National Uranium Resource Evaluation

CORDILLERAN METAMORPHIC CORE COMPLEXES
AND THEIR URANIUM FAVORABILITY

FINAL REPORT

Peter J. Coney and Stephen J. Reynolds

with contributions by
George H. Davis, Stanley B. Keith, Paula F. Trever, Steven H. Lingrey,
Charles F. Kluth, Diane C. Ferris, James F. DuBois and James J. Hardy

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Tucson, Arizona 85721

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Tucson, Arizona 85721

November 1980

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

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PREPARED FOR THE U.S. DEPARTMENT OF ENERGY
ASSISTANT SECRETARY FOR RESOURCE APPLICATIONS
GRAND JUNCTION OFFICE, COLORADO
UNDER CONTRACT NO. DE-AC13-76GJ01664
AND BENDIX FIELD ENGINEERING CORPORATION
SUBCONTRACT NO. 79-357
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SUMMARY

By

Peter J. Coney and Stephen J. Reynolds

Cordilleran metamorphic core complexes are unique centers of plutonism, metamorphism, and deformation that are distributed in a discontinuous zone through the interior of the western United States. They are characterized by a distinctive assemblage of rocks and structures which occur in broad arch-like or domal features. They exhibit a central crystalline core of plutonic and high-grade metamorphic rocks. In high structural levels, these rocks are overprinted by a gently inclined mylonitic foliation containing a distinctive lineation. This lineation is typically consistent in trend over an entire mountain range or region. Near the margins of the crystalline core, the mylonitic rocks have been converted into a chloritic breccia via jointing, brecciation, faulting and hydrothermal alteration or retrograde metamorphism. The chloritic breccia is accompanied and overlain by a curvi-planar dislocation surface or décollement. Above the dislocation surface are an assortment of tilted and rotated rocks which generally lack any mylonitic fabric or metamorphic texture. These upper-plate rocks range in age from Precambrian to middle Tertiary and are cut by numerous low-angle structures which are inferred to be listric-normal faults that merge with or terminate against the underlying dislocation surface. A variation on this general theme occurs in several complexes where a zone of highly tectonized metasedimentary rocks (commonly marble) occupies the footwall of the dislocation surface.

Available geochronologic studies indicate a prolonged geological history for plutonic and metamorphic rocks of the crystalline core. However, it can generally be documented that final cooling of the crystalline core and movement on the dislocation surface are Tertiary! A Tertiary age for mylonitization is demonstrated for some complexes and can be inferred for many others. The genetic relationship between plutonism, metamorphism, mylonitization, and dislocation is currently controversial.

The uranium favorability of Cordilleran metamorphic core complexes is a function of processes that are either intrinsic or extrinsic to evolution of the complexes. Intrinsic processes include plutonism, metamorphism, mylonitization and formation of the dislocation surface (brecciation, hydrothermal alteration, tilting, etc.). Extrinsic processes such as weathering and sedimentation might have operated on rocks of the complexes before, during, or after the main phases of development of the complexes. Under favorable circumstances, both intrinsic and extrinsic processes are able to concentrate uranium into economically viable deposits.
The uranium favorability of Cordilleran metamorphic core complexes, as a group, is low. However, the favorability of individual complexes is as variable as their internal geology and regional tectonic setting. The Kettle, Selkirk and Albion complexes have the highest uranium favorability, while the remainder of the complexes have very low to moderate favorability.

Cordilleran metamorphic core complexes are in general most favorable for pegmatitic, metamorphic, hydrothermal, authigenic, allogenic, and peripheral lacustrine uranium occurrences. The complexes may be significant sources of uranium for later redistribution and concentration. Of particular importance in this regard are dislocation zones on the flanks of the complexes which may have been permeable channels or depositional sites for ascending (hydrothermal) or descending (meteoric) uraniferous fluids. The uranium potential of such zones is unknown and will only be revealed by exploratory drilling down-dip from uraniferous core rocks. Additional detailed study is needed to further document the uranium favorability of individual metamorphic core complexes.
The Principal Investigators are very grateful for the cooperation and efforts of many individuals who participated in the completion of this report. George Davis and Paula Trever, Department of Geosciences, University of Arizona and Stanley Keith, Arizona Bureau of Geology and Mineral Technology made significant contributions to the main scientific thrust of the project. Steven H. Lingrey and Charles F. Kluth, Department of Geosciences, University of Arizona did much of the original map compilations which appear on the 1:1,000,000 and 1:250,000 tectonic maps. Dianne C. Ferris, Department of Geosciences, University of Arizona compiled the bibliography and prepared the uranium occurrence lists and was always ready to provide any assistance needed. James F. Dubois was responsible for the thin-section petrography. Victoria V. Schale, Karl S. Tsuji, Rex A. Knepp, and Nancy Riggs did the drafting for the report. Sandra Hallenbeck, Administrative Assistant, Department of Geosciences, University of Arizona organized the logistics for final preparation of this report. Pat Bougie patiently coordinated and performed the typing. Richard L. Armstrong of the University of British Columbia, Greg A. Davis of the University of Southern California, Ed DeWitt of the Pennsylvania State University, and Wesley Peirce and Robert Scarborough both of the Arizona Bureau of Geology and Mineral Technology were most cooperative in providing unpublished data and helpful discussion which improved significantly the final product of this project.

We are all deeply grateful to John Burger of Bendix Field Engineering Corporation who was always most helpful and supportive and very patient throughout the entire project. We deeply appreciate the opportunity to carry out this study which because of its timeliness has considerably improved our understanding of Cordilleran metamorphic core complexes.
INTRODUCTION

by

Peter J. Coney and Stephen J. Reynolds

GENERAL STATEMENT

Cordilleran metamorphic core complexes are characterized by shallow-dipping foliation-lineation arches and domes composed of metamorphosed, migmatized and anomalously deformed rocks whose protoliths can range from Precambrian to middle Tertiary in age. Most have plutonic bodies within them and late-stage granitic rocks, frequently of two-mica composition with aplitic and pegmatitic phases, are common. All the complexes are associated with high-level décollement or dislocation zones and listric normal faults in an unmetamorphosed cover terrane. Much of the deformation in the cover as well as in the metamorphosed core zone can be proven to have occurred in Tertiary time. Early to Middle Tertiary basinal continental sedimentary rocks and associated caldera related volcanic rocks are common on margins of the complexes. These sediments and volcanics are usually tilted to high angles by listric normal faulting. Geochemistry and geochronology in core-zone rocks is always complex and radiometric age determinations can range from 2 billion to as young as 11 million years. Cordilleran metamorphic core complexes are found in a rather narrow sinuous belt along the so-called "hinterland" of the eastern part of the North American Cordillera from southern Canada southward into Sonora, Mexico. Over 25 complexes are now recognized most of which have only been identified since 1970. The features are presently surrounded by considerable controversy and there exists conflicting opinion as to their timing and mechanism as well as to their regional tectonic significance.

As awareness of these newly recognized tectonic features became more widespread the question was asked as to what economic
significance they might have. One of the first things that came to mind was their possible uranium potential. The reasons for this were, in the minds of some, both theoretical and observational.

A theoretical basis for a high uranium potential seemed to rest largely on the evidence for remobilization of pre-existing continental crust and on the widespread presence of granitic plutons of a two-mica type, that are characterized by extensive late-stage, residual phases of alaskite, pegmatite, and aplite. This type of pluton was believed by some to have anatectic origins and a high potential for late-stage concentrations of uranium. Also, it is probable that hydrothermal and meteoric processes in the complexes were capable of further concentration of uranium in adjacent sedimentary basins.

The observational basis for a possible high uranium potential was suggested by the fact that many known occurrences of uranium are clearly associated with a metamorphic core complexes, although previous workers had not recognized the relationship, mostly because the complexes had not been recognized for what they were. Also, there has been much interest in the Rossing uranium deposit of southwest Africa which is classified by Bendix as "Anatetic" and which is associated with a "gneiss dome". Since some felt that the complexes were in fact gneiss domes interest was aroused in the complexes as possible sources of uranium. It was in the above context that the Department of Geosciences of the University of Arizona was approached by Bendix Field Engineering Corporation to provide a study of Cordilleran metamorphic core complexes and their uranium favorability as part of their World Class Studies program on uranium resources.

OBJECTIVES AND SCOPE OF THIS REPORT

The objective of this report is to provide a descriptive body of knowledge on Cordilleran metamorphic core complexes including their lithologic and structural characteristics, their distribution within the Cordillera, and their evolutionary history and tectonic setting. We also examine the occurrence of uranium in the context of possibility for uranium concentration.

Chapters 1-3 deal with the basic geologic data on the complexes and related features while Chapters 4-6 deal with uranium distribution and potential of the complexes in western United States. Chapter 1 is an overview of Cordilleran metamorphic core complexes which describes their physical characteristics, tectonic setting and geologic history. This overview is accompanied by a tectonic map at a scale of 1:1,000,000 of the core complex belt (Map 1-1). Chapter 2 is a discussion of the mantled gneiss dome concept. The purpose of including this work is to provide a basic
history of this concept and to describe the characteristics and distribution of gneiss domes throughout the world to enable one to compare and contrast them with the metamorphic core complexes as discussed in this report. Since some gneiss domes are known producers of uranium (as are also some core complexes) we felt it would be productive to include a discussion on them. Chapter 3 is an examination of the effects of the core complex process on adjacent sedimentary and volcanic cover terranes which can extend over 100 kms. beyond the exposed cores of the complexes themselves. Also included is a discussion of the kinematic significance of these cover terranes as they are related to process within the cores of the complexes. Since some of the cover terranes have uranium prospects in them we include a discussion of them. Chapter 4 is a detailed discussion of uranium in Cordilleran metamorphic core complexes and includes the conceptual basis for the various types of occurrences and the processes that might favor concentration of uranium.

