felsic volcanics, and volcaniclastic sedimentary rocks. The greenstone belts were deposited between 2950 and 2750 m.y. ago (Krogh and Davis, 1972) and were metamorphosed to greenschist facies during the Kenoran orogeny -- 2500 m.y. ago. Algoman granites were also emplaced during the Kenoran orogeny. After the Kenoran orogeny, basic dikes were emplaced into the newly stablized craton, then the craton was subjected to uplift accompanied by prolonged weathering and erosion. This period of erosion resulted in the formation of a regolith which is now preserved beneath outcrops of the basal Huronian Supergroup and it created many topographic irregularities which were important in localizing heavy minerals during deposition of the first sediments.

Age

The age of the Huronian Supergroup is bracketed between 2500 m.y. -- the age of Algoman granites which form the basement for the sediments (Van Schmus, 1965; Wanless and others, 1965) and 2160 m.y. -- the age of the Nipissing Diabase which intrudes the sediments (Van Schmus, 1965; Fairbairn and others, 1969). In addition, a date of about 2400 m.y. was obtained from volcanics below the Matinenda Formation (Fairbairn, 1965; Fairbairn and others, 1969) and a date of 2288 was obtained from the Gowganda Formation (Fairbairn and others, 1969). Thus, the time of deposition of the Huronian Supergroup is firmly established as earliest Proterozoic 2500 - 2150 m.y. Roscoe (1973) believed that deposition took place in the early part of this interval (2450 - 2250) whereas Van Schmus (1976) believed that deposition was slightly later (2400 - 2150).

Distribution and Structure

Metasedimentary rocks of the Huronian Supergroup crop out in a 440 km long belt which contains three segments, shown in Figure 2.2. The western segment extends from Sault Ste. Marie to the Elliot Lake Area. In the western part of this segment the Huronian rocks form a block faulted, south-dipping homoclinal succession and in the eastern part, they are folded into two broad, open folds: the Chiblow anticline and the Quirke syncline (Roscoe, 1969). The central segment extends from the Pronto Mine to Wanapitei Lake. This segment is more tightly folded than the western segment and has numerous folds with steeply dipping and overturned limbs. The northeastern segment extends from Lake Wanapitei to Noranda and is referred to as the Cobalt Embayment (Douglas, 1974). This is the largest of the three segments and contains northerly trending folds with dips less than 25 degrees (Roscoe, 1969).

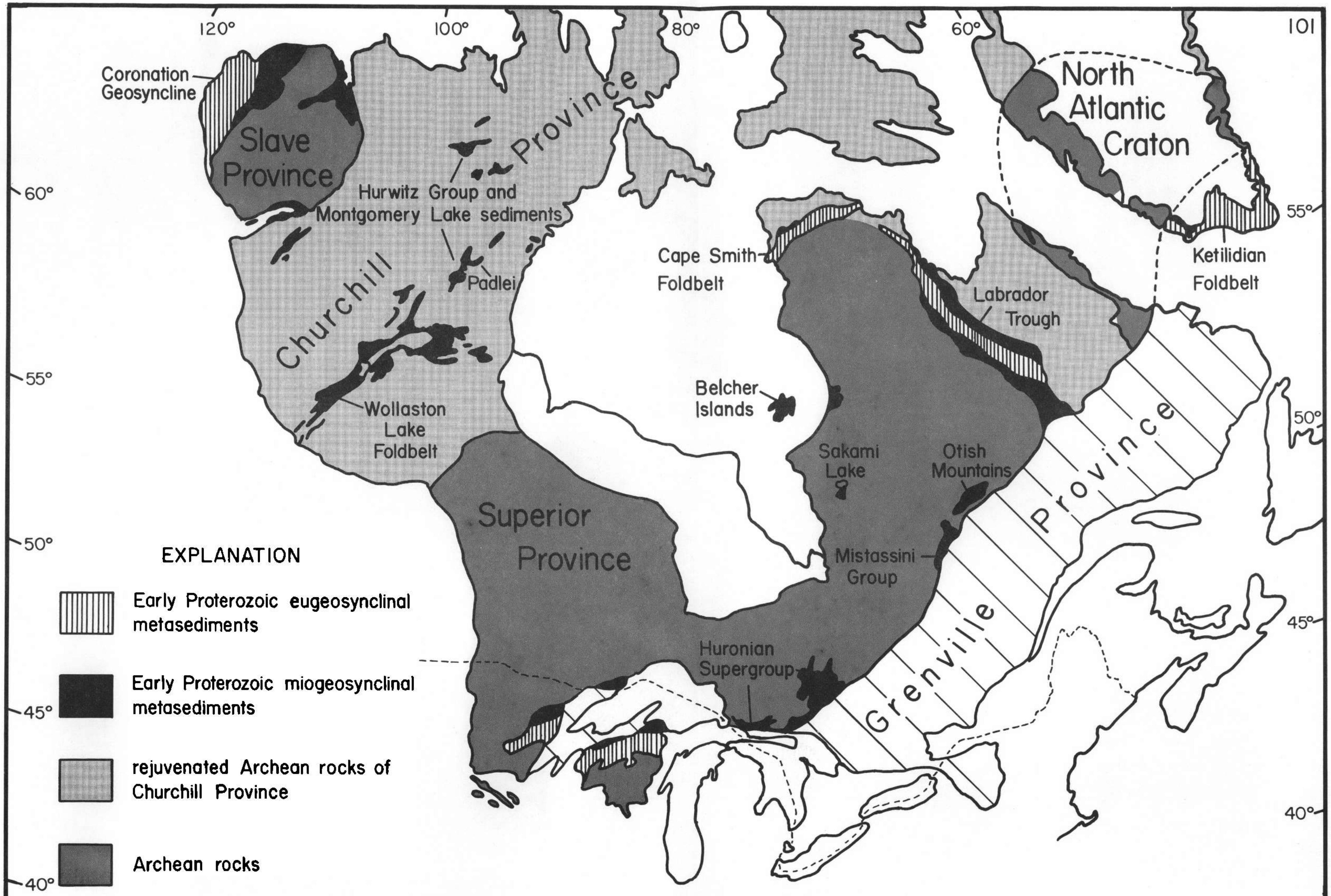

FIGURE 2.1. Index map of the Canadian Shield showing locations of Early Proterozoic metasediments

Economic concentrations of uranium occur in conglomerates of the lower Huronian Supergroup in the western and central segments, near Pronto mine, Elliot Lake, Quirke Lake, and Agnew Lake (Figure 2.2). There are also some radioactive conglomerates in the Cobalt Embayment but these are in the upper Huronian Supergroup and are not of economic importance.

Major east-west trending reverse faults are found in the western and central segments. The most important are the Murray and Flack Lake Faults (Figure 2.2) which are located in sedimentary hinge zones characterized by rapid facies changes (Robertson, 1976). These faults now form the boundaries of the area which contains economically important uranium deposits. Roscoe (1969) considered these faults to be reactivated older (Archean) normal faults on which reverse movement took place during folding of the Huronian Supergroup. He suggested that the faults may have been active during deposition of the Huronian Supergroup and that syn-depositional tectonic movements may have influenced thickness and facies changes in the Huronian Supergroup. He also suggested that volcanic extrusions along the Murray Fault may have obstructed drainage of the sedimentary platform to the north and could have been important in localizing deposition of uraniferous conglomerates in areas north of the fault.

Archean Regolith

The Huronian Supergroup was deposited nonconformably on an eroded Archean granite-greenstone terrain. In Canada it is possible to study exposures of Archean rocks that are believed to represent the weathered surface upon which the Proterozoic sedimentary rocks were deposited. Contacts between Archean and Proterozoic rocks have been studied as well as Archean rock surfaces where Proterozoic rocks have been removed locally. These contacts and surfaces were recognized as paleosol by Collins (1925) and they

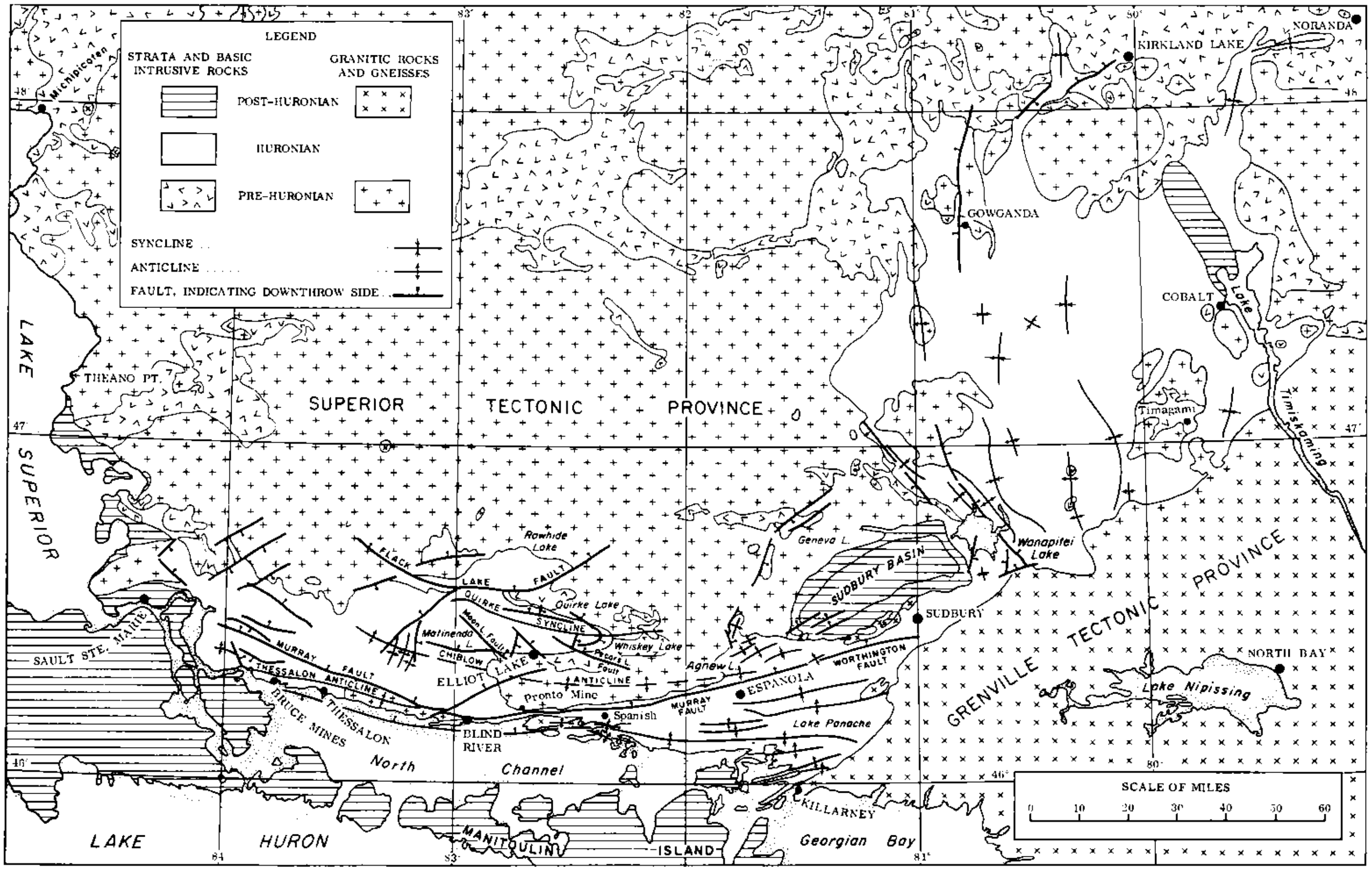

Figure 2.2. Distribution and structure of Huronian rocks. Taken from Roscoe (1969, p. 14). From Geological Survey of Canada, Paper 68-40, Figure 1.

have subsequently been studied and interpreted in the same manner by Roscoe (1957), McDowell (1957), Pienaar (1963) and Robertson (1961, 1966). Roscoe (1969, p. 70-71) gives an excellent description of one of these paleosols, as follows: "granitic rock, for example a pink biotite, granite, grades upward into a white rock with recognizable granitic texture, but containing highly altered mafic minerals and plagioclase almost entirely altered to sericite. Inclusions of unaltered rocks are found in the white-altered granitic material. Higher in the section the granitic texture disappears, microcline is partly replaced by sericite that commonly divides individual grains into fragments. The quartz grains remain similar in shape and size to those in the original granite. The ultimate development is a greenish rock containing quartz grains and relatively few, smaller, irregular grains of microcline 'floating' in a structureless mass of sericite (mica-illite)." Roscoe further states that small grains of hematite, magnetite, pyrite, rutile, zircon, monazite, thorogummite, garnet, and amphibole have been recognized in samples of the granitic saprolite. This ancient soil is similar texturally to that observed today where granites have been weathered except that clay minerals take the place of mica-illite which is undoubtedly a metamorphic mineral.

There are, however, some important differences in chemistry between this Archean soil and modern ones developed on granite. According to Roscoe (1969, p. 72) uranium was not lost during weathering; CaO, SrO_2 and MnO_2 was extensively leached, appreciable amounts of MgO_2, Na_2O, FeO, and Fe_2O_3 were removed, but loss of Fe_2O_3 exceeds that of FeO, and water, Rb_2O_3, and variable amount of K_2O are believed to have been added. The primary difference between this soil and a modern one seems to be the state of oxidation. The increase in FeO/Fe_2O_3 ratio in soil over unaltered rock is the opposite of that in modern soils and under most circumstances uranium would be oxidized and removed in solution and Mn would be oxidized and remain as an oxide resistate in modern soils. This Archean soil study supports the concept of reducing atmosphere during the Late Archean and Early Proterozoic and also indicates that uraninite would be a resistate mineral in the soils of this time (Roscoe, 1969; Pienaar, 1963; and Robertson, 1964).

Stratigraphy and Paleogeography

Rocks of the Huronian Supergroup are divided into four groups which are separated by local unconformities: Elliot Lake Group, Hough Lake Group, Quirke Lake Group, and Cobalt Group (Table 2.1). The Elliot Lake Group contains subaerial volcanics overlain by uranium-bearing conglomerates, quartzites, and argillites. The upper three groups contain paraconglomerate - argillite - quartzite sequences which are interpreted by many workers to be glacial sequences which represent regional climatic fluctuations (Roscoe, 1969; Frarey and Roscoe, 1970; Young, 1973; Roscoe, 1973). By this interpretation the paraconglomerate is considered to be glacial till representing advance of the ice sheet; argillites are interpreted as marine sediments representing marine transgressions following glacial retreat; and quartzites are interpreted as shallow marine to fluvial transgressive sequences which represent uplift of the craton due, in part, to isostatic rebound following ice melt. An alternative interpretation is that the repeating paraconglomerate - argillite - quartzite sequences could be explained in terms of tectonic cycles (Pienaar, 1963; Roscoe, 1969).

Elliot Lake Group

The Elliot Lake Group is a transgressive sequence composed of: the Livingstone Creek Formation (local), basic volcanics (local), the Matinenda Formation, and the McKim Formation (Table 2.1). According to Roscoe (1973),

Group	Formation	Lithology	Thickness (m); distribution	Sedimentary Features	Depositional Environment	Source	Mineralization	Age
Cobalt	Bar River	sandstone	300-1500	cross-beds, ripple marks, mud-cracks, metazoan(?) tracks	shallow marine	N, variable		minimum age of Huranian Super Group is 2150 m.y. which is age of intrusive Nipissing Diabase
	Gordon Lake	reddish siltstone green siltstone reddish siltstone	250-1000	current ripples, slump structures micro cross-beds, dessication cracks	shallow marine	N, variable	Th (monazite) and hematite where un-comformable on Archean rocks	
	Lorrain	sandstone conglomerate arkose	600-2100; fairly uniform	cross-beds, ball and pillow structure, breccias, conglomerate layers	beach and eolian fluvial shallow marine	NNW	trace Th in lower conglomerates	2287 ± 88 from Rb-Sr whole rock analysis
	Gowganda	argillite (sand-stone) paracon-glomerate	150-2200; average about 1000	striated basement, striated clasts, varves, dropstones	glacial in N. glacial-marine in S.	N		
Quirke Lake	Serpent	arkosic sandstone	0-1000; thickens to S.	cross-beds, ripple marks, mudcracks	shallow marine	NW	trace U	
	Espanola	limestone, dolomite, siltstone	0-460; average 150, uniform	breccias, clastic dikes, mudcracks, ripple marks	shallow marine with tectonic disturbance	NW	trace U, Cu in lime-stone at diabase contacts	
	Bruce	paraconglomerate	0-180; average 30, uniform	dropstones, varves	glacial-shallow marine	N		
Hough Lake	Mississagi	coarse subarkosic sandstone	0-3400; thickens to S., thickest near Sudbury	planar cross-beds, troughs, ripple marks, dessication cracks, mud chips	fluvial-braided rivers	WNW in W. NNE in E. N. in S.	U near basement highs; U and trace Au where un-conformable on Archean rocks	
	Pecors	argillite, siltstone	0-500; thickens to S., thickest S. of Esponola	varves, ripple marks, micro cross-beds, slump structures	marine	NNW	trace U near basement highs	
	Ramsey Lake	paraconglomerate	0-180; average 20, thickens to S.	dropstones, crude bedding, mudcracks, cross-beds, ripple marks	glacial-shallow water	NW?	trace U where uncon-formable on Matinenda	
Elliot Lake	McKim	argillite greywacke	0-2150; thickens to S., thickest near Sudbury	ripple marks, small scale cross-beds, quartzite lenses	shallow water-deltaic	NW	trace U near basement highs	
	Matinenda	sandstone arkosic grit oligomictic conglomerate	0-300; thickens to S., absent in Cobalt Embayment	trough cross-beds imbricated pebbles, conglomerate layers, pyritic, uraniferous conglomerate	fluvial-deltaic	NW	U, Th, rare earths, and pyrite in conglomerates in the basement depressions	
	Thessalon Pater, Stobie, Copper Cliff	basalt, andesite (rhyolite, sandstone)	0-1500; local volcanic piles	amygdules, brecciated flows, rare pillows, interlayered conglomerate	subaerial	Flack Lake, Murray Faults	Fe and Cu sulphides in volcanics; U, Th in conglomerate interbeds	questionable 2400 m.y. Rb-Sr date
	Livingstone Creek	arkose, conglomerate	0-500; absent E. of Thessalon, thickest north of Bruce Mines	cross-beds, lenses of conglomerate and argillite	fluvial	NW?	U, Th in conglomerate lenses	
Archean	Algoman Granite	red quartz monzonite, gray granodiorite, quartz diorite					pitchblende veinlets, uraninite in pegmatite	2500 m.y.
	greenstone belts	basalt, andesite, volcaniclastic sediments		pillows, amygdules	marine		iron formation, pyrite	2750-2950 m.y.

Table 2.1. Summary of stratigraphy of the Huronian Supergroup. Adapted from Robertson (1976) and Roscoe (1969).

this group is least extensive of the groups in the Huronian Supergroup and it shows the most rapid and pronounced lateral facies changes. It also contains the principal uranium ores.

The Livingstone Creek Formation contains the oldest exposed Huronian rocks between Sault Ste. Marie and Blind River and it nonconformably overlies Archean basement (Frarey, 1967). The formation contains fine-grained, arkosic to subarkosic quartzite with thin layers of siltstone and uraniferous, pyritic, quartz-pebble conglomerate. Uraniferous conglomerate layers are most commonly found in close proximity to, or actually interlayered with, the overlying basic volcanic rocks (Roscoe, 1969). The matrix of these conglomerates is dark and chloritic and contains appreciable amounts of pyrrhotite and magnetite. According to Roscoe (1969), these conglomerates are of submarginal economic importance because of their low overall grades (<.1 per cent) and lenticular distribution. Roscoe (1969) suggested that the Livingstone Creek Formation, which overlies basic volcanics in the eastern sector, may be a westward extension of the lower Matinenda Formation of the Elliot Lake area, which overlies basic volcanics. If so, the lowest Huronian rocks were deposited during a period in which basic volcanism and rapidly changing fluvial deposition of coarse clastic material were occurring simultaneously. This type of sedimentation marks a transition between the unstable, volcaniclastic deposition in the late Archean and stable, platform-type, miogeosynclinal deposition of the upper Huronian Supergroup.

Volcanic rocks of the Thessalon, Pater, Stobie, and Copper Cliff Formations crop out between Sault Ste. Marie and the Sudbury area and seem to be localized along the Murray and Flack Lake Faults. Robertson (1976) suggested that the flows were extruded in areas of crustal weakness and rapid facies change (i.e. the fault zones) and that their source was a series of gabbroic intrusives which cut the underlying Archean rocks. The volcanic sequences contain mainly massive metabasalt with minor amygdaloidal metabasalt, felsic flows, and felsic tuffs. Geochemical information indicates the volcanics are sub-alkaline tholeiites (Robertson, 1976). According to Roscoe (1969), the flows were very fluid and had extensive flat tops as is evidenced by thin amygdule layers which can be traced for considerable distances and by the presence of laterally extensive interlayers of conglomerate and quartzite in the volcanic sequences. The conglomerate layers are uraniferous and were probably deposited in fluvial channels whose distribution was strongly influenced by the volcanic topography. As a consequence, the conglomerate layers are most commonly found on the margins of the flows (Robertson, 1976). Robertson (1976) also suggested that the volcanism may have contributed sulfur which converted detrital magnetite to pyrite and pyrrhotite in the conglomerates. Efforts to date the felsic phases of the volcanics have yielded a questionable Rb-Sr whole rock age of 2400 m.y. (Fairbairn and others, 1969). Paleomagnetic pole positions from the Thessalon volcanics suggest an age of 2375 m.y. (Symons and O'Leary, 1978).

The Matinenda Formation is the main uranium-ore bearing unit in the Huronian Supergroup (Table 2.1). It crops out in the western segment between the Murray and Flack Lake Faults, and is especially well exposed in the Chiblow Anticline, the Quirke Syncline, and near Pronto (Figure 2.2). It also crops out in the Agnew Lake area of the central segment. Most of the Matinenda Formation is composed of moderately-well sorted, medium-grained subarkose (Roscoe, 1969). However, the uraniferous conglomerate layers are oligomictic quartz-pebble conglomerates and they occur within zones of poorly-sorted, coarse-grained, sericitic, subarkosic grit which have a characteristic greenish color. Roscoe (1969) suggested that these grit zones can be used to identify horizons in the Matinenda which are likely to contain uraniferous oligomictic conglomerates.

Roscoe (1969) divided the Matinenda Formation of the Quirke Syncline into three members: Ryan Member (lowest), Stinson Member, and Manfred Member. The stratigraphic relationship between these three members is shown in Figure 2.3. The Ryan Member is poorly-sorted grit containing radioactive conglomerate layers and is the unit mined in the Nordic zones, on the south limb of the Quirke Syncline. The Stinson Member is a cleaner, better sorted, subarkose which rests on the Ryan Member to the south and on Archean rocks to the north. The Manfred Member is lithologically similar to the Ryan Member and contains the uraniferous conglomerate layers which are mined in the Quirke Zone on the north limb of the Quirke Syncline. As shown in Figure 2.3, the Manfred Member was deposited on the Stinson Member in the central part of the Quirke Syncline and on Archean rocks to the north. According to Robertson (1976), this northward overlap of lithofacies occurs throughout the Matinenda Formation so that uranium-bearing conglomerate horizons get progressively older to the south.

Crossbedding measurements from the Matinenda Formation show that the quartzites and conglomerates were deposited by southwest directed paleocurrents. Most workers consider the paleocurrents to have been part of a fluvial system. (James and Joubin, 1956; Holmes, 1956; McDowell, 1957; Pienaar, 1963; Roscoe, 1969; Robertson, 1976). Evidence for fluvial deposition of the Matinenda is: 1) unimodal, low-variance paleocurrent distribution (McDowell, 1957; Pienaar, 1963); 2) poor sorting and coarse-grain sizes of Matinenda (McDowell, 1957); 3) strong downstream pebble imbrication in conglomerate (Pienaar, 1963), and 4) the interfingering and lenticular distribution of sand and gravel lithofacies shown in Figure 2.3 (Pienaar, 1963).

The pattern of fluvial deposition and placer accumulation of heavy minerals was profoundly influenced by local topographic irregularities in the depositional surface. The two main ore-bodies, the Quirke and Nordic Zones,

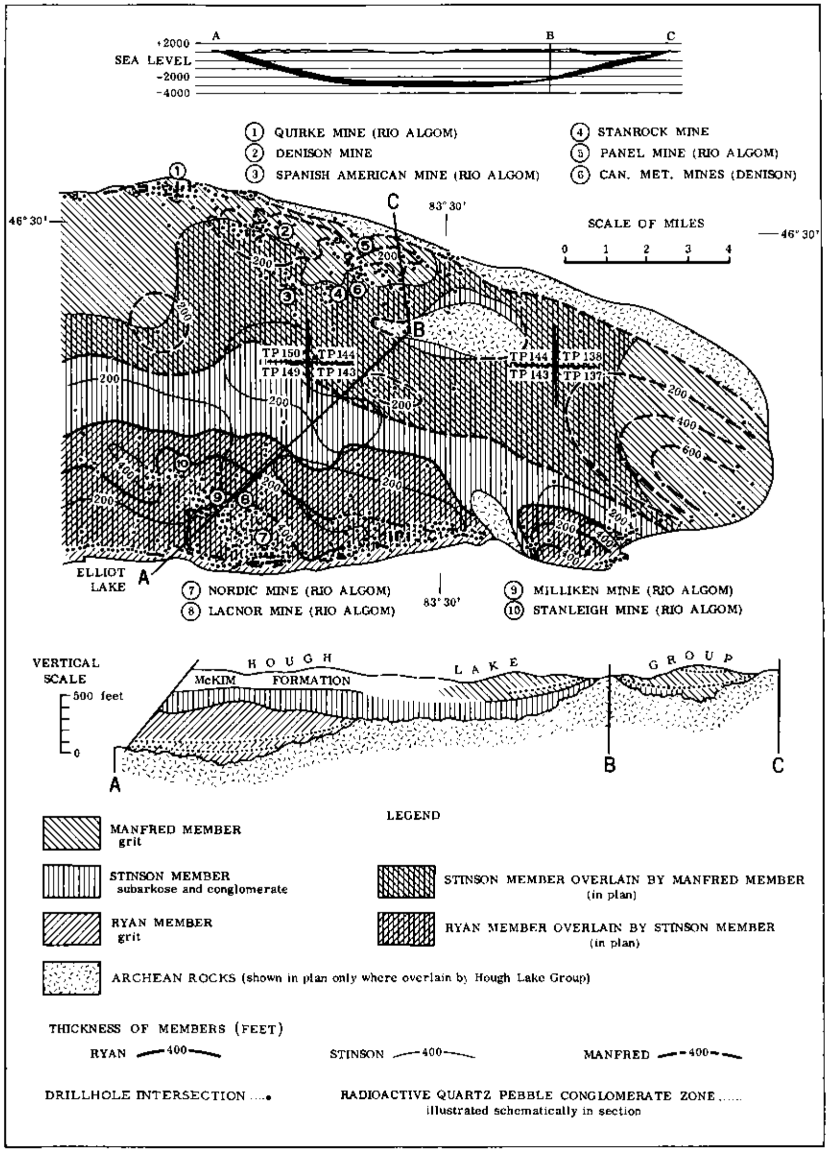

Figure 2.3a. Stratigraphy of the Matinenda Formation near Elliot Lake (from Roscoe, 1969, p. 43). From Geological Survey of Canada, Paper 68-40, Figure 3.

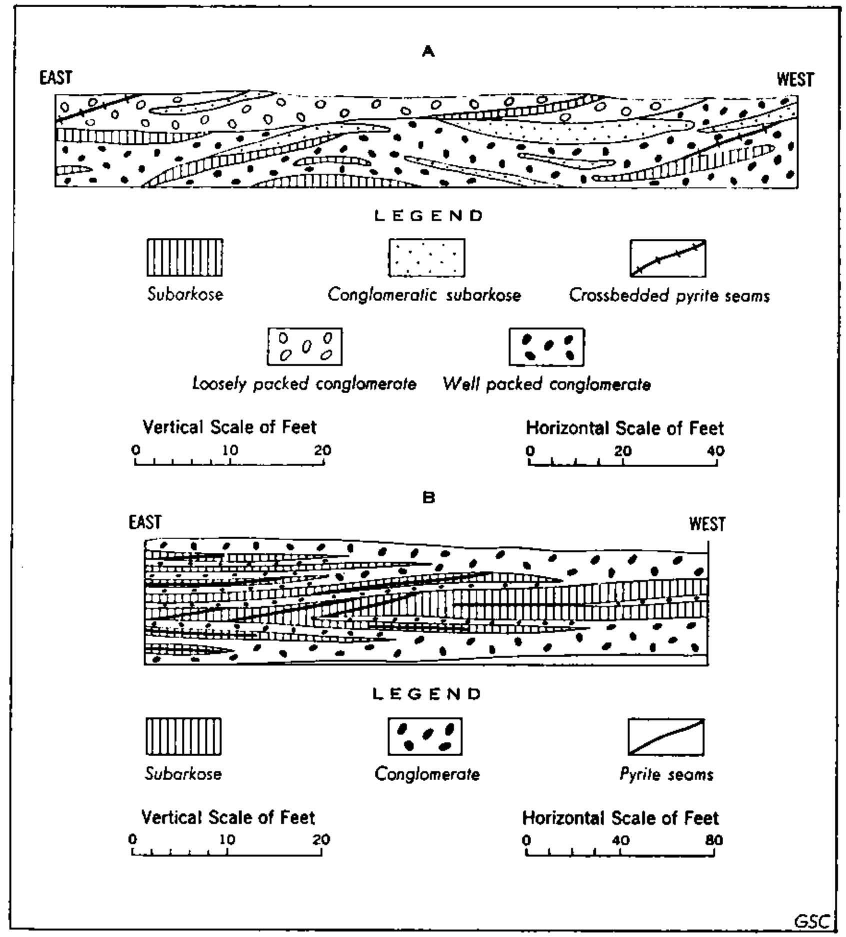

Figure 2.3b. Sections showing facies relationships in uranium-bearing rocks of the Matinenda Formation; A - Quirke Mine., B - Nordic Mine (from Pienaar, 1963, p. 64). From Geological Survey of Canada, Bulletin 83, Figure 11.

are both located in large depressions in the Archean erosion surface caused by preferential erosion of the weaker underlying greenstone belts relative to surrounding granitic rocks. Similarly, the Pronto ore-body is located in a depression in the Archean surface and local uraniferous conglomerates in the Livingstone Creek Formation are located in depressions marginal to volcanic flows. Other evidence for topographic control of conglomerate deposition is that the ore-bodies tend to be thickest in the central (deepest) part of the depressions, where detritus from the eroded basement was deposited by aggrading river systems (McDowell, 1957). Roscoe (1969) estimates that local vertical relief in these depressions was on the order of one foot per 100 feet of horizontal distance.

The McKim Formation crops out between Elliot Lake and Sudbury and contains mainly argillites and siltstones with local layers of coarser, cross-bedded arenite (Roscoe, 1969). Sedimentary structures include thin planar laminations, ripple marks, small-scale crossbedding and mudcracks (Card, 1965) and they indicate shallow water, perhaps deltaic deposition (Roscoe, 1969). This deltaic concept is supported by the presence of arenite intercalations in the argillite and by the fact that the McKim Formation interfingers with fluvial sediments of the Matinenda Formation to the north. To the south and east, the McKim is interlayered with mafic volcanic rocks of the Thessalon, Pater, Stobie, and Copper Cliff Formations. According to Roscoe (1969), the McKim Formation was deposited during marine transgression which began during a period of volcanism and fluvial sedimentation and continued after volcanism ceased.

Hough Lake Group

The Hough Lake Group contains the Ramsey Lake Formation (paraconglomerate), the Pecors Formation (argillite), and the Mississagi Formation (quartzite) which are shown in Table 2.1. Basal paraconglomerate of the Ramsey Lake Formation unconformably overlies rocks of the Elliot Lake Group in the area between Sault Ste. Marie and Lake Winapitei (Figure 2.2) and overlies Archean rocks north of the paleolimit of Elliot Lake Group deposition (Roscoe, 1969). This is evidence for continued northward overlap of sedimentation in the Huronian basin during deposition of the lower Hough Lake Group.

The Ramsey Lake Formation consists of matrix-supported, conglomeratic graywacke with clasts of gray granite, quartz, and mafic igneous rocks in a pyritic, gritty quartzose matrix. The lithology indicates deposition by mass transport mechanisms (Pettijohn, 1957). Robertson (1976) favored glacial transport because of the presence of graded bedding and dropstones in the paraconglomerate. Roscoe (1969) also favored glacial deposition instead of a mudflow interpretation (Pienaar, 1963) because of the wide lateral extent of the paraconglomerate, the lack of evidence of tectonic movements or steep paleoslopes, and the absence of beds of unsorted, immature sediment. The presence of mudcracks, ripple marks, and crossbedding in graywacke lenses in the Ramsey Lake Formation indicate shallow-water deposition, possibly in glacial-related environments.

The Ramsey Lake Formation is radioactive in areas where the paraconglomerate overlies radioactive conglomerates of the Matinenda Formation and it is lighter colored and more potassic where it overlies arkoses of the Matinenda. Robertson (1976) suggested that this is evidence that uranium mineralization in the Matinenda conglomerates predated deposition of the Ramsey Lake Formation and was, therefore, syngenetic.

The Pecors Formation (Table 2.1) conformably overlies the Ramsey Lake Formation and is composed mainly of argillite. The lower Pecors contains possible varved argillite with dropstone clasts which may have been deposited from floating icebergs (Robertson, 1976). Ripple marks, microcrossbedding, mudcracks, and slump structures higher in the Pecors Formation suggest shallow marine deposition. Argillites of the Pecors are locally radioactive (31 ppm U_3O_8; 51 ppm ThO_2) and contain disseminated pyrite and pyrrhotite (Roscoe, 1969).

The Mississagi Formation gradationally overlies the Pecors Formation (Palonen, 1973) and contains gray, coarse-grained subarkosic quartzite with minor subgraywacke and argillite. Adjacent to basement highs in the Elliot Lake area, the quartzite is a green sericitic, gritty and pebbly subarkose which is radioactive. The radioactivity reflects high potassium as well as the presence of uranium and thorium minerals. The uranium content is not high enough to be of economic importance.

Crossbedding is the most abundant sedimentary structure in the Mississagi Formation and is generally planar, although some trough crossbedding is also present. Other sedimentary structures include plane bedding, desiccation cracks, mud-chips, slump structures, and clastic dikes (Long, 1978). Most workers have interpreted the Mississagi Formation to be a fluvial sequence (Collins, 1925; McDowell, 1957; Pienaar, 1963; Young, 1968; Long, 1978). Long (1978) summarized the major lines of evidence supporting a fluvial (braided stream) depositional environment. These include: the abundance of planar and trough crossbedding, a general decrease in grain size and size of bed forms up-section, predominance of unimodal paleocurrent distributions, the abundance of clay size material in the arenites, the general immaturity of the sediment, and the absence of extensive mudstone units, ripple marks, flaser bedding, reactivation surfaces and other features of shallow marine environments.

Paleocurrent studies of crossbedding in the Mississagi Formation have been variously interpreted. McDowell (1957) found a mean paleocurrent of

109 degrees in the Elliot Lake area and Pienaar (1963) found a mean of 143 degrees for the same area. Both authors showed unimodal distributions which they interpreted as evidence for fluvial transport. In contrast, Palonen (1973) favored shallow marine deposition of the Mississagi Formation. He reported bimodal and polymodal paleocurrent distributions in areas south of the Murray Fault and he interpreted these rocks to be intertidal deposits. He also cited cylical upward-coarsening marine stratification sequences and a lack of fluvial features such as channeling and point bar sequences in favor of marine deposition. Long (1978) contradicated Palonen's observations and reported mainly unimodal distributions and fining-upwards sequences south of the Murray Fault. He interpreted his paleocurrent and petrographic data as evidence for two major braided river systems in the Missisagi Formation. The first flowed east-southeast from the Sault Ste. Marie - Elliot Lake area toward Lake Panache where it converged with a river flowing south-southwest off the Cobalt Embayment. The combined rivers flowed south.

The weight of the evidence favors fluvial deposition for the Mississagi Formation and this interpretation also fits the broader picture of marine regression during deposition of the Pecors and Mississagi Formations. Nevertheless, Palonen's concept of shallow marine deposition south of the Murray Fault cannot be ruled out, particularly in light of the rapid thickening and facies change in the Mississagi Formation south of the Murray Fault.

Quirke Lake Group

As shown in Table 2.1, the Quirke Lake Group is composed of the Bruce, Espanola, and Serpent Formations. These formations contain the most distinctive lithologies and best marker units in the Huronian Supergroup

(Roscoe, 1969). The basal formation of the Quirke Lake Group is the Bruce Formation, a paraconglomerate unit similar to the Ramsey Lake Formation, which unconformably overlies quartzites of the Mississagi Formations (Robertson, 1976). This conglomerate contains clasts of gray granite, quartzite, and greenstone in a moderately-sorted, coarse-grained graywacke matrix which is slightly pyritic (Robertson, 1976). Various workers (Collins, 1925; Pienaar, 1963; Cassyhap, 1966; Roscoe, 1969; Palonen, 1973; Robertson, 1976) have suggested that the Bruce Formation is a tillite. Palonen (1973) considered the Bruce Formation to be a terrigenous glacial till which represents continued marine regression following deposition of the Pecors Formation (shallow marine) and the Mississagi Formation (fluvial). Roscoe (1969) suggested that the Bruce Formation may be a terrigenous till in its lower part and a glacio-marine till in its middle and upper parts. Roscoe (1969) postulated that both the Ramsey Lake and Bruce Formations represent marine regressions caused by widespread glaciations.

The Espanola Formation contains marine sediments which represent a transgression following deposition of the Bruce Formation. In the Elliot Lake area, the Espanola consists of three members: the Bruce limestone, the Espanola graywacke, and the Espanola limestone (ferruginous dolomite). In the central sector, these members are not recognized. Sedimentary structures in the Espanola such as ripple marks, mudcracks, clastic dikes, intraformational breccias, and disharmonic drag folds indicate that the Espanola was deposited in shallow water and then was deformed while still in an unlithified state (Robertson, 1976). According to Robertson (1976), the presence of carbonate and iron in the Espanola Formation are indications of some free oxygen in the atmosphere.

The Serpent Formation is a well-sorted, feldspathic (plagioclase) quartzite with minor silstone and polymictic conglomerate. The quartzite contains crossbedding, ripple marks, and mudcracks (Pienaar, 1963) which indicate shallow water deposition by southeast directed paleocurrents (Robertson, 1976). According to Pienaar (1963), the presence of graywacke and polymictic conglomerate beds suggests intermittent rapid deposition due to local tectonic disturbances.

Cobalt Group

The Cobalt Group consists of the Gowganda, Lorrain, Gordon Lake, and Bar River Formations and it unconformably overlies the Quirke Lake Group (Table 2.1). The Cobalt Group differs markedly from the lower three groups of the Huronian Supergroup in three ways. First, the Cobalt Group is well exposed in the northeast sector, north of Sudbury and this represents a drastic northward increase in size of the Huronian depositional basin. Second, the Cobalt Group is thicker than the other groups and shows less dramatic lateral facies changes between segments or across major faults. Third, the Cobalt Group contains hematitic, red beds and thorium-magnetite heavy mineral assemblages instead of uranium-pyrite assemblages. These differences indicate an increase in the availability of free oxygen in the atmosphere. A fourth point of interest is that the Cobalt Group may contain evidence of rapid climatic change from glacial conditions (Gowganda Formation) to sub-tropical conditions (Lorrain Formation). Young (1973) has suggested that this was a regional climatic change which can be used in lithostratigraphic correlation of Early Proterozoic metasediments in North America.

The Gowganda Formation lies unconformably on rocks ranging from Archean granitic rocks to Quirke Lake Group rocks and contains a heterogeneous assemblage of polymictic paraconglomerate, graywacke, arkosic quartzite, and argillite. The lower Gowganda contains thick conglomerate layers, whereas the upper Gowganda is mainly argillite. Paraconglomerates of the Gowganda are distinct from the Ramsey Lake and Bruce paraconglomerates because they contain predominately red granite clasts instead of gray granite and their matrix has a higher Na/K ratio (Robertson, 1976).

The depositional environment of the Gowganda Formation is considered by most workers to have been glacial or periglacial (Collins, 1925; Arnold, 1954; Overshine, 1964, 1965; Cassyhap, 1966, 1968; Lindsey, 1966; Young, 1973). Roscoe (1969) emphasized that only part of the Gowganda is likely to actually be glacial till; the rest being glacial-fluvial, glacio-marine, and even non-glacial mass transport deposits such as mudflows and turbidites. The most convincing evidence for glacial-related deposition of the Gowganda is: 1) striated Archean basement directly below the Gowganda Formation, 2) dropstones in argillites which are interpreted to be ice-rafted pebbles, 3) varved conglomerates and argillites which probably indicate freeze-thaw cycles, 4) striated clasts, and 5) high Na/K ratios which indicate an absence of chemical weathering (Roscoe, 1969). If a glacial depositional environment is accepted, it is still necessary to distinguish terrigenous till from glacio-marine and other periglacial deposits. According to Robertson (1976), densely-packed boulder conglomerates, arkoses and argillites in the Gowganda Formation contain sedimentary features which indicate sub-aqueous deposition. Matrix-supported paraconglomerates, on the other hand, may be either terrigenous or sub-aqueous tillites or non-glacial mudflow or turbidite deposits.

Paleomagnetic evidence for the paleolatitude of deposition of the Gowganda Formation is contradictory. Symons (1975) reported a paleolatitude of 62° for deposition and development of primary magnetization in the Gowganda Formation of the Elliot Lake area, and interpreted this as evidence in favor of glacial deposition at polar latitudes rather than sub-polar deposition. Roy and Lapointe (1976), on the other hand, studied the upper Gowganda of the Cobalt area (Firstbrook Formation) and concluded that its primary remanent magnetization was acquired at paleolatitudes of 30 degrees, which supports low-latitude, possibly non-glacial deposition of the Firstbrook Formation. They also isolated a secondary remanence in the Firstbrook Formation which was introduced at polar latitudes at the time of intrusion of the Nipissing Diabase - 2160 m.y. ago (Van Schmus, 1965; Fairbairn and others, 1969). Perhaps these contradictions can be explained in terms of rapid rates of apparent polar wander during and after deposition of the Cobalt Group, in the interval 2300-2060 m.y.

The Lorrain Formation conformably overlies the Gowganda Formation and is divided by Young (1973) into three members which can be recognized throughout the Huronian basin. The lowest member contains pink, green, and white feldspathic quartzite and siltstone with planar bedding, cross-bedding, and abundant evidence of soft-sediment deformation. The middle member is a cross-bedded, sericitic quartzite containing quartz- and jasper-pebble conglomerates. The upper member is a pure quartz arenite, with minor quartz-pebble conglomerates, which becomes more mature up-section. Robertson (1976) reports the presence of thorium-rich radioactive conglomerates (Th:U > 10:1) within coarse-grained hematitic arkose of the lower Lorrain in the Elliot Lake area. According to Roscoe (1973), these conglomerates contain black sand-type heavy-mineral assemblages consisting of magnetite, hematite, monazite and zircon. This is a marked change from the pyrite-brannerite (or uraninite) heavy mineral assemblages found lower in the Huronian Supergroup and Roscoe interpreted this change as evidence for an increasing oxygen content of the Precambrian atmosphere.

Aluminous minerals such as kaolinite, sericite, pyrophyllite, diaspore, kyanite, and andalusite are found in the middle and upper Lorrain, generally in non-feldspathic layers. Textural evidence suggests that these minerals are diagenetic (and metamorphic) alteration products of the breakdown of detrital feldspars (Chandler and others, 1969; Young, 1970, 1973; Wood, 1970). This leads to the interpretation that the middle and upper Lorrain are fossil laterites which were extensively weathered in a tropical to subtropical climate (Young, 1973). Young (1973) interpreted the change from glacial deposition of the Gowganda Formation to tropical diagenetic weathering of the Lorrain Formation to be indicative of a very rapid, regional climatic change which can be recognized elsewhere in North America (e.g. southeastern Wyoming and Churchill Province). He suggested that this climatic transition should be considered a time line for lithostratigraphic correlation of Early Proterozoic metasediments in North America.

Young's (1973) interpretation of the depositional environments of the Lorrain Formation supports the proposed climatic transition. He suggested that the Lorrain represents a transition from glacial outwash deposition of the uppermost Gowganda, through shallow marine deposition of the lower Lorrain, to fluvial deposition of coarse conglomerates in the middle Lorrain. Fluvial deposition of the middle Lorrain is supported by a unimodal paleocurrent distribution showing southerly paleocurrents, variable grain size of lenticular sand and gravel lithofacies, and the paucity of shales (Young, 1973). The upper Lorrain is extremely mature, has a unimodal paleocurrent distribution, a bimodal size distribution, and 30-60 cm crossbed sets which Young (1973) considered as evidence that the upper Lorrain was deposited by aeolian currents and reworked by weak fluvial currents.

The Gordon Lake Formation conformably overlies the Lorrain and consists of siltstone, argillite, chertstone, and fine- to medium-grained quartzite which are evidence for marine transgression following deposition of the Lorrain. According to Robertson (1976), the Gordon Lake contains three members: red sandstone and siltstone with gypsum nodules (lowest); dark green siltstone, argillite and sandstone; and reddish siltstone, argillite, and chert. Mudcracks, current ripple marks, microcrossbedding, and slump features indicate deposition in very shallow water (Robertson, 1976).

The Bar River Formation is a massive quartz-arenite with infrequent ferruginous silt layers (Robertson, 1976). Desiccation features on ripple-marked surfaces of the silt layers are segmented and sinuous structures which could be biogenic-metazoans (Hoffman, 1967; Young, 1967; Donaldson, 1967; in Robertson, 1976). The presence of oolites in the ferruginous beds (Wood, 1970), and the sedimentary structures indicate shallow water deposition.

Post-Huronian Events

The end of Huronian sedimentation is marked by folding and faulting of the metasediments, followed by the intrusion of the Nipissing Diabase along pre-existing structures (Card and Pattison, 1973). The Nipissing Diabase is accurately dated at 2160 m.y. (Van Schmus, 1965; Fairbairn and others, 1969) and this places a minimum age on the Huronian Supergroup. Following the diabase intrusions, Huronian metasediments and diabases were metamorphosed under conditions ranging from lower greenschist to lower amphibolite facies (Card and Pattison, 1973). This occurred during the Penokean Orogeny 1700-1900 m.y. ago which metamorphosed Early Proterozoic rocks of the Marquette Range Supergroup in northern Michigan (Cannon, 1973). Deformation and metamorphism of the Elliot Lake uranium deposits was minimal (lower greenschist facies) during the Penokean Orogeny and uranium was not extensively mobilized.

Uranium Deposits

Distribution

The distribution of the uranium-bearing units of the Huronian Supergroup is shown in Figure 2.4. The only deposits of economic importance at the present time (1978) occur where oligomictic conglomerates of the Matinenda Formation unconformably overlie topographic depressions in the Archean basement. This situation is found in the Quirke, Nordic, and Pronto ore zones of the Elliot Lake-Blind River area and in the Agnew Lake area, north of Espanola (Figure 2.4).

The Quirke ore zone is 9700 m long by 2200 m wide (Robertson, 1976) and is located on the north limb of the Quirke Lake Syncline. It contains two ore-bearing reefs separated by barren arkose. The lower one, the Denison Reef, occurs at the base of the Manfred Member and unconformably overlies Archean rocks. The upper reef, the Quirke reef, occurs about 30 m above the Denison Reef. Both reefs are composite units containing intercalated lenses and fingers of conglomerate and quartzite. The Denison reef contains two major uraniferous conglomerate layers 1.8-3.6 m thick separated by 0.6-2.4 m of arkose (Robertson, 1976). The Quirke Reef contains three conglomerate layers with the middle layer being richest in uranium. Uranium ore from the Quirke Reef averages .17 per cent U_3O_8 in the New Quirke Mine and ore from the Denison Reef averages .14 per cent U_3O_8 in the Denison Mine (Robertson, 1976).

The Nordic Zone is 5800 m long by 1600 m wide and is located on the south limb of the Quirke Syncline. Ore occurs in a 30 m thick conglomeratic zone near the base of the Ryan member of the Matinenda Formation (Figure 2.3). This conglomeratic zone contains three reefs separated by barren arkoses. The richest reef, the Main Reef, is about 3 m thick and is laterally

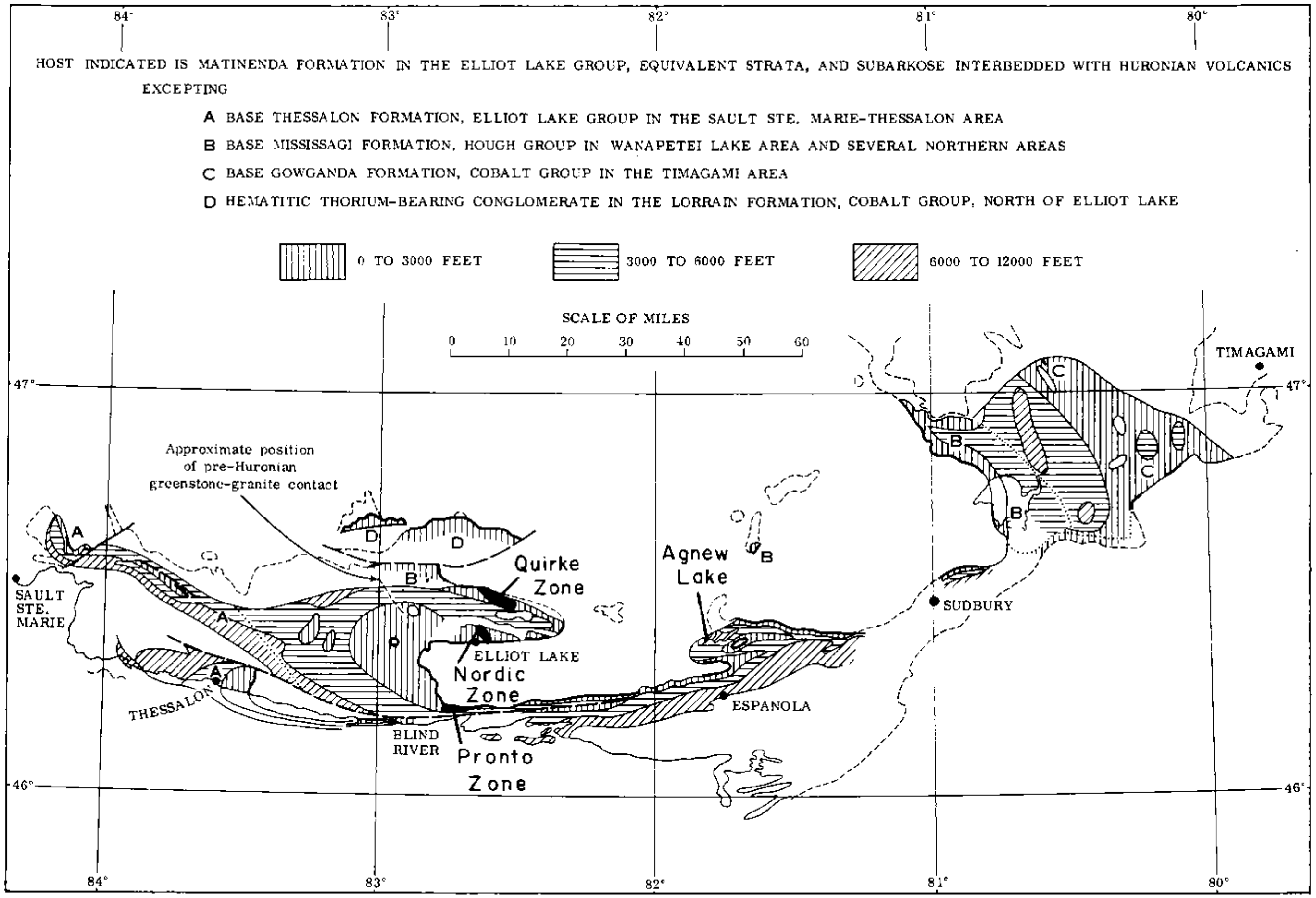

Figure 2.4. Distribution of radioactive rocks in the Huronian Supergroup showing locations of important ore zones (shown in black). Adapted from Roscoe (1969, p. 169). From Geological Survey of Canada, Paper 68-40, Figure 12.

quite continuous (Roscoe, 1969). Average ore grade from the Nordic Zone is .12 per cent U_3O_8 (Roscoe, 1969).

The Pronto Zone is located on the south limb of the Chiblow Anticline. Ores from this zone occur at the base of the Matinenda Formation, within a topographic depression in the underlying weathered granite. The ore conglomerate averages about 2 m thick and is significantly coarser than conglomerates of the ore zones farther north. Average ore grade is .13 per cent U_3O_8. The Pronto Zone dips south and is truncated by the Murray Fault.

Deposits in the Agnew Lake area occur in the lower Matinenda and conglomerates are finer-grained, more deformed, and have a higher Th:U ratio than the Elliot Lake ores. This area has a proven ore-body but, as of 1975, production had not been started (Robertson, 1976).

Other units with anomalously high radioactivity occur in the Huronian Supergroup but are presently uneconomical. The Livingstone Creek and Thessalon Formations contain conglomerate lenses interlayered with volcanic flows and the Mississagi and Lorrain contain radioactive conglomerates at the margins of the Huronian basin, where these units unconformably overlie Archean rocks. Argillite units of the Huronian Supergroup are often high in uranium in areas adjacent to basement highs and paraconglomerates may be uraniferous where they unconformably overlie radioactive conglomerates of the Matinenda Formation (Robertson, 1976).

Lithology and Mineralogy

Ore-bearing conglomerates of the Matinenda Formation are quartz-pebble conglomerates containing: well-rounded and well-sorted quartz pebbles 6-50 mm in diameter, a few chert and jasper pebbles, and rare pebbles of granite, greenstone, and argillite (Roscoe, 1957). The matrix is composed of poorly sorted grains of quartz and feldspar and interstitial sericite, chlorite and epidote. Pyrite also occurs in the matrix and may constitute up to 30 percent of the rock, although average values are closer to 15 percent (Robertson, 1976). Other heavy minerals in the matrix are monazite, zircon, brannerite, uraninite; with occasional anatase, amphibole, barite, apatite, cassiterite, chromite, diopside, garnet, epidote, gold, flourite, hematite, ilmenite, magnetite, rutile, scheelite, sphene, spinel, tourmaline, pyrrhotite, chalcopyrite, arsenopyrite, galena (probably radiogenic), cobaltite, xenotime and yttroflourite (Robertson, 1976).

The most important uranium-bearing minerals in the ore are brannerite, a uranium-titanium oxide, and uraninite, a uranium oxide. Brannerite is found as egg-shaped red-brown to black metamict grains with small inclusions of pyrrhotite and radiogenic galena (Thorpe, 1963; in Robertson, 1976). Ferris and Rudd (1971) and Theis (1973) suggested that the brannerite may have formed at low temperatures during diagenesis of the conglomerates by decomposition of ilmenite and introduction of uranium from solution. Uraninite occurs as black subhedral to rounded grains which are usually about 1 mm in diameter (Robertson, 1976). The uraninite has a high uranium to thorium ratio (10:1) and this suggests that it was derived from pegmatites instead of from hydrothermal uraninite veins (Robertson, 1976). Monazite also contains relatively high uranium concentrations, U_3O_8: ThO_2 = 1:4 (Thorpe, 1963; in Robertson, 1976) and is abundant enough in the Matinenda conglomerates that Roscoe (1969) considered it an ore mineral.

Pyrite grains have widely varying morphologies, even on the scale of a thin section. Grains range from anhedral grains 1/2 mm in diameter to euhedral grains 6 mm in diameter. Also, many pyrite grains are rounded to well-rounded. Roscoe (1969) considered the wide variation in morphology to indicate that the pyrite has several origins. Most of the pyrite was probably originally detrital grains .3-.5 mm in diameter (Roscoe, 1969).

However, some of the pyrite may have originated by secondary sulphidization of detrital magnetite and hematite (Arnold, 1954; Bottril, 1973; in Robertson, 1976), and some may have formed authigenically. In addition, much of the pyrite has been recrystallized during metamorphism.

Thucholite has been reported both from thin seams along bedding in the high-grade conglomerates and from secondary fractures. Ruzicka and Steacy (1976, in Robertson, 1976) propose a biogenic origin for the thucholite in the conglomerates and consider it to be primary.

Genesis

Most workers now consider the Blind River-Elliot Lake ore deposits to be fossil placer heavy mineral accumulations which were deposited in a fluvial environment under reducing conditions (Abraham, 1953; McDowell, 1957, 1963; Pienaar, 1963; Roscoe, 1969, 1973; D. S. Robertson, 1974; J. A. Robertson, 1961-1976). This syngenetic view of the genesis of the mineralization is overwhelmingly supported by detailed sedimentological, stratigraphic, and mineralogic evidence (Roscoe, 1969; Robertson, 1976). A modified view of this placer hypothesis was proposed by Derry (1960) and Joubin (1960) who suggested that uranium was carried in solution by the same streams that deposited the detritus so that some uranium could have been precipitated syngenetically or diagenetically, possibly by organic material. This idea is difficult to verify but should be seriously considered in view of work by Simpson and Bowles (1977) in the Witwatersrand which suggests that some of the uraninite was precipitated from solution. Also, the presence of brannerite may indicate that uranium-bearing solutions were present during and after deposition of the conglomerates. Nevertheless, most of the uranium minerals in the conglomerates were probably detrital so an understanding of the genesis of the mineralization involves understanding the provenance of the conglomerates, the method of transport of the uranium (paleogeographic setting and paleocurrent studies), and the depositional environment of the conglomerates.

Paleocurrent studies indicate that the provenance of the ore-bearing conglomerates was to the northwest and consisted of Archean granites and greenstones. There are two types of granites in the Archean terrain. The first type is non-radioactive gray granodiorite and the second is red quartz monzonite. The red quartz monzonite is slightly radioactive throughout and contains up to 100 ppm ThO_2 and 30 ppm U_3O_8 locally (Roscoe and Steacy, 1958). In some areas north and northwest of Elliot Lake, this quartz monzonite carries pitchblende coated joints or is associated with uraninite-bearing pegmatite veins (Robertson, 1976). This red quartz monzonite and associated pegmatite veins may well have been the source for detrital uraninite in the conglomerates. The absence of gold occurrences in the Archean greenstone belts north and northwest of Elliot Lake is reflected in the low gold content of the Huronian sediments.

The paleogeographic setting prior to deposition of the Huronian conglomerates involved deep weathering and erosion of the Archean source area and development of a system of southerly flowing rivers and streams. Roscoe (1969) studied the weathering of the Archean surface by examining the mineralogy and geochemistry of the preserved regolith. He concluded that the pre-Huronian surface was covered by a residual detrital debris formed by deep chemical and mechanical weathering of the Archean rocks. It was this debris which was the source of the detritus which formed the ore-bearing conglomerates and this explains the feldspathic and sericitic nature of the uraniferous conglomerate and surrounding arkoses. Detrital uraninite (and brannerite) was also apparently derived from this detrital debris, as is evidenced by the fact that the uranium content of the regolith is about the same as the uranium content of the granite from which it was derived (Roscoe and Steacy, 1958). This indicates that uranium minerals were resistates which were not dissolved by ground water. The detrital debris on the Archean surface was an extremely important factor in terms of the formation of uranium-bearing conglomerates, and it is significant that every formation of the Huronian Supergroup which unconformably overlies Archean basement contains some uranium mineralization.