The report is supported by a 5-part Appendix. Appendix A is a complete annotated bibliography on Cordilleran metamorphic core complexes, gneiss domes in world geologic literature, and any geologic-tectonic sources relevant to the core complex problem. Appendix B is an annotated bibliography on the uranium aspects of metamorphic core complexes, gneiss domes, and two-mica granites. Appendix C is a listing of uranium occurrence in the core complex belt of western United States and is accompanied by 1:1,000,000 scale maps of each of the states within which the belt occurs, showing the locations of the uranium occurrences included in the list. Appendix D is a geologic and uranium distribution summary of each of the core complexes accompanied by a tectonic-geologic map of each of the complexes at a scale of 1:250,000. In this section the uranium favorability of each complex is discussed. Appendix E is a data list of sample locations, geochemical data, and petrographic descriptions.

It is worth pointing out in conclusion that the majority of the core complexes discussed in this report either do not appear or are not recognizable on existing published geologic maps. They are, without question, the newest addition to the recognized architecture of the Cordillera. This report is an attempt to rectify the inadequacy of existing information and to provide a basis to access both their economic as well as scientific potential.
Chapter I
CORDILLERAN METAMORPHIC CORE COMPLEXES:
AN OVERVIEW

by

Peter J. Coney

INTRODUCTION

Cordilleran metamorphic core complexes are a group of usually domal or arch-like, isolated uplifts of anomalously deformed, metamorphic and plutonic rocks, overlain by a tectonically detached and distended unmetamorphosed cover. The features (see Figure 1-1) are scattered in a sinuous string along the axis of the eastern two-thirds of the North American Cordillera from southern Canada to northwestern Mexico. To date, over 25 of them are known and it is significant that over half of them have been recognized only since 1970. They are, without question, the newest and most controversial addition to the recognized architecture of the eastern two-thirds of the Cordillera since the discovery, in the early 1960s, of the Tertiary calderas and their associated vast ignimbrite sheets. During the past 10 years, considerable debate has centered on them, and they have been termed metamorphic core complexes (Coney, 1973a, 1976, 1978, 1979; Davis and Coney, 1979; Crittenden et al., 1978).

The first hint of the existence and potential significance of these complexes was from the early efforts of Peter Misch and his students who began work in the Great Basin in the 1950s. Misch (1960) discovered and emphasized scattered occurrences of a major sub-horizontal dislocation plane or "décollement", as it was called, separating an unmetamorphosed Paleozoic miogeoclinal cover from a usually metamorphosed substratum. He found this repeatedly
in many ranges along and west of the Utah-Nevada border about 200 kms west of the already known low-angle, east-verging Mesozoic thrust faults in central Utah.

To Misch (1960) the most important element of the eastern Nevada structural province, or hinterland, was the "décollement." This discontinuity was seen to separate Precambrian basement and metamorphosed late Precambrian-early Paleozoic sediments from an overlying unmetamorphosed allochthon. The type area for this regional décollement was the Snake Range of east-central Nevada. Misch reasoned that the décollement and the associated shearing and metamorphism along it were produced during mid-Mesozoic orogeny. He felt that the discontinuity was formed as cratonic basement moved westward into deep-seated thrust roots, peeling and shearing off the Phanerozoic cover as it moved. The implication was that the basal shearing-off plane, or décollement, was structurally continuous with pre-Laramide low-angle break-out thrust faults to the east in central Utah, along and west of the Wasatch line.

Working in the same region, Armstrong (1968b) and Armstrong and Hansen (1966) emphasized remobilization of basement rocks and metamorphism of the lowest part of the Phanerozoic cover during mid-Mesozoic orogeny. In an analogy to Caledonian systems, they termed the remobilized core zone an infrastructure. In contrast to Misch, they emphasized mobility below the décollement (or abschie- rung zone) and contrasted this domain of recumbent folds and planar fabrics to a less deformed, brittle suprastructure above.

The Price and Mountjoy (1970) model for the tectonic evolution of the eastern Cordillera of southern Canada was one of the best formulated and persuasive tectonic syntheses in the history of Cordilleran geologic thought. They proposed that a hot, mobile infrastructure rose buoyantly in the Shuswap axial metamorphic core zone and gravitationally spread eastward, propelling the Rocky Mountain foreland fold and thrust belt to the east. The deformation was seen as continuously evolving upward and eastward from late Jurassic to early Tertiary time.

The low-angle thrust faults along the Wasatch line in central and southern Utah are "older on younger" faults (Misch, 1960; Armstrong, 1968a) typical of foreland thrust belts the world over (Coney, 1973b). In the hinterland to the west, however, above the metamorphic rocks, low-angle faults are "younger on older" faults (Armstrong, 1972). Nothing quite like these widespread "younger on older" faults has been emphasized in the Canadian Rocky Mountains. In spite of this, the Price and Mountjoy model had great influence on workers in the western United States. As a result, subsequent syntheses attempted to apply modifications of the Canadian example to eastern Nevada and western Utah (Roberts and Crittenden, 1973; Hose and Danes, 1973). Hose and Danes, for example, interpreted
the décollement and younger on older faults in cover rocks as the result of extensional gravity-driven movement of the cover off an uplifted hinterland in eastern Nevada (see Figure 3-B). This cover terrane slid eastward to become the superficially telescoped older on younger thrust faults of central Utah.

This sets the stage for developments, mostly after 1970, which were to cast considerable confusion over the simplicity and beauty of the early models. Early workers noted certain data and relationships which were either troublesome or inconsistent with existing models. Damon (Damon and others, 1963; Mauger and others, 1968) found very young mid-Tertiary K-Ar "cooling ages" from metamorphic rocks in southern Arizona. Armstrong and Hansen (1966) found similarly young Tertiary ages from metamorphic rocks in Nevada. Misch (1960) was aware of Tertiary gravitational gliding and superficial brecciation in the Snake Range of eastern Nevada, and Drewes (1964) discussed multiple thrusting and gravity faulting extending into Tertiary time in the Schell Creek Range of eastern Nevada. Moores and others (1968) recognized Tertiary deformation and metamorphism in the White Pine-Grant Range of eastern Nevada. For many workers, most of these inconsistencies seem to have been explained as the result of Tertiary modifications of a mainly Mesozoic tectonic history, or as isolated events of only local significance. In some cases, the local importance was adequately emphasized, but the regional significance was not appreciated by others. The consensus seemed to be that these were minor Tertiary modifications of a basically Mesozoic tectonic regime of metamorphism and low-angle thrusting.

The first clear statement that initiated turn-around of this consensus was made by Armstrong (1972). He proposed that widespread "denudational" low-angle faulting in mid-Tertiary time was a possible explanation for the "regional décollement" of the eastern Nevada hinterland. Using geochronology and field relations, he reinterpreted existing published geologic mapping and structure sections. His results suggested that the extensional younger on older faults, which cut well-dated mid-Tertiary volcanic rocks as well as associated sedimentary rocks and flatten at depth to merge with the décollement plane were, at least in part, as young as the Tertiary volcanics they cut. The implication was clear. The décollement surface was, in part at least, mid-Tertiary in age and possibly had only little or perhaps nothing to do with Sevier age Mesozoic thrusting to the east. The work of Lee and others (1970) was very significant in regard to this problem. They showed that K-Ar ages from a well-dated Jurassic pluton below the décollement in the southern Snake Range were progressively reset to younger ages as one approached the décollement. The ages decrease to about 18 m.y. at the discontinuity. They concluded that the most recent
movement, at least on the surface, was that young and Armstrong (1972) entirely agreed with them.

Working independently, Coney (1974) in the Snake Range of eastern Nevada and Davis (1973, 1975) in the Catalina-Rincon Mountains of Arizona both advocated mid-Tertiary low-angle gravity sliding of unmetamorphosed cover rocks off metamorphic basement on a décollement surface. At about the same time this author heard (Crittenden, pers. comm., 1972) that Todd (1973), working in the Raft River-Grouse Creek range, had found that allochthonous sheets of Paleozoic cover rocks had moved off a metamorphic basement onto middle-to-late Tertiary sediments. Finally, Compton and others (1977) found evidence that the younger metamorphic fabric characteristic of the Albion-Raft River-Grouse Creek metamorphic complex was imprinted on a mid-Tertiary pluton.

All of this work implied that significant thermal disturbance, metamorphism, and deformation in the hinterland extended into mid-Tertiary time. This is much younger than, and well clear of, the proven age of foreland thrusting to the east. This time sequence cast suspicion on models that linked the metamorphic hinterland to the thrusting, and it raised the disturbing prospect that we were dealing with a very young, special, and enigmatic tectonic response of obscure significance.

We owe our readers an explanation of the term Cordilleran metamorphic core complex. There has been objection to the name and widespread wonderment about its meaning. The name was coined, quite accidentally, by me in 1973 (Coney, 1973a, p. 723) in reference to the Shuswap metamorphic complex and its relationship to the Canadian Rocky Mountain foreland thrust belt. I am quite sure it was an unconscious reference to the term "metamorphic core zone" used by Canadian workers (Wheeler and others, 1972) to describe the eastern Omineca crystalline belt in the Canadian Cordillera. Since then, the term has been applied by myself and others to the many metamorphic terranes and related features found southward in the Cordillera as they were recognized or re-evaluated. If for no other reason, the term has stuck because the complexes seemed so distinctively peculiar and hitherto not properly recognized that they needed a special name. For example, the metamorphic-plutonic basement terrane so typical of the complexes can be thought of as a "core" terrane. I make no apology for the term Cordilleran metamorphic core complex. It has proven useful for many of us and it simply indicates that, although variety exists, the objects so identified have a considerable commonality as to physical features and age.
CHARACTERISTICS

The metamorphic complexes occur in a discontinuous belt extending from southern Canada south through the Cordillera into Sonora, Mexico (Fig. 1). In general, these complexes are characterized by distinctly similar lithologies, structures, and fabric. These similarities are among the most remarkable aspects of the complexes and are noticed by anyone with more than a casual acquaintance with more than one of them.

Two distinct domains characterize the complexes (Fig. 2). These are a metamorphic-plutonic basement terrane or core zone, and an overlying or adjacent unmetamorphosed cover. Separating the two is a sharp discontinuity, or zone, marking rapid or abrupt change in lithology and structure. Rarely do lithology and structural fabric characteristic of either the basement or cover cross the discontinuity into the other domain.

The gross aspect of the complexes is domal or anticlinal, usually with an asymmetry such that one flank is slightly steeper than the other. Quite frequently, the complexes stand prominently in topography as the highest mountains in their respective regions. They are usually recognized from afar, the low-sweeping domal profile distinctive on the horizon.

The metamorphic basement is characterized by a low-dipping foliation whose attitude usually conforms to the overall domal or arch-like shape of the complex. This foliation imparts a distinct gneissic aspect to the rock. Within the foliation plane is an inevitable mineral lineation. The bearing of the lineation (but not necessarily the plunge) is often remarkably constant within a given complex and sometimes in adjacent complexes as well. In more than 15 complexes stretching over a distance of 400 km in southern Arizona and Sonora, for example, it is almost universally about N60°E (Davis, 1980). Northward the lineation is more variable, but it tends to be either east-west or northwest-trending (Coney, 1974; Misch, 1960). In some cases, the lineation lies close to the axes of the domes or arches, but in other cases it cuts across this trend. The dip of foliation on flanks of the domes is usually shallow, rarely exceeding 20° to 30°. Both the foliation and lineation are usually described as "cataclastic," but seem to involve recrystallization, particularly in quartz, as well as cataclasism (Todd, 1980; Reynolds and Rehrig, 1980). The rocks are best described as mylonitic gneiss. Strain is extreme, and axial ratios of mineral elongations and occasional stretched pebble conglomerates, can attain 8:2:1 (Coney, 1974; Compton, 1980; Compton et al., 1977; Davis, 1980). The overall strain picture is one of maximum shortening and flattening perpendicular to the sub-horizontal foliation plane and maximum extension parallel to the lineation (Compton, 1980). The direction of maximum elongation is
Figure 1-1

Schematic structural block diagram of typical domains of Cordilleran metamorphic core complexes. A-basement terrane; B-cover terrane; C-décollement zone; a-older metasediments; b-older pluton; c-younger pluton (Early to Middle Tertiary); d="cataclastic" foliation; e="cataclastic" lineation; f-marble tectonite (black); g-Early to Middle Tertiary sediments and volcanics.
frequently described as sub-parallel to minor recumbent flattened, and attenuated fold axes. There are also cases known (where lithology is favorable) of minor folds within the gneiss which suggest preferred vergence and simple shear in the same direction as the lineation. The inferred directions of movement are not universally northeast or east but form large domains which oppose one another over large regions (See discussion in Chapter 3 by Davis and Hardy). Davis has also found small late-stage normal faults whose strike is perpendicular to the lineation (Davis and others 1975; Davis, 1975, 1977, 1980). These faults seem to be the result of progression from ductile behavior to brittle failure. On the fault surfaces are slickensides whose bearing is sub-parallel to the lineation.

In those complexes where erosion has cut deeply into the uplifts, the above-mentioned foliated and lineated fabric can sometimes be seen to diminish downward into an earlier, usually steeper metamorphic fabric, or sometimes into a more homogeneous plutonic fabric (Coney, 1974; Todd, 1980; Reynolds and Rehrig, 1980). In some cases the superimposed mylonitic zone is only tens of meters thick. The deeper, earlier metamorphic fabrics are often quite complex and record poly-phase deformations and complex history (Reesor, 1970; Hyndman, 1968; Miller, 1980). The distinctive mylonitic foliation and lineation are therefore superimposed on either an earlier metamorphic fabric or on homogeneous plutonic fabrics.

The protoliths of the metamorphic basement are extremely variable both in lithology and in age. This is a point that cannot be overemphasized. In most cases, several protoliths exist in a single complex. At one place or another, the protoliths include proven older Precambrian metasedimentary basement (Reynolds and Rehrig, 1980), older Precambrian plutons which intrude into the metasediments (Banks, 1980; Compton and others, 1977; Shakel and others, 1977), Late Precambrian sedimentary rocks, Paleozoic sedimentary rocks (Misch, 1960; Howard, 1971; Thurman, 1970), probably Mesozoic sedimentary rocks (Rehrig and Reynolds, 1980; Hyndman, 1968), Laramide age plutonic rocks (Anderson and others, 1980), and even early to mid-Tertiary plutonic rocks (Reynolds and Rehrig, 1980; Keith and others, 1980). All of the above protoliths have had the distinctive late mylonitic foliation and lineation superimposed on them in one complex or another.

As one approaches the discontinuity separating the basement and cover terranes, all of the basement fabrics, including the late mylonitic fabric, are demonstrably brecciated and truncated at the discontinuity by a still later deformation apparently related to movement along the discontinuity (Coney, 1974). This latest deformation usually places unmetamorphosed cover rocks in direct contact with brecciated and locally truncated and disturbed
basement rocks. Regionally, however, the discontinuity, or "décollement" as it is often called, is subparallel or exactly parallel to the underlying mylonitic and gneissic fabric. This zone of brecciation is very distinctive and sometimes extends into non-mylonitic older basement. But it is always found below the décollement zone.

Granitic plutons are extremely common in the basement terrane. They are of special interest since some are described as of the two-mica garnet-bearing type (Chappell and White, 1976; Best and others, 1974; Miller and Engels, 1975). Other than the plutons, pegmatitic and migmatitic rocks are common, as are other late-stage differentiates and leucocratic phases. These smaller bodies usually form sheet-like or lensoid masses and fine stringers. These smaller bodies are frequently sub-parallel to the foliation, but can also cross it. Some have the mylonitic fabric superimposed on them, while others do not. The larger plutons are usually quite homogeneous at deeper levels, but usually progressively acquire the characteristic foliation and lineation approaching the décollement. The plutonic pegmatitic and migmatitic bodies never cross the décollement, and are frequently abruptly terminated at the décollement. In a few cases for which documentation is just emerging, even the largest plutons are apparently gigantic sill-like masses emplaced at or just below, and sub-parallel to the level of the décollement surface (Banks, 1980; Keith and others, 1980).

Metamorphic grade in the basement terrane is quite variable, but is usually moderate to high. In many complexes, the older and deeper metamorphic grade was quite high, and kyanite-sillimanite-andalusite are not uncommon (Reesor, 1970; Armstrong, 1968; Compton and others, 1977). The younger "cataclasis" presumably formed at more moderate conditions.

The overlying cover terrane consists of unmetamorphosed rocks separated from the metamorphic-plutonic core by the décollement, or by a zone of very steep metamorphic gradient. In many cases, very little of this superficial covering terrane remains at the surface since it has frequently been largely stripped away by erosion. Commonly, the only remnants are isolated "klippen," as they are usually termed, scattered around the margins of the complex.

Like the basement terrane, lithology and age of the cover rocks are highly variable. In one complex or another, the cover rocks consist of slivers of original Precambrian basement (Davis 1980; Drewes, 1977), late Precambrian sedimentary rocks, Paleozoic sedimentary rocks (Coney, 1974; Compton and others, 1977; Thorman, 1970), probable Mesozoic sedimentary rocks (Rehrig and Reynolds, 1980), and Tertiary volcanic and sedimentary rocks.
All of the above rocks can be demonstrated to have moved relative to the basement terrane along the décollement including, it is to be emphasized, Early to Mid-Tertiary volcanic rock and associated Tertiary sedimentary rocks (Davis et al., 1977; Davis et al., 1980).

In those complexes where sufficient cover terrane is preserved to make observations, structures in the cover are very complex. Workers are usually impressed by many low-angle, younger on older bedding plane faults and by many extensional listric normal faults (Coney, 1974; Hose and Danes, 1973). At times, the entire cover terrane can take on the character of a mega-breccia. It is extremely rare that faulting of any kind penetrates into the basement below the basal décollement. In other words, the décollement marks a discontinuity of extreme ductility contrast—brittle above, ductile below.

In some areas, detailed structural analysis of cover rocks reveals structures ranging from minor folds to slickensides on faults, or on the décollement surface itself, that can be interpreted to reflect movement of cover rocks down present dips of the décollement surface into adjacent basins (Coney, 1974; Davis, 1975). The movement directions inferred are in some cases at a high angle to the bearing of lineation in the underlying basement, while in other cases they are sub-parallel. Also, these earlier basement fabrics, including the mylonitic foliation and lineation, are re-deformed, usually brittlely, into geometries consistent with the movement picture derived from the cover (Coney, 1974). This late movement seems to have been associated with intense brecciation seen in both cover rocks and in basement terrane just below the décollement. Water was abundant, evidenced by clastic dikes and chloritic and hematitic fillings in the pervasive fractures.

The amount of extension in the cover terrane is often dramatic. In several complexes, attenuation of original stratigraphy is extreme (Compton and others, 1977; Todd, 1980; Davis, 1975) and lithologic units once very high in the original cover sequence have been brought down into contact with the décollement surface. In some cases, the stratigraphic separation is well over 2 km. It is not uncommon to find Tertiary rocks, the youngest of the original cover, brought down in tectonic contact with the basement terrane.

In many of the complexes, particularly in southern Arizona-Sonora, part of the cover sequence is an early to mid-Tertiary continental sequence composed of conglomerate, fanglomerate, sandstone, siltstone, lacustrine units, and evaporitic deposits (Pashley, 1966). These sedimentary rocks are often red and frequently associated regionally (usually below) with well-dated early to mid-Tertiary volcanics (Armstrong, 1972). All these rocks are
usually tilted to high angles. These distinctive continental sediments are so consistently found in association with the complexes in Arizona that they suggest a genetic link.