The Archean erosion surface contained numerous, northwest-southeast (Pienaar, 1963) or north-south (Roscoe, 1969) trending ridges and valleys prior to deposition of the Huronian Supergroup. These depressions were the major factor which controlled facies and thickness changes during deposition of the Matinenda Formation (Roscoe, 1969).

Fluvial transport and deposition of the detrital debris from the Archean source area were the major mechanisms of concentration of uranium minerals. During deposition of the Matinenda Formation, detritus was swept into streams and rivers which followed the local depressions. Placer accumulation of the coarse and heavy materials took place in the deepest parts of these depressions. The major ore zones are located in the deepest parts of the basement depressions and were probably deposited by aggrading streams. This type of deposition would deposit gravel bodies with lateral dimensions equal to the width of the meander system (McDowell, 1957) and could account for the blanket type deposits of the ore zones. The presence of uraninite as a detrital mineral in the conglomerate indicates that weathering, fluvial transport, and placer deposition took place under reducing conditions, possibly in a frigid climate, and is evidence for a low oxygen Early Proterozoic atmosphere.

Production and Reserves

Between the time of early uranium production from the Pronto Mine in 1955 and 1973, 12 mines in the Blind River-Elliot Lake area produced about 1.5 billion dollars worth of uranium (something in excess of 90,000 short tons of U_3O_8) with an average grade of 2 pounds U_3O_8 per short ton (.1 per cent). Two mines, the Denison and New Quirke Mines, continue to operate and produce about 4000 short tons U_3O_8 per year (Robertson, 1976).

Estimated reserves as of 1973 were 200,000 short tons of U_3O_8 with a grade of greater than 1.8 pounds per ton (.09 per cent) and an additional 150,000 short tons at grades between 1.4-1.8 (.07-.09 per cent) pounds per ton. The Blind River-Elliot Lake conglomerates also contain thorium reserves of greater than 100,000 short tons ThO_2. These reserves constitute about 75 per cent of Canada's uranium reserves and about 17 per cent of the free world's recoverable reserves (Robertson, 1976).

NORTHERN QUEBEC

Otish Mountains-Mistassini Lake Areas

Early Proterozoic metasedimentary rocks crop out in two areas along the south edge of the Superior Province in Quebec (Figure 2.1). The northern area is located in the Otish Mountains and the southern area is located near Mistassini Lake. The stratigraphy of the Early Proterozoic metasediments is shown in Table 2.2. The age of these metasediments is bracketed between the 2600 m.y. old basement and 1700-1800 m.y. metamorphic ages reported both from the upper Mistassini Group (Fryer, 1971) and from micas in the Archean regolith (Krough, *in* Chown and Caty, 1973).

The oldest metasedimentary rocks are conglomerates and sandstones of the lower Otish Group and the Papaskwasati Formation which lie unconformably on a regolith developed on Archean gneisses. According to Chown and Caty

(1973), this regolith is characterized by weathered feldspars and the absence of Fe-Mg minerals and appears to be thickest over basement highs and thinnest in basement troughs. A basal polymictic paraconglomerate locally overlies the regolith and its composition varies with the underlying basement (Chown and Caty, 1973). The most common type of conglomerate, however, is an oligomictic, quartz-pebble conglomerate which is thickest and coarsest in NE-SW trending valleys in the Archean basement. In the Papaskwasati Formation, those oligomictic conglomerates are greenish and radioactive; they contain abundant sericite, little hematite, and no reported pyrite (Roscoe, 1969). Oligomictic conglomerates of the lower Otish Group are also locally radioactive, but the conglomerates contain iron-oxides and the radioactivity results from a high thorium content related to detrital monazite. According to Roscoe (1973), the Otish Group and Papaskwasati Formation may be correlatives of the Lorrain Formation of the Huronian Supergroup. This correlation is based on the presence of magnetite- and monazite-dominated heavy mineral assemblages in the Otish conglomerates and the fact that the quartzite - conglomerate successions underlie the oldest red-beds in the area (Peribonca Formation) and probably overlie glacial units of the Chibougamau Group, which is correlated with the Gowganda Formation (Morris, 1977). This is similar to the Lorrain, which underlies red-beds of the Gordon Lake and Bar River Formations and overlies glacial units of the Gowganda Formation.

The Mistassini Group overlies the conglomerate - quartzite succession in the Mistassini Lake area. Chown and Caty (1973) suggested that the Mistassini Group is a deeper water facies of the Papaskwasati Formation and that both successions are therefore contemporaneous. However, Roscoe (1969) stated that the Papaskwasati Formation is folded into open, N-S trending folds whereas the Mistassini Group is flat lying and dips gently south

Otish Mountains Area					Mistassini Lake Area			
	Name (thickness)	Lithology	Source	Depositional Environment	Name (thickness)	Lithology	Source	Depositional Environment
	Kaniapiskau Supergroup	Matonipi iron formation red siltstone		marine	Mistassini Group (>330 m)	iron formation carbonaceous shale stromatolitic dolomite		marine
	?					?		
OTISH GROUP	upper Peribonca Formation (120 m)	red arkose, sandstone	NE	shallow marine or fluvial	Cheno Formation (150 m)	dolomite sandstone sandy dolomite		shallow marine
OTISH GROUP	lower Peribonca Formation (220 m)	stromatilitic dolomite arkose	ENE	marine shallow marine				
OTISH GROUP	Indicator Formation (300-800 m)	arkose	SSW?	fluvial	Papaskwasati Formation (500 m)	arkose	NE	
OTISH GROUP		arkosic conglomerate	NNE	fluvial		arkosic conglomerate		fluvial
OTISH GROUP		quartz pebble conglomerate	NE	fluvial		quartz-pebble conglomerate		fluvial
						?		
	Archean gneisses				Chibougamau Group	tillite-like paraconglomerate		glacial
					Archean gneisses			

Table 2.2. Stratigraphy and correlation of Early Proterozoic metasedimentary rocks of the Otish Mountains and Mistassini Lake areas. Adapted from Roscoe (1973) and Chown and Caty (1973).

(Nielson, 1951; Chown, 1960; in Roscoe, 1969). This suggests that the Mistassini Group unconformably overlies the Papaskwasati Formation and implies that the Mistassini Group may be significantly younger. Roscoe (1973) correlates the Mistassini with the Animikie Group of the Lake Superior area and iron formations found in the Labrador Trough. Roscoe's interpretation appears to explain the contrasting lithologies more satisfactorily than that of Chown and Caty.

Roscoe's (1969) description of the rocks in the Mistassini-Otish areas suggests that the potential for uraniferous conglomerates is limited. The paleogeographic setting characterized by a deeply weathered and dissected Archean erosion surface and by early, fluvial sedimentation of quartz-pebble conglomerates is favorable for deposition of uraniferous conglomerates. However, it appears that the conglomerates are too young. The Otish Group conglomerates contain magnetite and monazite instead of pyrite and uraninite suggesting that deposition took place in the presence of sufficient free-oxygen to make detrital pyrite and uraninite unstable. The conglomerates of the Papaskwasati Formation are interpreted by Chown and Caty to be correlative with the lower Otish Group, so they are probably also too young. The potential for gold in the conglomerates is also low because there are no important gold deposits in the Archean source area. However, the conglomerates could contain appreciable thorium reserves.

Sakami Lake

Uraniferous and pyritic quartz-pebble conglomerates crop out west of Sakami Lake (Figure 2.1) and are described by D. S. Robertson (1974). They unconformably overlie a tightly folded Archean greenstone succession and appear to have been metamorphosed during the Kenoran Orogeny - 2500 m.y. ago (D. S. Robertson, 1974). This suggests that the conglomerates are between 2750 and 2500 m.y. old. The conglomerates are unconformably overlain by red-beds of the Sakami Formation.

The conglomerates occur in thin beds within a succession of coarse-grained, yellowish-green to gray quartzites. The pebbles in the conglomerate tend to be small. According to Robertson (1974), heavy minerals include pyrite (up to 10 percent), uraninite (with 10 percent ThO_2), and a thorium silicate (thorite or allanite). Robertson (1974) interpreted these conglomerates to be similar in age and depositional environment to conglomerates of the Huronian Supergroup.

GRENVILLE PROVINCE

The Grenville Province is a northeast-trending zone of highly deformed and metamorphosed Archean and Proterozoic rocks which forms the southeastern boundary of the Superior Province. In the Lake Huron area, the Grenville Province is the southeast boundary of recognizable Huronian rocks and may include highly metamorphosed and deformed equivalents of the Huronian Supergroup (Quirke and Collins, 1930; Frarey and Cannon, 1969). No uraniferous conglomerates are present in the Grenville Province because of the high degree of deformation and metamorphism. However, it does contain important hydrothermal and metasomatic uranium deposits, especially toward the southwest (Ruzicka, 1971). In particular, uranium deposits occur along the northeast margin of Lake Huron and in the Bancroft area (Figure 2.1). These deposits occur in reworked granitic pegmatites within an area of reworked Early Proterozoic (possibly Huronian?) rocks and they are located within several hundred km of the area of outcrop of the Huronian Supergroup. This leads to the speculation that the original source for the uranium in the southwestern Grenville Province could have been low-grade Early Proterozoic rocks, similar to those found in the Huronian Supergroup. The uranium could easily have been remobilized and reconcentrated during the Grenville orogeny. The spatial relationship of epigenetic uranium deposits with uranium-bearing Early Proterozoic rocks in this and other areas (e.g. Wollaston Lake fold belt, northern Australia) may be of importance in understanding uranium districts in many parts of the world.

CHURCHILL PROVINCE

Hurwitz Group and "Montgomery Lake Sediments"

Early Proterozoic metasedimentary rocks crop out in a series of isolated basins in the Churchill Province west of Hudson Bay (Figure 2.1). These rocks include the Hurwitz Group (Wright, 1955) and a pre-Hurwitz Group succession called the "Montgomery Lake Sediments" (Bell, 1970). Both sequences nonconformably overlie an Archean basement which consists of metavolcanic and metasedimentary rocks (Kaminak Group of Davidson, 1970) and intrusive granites dated as 2500 m.y. old (Bell, 1970). Diabase dikes dated as 2330 m.y. (Davidson, 1970) form part of the basement for the Hurwitz Group but their relationship to the "Montgomery Lake Sediments" is not known (Bell, 1970). Hudsonian metamorphism at about 1800 m.y. affected Archean and Early Proterozoic rocks so that the age of the Hurwitz Group is between 2330 and 1800 m.y. and the age of the "Montgomery Lake Sediments" is between 2500 and 1800 m.y. (Table 2.3).

The "Montgomery Lake Sediments" consist of a discontinuous basal paraconglomerate, yellowish-green to gray quartzites, oligomictic conglomerate, and siltstone, which crop out in a small area near Padlei. Pyritic, radioactive quartz-pebble conglomerate from those sediments was described by Heywood and Roscoe (1967) and D. S. Robertson (1974). Heywood and Roscoe (1967) described a 1.3 m thick pyritic conglomerate layer which may occur about 100 m above basement, although stratigraphic relationships are complex. It consists of poorly-packed pebbles of quartz, and minor chert and jasper, about 12 mm in diameter in a sericitic, arkosic matrix containing 5 percent pyrite. One sample contained 13 ppm U_2O_8, 35 ppm ThO_2, and .7 ppb Au (Heywood and Roscoe, 1967). Similar conglomerates up to 2.5 m thick have been intersected by drilling and contain grades up to 260 ppm U_3O_8, 350 ppm ThO_2, and .1 ppm Au over a two foot interval (D. S. Robertson, 1974). According to Robertson (1974), the principal radioactive minerals are uranothorite and zircon. Radioactivity in the conglomerates increases with increasing pebble size and increasing thickness of conglomerate beds. Heywood and Roscoe (1967) and Robertson (1974) suggested a tentative correlation with the lower Huronian Supergroup on the basis of the presence of pyritic, radioactive conglomerate.

The Hurwitz Group unconformably overlies both "Montgomery Lake Sediments" and Archean rocks and has a much wider distribution than the "Montgomery Lake Sediments" (Figure 2.1). The stratigraphy of the Hurwitz Group is shown in Table 2.3. The Padlei Formation contains paraconglomerate, varved mudstone, and dropstones and may be a glacial unit (Bell, 1970; Young, 1973). Young (1973) suggests that it is correlatable with glacial units of the Gowganda Formation and represents tills deposited near the NW margin of an Early Proterozoic ice-cap. The overlying Kinga Formation contains thick aluminous quartzites which are similar to those of the Lorrain Formation (Young, 1973). Units above the Kinga Formation consist of red beds, argillite, graywacke, volcanics, thin iron formation, and quartzite and are considered by Bell (1970) to be younger than the Huronian Supergroup.

The paleogeographic setting for deposition of the Hurwitz Group and pre-Hurwitz sediments is difficult to evaluate because of the extensive remobilization of Archean and Early Proterozoic rocks during the Hudsonian

Group	Formation	Lithology	Thickness (m)	Sedimentary Features	Source	Depositional Environment	Age
Hurwitz	Hurwitz G.	quartzite, dolomitic quartzite, argillite	300-1200	coarsens upward	SW	non-marine	minimum age of 1800 from metamorphic dates
	post-Ameto complex	graywacke, argillite, dolomite	600-1540		?	marine	
	Ameto	volcanics, pillow basalt slate, graywacke	300 600	pillows		marine volcanism and volcanoclastic deposition	
	Kinga	fine-grained quartzite	300	ripple marks thin bedded	SE	shallow marine	
		coarse-grained aluminous quartzite, arkose, grit	300-1200	cross-bedded, thick bedded	SW	shallow marine	
	Padlei	mudstone, siltstone paraconglomerate	<500	varves?	SE?	glacial and glacial-fluvial	maximum age of 2330 from intrusive diabase
"Montgomery Lake Sediments"		quartzite, siltstone pyritic quartz-pebble conglomerate paraconglomerate	0-3000	cross-beds ripple marks	E?	continental-fluvial	1800-2500
Kaminak		Archean metavolcanic and metasedimentary rocks					cut by 2500 my. granites

Table 2.3 Stratigraphy of the Hurwitz Group of the Churchill Province. Adapted from Bell (1970), Young (1973), and D. S. Robertson (1974)

orogeny 1800 m.y. ago. Bell (1970) suggested that the Hurwitz Group was deposited on a "metastable" craton which was unlike the stable Archean craton on which the Huronian Supergroup was deposited. It seems more likely that the general paleogeographic setting characterized by a weathered granitic, Archean craton and early fluvial deposition is analogous to that of the Huronian Supergroup (D. S. Robertson, 1974) and that observed differences in preservation and metamorphism are related to remobilization during the Hudsonian orogeny.

D. S. Robertson considered the potential for important occurrences of uraniferous conglomerate to be small. The conglomerates of the "Montgomery Lake Sediments" appears to be the right age and lithology and to have been deposited in approximately the right paleogeographic setting. However, the conglomerates are poorly-sorted and the uranium content is low. Robertson (1974) suggested that the Archean metasediments in the area may not have been a favorable source for detrital uranium and gold (even though Lord (1953) reported gold in nearby Archean greenstones) and that the poor sorting of the conglomerates indicates that the fluvial system was not well enough developed to concentrate uranium and gold in mature-oligomictic conglomerates. Another severe limitation is the small lateral extent of the conglomerates. They have been largely eroded from the Churchill Province and are preserved in small synclinal remnants. Thus, even if cleaner and higher-grade uraniferous conglomerates could be found, it is questionable that economically significant quantities will be found.

The Hurwitz Group appears to be a bit too young and to have been deposited in the wrong depositional environments under oxidizing conditions. The presence of glacial deposits, aluminous quartzites, and redbeds indicate that the Hurwitz Group may be correlatable with the upper Huronian Supergroup and, hence, too young for important placer uranium occurrences.

Wollaston Lake Foldbelt

The Wollaston Lake foldbelt is a northeast-trending zone of Early Proterozoic metasedimentary rocks located in the central part of the Churchill Province (Figure 2.1). The metasedimentary rocks consist of arkosic quartzite, graywacke, pelitic schists, and metavolcanic rocks which have been folded into upright isoclinal folds and metamorphosed to amphibolite facies under low pressure conditions (Abukuma facies series). They are surrounded by granitic rocks which are mainly remobilized Archean granites but are partly syn-orogenic Hudsonian (1700-1800 m.y.) granites (Money and others, 1970). The age of the metasediments is bracketed between the 2500 m.y. age of the basement rocks and metamorphism of 1700-1800 m.y. (Ray, 1977). In addition Pb isotope data and a two point Rb-Sr isochron on quartzites (Cumming and Scott, 1976) suggest that the metasedimentary rocks are older than 2000 m.y.

Stratigraphic and sedimentological information on the metasedimentary rocks is incomplete because of the high degree of deformation and metamorphism. However, several attempts at stratigraphic subdivision have been made. Money and others (1970) divided the metasedimentary rocks south of Wollaston Lake into three assemblages separated by unconformities: Sandfly Lake Group (immature clastics such as arkose and graywacke); Meyers Group (mature quartzites and quartz-pebble conglomerate); and Daly Lake Group (both mature and immature clastics and calc-silicates). However, work by Gilboy (1975) and Ray (1975, 1976, 1977) in the Wollaston Lake area indicated that the Sandfly Lake and Daly Lake groups are, in part, lateral equivalents and they suggested that the entire metasedimentary assemblage be called the Wollaston Group. The Meyers Group, which crops out in the eastern part of the foldbelt, retains its identity but was considered by Ray (1977) to be a sub-group within the Wollaston Group. We will use the more recent terminology of Ray (1977). The total thickness of the Wollaston Group is 2500-3000 m near Wollaston Lake and may be as much as 11500 m farther north (Weber and others, 1975).

Ray (1977) divided the Wollaston Group into two broad units: a lower, locally graphitic, pelitic unit 200-500 m thick which rests unconformably on Archean basement and an upper unit consisting of thick arkoses 2000-3000 m thick with interbeds of pelite, quartzite, amphibolite, and calcareous metasediments. The lower unit was deposited on a weathered Archean surface during marine transgression and Ray (1977) considered the unit to represent organic-rich shales which were deposited in a stagnant, anaerobic, shallow-marine environment. Pelitic rocks of the lower unit contain uneconomic but widespread U, Cu, Pb, Zn, and Mo mineralization which Ray (1977) considered to be syngenetic. He suggested that the organic-rich anaerobic environment could have been favorable for syngenetic precipitation of uranium from solution. He also suggested that the lower pelitic rocks may have been an important source for the unconformity-type deposits found at Key Lake and Rabbit Lake, Saskatchewan. The upper arkosic unit is a heterogeneous assemblage which probably is mainly of shallow-marine origin (Money and others, 1970). The Meyers sub-Group consists of mature quartzites, oligomictic quartz-pebble conglomerate pelitic rocks, volcanics and calcareous rocks. Money and others (1970) considered the conglomerate to be a beach deposit on the basis of its wide areal extent (25 - 50 km) and they considered the other units to be mainly shallow marine on the basis of their lithologies.

Uraniferous quartz-pebble conglomerates seem to be entirely lacking in the Wollaston Lake foldbelt (Roscoe, personal communication, 1978). Nevertheless, these rocks have several important implications in terms of our model. The Wollaston Group was deposited on a deeply-weathered Archean erosion surface during the Early Proterozoic and the wide areal extent of the lower unit of the Wollaston Group suggests that deposition took place in a large sedimentary basin adjacent to a stable Archean craton. It appears that the paleogeographic setting and the age of the Wollaston Group were favorable for formation of uraniferous conglomerates. Also, the presence of syngenetic uranium in the lower pelitic unit indicates that the source area contained appreciable uranium. However, the depositional environment was wrong. The earliest sedimentation on the Archean surface was in stagnant, shallow basins instead of in fluvial channels. As a consequence, detrital uranium is absent and the only uranium in the sediments appears to have been transported in solution and precipitated by organic materials. Presumably, detrital uraninite was transported and concentrated by fluvial environments in the basin (to the east?) but is no longer preserved.

The presence of major unconformity-type uranium deposits in the Wollaston Lake foldbelt indicates that this area is a uranium district of the first order. It seems probable that the original source for much of the uranium was late Archean granitic rocks. Uranium was then weathered from the Archean terrain and was transported in solution and syngenetically concentrated in the pelitic rocks. Pelitic rocks may have provided an important source for the uranium now found in the unconformity-type deposits.

We would like to emphasize two aspects of this discussion which have a direct bearing on the formation of uraniferous conglomerates. First, deep weathering of an Archean granitic craton appears to have been important in providing a source for the uranium now found in the earliest Proterozoic sedimentary rocks in the Wollaston Lake foldbelt and elsewhere in North America. Second, uranium appears to have been transported in solution and syngenetically precipitated by organic materials in the lower pelites of the Wollaston Group. This indicates that uranium was probably being transported in solution and organically precipitated elsewhere as well and leads to the interpretation that both detrital and aqueous transport of uranium may have contributed to formation of the uraniferous conglomerates.

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AFRICAN PLATFORM

WITWATERSRAND SYSTEM

Geologic Setting

Archean rocks in southern Africa crop out within two ancient cratons: the Kaapvaal craton to the south and the Rhodesian craton to the north (Figure 2.5). Both cratons are granite-greenstone terrains characterized by irregular synclinal greenstone belts containing low-grade volcanics surrounded by large volumes of diapirically emplaced granitic material (Anhaeusser and others, 1969; Anhaeusser, 1973; Hunter, 1974; Kroner, 1976). The cratons are surrounded, or cross-cut, by mobile belts which are characterized by intense deformation, high-grade metamorphism, and granitization. According to Kroner (1976), these mobile belts represent zones of Early Archean sialic crust which were remobilized about 2700 m.y. ago. Kroner also suggested that, prior to 2700 m.y., the Kaapvaal and Rhodesian cratons were continuous and were part of a larger Archean protoshield in southern Africa which contained granitic gneiss, greenstone belts, intrusive granites, and small intracratonic basins filled with quartz-rich sedimentary rocks such as the Pongola, Messina, and Kheis sequences. This protoshield became the source area and depositional site for the sediments of the Witwatersrand system.

The Archean and Early Proterozoic history of southern Africa can be viewed in terms of a progressive evolution of the style of supracrustal deposition; a change from deposition of volcanics and volcaniclastic sediments in the Archean (greater than 2800 m.y.) to deposition of thick sequences of quartz-rich clastic sediments in the Late Archean and Early Proterozoic.

The oldest known rocks in southern Africa are 3600 m.y. old granitic gneisses from the Rhodesian craton (Hawkesworth and others, 1975). According to Kroner (1976), these gneisses prove the existence of a pre-greenstone belt sialic basement in southern Africa.

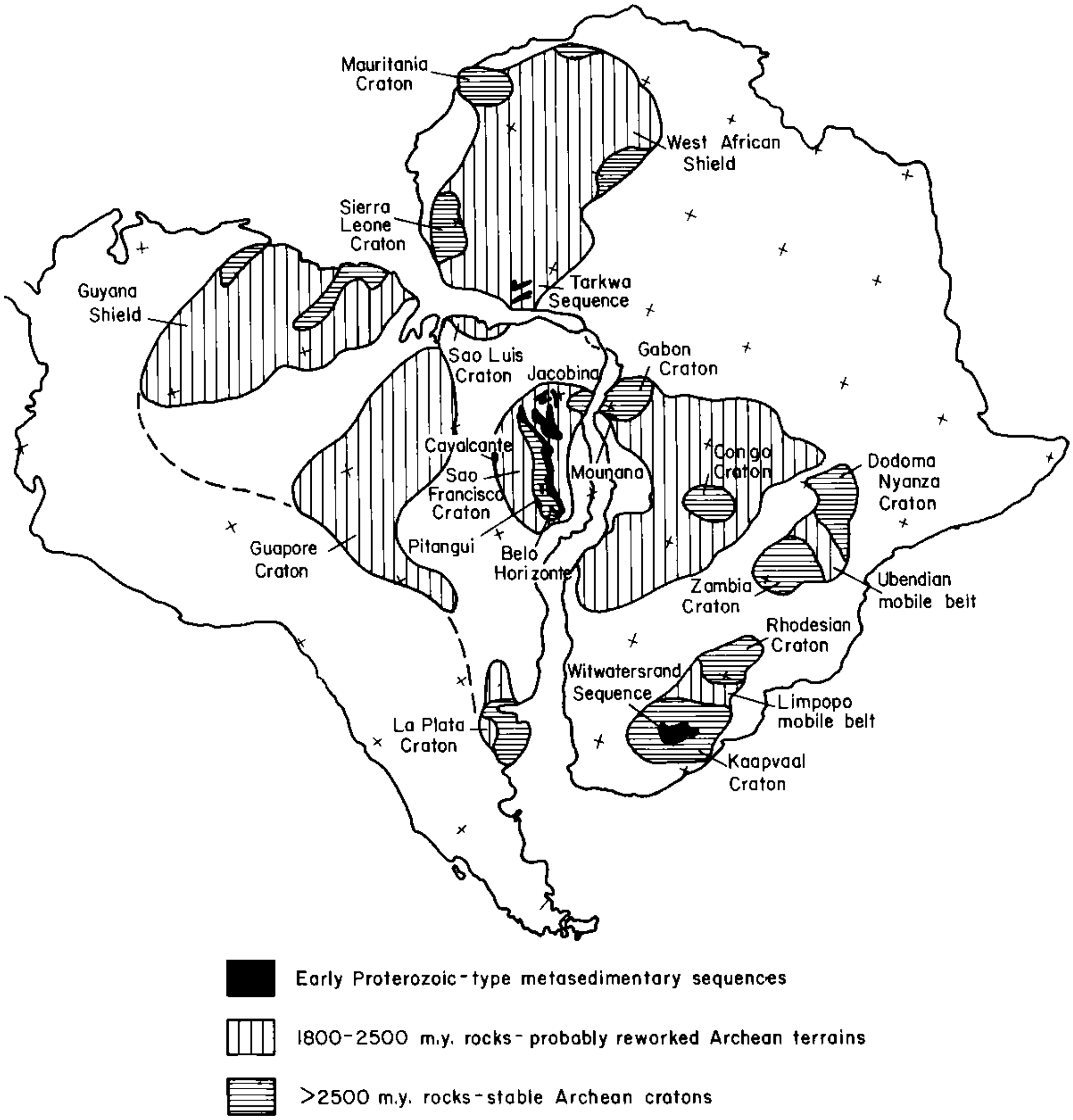

Figure 2.5. Distribution of Archean and Early Proterozoic rocks in Africa and South America, showing locations of Precambrian uranium- and gold-bearing placers; drawn on pre-Mesozoic reconstruction of Bullard and others (1965). Sources of data: Hurley and Rand (1969, 1973); Clifford (1974); Cordani and others (1973); Mason (1973); Anhaeusser (1973); Almeida and others (1973); Pflug and Scholl (1975); Pretorius (1976b); Sestini (1973); Da Costa and Angeiras (1971).

The earliest supracrustal rocks in southern Africa are greenstones of the Swaziland Sequence in the Kaapvaal craton. The lower part of this sequence was deposited about 3500 m.y. ago (Jahn and Shih, 1974) and volcanism and sedimentation continued episodically for at least 300 m.y. Anhaeusser (1973) considered these rocks to be the oldest in the Kaapvaal craton -- a point disputed by Hunter (1974) and Kroner (1976). As shown in Table 2.4, the Swaziland Sequence contains, in ascending order, ultramafic volcanics, mafic volcanics, felsic volcanics, marine volcaniclastic sediments and shallow marine clastic sediments (Anhaeusser and others, 1969; Anhaeusser, 1973). This has become the idealized greenstone belt stratigraphy and it reflects the evolutionary trend toward an increase in deposition of clastic sediments at the expense of volcanics. The Swaziland Sequence was diapirically intruded by tonalites which range in age from 3160 to 3400 m.y. (Anhaeusser, 1973).

The next stage in the sedimentary history of southern Africa was the deposition of quartz-rich clastic sequences in small intracratonic basins which developed on the granite-greenstone terrain. This stage represents a transition from the Archean permobile tectonic regime to the Early Proterozoic tectonic regime characterized by relative crustal stability. One such metasedimentary sequence is the Pongola Sequence of the Kaapvaal craton (Table 2.4). This sequence was deposited about 3100 m.y. ago (Van Eeden, 1972) and consists of volcanics, quartzite, and conglomerate. Some of the conglomerates contain fossil-placer gold deposits. Similar metasedimentary sequences were deposited north and southwest of the Kaapvaal craton and are called the Messina and Kheis Sequences respectively. These rocks were affected by deformation and metamorphism associated with early movements in the Limpopo and Namaqua mobile belts but Kroner (1976) maintained that the nature of these sequences indicates that they were deposited in intracratonic

	Sequence	Group	Lithology	Thickness from type areas (m)	Sedimentary Features	Depositional Environment	Mineralization	Approx. age range (m.y.); data
	Waterberg	Kronsberg						1750-2000;
		Nylstroom	red shale, siltstone, minor conglomerate	6500	regressive sequence	fluvial, shallow marine		$>$ 1790 $\pm$ 70 (U-Pb) based on intrusive granite
	Transvaal	Pretoria	shale, sandstone, diabase, tillite	2400	basal breccia, ripple marks	marine and shallow marine	gold, gold veins	2100-2300; 2224 $\pm$ 21 on andesite from the Pretoria Group; minimum age 1950 $\pm$ 150 from the intrusive Bushveldt Complex
		Dolomite	dolomitic limestone, chert, iron-formation	1520	stromatolites, ripple marks	shallow marine	fine-grained gold, chalcopyrite	
		Black Reef	shale, sandstone, conglomerate	30	transgressive sequence, basal conglomerate	fluvial-deltaic	gold, uranium, osmiridium	
	Ventersdorp		basalt, andesite, volcaniclastics, andesite, quartz porphyry, sandstone, conglomerate	5000	heterogeneous assemblage with rapid lithological variations	subaerial volcanism, marine volcanism, fluvial, eluvial	gold, uranium in basal conglomerate	2250-2500; 2300 $\pm$ 100 (U-Pb on zircons from upper Ventersdorp)
WITWATERSRAND TRAID	Upper Division Witwatersrand Sequence	Klipriviersberg (local)	andesite, pyroclastics, sandstone	3050		subaerial volcanism, high-energy fluvial despoition		
		Kimberly-Elsburg	sandstone, conglomerate, shale	1670	gritty sandstone, coarsening upward	upper mid-fan and fanhead parts of fluvial fan	gold uranium	
		Main-Bird	sandstone, conglomerate	1490	conglomerate reefs in arkosic sandstone	lower, mid-fan part of fluvial fan	major gold, uranium	
	Lower Division Witwatersrand Sequence	Jeppestown	shale, sandstone, lava	1380	amygdaloidal andesite marker bed, banded shale (varves?)	fan-base of fluvial fan — laucustrine system	miner gold, uranium	2500-2750
		Government	sandstone, shale, conglomerate	1970	tillite	fluvial lacustrine	minor gold, uranium	
		Hospital Hill	shale, sandstone, iron-formation	1620	mature sandstone re-worked from Moodies Group	fluvial-lacustrine		
		Dominion	rhyolite, andesite, tuff	2040				
			basic volcanics, minor sandstone	610	amygdaloidal flows, conglomerate in basement depressions	high-energy subaerial environment dominated by volcanism, tectonism		2820 $\pm$ 55 (Rb-Sr) 2800 $\pm$ 60 (U-Pb)
			basal conglomerate, sandstone	100			uranium, gold	
	Pongola	Mozaan	sandstone, conglomerate volcanics	10700	regressive sequence	fluvial, subaerial	gold, pyrite in basal conglomerate	2900-3100 based on intrusive granites
		Insuzi	shale, volcanics, sandstone			fluvial, subaerial	gold, pyrite	
	Swaziland	Moodies	conglomerate, sandstone, shale	3140	four transgressive cycles	shallow water		$>$ 3070 based on intrusive granites
		Fig Tree	greywacke, shale, iron-formation	2150	minor tuff, agglomerate	marine turbidites		
		Upper Onverwacht	mafic to felsic volcanics	7683	flows, pyroclastics, chert	marine volcanism	gold at upper contact, also pyrite, arsenopyrite, stibnite	3370 $\pm$ 20 (Rb-Sr)
		Lower Onverwacht	ultramafic and mafic volcanics	7530	ultramafic flows	marine volcanism	gold at upper contact, also pyrite arsenopyrite, stibnite, and minor copper, nickel, cobalt	3310 $\pm$ 40 (U-Pb); 3500

Table 2.4. Stratigraphy of Archean and Early Proterozoic Supracrustal Rocks in South Africa. Based on Pretorius (1976a), Brock and Pretorius (1964a); Whiteside and others (1976); Anhaeusser (1973); and Button (1976).

basins on an extensive sialic protoshield prior to 2700 m.y. ago. The Pongola Sequence contains about 45 percent clastic sediments and 60 percent volcanics and can be viewed as transitional between the deposition of Archean greenstone belts, where immature volcaniclastic sediments make up 20 percent and volcanics 80 percent of the section and the deposition of the Witwatersrand Sequence, where mature, quartz-rich sediments make up about 80 percent and volcanics 20 percent of the sequence (Anhaeusser, 1973).

There are two features of the Archean sedimentary history of southern Africa which should be emphasized. First, the occurrence of fossil-placer gold deposits in the Archean Pongola Sequence indicates that both Archean and Early Proterozoic quartz-rich clastic sequences can be hosts for fossil-placer gold deposits and that there is no older age limit for fossil-placer gold deposits except that which limits the formation of the quartz-rich clastic sequences (i.e. evolution of crustal stability). Thus, Archean clastic sequences in South Africa, Brazil, India, and elsewhere are potential targets for gold.

Second, the Archean crustal evolution in southern Africa appears to have been precocious relative to that of the Canadian Shield. The greenstone belts are significantly older and so are the early quartz-rich metasedimentary sequences. Also, detrital gold in quartz-pebble conglomerates appeared about 600 m.y. earlier in southern Africa (3100 versus 2500 m.y.) and detrital uranium in conglomerates appeared about 300 m.y. earlier (2800 versus 2500 m.y.). These facts suggest that the Kaapvaal craton of southern Africa, and perhaps other southern hemisphere cratons in the Brazilian and Indian shields were stabilized earlier than the cratons in the northern hemisphere.

Distribution and Structure

The aggregate thickness of Archean and Early Proterozoic quartz-rich metasedimentary rocks in South Africa is greater than 40 km (Table 2.4). These metasedimentary rocks form five sequences which were deposited between 3100 and 1800 m.y., a period of time twice as long as the Phanerozoic Era. These sequences are: the Pongola Sequence (3100-2900 m.y.), the Witwatersrand Sequence (2800-2500 m.y.), the Ventersdorp Sequence (2500-2250 m.y.), the Transvaal Sequence (2300-2100 m.y.), and the Waterberg Sequence (2000-1800 m.y.). These metasedimentary sequences were deposited in separate intracratonic basins which become more extensive with time (Pretorius, 1976b). Figure 2.6 shows that the depositional axes of the basins trend east-northeast and are gently warped about a northwest-trending fold axis. Figure 2.6 also shows that the axis of maximum sedimentation migrated northward with time. This was the result of a persistent tectonic pattern where the northwest side of each basin was fault-bounded and tectonically subsiding whereas the southeast sides were less active. This tectonic pattern caused the northwest side of an earlier basin to be overlapped by sediments deposited on the passive southeastern side of the succeeding basin (Pretorius, 1976b).

The most important of the five sequences is the Witwatersrand sequence because it contains the economically important gold- and uranium-bearing rocks of the Witwatersrand Sequence. This basin is about 320 km long and 160 km wide and contains about 14 km of sediments and volcanics. The basin has the overall configuration of an asymmetrical syncline with the steeper limb on the northwest side. This syncline trends east-northeast and is warped about a northwest-trending fold axis (Figure 2.6). Figure 2.7 shows that the detailed structural pattern of the Witwatersrand basin is characterized by egg carton-like, basin and dome folds resulting from

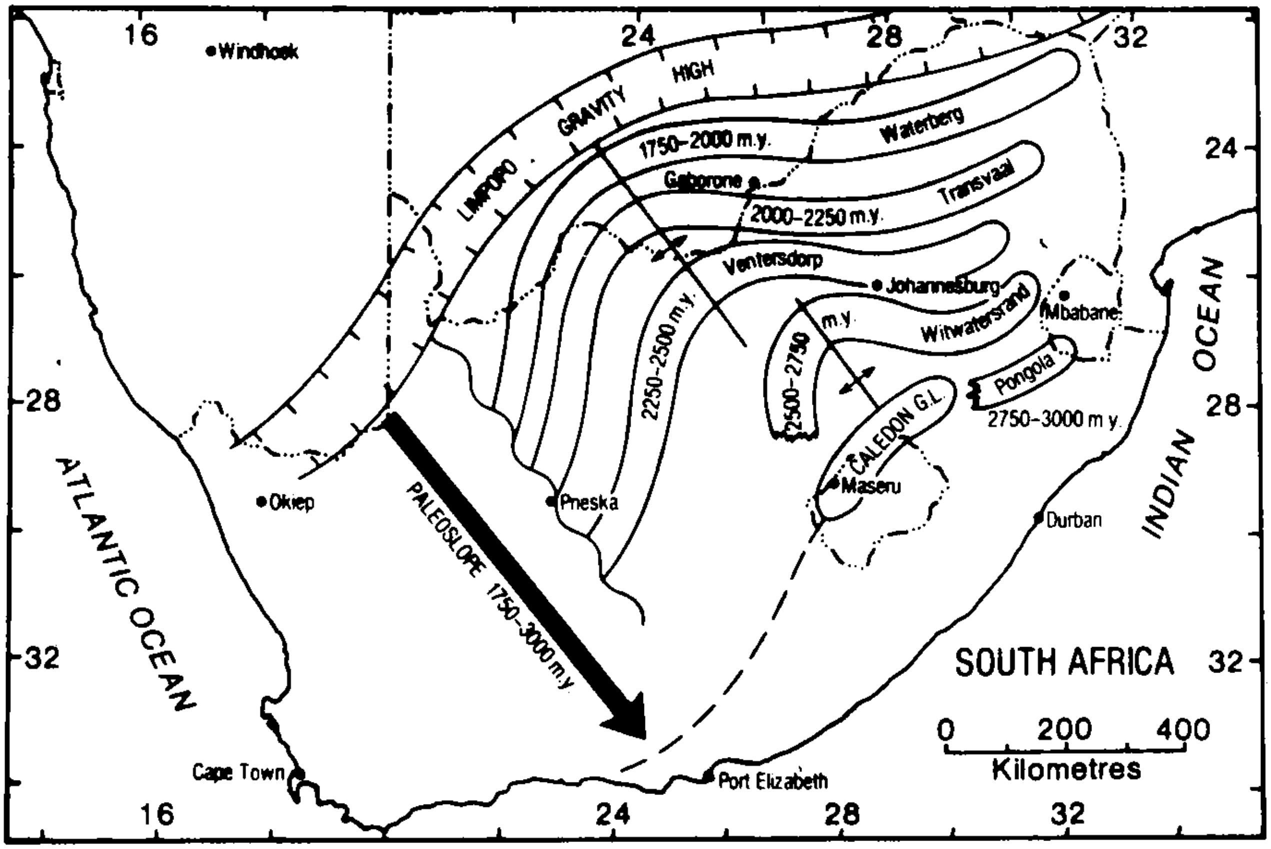

Figure 2.6. The relative positions of the depositional axes of progressively younger Proterozoic sedimentary basins on the Kaapvaal craton in the northeastern part of South Africa. The basins young up the regional paleoslope from the Caledon cavity low towards the Limpopo gravity high. The full extenions of the Pongola and Witwtersrand basins still remain to be determined beneath the Phanerozoic cover. Taken from Pretorius (1976b, p. 34). Reproduced with permission from Elsevier Publishing Company and the author.

interference of the northwest and east-northeast trending fold systems. According to Pretorius (1976b), both of these fold systems were active before, during, and after deposition of the Witwatersrand Sequence and the folding pattern, therefore, exerted a strong influence on the patterns of sedimentation. Topographic basins formed where synclinal traces intersected and these basins became the site of alluvial fans and fan-deltas in which detrital gold and uraninite were concentrated. The basins were surrounded by topographic highs which formed where anticlinal traces intersected. These domes are cored by granitic basement rock (Figure 2.7). One of these domes, the Vredefort Dome, is located in the center of the basin and is surrounded by Witwatersrand sediments. On the north side of the dome, the sediments are overturned and dip toward the dome. This dome was a structural high before and during deposition of the Witwatersrand Sequence as evidenced by gold-bearing conglomerates around its flanks (Brock and Pretorius, 1964b). The dome was also elevated after deposition of the sediments as is evidenced by overturned beds and higher grades of metamorphism around the dome.

Faulting in the Witwatersrand basin was closely related to the patterns and stresses generated by concentric folding. The most important faults are strike faults which parallel the edge of the depositional basin. These are mainly high-angle normal faults with the downthrown side toward the axis of the basin. The faults were active before, during, and after deposition of the Witwatersrand Sequence and were the major mechanism of basin subsidence.

The persistence of the tectonic pattern and the influence of tectonics on sedimentation are both keys to the understanding of the development and distribution of the major goldfields. Apparently, continued folding and faulting created enough relief so that fluvial fan-deltas could form in the basin depressions along the margins of the Witwatersrand basin. These fans

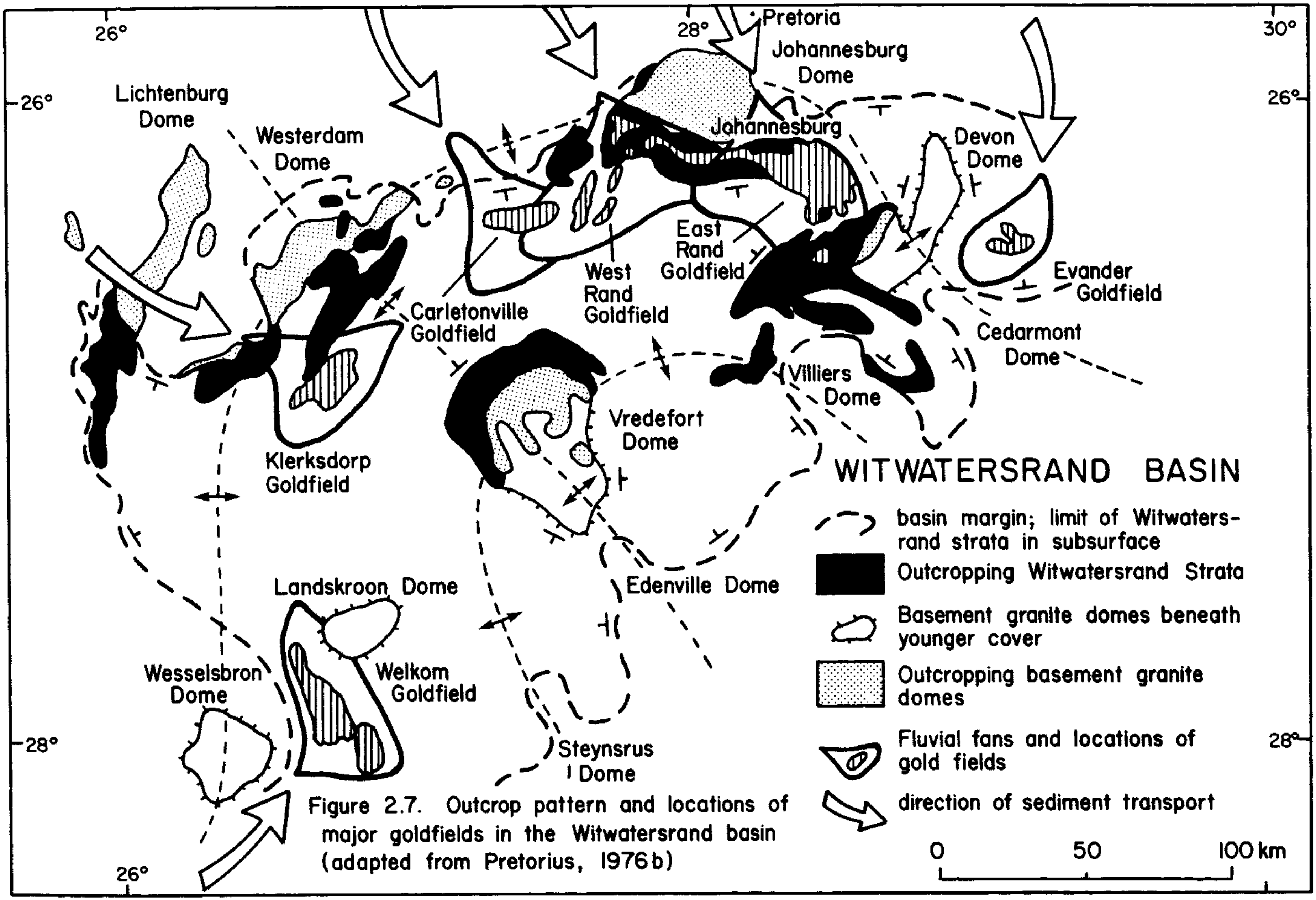

Figure 2.7. Outcrop pattern and locations of major goldfields in the Witwatersrand basin (adapted from Pretorius, 1976b)

were fed by detritus from the surrounding domes, which also supplied gold, uraninite, and other heavy minerals.

Stratigraphy

Witwatersrand Sequence

Dominion Reef Group. The Dominion Reef Group is the basal unit of the Witwatersrand Sequence (Table 2.4). It crops out only in the northwestern and central parts of the Witwatersrand basin and, in both areas, unconformably overlies Archean granites. The type area of the group is in the northwestern part of the basin. In this area, the group consists of three units: a lower clastic unit (100 m) containing quartzites and uraniferous and auriferous conglomerates, a middle unit (610 m) containing amygdaloidal basalts and andesites, and an upper unit (2040 m) containing felsic volcanic rocks (Brock and Pretorius, 1964a; Whiteside, 1970).

The basal clastic unit, where it is fully developed, consists of a basal conglomerate called the Lower Reef, a middle part consisting of medium- to coarse-grained, sericitic, pebbly arkose, and an upper conglomerate called the Upper Reef. The basal conglomerate has been mined for detrital gold and uranium and occurs in lenses which fill local depressions in the underlying basement. Basal conglomerates were not deposited outside basement depressions. Instead, Archean granites appear to grade up into arkosic quartzite indicating that a regolith had developed on the eroded Archean basement.

The upper Reef occurs about 18 m above the Lower Reef and is separated from it by barren, sericitic arkoses which contain lenticular conglomerate layers. The Upper Reef varies from 2 to 120 cm thick and is remarkably persistent throughout the area underlain by the Dominion Reef Group (Von Backstrom, 1976). This reef contains both placer uranium and gold but uranium values are much higher. Grades appear to be highest where the conglomerate is thin and grades as high as 1 percent uranium have been reported from areas where the Upper Reef is 2 to 5 cm thick (Von Backstrom, 1976).

The middle unit of the Dominion Reef Group consists mainly of gray, fine-grained andesite and amygdaloidal andesite. In addition, tuffs and tuff breccias occur throughout the formation at the tops of flows. Other distinctive horizons include a 30 m thick quartzite bed near the base and a quartz-feldspar porphyry higher in the section (Whiteside, 1970).

The upper unit of the Dominion Reef Group conformably overlies the middle unit and consists of rhyolites with subordinate andesites and tuffs. Thick rhyolite sections occur at the top and bottom of the unit and interbedded rhyolite and andesite occur in the middle part (Whiteside, 1970).

Lower Division of the Witwatersrand Sequence. The Lower Division of the Witwatersrand Sequence is a clastic sequence which is distinguished from the underlying Dominion Reef Group by a relative lack of volcanic rocks. Arkosic quartzite and shale are the dominant lithologies but quartz-pebble conglomerates, iron formation, tillite, and quartz-arenite are also present (Pretorius, 1964). In general, rocks of the Lower Division were deposited under transgressive conditions and fine-grained, distal-facies sediments predominate (Whiteside and others, 1976).

Table 2.5 shows that the Lower Division is subdivided into three groups: the Hospital Hill Group, the Government Group, and the Jeppestown Group. This figure also shows the most important marker beds within each group and the main mineralized horizons. The best descriptions of the Lower Division come from the central and eastern parts of the Witwatersrand basin (Pretorius, 1964; Whiteside, 1964; Whiteside and others, 1976) and the following summary is based on these descriptions.

The Hospital Hill Group unconformably overlies both Archean rocks and the Dominion Reef Group. The basal unit of the group is the Orange Grove Quartzite (Table 2.5) which contains light gray to white cross-bedded quartz arenite with a few thin layers of shale and conglomerate. This unit is overlain by a sequence of alternating subgraywacke and shale which includes distinctive marker beds such as iron formation in the Water Tower Slate, the Ripple Marked Quarzite, and deformed iron formation of the Contorted Bed. The uppermost part of the Hospital Hill Group contains quartz-arenite of the Hospital Hill Quartzite.

The Government Group contains fewer shales and more arenites than the other two groups in the Lower Division and consists mainly of arkose, subgraywacke, and conglomerate. The basal units, the Promise Beds, are shales which are split by a thin zone of quartzite and conglomerate called the Promise Reef. The Promise shales are overlain by the Coronation Beds which include, a succession of tillite; Coronation Shale (iron formation); and Coronation Reef (grits). The top of the Government Group is marked by the Government Reef beds, which contain alternating quartzite and shale with a conglomerate layer near the middle called the Government Reef. This reef contains auriferous and uraniferous conglomerates which have been mined in some areas (Pretorius, 1976b) but are usually of sub-economic importance.

The Jeppestown Group consists of shale, argillaceous quartzite, and quartzite with a volcanic marker unit near its middle. This unit, the Jeppestown Amygdaloid, contains amygdaloidal volcanics and volcanic breccia (Whiteside, 1970). The Jeppestown Group also contains a gold- and uranium-producing conglomerate in the Klerksdorp area, called the Jeppestown Reef (Pretorius, 1976b). The Jeppestown Group marks the change from transgressive sequences in Lower Division to regressive sequences in the Upper Division (Whiteside and others, 1976).

Sequence and total thickness (m)	Group and total thickness (m)	Main marker horizons	Thickness of marker horizon (m)	Mineralization
Lower Division Witwatersrand Sequence (4970)	Jeppestown Group (1380)	Jeppestown Amygdaloid	31	
		Jeppestown Reef		Au, u
	Government Group (1970)	Government Reef	2	Au, u
		Coronation Reef	2	
		Coronation (or West Rand) Shales	154	
		Promise Reef	2	
	Hospital Hill Group (1620)	Hospital Hill Quartzites	400	
		Comforted Bed	46	
		Speckled Bed	2	
		Ripple-marked Quartzites	6	
		Water Tower Slates	246	
		Orange Grove Quartzites	169	

Table 2.5. Main marker beds of the Lower Division of the Witwatersrand Sequence. Adapted from Brock and Pretonius (1964a).

Upper Division of the Witwatersrand Sequence. The Upper Division of the Witwatersrand Sequence differs in several important ways from the Lower Division. First, the Upper Division contains the vast majority of gold- and uranium-bearing conglomerates of the Witwatersrand Sequence. Second, the Upper Division is almost devoid of volcanic rocks and, instead, is dominated by continental and shallow water clastic sediments. Third, clastic sediments in the Upper Division are generally coarser and the sand to shale ratio much larger than in the Lower Division. This indicates that higher energy sedimentary conditions prevailed during deposition of the Upper Division.

The most characteristic feature of the Upper Division is the occurrence of auriferous and uraniferous quartz-pebble conglomerates which are developed on local unconformities. These conglomerates occur in relatively thin layers called reefs which appear in several stratigraphic levels within the Upper Division. The conglomerates tend to be richest in gold and uranium where they overlie unconformities (Sharpe, 1949) which suggests that the younger gold-bearing reefs represent sedimentary reworking of older reefs (Whiteside and others, 1976). Cousins (1965) suggested that the Upper Division was deposited during a series of cycles, each of which began with deposition of gold-bearing conglomerate and ended with erosion of the underlying sediments. These high-frequency cycles appear to be superimposed on a lower frequency trend toward overall regression and construction of the basin (Pretorius, 1976b). Evidence for overall regression in the Upper Division is the increasing thickness of conglomerate units up-section (Brock and Pretorius, 1964a). Thick conglomerates are indicative of a higher energy fluvial facies which was closer to the source than the thin conglomerates lower in section and this implies regression (Pretorius, 1976b).

The Upper Division is subdivided into three groups by Pretorius (1976a): the Main-Bird Group, the Kimberley-Elsburg Group, and the Klipriviersberg Group (Table 2.4). The lower two groups contain quartzite and conglomerate, including the main gold-bearing conglomerates of the Witwatersrand basin. In contrast, the upper group contains mainly andesite, tuff, and agglomerate which are often considered to be part of the Ventersdorp Sequence (Whiteside and others, 1976; Brock and Pretorius, 1964a). Figure 2.8 shows a series of stratigraphic columns for the Main-Bird and Kimberley-Elsburg Groups in various areas of the Witwatersrand basin.

The Main-Bird Group is a series of quartzites and conglomerates which occur between the top of the Jeppestown Group and the top of the Kimberley Shale. The quartzites are gray to green, sericitic, feldspathic quartzites consisting of nearly equal abundances of detrital quartz and chert with lesser amounts of muscovite, chlorite, pyrophyllite, and chloritoid (Brock and Pretorius, 1964a). The quartzites commonly contain trough crossbedding and ripple marks.

The Main-Bird Group contains the major gold- and uranium-producing horizons in the Witwatersrand basin. This include sets of conglomerate layers in the Main Reef, Main Reef Leader, Livingstone Reef, and Bird Reef and carbonaceous layers of the Carbon and Vaal Reefs (Figure 2.8). As shown in Figure 2.8, the conglomerate reefs may occur either as individual layers or as groups of more than ten conglomerate layers within the quartzite. About 80 percent of the conglomerate, by weight, is well-rounded pebbles which are set in a matrix of sericitic quartzite. The matrix tends to have less sericite, chlorite and rock fragments than the overlying and underlying quartzites, reflecting higher energy deposition and more extensive reworking. Nearly all the conglomerates contain pyrite, gold, uranium, and other heavy minerals in the matrix but only a small proportion contain mineable

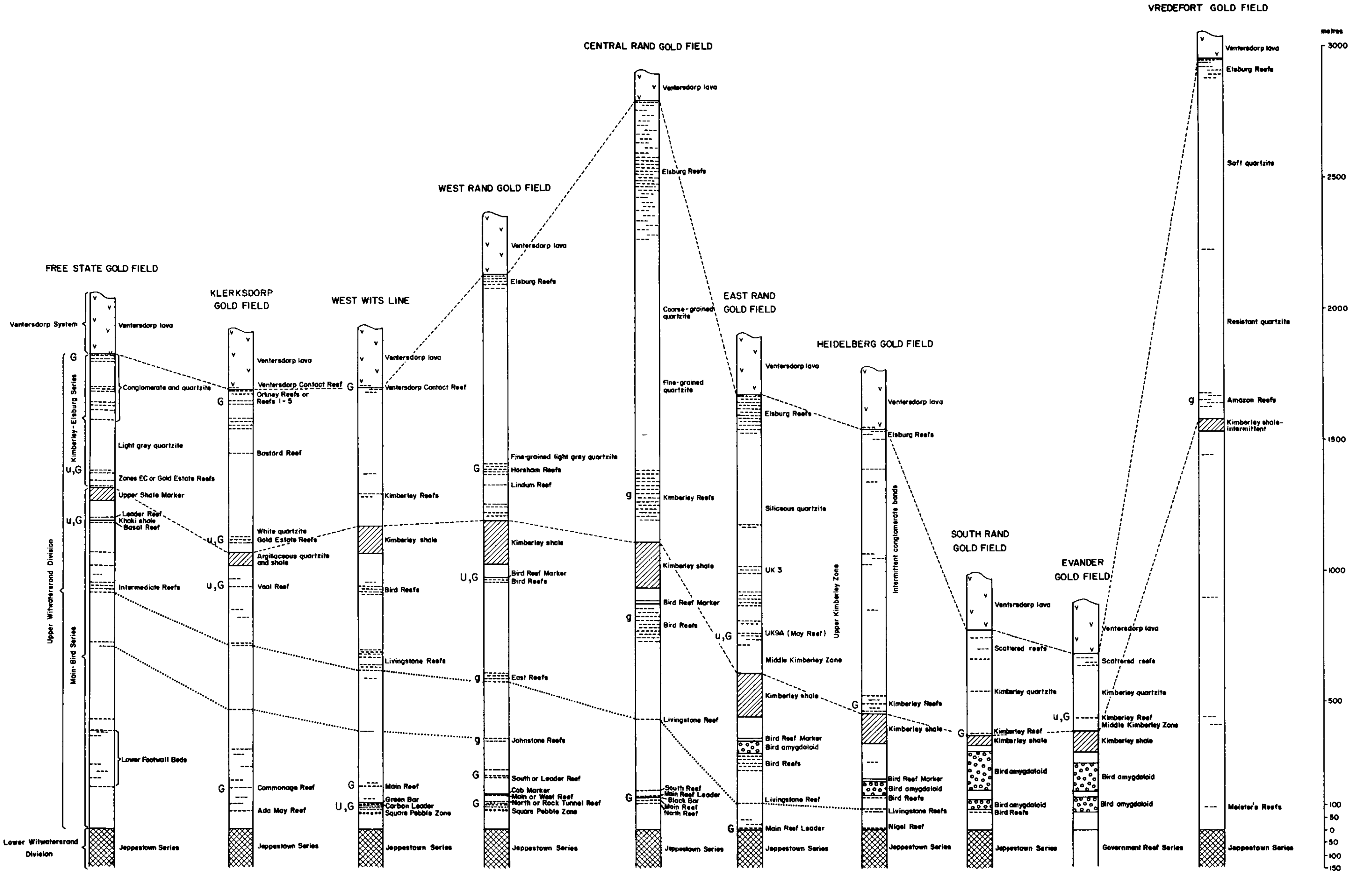

Figure 2.8. Comparative stratigraphic columns of the Upper Division of the Witwatersrand Sequence showing mineralized reefs (after C. B. Coetzee). Adapted from Whiteside and others (1976), p. 42). G,g and U,u are major and minor gold- and uranium-producing reefs respectively. Reproduced with permission from the Department of Mines and Geological Survey of South Africa.

concentrations. Pebbles in the conglomerates are mainly quartz and chert but occasionally clasts of quartzite, schist, and shale are present (Pretorius, 1976b). The pebble assemblages are highly variable between one conglomerate reef and another and can sometimes be used to identify particular reefs (Pretorius, 1976b).