The sediments are usually very thick, and can attain several kilometers. In those few cases where investigations have been made (Pashley, 1966), current indicators and pebble counts suggest that the fluvial systems did not drain from the core complexes, but may have actually flowed toward them. Only in upper levels of the deposits do streams appear to have drained off the complexes, and it is only in these youngest horizons that clasts of foliated-lineated basement rocks occur.

These distinctive Early to Middle Tertiary sediments and volcanics commonly extend over wide regions beyond the core complexes. They are particularly well preserved in southern Arizona and in southeastern California and southern Nevada. They are inevitably tilted to high angles and show conclusive evidence of deformation by listric normal growth faults. These listric normal faults can frequently be shown to flatten with depth and merge into a décollement-like zone placing the Tertiary rocks directly on crystalline basement. This relationship and geometry was first recognized by Anderson (1971, 1977, 1978) in the Lake Meade region and later extended into Arizona by Rehrig and Heidrick (1976), Eberly and Stanley (1978), Scarborough and Peirce (1978) and Shafiqullah and others (1978).

Quite remarkably, the direction of strike in these tilted sequences over thousands of square kms. is regionally perpendicular to the direction of lineation in the mylonitic gneiss core terrane of the complexes themselves. In other words, the movement picture derived from the faulted sediments and volcanics is the same as that derived from the mylonitic gneiss in the core complexes. This is in spite of the fact that there is usually evidence that the listric normal faulting is younger that the mylonitic gneiss. For a more complete discussion of these relationships see Chapter 3 by Davis and Hardy in this report.

The décollement, or dislocation surface is perhaps the most distinctive aspect of the core complexes thus far recognized (Misch, 1960; Coney, 1974; Whitebread, 1968; Nelson, 1966; Davis and others, 1980; Miller, 1972). Although variations exist, the general characteristics of the décollement are so remarkably similar from one complex to another that the feature is instantly recognized.

In some areas, the decollement is located quite close to either the Precambrian-Phanerozoic unconformity or in horizons directly above thick Late Precambrian-Early Paleozoic quartzite-siltstone sequences. If carbonates lie below the decollement
they are usually metamorphosed and attenuated and intensely deformed into a marble tectonite rarely over several tens of meters thick (Misch, 1960; Nelson, 1966; Coney, 1974; Whitebread, 1968). Davis (1980) has referred to this tectonite band as a metamorphic carapace. Unmetamorphosed Paleozoic carbonates can directly overlie their metamorphosed equivalents below the décollement. Where the basement is composed of plutonic rocks, the décollement usually lies directly above them and the plutons never cross the décollement. A remarkable example in the southern Snake Range of eastern Nevada (Whitebread, 1968), has a Jurassic pluton as basement. This pluton is overlain along a clearly tectonic low-dipping planar surface by up to 30 meters of marble tectonite. The décollement lies above the marble and is overlain by unmetamorphosed Cambrian limestone. Here K-Ar ages on the Jurassic pluton decrease to 18 m.y. approaching the décollement (Lee and others, 1970; Armstrong, 1972; Coney, 1974).

The décollement surface is locally very sharp and clearly visible in topography. It is generally polished and has slickensides, and rock directly below can have the aspect of a fused paste or welded small clast-sized breccia.

Rocks near the décollement, both above and below, are usually brecciated and show extensive alteration (Reynolds and Rehrig, 1980). Retrograde chlorite is very common in brecciated basement rocks, and development of a distinctive hematitic red-stained fracture filling between fragments is ubiquitous. These breccias have the appearance of "exploded rocks." Even in thin section, so-called mylonitic zones are clearly without planar fabric, but rather show an exploded micro-breccia aspect.

The décollement is typically best developed on one side of a particular complex, usually on the less steeply dipping flank of the commonly asymmetrical dome or arch (Reynolds and Rehrig, 1980). On the other, steeper side, the discontinuity can be less sharp or tectonically abrupt with only very steep metamorphic gradient and little, if any, evidence of major relative movement between cover and basement demonstrable. In at least two complexes, several slices, or thin packets of discontinuity-bound rocks, are found extending across the dome or arch. Three separate slices are recognized in both the Raft River-Grouse Creek and Rincon complexes. In Raft River-Grouse Creek (Compton, 1972, 1975; Compton, 1972, 1975; Compton and others, 1977; Todd, 1980), the two lower slices are metamorphosed, but the upper is not. In the Rincon Mountains (Davis, 1980, Drewes, 1977), the lowest slice is metamorphic Paleozoic rocks extremely attenuated, the middle slice is original Precambrian basement, and the upper slice is unmetamorphosed Paleozoic and Mesozoic rocks. The core in both examples is Precambrian granite and metasediments and early to middle Tertiary plutons.
Are the complexes basically classic gneiss domes (Escola, 1949)? This question is often asked. Certainly, some show certain characteristics of classic gneiss domes, particularly the extreme stretching across the top, the overall domal or arch-like geometry, steep metamorphic gradient, etc. There are, however, certain difficulties. First, the actual doming is apparently a late feature superimposed on the metamorphic fabric of the basement. Second, the actual structural relief on the domes is not particularly great. For example, the amplitude of most of the domes is no more than 4 km in a wavelength of as much as 50 km, giving an amplitude-wavelength ratio of .08—a fact reflected in the universal low-dipping foliation which rarely exceeds 20° to 30° (e.g., see Drewes, 1977). This is considerably less structural relief than that usually depicted in classic gneiss domes or in experimental or theoretical modeling (Ramberg, 1972; Dixon, 1975) where amplitude-wavelength ratios are .25-.5 or more. Third, it has been emphasized earlier in this paper that the cataclastic foliation and lineation so characteristic of the complexes in many cases actually seems to diminish and even disappear downward in the basement terrane. This argues for rigidity of this terrane in some cases since Precambrian time (Compton and others, 1977). It also argues against deep mobility characteristic of an "infrastructure" usually cited as an essential ingredient of classic gneiss domes.

In any event, some workers in the Cordillera have found it useful to informally and tentatively reject the classic gneiss dome concept if for no other reason than to aid objectivity and identification of the real characteristics of the Cordilleran complexes. Perhaps the main difference is that the Tertiary aspects Cordilleran core complexes evolved in a regional non-compressional, perhaps even extensional tectonic setting, whereas domes characteristic of other orogens did not. In fact, it is the overwhelming evidence for pervasive extension in basement fabric, structures in the cover and along the décollement, that dominates all characteristics of Cordilleran metamorphic core complexes. For a more complete discussion of the gneiss dome concept see Chapter 2 of this report.

REGIONAL TECTONIC SETTING

The regional tectonic setting of Cordilleran metamorphic core complexes is portrayed on the 1:1,000,000 tectonic map accompanying this report. (Map 1-1) To date, no specific regional tectonic relationship of the complexes has been demonstrated to everyone's satisfaction. Regardless, it has become clear that the complexes are an important element in the overall architecture of the North American Cordillera.

Distribution of metamorphic rocks in western North America,
excluding the cratonic Precambrian basement beneath the eastern margin of the orogen, grossly reveals two sub-parallel belts (King, 1969; Monger and Hutchison, 1971). A western belt is largely a metamorphic sheath below, within and adjacent to the great belt of Cordilleran batholiths. This belt extends through the Canadian coastal plutonic complex, the Idaho batholith, and the Sierra Nevada–Peninsular batholith southward into western Mexico. An eastern belt (Coney, 1978) follows the Omineca crystalline complex of the eastern Cordillera in Canada, culminating in the Shuswap terrane of southern British Columbia which extends into north-eastern Washington (Cheney, 1977, 1980; Fox and others, 1977). southward across Nevada, southeastward across Arizona, then southward into Sonora. All of the so-called Cordilleran metamorphic core complexes as defined here lie in this eastern belt, extending at least from southern Canada to northern Mexico.

Most of the eastern belt developed either on or very close to the edge of original North American Precambrian cratonic basement. In contrast, most of the western belt may have developed in magmatic arcs on oceanic crust or on crustal fragments, either recently of subsequently accreted against North America's continental margin (Dickinson, 1976; Coney, 1978). Finally, the apparent continuity of both belts does not imply historical continuity. The ages of batholiths in the western belt vary along strike, and much of the northern two-thirds of the eastern belt records a much more prolonged metamorphic history than the southern part. It is the most recent metamorphic–tectonic events of the eastern belt that we identify as characteristic of the core complexes, and these are superimposed on a variable pre-core complex history and tectonic evolution.

Pre-Mesozoic Tectonic Trends

Of the 25–odd Cordilleran metamorphic core complexes currently identified, all evolved in terrane underlain by North American Precambrian continental cratonic basement as defined by the .706 ^87Sr/^86Sr line (see Fig. 1) (Armstrong and others, 1977; Kistler and Peterman, 1973). Two possible exceptions are the Okanagan (Fox and others, 1976) and Kettle (Cheney, 1977, 1980) complexes in Washington which appear to lie west of proven North American Precambrian basement. In most complexes, the Precambrian basement is either proven or implied by isotopic dating (Wanless and Ressor, 1974; Clark, 1973; Armstrong and Hills, 1976; Compton and others, 1977; Shakel and others, 1977); and regional relations seem to demand its presence in most of the remainder. Looking more closely at this Precambrian basement, controls on evolution of metamorphic complexes are not obvious. Precambrian lithologic and structural trends, which are generally northeasterly, strike at high angles to the trend of the belt of complexes and basement ages crossed by the belt range from greater than two billion years in the north to
about one billion in the south (King, 1969). The Arizona complexes may parallel a northwest-southeast trend of late Precambrian right-shear just southwest of the Colorado Plateau, but so do all other post-Paleozoic tectonic trends.

From southern Nevada northward, all the complexes lie within the thick Cordilleran latest Precambrian-Paleozoic miogeoclinal prism. In Arizona and Sonora, on the other hand, the complexes evolved well inboard of the Paleozoic miogeoclinal on what had been a thin cratonic shelf (Peirce, 1976). As a result, relating the complexes to a zone of deep Paleozoic burial and thick Paleozoic deposition over the eventual site of the core complexes is unlikely. Paleozoic cover varies from almost 15 km in southern Canada, to about 10 km in Nevada, to only 2 km in southern Arizona.

Similarly, distribution of the complexes with respect to Paleozoic orogenic activity and metamorphism is variable. The northern complexes are found in a region that probably suffered mid- and late-Paleozoic Antler-Sonoma deformation and, in the case of the Shuswap complexes, some Paleozoic metamorphism (Okulitch and others, 1975; Read and Okulitch, 1977; Brown and Tippett, 1978). Southward in Nevada and Arizona, the complexes lie well east or south of any profound Paleozoic thermal or tectonic events.

Relationship to Mesozoic-Early Cenozoic Trends

Latest Paleozoic-early Mesozoic time was a major transition in Cordilleran tectonic evolution (Coney, 1973a; Burchfield and Davis, 1976). It marks inception of draping and accretion of magmatic arcs on the North American margin and the evolution of arc-rear thrusting and folding inboard from the magmatic belts.

Genetic linking of Cordilleran metamorphic core complexes to Cordilleran thrust belts has been the most persistent and persuasive argument put forward to explain the phenomena under study. The concept has manifested itself in several ways, and has influenced discussion of Cordilleran tectonics in general as well as discussion of the complexes themselves. No aspect of the complexes has generated more controversy than this.

From Nevada northward, the complexes lie within a belt about 200 km west of the thin-skinned foreland folds and imbricate thrust faults so characteristic of the North American Cordillera and other orogens (Coney, 1976). This belt of metamorphic rocks has been termed a "hinterland" behind the thrust faults where more deep-seated tectonics is supposed to have taken place with associated uplift, thermal disturbance, and remobilization. The assumption has been that the metamorphism accompanied the thrusting and was
thus genetically linked to it. It was this position that so heavily influenced early models such as those of Misch (1960) and Price and Mountjoy (1970).

In Canada, Paleozoic and lower Mesozoic rocks are clearly metamorphosed (Hyndman, 1968), and in Nevada Paleozoic rocks are involved (Misch, 1960). From regional relationships in Canada, some of the metamorphism is Paleozoic and most of it is at least as old as mid-Jurassic (Wheeler and Gabrielse, 1972; Brown and Tippett, 1978). Ironically, recent work in the southern Rocky Mountains of Canada suggests that much of the complex poly-phase deformation and metamorphism so characteristic of the Shuswap terrane is pre-late Jurassic (Brown, 1978; Wheeler and Gabrielse, 1972), thus earlier than the well-documented late Jurassic to early Tertiary folding and thrusting to the east. Furthermore, the increasing evidence that the mylonitic metamorphism in United States core complexes is Tertiary (Compton and others, 1977; Cheney, 1980; Reynolds and Rehrig, 1977, 1980; Anderson and others, 1980) and hence much later than Sevier or Laramide thrusting has clouded the issue of thrusting and metamorphism more than anything else.

In Arizona and Sonora, the complexes are not in a "hinterland" behind a thrust belt, but are in fact in the midst of a Laramide belt of deformation, although of a somewhat different character from the thin-skinned folds and thrusts of Sevier-Laramide age to the north (Davis, 1979). Here deformation was apparently more deep-seated, involving both basement and cover rock. Furthermore, the metamorphism and deformation associated with most of the complexes is clearly younger than Laramide tectonic features and is superimposed on post-thrusting Laramide plutons and even on some mid-Tertiary plutons (Anderson and others, 1980: Reynolds and Rehrig, 1980). As is the case with so many other regional aspects of Cordilleran metamorphic complexes, the Arizona-Sonora examples show a major departure from relationships historically so suggestive to the north.

In Nevada, the original battleground of metamorphism and thrusting in the Cordillera, the relations are less clear. An older, mid-Mesozoic, metamorphism has long been advocated in the hinterland (Misch, 1960); and certainly exists in the Ruby Range (Howard, 1971, 1980; Snoke, 1980), and probably in the Snake Range as well. As stated earlier, however, this is an older metamorphism exposed deep in the cores of some of the complexes. Superimposed on this earlier, generally steeper fabric is the shallow-dipping late "cataclastic" fabric so characteristic of the complexes throughout their extent. This late metamorphism and associated deformation are at issue here. That event appears to
be superimposed on plutons as young as Early to Middle Tertiary (Compton and others, 1977), thus much younger than the late Jurassic to late Cretaceous thrusting in the Sevier thrust belt to the east in central Utah.

The thrust belt-core complex concept has played another significant role in interpretations of the complexes. The concept has led to identification of the distinctive "décollement" surface and its associated cataclasis with the basal shearing-off plane of the thrust belts (Misch, 1960). This interpretation has been variably invoked in Idaho (Hyndman and others, 1975), Nevada-western Utah (Misch, 1960; Hose and Danes, 1973), and in Arizona (Drewes, 1976, 1978). Obviously, the age of shearing-off would have to be of Sevier age in Nevada, of Sevier-Laramide age in Idaho, and of Laramide age in Arizona. The fact that the décollement surface and the "cataclastic" deformation associated with it are now known to cut rocks as young as Early to Middle Tertiary in all these regions casts some doubt on this concept (See Fig. 5). As a result, some workers have more recently argued models of multiple thrusting (Drewes, 1976, 1978; Thorman, 1977) and/or mid-Tertiary gravitational sliding on a décollement surface originally made during regional low-angle thrusting during Sevier or Laramide time (see discussion in Compton and others, 1979 and Crittenden, 1979). These models seem in some cases at least geometrically difficult. Armstrong (1972), for example, has argued on geometric grounds that the décollement surface in the Snake Range of eastern Nevada is unrelated to the basal shearing-off plane of Sevier age thrusts to the east (Fig. 3):

In any event, the debate concerns whether the characteristics so typical of the core complexes, namely the décollement surface, the distinctive mylonitic or "cataclastic" foliated and lineated fabric, and younger on older "faults" in the cover, are genetically linked to regional thrusting of Mesozoic-early Cenozoic age. No one can deny deformation and even metamorphism of Sevier and/or Laramide age in some of the regions now occupied by core complexes. What is argued is that the younger features typical of the complexes are superimposed on earlier fabrics: that the younger features are probably, for the most part, Tertiary in age; and that they are seemingly unrelated to the thrust belts at least in any direct way.

Since the core complexes are clearly, in part, of thermal origin, distribution of magmatic activity is of interest in discussing their evolution. Compilations of Mesozoic-early Cenozoic magmatic activity reveal considerable complexity in both space and time. This activity encompasses the entire Cordillera and spreads over a far greater region than the rather narrow belt of complexes as defined here.

The main result of this magmatic activity was the eventual emplacement of a massive Cordilleran batholith system comprising
Two contrasting interpretations of structure from the Snake Range, Nevada to the Wah Wah Range, Utah based on a section by Armstrong, 1972. Figure 3A - deep-seated thrust fault model showing basal Sevier thrust in Wah Wah Range rooting beneath the Snake Range. This interpretation was favored by Armstrong. Figure 3B - shallow thrust fault model which connects the Snake Range décollement is not connected to the Sevier thrust faults to the east. This interpretation was favored by Hose and Danes, 1973. In Fig. 3A the Snake Range décollement is not connected to the Sevier thrust faults, but is interpreted as a Tertiary denudation fault (Armstrong, 1972) off the Snake Range core complex. 
a-Precambrian basement; b-Upper Proterozoic-Lower Cambrian clastic rocks; c-Paleozoic carbonates; d-Snake Range décollement. X-pattern-Mesozoic-Cenozoic plutons.
the Canadian coastal plutonic complex, the Idaho batholith, and Sierra Nevada–Peninsula batholith in California and western Mexico (King, 1969). These bodies are not all the same age. Furthermore, with the exception of the Idaho batholith, all are well west of the belt of metamorphic core complexes.

Lesser plutonic bodies extend eastward across the Cordillera into, and even east of, the core complex belt. In Canada, the Omineca crystalline belt has plutons of late Jurassic and Cretaceous ages (Gabrielse and Reesor, 1974), most of which clearly cross-cut the metamorphic fabrics typical of the Shuswap terrane (Hyndman, 1968).

In the United States, the Washington–Idaho complexes are superimposed on, or certainly co-existent with, the main batholith belt (Miller and Engels, 1975). Further south, in Nevada, the complexes lie well east of the batholith belt, but well within a field of scattered Jurassic to late Cretaceous plutons. These plutons are perhaps best explained as scattered intrusions parasitic to the main batholith belt which were produced either from deeper levels of the descending slabs or from transient variable dips in these slabs. The Arizona–Sonora complexes are within a broad belt of mainly late Cretaceous–early Tertiary scattered Laramide magmatic activity which swept inboard from the Cretaceous peninsular batholith along the coast after 80 m.y.b.p. (Silver and others, 1975; Coney and Reynolds, 1977).

In summary, Mesozoic–early Cenozoic magmatic arc activity spread over the entire Cordillera at one time or another, including the narrow core complex belt. Most of the volume of plutons was emplaced well west of the belt of core complexes. Finally, with the possible exception of the Omenica crystalline belt in Canada, no unique or distinct pre–Early–Middle Tertiary magmatic trend coincides with the belt of complexes, either in space or in time.