The gold-bearing conglomerate reefs are the most useful marker horizons in the Upper Division. However, another marker unit, the Bird amygdaloid, occurs in the eastern part of the Witwatersrand (Figure 2.8). This unit is an amygdaloidal basalt (or diabase) which sometimes includes one or two lenses of quartzite and is frequently overlain or underlain by poorly mineralized conglomerates.

The Kimberley-Elsburg Group disconformably overlies the Kimberley Shale and is lithologically similar to the Main-Bird Group. However, the quartzites of the Kimberley-Elsburg Group are somewhat less mature than those of the Main-Bird Group and typically contain more sericite and 10-15 percent less quartz (Brock and Pretorius, 1964a). Also, the individual conglomerate layers are thicker (up to 6-10 m), more closely spaced, and less richly mineralized. These differences are manifestations of marine regression accompanied by more frequent tectonic adjustments of the margins of the basin. The decreased maturity of quartzites reflects more proximal fluvial deposition -- hence regression; the thicker conglomerate layers reflect higher-energy fluvial deposition; the close spacing of conglomerates indicates frequent tectonic uplifts; and the decrease in mineralization reflects less sediment reworking -- hence rapid deposition.

The Kimberley-Elsburg Group can be subdivided into three units: the Kimberley Reef zone, the Kimberley quartzite zone, and the Elsburg Reef zone. The Kimberley Reef zone contains numerous conglomerate layers which produce significant quantities of gold and uranium (Pretorius, 1976b). The conglomerates are quartz-pebble conglomerates which are similar to those of the Main Reef except they are slightly coarser, less mature, and thicker. The Kimberley quartzite zone consists mainly of barren, medium-grained quartzite. The Elsburg Reef zone consists of numerous lenses of coarse quartz-pebble conglomerate within a quartzite succession. Some of those conglomerate layers have been mined for gold and uranium, especially in the western part of the Witwatersrand basin (Figure 2.8).

Klipriviersberg Group. The Klipriviersberg Group contains two formations: the Vaal Bend Formation which consists of andesites, pyroclastics, quartz-feldspar porphyry, conglomerates, quartzites and shales and the Langgeleven Formation which consists of thick andesites, agglomerates, and tuffs with minor sediments (Pretorius, 1976a). According to Pretorius (1976b), these rocks represent high-energy volcanics and clastics deposited at the end of the Witwatersrand cycle of sedimentation and should be included with the Witwatersrand Sequence. This position is based on a similar suggestion made by Winter (1965). In contrast, many workers lump these rocks with the volcanics of the Ventersdorp Sequence (Brock and Pretorius, 1964a; Whiteside, 1970; Whiteside and others, 1976).

Ventersdorp Sequence

The Ventersdorp Sequence contains volcanic and volcaniclastic rocks and minor clastic sediments, and appears to conformably overlie the Witwatersrand Sequence in the eastern part of the Witwatersrand basin (Whiteside and others, 1976). However, in the western part of the basin, the contact is marked by an unconformity and a basal conglomerate called the Ventersdorp Contact Reef. This reef is an auriferous and uraniferous conglomerate layer which unconformably overlies the Witwatersrand Sequence and unconformably underlies the volcanics of the Ventersdorp Sequence. The conglomerate contains reworked erosional products of the underlying Elsburg conglomerates and it often contains higher gold values than its parent conglomerates.

Table 2.4 shows that the Ventersdorp Sequence is about 5000 m thick and is composed of massive and amygdaloidal andesites and basalts, volcanic breccias, agglomerates, and tuffs. Much of the lava was probably erupted along fissures which are now represented by diabase dikes in the Witwatersrand basin (Pretorius, 1976b; Haughton, 1963). Sediments within the Ventersdorp Sequence are minor but conglomerate, arkose, and occasional dolomite, chert and shale lenses do occur (Haughton, 1963; Whiteside, 1970).

More detailed stratigraphies of the Ventersdorp Sequence have been proposed by Winter (1965) and Whiteside (1970) and will not be discussed here. However, it should be noted that there are several disconformities within the sequence and several horizons near the base which contain pillow lavas. Both of these facts indicate that there were fluctuations in sea level caused by transgressions and regressions during deposition of the Ventersdorp Sequence (Whiteside, 1970). In general, however, most of the Ventersdorp Sequence is probably of subaerial origin.

Transvaal Sequence

The Transvaal Sequence unconformably overlies rocks ranging in age from Archean basement to rocks of the Ventersdorp Sequence. The Transvaal Sequence is subdivided into three groups: the Black Reef Group (20 m), the Dolomite Group (1520 m) and the Pretoria Group (2440 m), shown in Table 2.4.

The Black Reef Group is the most important of these groups because it contains significant quantities of gold and uranium. It consists of fine-grained vitreous quartzite and pebbly quartzite with occasional lenses of basal and intraformational conglomerate which form zones as much as 15 m thick. The upper parts of the group contain interbeds of black carbonaceous shale which become more frequent up-section and grade into black shales of the overlying Dolomite Group (Papenfus, 1964). The average thickness of the Black Reef Group in the Witwatersrand basin is 15 to 30 m. Thickness variations reflect irregularities in the underlying erosion surface, with the Black Reef thickest in topographic lows and often absent above topographic highs.

Conglomerates of the Black Reef Group contain the erosion products of the underlying rocks. Pebbles in the conglomerate are mainly well-rounded quartz pebbles, usually less than 1.25 cm in diameter (Papenfus, 1964). The matrix of the conglomerates is quartzitic and contains pyrite, gold, and uraninite. These conglomerates are similar to conglomerates of the Ventersdorp Contact Reef in that their composition is strongly influenced by the nature of the underlying rocks. For example, pebbles of quartzite, porphyry, diabase, and shale occur in areas near suboutcrops of these lithologies. Also, the quantities of gold, uranium, pyrite, and other heavy minerals are strongly variable depending on the underlying source rocks. Economic concentrations of heavy minerals in the Black Reef occur only where it overlies auriferous and uraniferous conglomerates of the Witwatersrand Sequence -- usually the Kimberley Reefs (Whiteside and others, 1976).

The Dolomite Group gradationally overlies the Black Reef Group and consists mainly of chemically precipitated sediments. The majority of the group consists of dolomite with subordinate bands of chert, limestone, iron formation, quartzite, and shale. The dolomites contain stromatolites, oolitic structures, or ripple marks in some areas. Dolomites give way up-section to cherts of various colors, ferruginous cherts, and, finally, banded iron formation (Haughton, 1963). This iron formation contains important hematite ores, expecially in the eastern part of the Transvaal basin. Gold occurs in the shales, silty dolomites, and dolomites of this group and is commonly associated with pyrite and arsenopyrite (Hammerbeck, 1976). According to Pretorius (1976a), this gold was deposited in a low-energy facies of a deltaic environment and probably was precipitated from solution by organic (algal) agents. He calls this type of deposit a Transvaal-type goldfield to distinguish it from the high-energy fluvial deposits of the Witwatersrand-type goldfields. In contrast, Hammerbeck (1976) considered the gold deposits of the upper Transvaal Sequence to be structurally controlled epigenetic deposits on the basis of the occurrence of gold lodes which cross-cut bedding.

The Pretoria Group unconformably overlies the Dolomite Group and contains a basal chert breccia or conglomerate. The Pretoria Group contains four formations: the Timball Hill Formation consisting of conglomerate, quartzite, and shale; the Daspoort Formation consisting of shale, quartzite, tillite, andesite, and iron formation; the Magaliesberg Formation consisting of shale, graphitic shale, dolomite and quartzite; and the Smelter's Kop Formation consisting of quartzite, shale, andesite and tuff (Haughton, 1963). The Pretoria Group is overlain and intruded by igneous rocks of the Bushveldt complex. Gold occurs in the upper part of the Pretoria Group both in shales and as epigenetic deposits in fracture zones (Hammerbeck, 1976).

Waterberg Sequence

The Waterberg Sequence unconformably overlies the Transvaal Sequence with considerable hiatus. It consists of reddish conglomerate and quartzite, with minor shale, andesite, and pyroclastics. The rocks are mainly subaerial and fluvial in origin. The importance of the Watersberg Sequence to this discussion is that the quartzites are red in contrast to the gray and green quartzites in the Witwatersrand and lower Transvaal Sequences. This color change indicates that hematite (either syngenetic or diagenetic) had become stable and this indicates that there was a large increase in free oxygen in the atmosphere during deposition of the Transvaal and Waterberg Sequences -- between 2300 and 1800 m.y. ago. This is about the same time that the first red-beds appeared in the Huronian Supergroup (e.g. 2200 m.y. old upper Cobalt Group) which suggests that the first appearance of red-beds may have been a world-wide event controlled by atmospheric evolution (Roscoe, 1973).

Intrusive Rocks

The Witwatersrand basin contains numerous dikes and sills composed of diabase and other lithologies. These intrusives represent at least four different igneous events. The oldest dikes and sills are diabases which are considered to represent the fissures along which the Ventersdorp lavas were extruded (Pretorius, 1976b). If so, they are about 2300-2400 m.y. old. The next episode of dike emplacement was probably associated with intrusion of the Bushveldt complex to the north about 1950 m.y. ago. These dikes include diabases, quartz porphyry, gabbro, and pyroxenite (Pretorius, 1976b). Subsequent thermal episodes about 1300 and 150 m.y. ago emplaced dikes of carbonatite, diabase, syenite, kimberlite, and other lithologies.

Sedimentology and Paleogeography

Model of Source Area

Paleocurrent data indicate that the Witwatersrand Sequence was deposited in an intracratonic basin surrounded by an Archean granite-greenstone terrain. Sediments were brought into the basin by river systems which flowed down a southeast paleoslope. The majority of the sediments entered the basin along its northwestern margin, where large deltas formed (Figure 2.7). These deltas contain the major economic concentrations of gold- and uranium-bearing conglomerates. Sediments also entered the basin along its southwestern edge but these sediments are thinner, more argillaceous, and do not contain as much gold and uranium as those on the northwestern edge (Pretorius, 1976b). This paleogeographic evidence indicates that the source of gold, uranium, and most of the sediments in the Witwatersrand Sequence was to the northwest of the basin (Pretorius, 1976b).

The area to the northwest of the Witwatersrand basin is mainly covered by younger rocks of the Ventersdorp and Transvaal Sequences and the Bushveldt complex. However, where pre-Witwatersrand basement is exposed it consists of Archean granitic rocks and greenstone belts, similar to rocks exposed in the Barbarton Mountain Land, several hundred km to the northeast. Therefore, rocks of the Barbarton Mountain Land have been used as examples of what the source area of the Witwatersrand Sequence was like. The Barbarton Mountain Land contains two basic rock types: granitic rocks and supracrustal rocks of the greenstone belts. Both rock types contributed detrital components to the Witwatersrand Sequence.

The supracrustal rocks in the Barbarton Mountain Land are called the Swaziland Sequence and the stratigraphy of these rocks is shown in Table 2.4. The lowest unit is an ultramafic unit. This is overlain by a mafic to felsic volcanic unit, a graywacke unit, and a quartzite unit. The lower ultramafic and mafic rocks contain siderophile elements such as Au, Pt, Cr, Co, and Ni and diamonds, and these rocks are thought to have been the source of detrital gold, platinoids, chromite, cobalite, and diamonds found in the Witwatersrand Sequence (Pretorius, 1976b). The upper felsic volcanics contain larger proportions of chalcophile elements such as Ca, Zn, Pb, Bi, and Ag, and these rocks were probably the source of much of the detrital pyrite, chalcopyrite, and silver found in the Witwatersrand Sequence. The volcaniclastic and clastic materials in the upper part of the greenstone successions probably contributed quartz, feldspar, and rock fragments to the quartzites of the Witwatersrand Sequence.

The Archean granitic rocks include both older tonalites and granodiorites (3400-3150 m.y.) and somewhat younger granodiorites and granites (3100-2600 m.y.). All of these granitic rocks probably contributed quartz, plagioclase, zircon, monazite, cassiterite, and garnet to the sediments. However, uranium and potassium came primarily from the younger granites and hydrothermal quartz-uraninite veins associated with them (Shepherd, 1977). This is substantiated by dates of 3050 $\pm$ 50 m.y. from detrital uraninite of the Dominion Reef, Witwatersrand, Ventersdorp, and Transvaal Sequences (Rundle and Snelling, 1977) and by the fact that detrital uranium first appeared in southern Africa about 2800 m.y. ago, whereas detrital gold appeared 3100 m.y. ago.

The validity of this source area model can be assessed by examining the history of basin filling versus erosion of the source area. Figure 2.9 shows that the stratigraphy of the Swaziland Sequence is essentially inverted in the Witwatersrand basin as a result of progressive erosion and deposition. The quartzites of the Moodies Group were reworked and deposited as quartz-arenites in the Hospital Hill Group, which are the only quartz-arenites in the Witwatersrand basin. Iron formation of the upper Fig Tree Group was redeposited as ferruginous shales in the Hospital Hill Group. The graywackes of the Fig Tree Group were redeposited as subgraywackes in the Government and Jeppestown Groups. Gold occurs at the upper and lower contacts of the Upper Onverwacht Group in the Swaziland Sequence, and these horizons were the principal sources of gold deposited in the Main-Bird and Kimberley-Elsburg Groups respectively (Pretorius, 1976a).

There are several important implications of this source area model which pertain to exploration for fossil-placer deposits. First, gold is apparently supplied by the ultramafic and mafic components of greenstone sequences. This may explain why the Blind River-Elliot Lake deposits in Canada do not contain significant gold quantities. Greenstone belts in Canada do not contain thick ultramafic sections and, instead, felsic volcanics predominate. In particular, the greenstone belts in the source area of the Huronian Supergroup are devoid of gold-bearing mafic and ultramafic sequences. Thus, a favorable source area for gold should include greenstone belts containing thick ultramafic rocks or gold-bearing mafic rocks at their base.

Second, uranium is apparently supplied from potassium-rich granites in the source area which are generally 3100-2500 m.y. old. The presence of large volumes of these granites in the source area would be favorable for concentration of placer uranium. In southern Africa, these younger granites are not as voluminous as in the Canadian Shield because of the presence of large areas underlain by greenstone belts and older tonalites. This may explain the larger quantity of detrital uranium in the Huronian Supergroup than in the Witwatersrand Sequence.

Model of Sediment Transport

The majority of the clastic components of the Pongola, Witwatersrand, Ventersdorp, and Transvaal Sequences were carried into the depositional basin by rivers and streams. This has been well documented by the lateral distribution of lithofacies, paleocurrent studies, patterns of facies changes, patterns of grain size variation, study of sedimentary structures, and the patterns of distribution of detrital gold and uranium (Pretorius, 1976a). These rivers flowed down a southeast paleoslope which persisted in southern Africa from about 3100 to 1800 m.y. and carried detrital debris from an eroded Archean basement into a shallow intracratonic marine or lacustrine basin. Within the basin, clockwise longshore currents washed and reworked the detritus and chemical sediments were deposited in the deeper part. The distribution of the rivers and the configuration of the basin were strongly influenced

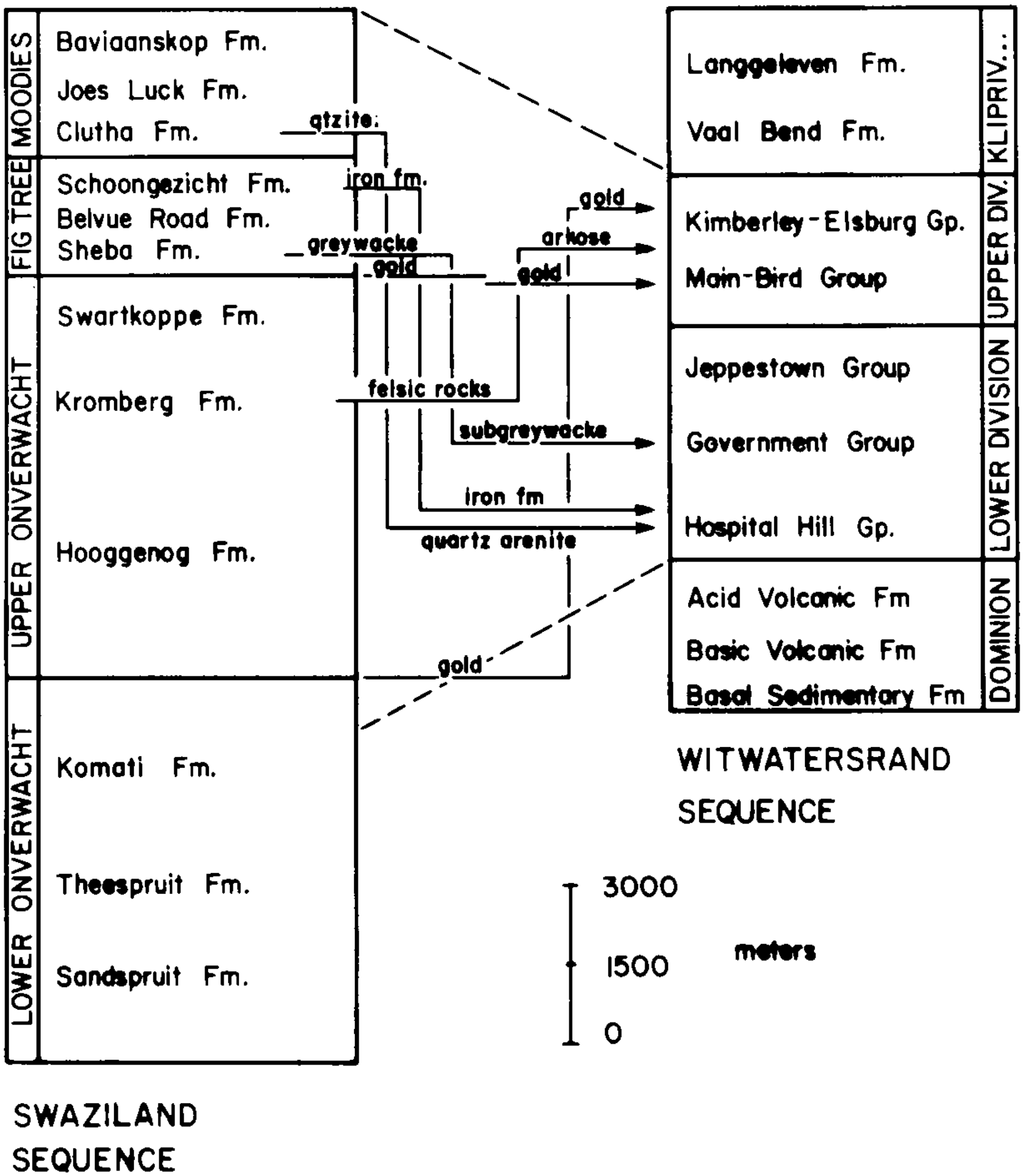

Figure 2.9. An inverted stratigraphy model of the history of L.... filling in the Witwatersrand basin. The source of most of the gold in the Upper Division of the Witwatersrand Sequence is believed to have been strata-bound ore bodies along the contacts between the Lower and Upper Onverwacht and between the Upper Onverwacht and Fig Tree Groups (redrawn from Pretorius, 1976a, p. 21). Reproduced with permission from Elsevier Scientific Publishing Company and the author.

by pre- and syn-depositional basin and dome warping and by block faulting--both in the source area and around the margins of the basin. Thus, the paleogeographic setting for deposition of the Witwatersrand Triad involved relatively high-energy fluvial transport of clastics combined with subaerial deposition of volcanics and volcaniclastics and reworking of sediment by lacustrine or shallow-marine longshore currents. The continental-lacustrine conditions were interrupted by at least one major period of marine transgression in which chemical sediments of the Transvaal Sequence were deposited.

Figure 2.10 is a schematic representation of the transfer systems which carried sediments into the Witwatersrand basin (after Pretorius, 1976a). The left side of the figure depicts the high-energy fluvial systems which deposited the coarse-clastics of the Pongola, Witwatersrand, Ventersdorp, and lower Transvaal Sequences. Pretorius (1976a) calls this the Witwatersrand-type transfer system. This type of fluvial system was characterized by: relatively high relief; short, linear transport; high energy transport; and a shallow lacustrine basin. Conglomerates, coarser sands, and heavy minerals were deposited in the high-energy regimes of this system and shales, silts, and algal mats were deposited in the lower energy regimes. Pretorius (1976a) suggested that a modern analogy would be something like the alluvial fans in the Basin and Range Province of the western United States except that large quantities of water were available. The Witwatersrand-type transfer system operated during regressive conditions and was responsible for much of the clastic deposition in the Pongola and Witwatersrand Sequences which took place between 3100 and 2500 m.y. ago.

The right side of Figure 2.10 depicts the lower-energy fluvial transport system which Pretorius (1976a) calls the Transvaal-type. This type of system was characterized by: more subdued relief in the source area; long

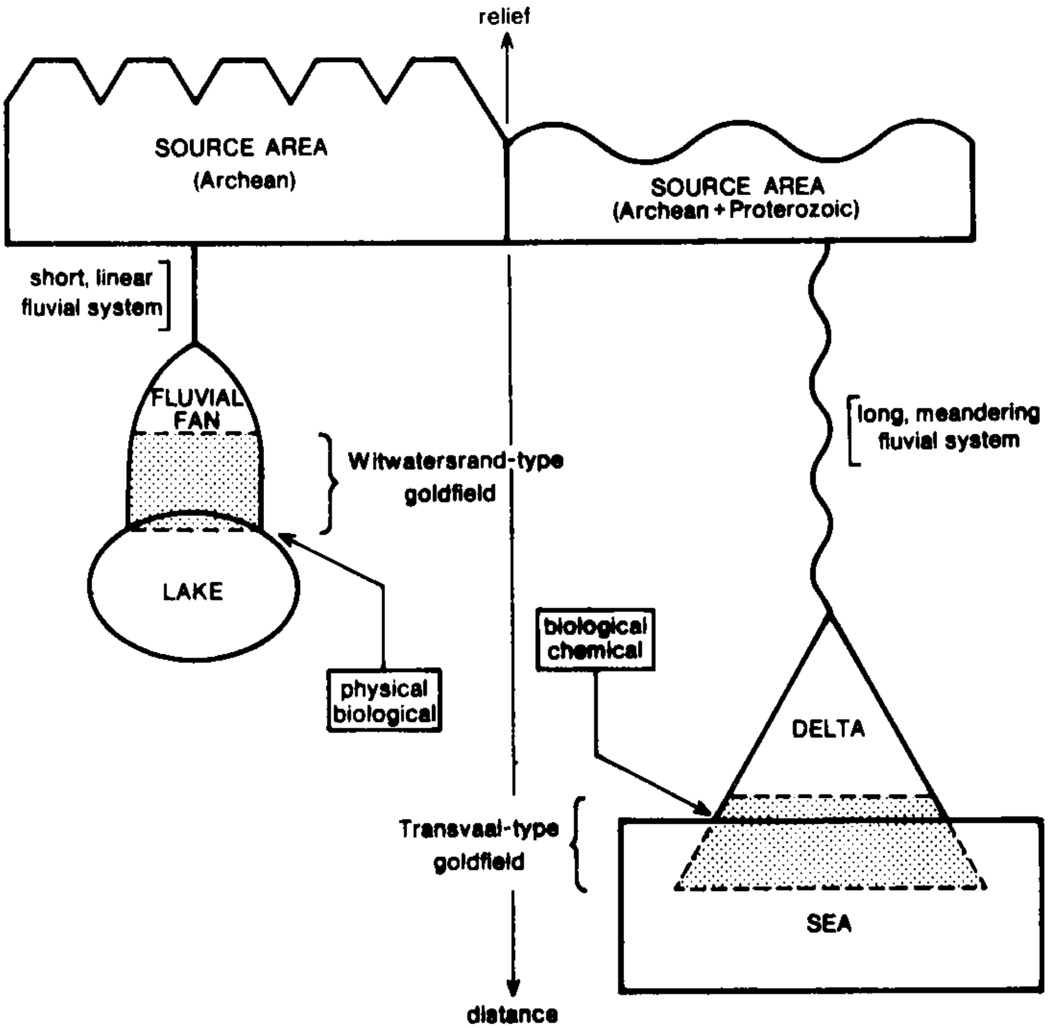

Figure 2.10. Conceptual model of the transfer systems which collect erosional debris, including gold in solid state and in solution, from the source area, and transport it to a depository, where it is concentrated by physical and biological agencies to form a Witwatersrand-type goldfield in coarse clastics and algal mats, and by chemical and biological agencies to form a Transvaal-type goldfield in fine clastics and non-clastics. Taken from Pretorius, 1976a, p. 11. Reproduced with permission from Elsevier Scientific Publishing Company and the author.

transport distances in a meandering river; lower energy transport conditions; and deeper water lacustrine or marine conditions. The sediments of this system are typically fine-grained clastics and chemical sediments such as dolomite and iron formation. The Transvaal-type transport system formed large deltas located well out into the depositional basin and would be analogous to continental-shelf type environments (Pretorius, 1976a). The Transvaal-type system operated during the major period of marine regression which affected the Witwatersrand basin during deposition of the Transvaal Sequence 2300-2100 m.y. ago.

Model of Sediment Deposition

The environments of deposition of various units of the Witwatersrand system are shown in Table 2.4. The majority of the clastics were deposited in fluvial-lacustrine or fluvial-deltaic environments and the majority of the volcanics were deposited subaerially. These interpretations of depositional environments are based on abundant stratigraphic and sedimentological data, especially for the Witwatersrand Sequence. The sedimentology of the Witwatersrand basin has, perhaps, been as extensively studied as any sedimentary basin on earth because of the gold and uranium deposits in conglomerates. The process of mining and exploration of the deposits has made available an enormous amount of subsurface information so that the configuration and distribution of lithofacies of conglomerate, sand, and shale are well known; paleocurrent data are abundant; and sedimentary structures and stratification sequences are exposed in three dimensions. The most recent interpretation of the sedimentological data is presented by Pretorius (1976a, 1976b) and Whiteside and others (1976) and is summarized below. Other important articles include: Minter (1976), Pretorius (1975), Brock and Pretorius (1964a, 1964b), and Haughton (1964).

The sedimentological data indicate that the Witwatersrand basin was an elongated asymmetrical depository with shorelines along its northwestern and southeastern edges. It is not known whether the ends were closed or open to a larger seaway. The majority of the sediment was carried into the basin by southeast-flowing rivers which formed deltas and fan deltas where they spilled into the basin. The deltaic sediments, especially the distal facies, were reworked by lacustrine (or shallow-marine) longshore currents which traveled in a clockwise direction and transported and deposited fine-grained clastic sediments between the deltas. Between deltas, algal colonies developed and chemical sediments were deposited. Outside of the basin, continental sedimentation and subaerial volcanism were the dominant styles of deposition. The northwest margin of the basin was fault-bounded and tectonically active during deposition and the majority of the sediments entered the basin on this side. These sediments are significantly thicker, coarser-grained, and more highly reworked than the sediments on the relatively passive, southeast side of the basin. Frequent tectonic adjustments along the northwestern marginal fault and in the source area created numerous regressions and transgressions during deposition of the Witwatersrand Sequence which resulted in the formation of disconformities overlain by reworked conglomerates. Constant reworking of deltaic sediments during the regressive episodes is probably the key to the accumulation of economically important concentrations of detrital heavy minerals.

Fluvial deposition in the Witwatersrand basin can be understood in terms of two end-member types of depositional environments shown in Figure 2.10, the Witwatersrand-type and the Transvaal-type. The Witwatersrand-type depositional environment was a high-energy fluvial system in which conglomerates and sands were deposited in a fan-delta at the edge of a shrinking basin. A more detailed representation of this system is shown in Figure 2.11 from Pretorius (1976a). The fan-deltas formed in this system were asymmetrical structures composed of two lobes of coarse-clastics surrounded by finer-grained clastics. The asymmetry was the result of lacustrine longshore currents which redistributed the sediments; the lobes represent the main channels of the braided streams. The components of the system were the fanhead, midfan, and fanbase (Figure 2.11). The fanhead was located near the tectonically-active basin edge and it contained cobbles and boulders in deeply incised channels. Tectonic movements along the edge of the basin usually caused the fanhead clastics to be uplifted, reworked, and redeposited farther down the fan-delta. The midfan contained gravels and sands deposited in braided, shallow channels. This part of the fan-delta contained the optimum conditions for accumulation of gold because it contained relatively coarse clastics which were depressed during tectonic adjustments, and therefore preserved. The fanbase contained finer-grained clastics which were deposited in shallow, braided channels and were strongly influenced by longshore currents in the basin.

Deposition in the Witwatersrand-type fan-delta took place by a series of pulses of sedimentation caused by regressive-transgressive cycles. The delta was built-out during regression due to fluvial deposition of open framework gravels and sands. This was followed by deposition of finer-grained clastics during transgression. Continued transgression caused clastic deposition to cease and the sediments were then eroded and reworked by lacustrine currents--forming an unconformity. A new pulse of gravels and sands were then deposited on the disconformity following regression caused by tectonic movements and the cycle was repeated. This type of cyclical deposition is well recorded in the conglomerates and quartzites of the Upper Division of the Witwatersrand Sequence.

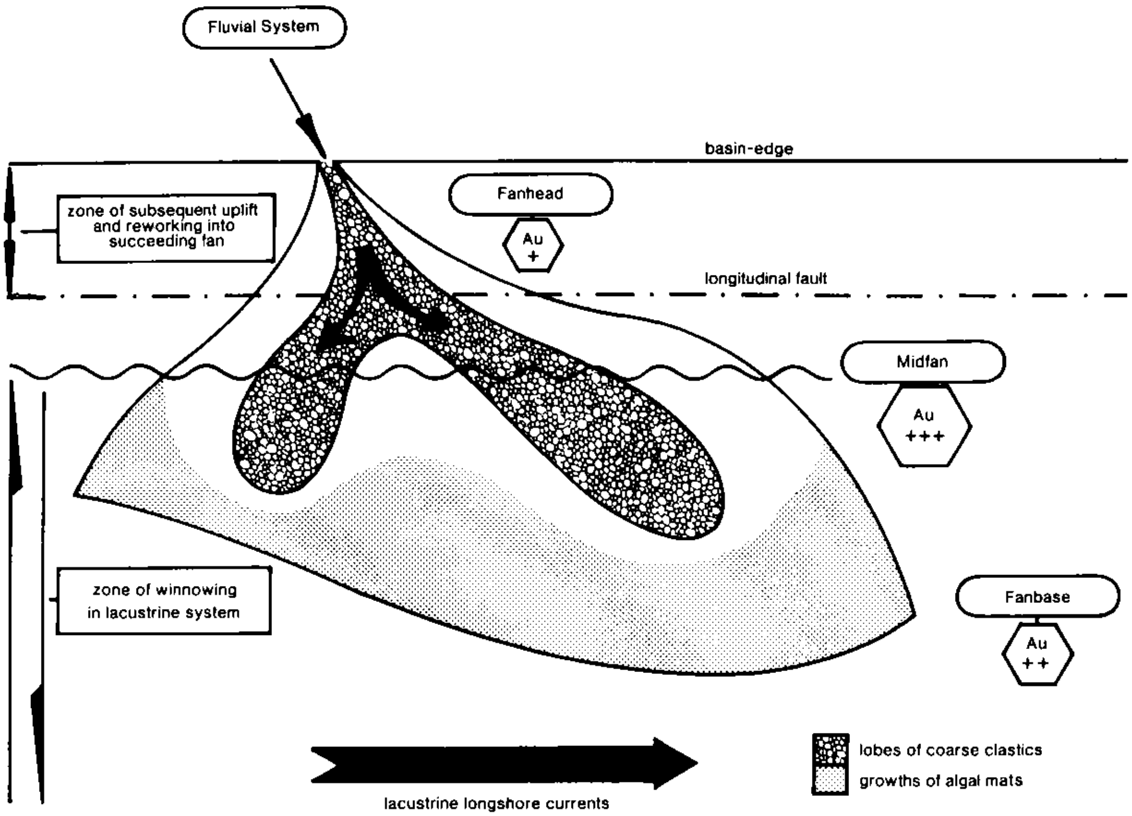

Figure 2.11. Conceptual model of a Witwatersrand-type goldfield. The fluvial system brings from the source-area unsorted erosional debris which undergoes sorting on the fluvial fan in accordance with a hydrodynamic regime radially decreasing in energy away from the apex of the fan. Because of the small grain-size of the gold particles, they are unable to settle, to any marked extent, in the fanhead facies. Optimum conditions for settling occur in the midfan facies. The energy level becomes too low to move detrital particles in any quantity to the fanbase environment. However, gold in solution is precipitated by the algae which grow preferentially in the non-turbulent conditions along the margins and base of the fan. Taken from Pretorius (1976a), p. 17. Reproduced with permission from Elsevier Scientific Publishing Company and the author.

The Transvaal-type fluvial depositional environment was a lower energy regime in which silts, muds, and dolomite were the main deposits. These fine-grained sediments were deposited in an environment which contained a long, meandering river system of low gradient feeding a large submarine delta located well out into the basin. Associated environments probably included alluvial flood plains, swamps, lagoons, and other intertidal environments. Sedimentation was less affected by tectonic movements at the edge of the basin than in the Witwatersrand-type system and the influence of shallow-marine currents was stronger. This type of depositional environment accounted for deposition of the Transvaal Sequence and probably part of the Lower Division of the Witwatersrand Sequence.

Model of Tectonic Influence of Sedimentation

The stratigraphy of the Witwatersrand Sequence can be viewed in terms of a series of cycles of varying frequency (Sharpe, 1949; Duff and others, 1967; Pretorius, 1965, 1966, 1975, 1976a). The lowest-frequency cycle is represented by the five sedimentary sequences which formed in separate intracratonic basins in southern Africa between 3100 and 1800 m.y. ago with a periodicity of about 250 m.y. Medium-frequency cycles are represented by the overall trend of deposition in each of the sedimentary sequence which generally began with volcanism, proceeded through clastic deposition, and ended with renewed volcanism. High-frequency cycles are represented by alternating lithologies within the major groups of the medium-frequency cycle. Figure 2.12 shows the details of the medium-frequency cycle of the Witwatersrand Sequence in relation to stratigraphy and tectonic stability. The Dominion Reef Group contains volcanics and coarse-clastics which were deposited in a tectonically unstable regime following a period of erosion of the Archean basement. The Dominion Reef Group was followed by the transgressive,

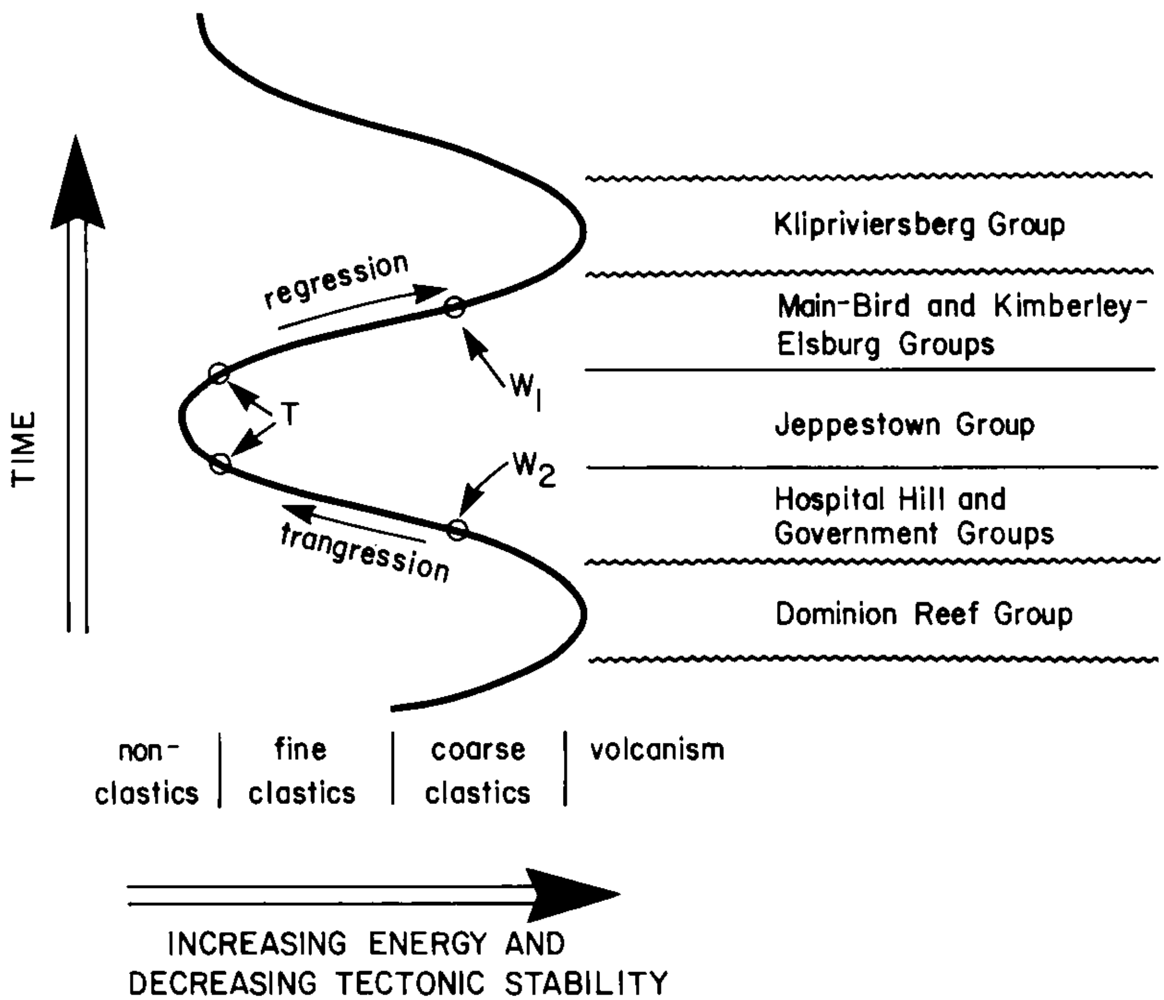

Figure 2.12. Conceptual model of the stratigraphic response to harmonic variations in energy of deposition in the Witwatersrand basin. The model requires a mirror-image repetition of the stratigraphic succession from the lower members of the depositional fill into the upper members. Because of this, ideally there are two optimum stratigraphic positions for the formation of an intermediate-energy Witwatersrand-type goldfield (W_1 and W_2) and two positions for a low-energy Transvaal-type goldfield (T). In practice, W_1 is more favorable than W_2 for gold concentration because regression causes continued sediment reworking whereas transgression causes burial of pre-existing sediments. From Pretorius (1976a, p. 14). Reproduced with permission from Elsevier Scientific Publishing Company and the author.

fining-upward clastic assemblage of the Hospital Hill and Government Groups. Transgression reached its peak during deposition of non-clastics and fine-clastics of the Jeppestown Group. This was followed by the regressive, coarsening-upward clastic sequence of the Main-Bird and Kimberley-Elsburg Groups. Finally, volcanism began again and the Klipriviersberg Group was deposited.

Higher-frequency cycles are superimposed upon the medium-frequency cycle. Sedimentation in the Lower Division consisted of alternating shales and sands and sedimentation in the Upper Division consisted of alternating sands and gravels (Figure 2.8). These alternating clastic sequences can be explained in terms of high-frequency transgressive-regressive cycles which were superimposed in the medium-frequency transgression and regression trends shown in Figure 2.12. These high-frequency transgressions and regressions were responsible for progressive build up of the fluvial fans-deltas and concentration of heavy minerals in conglomerates.

The major control of cyclical sedimentation in the Witwatersrand basin appears to have been tectonism (Pretorius, 1976a; Brock and Pretorius, 1964b). The low-frequency cycles were a response to changes in topographic elevation which reflected the large-scale tectonic trend toward crustal thickening and cratonic stability which was the prominent feature of the Archean-Proterozoic transition. The craton appears to have been slowly rising throughout deposition of the Swaziland, Pongola, Witwatersrand, and Ventersdorp Sequences (3100-2300 m.y.); then it subsided during deposition of the Transvaal Sequence (2300-2100 m.y.) and re-emerged for deposition of the Watersberg Sequence (2000-1750 m.y.). The medium-frequency cycles represent transgressions and regressions which occurred during the general trend toward cratonic stability. These transgressions and regressions reflect episodic tectonic activity which is probably related to basin and dome folding of the basin and the source area. The high-frequency cycles in the Witwatersrand Sequence appear to be related to faulting and differential subsidence near the margins of the basin. Movement on the faults repeatedly caused elevation of the source area relative to the basin which resulted in deposition of coarse clastics at the beginning of each cycle.

Mineral Deposits

Distribution

The major goldfields (and uranium producing areas) in the Witwatersrand basin are shown in Figure 2.7. In a clockwise direction these are: Welkom goldfield (also called the Orange Free State goldfield); Klerksdorp goldfield; Carletonville goldfield (also called the Far West Rand or West Wits Line goldfield); West Rand goldfield; East Rand goldfield; and Evander goldfield. The Central Rand goldfield, where gold was first discovered in the basin, is considered by Pretorius (1976a) to be the area where the East and West Rand goldfields overlap and not a separate goldfield. Each of the goldfields is interpreted to be a fluvial fan-delta which formed where a braided river system entered the basin. Their distribution on the northwest side of the depositional basin indicates that the majority of the Witwatersrand sediments were derived from a northwestern source and transported into the basin down a southeast paleoslope. Figure 2.7 also shows that the goldfields are located southeast of the trace of the Rand Anticline, in structural depressions located between domes cored by granitic material. This suggests that the northwest margin of the basin at the time of deposition of the Witwatersrand Sequence was located near the trace of the Rand Fault and that the structural pattern of basins and domes had already been established prior to deposition.

Gold and uranium occur in four different types of reefs within the fluvial fan-deltas: 1) as detrital grains in the matrix of quartz-pebble conglomerates; 2) on the foresets of cross-bedded pyritic sands which fill scoured channels; 3) in thin layers which overlie unconformities developed on sands or muds; and 4) in thin carbon seams which developed on unconformities. Each of these types of reefs represents some facies of cyclical fluvial-deltaic deposition. The first two were deposited in response to the initial sedimentation which followed a period of erosion. Conglomerates were deposited in thick and extensive bankets in areas where the main channels of a braided river coalesced. In other areas, where transport distance was less, sands were deposited on the eroded surface. In both of these cases, some heavy minerals were brought in with the coarse-clastics and others were simply incorporated into the basal clastics from debris left on the erosion surface. In the last two types of reefs, heavies were concentrated by winnowing processes which were active during the terminal, erosion phase of the cycle. Then, if sand deposition of the succeeding cycle was rapid enough, and of low-enough energy, the accumulated heavies on the erosion surface were buried undisturbed instead of being ripped up and incorporated into the succeeding coarse-clastic type of reef. The result was a layer of heavy minerals between two sand or mud units.

It should be emphasized that all of these types of gold- and uranium-bearing reefs in the Witwatersrand basin occur either directly below or directly above inter-cycle unconformities. As a consequence, the identification of such unconformities is of great economic importance.

Lithology and Mineralogy

The mineralized conglomerates of the Witwatersrand Sequence are relatively mature quartz- and chert-pebble conglomerates which are typically composed of 80 percent (by weight) pebbles and 20 percent matrix (Pretorius, 1976b). Quartz and chert pebbles usually constitute 90-95 percent of the pebble fraction; quartzite and quartz-porphyry often constitute 5-10 percent; and less competent clasts of rock fragments of shale, schist, and volcanics rarely make up more than 1-2 percent (Pretorius, 1976b). Most of the pebbles are smaller than 2.5 cm in diameter but they can range up to 10 cm. The larger pebbles are usually well-rounded to sub-rounded and oval shaped and the smaller pebbles are subangular (Whiteside and others, 1976). Pebbles are usually well-sorted due to cyclical reworking of the sediments and packing is usually fairly tight, but not always pebble-supported. In general, the highest grade conglomerates are the ones with the lowest percentage of labile pebbles such as quartzite, quartz-porphyry, shale, and schist, and the highest percentage of quartz and chert pebbles.

The matrix of the conglomerates consists mainly of quartz grains with variable amounts of phylosilicates, heavy minerals, rock fragments, and feldspar. The size of the quartz fraction is commonly in the range of .5 to 1 mm and the quartz grains are usually recrystallized. The percentage of heavy minerals is variable and tends to increase with increasing quartz content of the matrix. As an example, the Vaal Reef of the Main-Bird Group (Figure 1.8) in the Klerksdorp area is an important producer of both gold and uranium. Its matrix contains 93 percent quartz, 6 percent sericite and chlorite, and 1 percent rock fragments. In contrast, the quartzite units directly above the reef contain 62 percent quartz, 34 percent phylosilicates, and 4 percent rock fragments and are essentially barren of gold and uranium (Minter, 1976).

The heavy mineral fraction of the conglomerates of the Witwatersrand Sequence includes an impressive number of different minerals which are shown in Table 2.6. Of these, gold, platinum, uranium and pyrite (for sulphuric acid) are the major ores, with gold being by far the most important economically. The most abundant heavy mineral is pyrite, which commonly constitutes

15 percent of the matrix and 3 percent of the total rock in an ore-bearing conglomerate and can constitute up to 25 percent of the rock (Pretorius, 1976b). The most common non-economic heavy minerals are chromite, zircon, and leucoxene. The relative percentages of these three minerals is quite variable and is often diagnostic of a particular reef (Pretorius, 1976b).

Pyrite occurs in three distinguishable varieties: allogenic, concretionary authigenic, and reconstituted authigenic grains (Köppel and Saager, 1974). The allogenic pyrite is the most common type in the Dominion Reef and in many of the other reefs of the Witwatersrand Sequence. It is identified by its rounded and abraded outlines, its fresh homogeneous appearance, and a distinct suite of inclusions--most notably galena. It is interpreted to be detrital in origin on the basis of its morphology, because of its distribution in layers in the matrix of conglomerates and on the foresets of cross-beds, and because it is in hydraulic equivalence with uraninite, quartz, and gold in the matrix of the conglomerates. The second type is concretionary authigenic pyrite, and it occurs as rounded, porous grains 1-4 mm in diameter, some of which are layered or zoned with alternating sulphide and silicate layers (Simpson and Bowles, 1977). This type of pyrite was interpreted by Simpson and Bowles (1977) to be detrital on the basis of the heterogeneity of these pyrite grains on the scale of a hand specimen and the presence of fractured and abraded grains. They suggested that this type represents pyritized mudballs formed within the basin. The third type is reconstituted authigenic pyrite which occurs in euhedral shapes and is found both in the matrix of conglomerates and as overgrowths on detrital pyrite grains. Simpson and Bowles (1977) noted the presence of gold grains within reconstituted pyrite grains and suggested, on this evidence, that gold, pyrite, and uraninite were remobilized during post-depositional regional metamorphism to form these reconstituted grains.

Economic Minerals	Sulphides	Oxides	Silicates	Others
Gold	pyrrhotite	quartz	muscovite	calcite
Tellurium	leucopyrite		sericite	dolomite
	loellingite	cassiterite	pyrophyllite	
Silver	marcasite		chlorite	zenotime
Stromeyerite		chromite	chloritoid	monazite
Proustite	chalcopyrite		biotite	
Dyscrasite	chalcopyrrhotite	columbite		diamond
	cubanite		kaolinite	graphite
Platinum	chalcocite	corundum		
Platiniridium	neodigenite		epidote	
Osmiridium	covellite	magnetite		
Iridosmine	bornite	hematite	tourmaline	
Sperrylite	tennantite	goethite		
Braggite			garnet	
Cooperite	galena	rutile		
		leucoxene	zircon	
Uraninite	sphalerite	ilmeno-rutile		
Thucholite		ilmenite	sphene	
Brannerite	molybdenite	anatase		
Uranothorite	bismuthinite	brookite		
Pyrite				
	arsenopyrite			
	skutterudite			
	cobaltite			
	glaucodot			
	linnaeite			
	safflorite			
	gersdorffite			
	niccolite			
	millerite			
	pentlandite			
	bravoite			
	mackinawite			

Table 2.6. Minerals present in Witwatersrand auriferous horizons. Taken from Pretorius (1976b, p. 64). Reproduced with permission from Elsevier Scientific Publishing Company and the author.

Uraninite occurs in two forms in the Witwatersrand Sequence: as well-crystallized uraninite with a significant thorium content (∿ 6.6 percent) and as poorly-crystallized uraninite with a low thorium content (∿ 2 percent) which is associated with organic material (Simpson and Bowles, 1977). The well-crystallized uraninite occurs as discrete euhedral to rounded grains .06-.25 mm in diameter which are often fractured. Simpson and Bowles interpreted this type of uraninite to be allogenic on the basis of its morphology, distribution, and the fact that it is in hydraulic equivalence with pyrite and quartz grains in the matrix of conglomerates (Koen, 1961). The median uranium to thorium ratio in this type of uraninite is 8.5.

Poorly crystallized, low thorium uraninite is found associated with thin seams of carbonaceous material and in pyrite concretions. These carbonaceous seams are very rich in both uranium and gold and are considered by Hallbauer (1975) to be the fossilized residue of Precambrian algal colonies. The carbonaceous material occurs in granular and columnar varieties in seams which are about 1 cm thick. Gold and uranium are disseminated in the pores and interstitial areas in both types of carbonaceous material. According to Simpson and Bowles (1977), the uranium was incorporated into the carbonaceous layers in two ways: as fine-grained allogenic uraninite particles caught by the algal material, and as uranyl ions which were absorbed, reduced, and precipitated by the carbonaceous muds to form the low-thorium uraninite. They suggested that the low-thorium uraninite in the carbonaceous material is evidence that uranium was transported in solution and precipitated by organic agents in the lower-energy environments of the fluvial fan-delta. One example of a carbon seam is the Carbon Leader of the Main Reef Group, which is a maximum of 13 mm thick and contains important quantities of uranium and gold. The uraninite in pyrite concretions occurs in the pores or in the rims and probably was absorbed during transport of the concretion (Simpson and Bowles, 1977).

Gold occurs in three forms in the Witwatersrand Sequence: 1) with quartz and silicates in the matrix of conglomerates and in quartzites, 2) with carbonaceous material, and 3) as coatings on pyrite and other sulphides (Hallbauer and Joughin, 1973). The majority of the gold is of the first type. It occurs as platelets and irregular specks .005-.5 mm in diameter and is associated with quartz and phylosilicates (Pretorius, 1976b). This type of gold was transported as detrital particles in sands and was incorporated in the matrix of gravels either by incorporation of underlying sands during deposition of gravels or by downward filtration of sands into previously deposited open-framework gravels (Pretorius, 1976b). The second type of gold occurs as veinlets and patches in the pores of carbonaceous material and as a replacement of the algal filaments in the columnar variety of carbonaceous material (Hallbauer, 1972, 1975). The gold is associated and intergrown with fine-grained uraninite, pyrite, and galena. According to Pretorius (1976a), this type of gold was transported in solution as chloride and cyanide complexes and was precipitated by organic material and absorbed on colloidal particles of clay. Deposition of this type of gold occurred in the distal (fanbase) parts of the delta under low-energy, estuarine conditions. The third type of gold occurs as inclusions in or overgrowths on sulphide minerals such as pyrite, chalcopyrite, and galena (Liebenberg, 1973). Much of this type of gold probably represents small-scale, post-depositional remobilization of gold.

Genesis

The genesis of gold- and uranium-bearing conglomerates can be largely explained in terms of the sedimentological processes which transported and concentrated detrital heavy minerals in conglomerates and coarse sands of a braided river system. The nature of these sedimentological processes have been discussed in previous sections dealing with models of the source area, transfer system, and depositional environment of the conglomerates and will not be discussed here. However, there are two aspects of ore-genesis which need further discussion. These are the geochemical constraints and organic controls of ore genesis.

The most important question concerning geochemistry pertains to the composition of the atmosphere and waters in which the ores were deposited. Geological evidence is now interpreted to indicate that most of the gold, uraninite, pyrite, and other sulphide minerals were transported as detrital particles in braided rivers. Uraninite, pyrite, and sulphides are generally unstable minerals in modern fluvial systems because they tend to be dissolved in oxygenated waters. As a consequence, their presence in conglomerates of the Witwatersrand Sequence is usually interpreted as evidence for a low oxygen fugacity in the Archean and Early Proterozoic atmosphere (Pretorius, 1976a; Grandstaff, 1974). This concept fits well into the larger concept of atmospheric evolution which is based on evidence from biologic evolution (Cloud, 1968), and from the temporal distribution of redbeds (Roscoe, 1973), iron formation, and dolomites. These lines of evidence indicate that abundant free oxygen did not become available in the atmosphere until about 1800-2000 m.y. ago.

However, this model is difficult to reconcile with the relatively new geologic evidence of Simpson and Bowles (1977) and Pretorius (1976a) which indicates that gold, uranium, and sulphate were transported in solution and syngenetically deposited in the lower-energy regimes of the fluvial-lacustrine system by organic agents. Simpson and Bowles (1977) interpreted these data as an indication that uranyl and sulphate ions, and presumably gold complexes, were transported in <u>oxidizing</u> waters and precipitated in local reducing environments created by still-water conditions and organic activity. According to them, uraninite can be transported as a detrital mineral in oxygenated waters if it contains more than 1 percent thorium which all the Witwatersrand uraninite does. They propose that the atmosphere and waters during deposition of the Witwatersrand Sequence were oxidizing.

If all the interpretations of geological data are correct, then the atmosphere and water geochemistry was such that uranium, gold, and sulfur were transported both in detrital grains and in solution, and organic material was present to reduce, precipitate, and absorb the materials from solution. The exact Eh and pH of the atmosphere and waters in which this could occur is not known. Certainly some oxygen was present, but the question of how much remains enigmatic.

The exact role of organic material is another interesting problem in terms of ore genesis. Pretorius (1976a) showed that the peaks in development of gold mineralization in the Witwatersrand and Transvaal Sequences corresponds to peaks in algal activity within the Witwatersrand basin. He suggested that this correlation reflects sedimentological controls and that the braided river-lacustrine depositional environment was favorable both for placer gold deposition and algal activity. The presence of unmineralized carbonaceous seams (Simpson and Bowles, 1977) and carbon-free mineralized conglomerates indicates that there was not necessarily any direct dependence of mineralization on carbonaceous material or vice versa.

Economic Importance

The history of exploration in the Witwatersrand basin is summarized by Borchers (1964) and Whiteside and others (1976). The discovery of gold in conglomerates dates back to 1885 when the Struben brothers discovered gold-bearing reefs in the Lower Division of the Witwatersrand Sequence. Higher grade conglomerates of the Main Reef were discovered in the Central Rand in 1886 and Johannesberg was founded that same year. The Klerksdorp field was discovered in 1887. The large extent of the conglomerate ores in the subsurface was proven in the early 1890's by deep drillholes in the Central Rand which intersected the Main Reef at depths in excess of 1000 m. These drillholes showed that the dip of the beds decreased at depth and this led to the idea that the Witwatersrand was a basin. This idea led to the discovery of the West Rand in 1898. There was little exploration from 1907 to 1932, at which time South Africa went off the gold standard and the price of gold consequently jumped. This resulted in new activity in existing goldfields, such as the discovery of the Vaal Reef in Klerksdorp field in the 1930's, and the discovery of new goldfields such as Carletonville goldfield in 1935 and the Welkom goldfield in 1939. The last of the presently known goldfields, the Evander goldfield, was discovered in the middle 1950's. In 1953, recovery of uranium as a by-product of gold mining began. Uranium production reached its peak in 1960 and gold production was highest in 1970.

The economic importance of the Witwatersrand basin is nearly unparalleled. This one mining district has contributed about 55 percent of all the gold that the world has produced throughout history--about 29,000 metric tons of gold valued at more than 34 billion dollars. As of 1972, it was responsible for about 66 percent of the world's annual gold production, and it still contains the world's largest reserves of gold (Pretorius, 1976b). The Witwatersrand basin has also produced about 76,000 metric tons of uranium valued at about 2 billion dollars and it contains about 20 percent of the world's "reasonably assured" uranium reserves of \$10/pound U_3O_8 (Ninninger, 1974).

The most significant gold-bearing reefs are those in the lower part of the Main-Bird Group. These reefs include the Carbon Leader, Main Reef, Main Reef Leader, and South Reef which are shown in Figure 2.8. Of these, the Carbon Leader has the highest gold grades--20 ppm in the Carletonville goldfield. Other important gold-bearing reefs include the Bird, Kimberley, and Elsburg Reefs of the Witwatersrand Sequence, the Ventersdorp Contact Reef, and the Black Reef of the Transvaal Sequence. The average ore grade of all these reefs is about 10 ppm.

Uranium is distributed somewhat differently. The most important uranium ores come from the Bird Reef section of the Main-Bird Group, while the Main Reef section is relatively low in uranium. The highest grades of uranium occur in the Vaal and Basal Reefs, and the Carbon Leader (Figure 2.8), with values up to about .1 percent. The Dominion Reef, like the Bird Reefs, is generally higher in uranium than in gold. The Kimberley Reefs are the youngest reefs which contain significant uranium mineralization. According to Whiteside and others (1976), the main uranium mines occur in the western part of the Witwatersrand basin--in the Carletonville, Klerksdorp, and Welkom goldfields.

WEST AFRICA

Ghana and Ivory Coast

Geologic Setting

Auriferous quartz-pebble conglomerates occur in the Tarkwa Sequence of Ghana and Ivory Coast, in the southwest corner of the West African Shield (Figure 2.5). In the southern part of this shield, the oldest rocks are granitic gneisses and infolded supracrustal rocks which yield dates between 2600 and 3700 m.y. (Hurley and Rand, 1973; Cahen, 1961, Hurley and others, 1975; Hedge and others, 1975; Williams, 1978). These Archean rocks are overlain by the Birrimian Sequence containing a lower assemblage of slates, phyllites, graywackes, and volcanics and an upper assemblage of metavolcanics and minor phyllite, graywacke, and manganiferous shale (Junner and others, 1942, *in* Hurley and Rand, 1973). The Birrimian Sequence was metamorphosed and intruded by granites during a thermal episode about 2000 m.y. ago (Hurley and Rand, 1973; Kolbe and others, 1967). Thus, the age of Birrimian Sequence is between 2600 and 2000 m.y.

The Tarkwa Sequence unconformably overlies the Birrimian Sequence (Junner and others, 1942; Service, 1943, *in* Sestini, 1973) and consists of conglomerates and quartzites of fluvial origin. These rocks are at least 1645 m.y. old (Holmes and Cahen, 1955, *in* Sestini, 1973). The rocks are folded, along with the underlying Birrimian Sequence, into asymmetric northeast plunging folds with steep, overturned, or thrust faulted northwest limbs. Metamorhpic grade ranges from middle greenschist facies to middle amphibolite facies (Sestini, 1973).

Figure 2.13 shows the two major outcrop areas of the Tarkwa Sequence in Ghana and Ivory Coast. The southern outcrop, located near Tarkwa, Ghana is about 250 km long and 25 km wide. This outcrop contains gold mines along its southeastern margin. The northern outcrop is about 200 km long and straddles the border between Ghana and Ivory Coast. This outcrop contains only minor amounts of gold (Sestini, 1973).

Stratigraphy and Lithology

The Tarkwa Sequence near Tarkwa is about 2400 m thick and is subdivided into four formations: Kawere conglomerate (lowest), Banket quartzite, Tarkwa phyllite, and Huni sandstone (Junner and others, 1942, in Sestini, 1973). These units and cross-cutting diabase dikes and sills are shown in Figure 2.13. According to Sestini (1973), these formations represent lithofacies of a fluvial depositional environment so each of the formations is approximately contemporaneous with and interfingers with the other formations.