Relation to mid–Cenozoic Tectonic Trends

Starting about 55 m.y.b.p. and extending to about 20 m.y.b.p., a very complex pattern of post–Laramide magmatic activity spread over the entire southern Cordillera (Armstrong, 1974; Coney and Reynolds, 1977; Noble, 1972; Lipman and others, 1971). It was characterized by enormous outbursts of caldera–associated ignimbrites and the emplacement of shallow plutons. This massive thermal disturbance reset radiometric ages over thousands of square kilometers, and also caused much low–angle normal faulting. This enigmatic phenomenon is very important because the core complexes seem to have evolved just prior to and during the outburst (Coney, 1974, 1978). What is puzzling, however, is that the outburst covered a region far wider than that of the core complex belt itself.
In late Cretaceous time (Fig. 4A), a well-established magmatic arc terrane extended from Canada southward through the Idaho-Sierra Nevada-Peninsula batholiths into western Mexico (Coney, 1976, 1978; Armstrong, 1974). After 80 m.y., this Laramide magmatic activity swept rapidly eastward (Coney and Reynolds, 1977) across the southern Cordillera, then extinguished north of Arizona except for very scattered activity (Fig. 4B). In Arizona-New Mexico (Coney and Reynolds, 1977) and in Mexico (Clark and others, 1978), the eastward sweep did not extinguish and reached nearly 800 km inboard by Eocene time (Fig. 4C). At about the same time, the entire Pacific northwest erupted violently with Challis-Absaroka volcanism and shallow plutonism (Armstrong, 1974). This started a rapid return sweep of magmatic activity which eventually swept back toward the continental margin, becoming the vast ignimbrite flare-up (Fig. 4D) across Idaho-Washington-Oregon, Utah-Nevada, New Mexico-Arizona, and all of central and northern Mexico (Coney and Reynolds, 1977). Only the Colorado Plateau was spared. By 15-20 m.y. ago, the outburst reached the coast, formed the Cascade magmatic arc trend in the Pacific Northwest, but transformed to a widespread bi-modal basalt-rhyolite phase associated with Basin and Range rifting eastward and southward (Lipman and others, 1971) (Fig. 4E). There is considerable evidence, at least in the United States, that the core complexes developed during this massive retrograde sweep of magmatic activity between 55 and 15 m.y. Furthermore, some evidence suggests that the complexes north of the Snake River Plain evolved during the Eocene Challis-Absaroka outburst (Reynolds, pers. comm., 1977; Cheney, 1980); likewise, those south of the Snake River Plain evolved during the Oligocene-early Miocene ignimbrite flare-up there (Coney, 1978).

The post-Laramide to middle Tertiary is a most puzzling time (Coney, 1978). The events which took place are clearly post-Laramide compressive deformation and they seem to have begun by widespread erosion and beveling of Laramide landscapes (Epis and Chapin, 1975). Within the belt of core complexes, the relationship to the ignimbrites is often dramatic. In southern Arizona, volcanic ranges adjacent to the core complexes can be made up of vast ignimbrite sheets whose radiometric ages are essentially the same as cooling ages in metamorphic rocks within the core complex. Just why the core complexes should be restricted to such a narrow belt in this widespread panorama of ignimbrite eruption is not obvious.

Relationship to Late Cenozoic Tectonic Trends

One relationship on which almost all workers are agreed is that post 15-18 m.y. Basin and Range faulting seems to post-date most core complex activity (Fig. 4E). In many areas, the steep
Figure 1-4

Major features of Cordilleran tectonic evolution since Early Cretaceous time. A-Cretaceous; B-Laramide; C-Eocene; D-Middle Tertiary; E-Late Tertiary. In all figures heavy stipple-magmatic arc terranes; heavy barbed line – subduction zones; light barbed lines – thrust faults. Core complexes in solid black on figures C and D. Heavy black lines west of North America are approximate positions of Pacific spreading centers at times indicated from Coney (1978).
Middle Tertiary
Basin and Range
block faulting clearly cuts metamorphic rocks, the décollement, or cover rocks (Coney, 1974; Eberly and Stanley, 1978).

TECTORIC SIGNIFICANCE

The tectonic significance of Cordilleran metamorphic core complexes has been much debated during the past 20 years. Before their significance can be fully understood it is necessary to recognize two distinct, and perhaps largely unrelated, aspects of their history. The first aspect is the earlier, mostly Mesozoic, history experienced by many of the complexes, particularly those from western Arizona northward. The second aspect is the Early to Middle Tertiary history which is emerging as so important in all of the complexes, at least from northeastern Washington southward into Mexico. In my view, much of the confusion surrounding the complexes has been due to a lack of full appreciation of these younger Tertiary features and assignment of these features to results of events of Mesozoic age. In this regard, the full significance of the Arizona-Sonora examples emerged. This happened because they are not in the tectonic setting of the hinterland behind the Mesozoic thrust belts which are so characteristic of the setting of those complexes to the north. The fact that the Arizona-Sonora complexes so remarkably resembled the younger aspects of the northern complexes lent support to the growing recognition of the importance of the Tertiary events throughout the belt.

In other words, there is considerable evidence that much of the metamorphism, deformation and thermal disturbance so characteristic of Cordilleran metamorphic core complexes is Early to Middle Tertiary in age (Fig. 5). In many complexes, particularly those northward from Arizona, these mainly Tertiary features were superimposed on mainly Mesozoic metamorphic and deformational effects of the thrust belt hinterland but much, if not all, of the characteristic "cataclastic" or mylonitic fabrics, décollement zones, and chaotic structures in cover rocks are, in part at least, the result of Early to Middle Tertiary tectonics.

In the Shuswap complex of southern Canada something on the order of 40,000 km$^2$ of metamorphic rock are exposed. It is the largest of all the Cordilleran metamorphic core complexes and in the original conception of the problem was the type example and the source of the name. As outlined earlier, studies indicate some of the metamorphism is as old as Paleozoic and much of it is at least as old as Jurassic (Okulitch and others, 1975; Reed and Okulitch, 1977; Brown and Tippett, 1978; Wheeler and Gabrielse, 1972; Hyndman, 1968). There are dated Upper Jurassic to Middle Cretaceous plutons which cross-cut metamorphic rocks (Gabrielse and Reesor, 1974). There are also many Early Tertiary
(mostly Eocene) K-Ar apparent ages associated with scattered shallow plutons and a widespread resetting of isotopic clocks (Fox and others, 1977). Unfortunately, the age of the apparent later arching seen in the three distinctive "gneiss domes," and the narrow belt of late "cataclasism" along the eastern margin of the complex is not precisely known. The "cataclasism" is, however, apparently the youngest of the metamorphic fabrics (Reesor, 1970). In most ways, the late domes and the zone of east-dipping "cataclasism" most resemble those features characteristic of the complexes to the south considered here to be mainly Tertiary in age.

As a result of field work in northeastern Washington carried out in connection with this project, and building on earlier work by Reynolds, we have demonstrated that the Purcell trench, or fault zone, is the eastern margin of the Selkirk metamorphic core complex. This zone is a 30° east-dipping mylonitic front and is very similar, although off-set to the east, to the described east-dipping mylonitic zone along the east side of the Shuswap complex. Southward the Purcell mylonitic front terminates near Lake Coeur d'Alene and is apparently complexly transformed into the right-lateral Lewis and Clark fault zone which extends east-southeast to the north end of the east-dipping, north-trending Bitterroot mylonite zone in Idaho-Montana. The evidence in Montana and Idaho suggests the system is in part Early Tertiary in age and part of the core complex process.

In any event, the conclusion reached by some workers that much of the metamorphism in the Shuswap terrane is at least as old as Jurassic is of considerable importance for the Price and Mountjoy model relating the metamorphic hinterland to the Rocky Mountain thrust belt to the east. This model invokes a mobile infrastructure which buoyantly rose and propelled the thrusts to the east largely by gravity. This model has difficulties because the assignment of a Jurassic age to the metamorphism implies that most of the metamorphism was over before the thrusting began. Since the metamorphic hinterland exposes rocks once deeply buried and subjected to temperatures of 600°C and pressures approaching 4 kb, a post-metamorphic uplift of 10 kms or more is demanded. An uplift of this magnitude, particularly over a region as large as the Shuswap complex, is not easy to explain. I have argued elsewhere (Coney, 1979) that this massive uplift is perhaps best explained as being due to crustal telescoping in the thrust belt and resulting crustal thickening in the region of the hinterland during Sevier-Laramide time. To what degree Tertiary features similar to those found in the complexes southward in United States and northwestern Mexico have been superimposed on the results of mainly Mesozoic events described above is not yet known. It seems, however, that much of the gross metamorphic and structural character of the Canadian Shuswap core complex has an origin in
Figure 1-5

Age relationships in selected Cordilleran metamorphic core complexes. Column A-Northeastern Washington (Selkirk complex); B-Idaho Batholith (Bitterroot complex); C-Albion-Raft River-Grouse Creek complex; D-Western Arizona (Harcuvar complex); E-South Mountain complex south of Phoenix, Arizona; F-Catalina-Rincon complex near Tucson, Arizona; G-Sonora, Mexico. In all columns stipple pattern is for age range of plutonic rocks with dash pattern at top signifying mylonitic gneisses ("cataclastic") fabric superimposed on the pluton; V-pattern is age range of Early to Middle Tertiary volcanic rocks in vicinity of each complex; the vertical heavy arrow represents the possible age limits for formation of the overprinted mylonitic gneissic ("cataclastic") fabrics superimposed on the plutons and other basement terranes. The sub-horizontal heavy barbed arrow is the approximate older age limit on the movement of cover rocks over the basement terrane on "décollement" surfaces. In many cases this movement could be younger than indicated, but in most cases it can be shown to be older than latest Tertiary Basin and Range faulting. Vertical light stipple bands in each column are approximate durations of major compressional Sevier and/or Laramide deformation in thrust belts east of, or in vicinity of, respective core complexes.

Numbered plutons as follows: 1-Silver Point Quartz monzonite (Miller, 1972); 2- Mid Cretaceous plutons (Miller and Engels, 1975); 3-Eocene plutons (Reynolds and Rehrig, pers. comm., 1979); 4-Bitterroot lobe of Idaho Batholith (Chase and others, 1978); 5-Red Butte pluton (Compton and others, 1977); 6-Imigrant Pass pluton (Compton and others, 1977); 7-muscovite granite (Rehrig and Reynolds, 1980); 8-Tank Pass pluton (Rehrig and Reynolds, 1980); 9-South Mountain pluton (Reynolds and Rehrig, 1980); 10-Catalina granite (Shakel and others, 1977); 11-Wilderness granite (Shakel and others, 1977; Keith and others, 1980); 12-Sierra Mazatan, (Anderson and others, 1980).
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**Locations:**
- NE Wash.
- Raft R.
- W. Ariz.
- S. Mtn.
- Cat.-Rin.
- Sonora

**Time Periods:**
- Jurassic
- Cretaceous
- Paleocene
- Eocene
- Oligocene
- Miocene
- Pliocene
earlier Cordilleran tectonic history. The later Early Tertiary history is still not fully documented or evaluated.

A similar model of crustal telescoping and uplift of the hinterland can be applied southward in United States at least as far as southern Nevada, but telescoping and resulting uplift was probably less and mostly confined to mid-Cretaceous Sevier orogeny time (Armstrong, 1968a). The age of the older core complex metamorphism is not well-controlled, it has traditionally been described as simply "mid-Mesozoic" (Misch, 1960; Armstrong and Hansen, 1966; Armstrong, 1968b; Howard, 1971, 1980; Snoke, 1980), but may be as old as Jurassic (Compton et al., 1979). Superimposed on, or at least late in the history of, this earlier metamorphism are the shallow-dipping late mylonitic fabrics, associated décollements and related deformation of unmetamorphosed cover rocks. These late features appear to be, in part at least, superimposed on plutons as young as Early to Middle Tertiary age in Raft River-Grouse Creek Mountains, Kern Mountains, and Ruby Mountains (Compton and others, 1977; Best and others, 1974; Snoke, 1980; Todd, 1980), and similar relationships are emerging in north-eastern Washington (Cheney, 1980; Miller, 1972; Reynolds and Rehrig, pers. comm., 1978) and in the Idaho batholith (Chase and others, 1978). These events have reset K-Ar apparent ages to as young as 18 m.y. along the Snake Range décollement (Lee and others, 1970; Armstrong, 1972). They have produced field relationships such as "klippen" of Paleozoic rocks over Middle to Late Tertiary sediments in Raft River-Grouse Creek Mountains (Todd, 1973, 1980) and low-angle normal faults in mid-Tertiary volcanic rocks (Armstrong, 1972). Coney (1974) inferred that minor structures in rocks associated with the Snake Range décollement were consistent with Armstrong's (1972) model of Tertiary denudational faulting in the Nevada hinterland. All of these features are clearly pre-Late Tertiary Basin and Range faulting (Coney, 1974).

In Arizona and Sonora, the metamorphic core complexes are clearly Early to Middle Tertiary in age. In southwestern Arizona an older Mesozoic metamorphism is overprinted by mylonitic gneissic fabrics in several complexes (Rehrig and Reynolds, 1977, 1980), but for the most part the complexes of Arizona and Sonora are not complicated by widespread and complex pre-Tertiary metamorphism and deformation like those complexes to the north. More important, as already mentioned, the complexes here are not in a hinterland behind a thrust belt of any age but, instead, are in part in the the midst of a belt of deep-seated basement cored thrust uplifts of Laramide age accompanied by scattered Laramide plutons. Furthermore, the typical core complex fabrics and related décollement are superimposed on plutons ranging in age from about 55 m.y. in Sonora (Anderson and others, 1977, 1980) to as young as 25 m.y. at South Mountain near Phoenix (Reynolds and Rehrig, 1980). In the Catalina-Rincon complex near Tucson, a complicated history of
plutonism, deformation and metamorphism is recorded, but most workers now agree that features typical of the core complexes throughout the Cordillera are superimposed on plutonic rock as young as at least 50 m.y. (Shakel and others, 1977; Keith and others, 1980). Davis (1975) inferred Tertiary movement of cover rocks down décollement surfaces in the Rincon Mountains based on structural analysis. Finally, all these features are clearly pre-Basin and Range block faulting.

It cannot be denied that the later, mainly Tertiary, overprint so characteristic of Cordilleran metamorphic core complexes was perhaps influenced by the preceding Mesozoic history. This is particularly true of the complexes in Nevada and northward into southern Canada. Exactly what the influence was, however, has been difficult to identify. Low-angle faults, some certainly of younger on older type, undoubtedly formed in the hinterland during Mesozoic time. They may have served to localize the Tertiary décollements. The earlier Mesozoic metamorphism and deformation has already been mentioned. Perhaps one important influence was the Mesozoic uplift of the hinterland produced from crustal thickening behind the telescoping thrust belts to the east. This would permit the later thermal culminations and associated plutons of Early to Middle Tertiary age to likewise rise higher before being frozen in a reactive endothermic Phanerozoic cover. In any event, the near perfect tracking of the core complexes along the hinterland behind the thrust belt can hardly be a fortuitous accident. Even in Arizona and Sonora where the above relationships did not hold, the structurally and thermally battered ground inherited from particularly Laramide events could have prepared a weakened basement conducive to concentrating the Tertiary events.

The evidence for large-scale Tertiary extension in and adjacent to the complexes is extremely compelling. It was first acknowledged in the cover terranes (Armstrong, 1972; Anderson, 1971; Coney, 1974; Davis, 1973, 1975; see also Davis and others, 1980). Only more recently has it been proposed in the basement itself as manifested in the mylonitic foliation and lineation. It was certainly the work of George Davis (1975, 1973, 1977a, 1977b) in the Catalina-Rincon complex in Arizona that first suggested this basement extension as a major aspect of core complex evolution. Davis has subsequently (1977, 1980; Davis and Coney, 1979) expanded these observations into the provocative concept that the complexes are in fact mega-boudins. This concept is similar to recent interpretations of the Death Valley turtlebacks (Wright and others, 1974; Burchfiel and Stewart, 1966). Whatever they are, the evidence is that they were produced by extension and tectonic denudation. This is certainly the case with the cover terranes. Another early suggestion for regional extension was recognition of northwest-trending dikes cutting mid-Tertiary
plutons in southern Arizona (Rehrig and Heidrick, 1976). These dikes are oriented perpendicular to the lineation in the core complexes (Reynolds and Rehrig, 1980). This extension has also been recognized outside or adjacent to the main core complex belt in areas of low-angle pre-Basin and Range listric normal faults which cut Middle Tertiary volcanic and sedimentary rocks (Anderson, 1971; Eberly and Stanley, 1978; Rehrig and Heidrick, 1976; Davis and others, 1980).

It is remarkable how many of the complexes are characterized by Early to Middle Tertiary granitic plutons. The late cataclastic fabrics are superimposed on many of these plutons and evidence suggests that they were cooling and still partly mobile at the time of at least the earlier phases of the deformation recorded in the complexes. The granitic plutons precede, then become intimately associated with a massive thermal disturbance which is mainly of Eocene age (Fig. 4C) in the Pacific Northwest (Armstrong, 1974) and of Late Eocene to Middle Miocene age (Fig. 4D) in the south (Lipman and others, 1971). This vast ignimbrite flare-up has been interpreted as the result of the collapse and retrograde sweep of a flattened Laramide Benioff zone during Eocene through Miocene time (Coney and Reynolds, 1977). A significant enough number of granitic plutons, particularly the earlier ones, are of the two-mica garnet-bearing type to suggest a genetic association between this type and the evolution of the complexes. Perhaps the earlier two-mica plutons were generated during the late-Laramide period of maximum flattening of the Benioff zone when Farallon lithosphere was essentially plated beneath North American lithosphere.

Since 1970, the situation surrounding the core complexes has become one of considerable controversy and heated debate. In the process, some discoveries have been made and there have been some new insights into Cordilleran tectonics. Considerable confusion yet prevails and there is still substantial, justifiable disagreement. The most important aspect of the recent work is certainly the suspicion and realization that there had been a strongly manifested and enigmatic Tertiary overprint superimposed on the region. The overprint is most clearly recognized in young K-Ar apparent age dates from basement rocks and in the listric normal faulting and décollement zones characteristic of the unmetamorphosed cover terranes of the complexes. This activity followed Mesozoic thrusting and preceded Basin and Range block faulting, and was more or less contemporaneous with widespread Tertiary ignimbrite eruptions. This created a strange image of tectonic response not easily understood then or even today.

In retrospect, the following developments proved significant:

1) Publication of Armstrong's (1972) views on Tertiary denudational
faulting in the Nevada hinterland and subsequent tests which supported his views (Coney, 1974; Todd, 1973), 2) Discovery of the large number of, and great extent of, metamorphic complexes in Arizona and recognition of their Tertiary age (Davis, 1973). Also important was the realization that these metamorphic culminations in Arizona were not in a hinterland behind a thrust belt of any age. This cast suspicions on the presumed genetic link between the northern complexes and the thrust belts to the east, 3) Finally, the realization that most of the known complexes throughout the Cordillera had very similar lithologies and structures suggested a common origin.

It is quite remarkable how serendipitously the present stage of knowledge evolved. It is also instructive to note how scattered early work, which were seeds of what was to come, went largely unnoticed and have only recently flowered to significance and appreciation, albeit still not completely understood. Such work includes Anderson's (1971) study of Tertiary listric normal faulting south of Lake Mead and a number of early studies in Arizona and in the Nevada hinterland (for example see Damon and others, 1963; Drewes, 1964; Moores and others, 1968).

It will be obvious to all readers that considerable controversy still exists and there are major problems surrounding the complexes which need to still be resolved. This is only normal for a new and complicated large-scale tectonic feature. At the same time, this situation adds to the excitement and clearly indicates need for continued study. A number of novel and somewhat contrasting models have been recently proposed to explain the new data surrounding the complexes and there are certainly more to follow. This is also only natural and need not be taken too seriously at this time.