The Banket quartzite is of greatest interest to this discussion because it contains the major gold-bearing conglomerates of the sequence. The Banket quartzite is about 120-600 m thick and is underlain by less mature arkoses and arkosic conglomerates of the Kawere conglomerate. The base of the Banket quartzite is marked by a conglomeratic zone 30-100 m thick which contains at least three reefs separated by cross-bedded, fine- to medium-grained sericitic quartzite and pebbly quartzite. The reefs are called the Basal (or Main) reef, the Middle (or West) reef, and the Breccia reef (Sestini, 1973).

The Basal reef is 4-15 m thick and contains between two and ten distinct conglomerate lenses which pass laterally and vertically into relatively mature, trough cross-bedded grit, breccia, and pebbly quartzite characterized by hematite seams, sericite, and rare feldspar. The conglomerate layers occur as isolated lenses of tightly packed pebbles up to 1 m thick and as the basal part of fining-upwards, trough cross-bedded stratification sequences (Sestini, 1973). The conglomerates are usually poorly sorted. The average pebble size is 5-8 cm but they range up to 18 cm. The larger pebbles (8-18 cm) are well-rounded and smaller pebbles (35-60 mm) are sub-angular. More

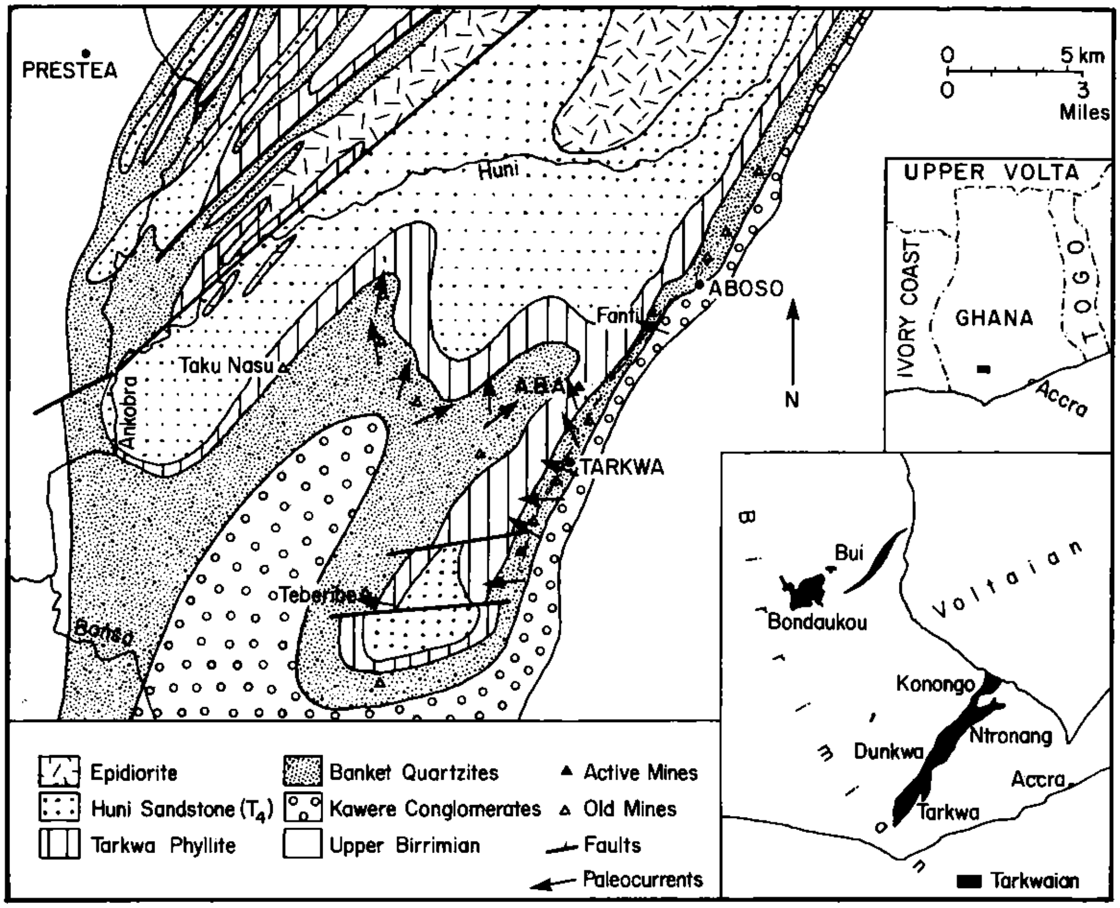

Figure 2.13. Geology of the Tarkwa region and index maps of the Tarkwaian outcrops of Ghana and Ivory Coast. Redrawn from Sestini (1973, p. 277). Reproduced with permission from Springer-Verlag.

than 90 percent of the pebbles are quartz, the rest being schist, phyllite, and chert. The matrix of the conglomerates consists mainly of quartz grains (.9-1.4 mm) with sericite, hematite (.4-.6 mm), magnetite, and accessory tourmaline, zircon, rutile, garnet, chloritoid, epidote, pyrite, bornite, and chalcopyrite. Hematite comprises 10-50 percent of the matrix of the conglomerates and gold values range up to 60 ppm, with an average of 5-15 ppm (Sestini, 1973).

The Middle reef is 1-15 m thick and usually consists of 2-5 conglomerate layers. It is characterized by abundant hematite seams and small-scale trough cross-bedding. It differs from the Basal reef in that it is finer grained and usually contains less gold (2-6 ppm).

The Breccia reef is 2-20 m thick and consists of angular lithic fragments of schist and phyllite up to 20 mm in diameter. These fragments are interpreted by Sestini (1973) to be fragments derived from mudstones and shales of the Tarkwa Sequence and from the underlying Birrimian Sequence. The Breccia reef is associated with quartzites and grits containing thin breccia bands and scattered lithic fragments. This reef usually contains less than 3 ppm gold (Sestini, 1973).

Sedimentology

The Banket quartzite is interpreted by Sestini (1973) to be a fluvial assemblage on the basis of sedimentary structures, stratification sequences, and the aerial distribution of lithofacies. His evidence for fluvial deposition is as follows: 1) the lenticular nature of conglomerate beds, with transitions from conglomerate to pebbly quartzite; 2) abundant trough cross-bedding; 3) fining-upwards cycles, from conglomerate to fine-grained quartzite, accompanied by decrease in depth of cross-bed sets; 4) poor sorting and abundance of sand-supported gravel; 5) abundance of tabular pebbles, many with their long axis normal to the current direction, and occasional upstream imbrication; and 6) low roundness of pebbles, indicating short transport distances.

Sestini envisioned a braided river system on a piedmont slope instead of a meandering river system in a flood plain. Evidence for braided river deposition includes: the wide aerial extent of the conglomerates, low variability in cross-bedding types and directions, an absence of fine-grained overbank sediments and other floodplain features such as lamination and mudcracks, and poor sorting. All of these factors suggest deposition by an overloaded, relatively high-gradient braided river system.

Paleocurrents measurements on crossbedding indicate that the source of the rocks near Tarkwa was to the east and south of the present outcrops (Figure 2.13). Sestini (1973) suggested that the sediments were deposited in a series of alluvial fans which were built up on the east and south margins of a depositional basin by west and north flowing, converging rivers. He also proposed that the present limit of outcrops of the Banket quartzite was the approximate location of the basin margin during deposition of the Basal reef. This idea is supported by the decrease in gold content of the basal conglomerate to the west--down the paleoslope.

Mineral Deposits

Gold is the only reported mineral of economic importance in the Tarkwa Sequence. It occurs as irregular grains 10-15 microns in diameter in the matrix of conglomerates. Several lines of evidence indicate that the gold is detrital in origin. These are: 1) gold is associated with conglomerates; 2) high gold values occur in strips which are elongated parallel to the paleocurrent directions; 3) gold values are highest in the lower 25-50 cm of the Basal reef; 4) gold tends to be richest in conglomerate lenses which are thin, very coarse-grained, and have tightly packed pebbles; and 5) there is a direct correlation between high gold values and high concentrations of other heavy minerals, especially hematite.

Unlike most Early Proterozoic fossil-placer conglomerates, hematite rather than pyrite is the most abundant heavy mineral. In the Tarkwa conglomerates, hematite may constitute more than 50 percent of the matrix of the conglomerates and commonly comprises greater than 20 percent. It occurs as euhedral to rounded grains .4-.6 mm in diameter which may be randomly distributed in the matrix of conglomerates or concentrated along foreset beds and in small scours in the quartzites. It is often associated with rounded grains of zircon and rutile which are obviously of detrital origin. This association, and the relationship of hematite grains to sedimentary features, indicates a detrital origin for the hematite. Recrystallization during metamorphism was responsible for the euhedral shapes and for remobilization of hematite to form stringers and patches of hematite which are oriented parallel to foliation.

Sestini's model for the origin of the gold- and hematite-bearing conglomerates is very similar to the model developed by Pretorius (1976a) for the origin of the Witwatersrand conglomerates. It involves high energy transport of detritus in a braided-river system, deposition of a fluvial fan-delta at the margin of a lacustrine or shallow marine basin, and concentration of heavy minerals in the middle portions of the fan-delta by repeated reworking of the sediments. In the Tarkwa Sequence, the observed correlation of high gold values with thin, tightly packed, very coarse conglomerates can be explained in terms of this model. Gold was deposited in moderate quantities with very coarse gravel near the apex of the fan-delta. Subsequent reworking of this gravel by fluvial currents removed the sand, improved the packing of the pebbles, made the unit thinner, and concentrated the heavy minerals. The resulting open-framework lag gravels were subsequently covered and infiltrated with sand and heavy minerals during the next pulse of sedimentation. The heavy minerals were concentrated in the gravels through settling processes and the sand was winnowed out by continued reworking. Eventually, a tightly packed conglomerate with appreciable concentrations of gold and other heavy minerals developed (Sestini, 1973).

There are two other lines of evidence indicating continued reworking of the conglomerates in a fluvial fan-delta environment. First, the auriferous reefs are richest near the east margin of the present outcrop area. This is also where the conglomerate layers are thinnest. To the west, the reefs get thicker, they contain more rock fragments, and gold values decrease. This suggests that reworking of conglomerates near the basin edge was important in concentrating gold. Second, in the eastern part of the Tarkwa Sequence, gold in the Basal reef zone is found mainly in the basal conglomerate, whereas to the west, significant gold quantities are found in the upper conglomerate layers of the zone. Similarly, the Middle reef appears to get richer toward the west. Both of these facts suggest that uplift, erosion, and reworking near the apex of the fan-delta, after deposition of the basal conglomerate, provided a source of gold which was transported and reconcentrated farther down the fan with time. This suggests that the Tarkwa Sequence is a regressive sequence deposited in a shrinking sedimentary basin; a situation very similar to conditions during deposition of the Upper Division of the Witwatersrand Sequence.

The origin of the Tarkwa ore deposits seems to be closely analogous to that of the Witwatersrand Sequence in many respects. The paleogeographic environments were similar in that both metasedimentary sequences were deposited in intracratonic basins within stable Archean shield areas. Also, the lithologies and depositional environments appear to be the same. However, there is one major difference, and that is the heavy mineral assemblages. The Witwatersrand conglomerates contain gold, uraninite, and pyrite, whereas the Tarkwa conglomerates contain gold and hematite. One possible explanation is that the Archean source areas were drastically different, containing, perhaps, a greenstone belt with associated iron formation and older tonalitic granites. A source area of this sort could supply sufficient quantities of gold, hematite, and quartz plus accessory zircon, rutile, garnet, chloritoid, pyrite, and chalcopyrite. If younger, potassium-rich granites were absent, there may not have been a source for detrital uraninite. However, the lack of abundant pyrite might be difficult to explain in terms of source area controls.

The alternative explanation, and the one favored by us, is that the Tarkwa conglomerates were deposited in a more oxygenated atmosphere than the Witwatersrand conglomerates and are several hundred million years younger than the Witwatersrand conglomerates. In an oxidizing fluvial system, uraninite and pyrite are unstable as detrital minerals (Pretorius, 1976; Liebenberg, 1955; Grandstaff, 1975; Roscoe, 1969), whereas gold, hematite, zircon, rutile and other heavy silicates are stable. According to Roscoe (1973), an oxidizing environment appeared on earth about 2200 m.y. ago, as shown by the change from pyrite-dominated heavy mineral assemblages (yellow sands) in the Lower Huronian Supergroup to hematite-dominated heavy mineral assemblages (black sands) in the Upper Huronian Supergroup which took place about this time. We speculate that the Tarkwa Sequence was deposited sometime after the appearance of appreciable free-oxygen in the atmosphere, perhaps less than 2200 m.y. ago and that, except for this age difference ($<$ 2200 versus 2700 m.y.), the Tarkwa Sequence is closely analogous to the Witwatersrand Sequence.

Gabon

Precambrian uraniferous conglomerates have also been reported from Mounana, Gabon, on the west coast of Africa (Bernazeaud, 1959; des Ligeris and Bernazeaud, 1960; Davidson, 1964; De Kun, 1965; Krendlev and others, 1967). Figure 2.5 shows that this area is near the southern margin of the Archean Gabon Craton of Hurley and Rand (1969) and is not far from Jacobina on the pre-Mesozoic continental reconstruction of South America and Africa.

According to Davidson (1964, p. 173), this area contains widespread radioactive conglomerates and sandstones of Proterozoic age which crop out close to an Archean granitic basement. De Kun (1965, p. 340) shows that the conglomerates are in fault contact with granitic basement and are overlain by shales, a situation broadly similar to that at Jacobina, Brazil. The radioactive minerals are reported to be black oxides of uranium and vanadium and they are associated with pyrite, galena, chalcopyrite, sphalerite, bornite, barite, calcite, and wulfenite--but no gold (Davidson, 1964). These minerals apparently form "bedded and irregular impregnations in the sediments" (Davidson, 1964) and Davidson suggested this occurrence is, perhaps, more similar to a Colorado Plateau-type uranium-vanadium deposit that to the Witwatersrand-type deposit. Certainly the presence of vanadium minerals, the porous nature of the host conglomerates, and the apparent fault control of mineralization suggest that the uranium is probably secondary. However, De Kun (1965, p. 340) reported that there was no "sympathetic relationship" between the vanadium and uranium and the minerals map of Africa (UNESCO, 1969) shows the Mounana deposit as a lateritic or eluvial deposit. If so, this would suggest that the uranium is primary and older than 2000 m.y. because after this time uranium ceased to be a resistate mineral because of increasing oxygen in the atmosphere. Nevertheless, we are inclined to believe the deposit is not a typical Early Proterozoic fossil placer but is probably secondary in origin.

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SOUTH AMERICAN PLATFORM

GEOLOGIC SETTING

Radiometric age dating combined with mapping of structural provinces have defined five cratonic areas in South America (Cordani and others, 1973). These are shown in Figure 2.14 and are named: the Guyana Shield, Guapore Craton, São Luis Craton, São Francisco Craton, and La Plata Craton (Cordani and others, 1973). All of these areas contain rocks which were affected by a regional metamorphic episode 1800-2200 m.y. ago (Trans-Amazonian orogeny of Cordani and others, 1973). As shown in Figure 2.14, three of the areas also contain Archean rocks. It seems likely that all of the areas will yield more pre-2500 m.y. dates in the future.

The largest area is the Guyana Shield. This area contains the oldest rocks so far reported from South America--the 2700-3400 m.y. Imataca granulite complex of Venezuela. Other Archean rocks are found to the southeast, in Guyana. The Guyana Shield was interpreted by Hurley and Rand (1973) to be a western extension of the West African craton (Figure 2.14). The southern border of the Guyana Shield is marked by onlapping Phanerozoic sedimentary rocks of the Amazon Basin. Kovach and others (1976) found rocks older than 1600 m.y. in drillcore which penetrated basement of the Amazon Basin and they suggested that the Guyana Shield may be continuous with the Guapore Craton to the south, which yields pre-2000 m.y. dates in its northeastern part (Cordani and others, 1968).

The São Francisco Craton of eastern Brazil contains the only uraniferous conglomerates that we are aware of in South America. Figure 2.14 shows that Archean rocks crop out in the northeast (Hurley and Rand, 1973) and southwest (Herz, 1970) corners of the São Francisco Craton and in a north-trending strip which underlies metasedimentary rocks (Pflug, 1975; Pflug and Scholl, 1975). These metasediments are called the Minas Series to the south

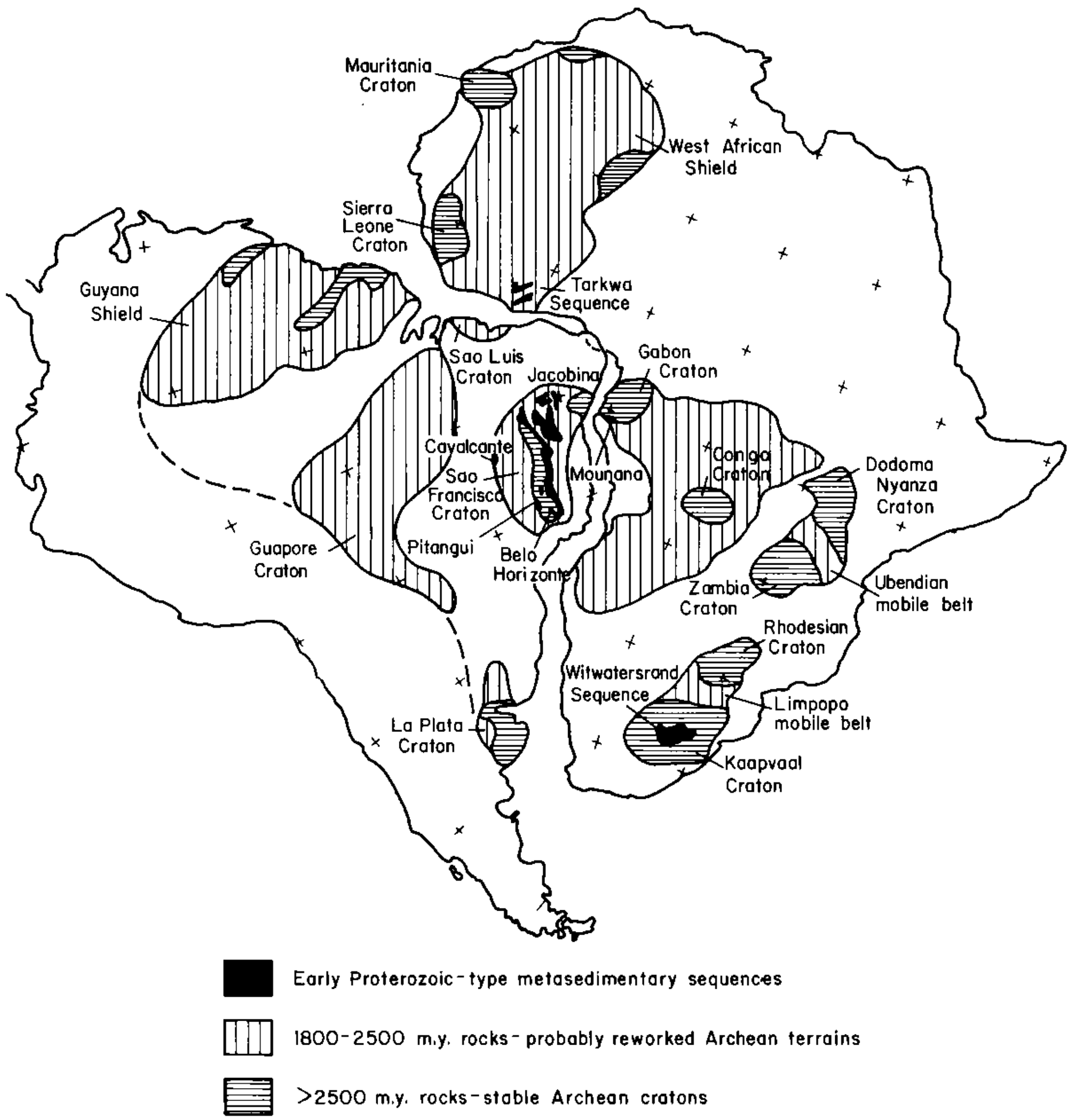

Figure 2.14. Distribution of Archean and Early Proterozoic rocks in Africa and South America, showing locations of Precambrian uranium- and gold-bearing placers; drawn on pre-Mesozoic reconstruction of Bullard and others (1965). Sources of data: Hurley and Rand (1969, 1973); Clifford (1974); Cordani and others (1973); Mason (1973); Anhaeusser (1973); Almeida and others (1973); Pflug and Scholl (1975); Pretorius (1976b); Sestini (1973); Da Costa and Angeiras (1971).

and the Jacobina Series to the north and are between 2700 and 1350 m.y. old. The presence of uraniferous and pyritic quartz-pebble conglomerates near the base of the metasediments suggests an age between 2700 and 2200 m.y. (D. S. Robertson, 1974). Paleocurrent data indicates that the source of Minas Series was to the west (Dorr, 1969) whereas the source of the Jacobina Series was to the east (Gross, 1968). These data, plus the position of the metasediments between Archean age provinces (Figure 2.14) suggests that the metasediments were deposited in an intracratonic basin. This paleogeographic situation may be similar to that of the Witwatersrand Sequence of South Africa.

The other cratonic areas in South America, the São Luis and La Plata Cratons, contain rocks older than 2000 m.y. and appear to be extensions of Precambrian cratonic areas in West and South Africa.

SÃO FRANCISCO CRATON

Jacobina

The best known occurrence of auriferous and uraniferous conglomerate in Brazil occurs at Jacobina, in the northeastern part of the São Francisco Craton (Figure 2.14). In this area, Early Proterozoic rocks crop out in a north-trending, east-dipping fault block which is over 100 km long and is bounded on the east and west sides by granitic rock of probable Archean age (Cox, 1967; Bateman, 1958). Gold has been intermittently mined from the area since the late 17th century.

Metasedimentary rocks in the Jacobina area are called the Jacobina Series (Branner, 1910) and have been divided by Leo and others (1962) and Cox (1967) into four formations shown in Table 2.7. The oldest of these, the Bananeira Formation, consists of pelitic schists and is only locally preserved. It is unconformably overlain by the Serra do Côrrego Formation which consists of auriferous quartz-pebble conglomerate and quartzite. This formation lies against Archean (?) granitic gneisses along the west edge of the Serra de Jacobina. The contact is reported both as an east-dipping thrust fault (Cox, 1967) and as an unconformity (Gross, 1968). Cox (1967) interpreted a quartz-kyanite rock which locally separates the Jacobina Series from the underlying gneisses as a metamorphosed regolith which developed on a weathered Archean surface prior to deposition of the metasediments. He interpreted the contact as a faulted unconformity. The Rio do Ouro Formation conformably overlies the Serra do Córrego Formation and consists of current-bedded quartzite. It is, in turn, overlain by the Cruz das Almas Formation which consists of pelitic schists containing hematite and local layers of iron formation.

Auriferous conglomerates crop out in a prominent ridge on the west side of the Serra de Jacobina. They occur in lenses and layers which vary in thickness from a single pebble to 20 m thick (Gross, 1968). The conglomerates are interlayered with coarse-grained quartzite, sericitic quartzite, and arkose and show abundant evidence of scour and fill structures and cross-bedding (Gross, 1968). Some of the beds of conglomerate are continuous for lateral distances greater than 1 km but most are more lenticular. This lenticular distribution of the conglomerates within a quartzite succession suggests fluvial deposition (Bateman, 1958).

Conglomerates in the Serra do Córrego Formation vary from a clean, arenaceous quartz-pebble conglomerate containing quartz pebbles up to 15 cm in diameter to a dirtier conglomerate with abundant pyrite, sericite, fuchsite and chlorite in the matrix and with quartz-pebbles averaging 3 cm in diameter. All of the conglomerates contain at least small quantities of gold in the matrix. Gross (1968) reported an average of .5 ppm gold in non ore-bearing conglomerates. The dirtier, pyritic conglomerates are significantly

	Name	Lithology	Thickness (m)	Sedimentary Features
Jacobina Series	Cruz das Almas Formation	pelitic schist, micaceous quartzite, conglomerate, local iron formation	>2100	
Jacobina Series	Rio do Ouro Formation	white to green quartzite, minor pelitic schist	2300	well-bedded, ripple marks, cross-bedding
Jacobina Series	Serra do Corrego Formation	quartzite, auriferous quartz-pebble conglomerate	2000	scour and fill structures, cross-bedding, ripple marks
Jacobina Series	Bananeira Formation	pelitic schist, micaceous quartzite	>1000	
	granitic gneiss			

Table 2.7. Stratigraphy of the Jacobina Series. Adapted from Cox (1967) and Gross (1968).

richer, containing an average of 10 ppm gold. Uranium is also more abundant in the pyritic conglomerates which contain an average of .1 percent U_3O_8 and grades up to .6 percent (Gross, 1968). Pyrite ranges from 2-5 percent in the dirtier conglomerates and sericite ranges from 15-25 percent (Bateman, 1968). White (1961) identified uraninite as the main uranium mineral by X-ray diffraction. Other heavy minerals in the matrix of the conglomerates include: chalcopyrite, pyrrhotite, sphalerite, ilmenite, molybdenite, rutile, brannerite, pitchblende, magnetite, tourmaline, and zircon (Ramdohr, 1958, in Cox, 1967; Gross, 1968).

Conglomerates of the Jacobina Series fit the model for fossil-placer deposits very well. They were deposited in a fluvial environment on an eroded Archean gneissic and metasedimentary terrain. The paleogeographic setting also appears favorable. Gneisses of possible Archean age crop out both east and west of the metasediments and paleocurrent data (Gross, 1968) indicate that no source was to the east. This implies that the metasediments were deposited in an intracratonic basin or trough, perhaps similar to the Witwatersrand depositional basin. The age of the Jacobina Series is not known but the rocks are correlated with rocks of the Minas Series to the south which are between 2700 and 1350 m.y. old.

The Jacobina Series differs from the Huronian Supergroup and Witwatersrand Sequence in that it has been metamorphosed to amphibolite facies. As a result, gold and uranium minerals have been remobilized. Gross (1968) cited the occurrence of gold and pitchblende in fracture systems, high concentrations of gold in the top of the conglomerate reefs, and gold at the contacts of mafic intrusives. This evidence was interpreted by Bateman (1958) and White (1961) as an indication of hydrothermal origin of the mineralization. However, Cox (1967), Gross (1968), and most modern workers consider that minor remobilization of the placer heavy minerals accompanied metamorphism and intrusion of mafic sills.

Belo Horizonte

Uraniferous and auriferous conglomerates have been reported by Ramos (1972) from the basal part of the Minas Series at the south edge of the São Francisco Craton near the city of Belo Horizonte (Figure 2.14). This area has been mapped in detail by the United States Geological Survey because of deposits of iron formation in the area. As shown in Table 2.8, the uraniferous conglomerates form the basal part of the Moeda Formation of the Carasa Group and they unconformably overlie the Rio das Velhas Series, an older metasedimentary and metavolcanic sequence. The age of the Minas Series is not well known but a maximum age of 2700 m.y. is given by Rb-Sr ages on muscovite from the underlying Rio das Velhas Series and by Rb-Sr whole-rock ages from basement granodiorites (Herz, 1970). A minimum age for the Minas Series is given by a 1350 m.y. old post-orogenic granite which intrudes the upper Minas Series (Herz, 1970). Closer brackets are questionable. Herz (1970) interpreted a 1930 m.y. date (DTM-6) for porphyritic granodiorite of the Moeda Complex which underlies the Minas Series to be a maximum age for the metasediments. However, he also reported ages ranging from 2300 m.y. (DTM-5) to 462 m.y. (Ha-24) from gneisses of the Moeda Complex in the same area. All the dates are mineral dates and we have no confidence that they reflect the time of crystallization of the granodiorite so we hesitate to use them to define the age of the metasediments. Of the reported dates, the 2300 m.y. date could be primary because it is the oldest one reported in this area and it is consistent with the model that the presence of detrital pyrite and uranium in the overlying conglomerates indicates an age greater than 2000 m.y. (Roscoe, 1969, 1973; D. S. Robertson, 1974).

Series	Group	Name	Lithology	Thickness (m)	Sedimentary Features	Depositional Environment (source)	Age
		Itacolami Series	quartzite, phyllite, conglomerate	2000		deltaic or littoral	
Minas Series		Piracicaba Group	phyllite, schist, greywacke	3500	volcanoclastic sediments	marine turbidites	maximum age of 1350 from intrusive granite
			graphitic phyllite	120			
			quartzite	120	irregular brown cavities		
			dolomite phyllite	410	locally ferruginous	marine	
			ferruginous quartzite	400	locally conglomeratic		
	Itabira Gp	Gandarela Formation	iron formation, dolomite, phyllite	300	locally ferruginous	marine	
		Cauê Itabirite	iron formation (oxide facies)	250	alternating quartz and hematite layers	marine	
	Caraca Gp	Batatal Fm.	fine-grained phyllite	100	pyritic, ferruginous	marine	
		Moeda Formation	quartzite, conglomerate quartz phyllite	100-1000		fluvial-deltaic (W)	
			uraniferous conglomerate		brannerite, Au, zircon in matrix		
Rio das Velhas Series		Tamanduá Gp.	quartzite, phyllite	1200	local iron formation	paralic (W)	
		Maquiné Gp.	quartzite, conglomerate sericitic quartzite	1600		shallow marine (W)	
		Nova Lima Gp.	phyllite, mica schist greenstones	>4000	minor iron fm., greywacke, paraconglomerate, graphite contains Au, Mn deposits	marine	>2700
			gneisses and migmatites				>2800

Table 2.8. Early Precambrian stratigraphy of Minas Gerais, Brazil. Adapted from Dorr, 1969.

Basal conglomerates of the Moeda Formation were deposited on an erosion surface of moderate relief. Dorr (1969) stated that maximum relief was about 50 m and that there were many depressions with relief less than several tens of meters. Guild (1957) suggested that there was a regolith separating the underlying metasedimentary rocks from the Minas Series on the basis of a parallel foliation on both sides of the lower contact of the Minas Series and on the basis of the quartz-rich character of the upper Rio das Vehlas Series rocks. However, Wallace (1965) interpreted the same evidence to be the result of granitization (Herz, 1970).

Two types of conglomerates are present in the Moeda Formation. The first is a polymictic conglomerate which occurs in lenses up to 30 m thick at the base of the Moeda Formation and fills local depressions in the underlying erosion surface. These conglomerates were locally derived and contain angular fragments of the Rio das Vehlas rocks. Gold, which was probably derived from veins cutting carbonaceous iron-formations of the Rio das Vehlas Series is also present in the conglomerates and occurs in sufficient quantities (1-3 ppm) that it was mined in the 18th and 19th centuries (Dorr, 1969). The second type of conglomerate is a greenish-gray quartz-pebble conglomerate which occurs in lenses throughout the lower 300 m of the Moeda Formation.

Recent drill data from several holes greater than 400 m deep is reported by Ramos and Fraenkel (1974). Their preliminary conclusions are: 1) there is a large volume of pyritic conglomerate but grades are low. Some holes showed thicknesses of 100 m with grades of U_3O_8 higher than 10 ppm. Gold grades are generally less than 1 ppm in the conglomerates; 2) mineralized lenses have a maximum thickness of 1.5 m and are not laterally continuous; 3) uranium and pyrite always occur together but there appears to be no direct relationship between amounts of these minerals; 4) there is a direct relationship between amounts of uranium and titanium; 5) $ThO_2:U_3O_8$ ratio in the subsurface ranges from .1 to 2.32 and is about 10 on the surface. This indicates appreciable surface leaching of U_3O_8; and 6) there are at least two generations of pyrite. The first occurs in large (2 cm) euhedra in many different rock types and is probably secondary. The second occurs as fine, disseminated particles in the matrix of conglomerates, is associated with other heavy minerals, and is probably of detrital origin.

The deposition of the Moeda Formation was interpreted by Dorr (1968) and Bain (in Ramos and Fraenkel, 1974) to be fluvial-deltaic on the basis of the lenticular distribution of basal conglomerates in the basement depressions and the intercalated lenses of quartz-pebble conglomerate in coarse-grained quartzite. Dorr (1969) stated that the source of the Moeda Formation was to the west because the coarse lower facies thins to the east and clasts in the conglomerate get smaller to the east.

Units above the Moeda Formation contain phyllite, iron formation, and marine sediments. Dorr considered these rocks to represent a marine transgressive sequence.

The uraniferous conglomerates near Belo Horizonte seem to fit the fossil-placer model very well. They are about the right age, were deposited on an eroded surface by fluvial processes, and are part of an Early Proterozoic sequence of clastic sediments. In some ways, the paleogeographic setting for deposition of the Moeda Formation is similar to that of the Witwatersrand Sequence. In both cases, the eroded basement on which conglomerates were deposited consists of older (gold-bearing) metasedimentary rocks. The Witwatersrand Sequence was deposited on quartzites and volcanics of the Dominion Reef Sequence and most of the detrital gold came from nearly greenstone belts. Similarly, the Moeda Formation was deposited on quartzites of the Maquine and Tamandua Groups and its source of gold appears to be greenstone-like sequences in the Nova Lima Group. Other similarities with the Witwatersrand may include age (the Moeda Formation could be as much as 2700 m.y. old) and deposition in an intracratonic basin at similar paleolatitudes.

The implications of this analogy are twofold. First, the nature of the source terrain for the Moeda Formation suggests that, like the Witwatersrand, gold may be economically more important than uranium in the conglomerates. Second, the units below the Moeda Formation should be investigated for conglomerates bearing detrital gold, such as occurs in the Dominion Reef Sequence. In particular, the Maquine Group appears to be a favorable host for detrital gold because it contains fairly mature clastic sediments which unconformably overlie gold-bearing rocks of the Nova Lima Group.

One unique feature of the Minas Series is that the conglomerates of the Moeda Formation are conformably overlain by a thick section of iron formation (Dorr, 1969). Other known uraniferous conglomerates are in thick quartzite successions which are usually interpreted to be appreciably older than the Early Proterozoic banded iron formations. For example, the Huronian Supergroup is generally thought to be about 200 m.y. older than the main Lake Superior iron formations, although the geochronologic evidence is not definitive (Goldich, 1973). The close temporal and spatial relationship of uraniferous conglomerate and banded iron formation in Brazil indicates that deposition of these two rock types could take place under the same low oxygen atmospheric conditions and that the two rock types were deposited nearly contemporaneously in Brazil. This may provide a partial explanation for the absence of detrital magnetite in the conglomerates. Perhaps magnetite was leached and iron was transported in solution into deeper parts of the depositional basin where it was precipitated as iron formation--at the same time that fluvial processes were concentrating pyrite, uraninite, and gold at the edges of the basin.

Other Areas in the São Francisco Craton

D. S. Robertson (1974) reported another occurrence of uraniferous and auriferous quartz-pebble conglomerate near Pitangui, northwest of Belo Horizonte (Figure 2.14). These conglomerates occur in a succession of the Proterozoic quartzites and arkoses and they unconformably overlie highly deformed schists and gneisses of the Nova Lima Group.

Uraniferous conglomerates were also reported by Ramos (1972, *in* Robertson, 1974). These conglomerates crop out near Cavalcante, Brazil, along the northwest boundary of the São Francisco Craton (Figure 2.14). The conglomerates overlie Archean (?) mica-schists and are correlated with the Minas Series of the Belo Horizonte area.

Conglomerates containing fossil-placer mineral deposits also may occur elsewhere in the São Francisco Craton. Figure 2.14 shows that quartzites which are believed to be correlative with the Minas Series and the Jacobina Series extend in a 1200 km long strip in the central part of the craton. In one area, near Diamondtina, intraformational polymictic boulder conglomerates containing diamonds were reported by Pflug and Scholl (1974). They interpret these conglomerates to have formed during a tectonic episode which interrupted a long period of deposition of shallow marine quartzites. They correlate the quartzites and conglomerates with the Minas Series. Quartzites like these should be investigated for heavy minerals in areas where they unconformably overlie either granitic rocks or older metasediments.

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INDIAN SHIELD

GEOLOGIC SETTING

Archean and Early Proterozoic rocks crop out in three areas in India (Figure 2.15). The largest and best studied area is in southern India and is called the Karnataka (or Dharwar) craton. The other two areas are in northeastern and northwestern India and have been named the Singhbhum and Aravalli cratons respectively (Naqvi and others, 1974). The cratons are separated by areas covered by Paleozoic Gondwana Group sediments, Cretaceous-Eocene Deccan flood basalts, and Quaternary alluvium. These areas conceal deep-seated crustal features which Naqvi and others (1974) interpreted as rift valleys located in zones of Early Proterozoic faulting, perhaps similar to the mobile belts of southern Africa. The Karnataka craton includes, in the south, the South Indian mobile belt, an area of granulite grade gneisses, charnokites, mafic rocks, and carbonatites.

All three cratonic blocks in India contain remnants of rocks older than 3000 m.y. (Crawford, 1969; Sarkar, 1968; Naqvi and others, 1974), and, as on other continents, it is not clear whether the oldest rocks are ultramafic and mafic volcanics or tonalitic gneisses. In Karnataka, high-grade schists of the Sargur Group are dated as 3010 $\pm$ 90 m.y. (Jayarem and others, 1976) and are cross-cut by gneisses dated as 2950 $\pm$ 150 m.y. (Radhakrishna and Vasudev, 1977). This thermal event at 3000 m.y. also affected gneisses on the other Indian cratons.

Archean greenstone belts, similar to the Keewatin greenstone belts in the Canadian Shield, crop out in north-trending arcuate belts in the eastern part of the Karnataka craton (Viswanatha and Ramakrishnan, 1975). These greenstone belts are dominated by mafic and intermediate volcanics, with subordinate felsic volcanics, graphite schist, chert, and iron formation

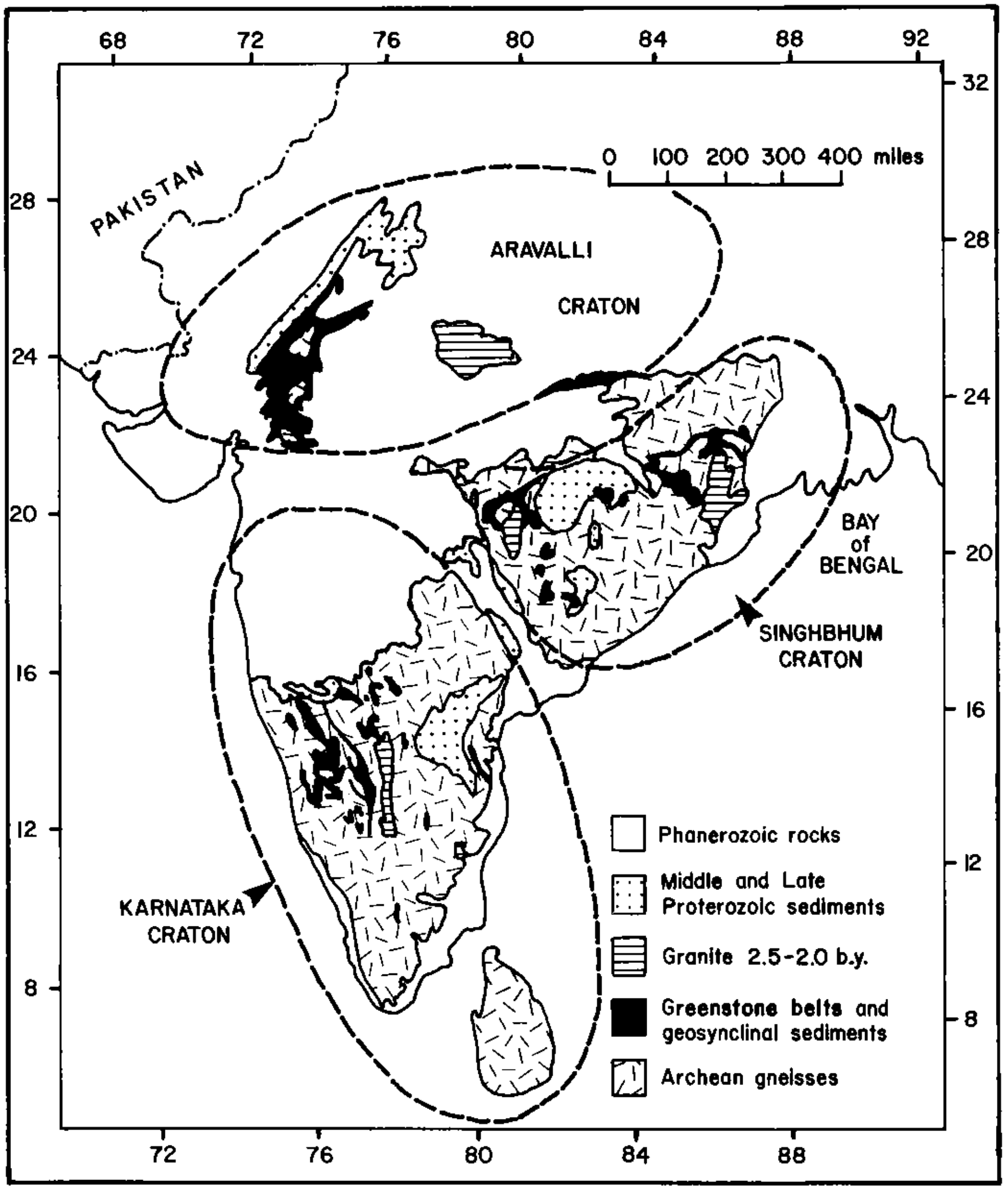

Figure 2.15. Simplified geological map of India showing distribution and strike of Archean and Proterozoic rocks and Archean protocontinents. Adapted from Naqvi and others (1974, p. 355).

and they contain important gold deposits associated with quartz veins (Radhakrishna and Vasudev, 1977). The greenstone belts are surrounded by granitic gneisses and are intruded by granites which Ramakrishnan and others (1976) suggest were mobilized during a major thermal episode 2500-2600 m.y. ago.

The best studied Early Proterozoic metasedimentary rocks crop out in the western part of the Karnataka craton and are called the Dharwar Supergroup. These rocks are metavolcanics and metasedimentary rocks which include thick quartzite successions with auriferous and uraniferous basal conglomerates. Most Indian geologists consider the rocks of the Dharwar Supergroup to be Archean and to be the same age as the Keewatin-type greenstone belts in eastern Karnataka (Radhakrishna and Vasudev, 1977). However, the fact that these sequences contain thick cross-bedded quartzites and uraniferous basal conglomerates which lie unconformably on Archean gneiss suggests that the Dharwar Supergroup is a clastic platform sequence more closely analogous to the Early Proterozoic rocks of the Witwatersrand Sequence and the Huronian Supergroup than to any known Archean greenstone belts.

Radiometric dates from the Dharwar Supergroup permit either interpretation. The only date from the metasediments is a questionable 2345 $\pm$ 60 m.y. date for lavas of the "Dharwar Schist" which was obtained by Crawford (1969) by selecting only favorable points. Bracketing ages are not much better. A maximum age of 2500-2700 m.y. comes from dates on gneisses of the Peninsular Gneiss complex on which basal conglomerates rest unconformably (Radhakrishna and Vasudev, 1977). A minimum age is obtained from the Closepet granite which intrudes the Dharwar Supergroup. Crawford (1969) obtained an age of 2380 m.y. on this granite whereas Venkatasubramanian (1974) obtained an age of 2000 $\pm$ 80 m.y. Thus, all existing evidence suggests that the

Dharwar Supergroup was deposited between 2700 and 2000 m.y., during the same interval that Early Proterozoic platform sequences were being deposited elsewhere in the world. There seems to be little justification for Windley's statement (Windley, 1977, p. 30) that the auriferous and uraniferous conglomerates of the Dharwar Supergroup "are the earliest conglomerates of this type recorded from any continent."

Problems in the Precambrian Stratigraphy of India

Despite nearly one hundred years of geological investigations, there are still major gaps in the understanding of Archean and Early Proterozoic stratigraphy in the Indian Shield. These gaps are largely the result of confused terminology, a lack of detailed field mapping of stratigraphical relationships, and a paucity of radiometric dates.

The major problem has been in the use of terms "Dharwar Schist" and "Peninsular Gneiss." According to Radhakrishna (1974), the term Dharwar Schist has been used "to include all schistose rocks based purely on their supposed lithologic similarity." Thus all schists, regardless of age, have been called Dharwar Schists. Similarly, the term Peninsular Gneiss has been used to refer to a whole complex of granitic gneisses and intrusive granites of various ages which enclose the Dharwar Schist belts (Figure 2.15). The use of the terms Dharwar Schist and Peninsular Gneiss in this purely lithologic context, instead of in a stratigraphic context, has resulted in numerous arguments among Indian geologists concerning the relative age of schists and gneisses.

Many of the early geologists thought that most of the Dharwar schists were younger than the Peninsular Gneiss (Foote, 1888; Oldham, 1893; Holland, 1907; Middlemiss, 1917). Their evidence was that: 1) Dharwar Schists rest with angular unconformity on Peninsular Gneisses locally, 2) there is a conglomerate at the base of the Dharwar Schists locally, and 3) top and bottom criteria in the schists consistently show the schists on top of the gneisses. This view is upheld by Radhakrishna, Ramakrishnan, Viswanatha, Swami Nath, and other members of the Geological Survey of India.

The opposing idea, that most of the Peninsular Gneisses are younger than most of the Dharwar Schists was brought forward to explain the schistose inclusions in the gneisses which are lithologically similar to and were thought to have been derived from the Dharwar Schists (Newbold, 1850; Smeeth, 1901; Fermor, 1936). Additional evidence for this interpretation is that there are areas where granitic gneisses cross-cut the schist belts (Srinivasan and Screenivas, 1972). Proponents of this view include Rama Rao (1962), Pascoe (1965), and Pichamuthu (1967).

The most reasonable interpretation of available evidence is that the Peninsular gneiss complex contains both intrusive granites which are younger than the Dharwar Supergroup and basement gneisses which are older (Srinivasan and Screenivas, 1972; Naqvi and others, 1974). Therefore, the solution to the arguments must involve detailed field mapping aimed at subdividing the Dharwar schists and Peninsular gneisses into correlatable lithostratigraphic units. This approach has been initiated by the Geological Survey of India (Viswanatha and Ramakrishnan, 1975; Ramakrishnan and others, 1976; Radhakrishna and Vasudev, 1977). These reports have subdivided the schists into several groups, shown in Table 2.9.

EARLY PROTEROZOIC METASEDIMENTARY ROCKS

Karnataka Craton

Figure 2.16 shows the distribution of Early Proterozoic rocks in the Karnataka craton and Table 2.9 summarizes the stratigraphy of the metasedimentary groups. This stratigraphic subdivision is based on recent (1975-1977)

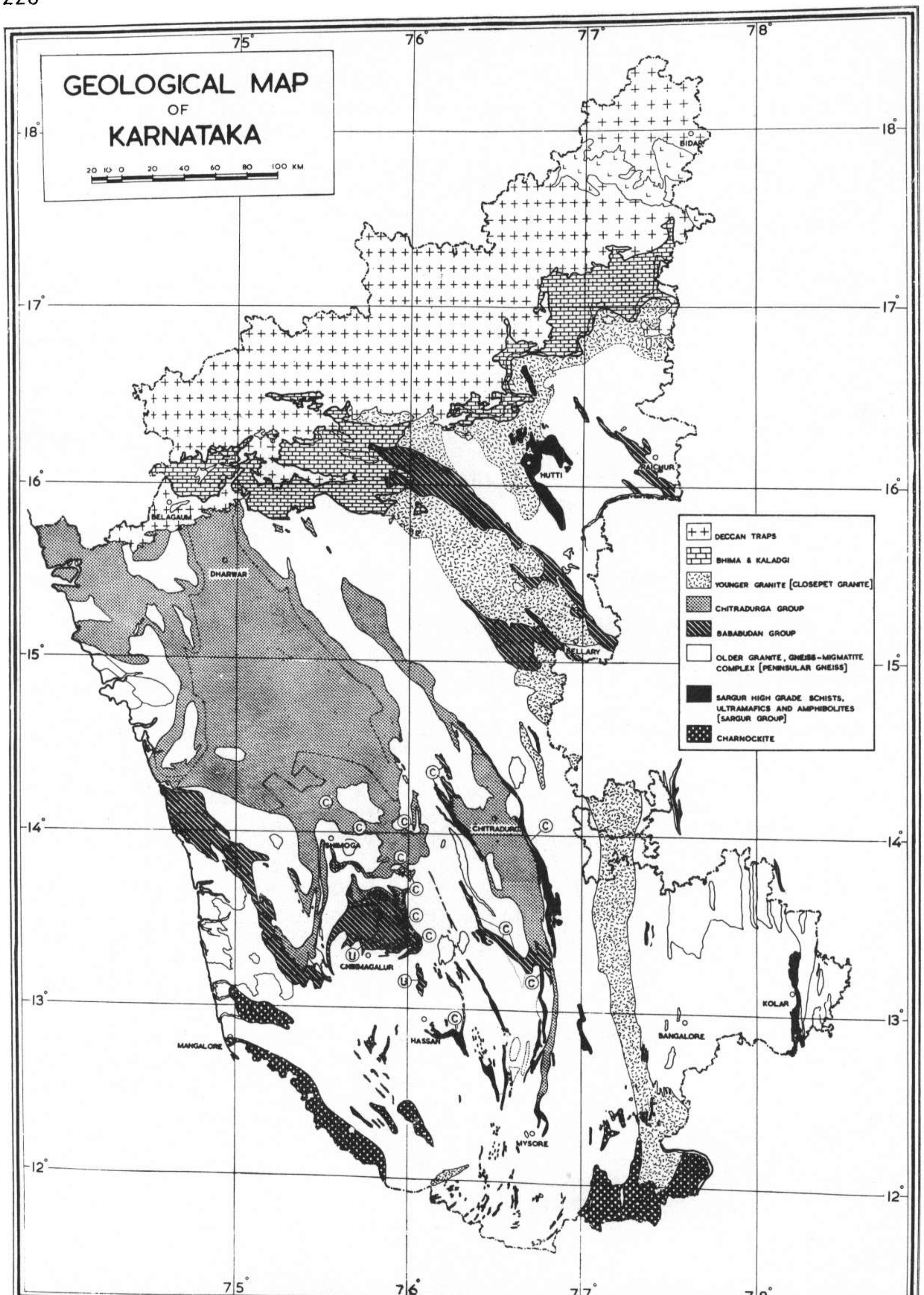

Figure 2.16. Geological map of the Karnataka Craton showing locations of uranium-bearing conglomerates U and other conglomerates with potential for fossil-placer mineralization C . Adapted from Radhakrishna and Vasudev (1977).

	Name	Lithology	Mineralization	Age
Cuddapah System		argillite, quartzite, limestone		<1700
Closepet Granite		porphyritic microcline granite		2000-2380
DHARWAR SUPERGROUP	Ranibennur Group	manganiferous phyllite ankeritic limestone greywacke chlorite phyllite	manganese	
	Chitradurga Group	greywacke, phyllite pillow basalt, iron formation iron formation, phyllite manganiferous phyllite limestone, dolomite phyllite, quartzite polymictic, oligomictic congl.	 iron iron manganese placer gold	2350?
	Bababudan Group	carbonate, graphite, phyllite magnetitic quartzite argillite, felsic volcanics mafic volcanics cross-bedded quartzite mafic volcanics quartzite, quartz-mica schist oligomictic conglomerate	 iron U, Au	
granites and granodiorites of Peninsular Gneiss				2500-2700
Keewatin-type greenstone belts		mafic and intermediate volcanics	gold in quartz-veins	
tonalites and gneisses of Peninsular gneiss		tonalite-migmatite, amphibolite, granitic gneiss		2600-3250
Sargur Group		amphibolite iron-formation garnet amphibolite mica schist calc-silicates mica-schist, paragneiss quartzite, quartz-mica schist	 iron gold in quartz veins?	>3000
Older gneiss?		migmatites		3000

Table 2.9. Early Precambrian stratigraphy of the Karnataka Craton. Adapted from Viswanatha and Ramakrishnan (1975); Radhakrishna and Vasudev (1977).

reports of the Geological Survey of India which divide the Dharwar schists into an older group of high-grade schists called the Sargur Group (Viswanatha and Ramakrishnan, 1975), a group of Keewatin-type greenstone belts dominated by metavolcanic rocks, and a sequence of metasedimentary and metavolcanic rocks called the Dharwar Supergroup.

This stratigraphic subdivision is new and there is still confusion regarding which schist belts belong to which sequence of metasediments. According to Ramakrishnan and others (1976), rocks of the Sargur Group and the Dharwar Supergroup are confined to western Karnataka and Keewatin-type greenstone belts are confined to eastern Karnataka (Figure 2.16). In contrast, Radhakrishna and Vasudev (1977) considered some of the schist belts in eastern Karnataka to be part of the Sargur Group and some to be Keewatin-type greenstone belts and they equated the Keewatin-type greenstone belts in eastern Karnataka with the Bababudan Group in western Karnataka. However, this second interpretation obscures the important distinction between the metavolcanics-dominated greenstone belts and clastic metasedimentary sequences of the Dharwar Supergroup. In this discussion, the Keewatin-type greenstone belts are considered to be older than the metasedimentary sequences of the Dharwar Supergroup (Table 2.9).

The Dharwar Supergroup is subdivided into three groups, shown in Table 2.9. The Bababudan Group includes basal uraniferous quartz-pebble conglomerate, vesicular mafic volcanics, quartzite and iron formation and was deposited in presumed subaerial to shallow water conditions. The Chitradurga Group is a miogeosynclinal conglomerate, quartzite, phyllite, stromatolitic limestone, iron formation, pillow basalt, and volcaniclastic graywacke. The Ranibennur Group contains thick sections of graywacke and chlorite phyllite.

As shown in Table 2.9, conglomerates of the Bababudan Group rest unconformably on granites and gneiss of the Peninsular gneiss complex. In the Bababuden schist belt, the conglomerates trend E-W and truncate the north-trending foliation in the Chikamagular granite. Similarly, N-S trending conglomerate in the Sigegudda schist belt is discordant to the N-W trending underlying migmatites (Ramakrishnan and others, 1976). According to Ramakrishnan and others (1976) field relationships in these and other areas are unequivocal evidence that rocks of the Peninsular gneiss complex formed the basement on which conglomerates of the Bababudan Group were deposited.

The basal conglomerate occurs in layers and lenses 5-25 m thick which are intercalated with quartzites and quartz-sericite-fuchsite schists. Pebbles in the conglomerate are well rounded and are composed of vein quartz and quartzite. The matrix of the conglomerate contains sporadic disseminations of pyrite, chalcopyrite, gold, silver, and uranium minerals. According to Viswanatha and others (1977) "significant concentrations of chalcopyrite and uranium ores are known in the Kalasapura and Devagondonahalli areas."

The basal conglomerate and quartzite of the Bababudan Group is overlain by amygdaloidal metavolcanics which are interlayered with quartzites. Cross-bedding and ripple marks in the quartzites and amygdules in the volcanics suggest subaerial to shallow water deposition. The upper part of the Bababudan Group contains iron formation and phyllite. The iron formation is a magnetitic quartzite 200 m thick which is of economic importance.

The Chitradurga Group is a miogeosynclinal sequence which has a higher proportion of sediments than the underlying Bababudan Group. The basal unit is a polymictic conglomerate called the Kaldurga conglomerate which shows graded bedding and contains rounded and moderately-sorted granite clasts

(Radhakrishna and Vasudev, 1977). This conglomerate grades both laterally and vertically into an oligomictic quartz-pebble conglomerate (Ramakrishnan and others, 1976) which may be a good target for exploration for fossil-placer uranium and gold deposits. According to Viswanatha and others, 1977, the basal conglomerate indicates rapid tectonic uplift during early deposition of the Chitradurga Group. The conglomerate is overlain by a transgressional sequence of quartzite, phyllite and limestone followed by manganiferous and ferruginous cherts, volcanics, and volcaniclastic sediments. The overlying Ranibennur Group contains mainly volcaniclastic phyllites and graywackes.

The three groups of the Dharwar Supergroup represent the same trends in Early Proterozoic sedimentation which are seen elsewhere in the world. First, an Archean granite-greenstone terrain was stabilized, uplifted, and eroded. Then, a period of subaerial fluvial deposition and sporadic subaerial volcanism ensued which deposited rocks of the Bababudan Group. This was followed by a broad trend of marine transgression accompanied by a decrease in volcanic activity which resulted in deposition of shallow marine miogeosynclinal sediments, including limestones and iron formation of the Chitradurga Group. Continued transgression resulted in deposition of deeper water sediments of the Ranibennur Group.

Singhbhum Craton

Early Proterozoic rocks in the Singhbhum Craton are called the Iron-ore Series (Jones, 1922) because of the presence of economically important iron formation. This series lies unconformably above an assemblage of schists which is probably of Archean age and may be correlative with the Sargur Group of the Karnataka Craton. A minimum age for the Iron-ore Series is 2038 m.y., which is the age of the intrusive Singhbhum granite (Pichamuthu, 1967).

The Iron-ore Series contains, in ascending order, conglomerate, sandstone, mafic volcanics, shales, hematitic iron formation, and shales (Sarker and Saha, 1962, in Pichamuthu, 1967. These rocks are essentially unmetamorphosed. The predominance of iron-formation and shales in the Iron-ore Series suggests that the Iron-ore Series may be correlatable with the Chitradurga Group of the Karnataka Craton. To our knowledge, no uraniferous quartz-pebble conglomerates have been reported from the Singhbhum Craton.

Aravalli Craton

Early Proterozoic rocks in the Aravalli Craton are called the Aravalli Group. These metasediments appear to rest unconformably on a granitic basement which Crawford and Compston (1970) dated as 2585 m.y. old and they are intruded by granites which yield a questionable date of 2275 m.y. (Crawford and Compston, 1970).

The lower units of the Aravalli Group, in areas where they are not migmatized, include thin quartzites and conglomerates. These are overlain by argillaceous rocks and limestone which, in turn, are overlain by quartzites and grits of the upper Aravalli Group. These quartzites grade up into calcareous rocks of the Raialo Group (Pichamuthu, 1967).

MINERAL DEPOSITS

Conglomerates of the Bababudan Group of the Dharwar Supergroup are the only known rocks in India which contain uranium- and gold-bearing conglomerates. These rocks fit the fossil placer model well. They were deposited on an eroded Archean terrain which contained sources for both gold (greenstone belts) and uranium (granites), and they appear to be fluvial deposits of about the right age (2700-2000 m.y.).

Another target for Precambrian fossil placer minerals in the Karnataka

Craton is the conglomerate unit at the base of the Chitradurga Group. This is mainly a polymictic conglomerate but also contains lenses of oligomictic conglomerate. The conglomerates were deposited unconformably on the Bababudan Group and are the basal deposits in a transgressional sequence of quartzite-phyllite-limestone. This suggests that the conglomerates may be fluvial deposits. The conglomerates contain gold (Radhakrishna and Vasudev, 1977, p. 538) and should be investigated in detail, particularly in areas where the Chitradurga conglomerates crop out close to a known source of gold and uranium such as the Bababudan conglomerate (e.g. south of Chitradurga).

Early Proterozoic rocks in the Singhbhum and Aravalli Cratons are broadly contemporaneous with the Dharwar Supergroup. However, the rocks in both of these areas are predominantly marine sediments such as iron-formation, limestone, and phyllite with only minor quartzite. Therefore, the depositional environments of the majority of the metasediments apparently were not favorable for concentration of heavy minerals. Nevertheless, these metasedimentary sequences should be investigated more closely in areas where they unconformably overlie Archean basement.

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AUSTRALIAN SHIELD

GEOLOGIC SETTING

The main tectonic provinces in Australia are shown in Figure 2.17. These are: the Archean cratons of western Australia, Early Proterozoic supracrustal rocks of western and northern Australia, Middle Proterozoic infracrustal rocks in central and southern Australia, and Phanerozoic rocks of the Tasmanian orogenic province in eastern Australia (Rutland, 1973; Fisher and Warren, 1975). Early Proterozoic metasedimentary platform sequences occur adjacent to the Archean cratons in western Australia, and in isolated foldbelts in northern Australia, and these areas will be emphasized. More general discussions of the geology of Australia, including discussions of the Tasmanian province and the Middle Proterozoic mobile belts, are in Rutland (1973; 1976), Knight (1975), Parkin (1969), Brown and others (1968), Sprigg (1967), McWhae and others (1956), and David (1950).

Archean rocks crop out in the Yilgarn and Pilbara cratons which, together, make up the Western Australian Shield (Gee, 1975). Archean rocks also crop out in a few isolated basement domes in northern Australia. In addition, geophysical and sedimentological data (Gravity map of Australia; Gellatly, 1971) suggest that Archean rocks underlie most of western and northern Australia.

The Western Australian Shield contains two Archean cratons separated by Proterozoic sedimentary rocks (Figure 2.17). The Yilgarn craton to the south is the larger of the two and consists of two provinces. The southwestern province contains granulite-grade metamorphic rocks of the Wheat Belt which yield dates ranging from 2800 to 3100 m.y. (Arriens, 1971). These rocks are mainly gneisses and migmatites and are described by Wilson (1969) and Glikson and Lambert (1973). The northeastern province contains linear,

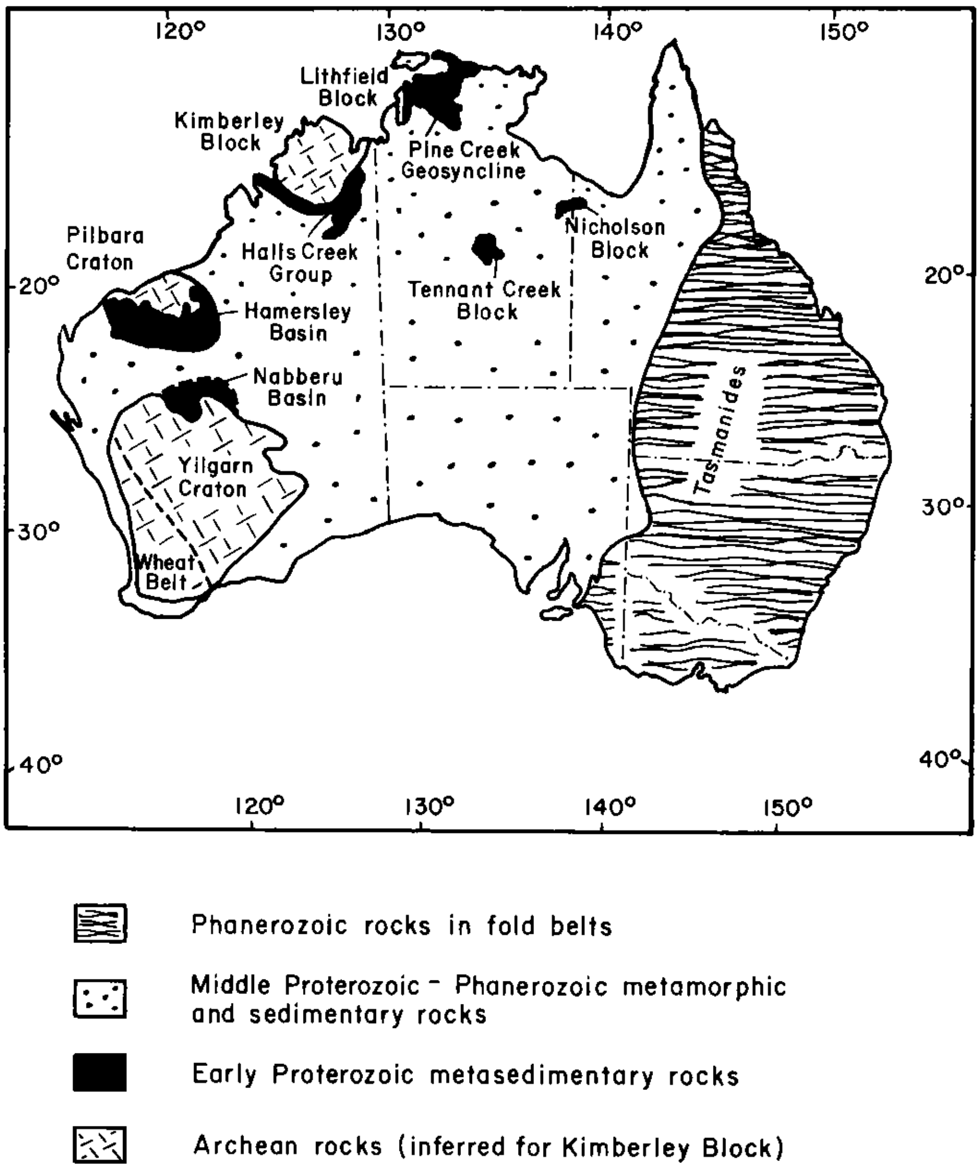

Figure 2.17. Distribution of Archean and Early Proterozoic rocks in Australia. Adapted from Rutland (1976).

low-grade greenstone belts surrounded by intrusive granites which yield 2600-2700 m.y. dates (Compston and Arriens, 1968). The greenstone belts contain basic and intermediate volcanics and volcanogenic sediments such as chert, graphitic schist, iron formation, and conglomerates as well as extensive gold and sulphide mineralization in quartz veins. The greenstone belts are discussed by Glikson and Lambert (1973), McCall (1971), Rutland (1976), and Gee (1975). According to Rutland (1973; 1976), the greenstone terrain of the northeastern Yilgarn craton is broadly synclinal and may be younger than the bordering granulite terrain. However, this view is disputed by Glikson and Lambert (1973; 1976) who interpreted the granulite terrain to have formed at deeper crustal levels at about the same time as the greenstone terrain.

The Pilbara craton is also a granite-greenstone terrain. However, it differs from the northwestern Yilgarn craton in several respects. First, most of the granitic plutons are 2900-3100 m.y. old instead of 2600-2700 m.y. This suggests that the greenstone belts may be significantly older in the Pilbara craton. Second, the Pilbara greenstone belts are unconformably overlain by Archean platform-type sedimentary successions containing iron formation and clastic sediments which may have been deposited at the same time as the greenstone belts in the Yilgarn craton to the south. Third, late Archean (2500-2700 m.y.) granites in the Pilbara craton are post-orogenic and equant in plan view whereas the Yilgarn granites are syn-orogenic and elongated. This supports the previous differences in suggesting that the Pilbara craton was stabilized before the Yilgarn craton, a situation which may be analogous to that of the Kaapvaal and Rhodesian cratons in South Africa, where the former was stabilized about 500 m.y. before the latter.

Archean rocks are also known from northern Australia, where they form the basement for Middle Proterozoic sediments of the Kimberley Block and crop out in isolated inliers of the Rum Jungle Complex and the Litchfield Block (Figure 2.17). These granites are about 2400-2500 m.y. old and have a high initial Sr^{87}/Sr^{86} ratio, which suggests that they represent remobilized older crustal material (Richards and others, 1966; Richards and Rhodes, 1967).

In terms of fossil-placer deposits, the Yilgarn Archean craton appears to be a very favorable source. It contains significant gold mineralization in greenstone belts and an abundance of late Archean potassium- and silica-rich granites to supply uranium. Unfortunately, with the possible exception of the Nubberu Basin of Hall and Goode (1975), there are no major Early Proterozoic platform metasediments adjacent to the Yilgarn craton. The Pilbara craton was, perhaps, a slightly less favorable source area. It does contain gold-bearing greenstone belts (Woodall, 1975), but these are minor compared to the abundant gold areas in the Yilgarn craton. Also, the majority of the granitic rocks in the Pilbara craton are older, sodium-rich granites which are less likely to supply appreciable uranium than younger, potassium-rich granites. Nevertheless, the Pilbara craton apparently did supply both gold and uranium to basal clastics of the Fortesque Group, which crops out on the south margin of the craton, but values reported so far are generally uneconomic. The granites of the Rum Jungle Complex are also a favorable source for uranium and contain up to 28 ppm. These granites may have been one source for the uranium now concentrated in the unconformity deposits of the Pine Creek Geosyncline (Heier and Rhodes, 1968).

EARLY PROTEROZOIC METASEDIMENTARY ROCKS

Early Proterozoic metasedimentary rocks in Australia are called the Nullaginean System. These rocks occur as relatively undeformed platform assemblages only in Hamersley Basin. In all other areas of Australia, Early

Proterozoic rocks crop out in foldbelts which are characterized by amphibolite facies metamorphism and open, upright folds with steeply inclined axial plane cleavage (Walpole and others, 1968). The deformation which formed these foldbelts was associated with intrusion of granites and took place about 1800-2000 m.y. ago, giving a minimum age for the metasediments.

Hamersley Basin

The most likely target for fossil-placer mineralization is the relatively undeformed metasedimentary rocks of the basal units of the Mount Bruce Supergroup in the Hamersley Basin. These rocks are best exposed on the north side of the basin. However, isopachs show that the basin was a closed, intracratonic basin (Ryan and Blockley, 1965), and outcrops of the Mount Bruce Supergroup have been reported from the southern part of the basin as well (Hall and Goode, 1975). The degree of deformation increases to the south so the northern edge of the basin is probably the most favorable area for preservation of fossil-placer uranium deposits.

The Mount Bruce Supergroup is more than 12 km thick and is subdivided into the Fortesque Group (4.3 km), Hamersley Group (2.5 km), and the Wyloo Group (> 5 km). The Fortesque Group consists of epicontinental volcanics and clastic rocks and crops out along the south margin of the Pilbara craton. According to Robertson (1974), the basal unit in the Fortesque Group is a conglomerate-arkose unit which occurs in depressions in the Archean erosion surface. Near Nullagine, these conglomerates are polymictic boulder conglomerates that contain pyrite and rounded nuggets of gold (Robertson, 1974). The conglomerate-arkose unit is overlain by about 1000 m of andesitic volcanics which yield a Rb-Sr age of 2200 m.y. (Arriens, in Trendall, 1968). The next unit in the Fortesque Group is the Glen Herring Formation which contains 1500 m of argillite with interbeds of arenite. This is overlain by the

Green Hole Formation which consists of coarse clastics, including poorly-sorted, uranium-bearing quartz-pebble conglomerates in layers several meters thick which are laterally continuous over distances of 100 km. The pebbles are mainly vein quartz 1-10 cm in diameter and the matrix is greenish-gray arkose with pyrite, monazite, and zircon on bedding planes. Robertson (1974) reported average uranium values of 50-100 ppm in this unit with a value of 425 ppm over a two meter interval reported from one drill core. In this core, the uranium was associated with thucholite, pyrite, monazite, zircon, and anatase. The Green Hole Formation is overlain by a thick sequence of interbedded basic volcanics and dolomite.

Above the Fortesque Group is the Hamersley Group, which is about 2.5 km thick and includes a 1000 m thick section of economically important hematitic banded iron formation. Volcanics from the Hamersley Group are dated as 2000 m.y. old (Arriens, 1971). Discussions of the Hamersley Group are in Trendall (1968; 1973), Trendall and Blockley (1970), and Macloed (1966).

The Wyloo Group overlies the Hamersley Group with local unconformity. It is about 10 km thick and consists of red-beds and coarse-grained terrigenous clastics. The Wyloo Group is cross-cut by granites dated as 1700 m.y. old.

Nabberu Basin

The Nabberu basin is on the northeast margin of the Yilgarn Craton (Figure 2.17) and is considered by many workers to be the southern extension of the Hamersley basin (Hall and Goode, 1975; Horwitz and Smith, 1978). It contains Early Proterozoic clastic metasediments which lie unconformably on Archean rocks and which underlie banded iron formation similar to the Hamersley Group to the north. No uraniferous conglomerates have been reported but quartz-pebble conglomerates occur within a 440 m thick section of otherwise fine-grained clastics called the Malmac Formation (Horwitz and Smith, 1978), and within a coarser-grained, but thinner, succession of sandstones which crop out in a similar stratigraphic position to the south (Hall and Goode, 1975). These quartz-pebble conglomerates, and correlative units, would appear to be favorable hosts for either uranium or gold placers (the Yilgarn craton is a good source of both metals) if the depositional environments were favorable for placer accumulation.

Halls Creek Province

The Halls Creek Province contains metasediments and metavolcanics of the Halls Creek Group which crop out in foldbelts on the southwest and eastern margins of the Kimberley Block (Figure 2.17). These rocks were deformed and intruded by granites about 1960-2100 m.y. ago (Bofinger, 1967) and they are cross-cut by a pegmatite which yielded an age of 2700 m.y. (Plumb and Derrick, 1975). If this age on the pegmatite is correct, the Halls Creek Group is Archean. However, most workers consider the rocks to be Early Proterozoic in age (Rutland, 1973, 1976; Plumb and Derrick, 1975; Page and others, 1976). For example, Gellatly (1971) suggested that the Archean rocks occur in the subsurface below Carpentarian (1800-1400 m.y.) strata and that these Archean rocks may have been the basement on which the Halls Creek Group was deposited. This is supported by Dow and Gemuts (1969) who suggested that the Halls Creek Group is an Early Proterozoic sequence which originally was much more widely distributed than now.

The Halls Creek Group is about 11 km thick and is essentially a eugeosynclinal succession composed of basic volcanics passing up-section into graywacke and turbidites. The base of the group is not exposed but the lowest unit is a 300 m thick basic volcanic unit containing copper mineralization. This unit is overlain by a 200 m thick sequence of quartzite and quartz-graywacke which reportedly contains uranium (Plumb and Derrick, 1975). No details of this uranium mineralization were reported. Above the quartzite is more than 10 km of turbidites which contain minor interbeds of conglomerate and dolomite and which are hosts to gold mineralization (Plumb and Derrick, 1975).

The Halls Creek Group is unconformably overlain by volcanics of the Lamboo complex (1800-1900 m.y.) and these, in turn, are unconformably overlain by metasediments of the Carpentarian System (1800-1400 m.y.) in the Kimberley Basin to the north.

Within the basal units of the Carpentarian system is the radioactive King Leopold conglomerate of the Kimberley Group. This conglomerate contains pebbles and cobbles of quartz, quartzite, and chert, and occurs in lenses about one meter thick within sections of reddish coarse-grained arkose and grit. The conglomerate is interpreted by Hughes and Harms (1975) to be of fluvial origin and to have formed near a marine shoreline. The finer-grained conglomerates contain predominantly quartz pebbles and are notably radioactive. They typically contain 20 ppm uranium and 550 ppm thorium with values up to 100 ppm uranium and 1000 ppm thorium. These high Th/U values persist at depth and do not represent significant surface leaching of uranium. The radioactive minerals are thorogummite, florencite, thorite, and cheralite, and they are interpreted by Hughes and Harms (1975) to be detrital in origin. Other heavies include ilmenite, zircon, leucoxene, monazite, magnetite, hematite, and cassiterite.

According to our model, these conglomerates are certainly too young to contain significant fossil-placer uranium deposits. Nevertheless, the processes of concentration of heavy minerals in this conglomerate appear to be analogous to those of older fossil-placer conglomerates; the unconformity and fluvial deposition in basement depressions are both familiar features. Thus, the conglomerates could contain significant thorium or gold deposits. Paleocurrent data from Gellatly and others (1970) indicate that the source of the conglomerate was to the north, an area now concealed by Carpentarian cover rocks. This source probably was mainly an Archean granitic terrain, but it is also possible that uranium and thorium were derived from concealed Early Proterozoic platform-type metasedimentary successions on the Kimberley Block.

Pine Creek Geosyncline

The Pine Creek Geosyncline is a large area of deformed greenschist to upper amphibolite facies Early Proterozoic metasedimentary rocks in northern Australia (Figure 2.17). Metasedimentary rocks unconformably overlie Archean rocks which crop out in gneiss domes of the Rum Jungle complex and Litchfield Block (Walpole and others, 1968; Plumb and Derrick, 1975). The Rum Jungle complex is 2400-2500 m.y. old (Richards and Rhodes, 1967) and this provides a maximum age for the metasediments. Intrusive granites, 1944 m.y. old (Webb, in Plumb and Derrick, 1975), provide a minimum age.

The Early Proterozoic stratigraphy of the Pine Creek Geosyncline has been studied in detail by Walpole and others (1968) and Smart and Wilkes (1975). The most prominent features of the stratigraphy of the Pine Creek Geosyncline are the rapid facies changes within the depositional basin which are interpreted by Walpole and others (1968) to be a reflection of northwest trending tectonic zones and elongated depositional basins.

The lowest unit in the Pine Creek rocks is the Stag Creek mafic volcanic sequence which crops out in the central part of the basin. This unit is overlain by coarse-grained immature arkose of the Mount Partridge

Formation in the eastern trough and by argillaceous arenites of the Masson Formation in the western trough. In the western fault zone, the equivalent strata are alternating arkose and dolomite of the Batchelor Group which unconformably overlie Archean rocks of the Rum Jungle complex. The Batchelor Group and Masson Formation are gradationally overlain by pyritic dolomite and pelitic rocks of the Golden Dyke Formation and the Mount Partridge Formation is gradationally overlain by dolomites and pelitic rocks of the Koolpin Formation. The Golden Dyke and Koolpin Formations are the hosts of the major unconformity-type uranium deposits in northern Australia. Continued marine deposition of the South Alligator Group in the eastern trough was accompanied by tectonic movements and deposition of turbidites of the Finiss River Group in the western fault zone. The final sedimentation in the Pine Creek Geosyncline was deposition of the Chilling Sandstone in the western margin of the basin.

The metasediments were metamorphosed and deformed about 1800-1900 m.y. ago. Deformation was most intense in the eastern part of the basin where metamorphism caused anatexis and migmatization of sediments in the Nananbu and Nimbuwah metamorphic complexes. This orogenic deformation was followed by platform deposition of the Carpentarian Sequence between 1800 and 1400 m.y. ago.

Uranium deposits occur in dolomite of the upper Batchelor Group, in black pyritic and carbonaceous shales of the Koolpin Formation, and in black shale and chlorite schist of the Golden Dyke Formation. The salient characteristics of these deposits are as follows: they are localized in specific lithographic units (black shales), they are adjacent to granitic complexes, the deposits occupy fractures, folds, and shear zones, the deposits are usually less than 50 m below the Carpentarian unconformity, and the most important deposits are adjacent to high-grade metamorphic complexes such as the Namabu and Nimbuwah complexes (Dodson and Prichard, 1975; Needham and Smart, 1972). It seems clear that the uranium deposits had a complex history which probably involved the following stages: 1) original syngenetic deposition of uranium in black shales by reduction of uranium from solutions which leached the underlying Archean granites; 2) remobilization of uranium during metamorphism about 1700-1900 m.y. ago; 3) introduction of supergene uranium from volcanics, sandstones, and arkoses of the Carpentarian System; and 4) renewed mobilization of uranium about 850 and 500 m.y. ago which are the dates obtained by Hills and Richards (1972) and Cooper (1973) for the uranium mineralization.

These deposits are clearly unrelated to the fossil-placer model. Deposition of sediments was in shallow marine instead of fluvial environments and uranium was transported in solution instead of as detrital particles. The concentration of uranium was evidently mainly an epigenetic, hydrothermal process (Eupene and others, 1976) and not related to sedimentary processes. Nevertheless, one pertinent aspect of the northern Australian deposits is that the most important deposits such as Jabiluka, Ranger, and Nabalek are adjacent to areas of intense metamorphism (Plumb and Derrick, 1975; Warren 1972a). This suggests that high-grade metamorphism literally expels uranium and concentrates it in near-by lower grade rocks. This concept could be very useful in exploration for uranium in metamorphosed quartz-pebble conglomerates.

Other Areas

Early Proterozoic rocks also crop out in the Nicholson and Tennant Creek Blocks shown in Figure 2.17. The Nicholson Block includes pelites, graywackes, and volcanics which are metamorphosed to greenschist facies (Roberts and others, 1963). The volcanics contain uranium and copper deposits which are related to the overlying unconformity with Carpentarian sediments (Warren, 1972b). The Tennant Creek Block contains marine graywackes and siltstones which apparently overlie Archean basement (Warren, 1972a).

MINERAL DEPOSITS

There are no known major Early Proterozoic fossil-placer conglomerates in Australia (1979), a rather surprising fact considering that the extensive Archean cratons of western Australia provide a good source for metals and the Early Proterozoic rocks on the margins of the cratons provide a suitable host. However, there are uneconomic uraniferous conglomerates and sub-economic auriferous conglomerates reported from the Hamersley Basin, where Early Proterozoic metasediments onlap the southern margin of the Pilbara craton (Robertson, 1974; Richards, 1972; Ryan, 1977). Auriferous conglomerates have been mined since 1888 from the base of the Fortesque Group of the Mount Bruce Supergroup (Robertson, 1974; Blockley, 1975) and uraniferous conglomerates occur in the same group some 2000 m higher in the section--in the Green Hole Formation. These conglomerates are 1-5 m thick and laterally continuous, and they contain values up to 425 ppm uranium. Either of these fossil-placer bearing conglomerates could prove to be important sources of uranium or gold if the right sedimentary facies could be found. The length of outcrop of the Fortesque Group is more than 600 km along the south margin of the Pilbara craton and it seems likely that somewhere within this distance, there were one or more deltas or alluvial fan-deltas in which heavy minerals and coarse-clastics could be continually reworked and concentrated by braided rivers. Paleocurrent and depositional environment studies of the lower Fortesque Group would be very helpful in locating such deltas if they exist.

Other Early Proterozoic metasediments in Australia appear less likely to contain fossil-placer deposits. The Halls Creek Group is mainly a eugeosynclinal sequence so depositional environments of the sediments were not favorable for fossil-placer accumulations. However, uranium is reported from a lower clastic unit of the group, and it remains possible that uranium could occur in miogeosynclinal or platform-type equivalents of the Halls Creek Group beneath the Carpentarian cover rocks on the Kimberley Block.

Rocks of the Pine Creek Geosyncline are mainly marine, and fluvial clastics are not abundant so the possibility for fossil-placer deposits in these rocks is also limited. There are radioactive conglomerates in the Crater Formation of the Batchelor Group, but the radioactivity is weak and is due to detrital thorite and monazite (Walpole and others, 1965; Ryan, 1977). This unit, and others like it, may have contributed uranium to the unconformity deposits but are probably not of economic significance themselves.

Rocks of the Tennant Creek and Nicholson Blocks are similar to the Pine Creek Syncline and are unlikely to contain fossil-placer ore-deposits.

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EAST EUROPEAN PLATFORM

BALTIC SHIELD

Geologic Setting

The Baltic Shield is located in Scandinavia and northwestern Russia and contains an Archean granite-greenstone nucleus, Early Proterozoic metasedimentary rocks, and a large area of Proterozoic metamorphic and igneous rocks. The Archean craton of the Baltic Shield contains: granulite facies supracrustal rocks (Kola Series) which are older than 3300 m.y.; granodioritic and tonalitic intrusives about 3000 m.y. old; greenstone belts which were deposited 3000-2800 m.y. ago and are preserved in a few north- to northwest-trending, isolated inliers surrounded by granitic rocks; intrusive trondhjemites and granodiorites 2800-2700 m.y. old; and potassium-rich (microcline) granites about 2700-2500 m.y. old (Gaál and others, 1978; Bowes, 1976; Salop, 1977). As shown in Figure 2.18, the Archean craton is bounded on the northeast by the Arctic Ocean, on the northwest by deformed Paleozoic sediments of the Scandinavian Calidonides, and on the southeast by Paleozoic sedimentary rocks of northwestern Russia which overlie and conceal large areas of Precambrian rocks of the East European Platform. On the southwest, the Archean craton is bounded by Proterozoic metasedimentary and plutonic rocks, including a sequence of Early Proterozoic quartz-rich clastics called the Karelian Supergroup. The Karelian Supergroup also crops out in isolated intracratonic basins within the Archean craton (Figure 2.18).

The Proterozoic rocks in the western and southwestern parts of the Baltic Shield are mainly 2100-1750 m.y. old (Lobach and others, 1972, *in* Bowes, 1976) and include schist belts which are referred to as the Svecofennian schist belts in the older literature. These schists were metamorphosed, deformed, and intruded by numerous granitic plutons during the Svecokarelian

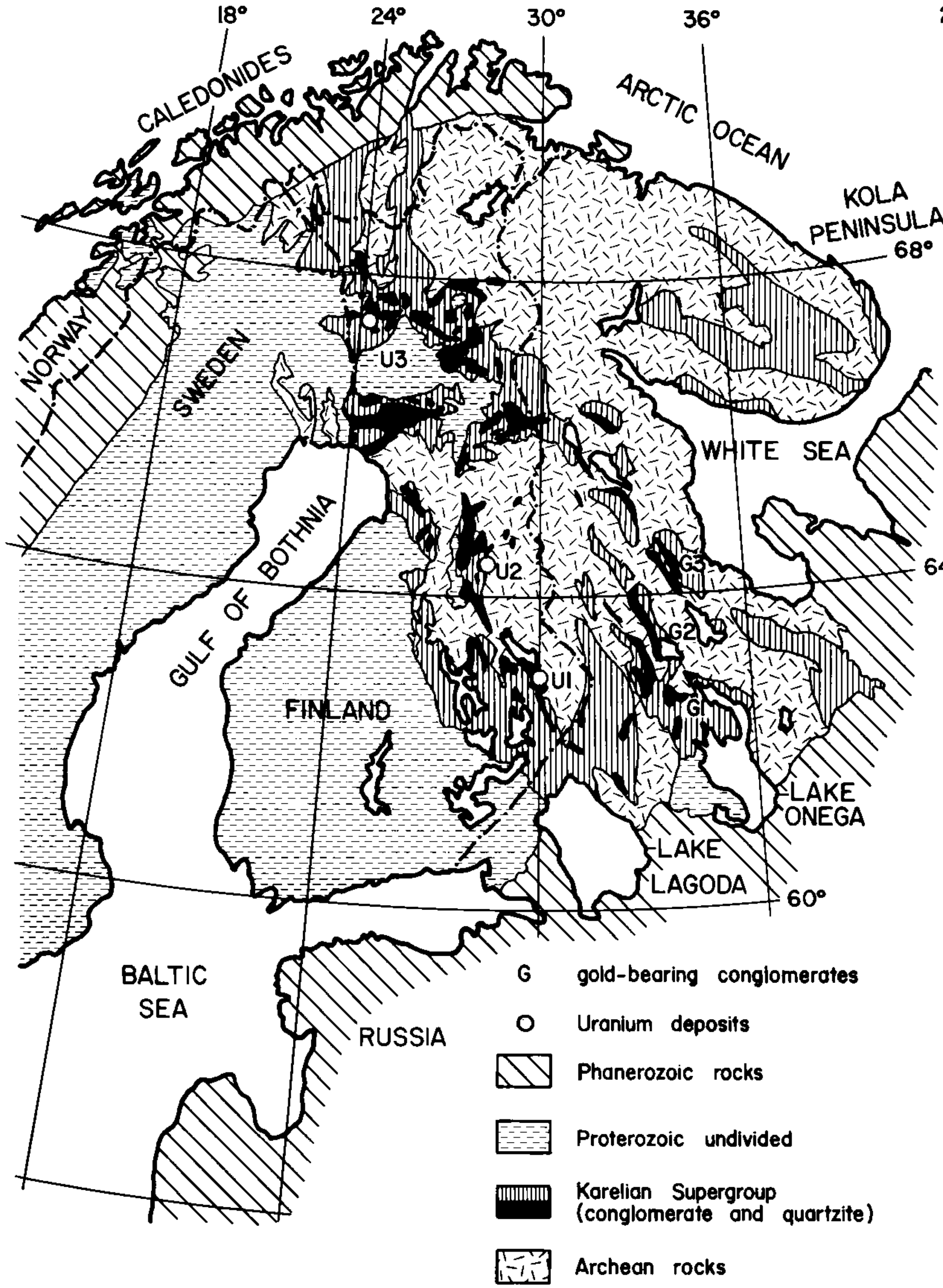

Figure 2.18 Generalized geology of the Baltic Shield. Adapted from Nalivkin, 1960; Simonen, 1971; Bowes, 1976; Gaál and others, 1978; and Kahma, 1973. Numbered mineral occurrences refer to localities discussed in text.

orogeny, about 1800-2000 m.y. ago (Eskola, 1963; Stromberg, 1976), and include eugeosynclinal counterparts of the miogeosynclinal Karelian quartzites (Kahma, 1973; Eskola, 1963) plus younger metasediments (Bowes, 1976). The Svecokarelian orogeny is similar to the Hudsonian orogeny of the Canadian Shield in that it remobilized large areas of older basement in Sweden and Norway about 1800 m.y. ago (Stromberg, 1976).

The tectonic fabric of the Baltic Shield is dominated by a series of northwest-trending wrench faults and shear zones which form the boundaries of the major rock provinces (Touminen and others, 1973). According to Stromberg (1976, 1978) and Kahma (1973) these fault zones are the locus of important sulphide mineralization in Scandinavia and may be related to Archean structural patterns.

Post-Svecokarelian events include anorogenic plutonism and volcanism 1700-1500 m.y. ago, including emplacement of anorthosites and rapikivi granites, arenaceous sedimentation 1300-1200 m.y. ago, metamorphism about 1000 m.y. ago, and the Caledonian orogeny about 400 m.y. ago.

Karelian Supergroup

Distribution and Age

Figure 2.18 shows that most of the outcrops of the Karelian Supergroup occur in a northwest-trending zone which extends from Lake Onega, in the southeast, to the Caledonides in the northwest. Other outcrops occur in the Kola Peninsula of Russia. Rocks of the Karelian Supergroup unconformably overlie the Archean granite-greenstone terrain and are preserved both on the southwest margin of the Archean craton, where they dip away from basement, and as isolated synclinal inliers within the Archean craton (Ojakangas, 1965). This distribution suggests that the Karelian Supergroup originally covered most of the Archean craton (at least 400,000 sq. km) and that the preserved inliers represent the deepest parts of, once larger, depositional basins on the craton (Ojakangas, 1965). This idea is compatible with Salop's (1977) observations that the Karelian rocks within the craton tend to be thin and only slightly metamorphosed platform sediments whereas the Karelian rocks which border the Archean craton (e.g. near Lake Ladoga, in northern Finland, and near the White Sea) are thicker, more highly metamorphosed miogeosynclinal strata. In both platform and miogeosynclinal sequences, the general stratigraphic sequences are similar but thickness and names differ (Salop, 1977).

The age of the Karelian Supergroup is bracketed between 2600 and 2000 m.y. (Table 2.10). A maximum age is given by 2600 m.y. old granites from Russia (Salop, 1977) and 2500 m.y. old granites from Finland (Gaál and others, 1978), both of which unconformably underlie the Karelian Supergroup. A minimum age comes from 2000-2150 m.y. old diabases which intrude the Karelian Supergroup in Russia and Finland (Table 2.10), from 1900 m.y. dates from metamorphic minerals in the Karelian Supergroup, and from 1900-2000 m.y. old intrusive syn-orogenic granites related to the Karelian (Svecokarelian) orogeny (Salop, 1977; Wetherill and others, 1962). Direct dates on the Karelian rocks include a 2500 m.y. K-Ar date from basic lavas of the Segozero Group in Russia (Svetov, 1972, <u>in</u> Salop, 1977) and a 2300 $\pm$ 120 m.y. Pb isotope date on carbonate rocks in the overlying Onega Group (Salop, 1977).

Stratigraphy and Lithology

Table 2.10 summarizes the stratigraphy of the Karelian Supergroup in Russia and Finland. In both areas, the Karelian Supergroup is divided into groups which are generally composed of a lower sedimentary sequence, dominated by clastics, and an upper volcanic sequence. As in other Early Proterozoic sequences such as the Witwatersrand Sequence, the ratio of clastic

KARELIAN SUPERGROUP OF RUSSIA				KARELIAN SUPERGROUP OF FINLAND			
Group (Mineralization)	Lithology	Thickness (m)	Age (method)	Group (Mineralization)	Lithology	Thickness (m)	Age (method)
Vespian	red, gray sandstone, slate dark quartzite, slate	1650 600	intruded by 2000-2150 m.y. diabase (K-Ar)	Koli	quartzite conglomerate		
Bessovets (Ladoga)	fm.-gr. sandstone, siltstone	1000		Kalevian	phyllite, mica schist quartzite, conglomerate		
Onega (Fe)	basic volcanics dolomite, siltstone, iron fm., quartzite, quartz-pebble congl.	300-500 800-1500	2300±120 (Pb-Pb)	JATULIAN: Marine Jatulian	amphibole schist dolomite, phyllite	>300 50-100	2050±50 (U-Pb zircon
Segozero (Au, Fe)	tuff-sandstone, schist metadiabase dolomitic sandstone, siltstone, quartz-pebble congl., arkose	200-400 700-1000	2500(K-Ar)	JATULIAN: Kainuu (Au, U)	quartzite basic volcanics schist, siltstone sericitic quartzite, quartz-pebble congl.	600-800 200 350-450 100-200	cross-cut by 2160 m.y. diabases
Sariolian	basic volcanics, tuff polymictite congl., arkose, grit, quartz-pebble congl.	0-400 0-800		JATULIAN: Sariolian	basic volcanics conglomerate, arkose	0-500 0-50?	
Tunguda-Nadvoitsa (U, pyrite)	acid volcanics, tuff, basic volcanics, tuff quartzite, pyritic quartz-pebble congl., arkose, polymictic congl.	0-1800 0-120		Belomoride and Tuntsa-Savukoski	mica schist, quartzite paragneiss, volcanic agglomerate	0-?	
Pebozero	amphibolite, tuff, diabase gneiss, schist, quartzite	0-3000					
Tiksheozero Keyvy, Chartvert	conglomerate, quartzite, schist	0-1000					
Archean granite - greenstone rocks			2600-2800	Archean granite - greenstone rocks			>2500

Table 2.10. Comparative stratigraphies of the Karelian Supergroup in Russia and Finland. Adapted from Salop (1977); Simonen (1960); Silvennoinen (1972).

sediments to volcanics in the various groups increases up-section.

The lowest groups in both columns of Table 2.10 crop out mainly in the northern part of the Baltic Shield and are only locally developed (Salop, 1977; Simonen, 1960). These sequences are poorly known and may be Archean in age (Bogdanov and others, 1968; 1971, in Salop, 1977). However, Salop (1977) stated that they appear to underlie units of the Karelian Supergroup without marked angular unconformity and should, therefore, be considered part of the Karelian Supergroup.

The Sarolian Group is more widespread but it also is only locally developed. It occurs in isolated outcrops which appear to have filled valleys and depressions in the underlying Archean granitic basement (Simonen, 1960; Eskola, 1963). This group consists of basal conglomerates and arkosic quartzites which were derived from weathering products of the underlying basement (Eskola, 1963). The basal conglomerates are polymictic and contain subangular pebbles of gneiss, schist, and amphibolite up to 50 cm in diameter in a matrix of quartz, albite, mica, and carbonate (Silvennoinen, 1972; Eskola, 1963). The conglomerates get finer-grained and pebbles get more rounded up-section, eventually grading into coarse-grained arkose. This unit is, in part, a regolith developed on the Archean terrain and, in part, immature fluvial deposits which filled local depressions (Gaál and others, 1975; Ojakangas, 1965; Eskola, 1963; Simonen, 1960).

The Segozero Group of Russia and the Kainuu Group (Jatulian quartzites) of Finland are thick (1000-1500 m) and laterally persistent sequences of conglomerates, quartzites, and metavolcanics which are readily identified throughout the Baltic Shield. The distribution of the quartzites and conglomerates of this sequence is shown in Figure 2.18. The sequence is interpreted by Simonen (1960) and Eskola (1963) to be a transgressive epicontintal assemblage with a basal sericitic oligomictic quartz-pebble conglomerate, sericitic quartzite, sericite schist (and kaolinitic quartzite), and finally a pure quartz-arenite (Simonen, 1960). The lower units of the sequence contain gold- and uranium-bearing conglomerates in Finland and Soviet Karelia (Salop, 1977).

The Onega and Marine Jatulian Groups consist of marine sediments up to 2000 m thick such as stromatolitic dolomite, phyllite, iron formation, and pillow basalt. These sediments represent continued marine transgression following deposition of the Segozero/Kainuu sequence.

The Kalevian and Bassovets Groups unconformably overlie the Jatulian/Onega Groups in most areas (Simonen, 1960; Salop, 1977). The base of the unit is marked by a conglomerate containing pebbles of the underlying quartzites. The majority of the rocks in this sequence are phyllites and mica schists, including graphitic schists and rhythmically bedded graywacke of probable turbidite origin.

The highest units of the Karelian Supergroup are the Vepsian and Koli Groups. These units consist of terrigenous clastics which were deposited unconformably on the underlying marine rocks. Some of these clastics are redbeds which contain microphytolites (Salop, 1977).

Paleogeography

The paleogeographic setting prior to deposition of the lowest sediments of the Karelian Supergroup was characterized by an uplifted and deeply eroded block of Archean granitic gneisses and greenstones which was experiencing intermittent subaerial volcanism. The erosion surface on top of the block was irregular and probably contained several small intracratonic basins in which isolated sedimentary successions like the Tiksheozero, Pebozero, Tunguda-Nadvoitsa, and Sarolian Groups were deposited. These sequences contain epicontinental to shallow-water detrital deposits and subaerial volcanics. For example, the Sarolian Group is probably a fluvial deposit and consists of poorly-sorted, oligomictic conglomerates which were derived from mechanical weathering of nearby topographic highs in the gneissic basement, transported a short distance, and deposited in local depressions (Simonen, 1960). These conglomerates contain heavy mineral assemblages but the distance of fluvial transport and amount of fluvial reworking were probably not adequate to form significant concentrations of heavy minerals.

The Kainuu/Segozero Groups represent the first widespread deposition of metasediments on the Archean craton. These groups are interpreted by Simonen (1960) and Ojakangas (1965) to be transgressive sequences, starting with fluvial deposition of conglomerate and aluminous quartzite, both of which were derived from deep mechanical and chemical weathering of the granitic source terrain. The aluminous quartzites grade up-section into more mature, marine quartzites. According to Ojakangas (1965), the Jatulian Sea transgressed eastward over the Archean craton and the quartzites thin to the east. The fluvial deposits of the lower part of this sequence contain uranium- and gold-bearing conglomerates but these disappear up-section, as marine deposition predominated.

Crossbedding in the quartzites of the Kainuu Group is common and includes both trough and planar types. Ripple marks are also present and some argillaceous units contain mudcracks (Ojakangas, 1965; Eskola, 1963). Paleocurrent data from crossbedding in the Finnish quartzites indicate that the dominant paleocurrent directions were to the west and south and that the source of the Jatulian quartzites of Finland was the Archean craton to the northeast (Ojakangas, 1965).

Mineral Deposits

There are several reports of gold- and uranium-bearing conglomerates in the Karelian Supergroup but unfortunately very few detailed descriptions are available. Salop (1977) reported auriferous and uraniferous conglomerates in the Kainuu Group of Finland and in the corresponding Segozero Group of Russia and he reported pyritic quartz-pebble conglomerates in the underlying Tunguda-Nadvoitsa Group. According to our model, these units are the right age and lithology to contain fossil-placer uranium and pyrite. Also, the paleogeographic setting is favorable. The conglomerates were deposited on a deeply eroded Archean granitic basement by fluvial processes.

Kahma (1973) showed only three small uranium deposits in Finland. He interpreted them all to be stratiform deposits and they are all located within rocks of the Karelian Supergroup (Figure 2.18). This distribution of uranium within rocks of the Karelian Supergroup suggests that these rocks contain significant, if low-grade, concentrations of uranium.

The best known of the Finnish uranium deposits is the Paukkajanvarra mine in the Kola area of southeastern Finland (U1, Figure 2.18). This deposit is described by Tyni (1962; in Kaliokoski and others, 1978) and Wennervirta (1960). Uranium mineralization occurs near the contact between Archean gneisses and the overlying Kainuu Group metasediments. In ascending order, the Kainuu Group consists of sericite schist, sericitic quartzite, conglomerate, and fine-grained quartzite. The lower sericite schist is gradational with granitic basement and may be, in part, a regolith. The conglomerate contains mainly quartz pebbles with minor quartzite and sericite schist pebbles. The pebbles get finer grained up-section and the conglomerate grades into a light-gray, strongly fractured quartzite. The matrix of the conglomerate is sericitic at the base, but higher up it consists of biotite, chlorite, and magnetite. According to Ojakangas (1965) the most abundant heavy minerals in the Jatulian quartzites are apatite, tourmaline, and zircon with minor magnetite, ilmenite, hematite, pyrite, chalcopyrite, and other silicates.

Uranium ore is associated with diabase dikes which intrude both basement and metasediments. In most cases, the ore is within quartzites and conglomerates of the Karelian Supergroup which are in contact with intrusive diabase. The highest grades of ore are in conglomerate but quartzite and diabase also contain ore-grade mineralization and sub ore-grade uranium also extends into the underlying basement (Wennervirta, 1960).

According to Tyni (1962), the uranium is epigenetic in origin. The most common mineral is uranophane which occurs as soft masses which fill fractures. In the larger fractures, pitchblende is also observed. The best ore occurs in the highly-fractured areas of the metasediments adjacent to certain diabase dikes. Tyni suggested that the origin of the ore is related to the mineralogy of the diabase. However, it seems more likely to us that the uranium originated from the conglomerates and quartzites of the Kainuu Group and was reconcentrated by diabase intrusions which intersected paleochannels in the fluvial sediments. This would explain the association of uranium with only a few of diabase dikes and the preferred distribution of uranium ore with the basal conglomerate of the Karelian Supergroup.

The other uranium deposits in Finland are the Nuottijärva deposit in central Finland (U2 of Figure 2.18) and the Kesänkitunturi deposit (U3 of Figure 2.18) in northwestern Finland (Kahma, 1973). The Nuottijärva deposit contains uranium and thorium and is located near the basal contact of the Karelian Supergroup. The Kesänkitunturi deposit contains mainly uranium and is located entirely within metasediments (Figure 2.18). The nature and origin of these deposits is not known to us.

Radioactive conglomerates have also been reported from Soviet Karelia (Lobanov, 1961; Shkvorov and Kovalev, 1961; Davidson, 1964; Salop, 1977) but locations and detailed geologic information are not available. Shkvorov and Kovalev (1961, in Davidson, 1964, p. 172) reported uraninite-, brannerite-, and gold-bearing conglomerates in an unspecified locality which show a "zoning" between a uranium-pyrite-rutile assemblage and a brannerite-magnetite-davidite assemblage. Details on these rocks would be exceedingly interesting because the gradation between assemblages would suggest that the conglomerates were deposited during a period of transition between oxygenic and anoxygenic conditions.

One clue to the location of these conglomerates is the distribution of gold-bearing Proterozoic conglomerates in Soviet Karelia which, it appears, Russian geologists have discussed more freely. Yakoleva and others (1969) reported gold-bearing conglomerates from basal layers of clastic successions in three areas, shown in Figure 2.18. In the Koikar area (G1 of Figure 2.18), gold occurs in quartz-pebble conglomerates interpreted to be epicontinental sediments deposited in intermontane basins; in the Lekhtin area (G2 of Figure 2.18) gold, pyrite, and carbonaceous material are reported from presumed coastal-marine conglomerates; and the Yangozero area (G3 of Figure 2.18) contains a 300-350 m thick gold-bearing conglomerate horizon (Yakoleva and others, 1969). These rocks contain generally less than 2 ppm gold. However, they are reported to be similar in lithology to the Witwatersrand conglomerates and may well contain appreciable but unreported uranium as well as gold.

UKRANIAN SHIELD

Geologic Setting

The Ukranian Shield is located in southwestern Russia, north of the Black Sea, and includes Archean and Proterozoic rocks which are surrounded and partially concealed by Tertiary deposits. Drill data indicate that the Ukranian Shield and Baltic Shields are part of a continuous Precambrian block, the East European Platform, which is concealed by Phanerozoic rocks in western Russia. The oldest rocks in the Ukranian Shield are granulite facies supracrustal rocks of the Bug Group and equivalents which yield a Pb-model age of 3600 m.y. (Salop, 1977). This group includes gneisses, migmatites, and amphibolites with layers of calc-silicate, marble, garnet schist, sillimanite gneiss, quartzite, and iron formation. These gneisses were intruded by granitic bodies about 3000 m.y. ago (Vinogradov and others, 1960). Greenstone belt assemblages were deposited unconformably on the gneisses between 2800 and 2600 m.y. ago (Ladieva, 1965, in Salop, 1977). These belts are complex synclines nestled between granite domes which intruded and metamorphosed the greenstone belts about 2700-2500 m.y. ago (Salop, 1977).

Stratigraphy

Early Proterozoic rocks were deposited in the Ukranian Shield between 2700 and 2000 m.y. ago (Salop, 1977). These rocks are called the Krivoy Rog and Frunze Mine Groups in the central Ukranian Shield and the Belozerka and Pereversevsk Groups farther east. The Krivoy Rog Group (and equivalents) consist of two formations: a lower, 500 m thick clastic sequence containing conglomerate, arkose, quartz-arenite, and phyllite; and an upper, 1400 m thick, sequence of schists, phyllites, and economically important iron formation (Salop, 1977). Volcanic rocks are generally lacking. The basal conglomerate of this group is a paraconglomerate containing dropstones of possible glacial origin. The matrix of the conglomerate contains uranium-bearing pyrite grains which give a Pb-isotope age of 2600-2700 m.y. The Krivoy Rog Group is considered by Salop (1977) to be correlative with the Tunguda-Nadvoitsa and Sarolian Groups of the Karelian Supergroup.

The Frunze Mine Group (and equivalents) unconformably overlie the Krivoy Rog Group and consist of quartzite (40-120 m), slaty carbonate (300-900 m), a sandstone-conglomerate unit (1000-1500 m), and a slate unit (300-1000 m). These rocks are considered by Salop to be equivalent to the Segozero and Onega Group of the Karelian Supergroup on the basis of the similarities in stratigraphy and lithologies and the presence of organic remains in the carbonaceous rocks. A minimum age of 2000 m.y. for this group comes from 1700-2000 m.y. dates on metamorphic minerals in the sequence.

Mineral Deposits

We know of no reports of economic concentrations of uranium or gold in the Ukranian Shield. However, Salop (1977, p. 135) reported the presence of syngenetic uranium-bearing pyrite grains in the matrix of the basal conglomerates of the Krivoy Rog Group and in correlative rocks of the Voronezh Anticline, to the northeast, and conglomerates in both of these areas also contain gold. Rozhov (1969) and Pisemskij and others (1969) reported gold, pyrite, and "other heavy minerals of clastic origin" in the basal parts of sericitic, quartz-pebble conglomerate layers. These layers are 1 cm to 2 m thick in the southern part of the Krivoy Rog Syncline and occur within a 5-50 m thick quartz sandstone unit. Farther north, the basal 60-100 m of the Krivoy Rog Group is made up of interlayered polymictic boulder conglomerates and oligomictic quartz-pebble conglomerates. Fire assays of over 300 samples of conglomerates were reported by Rozhov (1969). Of the samples, 32 percent were barren, 46 percent contained trace amounts of gold, 17 percent contained .1 ppm, 4 percent contained .2 ppm, and one percent contained more than .3 ppm. The maximum reported gold content was 1 ppm. The gold occurs as fine-grained angular particles in the matrix or associated with pyrite and pyrrhotite grains. According to Rozhov (1969), the highest gold assays came from thin, uniform layers of quartz-pebble conglomerate which directly overlie coarser, polymictic boulder conglomerates. Rozhov also reported that the layers richest in gold show the highest gamma-ray readings of any rocks in the section, which suggests they may also contain some uranium-bearing detrital minerals.

The conglomerates in the Krivoy Rog Group are interpreted to be alluvial fan or intermontane fan-delta deposits (Rozhov, 1969), they were deposited unconformably on an eroded amphibolite terrain prior to 2000 m.y. ago, and they extend for about 80 km in a north-south direction. Thus, they appear to have been deposited in favorable depositional environment, they are the right age, and they are extensive enough to contain appreciable fossil-placer uranium and gold mineralization if the source area of the sediments contained uranium minerals and gold.

One interesting feature of the Krivoy Rog Group is the presence of economically important iron formation directly overlying the basal Proterozoic clastic sequence. This situation is analogous to the stratigraphic sequence seen in the Minas Series of Brazil. In both cases, the transgressive, and dominantly marine nature of the bulk of the sedimentary sequence suggests that fossil-placer deposits will be restricted to the basal clastic units and that mineral concentrations are likely to be of lower tenor than in Early Proterozoic cyclic fluvial sequences such as the Witwatersrand, where constant reworking of sediments during regressive cycles was responsible for concentrating heavy minerals.

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SIBERIAN PLATFORM

GEOLOGIC SETTING

Archean rocks of the Siberian Platform underlie most of Asia and constitute one of the largest Archean platforms in the world. As shown in Figure 2.19, Archean rocks crop out in three parts of the Siberian Platform: the Anabar Craton in the north, Aldan Craton in the southeast, and the Yenisei Anticline in the southwest. Drill data and geophysical data indicate that the areas in between contain Archean rocks in sub-outcrop, beneath late Proterozoic and Phanerozoic platform sediments (Salop, 1977; Kosygin and Parfenov, 1975). The Anabar Craton and Yenisei Anticline are continuous in sub-outcrop and form part of the Angara Shield, which comprises the entire western part of the Siberian Platform (Figure 2.19). The Angara Shield is separated from the Aldan Craton to the east by a north-trending trough which contains Archean rocks and Early Proterozoic metasediments and is considered by Salop (1977) to be two aulocogens separated by a small Archean block (Figure 2.19). The Siberian Platform is bounded on the west by Quaternary cover, on the northeast by folded Mesozoic sedimentary rocks and Mesozoic granites, and on the south by folded Proterozoic and Paleozoic sediments and Paleozoic (Caledonian) granitic intrusives (Figure 2.19).

The oldest rocks of the Siberian Platform are granulite-facies paragneisses and migmatites of the Aldan, Anabar, and Kansk Groups and equivalents. The Aldan Group, as an example, is about 16 km thick and is subdivided into three subgroups: a lower sequence containing metavolcanic gneisses and quartzites, a middle sequence containing amphibolite, mafic schist, and carbonate, and an upper sequence containing garnet paragneiss and calc-silicates. These rocks yield a Pb-isochron date and K-Ar mineral ages of about 3500 m.y. which are interpreted by Salop (1977) to represent the time of

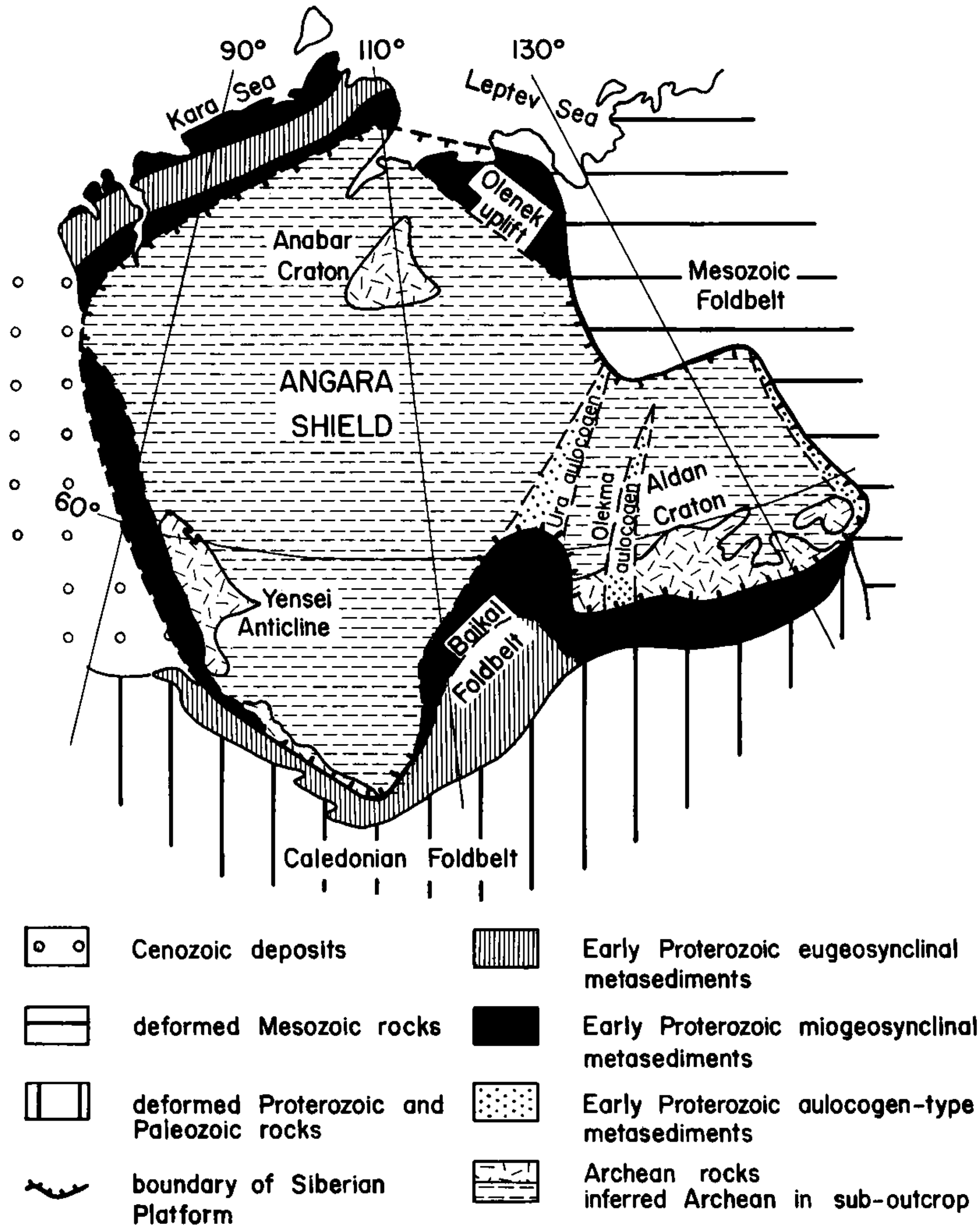

Figure 2.19. Generalized geology and Early Proterozoic structural elements of the Siberian Platform. Adapted from Salop (1977).

granulite grade metamorphism. A minimum age for the Aldan Group is 2600-2800 m.y., which is the date on granites that cross-cut the greenstone belts which overlie the Aldan Group. Other Early and Middle Archean rocks crop out in the Anabar Craton, Yenisei Anticline, and in the fold belts which surround the Siberian Platform. These fold belts contain granulite and amphibolite facies rocks which were remobilized and retrograded during Proterozoic orogenies.

Archean greenstone belts overlie the granulite facies gneisses. These sequences consist of volcanic and sedimentary rocks, including iron-formation, which are similar to greenstone belts on other cratons except they may contain a higher percentage of volcaniclastic and sedimentary rocks relative to volcanics. Also, in several areas they are reported (Salop, 1977) to be continuous with Early Proterozoic metasedimentary successions, suggesting that there was no abrupt change in style of sedimentation at the end of the Archean in the Siberian Platform. Another interesting feature of the Archean gneissic sequences is that they sometimes occur in narrow fault-bounded troughs which contain strongly folded and metamorphosed metasediments and intrusive gabbros and granites. According to Salop (1977), these troughs bounded discrete Archean blocks and may have been the precursors of mobile belts and Proterozoic aulocogens.

EARLY PROTEROZOIC METASEDIMENTARY ROCKS

There are four types of Early Proterozoic metasedimentary sequences in the Siberian Platform and in surrounding fold belts: 1) platform-type sequences, 2) miogeosynclinal sequences, 3) aulocogen-type sequences, and 4) eugeosynclinal sequences.

Early Proterozoic platform-type sequences of epicontinental clastic rocks are not common in the Siberian Platform. The only possible sequence of this type is the Eyekit Group of the Ohenek uplift in the northeastern corner of the Siberian Platform. These rocks are terrigeneous clastics and are probably Early Proterozoic (Kosygin and Parfenov, 1975), although Salop considered them to be late Archean. Salop (1977) suggested that other Early Proterozoic epicontinental sequences may be concealed by the thick cover of younger sediments on the Siberian Platform.

Miogeosynclinal sequences occur on the south, west, and north margins of the Siberian Platform (Figure 2.19). The best example is the Udokan Group of the Baikal fold belt, on the southern margin of the platform. This group is about 13 km thick and contains three subgroups: a lower sequence of fine-grained marine clastics, a middle sequence of presumed shallow marine quartzites and stromatolitic carbonates, and an upper sequence of coarse-grained arkoses and quartzites containing red-beds and detrital magnetite. An Early Proterozoic age for the Udokan Group is verified by the fact that it lies on 2600-2800 m.y. old greenstone sequences and is intruded by 1900 m.y. old granites (Salop, 1977).

Aulocogen-type sequences occur in the central Siberian Platform, between the Angara and Aldan Cratons, and on the eastern edge of the Aldan Craton (Figure 2.19). The best example is the Olekma sequence. This sequence is about 2.5 km thick and is divided into four formations: a lower unit consisting of basal conglomerate, arkose, and quartzite, a dolomite unit, a dark-gray siltstone unit, and an upper unit containing thick cross-bedded quartzite. This sequence is a transgressive sequence which unconformably overlies Archean granitic rocks. It is comparable to the miogeosynclinal sequence except units are thinner.

Eugeosynclinal sequences occur on the north, west, and south margins of the platform. One example is the Zama Subgroup of the Baikal fold belt. This subgroup contains graywacke, tuffaceous sandstone, dolomite, and volcanic rocks and is considered by Salop (1977) to be a continuation of Archean sedimentation in the Kilvana Subgroup.

MINERAL DEPOSITS

There are no reported occurrences of uraniferous conglomerates in Early Proterozoic rocks of the Siberian Platform that we are aware of. However, Levin and others (1970) do report gold-bearing Proterozoic conglomerates in Aldan craton. It seems likely that uranium may be associated with this and other similar conglomerates, particularly the basal conglomerates of aulocogen-type sequences, and possibly in basal units of the miogeosynclinal successions. Also, if other platform sequences are found beneath the younger cover rocks, these would be good targets for fossil placer deposits.

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PART III. UNITED STATES RESOURCE POTENTIAL

INTRODUCTION

The preferred exploration model discussed in Part I restricts the occurrence of uranium-bearing, quartz-pebble conglomerates to Precambrian clastic rocks older than about 2000 m.y. This limits prospecting to two major regions of the United States--the Lake Superior region and the region surrounding the Wyoming Province. These areas are shown in Figure 3.1.

We will discuss the Lake Superior region first because this region borders the Canadian Shield and is a logical area to look for possible extensions of the uranium-bearing units of the Huronian Supergroup, which crop out only 200 km east of the Lake Superior region of the United States (Figure 3.1).

LAKE SUPERIOR REGION

The Lake Superior region includes rocks of Archean age which form the southern extension, into the United States, of the Superior Province of the Canadian Shield. Archean rocks crop out in Minnesota, Michigan, and Wisconsin, as far west as the Minnesota-North Dakota border, as far south as about latitude 44° N in Wisconsin, and as far east as the vicinity of the city of Marquette, Michigan (Figure 3.2). Sims (1976) has subdivided the Archean into two terrains: a granite-greenstone terrain located primarily in northern Minnesota and northwestern Michigan and a high-grade gneiss terrain located in central Minnesota and Wisconsin (Figure 3.2). Two subparallel troughs of Proterozoic age (younger than 2500 m.y.) developed on this Archean basement. The older trough, shown in Figure 3.2 as the Marquette Range Supergroup and equivalents, is a geosyncline that extends roughly around the margin of Lake Superior and appears to be warped so that the axis strikes northeast in central Minnesota and western Wisconsin and then bends around

Figure 3.1. Geology of the Archean nucleus of North America showing Archean cratons (> 2.5 b.y.), rejuvenated Archean terrains (1.7 - 2.5 b.y.), and Proterozoic foldbelts (< 1.8 b.y.).

EXPLANATION

Grenville and Keweenawan rocks (~1.0 b.y.).

Early and Middle Proterozoic eugeosynclinal rocks

Early Proterozoic miogeosynclinal rocks

Archean terrains rejuvenated during Hudsonian orogeny

Archean rocks

Positive gravity anomaly

Known or inferred boundaries of geologic provinces

Electrical conductivity anomaly

2.5-3.2 Range of radiometric ages in billions of years

LABELED LOCALITIES

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

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Geophysics - Alabi and others (1975); Gibb and Thomas (1977); Horner and Hasegawa (1978); Kent and Simpson (1973); Kreary (1976); Lidiak (1971); Thomas and Gibb (1977).

Geochronology - Duncan (1978); Goldich and Hedge (1974); Goldich and others (1966); Hurst and others (1975); Hills and others (1968); Hills and Armstrong (1974); Hills and Houston (1979); King (1976); Moorbath and others (1972); Peterman and Hedge (1968); Price and Douglas (1972); Van Schmus (1976).

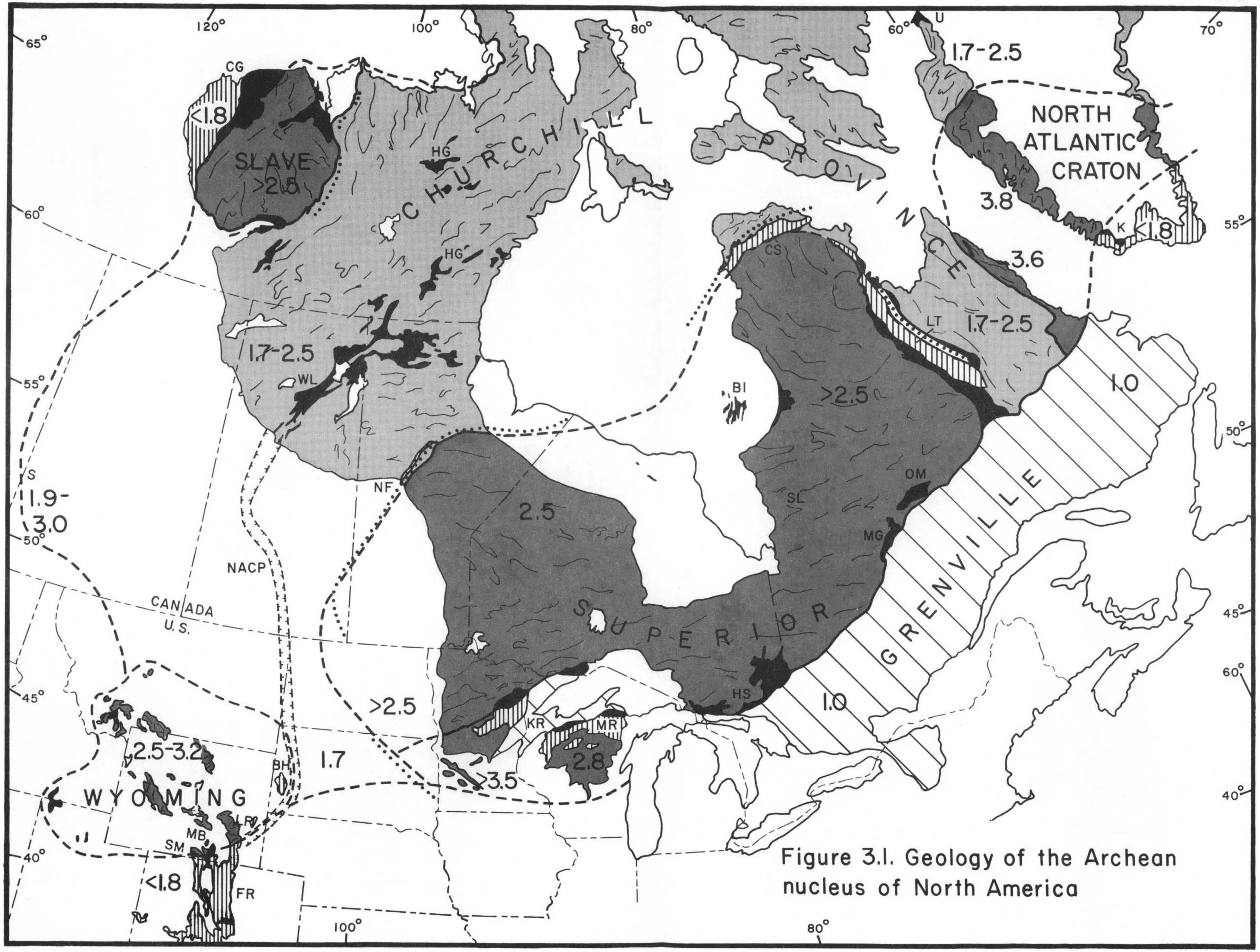


Figure 3.1. Geology of the Archean nucleus of North America

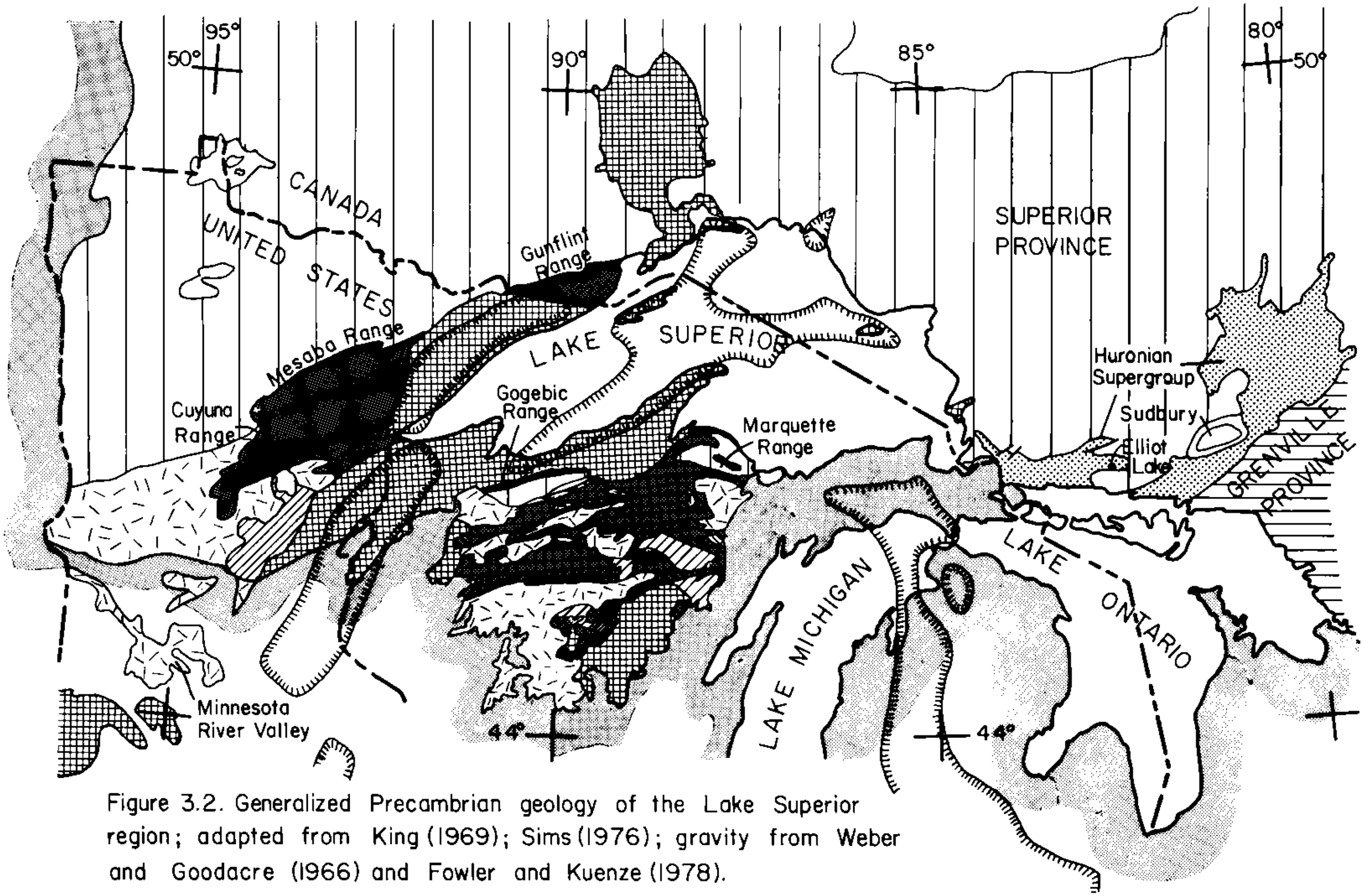
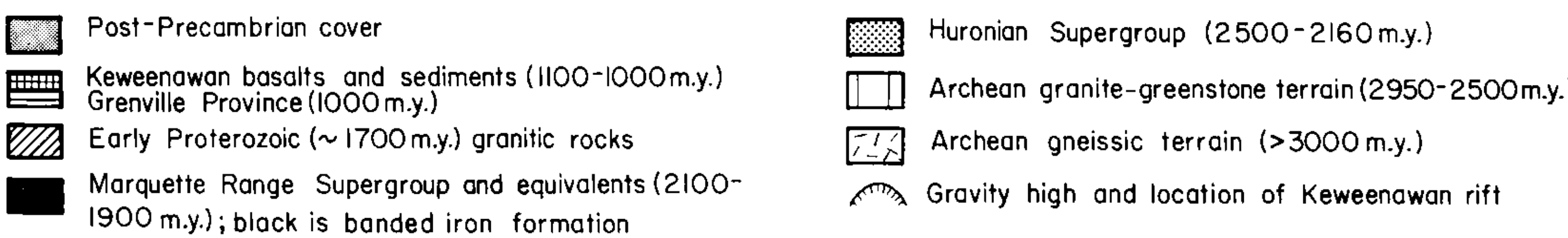


Figure 3.2. Generalized Precambrian geology of the Lake Superior region; adapted from King (1969); Sims (1976); gravity from Weber and Goodacre (1966) and Fowler and Kuenze (1978).



so that it strikes west to northwest in eastern Michigan. This older trough contains thick successions of metasedimentary rocks that are primarily miogeosynclinal in the north and eugeosynclinal in the south (Sims, 1976). The younger trough is marked by the Keweenawan rocks of Figure 3.2 and by the mid-continent gravity high. Together they form a remarkable rift system (Chase and Gilmer, 1973) that extends northeasterly from eastern Kansas, through eastern Nebraska, central Iowa, to eastern Minnesota and western Wisconsin, where it begins to change strike from northeast to east-west (Figure 3.2). The rift continues to the east, roughly conforming to the outline of Lake Superior, and at the east margin of Lake Superior, it apparently changes strike again to develop a southeast trend (Bayley and Muehlberger, 1968). This horseshoe-shaped structure is filled with metavolcanic and metasedimentary rocks; primarily basalt and gabbro in the lower part and clastic sedimentary rocks in the upper part. The metasedimentary-metavolcanic rocks in the rift are too young (1100-1000 m.y.) to be hosts for placer uranium minerals, but the rift itself is critical to our discussion because it separates rocks of the older Proterozoic geosyncline which are of possible interest.

Metasedimentary and metavolcanic rocks that fill the older Proterozoic geosyncline of the Lake Superior region probably range in age from 2100 m.y. to 1900 m.y., so the older metasedimentary rocks in this geosyncline are a reasonable target for uranium exploration. As shown in Figure 3.2, metasedimentary rocks of the Huronian Supergroup, which contain the Canadian Blind River-Elliot Lake deposits, crop out about 200 km east of the easternmost outcrop of metasedimentary rocks in the older Proterozoic geosyncline. If we were to remove the late Proterozoic rift, assuming E-W rifting perpendicular to the line of gravity highs, and fit these areas as they may have been about 1800 m.y. ago we would find that the Canadian metasedimentary rocks should either fit exactly with the United States section or be slightly north of the main United States section (if the strike of the Lake Superior geosyncline changes to the southeast as it appears to in Figure 3.2). In any event, on these grounds it would appear reasonable to assume that a corelation can be made between these two Early Proterozoic successions of metasedimentary rocks.

This correlation was indeed made by Van Hise (1892) and Van Hise and Leith (1911) and the term Huronian was used for the metasedimentary successions in both Canada (Logan and Hunt, 1855) and the United States. Following his classic study of the metasedimentary rocks on the North Shore of Lake Huron, W. H. Collins (1925, p. 112-114) refined this correlation and suggested that the "Lower Huronian" of the Marquette District of the United States (Van Hise and Leith, 1911) was correlative with his lower Bruce series of the North Shore of Lake Huron. However, he noted that no reasonable correlation could be made between overlying rocks, and it is clear from his discussion that he had some doubt about the overall correlation. After fifteen years of re-study of the iron-bearing districts of Michigan by geologists of the United States Geological Survey, James (1958, p. 33-35) suggested that the term Huronian be abandoned for the metasedimentary successions in the United States because of substantial uncertainty in correlation between the type Huronian section on the North Shore of Lake Huron and the metasedimentary succession in the United States. James (1958) justly pointed out that on lithologic grounds the Bruce series (Lower Huronian Supergroup of Canada) does not correlate very well with the "Lower Huronian" as defined in the Marquette District by Van Hise and Leith, and that a more reasonable correlation is between the Fern Creek Formation of Dickinson County, Michigan and the Gowganda Formation (tillite) of the Huronian Supergroup of Canada.

This would make the entire sequence of Michigan younger than uranium-bearing units of Canada.

Following James' 1958 proposal, terminology for the Proterozoic metasedimentary successions have been completely revised as shown in Table 3.1. The term Marquette Range Supergroup (Cannon and Gair, 1970) is now used to describe the Proterozoic succession in Michigan, the term Animikie Group is used for equivalent rocks in Minnesota, and the term Huronian Supergroup is now used only to describe the Proterozoic succession in Canada. As stated in 1970 by Cannon and Gair (1970, p. 2844), detailed investigations of both the Huronian Supergroup and Marquette Range Supergroup to that time had not produced sufficient evidence for correlation of the two sequences. This indicates that the assumption of east-west rifting across the Keweenawan trough is overly simplistic and does not satisfactorily explain the pre-rifting paleogeography.

Geochronological studies of the past decade have not entirely solved the above correlation problem, but these studies indicate that rocks of the Huronain Supergroup are older than those of the Marquette Range Supergroup. The rocks of the Huronian Supergroup are cut by the Nipissing diabase dated as approximately 2160 m.y. (Van Schmus, 1965; Fairbairn and others, 1969) and they lie on a basement of Archean rocks dated as 2400-2600 m.y. They are therefore bracketed in age between about 2500 m.y. and 2150 m.y. In contrast, Banks and Van Schmus (1971; 1972) have shown that metamorphic rocks that underlie basal beds of the Marquette Range Supergroup in the Felch Trough of northern Michigan were subjected to a severe thermal event 2100-2000 m.y. ago; an event that is not recorded in the overlying metasedimentary rocks. Thus, Van Schmus (1976) proposed that rocks of the Marquette Range Supergroup are younger than approximately 2100-2000 m.y. Also, rhyolite tuff from the Hemlock Formation which is about in the middle of the Marquette Range Supergroup

		Gunflint range Minnesota and Ontario Goldich et al. (1961) Tanton (1931)	Mesabi range White (1954)	Cuyuna range Schmidt (1963) Marsden (1972)	Gogebic range Leith et al. (1935)	Western Marquette range Cannon and Klasner (1972)	Eastern Marquette Gair (1975), Gair and Thaden (1968)	Iron and Dickinson counties, Michigan James (1958)	Subdivisions of Cannon and Gair (1970)	
		(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	(No equivalent rocks)	Fortune Lakes Slate Stambaugh Formation Hiawatha Graywacke Riverton Iron-formation Dunn Creek Slate	Paint River Group	Marquette Range Supergroup Middle Precambrian (X)
Middle Precambrian	Animikie Group	Rove Formation	Virginia Formation	Rabbit Lake Formation	Tyler Slate	Michigamme Formation: Upper slate mbr. Bijiki Iron-fm. mbr. Lower slate mbr. Clarksburg Volcanics mbr. Greenwood Iron-formation mbr.		Badwater Greenstone Michigamme Slate Fence River Formation Hemlock Formation	Baraga Group	
						Goodrich Quartzite	Goodrich Quartzite	Goodrich Quartzite		
						Unconformity	Unconformity	Unconformity		
		Gunflint Iron-formation	Biwabik Iron-formation	Trommald Formation	Ironwood Iron-formation	Negaunee Iron-formation Siamo Slate	Negaunee Iron-formation Siamo Slate	Vulcan Iron-formation	Menominee Group	
		Kakabeka Quartzite	Pokegama Quartzite	Mahnomen Formation	Palms Quartzite	Ajibik Quartzite	Ajibik Quartzite	Felch Formation		
				Unconformity	Unconformity		Unconformity	Unconformity		
		(No equivalent rocks)	(No equivalent rocks)	Trout Lake Formation Slate and quartzite	Bad River Dolomite Sunday Quartzite	(No equivalent rocks)	Wewe Slate Kona Dolomite Mesnard Quartzite Enchantment Lake Formation	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	Chocolay Group	
		Unconformity	Unconformity	Unconformity	Unconformity	Unconformity	Unconformity	Unconformity		
Lower Precambrian		Granite; greenstone	Granite; greenstone	Gneiss	Granite; greenstone	Gneiss	Gneiss; Mona Schist	Dickinson Group	Lower Precambrian (W)	

Table 3.1. Correlation chart of Early Proterozoic metasedimentary rocks in the Lake Superior region. Taken from Sims (1976, p. 1097). Rocks in Michigan are called the Marquette Range Supergroup; rocks in Minnesota are called the Animikie Group. Precambrian W = Archean; Precambrian X = Early Proterozoic. Reproduced from Economic Geology, 1976, vol. 71, p. 1097.

(Table 3.1) yielded a date of 1950 m.y. and a minimum age for the succession is given by 1850 m.y. metamorphic and igneous events of the Penokean orogeny (Banks and Van Schmus, 1972). Van Schmus thus placed deposition of the Marquette Range Supergroup between about 2000 and 1850 m.y.--at least 150 m.y. later than deposition of the Huronian Supergroup. The critical date is the age of deformation of reactivated basement underlying the Marquette Range Supergroup. Sims (1976) suggested on geologic evidence that the basement gneiss was reactivated after deposition of the Marquette Range Supergroup, and if this could be demonstrated conclusively beds below the Hemlock Formation might be anywhere between 2500 m.y. and 1910 m.y. in age.

Thus, there is a legitimate question concerning the age of the lower part of the Marquette Range Supergroup, and inasmuch as prospecting for uranium-bearing, quartz-pebble conglomerates in the Lake Superior region depends on recognition of metasedimentary rocks older than 2000 m.y. we will consider two alternatives. First, that all rocks of the Marquette Range Supergroup are indeed younger than rocks of the Huronian Supergroup. Second, that some of the units of the Marquette Range Supergroup below the Hemlock Formation may be older than 2000 m.y. and correlative with rocks of the Huronian Supergroup.

If all rocks of the Marquette Range Supergroup are younger than 2000 m.y., we believe that the best possibility for prospecting for uranium would be in any metasedimentary succession that underlies the Marquette Range Supergroup. Table 3.1 shows that in most of the Lake Superior area Marquette Range Supergroup rocks are underlain by gneiss, granite, and greenstone of Archean (?) age. An exception is in Iron and Dickinson Counties, Michigan where James and others (1961, p. 12-21) described a sedimentary-volcanic sequence named the Dickinson Group. The metasedimentary rocks of the Dickinson Group unconformably underlie metasedimentary rocks of the Marquette Range Supergroup, and they have been classed as Lower Precambrian by James and others (1961). Recent geochronological studies by Van Schmus (1976) suggested that rocks of the Dickinson Group may be Archean because zircons from a gneiss that intrudes the Dickinson Group have been dated as 2600 m.y. However, zircons from a gneiss that is said to underlie the Dickinson Group have been dated as 2400 m.y. In any event, it appears that the Dickinson Group rocks are older than 2000 m.y. and are a reasonable target for uranium exploration.

The Dickinson Group is a three-fold unit consisting of the East Branch Arkose (oldest), the Solberg Schist, and the Six-mile Lake Amphibolite (youngest). The East Branch Arkose is a thick-bedded arkose with numerous beds of conglomerate. Locally, the arkose and conglomerates are interbedded with volcanic rocks. Vitreous quartzite is said to be the most abundant clastic in the conglomerate which also contains clasts of quartz, granite gneiss, slate or schist (James and others, 1961, p. 14-15). The East Branch Arkose is probably a fluvial deposit and appears to be a better than average target for uranium exploration.

The second possibility is that beds below the Hemlock Formation of the Marquette Range Supergroup may correlate with rocks of the Huronian Supergroup. The best lithologic correlation is probably the one suggested by James (1958) and Young (1966; 1970), who pointed out that the Fern Creek Formation, which is at the base of the Marquette Range Supergroup, might correlate with the Gowganda Formation of the Huronian Supergroup (Table 3.1). The Fern Creek Formation was defined by Pettijohn (1943) who believed that the unit should not be considered simply a basal phase of the overlying Sturgeon Quartzite as was done by Bayley (1904) because the Fern Creek Formation contained a graywacke conglomerate that was probably tillite and laminated argillite with rafted cobbles that were probably also of glacial origin. The glacial origin is questionable (Bayley and others, 1966) and as noted by James and others (1961) it is not a very thick (~ 85 meters) or widespread unit. The Fern Creek Formation was apparently deposited on an erosional surface and is confined to topographic lows on the old surface. However, if it is a glacial deposit it may indeed correlate with the Gowganda Formation of the Huronian Supergroup which is clearly of glacial origin. Furthermore, the overlying Sturgeon Quartzite is an aluminum-rich quartzite which is similar to and may correlate with the Lorrain Quartzite of the Huronian type-section (Church and Young, 1970).

Even if the above correlation between the Fern Creek and Gowganda is correct, there may not be too much promise for uranium-bearing, quartz-pebble conglomerates because all significant uranium deposits in the Huronian Supergroup are several thousand feet lower in the section than the Gowganda Formation and quartz-pebble conglomerates in the Lorrain Formation, which overlies the Gowganda in the type area, contain iron-oxides and monazite as the principal heavy minerals and were considered by Roscoe (1973) to have been deposited under oxidizing conditions. Thus, the Fern Creek Formation and Sturgeon Quartzite, as well as equivalent quartzite units in Minnesota (Table 3.1) are not especially favorable targets for fossil-placer uranium deposits.

Fossil-placer gold deposits are not restricted in age and could be present in any fluvial quartzite with a favorable source area. With this in mind, it is interesting to note the similarity between the metasedimentary successions of the Gunflint and Mesabi Ranges of Minnesota (Table 3.1) and the Jacobina Series and Minas Series of Brazil. In both areas, a basal arkose-conglomerate unit overlies an erosional unconformity and gradationally underlies units containing banded iron formation and, in Brazil, the arkose-conglomerate units contain fossil-placer gold and uranium deposits. In Minnesota, the Pokegama and Kakabeka quartzites nonconformably overlie an Archean granite-greenstone terrain. The erosion surface has up to 30 m of relief. Quartzites which fill the depressions reach thicknesses of 60 m and grade up-section into banded iron formation (Sims, 1976). According to Sims (1976), these rocks were deposited in a miogeosynclinal zone and were derived from the Archean granite-greenstone terrain to the north. This type of terrain is favorable as a source of gold and uranium and the gold would be transportable even in an oxidizing environment of deposition. Thus, conglomerates and quartzites in these basement depressions might contain gold even if they are too young to contain detrital uranium.

Our best opinion in this review of the Lake Superior region is that there are no rocks in this area that correlate with the Canadian Huronian section on the North Shore of Lake Huron. As shown schematically in Plate I, we consider that the Lake Superior region was probably elevated during deposition of the Huronian strata and that the bulk of the rocks in the United States Lake Superior area are either older or younger than the Canadian Huronian.

There are many unanswered questions on age of the various metasedimentary rocks, however, and individuals who wish to prospect in this general area might keep track of the rapidly developing geochronological studies. For the present, it seems to us that rocks equivalent to the Dickinson Group are the best possibility for prospecting for uranium-bearing, quartz-pebble conglomerates and that basal Proterozoic quartzite-conglomerate units in the Gunflint, Mesabi, and Cuyuna ranges might contain fossil gold placers.

WYOMING PROVINCE

GEOLOGIC SETTING

The Wyoming Province is a geochronological province containing rocks of Archean age that are similar to rocks of the Archean Superior Province of Canada and the Great Lakes region. Engel (1963) suggested that the Wyoming Archean rocks were continuous with those of the Superior Province and he used the name Superior-Wyoming Province for the combined entity. However, as shown in Figure 3.1, this continuity has not been substantiated by more recent subsurface data in North and South Dakota and southern Canada. Instead, it appears that Archean rocks in Wyoming are separated from the main mass of Archean rocks in the Superior Province by rocks which were strongly deformed about 1700 m.y. ago (Goldich and others, 1966; King, 1976). The Wyoming area is therefore much like the Slave Province of Canada, which is also an isolated block of Archean rocks surrounded by rocks deformed in later orogenic events (Figure 3.1). Thus, several workers (including Houston and others, 1968; Condie, 1969b; Hills and Armstrong, 1974) have modified Engel's terminology and have defined the Wyoming Province as the area in Wyoming and adjacent states which is underlain by Archean rocks. Like the Canadian Superior Province, the Wyoming Province consists of a gneissic terrain containing isolated greenstone belts. Also like the Superior Province, the Wyoming Province Archean basement has younger Proterozoic sedimentary basins near its margins that contain thick piles of metasedimentary and metavolcanic rocks.

As shown in Figure 3.3, the boundaries of the Wyoming Province are only approximately known. This is because Archean rocks are exposed only in the cores of the Laramide uplifts and these constitute less than ten percent of the outcrop area in Wyoming and far less than ten percent of the area in Montana, South Dakota, Utah, Idaho, and Colorado--the rest being covered by a thick sequence of Phanerozoic rocks. Also, the geology along the western

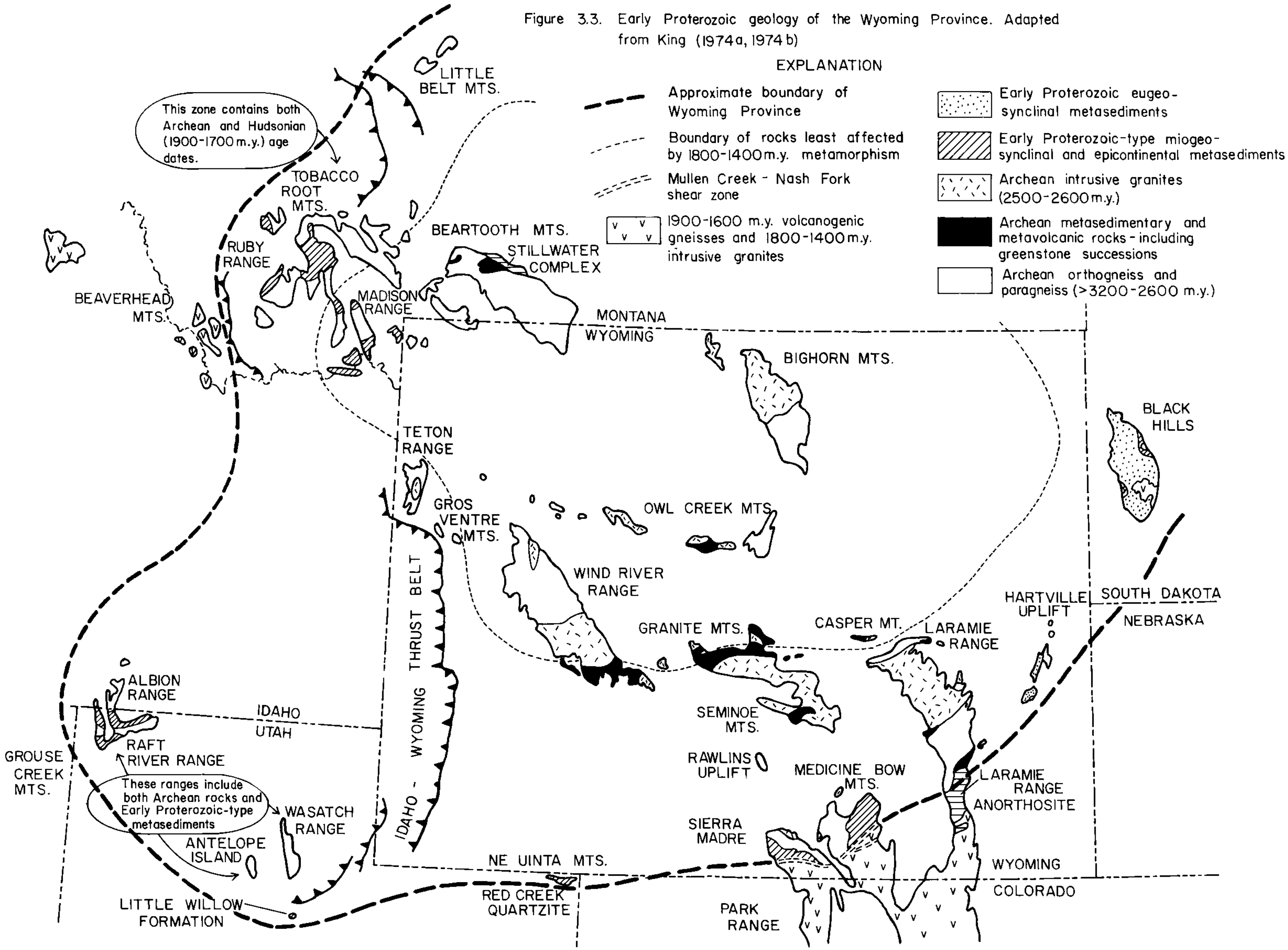


Figure 3.3. Early Proterozoic geology of the Wyoming Province. Adapted from King (1974a, 1974b)

boundary is drastically complicated by deformation associated with the Idaho-Wyoming and Montana thrust belts beginning in the late Mesozoic. This means that interpreting the Early Precambrian geology is like working with a jig-saw puzzle with less than ten percent of the parts and only a few edge pieces. Geologists working in the Wyoming Province do have one major advantage, however, and that is the generally superb exposures that characterize the outcrop areas.

The northern and northwestern margins of the Wyoming Province are poorly defined because of the scarcity of outcrops. However, Archean rocks are known as far north as the Little Belt Mountains of Montana (Figure 3.3). Here, zircons in gneisses give an age of 2450 m.y. (Catanzaro, 1966) while whole rock ages were reset to about 1900 m.y. during a Middle Proterozoic thermal episode probably related to the Hudsonian orogeny of Canada. King (1976) suggested that Archean and Hudsonian dates were intermingled in the Little Belt Mountains and in the ranges of southwestern Montana so that the northwestern boundary of the Wyoming Province is a wide zone which exhibits a gradational change from Archean dates in the Wyoming Province to reset Hudsonian dates to the northwest. This zone is shown in Figure 3.3. According to King (1976, p. 48), this gradational boundary between two Precambrian provinces contrasts to the boundary between the Superior and Churchill Provinces of the Canadian Shield which is marked by an abrupt structural change.

The southwestern and southern margins of the Wyoming Province are also poorly defined. However, Archean rocks have been reported from the Albion Range (Armstrong and Hills, 1967) and from the Farmington Canyon complex of the Wasatch Range (Carl Hedge, personal communication, 1978). The Archean rocks were strongly overprinted in the Wasatch Range during the 1700 m.y. Hudsonian orogeny so most rocks yield 1700 m.y. dates (Crittenden and others, 1971). This suggests that the southwestern margin of the Wyoming Province may also be a zone of intermingled Archean and Hudsonian dates. The farthest south rocks which are interpreted to be in the Wyoming Province are metasedimentary rocks of the Red Creek Quartzite and Little Willow Formation (Figure 3.3). Hansen (1965) reported a minimum age of 2320 m.y. for the Red Creek Quartzite and a subsequent thermal episode at 1520 m.y. and King (1976, p. 58) suggested that the Little Willow Formation resembles the Red Creek Quartzite in that it is a paraschist without cross-cutting intrusives.

The eastern margin of the Wyoming Province is similar to the northwestern margin in that it appears to be a zone of transition between the Archean ages to the west and rejuvenated Hudsonian metamorphic ages to the east, in South Dakota. Archean rocks crop out in two small domes in the Black Hills of South Dakota and are overlain by miogeosynclinal, then eugeosynclinal metasediments (Figure 3.3). The metasediments were strongly metamorphosed and deformed during the 1700-1800 m.y. Black Hills orogeny of Goldich and others (1966). In addition, subsurface and geophysical data (Lidiak, 1971; King, 1976) indicate that the Precambrian rocks which underlie central and western South and North Dakota were probably also intensely deformed and rejuvenated during the Hudsonian (1700-1800 m.y.) orogeny. Thus, as shown in Figure 3.1, the area of North and South Dakota is considered to be a southern extension of the Canadian Churchill Province.

The best exposed margin of the Wyoming Province--in fact, the only edge piece of the puzzle, is in southeastern Wyoming (Figure 3.3). Here, outcrops of Precambrian rocks are continuous from Wyoming south into Colorado in three separate ranges: the Sierra Madre, the Medicine Bow Mountains, and the Laramie Range. In each range, there is an abrupt transition between Archean and Early Proterozoic rocks to the north and Middle Proterozoic rocks to the south. In the Medicine Bow Mountains, this boundary or transition zone is the northeast trending Mullen Creek - Nash Fork shear zone (Houston and McCallum, 1961). This shear zone continues southwest into the Sierra Madre (Houston and others, 1975), where it either changes strike to east-west or is offset by a later fault system. In the Laramie Range the projected position of the shear zone is occupied by an anorthosite complex of Middle Proterozoic age (Hills and Armstrong, 1974). Little is known about this boundary east or west of these ranges but inferred extensions will be discussed in a later section.

The Mullen Creek - Nash Fork shear zone is up to 7 km wide in the Medicine Bow Mountains and contains cataclastic rocks ranging from mylonite and blastomylonite to augen gneiss. This zone is both a geologic-structural boundary and a geochronologic boundary (Houston and others, 1968). Rocks north of the shear zone are Archean rocks and low-grade Early Proterozoic miogeosynclinal metasediments (Figure 3.3). These rocks have many characteristics like those of the Lake Superior region of the United States and the Superior and Southern Provinces of the Canadian Shield. In fact, reasonable lithologic correlations can be made between Early Proterozoic rocks in southeastern Wyoming and the Great Lakes area (Plate 1). This analogy suggests that the metasediments north of the shear zone are an attractive target for uranium-bearing conglomerates.

In contrast, rocks south of the shear zone are mainly eugeosynclinal facies metasedimentary and metavolcanic rocks and associated intrusive bodies. These rocks are intensely deformed and metamorphosed to amphibolite facies. None of the rocks south of the shear zone has yielded an Archean age and the oldest ages found so far are 1900-1700 m.y. Rocks south of the shear zone probably represent island arc-type successions and current interpretations are that island arcs located in Colorado and other areas were accreted on to the Wyoming Province possibly about 1700 m.y. ago (Hills and others, 1975; Hills and Armstrong, 1974; Hills and Houston, 1979) and that the shear zone represents the suture along which this accretion took place. The important fact for the uranium geologist is that rocks south of this shear zone appear to be too young to be of interest as an exploration target for uranium-bearing quartz-pebble conglomerates.

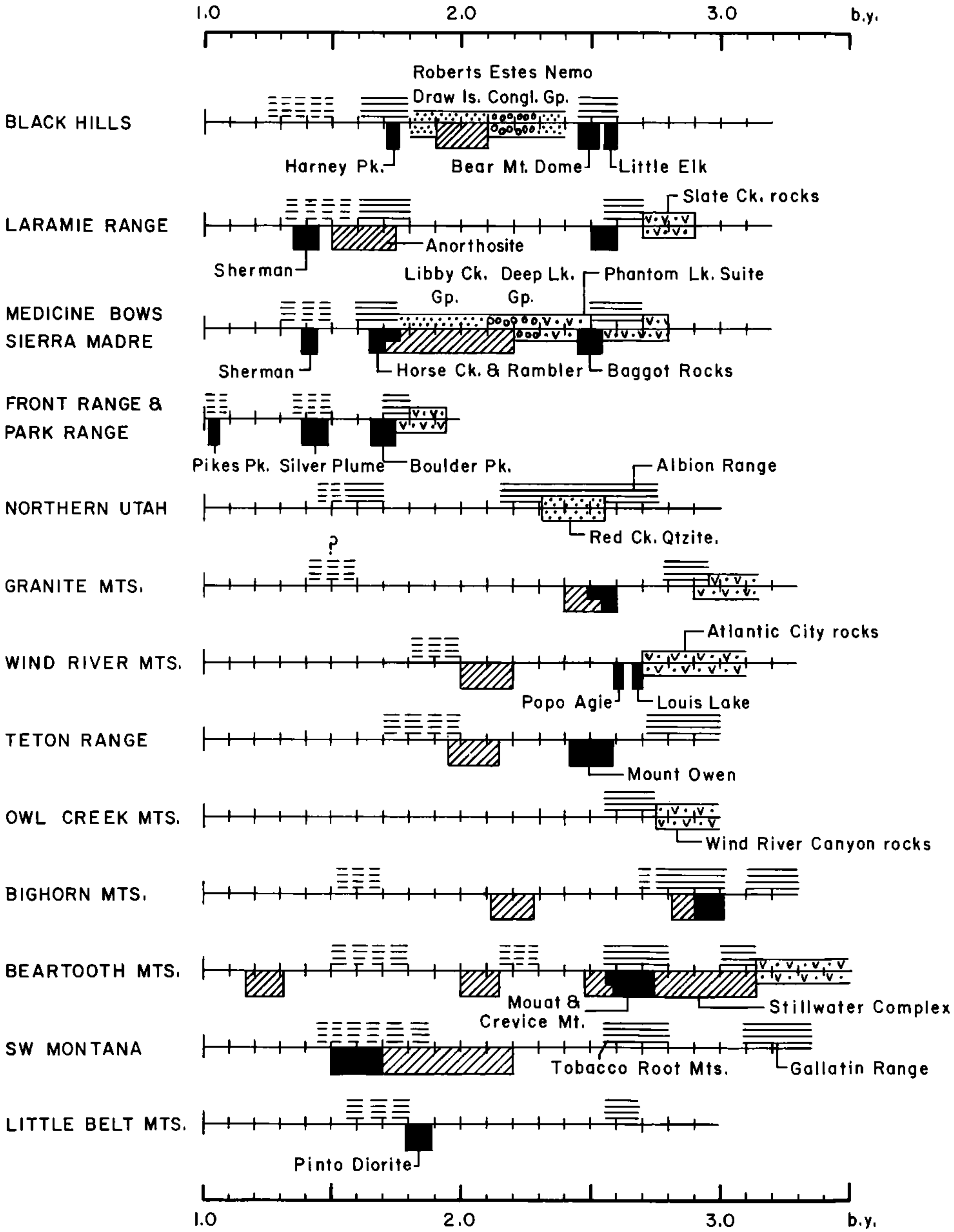
In the following review of the Wyoming Province we will emphasize Late Archean and Early Proterozoic metasedimentary rocks of the Wyoming Province that might be hosts for uranium-bearing, quartz-pebble conglomerate, but we will also discuss other aspects of the Precambrian history so the reader can better understand the geologic setting of the metasedimentary rocks.

ARCHEAN ROCKS

Figure 3.3 shows that there are three main types of Archean rocks in the Wyoming Province: granulite to amphibolite facies paragneisses and orthogneisses (most common); granitic to tonalitic batholiths; and a few Archean metasedimentary successions, including greenstone belts. The geologic history of these rocks is summarized in Figure 3.4.

Gneissic Terrain

The gneissic terrain of the Wyoming Province includes some of the most ancient rocks in North America. As shown in Figure 3.4, rocks older than 3000 m.y. have been reported in the Beartooth Mountains (Catanzaro and Kulp, 1964), Wind River Range (Cannon and others, 1966), Laramie Mountains (Johnson and Hills, 1976), and Granite Mountains (Peterman and Hildreth, 1978) and gneissic rocks 3000-2600 m.y. are present in most of the uplifts of the Wyoming Province. One of the most complete summaries of the geologic and geochronologic history of this gneissic terrain is that of Reed and Zartman (1973),



EXPLANATION

- Heating without significant metamorphism
- Principal regional metamorphic episodes
- Intrusion of mafic magmas
- Intrusion of granite magmas
- Deposition of clastic sediments
- Deposition of uraniferous conglomerates
- Deposition of volcanics and volcaniclastic sediments

note - <u>width of boxes implies uncertainty in dating events - not duration of events</u>

Figure 3.4. Schematic summary of the Precambrian history of the Wyoming Province, adapted from Reed and Zartman (1973) and Condie (1977). Width of boxes represents uncertainty in dating an event <u>not</u> the duration of the event. Sources of data:

<u>Black Hills</u> - Zartman and Stern (1967); Bayley (1972); Kleinkopf and Redden (1975); Hills (1977; 1979).
<u>Laramie Range</u> - Hills and Armstrong (1974); Subbarayudu and Hills (1974); Johnson and Hills (1976); Holden (1978).
<u>Medicine Bows and Sierra Madre</u> - Hills and others (1968); Hills and others (1975); Divis (1977); Hills and Houston (1979).
<u>Front Range and Park Range</u> - Peterman and others (1968); Snyder and Hedge (1978).
<u>Northern Utah</u> - Hansen (1965); Armstrong and Hansen (1966); Armstrong and Hills (1967); Armstrong (1968).
<u>Granite Mountains</u> - Peterman and others (1971); Rosholt and others (1973); Houston (1973); Peterman and Hildredth (1978).
<u>Wind River Mountains</u> - Basset and Giletti (1963); Condie and others (1969); Naylor and others (1970); Bayley and others (1973).
<u>Teton Range</u> - Reed and Zartman (1973).
<u>Owl Creek Mountains</u> - Giletti and Gast (1961); Granath (1975).
<u>Bighorn Mountains</u> - Heimlich and Banks (1968); Heimlich and Armstrong (1972); Stueber and others (1976); Barker (1976); Stueber and Heimlich (1977).
<u>Beartooth Mountains</u> - Catanzaro and Kulp (1964); Brookins (1968); Powell and others (1969); Nunes and Tilton (1971); Baadsgaard and Mueller (1973); Reid and others (1975); Page (1977).
<u>SW Montana</u> - Scholten and others (1955); Giletti (1966); King (1974); Spencer and Kozak (1975); Condie (1977).
<u>Little Belt Mountains</u> - Catanzaro and Kulp (1964); Woodward (1970).

who studied rocks in the Teton Range of western Wyoming. The oldest rocks of the Teton Range are interlayered biotite gneiss, plagioclase gneiss, amphibole gneiss, and amphibolite. These gneisses contain conformable bodies of quartz monzonite gneiss and a coarse metagabbro that is intrusive into but was metamorphosed and deformed with older units. The quartz monzonite gneiss and metagabbro were metamorphosed approximately 2700 to 3000 m.y. ago (Figure 3.4); therefore the interlayered gneiss and amphibolite are even older. These older rocks are cut by plutons and swarms of undeformed dikes of quartz monzonite and pegmatite dated as about 2500 m.y. old (Figure 3.4), and the entire complex is cut by undeformed but slightly altered tholeiitic diabase dikes of uncertain age. Reed and Zartman (1973) suggest that the diabase dikes may have been emplaced prior to or during a widespread thermal event, 1300 to 1500 m.y. ago, because biotite in the wall rocks of one major dike has a K-Ar age of about 1500 m.y.

Elsewhere in the Wyoming Province Archean gneissic rocks have had an equally complex or even more complex geologic history. For example, the Beartooth Mountains, mapped by Arie Poldervaart of Columbia University and his students and more recently by Reid and coworkers (1975) and Page (1977), are composed largely of granitized and metamorphosed quartzo-feldspathic gneiss with inliers of metasedimentary rocks which are more common to the northwest. According to Reid and others (1975), the northwestern Beartooths have been affected by four separate orogenies about 3100, 2600, 2200, and 1700 m.y. ago. These are shown in Figure 3.4 as regional metamorphic episodes. The earliest of these orogenies metamorphosed pre-existing sedimentary rocks indicating a pre-3100 m.y. geologic history. Then, sometime between 3150 and 2750 m.y. ago, the Stillwater layered mafic complex was emplaced (Page, 1977) and this was intruded by granitic rocks about 2750 m.y., during the second major orogeny (Figure 3.4). Mafic dikes and sills were emplaced during several different geologic events and are both syn- and post-tectonic in origin. Structural patterns in the Beartooths are extremely complex--Reid and others (1975) postulated seven superimposed deformational events, but the major structural pattern is of south- and southeast-plunging folds (Figure 3.1).

In the Bighorn Mountains Osterwald (1955; 1959) and Hoppin (1961) have shown that the Precambrian core consists of older gneiss and schist, amphibolite, and rare layers of tuff and agglomerate, and Heimlich (1969) suggested that granitic rocks in the northern Bighorns were formed by granitization of the metamorphic rocks. The gneiss has a northeast-trending fabric and the granitic rocks occupy a northwest plunging synform. These granitic rocks are cut by diabase dikes, some of which, in turn, were cut by and invaded by the same granite they cut (Figure 3.5). Heimlich and Armstrong (1972) have dated a large number of these rocks in the Bighorn Mountains by K-Ar methods, and they suggested that the event that formed the crystalline rocks (both gneiss and granitic rocks) ended about 2750 m.y. ago. Stueber and Heimlich (1977) suggested on the basis of Rb-Sr whole rock dates that this major period of metamorphism and granitization took place about 2850 m.y. ago (Figure 3.4). This date agrees with the 2830 m.y. date obtained by Stueber and others (1976) on the syn-tectonic mafic dikes. The last igneous event in the Bighorns was emplacement of post-tectonic mafic dikes about 2200 m.y. ago (Stueber and others, 1976).

More recent geologic mapping by Barker (1976) of Archean rocks in the southwestern Bighorn Mountains revealed two major episodes of magmatism, deformation, and metamorphism. Rocks formed in the early episode are felsic gneisses with minor amphibolite both of which may be volcanic. These gneisses

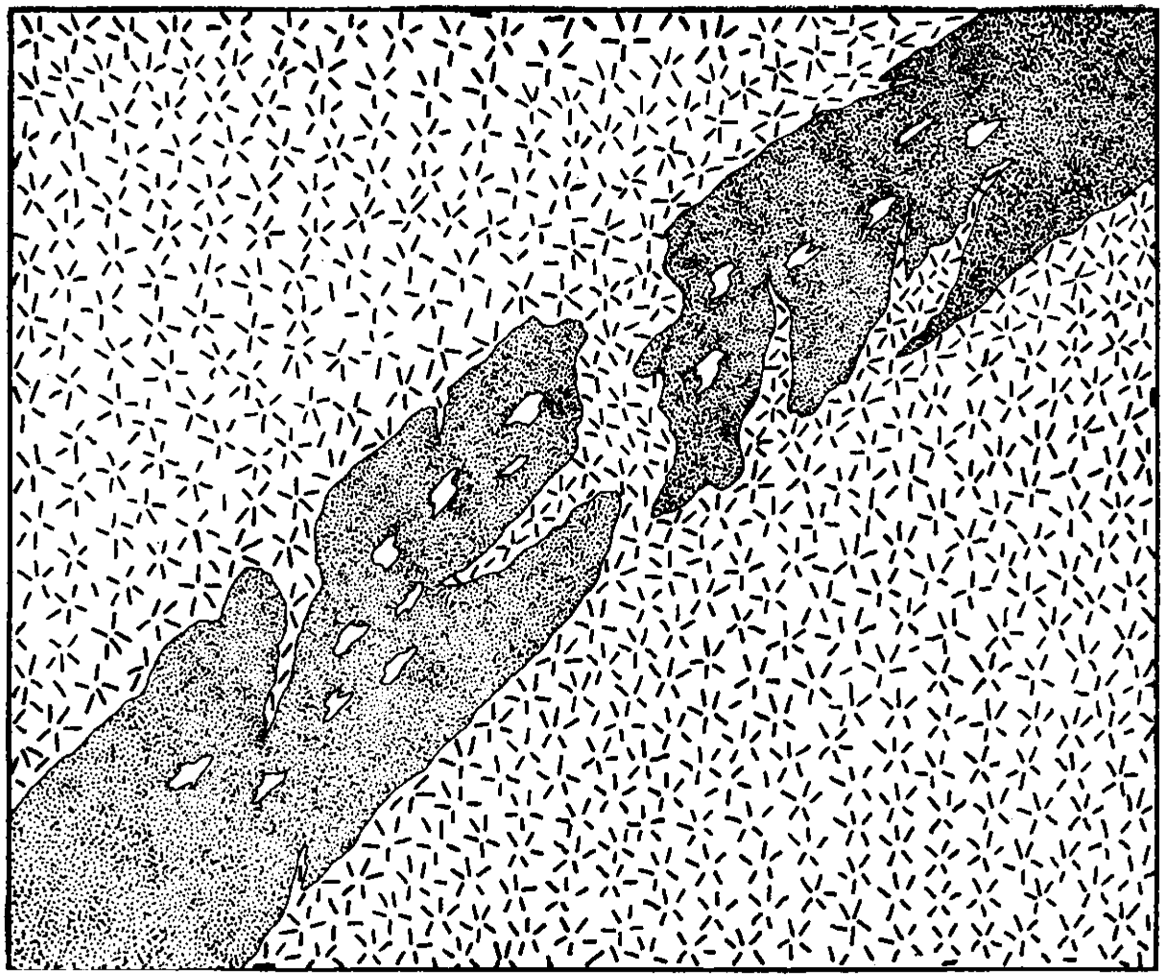


Figure 3.5. Sketch of altered mafic dike from the Archean of Wyoming showing dike that presumably cut granite; partially engulfed by the same granite.

were intruded by a syn-tectonic quartz monzonite, and the early episode ended with the intrusion of tonalite dikes about 3000 m.y. ago (Barker, 1976). The late episode began with the emplacement of syn-tectonic felsic igneous rocks that were metamorphosed to middle amphibolite facies and were mildly deformed. The late episode ended with intrusion of dikes of metabasalt. According to Barker (1976), these late episode rocks may correlate with northern Bighorn units dated as 2850 m.y. by Heimlich and Banks (1968) and Stueber and Heimlich (1977).

In most other ranges of the Wyoming Province, the earliest history is recorded by remnants of hornblende gneisses, felsic genisses, metamorphic schists of probable volcanic origin, rare layers of iron formation, quartzite, and marble. These rocks have been so intensely metamorphosed and granitized that little remains of the original characteristics of the rocks.

Nevertheless, Figure 3.4 shows some common elements in the geologic history of the various ranges. First, there appear to be very early periods of high-grade metamorphism and granite formation about 3000 m.y. ago and clearly many of the Wyoming Province rocks are older than this metamorphic event. These pre-3000 m.y. rocks are mainly restricted to the central core of the Wyoming Province--the Beartooth Mountains, Teton Range, Bighorn Mountains, Wind River Range and Granite Mountains. This is essentially the area shown in Figure 3.3 as the area least affected by later (∿ 1700 m.y.) metamorphism and it may represent an Early Archean nucleus of the Wyoming Province. A second widespread geologic event was the introduction of syn-orogenic granites and tonalites in the interval 2700-2500 m.y. This event took place throughout the Wyoming Province and was approximately synchronous with a world-wide period of granite emplacement which took place near the end of the Archean (∿ 2600 m.y. ago). A third common thread in the geologic history of the Wyoming Province appears to be an episode of emplacement of post-tectonic mafic intrusives about 2200-2000 m.y. Admittedly, only a few ranges (Beartooths, Bighorns, Wind Rivers) have even moderately well dated dikes. Nevertheless, other ranges have mafic dikes which may be in this age range (Black Hills, Medicine Bows, Sierra Madre, Tetons). Also, the Nipissing Diabase and equivalents of Canada (which may be analogous rocks in a similar tectonic environment) yield ages of about 2150 m.y. A fourth common event is metamorphism and emplacement of granitic rocks about 1700 m.y. ago. This event strongly affected the edges of the Wyoming Province as shown in Figure 3.3 and had least affect toward the center of the Province.

Archean Metasedimentary Successions

Greenstone Belts

Greenstone belts have been recognized in the southern Wind River Mountains (Bayley and others, 1973), Seminoe Mountains (Bishop, 1963; Bayley, 1968), Casper Mountain (Beckwith, 1939), Owl Creek Mountains (Gliozzi, 1967; Duhling, 1971; Houston, 1973; Granath, 1975), and the Granite Mountains (Houston, 1973; Peterman and Hildreth, 1978). All of these greenstone belts contain iron formation, quartzite, amphibole gneiss, amphibolite, and fuchsitic quartzite in variable proportions. In some areas, primary textures and structures in metamorphic rocks suggest pillow basalts, tuffs, conglomerates, agglomerates and graywackes.

The age relationships between rocks of the greenstone belts and surrounding gneissic terrains is questionable here as it is elsewhere in Early Precambrian shields. Bayley and others (1973) have made the most careful study of one of these belts, which is located in the southern Wind River Mountains, and they concluded that rocks of the greenstone belt are older than a post-tectonic batholith, the Louis Lake Batholith, which is approximately 2700 m.y. old. Bayley and others (1973, p. 28) therefore correlated the rocks of this greenstone belt with early Precambrian rock groups "such as the Couchiching Series, Keewatin Series, and Knife Lake (Seine) Group, and the Dickinson Group of the Felch District of Michigan," all of which they believed predated the 2700 m.y. old Algoman granite "sea" that makes up the major part of the Superior Province of the Canadian Shield.

We agree that the rocks of the Wyoming greenstone belts are indeed old, and although individual belts may vary in age, they are probably older than the associated Late Archean granites which, as Bayley and others have demonstrated, cut the greenstone belts. Also, these greenstone belts may not be as old as some of the gneissic rocks and remnants of metasedimentary and metavolcanic rocks contained within them. It is possible that much of the gneissic terrain formed about 3000 m.y. ago, or earlier, from rocks already in existence, and that the major greenstone belts were deposited in the interval 3000 m.y. to 2700 or 2500 m.y. on this gneissic terrain. Individual greenstone belts of various ages may then have been invaded by Late Archean post-tectonic granitic rocks emplaced 2700-2500 m.y. ago, during and slightly after the time of deformation of the greenstone belts.

We consider these greenstone belts to be generally poor targets for uranium exploration. Quartzites are rare and of limited extent and the environments of deposition of these rocks does not appear to have been favorable for concentration of uranium.

Beartooth Mountains

Archean metasedimentary rocks also crop out in several areas of the Beartooth Mountains. These are definitely Archean in age but are not typical greenstone successions. Instead, they appear to be a unique metasedimentary assemblage in the Wyoming Province. These rocks underlie a large area southwest of the Stillwater Complex in the northern part of the mountains (Page, 1977); they crop out in the North Snowy Block, in the extreme northwest corner of the mountains (Reid and others, 1975); and they occur as small inliers in gneisses in the eastern and southeastern Beartooths (Casella, 1969).

The metasedimentary rocks at the base of the Stillwater Complex are primarily fine-grained cordierite hornfels, but Page (1977) also described blue quartzite, iron formation, and a diamictite or paraconglomerate as interlayered with the hornfels. This metasedimentary succession probably contains both clastic sedimentary rocks and chemically precipitated sedimentary rocks and Page (1977) suggested that the bulk of the succession may be turbidites or marine clastic sedimentary rocks of some type. However, he also reported that the paraconglomerate contains dropstones and could be glacial in origin. The blue quartzite is a quartz-arenite (> 98 percent quartz) which is locally conglomeratic, containing rounded quartz pebbles up to 1.5 cm in diameter. This quartzite is interlayered with chlorite-quartz schists and reportedly grades up-section into magnetite-bearing iron formation.

Nunes and Tilton (1971) reported a Pb-isotope age of 3140 m.y. for zircons in the metasediments and they interpreted this age to be a minimum age. Also, the metasediments were thermally metamorphosed during intrusion of the Stillwater Complex which is greater than 2750 m.y. Thus, the metasediments are obviously Archean in age and appear to be older than 3000 m.y., at least 250 m.y. older than the greenstones of the southern Wind River Mountains.

There is no mention of radioactivity or uranium or thorium in any of these rocks (Page, 1977) and we consider the potential to be low because

Archean supracrustal rocks in high-grade terrains elsewhere in the world are generally uranium-poor. Nevertheless, a reconnaissance examination of the blue quartzite might be worthwhile for uranium or gold. Page (1977) suggested that the source area for the metasediments contained abundant mafic and ultramafic rocks and these are a good potential source of detrital gold. Excellent geologic maps are available (Page and Nokleburg, 1974) that show the distribution of the blue quartzite so reconnaissance prospecting for uranium and gold in this area should not be difficult.

Metasedimentary rocks in the North Snowy Block of the northwestern Beartooths were named the North Snowy Group by Reid and others (1975). They divided the group into seven formations: Barney Creek Amphibolite, George Lake Marble, Jewel Quartzite, Davis Creek Schist, Mount Delano Gneiss, Mount Cowen Gneiss, and Falls Creek Gneiss. According to Reid and others (1975), the Mount Delano Gneiss displays a minimum Pb-isotope age of 3000 m.y. and the Barney Creek Amphibolite, George Lake Marble, and Jewel Quartzite are older than the gneiss. Thus, these metasediments are older than 3000 m.y. and probably correlate with metasediments in the Stillwater block. Reid and others (1975) suggested that the metasediments were derived from a granitic source terrain and deposited as muds, quartz sands, and dolomitic carbonates which were intercalated with large volumes of basaltic tuffs and flows.

There are no reports of radioactive materials in the succession and the great age of the sediments suggests that the source area contained tonalites and other Early Archean granitic rocks which are not especially favorable sources of uranium. Also, even if originally present, uranium is likely to have been remobilized during the intense deformation and metamorphism which took place about 2700 m.y. ago (Reid and others, 1975). Thus, we consider these rocks to be a poor target for uranium. Nevertheless, a reconnaissance survey of the Jewel Quartzite would be relatively simple, using the maps of Reid and others (1975), and could conceivably turn up some gold or uranium. It should be noted that Reid and others reported accessory pyrite and pyrite replacing iron-oxides in the quartzite and inasmuch as pyritic quartzites are sometimes associated with uranium and gold, it might be fruitful to investigate the quartzites.

Metasedimentary rocks are also present in the eastern Beartooths. Casella (1969) described these as small, discontinuous units of quartzite, schist, amphibolite, gneiss and iron formation within a migmatite terrain. We consider these to be an unlikely target for uranium because of their limited extent and high degree of metamorphism (sillimanite is reported in some of the schists).

In summary, we interpret the metasediments in the Beartooths to be isolated remnants of ancient sediments which somehow escaped the intense metamorphism and deformation which the rest of the high-grade migmatite terrain experienced. Thus, they are more similar to supracrustal rocks in Archean high-grade terrains such as the Isua supracrustals of Greenland (Allaart, 1976) or the rocks of the Scourian Complex of Great Britain (Sutton, 1973) than to typical volcanogenic greenstone belts such as the Swaziland Sequence of South Africa (Anhaeusser, 1973) or greenstones of the Canadian Shield (Goodwin and Ridler, 1970). We consider them to be an unlikely host for uranium mineralization of any sort.

Northeast Laramie Range

Barlow (1950) described quartzite and mica schist in the LaPrele Creek-Boulder Creek area (T. 32 N., Rs. 73 and 74 W) of the northeastern Laramie Mountains. The Precambrian metasedimentary rocks may not be very extensive in this area inasmuch as Barlow shows quartzite in only one locality (sec. 24,

T. 32 N., R. 74 W.). We have shown these rocks as Archean metasediments in Figure 3.3 because of their limited distribution, their association with Archean granites and gneisses of the northern metamorphic complex of Condie (1969a), and the fact that these rocks are essentially on-strike with greenstones on Casper Mountain. However, very little is known of the Precambrian rocks in this general area so a reconnaissance study might reveal metasedimentary rocks of interest to the uranium geologist.

Central Laramie Range

Metasedimentary and metavolcanic rocks occur in several areas of Condie's (1969a) central metamorphic complex. Figure 3.3 shows the two largest of such outcrops. The northern outcrop is described by Condie (1969a) as an area of amphibolites which occurs near the boundary between the Laramie batholith and the central metamorphic complex. He suggested that these rocks are similar to metasedimentary rocks on Casper Mountain and in the Owl Creek Mountains and we interpret them, on this basis, to be part of a greenstone succession.

The southern outcrop is located along Slate Creek, just north of the Laramie anorthosite and the inferred extension of the Mullen Creek-Nash Fork shear zone (Figure 3.3). The generalized geology of this area is shown in Figure 3.6. The most abundant metasedimentary rocks are hornblende gneisses, amphibolites and quartzo-feldspathic gneisses, but there are also quartzites, conglomerates, marbles, calc-schists, and pelitic rocks (Toogood, 1967; Holden, 1978). Holden (1978, p. 7) suggested that the most likely protolith for the hornblende gneiss, amphibolite, and quartzo-feldspathic gneiss was a sequence of mafic and intermediate lavas and volcaniclastic sediments. However, the entire sequence of rocks is metamorphosed to highest almandine amphibolite facies (sillimanite is present in pelites) so primary features are

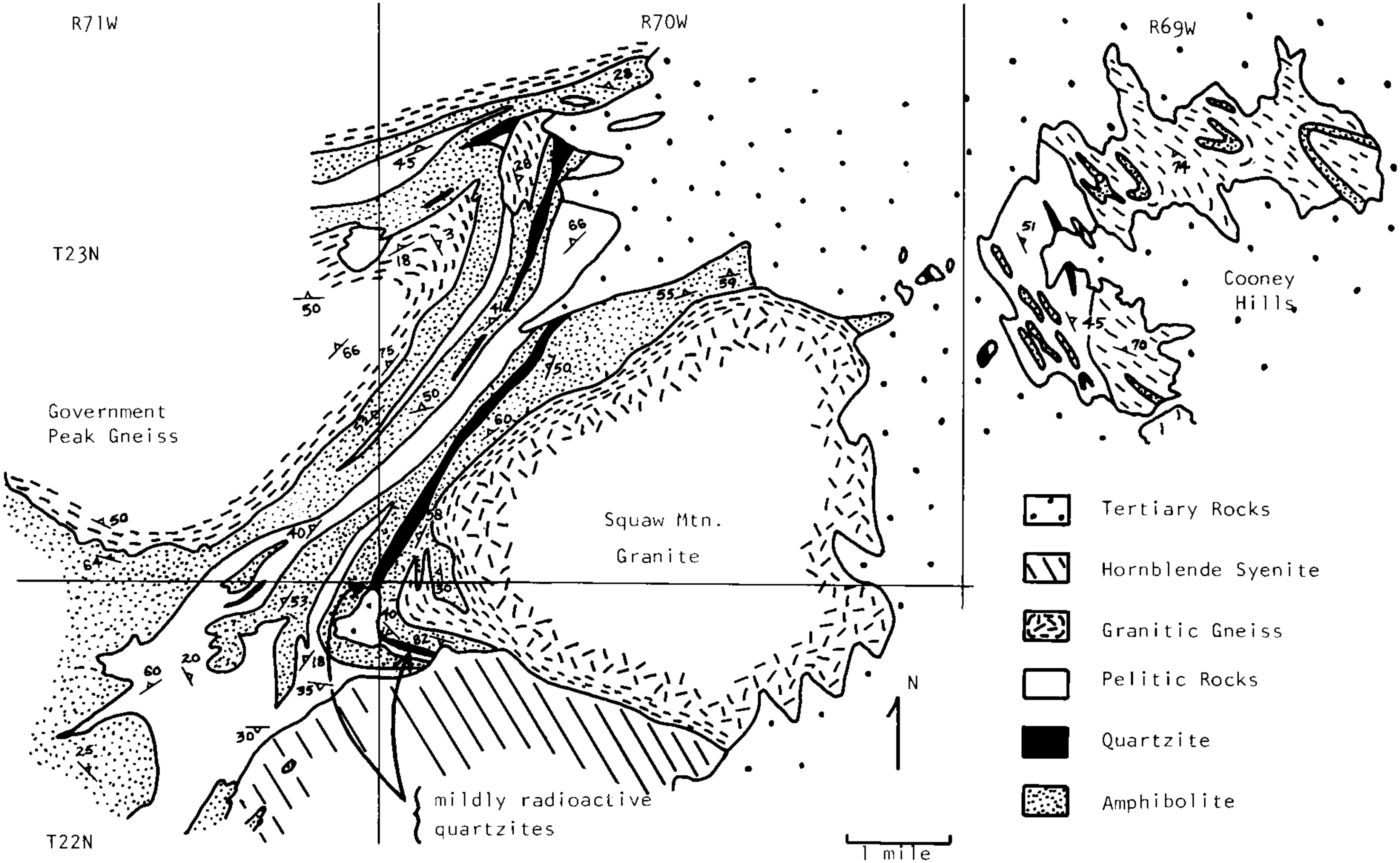


Figure 3.6. Generalized geology of the Slate Creek metamorphic terrain, adapted from Holden (1978), Hodge (1966), Toogood (1967) and Bothner (1967). Known radioactive quartzites occur on the SW side of the Squaw Mountain Granite.

absent and the origin of the rocks is difficult to ascertain (Fields, 1963; Bothner, 1967; Smith, 1967; Toogood, 1967; Hodge, 1966; Holden, 1978). We interpret these rocks to be a greenstone succession because of the predominance of amphibolite and pelitic sediments and the structural setting of the rocks--in a synform between two granite domes (Figure 3.6). However, the high metamorphic grade of these rocks is not typical of greenstone belts and this sequence could be a supracrustal succession which is transitional between true Archean greenstone belts and Early Proterozoic clastic sedimentary successions.

The age of these rocks has not been determined; the only date is a 1700 m.y. K-Ar date on hornblende reported by Hills and Armstrong (1974) and this is a minimum age. However, Holden (1978) suggested from field evidence that granitic rocks of the Squaw Mountain and Government Peak domes (Figure 3.6) intruded and deformed the metasediments. These granites yield dates of about 2600 m.y. (Hills and Armstrong, 1974) so if this is true, the metasediments are older than 2600 m.y.

Reconnaissance examination of the Slate Creek metasediments by the authors with a portable scintillometer has shown that mildly radioactive quartzite and conglomerate are present in several localities and may occur in one or more semi-continuous quartzite-conglomerate zones within the Slate Creek synform (Figure 3.6). The radioactivity of these rocks is only 3-4 times background and the quartzites are relatively thin units. Nevertheless, very few outcrops have been examined to date and therefore this area should be investigated more fully as a potential target for uranium-bearing rocks.

LATE ARCHEAN(?) OR EARLY PROTEROZOIC(?) METASEDIMENTARY SUCCESSIONS

Figure 3.3 shows that metasedimentary rocks crop out in several isolated ranges near the northwestern and southwestern margins of the Wyoming Province.

We have, somewhat arbitrarily, classified them in Figure 3.3 as Early Proterozoic-type successions because most of them contain abundant quartzites, marbles, pelitic rocks and other lithologies commonly found in Proterozoic successions, and we believe they should be investigated for uranium mineralization. However, these successions have been intensely deformed and metamorphosed and their age and original sedimentary characteristics are not well understood so it is also quite possible that some or all of these metasediments are Archean in age.

Southwestern Montana

Metasedimentary rocks are present in the Tobacco Root Mountains, Ruby Range, Gravelly Range, and Madison Range (Figure 3.7). These rocks are mainly paragneisses, mica schists and amphibolites, but there are also quartzites, marbles, phyllites and iron formation. The metasedimentary rocks in southwestern Montana are sometimes referred to as the Cherry Creek Group and this is the terminology used in Figure 3.7. Originally, the name "Cherry Creek Group" was used by Peale (1896) for a several thousand foot thick succession of marbles, mica schists, quartzites, and gneisses which crop out in Cherry Creek, in the eastern foothills of the northern Gravelly Range (Figure 3.7). Since then, however, many workers have applied the term to any pre-Beltian rocks in southwestern Montana which contain marble and quartzite and are of obvious sedimentary parentage. For example, Tansley and others (1933) and Reid (1957; 1963) have used the name to refer to the marble-bearing metasedimentary successions in the Tobacco Roots and Heinrich (1960) and Hadley (1969a; 1969b) applied the name to similar rocks in the Ruby Range and Gravelly Range. It should be noted, however, that the term is not universally accepted. For example, Witkind (1964; 1969; 1972; 1976) preferred not to use the term for metasedimentary rocks in the southern Madison Range

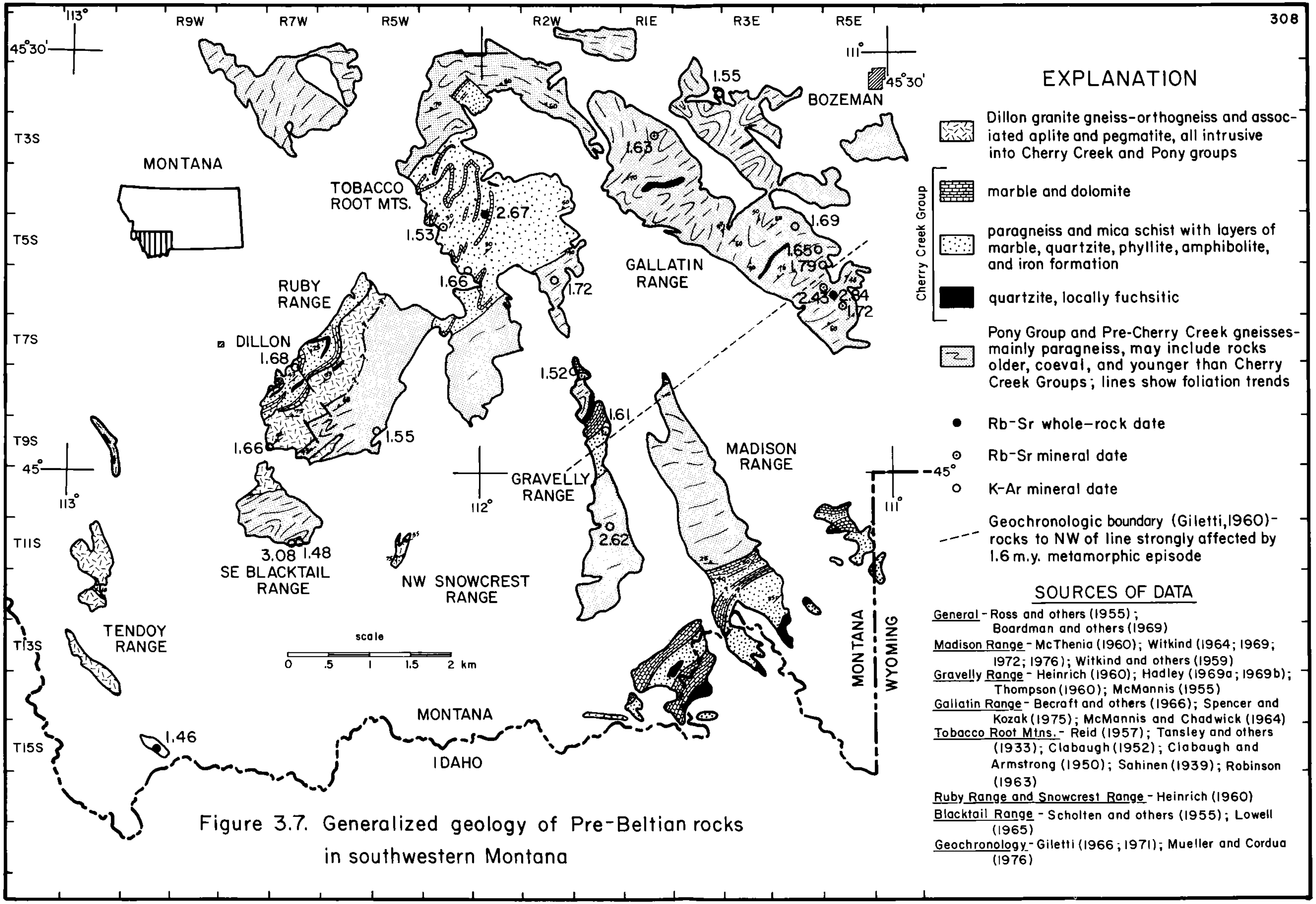


Figure 3.7. Generalized geology of Pre-Beltian rocks in southwestern Montana

and instead mapped the units according to age (Early Proterozoic) and lithology. Thus the designation of the rocks near the Montana-Idaho border in Figure 3.7 as Cherry Creek Group is ours, not Witkind's.

The relationship of the Cherry Creek Group to other gneissic rocks in southwestern Montana is not well understood. In the Tobacco Root Mountains, Tansley and others (1933) believed that the Cherry Creek Group unconformably overlies an older series of gneisses and schists, because they found inclusions of gneisses in Cherry Creek schists. They call these gneisses the Pony Series. A similar picture was presented by Heinrich (1960) in the Ruby, Blacktail, and Gravelly Ranges. He reported a series of pre-Cherry Creek gneissic rocks which are more intensely deformed and metamorphosed than the Cherry Creek rocks. However, this interpretation was disputed by Reid (1963), who suggested that the inclusions cited by Tansley and others (1933) were actually rotated boudins and that the Pony Series and Cherry Creek Group are part of the same depositional succession and were deformed together into recumbant isoclinal folds. In fact, Reid (1963) believed that the Pony Series in the Tobacco Roots is younger than the Cherry Creek Group because it is in a structurally higher position.

In Figure 3.7, we have reluctantly used the Cherry Creek--pre-Cherry Creek subdivision of Tansley and others (1933) and Heinrich (1960) because we know of no new regional interpretations to supersede it and because it seems likely to us that the marble-quartzite successions in the Gravelly Range and southern Madison Range will prove to be younger than the quartzofeldspathic gneisses. We are not confident this will prove to be true in the Tobacco Root Mountains and the Ruby Range. In these areas, thin marble and quartzite lenses appear to be infolded in units mapped both as the Cherry Creek Group and as pre-Cherry Creek gneisses (Figure 3.7) and the age relations between the two rock-types is still debatable. It seems probable that Reid's interpretation is correct--that both units are part of a paragneiss terrain which has been variably deformed and metamorphosed. If this is true, the metasediments in the Tobacco Roots and Ruby Range could be Archean. This appears to be supported by a 2670 m.y. date reported by Mueller and Cordua (1976) for Cherry Creek gneisses in the Tobacco Root Mountains (Figure 3.7) which is similar to Archean ages reported by Giletti (1971) for Pony Series gneisses in the Gallatin Range.

The third major unit shown in Figure 3.7 is the Dillon granite-gneiss, an intrusive, syn-orogenic granite which yields K-Ar dates of about 1400-1700 m.y. (Giletti, 1966). Giletti (1966) suggested that the Dillon granite intruded the older gneisses and metasediments during or shortly after a period of regional metamorphism about 1600 m.y. ago which also reset K-Ar dates in the gneisses and metasediments. Dillon granite-gneiss appears to crop out mainly in the western part of Figure 3.7. This led King (1976) to suggest that the Dillon granite may represent a Hudsonian plutonic event that was, at least partly, responsible for isotopically reconstituting Archean rocks within a zone extending from the geochronologic boundary shown in Figure 3.7 to the edge of the Wyoming Province (Figure 3.3).

In spite of the uncertainty about the age and contact relationships of the Cherry Creek Group, we consider it to be a reasonable target for uranium exploration. It does not appear to be a typical Archean greenstone belt because of the abundance of quartzites, marbles, and quartzo-feldspathic paragneisses; instead it is perhaps more reminiscent of Early Proterozoic-type successions. A good starting point for exploration might be the quartzites in the Gravelly Range and southern Madison Range because quartzites in these areas appear to be more extensive than quartzites reported from the Ruby Range and Tobacco Root Mountains (Figure 3.7). The maps of Hadley (1969a, 1969b) in the Gravelly Range and Witkind (1964, 1972, 1976) near the Montana-Idaho border show the distribution of quartzites and should greatly facilitate reconnaissance exploration.

Northern Utah and Southern Idaho

Archean and/or Early Proterozoic rocks of the Wyoming Province crop out in three areas in northern Utah and southern Idaho: the Albion-Raft River Range area; the Wasatch Range-Antelope Island area; and the northeastern Uinta Mountains (Figure 3.3). None of these areas contain known radioactive conglomerates and most of the rocks have been strongly deformed more than once since they were deposited--suggesting that even if uranium-bearing placers were deposited in these areas it is questionable that they would be preserved. Nevertheless, these areas deserve some attention from uranium explorationists because some of the metasediments appear to be the right age for placer uranium deposition and they are located in a favorable tectonic position--near the southwest margin of the Wyoming Province.

Albion and Raft River Ranges

The Albion Range in Idaho and the Raft River Range and Grouse Creek Mountains in Utah contain Precambrian rocks that were involved in a period of Mesozoic deformation and metamorphism associated with the Sevier orogeny (Armstrong, 1968). This area is believed to be part of a Cordilleran core complex or infrastructure analogous to the Shuswap terrain of the Canadian Cordillera. These core complexes are interpreted to be zones of updoming of plasticized material which instigated gravitational spreading and imbricate thrust faulting in the suprastructure to the east (Price and Mountjoy, 1970). Progress has been slow in unraveling the Precambrian history of these complexes because the intense Mesozoic deformation has strongly overprinted Precambrian events. However, geochronologic work has, at least, demonstrated the existence of Archean basement in several core complexes (Armstrong and Hills, 1967; Compton and others, 1977; Duncan, 1978).

In the Albion and Raft River Ranges, Archean rocks crop out in a series of domes. Armstrong (1968) referred to these as mantled gneiss domes in the Albion Range, implying that in-situ remobilization of basement rocks domed and deformed the mantling sediments (Eskola, 1949). However, Compton and others (1977) have presented a more detailed geologic picture for the Raft River Range and Grouse Creek Mountains and here, it appears that the Archean rocks are part of a large autochthon which has been partly concealed by three generations of allocthonous thrust sheets containing variably deformed Paleozoic and Mesozoic metasediments. Figure 3.8 shows the distribution of autochthonous areas (or gneiss domes) in both ranges.

Archean rocks in the Albion Range were named the Green Creek Complex by Armstrong (1968). This complex includes quartzo-feldspathic gneisses, schists, amphibolites, and quartzites which yield a poor Rb-Sr whole rock isochron of 2.46 b.y. (Armstrong and Hills, 1967). Correlative rocks in the Raft River Range and Grouse Creek Mountains were divided by Compton (1972, 1975) into four mappable units: an older quartz-schist; amphibolite; metamorphosed trondhjemite and pegmatite; and metamorphosed adamellite. According to Compton and others (1977) the adamellite yields a Rb-Sr age of 2510 $\pm$ 170 m.y.

The older schist of the Raft River Range is one possible target for fossil placer uranium or gold deposits. According to Compton (1972, 1975) this unit was originally shale, siltstone, arkose, and pebbly mudstone. The older schist is cross-cut by adamellite and therefore represents an Archean clastic metasedimentary succession (Figure 3.8). On the northeastern flank of the

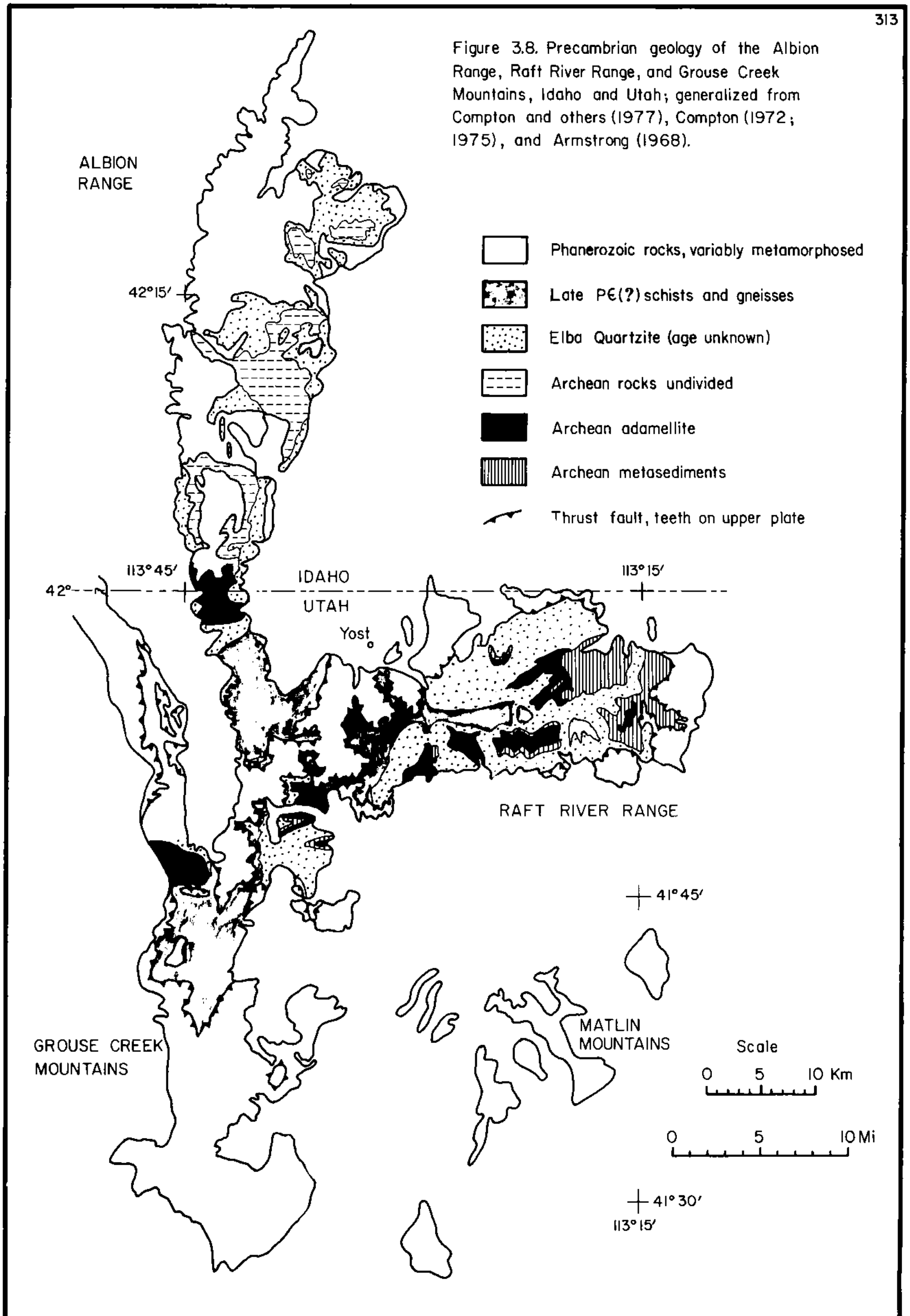


Figure 3.8. Precambrian geology of the Albion Range, Raft River Range, and Grouse Creek Mountains, Idaho and Utah; generalized from Compton and others (1977), Compton (1972; 1975), and Armstrong (1968).

Raft River Range, the lowest 70 m of this unit is semischistose cross-bedded quartzite and this area might be a good place to begin reconnaissance sampling.

Another possible target for fossil-placer mineralization is the Elba Quartzite. This unit was named by Armstrong (1968) and its distribution is shown in Figure 3.8. The Elba Quartzite unconformably overlies the Archean schist and adamellite and the unconformity shows several meters of relief. There is commonly a gradation between adamellite and the basal Elba Quartzite through a mica-schist zone interpreted to be a regolith (Armstrong, 1968; Compton, 1975). A basal quartz- and quartzite-pebble and cobble conglomerate up to 15 m thick locally fills depressions in the erosion surface. This is overlain by 30-50 m of white, cross-bedded micaceous quartzite then by 30-80 m of quartz-schist which contains hematitic quartzite layers.

The age of the Elba Quartzite is not known. It is obviously post-Archean but Armstrong (1968) and Compton (1972, 1975) interpreted it to be Cambrian in age because they found no important stratigraphic break between the Elba and overlying fossiliferous Paleozoic rocks. Also, Armstrong (1976, p. 21) reported a Rb-Sr whole rock date of 570 $\pm$ 80 m.y. from quartz-schists in the Elba Quartzite. In contrast, Crittenden (1979) suggested that the Elba could be substantially older because it contains distinctive lithologies such as fuchsitic quartzite, hematitic schist, amphibolite, and metatuff which are not found in Paleozoic sections elsewhere in the region.

In a brief visit to the Raft River - Albion area, we found no indication of above-background radioactivity in the Elba. Our tentative conclusion was that the Elba is too young to contain fossil-placer uranium. This conclusion is supported by the presence of hematite, rather than pyrite, in the conglomerates and quartzites plus the Rb-Sr date and stratigraphic evidence mentioned earlier and summarized by Compton and Todd (1979). The older schist, however, is still a reasonable target for uranium exploration.

Wasatch Range-Antelope Island area

Precambrian crystalline rocks crop out on the southeastern margin of the Great Salt Lake and along the western margin of the Wasatch Range in north-central Utah (Figure 3.3). These rocks are thought to be metasedimentary in origin and may be Archean or Early Proterozoic in age. These sequences are the Farmington Canyon Complex, the Little Willow Series, and the Formation of Facer Creek.

The Farmington Canyon Complex has been described by Eardley and Hatch (1940), Larsen (1957), Bell (1952), and Cohenour and Thomson (1966). It crops out in the Wasatch Range, from Salt Lake City to Ogden, and on Antelope Island, in the southwestern part of the Great Salt Lake. The complex is composed mainly of felsic paraschists and paragneisses which are permeated by sills and dikes of granitic and pegmatitic material. It also contains quartzites, quartz schists, arkosic quartzites, meta-graywacke, and quartzite-pebble conglomerates. The entire complex has been folded into north and northwest trending folds and metamorphic grades range from greenschist facies to granulite facies, with amphibolite facies metamorphism most common (Bell, 1952). According to Eardley and Hatch (1940, p. 61), the complex has a minimum thickness of 3600 m.

The age of the Farmington Canyon Complex is not precisely known. It is at least 1600-1700 m.y. old, which is the age of some of the felsic igneous rocks which intrude the paragneisses and the age of the major metamorphism (Whelan, 1970). However, more recent geochronologic work by Carl Hedge (personal communication, 1978) suggests that at least some of the gneisses may be Archean in age.

Whether Archean or Early Proterozoic, the Farmington Canyon complex is a reasonable target for reconnaissance uranium exploration. The complex as a whole is thought to have been a sedimentary sequence containing second cycle sediments (the conglomerates contain lithic fragments) and first cycle arkoses and deposition is interpreted to have taken place close to the source of the detritus (Eardley and Hatch, 1940; Cohenour and Thomson, 1966). Thus the Farmington Canyon Complex appears to have some similarities to Early Proterozoic-type clastic successions. Unfortunately, the degree of deformation and metamorphism is appreciably higher than for known uranium-bearing Early Proterozoic successions so, even if uranium was originally present, it may well have been remobilized during the 1600-1700 m.y. metamorphism. With this in mind, the explorationist should not overlook the possibility for secondary uranium deposits.

Little Willow Series

The Little Willow Series is a sequence of folded metasedimentary rocks which crops out in a small area (less than two km in diameter) in the core of the Uinta Arch, just north of Little Cottonwood Canyon near Salt Lake City (Crittenden, 1965; Granger and others, 1952). It consists of quartzites, quartz-mica schists, and stretched-pebble schists cut by mafic igneous rocks. According to Cohenour (1959, p. 36), the lower grade of metamorphism of the Little Willow Series, the absence of migmatites and pegmatites, and the abundance of mafic igneous rocks distinguishes it from the Farmington Canyon Complex to the north. The age of the Little Willow Series is unknown and various correlations have been suggested. Blackwelder (1949, p. 27) suggested it may be equivalent to the Red Creek Quartzite of the northeast Uinta Mountains, the metasediments in the Raft River and Albion Ranges, and the Farmington Canyon Complex; Granger and others (1952) agreed that it may be equivalent to the Farmington Canyon Complex; Cohenour (1959, p. 37) suggested that the Little Willow Series is younger than the Farmington Canyon Complex because it lacks injected felsic material and he correlated it with rocks in the Raft River Range; and Cohenour and Thomson (1966, p. 39) suggested the Little Willow Series was older than the Farmington Canyon Complex. Obviously the available data permit several interpretations. We agree with Cohenour (1959) and King (1976) that the absence of felsic intrusive material and the abundance of mafic intrusive material in the Little Willow Series may be significant. Using this as a criterion, it is reasonable to speculate that the Farmington Canyon Complex may be an older, possibly Archean, basement and that the Little Willow Series may be a relic of a somewhat younger metasedimentary succession.

If so, the quartzites and pebble schists of the Little Willow Series are interesting from a uranium exploration viewpoint. Indeed, a one day visit to the area by Karlstrom and Rex Cole of Bendix Field Engineering Corporation revealed quartzites and conglomerates with 2-4 times background radioactivity. The most radioactive unit found was a laminated, fine-grained, pyritic quartzite on the south side of Little Willow Creek. This was structurally overlain by a conglomeratic graywacke unit which is shown on Crittenden's (1965) map. The conglomeratic unit is also radioactive (2½ times background). The conglomerate is matrix supported and contains sparse pebbles of quartz, granite, schist, and mafic rock in a pyritic, schistose matrix. The abundance of basaltic clasts in the conglomerate plus the immature nature of most of the clastic rocks suggests to us that the Little Willow Series represents a volcano-sedimentary succession, possibly of Late Archean or Earliest Proterozoic age. If so, it may be analogous to the Phantom Lake Metamorphic Suite in southeastern Wyoming or the Livingstone Creek-Thessalon Formations in the Huronian Supergroup and may contain thin radioactive conglomerate units.

Formation of Facer Creek

Metasedimentary rocks also crop out in the upper plate of the Willard Thrust Sheet, north of Huntsville, Utah (Crittenden and others, 1971). These rocks unconformably underlie upper Precambrian diamictites and have been informally referred to as the Formation of Facer Creek by Crittenden and Wallace (1973, p. 125). The unit consists of vitreous quartzite, fuchsitic quartzite, pelitic schist, amphibolite, mica-schist, and rare pegmatite and Rb-Sr dates on muscovite yield a minimum age of 1.6-1.7 b.y. (Crittenden and others, 1971). Crittenden (1979) suggested that the rocks resemble the Elba Quartzite of the Raft River Range. Geologic maps of this unit are not available to us so we do not know how extensive it may be. Nevertheless, it is a reasonable target for uranium exploration.

Red Creek Quartzite of the northeast Uinta Mountains

Hansen (1965, p. 22-31) described intricately folded amphibolite facies metasedimentary rocks located in five horst-like blocks exposed along the northeastern margin of the Uinta Mountains of Utah and Colorado (Figure 3.3). They are collectively referred to as the Red Creek Quartzite, but a variety of metasedimentary rocks are present. Unfortunately, the Red Creek Quartzite is apparently too deformed and fragmented to establish a stratigraphic succession, but rock types include feldspathic quartzite, nearly pure quartzite, mica schist, and metalimestone. The quartzites contain kyanite and chrome-micas, and some of the more massive, cleaner quartzites resemble the Medicine Peak Quartzite of the Medicine Bow Mountains. However, the Red Creek Quartzite is more highly deformed than the Medicine Peak Quartzite and apparently is older, because muscovite from the mica schist yields a Rb-Sr metamorphic age of 2320 m.y.

Hansen (1965) also suggested that the Red Creek Quartzite is similar to Precambrian rocks found in the Albion and Raft River Ranges, the Farmington Canyon Complex, and the Little Willow Series beacuse all of these areas contain green fuchsitic quartzites, pelitic schists and amphibolites. However, none of these correlations has been proven and the relationship of the Red Creek Quartzite to metasedimentary successions in southeastern Wyoming and central Utah is still not understood. The 2320 m.y. minimum age, if correct, indicates the Red Creek Quartzite is either Archean or Earliest Proterozoic. At present, we lean toward an Archean age because, as Hansen (1965, p. 31) suggested, the 2320 m.y. date may be a relic of the 2500 m.y. metamorphic episode recorded elsewhere in the Wyoming Province which was down-graded by local open-system conditions. By this interpretation, the Red Creek Quartzite would represent an Archean clastic succession analogous to Compton's (1972, 1975) older schist of the Raft River Range and to the Farmington Canyon Complex of the Wasatch Range.

No uraniferous rocks have been reported from the Red Creek Quartzite. Hansen (1965, p. 25) reported analyses of two metaquartzites and neither of them contained detectable uranium, thorium, or gold. Nevertheless, the uranium potential of the Red Creek Quartzite cannot be established until enough work is done on the stratigraphy of the unit to determine whether it contains fluvial sedimentary rocks. If it does, it appears to be the right age to have potential for uranium-bearing fossil placers.

EARLY PROTEROZOIC-TYPE METASEDIMENTARY SUCCESSIONS

Metasedimentary successions of known Early Proterozoic age crop out near the eastern and southeastern margins of the Wyoming Province (Figure 3.3)--in the Black Hills of South Dakota, and the Hartville Uplift, Medicine Bow Mountains and Sierra Madre of Wyoming. These Proterozoic metasedimentary rocks have many characteristics like those of the Huronian Supergroup of Canada and the Marquette Range Supergroup of the United States. We will discuss the Proterozoic succession in the Medicine Bow Mountains of Wyoming first because this area has the most complete section and contains rocks that probably correlate with both the Huronian and Marquette Range Supergroups.

Medicine Bow Mountains, Wyoming

This section of the report is a brief summary of on-going research by the authors in the Medicine Bow Mountains. More detailed accounts can be found in Houston and others (1968); Karlstrom (1977); Houston and others (1977); and Karlstrom and Houston (1979a, 1979b).

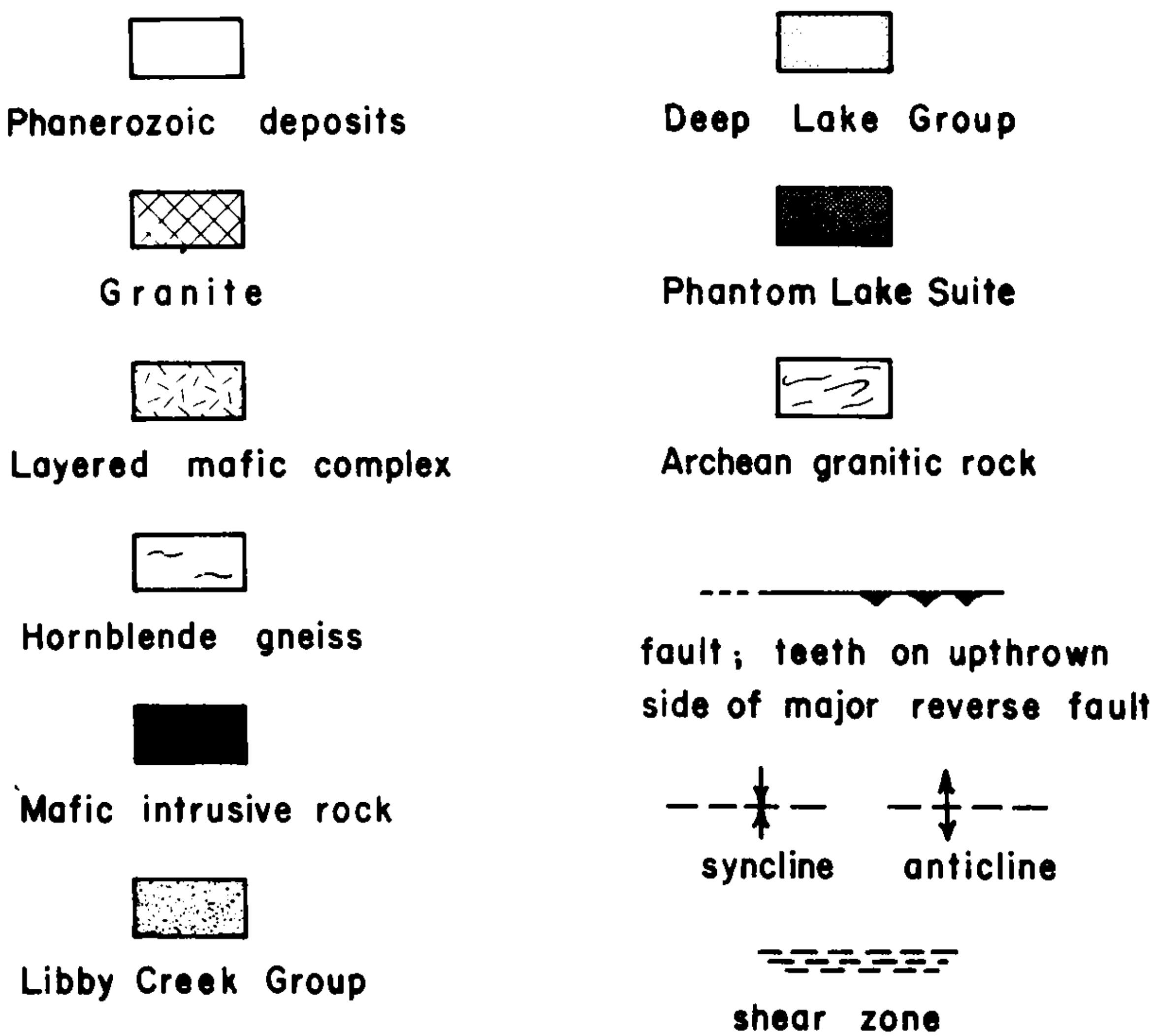
Geologic Setting

The Medicine Bow Mountains are a north-trending anticlinal uplift located in southeastern Wyoming about 50 km west of the city of Laramie (Figure 3.3). The mountains were uplifted during the Laramide orogeny (Late Cretaceous) and consist of a core of Precambrian rocks flanked by upturned Paleozoic and Mesozoic rocks and surrounded by basins filled with Tertiary sediments.

Precambrian rocks in the core of the mountains are divided into two geological provinces by the northeast-trending Mullen Creek - Nash Fork shear zone (Figure 3.9), which is a zone of mylonites and cataclastic augen gneisses up to 7 km wide (Houston and others, 1968). Precambrian rocks north of the shear zone are part of the Wyoming Province (Figure 3.3) and consist of an Archean basement overlain by Early Proterozoic metasedimentary rocks. The Archean basement is mainly quartzo-feldspathic gneiss and intrusive granite (Houston and others, 1968), both of which yield dates in excess of 2500 m.y. (Hills and others, 1968; Divis, 1977; Hills and Houston, 1979). The metasedimentary rocks form a 13 km thick section of epicontinental and miogeosynclinal clastic rocks which is divisible into three successions, shown in Figure 3.9: the Phantom Lake Metamorphic Suite, the Deep Lake Group, and the Libby Creek Group (Karlstrom and Houston, 1979a, 1979b). These rocks were deposited between 2500 and 1700 m.y. ago (Hills and others, 1968) and will be discussed in detail below.

Precambrian rocks south of the Mullen Creek-Nash Fork shear zone include quartzo-feldspathic and hornblende gneisses, mafic to ultramafic complexes, and intrusive granites (Figure 3.9). The gneisses appear to be eugeosynclinal facies metasediments which were deposited between 2000 and 1800 m.y. ago (Houston and others, 1968) and the intrusive rocks yield dates of around 1700 m.y. (Hedge and others, 1978; Hills and Houston, 1979). It should be emphasized that rocks south of the shear zone consistently yield dates younger than 1800 m.y. and are, therefore, not part of the Wyoming Province. Hills and Houston (1979) believe these rocks are island-arc sediments which were accreted onto the Wyoming Province during one or more craton-island arc collisions around 1700 m.y. ago. By this interpretation, the metasedimentary successions north of the shear zone represent a clastic wedge sequence deposited on an Atlantic-type continental margin. These sediments were deformed and the shear zone developed about 1700 m.y. ago, when an island arc collided with the clastic succession. Mafic rocks south of the shear zone are

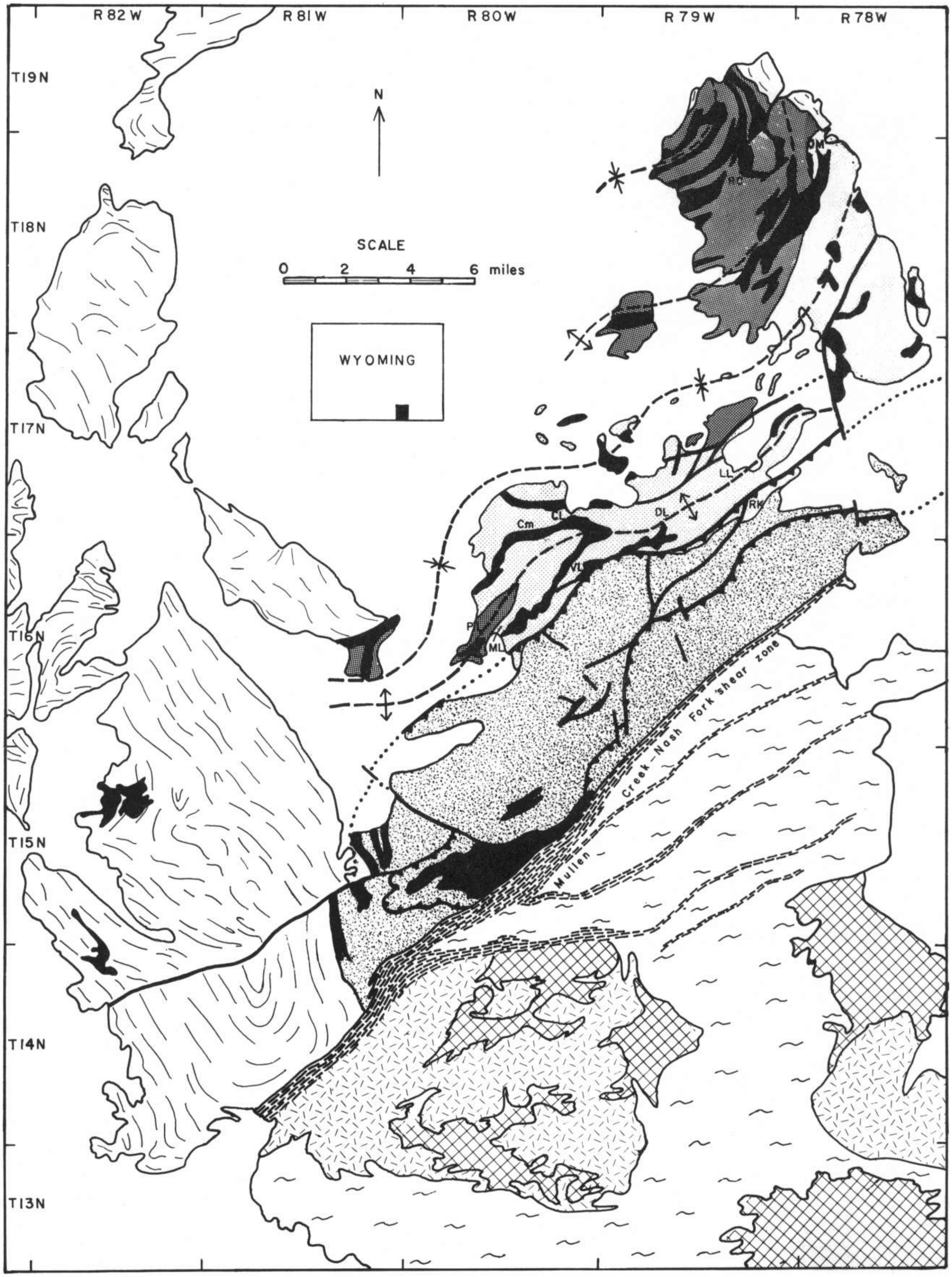
EXPLANATION



Explanation to Figure 3.9 -- Generalized geology of the Medicine Bow Mountains; adapted from Houston and others (1968) and Karlstrom (1977).

LABELED LOCATIONS

CL - Cascade Lake; Cm - Campbell Lake; DL - Deep Lake; LL Lindsey Lake; ML - Magnolia Lake; OM - Onemile Creek; PL - Phantom Lake; RC - Rock Creek; RK - Rock Creek Knoll; VL - Vagner Lake.


R82W
R81W
R80W
R79W
R78W
T19N
T18N
T17N
T16N
T15N
T14N
T13N
N
SCALE
0 2 4 6 miles
WYOMING
Cm
CL
DL
LL
RK
VL
ML
Mullen Creek-Nash Fork shear zone

interpreted by Hills and Houston (1979) as accreted oceanic crust, gneisses are interpreted to be volcanogenic sediments, and granodiorites are thought to have been derived from a southeast dipping subduction zone.

Distribution and Structure

Figure 3.9 shows the distribution and generalized structure of Precambrian rocks in the Medicine Bow Mountains. Archean basement rocks crop out mainly in the western and northwestern parts of the mountains. The dominant structures in Archean rocks are north-trending antiforms and synforms and superimposed northeast-trending folds (Karlstrom and Houston, 1979b).

Phantom Lake Metamorphic Suite rocks crop out mainly in the northeastern Medicine Bows (Figure 3.9). These rocks are folded into isoclinal folds with axial plane traces varying from east-west to north-south. Formation of these isoclinal folds and subsequent warping of axial plane traces are both considered to be the result of deformations which pre-dated deposition of the Deep Lake Group.

Rocks of the Deep Lake Group unconformably overlie the Phantom Lake Suite and occupy the north-central part of the Medicine Bows (Figure 3.9). This group has been folded into east-west and northeast trending open anticlines and synclines. These folds, in turn, have been warped by a later northwest-trending fold system (Karlstrom and Houston, 1979b).

Rocks of the Libby Creek Group form a south-dipping homoclinal succession in the central Medicine Bows. These rocks are interpreted to be miogeosynclinal rocks which were deposited farther off-shore than the epicontinental rocks of the Deep Lake Group and then brought into contact with Deep Lake Group rocks by movement along major northeast-trending thrust faults (Figure 3.9). This thrust faulting, and the northeast folding of the Libby Creek Group, are interpreted by Karlstrom and Houston (1979b) to be the result of plate margin tectonics associated with early movements along the Mullen Creek-Nash Fork shear zone about 1700 m.y. ago. The last recognizable deformation north of the shear zone was left lateral strike-slip movement on the shear zone and folding of earlier thrusts in the southwestern part of area of outcrop of the Libby Creek Group (Figure 3.9).

Thus, metasedimentary rocks in the Medicine Bow Mountains have undergone at least six folding episodes accompanied by one or more regional metamorphic episodes of greenschist to upper amphibolite grade and one or more episodes of intrusion of gabbroic sills and dikes. The sum of these deformations, coupled with the discontinuous nature of outcrop, has made stratigraphic and sedimentological reconstructions and the search for stratiform uranium mineralization unusually difficult.

Stratigraphy and Paleogeography

Phantom Lake Metamorphic Suite. The Phantom Lake Metamorphic Suite is a sequence of metasedimentary and metavolcanic rocks that has been deformed into isoclinal folds and metamorphosed to amphibolite facies. It is different from the overlying Deep Lake and Libby Creek Groups in that it is more highly deformed and metamorphosed and it contains a higher percentage of metavolcanic rocks. Phantom Lake Metamorphic Suite rocks crop out in the northeastern and west-central Medicine Bow Mountains, where they apparently lie unconformably on Archean basement and are overlain unconformably by rocks of the Deep Lake Group. In the west-central Medicine Bow Mountains a mafic sill lies between Phantom Lake Metamorphic Suite rocks and the Archean basement (Figure 3.9) and in the northeastern Medicine Bow Mountains rocks previously mapped as Archean basement (Houston and others, 1968) appear to be complexly interlayered with Phantom Lake Suite rocks so that the exact age relationships are unclear--although the Phantom Lake rocks are in a structurally higher (younger) position. In any event, from the viewpoint of the uranium geologist, no basal conglomerate has been recognized that marks the boundary between Archean basement and rocks of the Phantom Lake Metamorphic Suite. Phantom Lake Metamorphic Suite rocks include paraconglomerate, quartzite, phyllite, metavolcanic rocks and marble (Table 3.2). The metavolcanic rocks are largely basalt, most of which appears to be surface flows with amygdaloidal tops, but some basaltic pillow lava is present, and meta-agglomerates and metatuffs have been recognized. Quartzite of the Phantom Lake Metamorphic Suite is typically a fine-grained feldspathic variety but coarse-grained quartzite with local layers of quartz-pebble conglomerate also occurs and this is pyritic in places and is usually radioactive (2-4 times local background) especially where the pyrite is present. Amphibolite, hornblende gneiss and schist, and felsic gneiss are also interlayered with Phantom Lake Metamorphic Suite rocks. These metamorphic rocks are probably of volcanic origin and they are more common in the northeastern Medicine Bow Mountains where the rank of metamorphism is upper almandine amphibolite facies.

In the northeastern Medicine Bow Mountains, rocks of the Phantom Lake Metamorphic Suite are tightly folded and primary sedimentary structures necessary to determine the top of the succession are not always preserved. Nevertheless, a tentative stratigraphic succession has been put together from partial sections (Table 3.2), and the best current guess is that the total thickness of Phantom Lake Metamorphic Suite rocks exceeds 3,000-5,000 meters.

The age of metasedimentary and metavolcanic rocks of the Phantom Lake Metamorphic Suite has not been determined. Two samples of a staurolite-garnet-muscovite schist from a locality three kilometers southwest of Arlington (in T. 18 N., R. 79 W.) were analyzed by Hills and others (1968, p. 1771). These two points gave an isochron solution of 1913 m.y. which is probably a

Group	Cycle	NAME	THICKNESS METERS	LITHOLOGY	SEDIMENTARY FEATURES	SOURCE	HEAVY MINERALS	DEPOSITIONAL ENVIRONMENT
LIBBY CREEK GROUP		Headquarters Formation	650	phyllite quartzite phyllite paraconglomerate	varves trough cross-beds dropstones	NE		glacio-marine
		~~ unconformity ~~~						
DEEP LAKE GROUP	Cycle 3	Rock Knoll Formation	380	conglomerate quartzite	quartzite, pebbles, clay galls, cross-beds, ripple marks	ESE		fluvial shallow marine
DEEP LAKE GROUP	Cycle 3	Vagner Formation	120 - 400	phyllitic quartzite marble paraconglomerate	ripple marks, cross-beds dropstones?	ENE	pyrite	glacio-marine
		~~ unconformity ~~~						
DEEP LAKE GROUP	Cycle 2	Cascade Quartzite	150 - 1530	pebbly arkose pebbly quartzite quartzite	black chert pebbles, tabular cross-beds, trough cross-beds	ENE	apatite pyrite zircon	fluvial
DEEP LAKE GROUP	Cycle 2	Campbell Lake Formation	75	phyllite paraconglomerate	faint stratification		zircon apatite	marine glacial?
DEEP LAKE GROUP	Cycle 1	Lindsey Quartzite	440	quartzite	trough cross-beds	NE	zircon pyrite	fluvial-deltaic
DEEP LAKE GROUP	Cycle 1	Magnolia Formation (quartzite member)	420	quartz-granule conglomerate	trough cross-beds	N	zircon pyrite apatite	fluvial
DEEP LAKE GROUP	Cycle 1	Magnolia Formation (radioactive conglomerate member)	150	quartz-pebble conglomerate	planar bedding	NE	pyrite, zircon rutile, monazite	fluvial
		~~ unconformity ~~~						
PHANTOM LAKE METAMORPHIC SUITE		Upper Phantom Lake Metamorphic Suite	1500 - 2000	paraconglomerate metabasalt micaceous quartzite phyllite quartzite quartz-pebble conglomerate	angular clasts amygdules trough cross-beds tabular cross-beds channels	NE	pyrite monazite, zircon garnet, apatite	continental shallow-marine marine fluvial
PHANTOM LAKE METAMORPHIC SUITE		Lower Phantom Lake Metamorphic Suite	800	paraconglomerate quartzite metavolcanic				continental?
		~~ nonconformity ~~~						
		Archean basement		granitic rocks				

Table 3.2. Stratigraphy of the Phantom Lake Metamorphic Suite and Deep Lake Group of the Medicine Bow Mountains, Wyoming.

metamorphic date. Certainly two points are inadequate to establish a reliable isochron, but the date of 1913 m.y. is about two hundred million years older than ages determined by whole-rock Rb/Sr methods for schists of the Libby Creek Group. This information, though not conclusive, supports the concept that rocks of the Phantom Lake Metamorphic Suite were deformed and metamorphosed earlier than the metasedimentary rocks that overlie them.

Granite which resembles Archean granite of this area is locally in contact with rocks of the Phantom Lake Metamorphic Suite. In some areas, field relationships suggest that Phantom Lake Metamorphic Suite rocks lie unconformably on this granite, but in other areas granite has gradational or cross-cutting contacts with rocks of the Phantom Lake Metamorphic Suite. There are no reliable age determinations of these granitic rocks. Three samples were analyzed by Hills and others (1968, p. 1763) from gneissic granite near Arlington; these samples formed an isochron with an apparent age of 2350 m.y. with an error of $\pm$ 84 m.y. This is again too few samples for a reliable isochron, but it does suggest that some of the rocks of the Phantom Lake Metamorphic Suite are earliest Proterozoic or Archean in age.

Deep Lake Group. The Deep Lake Group is defined in the north-central Medicine Bow Mountains where the rocks of this group are best exposed and where they are least deformed and metamorphosed. A tentative stratigraphic succession has been established by Karlstrom (1977, p. 10) and Karlstrom and Houston (1979a, 1979b), who have subdivided the Deep Lake Group into six formations: the Magnolia Formation, Lindsey Quartzite, Campbell Lake Formation, Cascade Quartzite, Vagner Formation, and Rock Knoll Formation. Characteristics of formations are summarized in Table 3.2 and described below.

Magnolia Formation. The Magnolia Formation is composed of two members, a basal conglomerate member and an upper quartzite member. The basal conglomerate member is a radioactive, coarse-grained quartzite containing layers of quartz-pebble conglomerate. These basal beds of the Magnolia Formation vary from one locality to another. For example, in the Medicine Bow Mountains east of Arrastre Lake (near ML of Figure 3.9) a lenticular relatively coarse-grained, polymictic conglomerate with clasts up to 16 cm in diameter is present at the base of the unit. This conglomerate is up to 3 meters thick in places and is radioactive (7-8 times background). It is green to yellowish-red and contains rounded clasts of quartzite, micaceous quartzite, phyllite, metavolcanic rocks, and granite in a pyritic, arkosic matrix. The basal conglomerate is overlain by a more continuous quartz-pebble conglomerate that contains pebble-sized clasts of quartz, quartzite, fine-grained sericitic quartzite, phyllite, and altered rocks of probable volcanic origin in a sericitic, feldspathic-quartzite matrix. In the Arrastre Lake area the conglomerate member is not continuous; it pinches out west of Arrastre Lake where it apparently grades laterally into fine-grained quartzite. The conglomerate member that crops out near the head of the North Fork of Rock Creek (north of LL, T. 17 N., R. 79 W.) consists of arkosic quartzite interbedded with dark colored, conglomeratic quartzite. The dark colored conglomerate layers are more radioactive than the arkosic quartzite.

The most significant outcrop of the conglomerate member of the Magnolia Formation is in exposures along One-Mile Creek approximately 3 kilometers south of the village of Arlington in the northeastern

Medicine Bow Mountains (OM of Figure 3.9). Information on this area comes from outcrop study and the examination of cores made available by the Exxon Company. The conglomerate member of this area is a sericitic, coarse-grained quartzite, with pebble layers. Maximum thickness of the radioactive unit measured in subsurface is about 150 meters. The pebble layers range from beds with the thickness of a single pebble to compound zones up to 20 meters thick that include pyritic quartz-pebble conglomerate, fine-grained pyritic quartzite, and coarse-grained pyritic quartzite, in varying combinations. There are four compound zones ranging from 0.5 to 3 meters in thickness, and these are the most pyrite-rich and radioactive rocks in the section. In addition, there are more than eleven other pyritic quartz-pebble layers less than 0.5 meters thick scattered through the 130 meter section. The entire section has radioactivity above background for the area, but the most radioactive units are the conglomerate layers. These conglomerate layers are several orders of magnitude more radioactive in subsurface samples (at depths below 35 meters) than they are on the surface and they contain up to 20 percent pyrite which is almost entirely leached from surface outcrops.

Subsurface samples show that the most intense radioactivity is characteristically at the base of conglomerate suggesting that there is a concentration of uranium and thorium bearing minerals at the base of channels. The pyritic quartz-pebble conglomerate layers are trimodal and consist of a major clay-silt-sized fraction, sand-sized fraction, and pebble-sized fraction. The silt-sized fraction is largely sericite, the sand-size fraction is largely quartz with subordinate feldspar, and the rounded pebbles consist chiefly of vein quartz, quartzite, altered felsic volcanic rocks (?) and some altered feldspar of possible pegmatite origin.

The source of the clasts in the radioactive unit is not known. A few paleocurrent measurements have been made on crossbedding sets in quartzite above the conglomerate member and they indicate a source to the north or north-northeast. However, these measurements are not from rocks within the conglomerate member proper. Nevertheless, field observations of the size of clasts within the conglomerate member indicate that the coarsest fraction is found in the northernmost outcrop; so the available evidence suggests a north to north-northeast source.

The conglomerate member of the Magnolia Formation must underlie a large area of the Medicine Bow Mountains because outcrops of the conglomerate member have been found in the cores of anticlines and on the flanks of synclines from the Arrastre Lake area to One-Mile Creek, a distance of approximately 30 kilometers. The width of the outcrop belt of the conglomerate member varies, but probably averages about 8 kilometers. So far, the most promising uranium-thorium prospects have been found in the One-Mile Creek area, but very little of this 240 square kilometer area has been prospected by drilling. Thus, the conglomerate member of the Magnolia Formation remains one of the most significant prospects for uranium-bearing, quartz-pebble conglomerate in the United States.

The conglomerate member of the Magnolia Formation grades upward into a unit referred to as the quartzite member of the Magnolia Formation. The quartzite member is a coarse-grained, moderately-sorted quartzite consisting largely of well-rounded quartz granules that average 1-2 mm in diameter and less common pebbles that are up to one cm in diameter. The quartzite member contains less matrix than the underlying conglomerate (< 5%) and is lower in feldspar. The quartzite member has well-developed trough crossbeds with troughs approximately 5 to 15 cm deep. It is more consistent in thickness than the conglomerate member; averaging 425 meters.

In total, the Magnolia Formation is a fluvial succession that has a more proximal facies at its base and a more distal facies at its top. The conglomerate member must represent braided stream or river deposits not close enough to source to represent alluvial fan facies, but within the true braided stream depositional environment. The quartzite member is probably also a braided stream deposit, but the finer grain size of clasts, abundance of trough crossbed sets, that may have even finer grained material overlying such crossbed sets, suggests lower energy deposition than for the conglomerate member. A change in facies of this type may be a function of changing climate, changing supply of clastics in the source area, or it might represent a minor marine transgression.

The Magnolia Formation certainly represents a major change in character of sedimentation from rocks of the underlying Phantom Lake Metamorphic Suite. It lies unconformably on metavolcanic rocks of the Phantom Lake Metamorphic Suite in the Arrastre Lake area, on quartzite in the outcrop area along the Medicine Bow River, on paraconglomerate in the Deep Creek area, and on granite in the One Mile Creek area. We have not detected an angular unconformity at contacts between the Magnolia Formation and rocks of the Phantom Lake Metamorphic

Suite, but inasmuch as the rocks of the Phantom Lake Metamorphic Suite are tightly folded as compared with rocks of the overlying Deep Lake Group and Libby Creek Group and inasmuch as the Magnolia Formation lies on a great variety of Phantom Lake Metamorphic Suite rocks--we believe that there is a major unconformity at the base of the Magnolia Formation and that the rocks of the Phantom Lake Metamorphic Suite were deformed and perhaps metamorphosed prior to deposition of the Magnolia Formation. This interpretation indicates that the unconformity at the base of the Magnolia Formation is profound and therefore may be of tremendous significance in uranium exploration.

Lindsey Quartzite. The quartzite member of the Magnolia Formation is gradationally overlain by the Lindsey Quartzite (Karlstrom and Houston, 1979a, 1979b). West of Vagner Lake (VL of Figure 3.9), the Lindsey Quartzite is a light gray, medium-grained quartzite with layers of pebbles concentrated on foreset beds of cross-beds and in small scours. In the area between Deep Creek and the middle fork of Rock Creek of the Medicine Bow Mountains (near LL), the Lindsey Quartzite is composed of typical fining upward sets having pebble conglomerate at their base and trough crossbedded quartzite at their top. The pebble conglomerate is slightly radioactive (2-3X local background). The overall characteristics of the Lindsey Quartzite suggest a fluvial origin for the formation.

Campbell Lake Formation. The Lindsey Quartzite is overlain by the Campbell Lake Formation which consists of a paraconglomerate overlain by a quartz-rich phyllite. The paraconglomerate has an arkosic matrix containing rounded clasts of white granite, phyllite, quartzite, and metabasalt that are up to 76 cm in diameter. The conglomerate grades upward into a dark gray to black phyllite which becomes quartz-rich at the top. The Campbell Lake Formation is exposed over a wide area of the Medicine Bow Mountains but it is not continuous; where it is absent, this is thought to be non-depositional. The paraconglomerate of the Campbell Lake Formation may be glacial in origin, but no definitive glacial criteria such as dropstones have been found in the unit.

Cascade Quartzite. The Campbell Lake Formation is overlain by the Cascade Quartzite. The Cascade Quartzite has a distinctive lithology; it is marked by pebble layers ranging from a thickness of one pebble to channels 20 meters thick that contain pebbles of quartz, quartzite, black chert, and pink granite. The black chert is distinctive and aids in the identification of the Cascade. For the most part, the Cascade Quartzite is a more mature rock-type than most quartzites that underlie it. For example, it contains more resistant clasts and it is not feldspathic except at the very top of the unit where feldspathic quartzite is locally developed. No radioactive conglomerate has been identified in the Cascade quartzite, but like the underlying Lindsey Quartzite and Magnolia Formation it is considered fluvial (Karlstrom and Houston, 1979a, 1979b).

Vagner Formation. The Vagner Formation overlies the Cascade Quartzite unconformably. The Vagner Formation is a three-fold unit consisting of a basal paraconglomerate, a middle metalimestone, and and upper phyllite. The matrix of the paraconglomerate is arkosic and clasts in order of decreasing abundance are red granite, white granite, quartzite, phyllite and rare metavolcanic (?) rocks. This paraconglomerate and associate phyllite have features suggestive of a glacial origin. As first noted by Blackwelder (1926) and later by Sylvester (1973) these paraconglomerates are poorly sorted and have clasts which are more angular (some boulder-sized) than paraconglomerates described previously. Furthermore, associated phyllites have dropstones and this, along with other evidence (Sylvester, 1973), strongly suggests a glacial or possibly glacio-marine origin for the paraconglomerate and phyllite. The origin of the associated metalimestones is uncertain, but they may represent calcium carbonate brines deposited in response to the retreat of dry-based glaciers (Carey and Ahmad, 1961, p. 886).

Rock Knoll Formation. The Vagner Formation is in fault contact with the overlying Rock Knoll Formation. This unit is chiefly gray, fine- to medium-grained quartzite with thin layers of phyllite. Near the base of the Rock Knoll Formation planar cross-bedding in sets 0.6 to 1.5 m thick is common. This cross-bedded sequence is overlain by quartzite with phyllitic partings. The partings weather to olive patches on bedding surfaces and are distinctive and characteristic of this part of the Rock Knoll Formation. In addition, Blackwelder (1926, p. 627) and Houston and others (1968, p. 17-22) reported clay galls in the quartzites. The upper part of the Rock Knoll Formation is quartzite with abundant thin (6-10 cm) beds of conglomerate. Clasts are vein quartz, quartzite, and granite. As suggested by Blackwelder (1926, p. 626-627), this part of the Rock Knoll Formation may be fluvial.

Libby Creek Group. The Libby Creek Group has been described by Blackwelder (1926), Houston and others (1968), and Lanthier (1979). As defined by Houston and others (1968), it includes all rocks above the unit then referred to as the Deep Lake Formation and now referred to as the Deep Lake Group. As shown in Table 3.3, the formations of the Libby Creek Group include the Headquarters Formation (paraconglomerate and laminated phyllite of possible glacial origin), Heart Formation (largely quartzite with primary structures suggestive of shallow water origin), Medicine Peak Quartzite (thick, massive quartzite probably including fluvial deposits and high energy marine shoreline deposits), Lookout Schist (interlayered quartzite, phyllite and schist with structures characteristic of distal turbidites; beds of iron formation are present at the base of the Lookout Schist), Sugarloaf Quartzite (high silica orthoquartzite of probable marine origin), Nash Formation (stromatolitic dolomite and black graphitic phyllite, both of probable shallow marine origin), Towner Greenstone (mafic igneous rocks of uncertain origin), and French Slate (gray slate with rare thin hematitic quartzite layers, both of probable marine origin). These formations are described in some detail in the reports cited above and in papers by Knight (1968) on the Nash Formation and Wilson (1975) on the Lookout Schist.

It has been necessary to redefine the Headquarters Schist of the Libby Creek Group, because part of this unit as originally defined by Blackwelder (1926, p. 627-631) has been included in the Vagner Formation of the Deep Lake Group of this report (Table 3.2). The name Headquarters Formation replaces Headquarters Schist as three lithologic members are now recognized in this unit (Lanthier, 1978, 1979).

Several rocks within the Libby Creek Group have been dated by Hills and others (1968). The Rb/Sr whole rock ages determined for these metasedimentary

	Formation	Lithology	Primary features	Source	Depositional Environment	Thickness [m]
LIBBY CREEK GROUP	French Slate	black slate	thin qtzite layers	?	marine	600
	Towner Greenstone	chlorite schist	relict igneous textures	?	?	0-480
	Nash Formation	lenticular phyllite metadolomite	graphitic layers stromatolites	?	marine shallow marine	1900
	Sugarloaf Qtzite	massive qtzite	cross-beds, ripple marks	?	marine	0-580
	Lookout Schist	laminated schist with qtzite layers	sole markings clastic dikes cross-bedding	?	marine, turbidites	365
	Medicine Peak Qtzite	qtzite; congl. qtz-pebble congl qtzite qtzite	finning upward sequences massive tabular cross-beds	ENE	fluvial and shallow marine pro-deltaic? near shore marine	1585
	Heart Formation	qtzite phyllite qtzite	massive laminated massive, cross-beds	NE	shallow marine marine shallow marine; fluvial	650
	Headquarters Formation	phyllite arkosic qtzite paracongl	varves? cross-beds dropstones	NNE	glacial or glacio-marine	650

Unconformity

Table 3.3. Stratigraphy of the Libby Creek Group of the Medicine Bow Mountains, Wyoming.

rocks are probably dates of metamorphism, but they set an upper limit on the age of the metasedimentary succession. The most reliable metamorphic date is a Rb/Sr whole rock age for the Lookout Schist of 1710 ± 60 m.y. Inasmuch as the basal units of the Libby Creek Group lie on basement gneiss (southwestern area of Figure 3.9) dated as older than 2500 m.y. (Hills and others, 1968, p. 1762-1763) and the metamorphic date is approximately 1710 ± 60 m.y., these rocks must be younger than about 2500 m.y. and older than about 1700 m.y. Clearly we need a much closer age bracketing to evaluate the economic potential of these metasedimentary rocks.

In general, however, there are several indications that the quartzites of the Libby Creek Group are probably not a good target for uranium mineralization. First, they are interpreted to be mainly marine deposits which are less likely than fluvial deposits to contain uranium-bearing fossil-placers. Second, they contain magnetite and hematite rather than pyrite as their main iron phase. This suggests that they were deposited in oxygenated conditions which were unfavorable for transport of detrital uraninite (Roscoe, 1973). Third, geochemical assays to date have not revealed anomalously high uranium content in conglomerates of the Medicine Peak Quartzite. Fourth, lithostratigraphic correlation of the Medicine Bow metasediments with other Early Proterozoic rocks in North America (see Figure 3.10), especially the Huronian Supergroup, suggests that the Libby Creek Group may be too young to contain significant fossil placer uranium (Young, 1973).

Regional Stratigraphic Correlations

A remarkable aspect of the metasedimentary rocks of the Medicine Bow Mountains is their close lithologic resemblance to formations of the Huronian Supergroup of Canada. As shown in Figure 3.10, both areas contain volcanics, quartz-pebble conglomerates, drab quartzites, glacial (?) paraconglomerates,

SIERRA MADRE
Wyoming
AFTER GRAFF (1978)

		YOUNGER IGNEOUS AND METAMORPHIC ROCKS		
		SHEAR ZONE		
LIBBY CR. GP.		Slaughterhouse Fm.	Dolomite Phyllite	Carbonate bank
DEEP LAKE GROUP	CYCLE 4	Copperton Qtzite	Sandstone	Shallow marine
		Vagner Fm.	Sandstone Shale Limestone Paraconglomerate	Glacio-marine
	CYCLE 3	Cascade Qtzite	Arkose Sandstone	Fluvial
		Campbell Lake Fm.	Sandstone Shale Paraconglomerate	Glacial?
	CYCLE 2	Singer Peak Fm.	Sandstone Shale	Marine
		Magnolia Fm.	Sandstone Qtz-granule ss. Qtz-peb cgl.	Fluvial
PHANTOM LAKE SUITE	CYCLE 1	Silver Lake Conglomerate	Graywacke Boulder cgl.	Marine?
		Spring Lake Volcanics	Flows, tuffs	Subaerial-shallow marine
		Upper Jack Creek Fm.	Shale, sandstone Paraconglomerate Qtz-peb cgl.	Shallow marine Fluvial
		Lower Jack Creek Fm.	Sandstone Paraconglomerate Shale, flows, tuffs Graywacke	Subaerial, fluvial, and shallow marine

MEDICINE BOW MOUNTAINS
WYOMING
AFTER KARLSTROM AND HOUSTON (1979a, 1979b)

Medicine Peak Qtzite.	Sandstone Qtz-pebble cgl.	Shallow marine
Heart Fm.	Sandstone Siltstone Sandstone	Marine?
Headquarters Schist	Shale Arkose Paraconglomerate	Glacio-marine
Rock Knoll Fm.	Conglomerate Sandstone, shale	Shallow marine
Vagner Fm.	Shale Limestone Paraconglomerate	Glacio-marine
Cascade Qtzite	Pebbly arkose Pebbly sandstone	Fluvial
Campbell Lake Fm.	Shale Paraconglomerate	Glacial
Lindsey Qtzite	Sandstone	Fluvial
Magnolia Fm.	Qtz-granule ss. Qtz-pebble cgl.	Fluvial
Upper Phantom Lake Metamorphic Suite	Paraconglomerate basalt micaceous ss. argillite fine-gr. ss. quartz-pebble conglomerate	Subaerial Shallow marine Marine Fluvial
Lower Phantom Lake Metamorphic Suite	Volcanoclastic graywackes, flows and tuffs	Subaerial?

HURONIAN SUPERGROUP
ONTARIO, CANADA
AFTER ROSCOE, 1969

COBALT GROUP	Lorraine Fm.	Sandstone Qtz-pebble cgl.	Shallow marine
	Gowganda Fm.	Graywacke Siltstone Sandstone Paraconglomerate	Glacial, glacial-marine
QUIRKE LAKE GP.	Serpent Fm.	Sandstone	Fluvial shallow marine
	Espanola Fm.	Dolomite Limestone Sandstone	Shallow marine
	Bruce Fm.	Paraconglomerate	Glacio-marine
HOUGH LAKE GP.	Mississagi Fm.	Sandstone	Shallow marine
	Pecors Fm.	Siltstone	Shallow marine
	Ramsey Lake Fm.	Paraconglomerate	Glacial-shallow water
ELLIOT LAKE GP.	McKim Fm.	Siltstone	Shallow marine
	Matinenda Fm.	Sandstone Qtz-pebble cgl.	Fluvial
	Thessalon Fm.	Basalt	Subaerial
	Livingstone Creek Fm.	Subarkose	Marine

?

ARCHEAN METASEDIMENTS, GNEISSES, AND GRANITE

Figure 3.10. Lithostratigraphic correlation of metasedimentary rocks in the Sierra Madre, Medicine Bow Mountains, and Huronian Supergroup. Adapted from Graff (1978, p. 71).

siltstones, and thick aluminous quartzites. In addition, the distinctive lithologic units in both areas occur in the same stratigraphic order. Figure 3.10 shows that it is possible to correlate the Huronian Supergroup with rocks of the upper Phantom Lake Suite, Deep Lake Group and lower Libby Creek Group almost formation by formation. The Magnolia Formation corresponds to the ore-bearing Matinenda Formation and paraconglomerates of the Campbell Lake, Vagner, and Headquarters Formations correspond to the Ramsey Lake, Bruce, and Gowganda Formations respectively. By this correlation, most of the Phantom Lake Metamorphic Suite is interpreted to be older than the Huronian Supergroup and most of the Libby Creek Group is interpreted to be younger. Plate 1 shows that the upper Libby Creek Group may correlate with the Marquette Range Supergroup of the Lake Superior area.

This type of lithostratigraphic correlation between widely separated sequences can be strongly criticized because there are no fossils or precise radiometric age determinations to prove that the rocks are the same age. Nevertheless, we believe that lithostratigraphic correlation is a valid procedure, especially in Early Proterozoic metasedimentary successions, for several reasons. First, the appearance of radioactive conglomerates on earth reflects increased cratonic stability following the end of the Archean (about 2500 m.y. in North America) and these units persisted only about 300-500 m.y.--until the atmosphere became rich enough in oxygen to prevent transport of detrital uraninite. Second, if the paraconglomerate units are glacial in origin, they may represent continent-wide climatic episodes which could be correlated throughout Early Proterozoic rocks in North America. Third, the sequence tillite-aluminous quartzite may represent a widespread climatic change from subpolar to subtropical latitudes (Young, 1973) which can be used for correlation. Fourth, both the Medicine Bow rocks and the Huronian

Supergroup appear to be older than nearby Proterozoic banded iron formations (Plate 1) so they both presumably were deposited prior to the development of appreciable free oxygen in the atmosphere.

The significance of this correlation with respect to uranium exploration is that it suggests that the overall character of the Medicine Bow succession is close enough to uranium producing sequences of the Huronian Supergroup to be of economic interest simply on this basis. The significance of the correlation with respect to ideas on crustal and atmospheric evolution is discussed in a later section.

Mineral Deposits

History of Exploration. The lithologic resemblance of the metasedimentary rocks of the Medicine Bow Mountains to metasedimentary rocks on the North Shore of Lake Huron was recognized as early as 1926 by Blackwelder and was further emphasized by Houston and others (1968) and by Young (1970). Because of this similarity Houston and others (1968, p. 159) recommended that quartz-pebble conglomerate of the Deep Lake Formation (Deep Lake Group of this report) be examined for gold and other heavy elements. This recommendation was discussed further in a report on the Sierra Madre (Houston and others, 1975) and in 1975 a joint study of the Medicine Bow Mountains and Sierra Madre was undertaken by the Geology Department of the University of Wyoming; U.S. Geological Survey; and Wyoming Geological Survey to study both mountain areas and assess their potential for uranium. The U.S. Geological Survey project in the Medicine Bow Mountains was partially funded by a grant from the Department of Energy, then the Energy Research and Development Agency.

To the writers' knowledge the first discovery of radioactive conglomerate in the Medicine Bow Mountains was made by Stuart Roscoe of the Canadian Geological Survey who had done much of the original work on the Canadian, Blind River occurrences. Roscoe examined a quartz-pebble conglomerate locality near the Arrastre Lake area which was brought to his attention by one of the authors (Houston) and noted that it was radioactive. Additional localities were found as a result of the joint study during the 1975-78 field seasons, and the regional extent of the radioactive conglomerate at the base of the Magnolia Formation was finally determined during the 1978 field season by the writers. During this period, 1975-1978, geologists of the Exxon Company discovered the One Mile Creek locality by use of aerial radiometric surveys and surface geochemical surveys and subsequently most of the favorable areas in the Medicine Bows have been staked by mining companies of the United States and Canada. Geologic reports on uranium in the Medicine Bow Mountains include Miller and others (1977), Houston and others (1977), Houston and others (1979), and Karlstrom and Houston (1979a, 1979b).

Uranium Exploration Possibilities. It is clear from the above discussion that the basal conglomerate member of the Magnolia Formation of the Deep Lake Group is a prime target for exploration for uranium-bearing, quartz-pebble conglomerate. The exact age of this unit is unknown but it is bracketed between about 2700 m.y. and 1700 m.y. so it is probably Early Proterozoic. As indicated in the discussion of stratigraphy of the Deep Lake Group, uranium- and thorium-bearing pyritic conglomerate has been discovered in the conglomerate member of the Magnolia Formation at several localities in the Medicine Bow Mountains. This uranium-thorium-bearing, quartz-pebble conglomerate has many features in common with classic conglomerate occurrences in the Blind River-Elliot Lake area of Canada and in the Witwatersrand of South Africa. The conglomerate occurs in fluviatile sedimentary rocks, it is pyritic, and radioactive zones are thick enough (> 3 meters) to be of economic interest provided grade of the ore is high enough and tonnages are sufficient to be mineable.

Geochemistry and Mineralogy. The quartz-pebble conglomerate occurrences in the conglomerate member of the Magnolia Formation are highly altered on the surface. Pyrite is almost totally leached from surface outcrops leaving vuggy limonitic zones as evidence of its former presence. Some of these outcrops are strongly radioactive, despite leaching of pyrite and uranium minerals, because of the thorium content of the conglomerate. Conglomerate samples in outcrop contain up to 1000 ppm thorium and up to 150 ppm uranium and thorium/uranium ratios average about 4:1. Relative radioactivity of surface and subsurface (core) samples indicates that the uranium content of core samples taken below the zone of weathering is much higher than on the surface, perhaps an order of magnitude or more.

We have not been able to identify the primary uranium and thorium minerals in samples studied to date because most core samples come from the One Mile Creek locality of the Medicine Bow Mountains where the metamorphic rank is almandine amphibolite facies. Pyrite, for example, is completely recrystallized and there is no evidence of a rounded shape for the mineral except in some round clusters of euhedral crystals that may have been rounded grains originally.

Desborough and Sharp (1978) have identified 0.1 to 0.2 mm grains in pyritic conglomerate that have the chemical composition of coffenite and thorite. The "coffenite" contains 45-61 weight percent uranium and a few tenths to several weight percent of thorium and lead. Thorite contains 30-45 weight percent thorium, and a few tenths to several percent of uranium, lead, and yttrium. According to Desborough and Sharp these two uranium and thorium silicate minerals are anhedral, but are more like blebs than rounded grains that may have had a detrital history. These minerals are probably secondary.

In addition to pyrite, coffenite, and thorite, other minerals that appear to be secondary or recrystallized are zircon, a monazite containing up to 6 weight percent thorium and an ilmenorutile that Desborough and Sharp state is a recrystallized Nb-rich detrital mineral because it occurs as rounded spongy aggregates of euhedral to subhedral crystals supported interstitially by quartz. There is also carbon material in the conglomerate and Desborough and Sharp believe it is of organic origin because it contains sulphur. Other minerals of uncertain origin identified by Desborough and Sharp are Ni-Co-sulfides, marcasite, chalcopyrite, and columbite.

Heavy minerals that are clearly of detrital origin are sphene, ilmenite, spessertine, and zircon. There is also sparse magnetite that has the morphological characteristics of detrital grains.

Reserves. Extensive drilling programs will be required before ore reserves, if any, are outlined in the Medicine Bow Mountain area. The entire conglomerate member of the Magnolia Formation which underlies an area about 30 by 8 kilometers is not likely to be uniformly mineralized inasmuch as the radioactive conglomerate layers show wide variations in counts per minute from one locality to another. The uranium-thorium minerals were probably deposited in channels of braided streams or river systems and their extent depends on the width of these systems and on the proximity to source. Information on paleocurrent directions suggests that these braided river systems flowed south or south-southwest, but this has not been verified statistically.

Sierra Madre, Wyoming

Geologic Setting

The Sierra Madre of Wyoming is the northern extension of the Park Range of Colorado and is located about 10 km southwest of the Medicine Bow Mountains

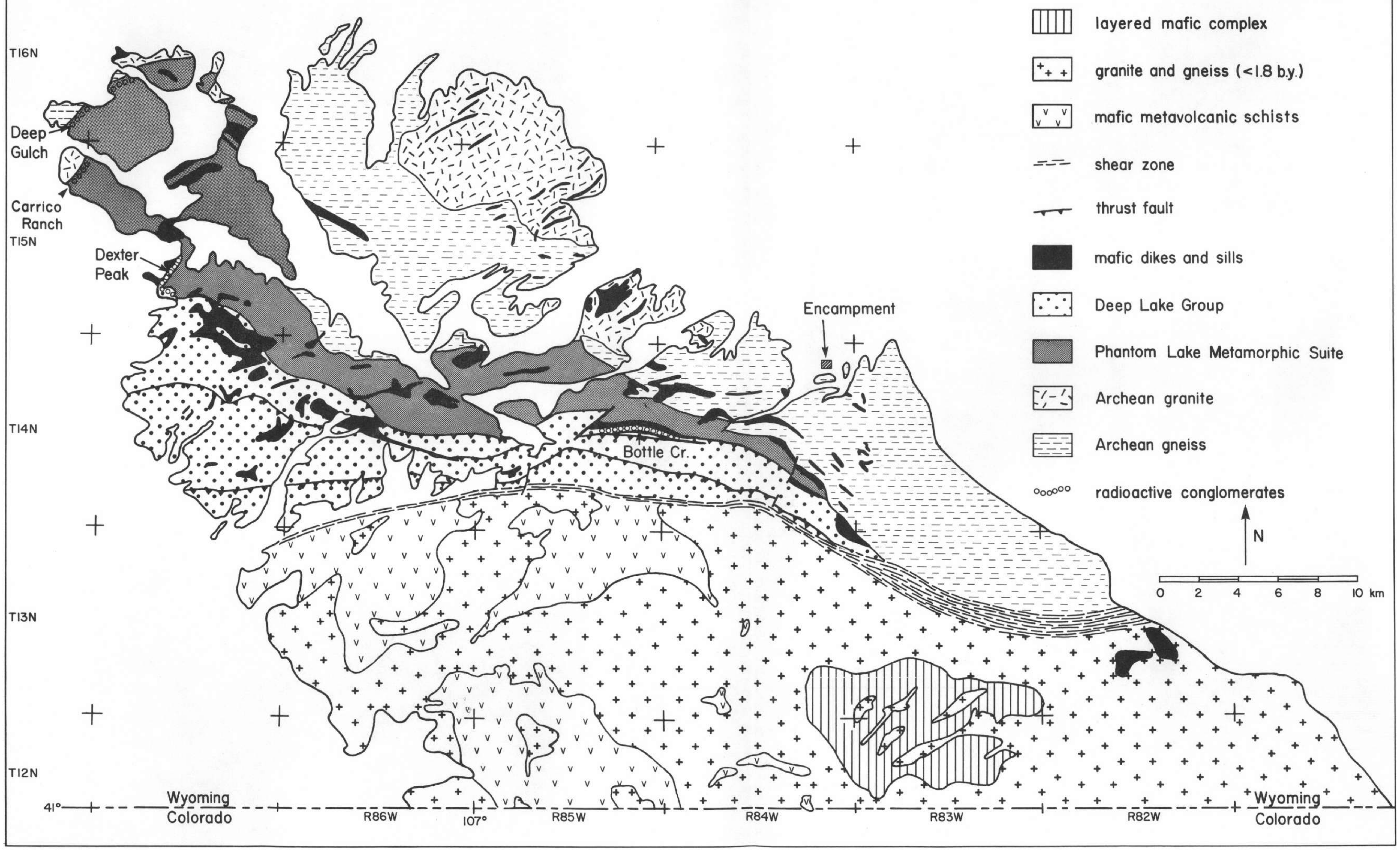
(Figure 3.3). The overall geology of the Sierra Madre is similar to that of the Medicine Bows in that the northern Sierra Madre consists of Archean granitic basement nonconformably overlain by a thick sequence of Early Proterozoic miogeosynclinal metasedimentary rocks and the southern Sierra Madre consists of a metavolcanic succession cut by mafic and felsic intrusive rocks (Houston and others, 1975). As shown in Figure 3.11, the northern and southern areas are separated by a shear zone which is thought to be the western extension of the Mullen Creek-Nash Fork shear zone of the Medicine Bow Mountains (Graff, 1978, 1979).

Distribution and Structure

Figure 3.11 shows that Early Proterozoic metasedimentary rocks of the Phantom Lake Metamorphic Suite and Deep Lake Group are preserved in a west-thickening wedge in the central Sierra Madre. The overall structure of the metasediments was interpreted by Spencer (1904) to be a west plunging, overturned synclinorium and by Houston and others (1975) to be a south dipping homoclinal succession. Graff (1978, 1979) reconciled these views by suggesting that the metasedimentary rocks form the north limb of a faulted synclinorium, the south limb having been removed by movements along the major shear zone.

Graff (1978) identified several generations of folds in the metasediments: a major east-west fold system, less well developed northeast- and northwest-trending folds, and a late episode of nearly vertical rotation. These four deformational events are similar in style and sequence to deformations identified by Karlstrom (1977) in the Medicine Bow Mountains. Graff (1978) also identified major east-trending faults in the southern part of the metasedimentary wedge, shown in Figure 3.11, which he interpreted to be major thrust faults. These faults have greatly attenuated the stratigraphic section of

Figure 3.11. Generalized geology of the Sierra Madre, Wyoming, showing locations of radioactive conglomerates.



the Phantom Lake Suite and Deep Lake Group in the eastern Sierra Madre.

Figure 3.11 shows that rocks of the Phantom Lake Metamorphic Suite crop out in the northern half of the metasedimentary wedge and rocks of the Deep Lake Group in the southern half. Rocks equivalent to the Libby Creek Group of the Medicine Bow Mountains, if they were originally present, have been removed by faulting along the shear zone.

The age of the metasedimentary rocks in the Sierra Madre is bracketed between about 2500 m.y., the age of the granitic basement (Divis, 1976; Hills and Houston, 1979) and 1700 m.y., the age of metamorphism of the sediments and movement along the shear zone (Hills and Houston, 1979).

Stratigraphy and Paleogeography

Phantom Lake Metamorphic Suite. The Phantom Lake Metamorphic Suite in the northern and northwestern Sierra Madre contains metavolcanic and metasedimentary rocks that are very similar lithologically to the rocks of the Phantom Lake Suite in the Medicine Bow Mountains. The stratigraphy of these rocks is not completely understood, but rock types include fine-grained feldspathic quartzite, paraconglomerate, metabasalt, pillow lava, metatuffs, meta-agglomerate, phyllite, quartz-pebble conglomerate and marble. Locally these rocks are interlayered with quartzo-feldspathic gneiss, hornblende schist and amphibolite. Graff (1978, 1979) has divided the Phantom Lake Suite into three formations, shown in Table 3.4. These are the Jack Creek Formation, Spring Lake Volcanics, and Silver Lake Conglomerate. The total thickness of the Phantom Lake Suite in the Sierra Madre is more than three kilometers.

Graff's (1978) interpretation of the depositional environments of the Phantom Lake Suite are shown in Table 3.4. In general, the sequence is believed to be an epicontinental succession which contains mainly fluvial and subaerial deposits.

SNOWY RANGE SUPERGROUP	Libby Cr. Gr.?	Slaughter-house Fm.	Thick calc-silicate and calcareous phyllites. Buff to tan, arenaceous, silica-seamed metadolomite.
	Deep Lake Group	Copperton Quartzite	White, massive to mylonitized completely recrystallized quartzite.
		Vagner Fm.	Interbedded, fine-grained, clean quartzite and green phyllite. Tillite (poorly sorted, angular-clast paraconglomerate) pink granite and rare quartzite clasts.
		Cascade Quartzite	Thick, massive, clean quartzite with occasional thin lenses of quartz-pebble conglomerate. Quartzites contain channels, cross-beds, and graded beds. Rare beds of sandy phyllites or green and blue phyllites.
		Campbell Lake Fm.	Paraconglomerates with angular, red granite clasts. Thin, clean, fine-grained quartzites. Thin green and blue phyllites.
		Singer Peak Fm.	Thick green and silver phyllites. Thin phyllitic quartzite with rare pebbly layers.
		Magnolia Fm.	Coarse to fine-grained quartzite (fining upward), generally phyllitic. Very coarse-grained quartz-granule conglomerates, slightly radioactive. Feldspathic quartzites with cross-beds and channels. Uraniferous quartz-pebble conglomerate and feldspathic, micaceous quartzite.
	Phantom Lake Metamorphic Suite	Silver Lake Conglomerate	Metagraywacke (cleaning upward). Boulder paraconglomerates with arkosic or amphibolitic matrices, clasts of granite, quartzite, and phyllite.
		Spring Lake Volcanics	Amphibole schists. Garnet-amphibole and micaceous schists. Pillow basalts, amygdaloidal basalts. Metatuffs, flows and peridotites.
		Jack Creek Fm.	Phyllitic quartzites and radioactive, silver (pebbly?) phyllites. Stromatolitic metalimestones, metadolomite, calcarenite. Fine-grained quartzite, feldspathic, very coarse-grained quartzite. Uraniferous quartz-pebble conblomerates, phyllitic fine-grained feldspathic quartzite.
ARCHEAN GNEISSES AND METASEDIMENTS			

Table 3.4. Stratigraphy of Early Proterozoic metasediments of the Sierra Madre, Wyoming. Taken from Graff (1978, p. 18).

The Deep Lake Group. Metasedimentary rocks believed to be equivalent to rocks of the Deep Lake Group of the Medicine Bow Mountains crop out in the southern part of the wedge of metasedimentary rocks in the Sierra Madre (Figure 3.11). The basal beds of the Deep Lake Group are like the Magnolia Formation of the Medicine Bow Mountains, but formations that overlie the Magnolia Formation are not always lithologically identical to those in the Deep Lake Group of the Medicine Bow Mountains. Figure 3.10 shows a tentative correlation of the metasedimentary rocks of the Sierra Madre and Medicine Bow Mountains. This correlation is based upon distinctive lithologic units such as radioactive conglomerates, boulder paraconglomerates, and glacial (?) paraconglomerates; and on similar cyclic stratigraphic sequences, shown in Figure 3.10 (Graff, 1979). In general, metasedimentary rocks of the Sierra Madre are finer grained than those of the Medicine Bow Mountains and apparently were deposited a greater distance from the source.

Mineral Deposits

Radioactive Conglomerate Units. There are two major radioactive quartz-pebble conglomerate units in the Sierra Madre. In the northwestern Sierra Madre radioactive quartz-pebble conglomerate has been recognized in the Phantom Lake Metamorphic Suite (Graff and Houston, 1977). This radioactive quartz-pebble conglomerate crops out in the Deep Gulch area where it can be traced for approximately 5 kilometers, and it also crops out to the northeast in sec. 6 and 7, T. 15 N., R. 87 W. where it can be traced for about 3 kilometers (Figure 3.11). The quartz-pebble conglomerate is present as beds within a radioactive coarse-grained sericitic quartzite, and is very similar to the conglomerate in the Magnolia Formation of the Medicine Bow Mountains. Conglomerate layers are parts of fining upward sets in the quartzite unit and they range from beds the thickness of a single pebble to compound beds a meter or more thick. The coarse-grained quartzite with layers of conglomerate grades upward into finer-grained quartzite with trough cross-beds and this quartzite grades into phyllitic quartzites with layers of marble. The succession is believed to represent braided stream or river deposits that grade upward into marine quartzites.

Radioactive quartz-pebble conglomerates have also been recognized in the Magnolia Formation of the Sierra Madre. Major occurrences are in the Dexter Peak area (secs. 21, 28, and 34, T. 15 N., R. 87 W.) where radioactive quartzite with conglomerate layers has been traced about 5 kilometers (Figure 3.11). The radioactive conglomerate of the Dexter Peak area is like that described in other occurrences of the Sierra Madre and Medicine Bow Mountains and is also interpreted as braided stream deposits.

In the eastern Sierra Madre radioactive, quartz-pebble conglomerate has been mapped for a distance of about 5 kilometers along strike near Bottle Creek, from sec. 14, T. 14 N., R. 85 W. east to sec. 18, T. 14 N., R. 84 W. (Figure 3.11). This conglomerate is cleaner, better sorted, and not as radioactive as that in the western Sierra Madre.

Uranium Exploration Possibilities. The radioactive, quartz-pebble conglomerates of the Sierra Madre are similar in most respects to those of the Medicine Bow Mountains and the radioactive conglomerate in the basal Magnolia Formation is considered to be the same in both areas. The Sierra Madre has additional radioactive conglomerate beds in the Phantom Lake Metamorphic Suite.

There has been no drilling in the Sierra Madre so very little is known about the uranium content of the conglomerate. In the western Sierra Madre, however, radioactive beds have about the same above background count as beds of the One Mile Creek area of the Medicine Bow Mountains, and surface samples contain up to 128 ppm uranium and 718 ppm thorium.

In mapping done to date, strongly radioactive beds can be shown to crop out over a greater area in the Sierra Madre than in the Medicine Bow Mountains, if we consider that the strongly radioactive beds of the Medicine Bow Mountains are mainly in the One Mile Creek area. Certainly the Sierra Madre seems to show equal promise as a uranium prospect, but extensive drilling will have to be done before an economic appraisal can be made.

Hartville Uplift, Wyoming

The Hartville uplift is a north striking uplift of Precambrian rocks located in eastern Wyoming between the towns of Guernsey and Lusk (Figure 3.3). Metasedimentary and metavolcanic rocks crop out on the western margin of the uplift, but their relationship to Archean rocks of the Wyoming Province is not known. These metasedimentary and metavolcanic rocks are more eugeosynclinal in aspect than those of either the Sierra Madre or Medicine Bow Mountains. Metasedimentary rocks include quartzite, iron formation, metadolomite, slate, phyllite and graywacke. Metavolcanic rocks are primarily basalt--some showing pillow structures. The stratigraphy of metasediments in the Hartville uplift is shown in Plate 1.

This succession of rocks was first defined by Smith and Darton (1903) and was subsequently studied by Ebbett (1956) and Millgate (1965). The entire uplift is currently being studied by George L. Snyder of the U.S.G.S. The writers had the opportunity to review this area with Snyder in August 1978, and it is clear that this succession of metasedimentary and metavolcanic rocks is not similar to that of the Sierra Madre or Medicine Bow Mountains; in fact, it more closely resembles rocks of the eugeosynclinal or younger parts of the Marquette Range Supergroup of the Lake Superior Region (Plate 1).

We believe that the metasedimentary rocks of the Hartville uplift are not good prospects for quartz-pebble conglomerate-type uranium deposits because they are probably wholly marine and are perhaps younger than 2000 m.y.

Black Hills Uplift, South Dakota

The Black Hills uplift of South Dakota is a broad anticlinal uplift located in the southwestern part of the state (Figure 3.3). The core of the uplift consists of Precambrian metasedimentary and metavolcanic rocks that are highly deformed and metamorphosed (Noble and others, 1949). The metasedimentary rocks are cut by amphibolite, foliated granite, and pegmatite. The major granitic bodies are in the south and are referred to as the Harney Peak Granite.

The metasedimentary rocks of the Black Hills have been studied in detail only in certain areas and the relationship between areas is not fully understood. A modern compilation of the Precambrian geology of the entire uplift by Kleinkopf and Redden (1975) and a review of the general geology by Bayley and James (1973) serve as the basis for this discussion.

Figure 3.12 shows that there are two areas in the Black Hills were Archean basement is exposed: the Nemo district of the northeastern Black Hills and the Bear Mountain area in the southwestern Black Hills. In the Nemo district, contacts between units are poorly exposed or covered by Phanerozoic sedimentary rocks, but the Archean Little Elk Granite is believed to be overlain unconformably by quartzite which is referred to as the Boxelder Creek Quartzite by Bayley (1972). This quartzite has interbedded quartz-mica schist and also contains layers of quartz-granule conglomerate. It is overlain conformably by layered oxide-facies iron formation referred to as the Benchmark Iron-Formation (Plate I). The Benchmark Iron-Formation and Boxelder Creek Quartzite are overlain unconformably by the Estes Conglomerate of Bayley (1972).

The Estes Conglomerate is a complex unit which Bayley considered a boulder conglomerate that grades into finer-grained facies including quartzite, quartz-mica schist, meta-arkose, and pebbly chlorite schist. All facies have pebbly layers. The boulder conglomerate is an open framework conglomerate with a chloritic matrix; Bayley's descriptions suggest that this boulder conglomerate is a proximal or perhaps an alluvial fan facies that grades laterally into a braided stream facies. Bayley (1970, p. 197) notes that the Estes Conglomerate contains abundant fragments of underlying rocks such as the Benchmark Iron-Formation and states that "each of the several facies rest unconformably on the Nemo Group (Boxelder Creek Quartzite and Benchmark Iron-Formation) at different places--each facies is basal in its own area--which probably indicates that the pre-Estes topography had considerable relief."

The Estes Conglomerate therefore fits the quartz-pebble conglomerate model quite well inasmuch as it is a fluviatile succession lying unconformably on Archean rocks. The age of the Estes Conglomerate is not precisely known; it lies on basement dated as about 2500 m.y. (Zartman and Stern, 1967), and although no upper age limit can be set, Black Hills metasedimentary rocks younger than the Estes Conglomerate are cut by the Harney Peak Granite dated as about 1600 to 1740 m.y. (Wetherill and others, 1956; Giletti and Gast, 1961; Riley, 1970).

F. A. Hills (1977, 1979) of the United States Geological Survey examined the Estes Conglomerate because he believed it might fit the quartz-pebble conglomerate model, and thus might contain detrital uranium minerals or other heavy metals. Hills' study resulted in the discovery of radioactive quartz-pebble conglomerate in the Nemo District. Hills (1977, 1979) found that the Estes Conglomerate contains beds of quartzite and quartz-pebble conglomerate with matrices of micaceous quartzite that had up to 25 percent pyrite. The conglomerate layers contain uranium (10 to 40 ppm) and thorium (20 to 800 ppm), and inasmuch as the Black Hills area is an important gold producing province, Hills considered that the Estes Conglomerate should have potential for gold (Hills, 1979).

After the announcement of the discovery of uranium-bearing, quartz-pebble conglomerate in the Black Hills, the area has been more thoroughly prospected by private companies and drilling has begun to determine if ore-grade uranium is present in the Estes Conglomerate. However, the results of these recent efforts are not available to us.

The Bear Mountain area of the southwestern Black Hills is a dome of Archean granite dated as about 2500 m.y. (Ratte and Zartman, 1970) overlain unconformably by metasedimentary and metavolcanic rocks (about 1000 meters thick) that are the dome's mantle (Ratte and Zartman, 1970). These mantle rocks are in turn separated by an unconformity from an even thicker overlying metasedimentary-metavolcanic succession. According to Kleinkopf and Redden (1975, p. 3) the mantling rocks include light-colored, pure quartzites, pebble conglomerates, and arkose. These rocks are correlated with the Estes Conglomerate by Kleinkopf and Redden (1975). They appear to be a good target for uranium exploration.

The uranium potential of other metasedimentary rocks of the Black Hills does not appear good. The Black Hills area has not been mapped in the detail necessary to fully appraise the uranium potential, but the Black Hills Precambrian can be viewed as a north-northwest striking synclinorium of metasedimentary rocks having two uplifts or domes of Archean basement on either side (Figure 3.12). The older metasedimentary successions that have uranium potential are apparently confined to the vicinity of the basement domes and

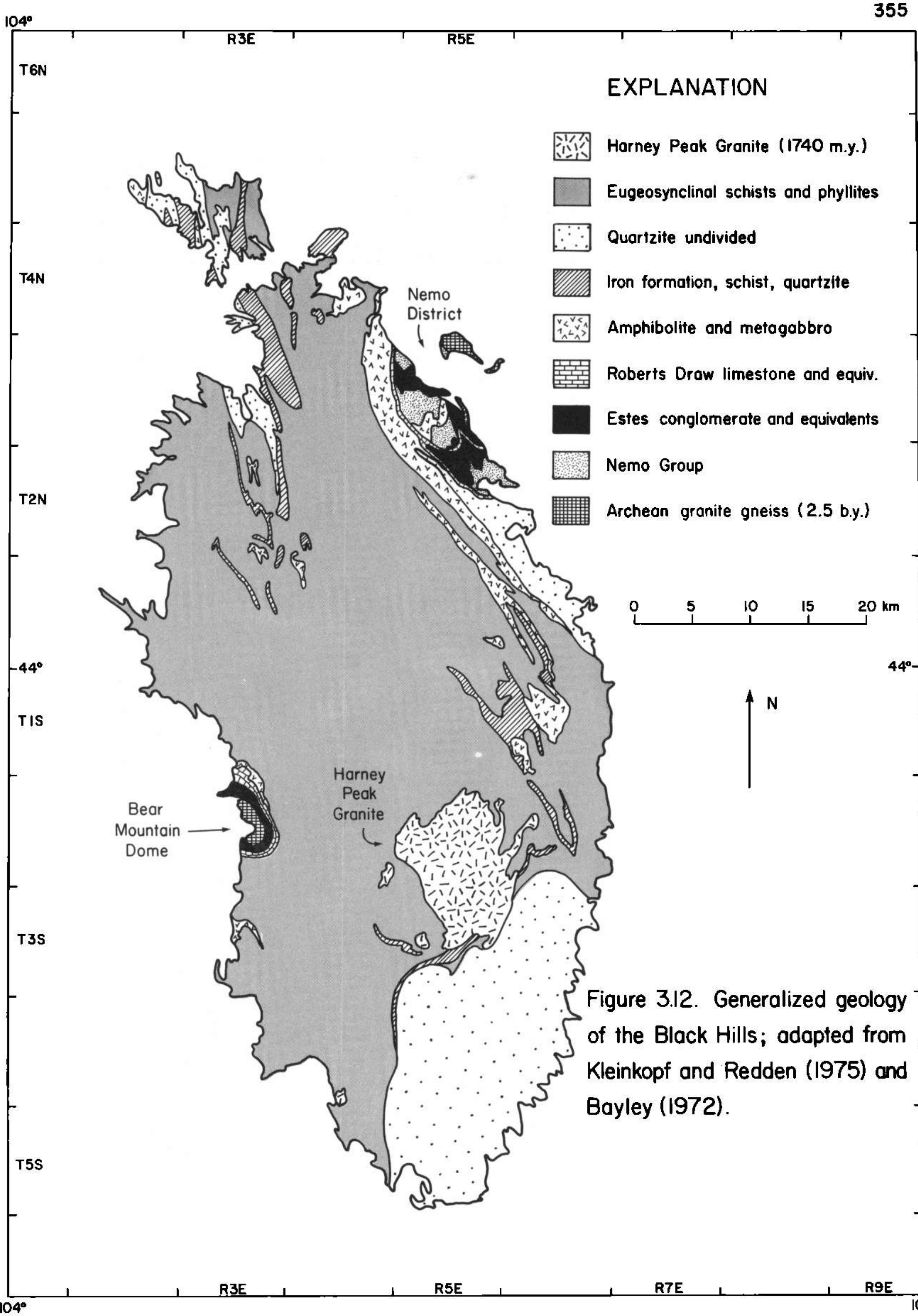


Figure 3.12. Generalized geology of the Black Hills; adapted from Kleinkopf and Redden (1975) and Bayley (1972).

younger metasedimentary rocks are largely marine and eugeosynclinal. These younger metasedimentary rocks contain iron formation and as Bayley and James (1973, p. 948) have suggested they are somewhat like the younger eugeosynclinal phase of the Marquette Range Supergroup. If so, they are too young and are the wrong geological environment to contain uranium-bearing quartz-pebble conglomerates.

SUMMARY AND CORRELATION OF EARLY PROTEROZOIC-TYPE METASEDIMENTARY SUCCESSIONS.

Perhaps the best way to summarize the potential for Early Proterozoic uranium-bearing conglomerates in the United States is via a regional litho-stratigraphic correlation of Early Proterozoic successions, such as is shown in Plate 1. Part I of this report, the exploration model, has emphasized that uranium-bearing conglomerates are most likely to be found in thick quartz-rich miogeosynclinal sedimentary successions that are older than about 2000 m.y. Within such successions, the richest conglomerates are likely to be fluvial, braided-river deposits which are either in the basal, clastic part of the succession (Huronian-type deposits) or within the coarse clastics of cyclic (transgressive-regressive) sedimentary sequences (Witwatersrand-type deposits). Clearly, age and stratigraphic-sedimentologic information are the most important factors in evaluating the uranium potential of a sequence, and this information can be portrayed in a lithostratigraphic correlation chart.

However, it is necessary to evaluate the justifications for lithostratigraphic correlation in Early Proterozoic successions. Without the benefit of guide fossils and with only a scattering of radiometric age determinations, what is the basis for correlating units in widely separated Proterozoic sequences?

The key to this question is that the Early Proterozoic (2500-1700 m.y.) appears to have been a time of evolutionary changes in tectonic conditions, atmospheric and hydrospheric composition, and perhaps climatic conditions, and these factors all contributed to recognizable patterns in sedimentation. For example, Late Archean metasedimentary successions generally contain an appreciable percentage of volcanic rocks, reflecting the change from permobile tectonics in the Archean to cratonic stabilization in the Early Proterozoic. Somewhat younger, Early Proterozoic, successions tend to be richer in quartz and often contain fluvial, uranium- and gold-bearing conglomerates. Such conglomerates appear to be restricted in geologic time between 2500 and 2000 m.y. in North America, reflecting both tectonic and atmospheric evolution. Still younger rocks include widespread glacial (?) paraconglomerates and highly aluminous quartzites, which may reflect regional climatic changes; stromatolitic dolomites, which reflect biologic and atmospheric evolution; and Proterozoic banded iron formations, which reflect the early stages of a change toward an oxygenated atmosphere.

No one of these distinctive lithologies should be used for correlation by itself because of the diachronous nature of evolutionary processes. However, in North America, several of these lithologies occur in the <u>same stratigraphic order</u> within widely separated Early Proterozoic successions and this suggests to us that sedimentation in North America during the Early Proterozoic was strongly influenced by regional changes in tectonic, climatic, and atmospheric conditions. These regional changes, we believe, are justification for the lithostratigraphic correlation and idealized Early Proterozoic stratigraphy shown in Plate 1.

We have divided the Early Proterozoic rocks in North America into four broad depositional sequences, shown in Plate I. (1) The Volcano-sedimentary sequence includes Late Archean and Earliest Proterozoic metasediments and is characterized by interbedded volcanic rocks and clastic rocks; including volcanic flows, pyroclastics, thin iron-formation, and sometimes thin beds of radioactive quartz-pebble conglomerate. Examples include the Phantom Lake Metamorphic Suite of the Medicine Bows and Sierra Madre; the Nemo Group of the Black Hills; and possibly the Red Creek Quartzite and Little Willow

Series of northern Utah, the Dickenson Group of the Lake Superior region, and the Thessalon and associated rocks of the Huronian Supergroup. These sequences were deposited during the transition between permobile Archean tectonic conditions and Early Proterozoic stable craton conditions--probably between 2700 and 2300 m.y. ago.

(2) The radioactive conglomerate-tillite sequence is characterized by thicker radioactive conglomerate beds, several glacial (?) paraconglomerates (tillites), thick clastic successions of quartzites and shales, and thin limestones. Another characteristic of this sequence is that the rocks appear to occur within cyclic stratification sequences which are interpreted to reflect climatic changes associated with glacial advance and retreat. As shown in Plate 1, examples of this type of sequence are the Deep Lake Group of the Medicine Bows and Sierra Madre, the Estes Conglomerate of the Black Hills, and the Huronian Supergroup of Ontario. These sequences were deposited on newly stabilized cratonic blocks in an atmosphere which was low enough in oxygen to permit detrital transport of uraninite and pyrite. This was probably between 2500 and 2000 m.y. ago.

(3) The stromatolitic dolomite, quartzite, argillite sequence is characterized by stromatolitic dolomites and by the presence of non-cyclic miogeosynclinal sediments which tend to be thicker and laterally more extensive than units of the underlying sequences. The appearance of thick sections of carbonates containing stromatolites in the miogeosynclinal sequence, often above aluminous quartzite, is probably related both to biologic and atmospheric evolution and is, thus, a valid unit for lithostratigraphic correlation. These dolomites occur in the Medicines Bows, Hartville uplift, and throughout the Lake Superior region of the United States (Plate 1). This sequence was probably deposited between 2300 and 2000 m.y. ago.

(4) The iron formation-volcaniclastic sequence contains mainly eugeosynclinal clastic sediments plus major Proterozoic banded iron formations. The appearance of the banded iron formations is usually interpreted as an indication of increasing free oxygen in the atmosphere--which is certainly a regional event. Examples of this sequence are in the Hartville Uplift, Black Hills and the Lake Superior area of the United States (Plate 1). This sequence was probably deposited between 2100 and 1800 m.y. ago.

From the uranium explorationists' viewpoint, the lower two sequences are the ones of interest. Thus, if the correlation shown in Plate 1 is correct, the Medicine Bows, Sierra Madre and Black Hills are of greatest economic interest because they contain rocks of the radioactive conglomerate-tillite sequence as well as the volcano-sedimentary sequence. However, the Cherry Creek Group of Montana, the Red Creek Quartzite and Little Willow Series of northern Utah and the Dickenson Group are also of interest and may contain thin conglomerate units interbedded with volcanic rocks.

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PART IV. BIBLIOGRAPHY OF PRECAMBRIAN URANIUM-BEARING FOSSIL PLACERS.

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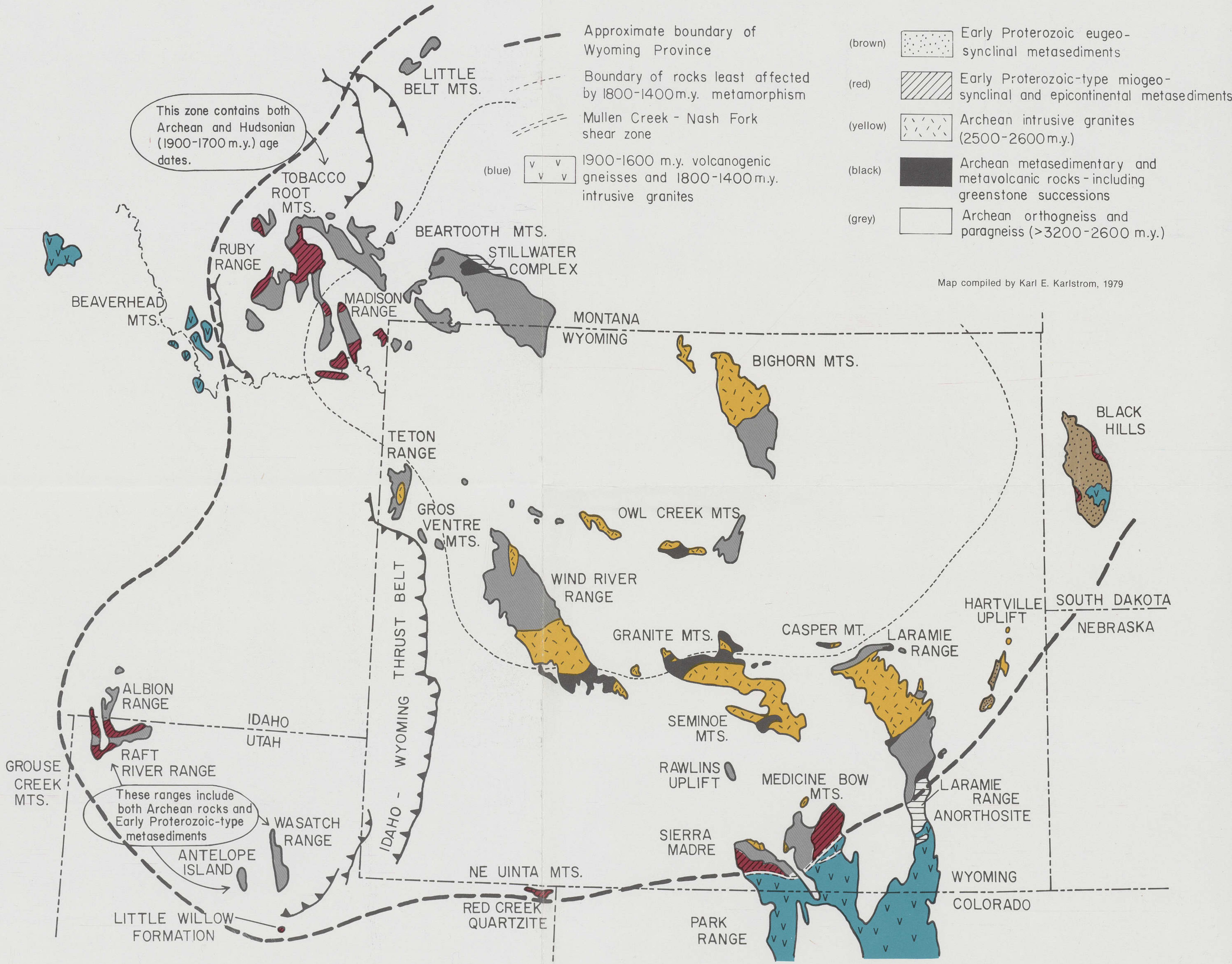
PLATE 1. Regional lithostratigraphic correlation and idealized stratigraphy chart of Early Proterozoic rocks along the south margin of the Archean nucleus of North America (see Index map).

(Fe) = iron formation (S) = stromatolitic dolomite (A) = aluminous quartzite = thrust fault

(T) = tillite and paraconglomerate (R) = radioactive quartz-pebble conglomerate (V) = volcanic rocks



Approximate boundary of Wyoming Province
Boundary of rocks least affected by 1800-1400 m.y. metamorphism
Mullen Creek - Nash Fork shear zone
(blue) 1900-1600 m.y. volcanogenic gneisses and 1800-1400 m.y. intrusive granites
(brown) Early Proterozoic eugeo-synclinal metasediments
(red) Early Proterozoic-type miogeo-synclinal and epicontinental metasediments
(yellow) Archean intrusive granites (2500-2600 m.y.)
(black) Archean metasedimentary and metavolcanic rocks - including greenstone successions
(grey) Archean orthogneiss and paragneiss (>3200-2600 m.y.)
Map compiled by Karl E. Karlstrom, 1979
This zone contains both Archean and Hudsonian (1900-1700 m.y.) age dates.
LITTLE BELT MTS.
TOBACCO ROOT MTS.
RUBY RANGE
BEAVERHEAD MTS.
MADISON RANGE
BEARTOOTH MTS.
STILLWATER COMPLEX
MONTANA
WYOMING
BIGHORN MTS.
BLACK HILLS
TETON RANGE
GROS VENTRE MTS.
OWL CREEK MTS.
WIND RIVER RANGE
IDAHO - WYOMING THRUST BELT
HARTVILLE UPLIFT
SOUTH DAKOTA
NEBRASKA
GRANITE MTS.
CASPER MT.
LARAMIE RANGE
ALBION RANGE
IDAHO
UTAH
RAFT RIVER RANGE
GROUSE CREEK MTS.
These ranges include both Archean rocks and Early Proterozoic-type metasediments
SEMINOE MTS.
RAWLINS UPLIFT
MEDICINE BOW MTS.
LARAMIE RANGE ANORTHOSITE
WASATCH RANGE
ANTELOPE ISLAND
SIERRA MADRE
NE UINTA MTS.
WYOMING
COLORADO
LITTLE WILLOW FORMATION
RED CREEK QUARTZITE
PARK RANGE