The major contribution of all the recent work is to signal an anomalous petro-tectonic assemblage, now recognized from southern Canada to northwestern Mexico, which has an almost certain Early to Middle Tertiary evolutionary history. The evidence seems to indicate that a significant part of this assemblage was produced by extensional tectonics and that, whatever the process was, it post-dates Sevier-Laramide compressional thrusting and pre-dates Late Tertiary Basin and Range extensional block faulting. It has clearly been superimposed on and confused with earlier, mainly Mesozoic tectonic events.

The most important controversy still remaining is the origin and significance of the mylonitic gneiss fabrics so characteristic of the basement cores of the complexes. Without question, the dramatic resemblance of these fabrics to those produced along deep-seated thrust faults in other parts of the world is an obstacle to what might be called a new consensus. Some workers
have concluded that even this aspect of the complexes was produced by Tertiary regional extension, but other workers conclude that the metamorphic basement and all its fabrics are significantly older than the upper-plate dislocations and listric normal faulting and genetically unrelated to them. As has been mentioned earlier in this chapter, many prefer to relate the metamorphism and the mylonitic fabrics to earlier periods of major Sevier-Laramide compressional tectonics and would even argue that most of the décollement and the upper-plate structures date from these earlier deformations as well.

As of this writing the main interpretations of the origin and significance of Cordilleran metamorphic core complexes can be summarized as follows:

1. The complexes are mainly Jurassic to Sevier-Laramide (mostly Mesozoic) in age. The metamorphic basement terranes are a deep-seated infrastructure (gneiss domes) and the décollements are simply bedding-plane thrust faults both of which are genetically related to Sevier-Laramide foreland thrust faulting. This is the classic interpretation of the complexes and of the Cordilleran hinterland. Most who take this view acknowledge minor later Tertiary disruption and local gravity slides along the earlier fault planes, but prefer to minimize their importance. This position is perhaps best represented by the early work of Misch (1960), Hose and Danes (1973), and Roberts and Crittenden (1973).

2. The complexes are mainly Early to Middle Tertiary in age. In this view both the mylonitic fabrics and late metamorphism and the décollements and listric normal faulting are genetically related to Early-Middle Tertiary mainly extensional tectonics and post-date Sevier-Laramide compressional tectonics. Those who take this position acknowledge an older metamorphic and structural history recorded in most of the complexes as certainly of mainly Mesozoic origin, but feel it is significantly overprinted by the Tertiary events. This is the new position that began to emerge after 1970 and has been particularly emphasized in the publications of Davis and Coney (1979), Coney (1979) and Rehrig and Reynolds (1980).

3. The complexes are truly hybrid and of both Sevier-Laramide and Early-Middle Tertiary age. In this view all the basement fabrics including the later mylonitic fabrics are of compressional origin, such as deep-seated thrust faults, and mainly related to Sevier-Laramide events, or perhaps to some late-Laramide "compressional" event. The décollements and listric normal faulting in the cover terrane are acknowledged as Tertiary, but genetically unrelated to the earlier basement terrane. In this view the co-linearity
of the basement lineation and its interpreted movement picture and the movement picture derived from upper plate structures is purely accidental. This position has been particularly emphasized most recently by Davis and others (1980).

As has been mentioned earlier most are agreed that most of the dislocation and décollement, particularly the listric normal faulting in cover rocks, clearly brittlely disrupt the more ductile mylonitic basement fabrics (Coney, 1974). Where the upper plate structures include Tertiary rocks all will have to agree that some of this late brittle deformation is Tertiary in age. It is to be emphasized that this relationship of involvement of Tertiary rocks has been demonstrated in the majority of the complexes. The main dilemma is whether the earlier mylonitic basement fabrics, in spite of being clearly older, are genetically related or totally unrelated to the later brittle Tertiary events.

Those who choose to genetically relate the two terranes emphasize the co-linearity in kinematics of the two terranes and also emphasize the fact that the mylonitic fabrics are superimposed on young rocks, particularly the Tertiary plutons emplaced in the basement terrane. As has been discussed earlier, many of these plutons are rather well dated as early Tertiary and some appear to be as young as middle Tertiary. Those who choose to not genetically relate the two terranes emphasize the age differences and contrast in mechanical behavior. The co-linearity in kinematics is thus a fortuitous accident.

Perhaps the trap is in what might be called argument based on classic geologic reasoning from age data. If all the mylonitic basement is mainly as old as some of the older Tertiary plutons affected and if all of the brittle deformation is younger than the youngest cover rocks, we are seemingly dealing with two separate and presumably unrelated events. If the mylonitic fabrics are as old and as young as the age-range in affected Tertiary plutons and if the listric normal faulting is a growth faulting and as old and as young as the age-range in Tertiary sediments and volcanics affected, the seemingly separate events could over-lap and merge into a continuum of deformation ranging from Eocene to Late Miocene in age.

This is not the place to attempt to resolve the genetic dilemma surrounding Cordilleran metamorphic core complexes. Our main concern in this report is to present the distribution, character, and geologic-tectonic setting and history of these enigmatic and newly recognized features in the North American Cordillera. Hopefully this will provide a base upon which the
assessment of their economic as well as scientific potential can be carried out. There is clearly much work yet to do. It is particularly obvious that much more work needs to be done on the petrology, structure, and geochronology of the basement terranes of the complexes. There is also much need for detailed sedimentologic and stratigraphic studies in the Tertiary sediments and volcanics of the cover terrane to better document its role in the core complex process. Similarly, the plutonic manifestation of the complexes warrants much detailed petrologic and geochemical study. This is particularly true of the so-called "two-mica" granites which are so common in the core complex belt.

If the evidence for extension is recognized and acknowledged, the belt of complexes takes on the character of an irregular, elongate and sinuous large-scale pull-apart terrane extending the length of the Cordillera at least from southern Canada into northwestern Mexico. The only large-scale phenomenon it seems to be associated with is the region that lies above the first flattened, low-dipping, Laramide Benioff zone which then steepened and collapsed during Early to Middle Tertiary time. The entire metamorphic core complex process, as either separate distinct phases or as a continuum, probably endured 30 to 40 m.y. between about 55 m.y. and 15 m.y. in Early to Middle Tertiary time.

Just why this apparent extension on a regional scale occurred when it did is not entirely clear. It certainly began during Farallon-North America plate convergence at least 10 to 30 m.y. before even initial contact between Pacific and North America plates and resulting growth of the San Andreas-Basin and Range transform-transpressive rifting (Fig. 4). It did not take place behind a magmatic arc; it took place within a magmatic arc. But this magmatic arc was a very special one, perhaps the result of extreme flattening of Laramide Benioff zones followed by a massive collapse and resulting retrograde sweep of arc activity across the Cordillera from the Pacific Northwest southward into Mexico. The thermal upwelling and instability implied in such a model gives at least an intuitive rationale for all that transpired.
CHAPTER 2

MANTLED GNEISS DOMES

by

Paula F. Trever

PART I: A REVIEW OF THE LITERATURE

INTRODUCTION

The recognition of metamorphic rocks in the hinterland of the North American Cordillera was accompanied by a renewed interest in the classic concepts of orogenic development. Did these rocks represent a classic "metamorphic core," a zone in which formerly mobile orogenic infrastructure was raised to view (Armstrong and Hansen, 1966; Price and Mountjoy, 1970)? As the model of the Cordilleran "metamorphic core complex" developed, infrastructural imagery was superseded by an emphasis on a superimposed Tertiary mylonitic-cataclastic effect, unrelated to earlier orogenesis. The model, as presently expounded (Davis and Coney, 1979), does not emphasize the conclusions of local studies which indicate that mobile behavior was necessary for the structural development of some of the complexes (McMillan, 1973; Reesor and Moore, 1971; Fox and others, 1977; Armstrong, 1968; Wagg, 1968).

The concept of mobilization, somewhat foreign to Cordilleran geologists, has been reviewed by Watson (1967), who noted the contributions of Sederholm (1926), Wegmann (1935), and Eskola (1949). The terminology of Wegman is familiar to those who are acquainted with the later work of Haller (1955). However, it is the work of Eskola, with his formulation of the mantled gneiss dome concept, that is best known to North American geologists and has the most frequently been applied to the metamorphic terranes of the Cordillera. This chapter will provide a basis on which to assess such usage.
Eskola described from the Karelide (early Proterozoic) zone of East Finland gneissic domes overlain by sedimentary strata in which the layering was parallel to both the dome contacts and the foliation of the gneiss. In some of the domes, the basal layer of the mantle was a conglomerate which contained boulders of the underlying gneiss; although in other domes, quartzite or dolomite formed the basal layer. According to Eskola (p. 461):

In some domes the gneiss, or rather granite, has apparently been preserved as it was when the sediments were deposited upon the eroded surface of the plutonic mass. In most cases, however, it has become migmatized and granitized during the doming, and shows a veined structure and has a potash-rich ideal-granitic composition, although its original composition may have been granodioritic or quartz dioritic. In some cases massive granites break through the domes, and at the contacts the paligenic gneissose granite may display an intrusive relation to the mantle rocks.

By creating the mantled gneiss dome model, Eskola reconciled two contradictory bodies of geologic evidence. The position of the gneiss-sediment contact within a given domal complex at a fairly constant stratigraphic horizon, and the occurrence of basal conglomerate containing gneissic cobbles, supported the conclusion that the gneiss was basement and that the gneiss-mantle contact was an unconformity. However, banding and foliation in the outer part of the gneiss were nearly everywhere parallel to the base of the mantling strata, a relationship unlikely to result from deposition above an unconformity. The structural concordance, along with the presence of marginal facies of gneiss which invaded the mantle as dikes, suggested that the gneiss had an intrusive origin.

Eskola resolved the geologic paradox by proposing a polyorogenic history for the gneiss (p. 461):

The mantled domes apparently represent earlier granitic intrusions related to an orogenic period. The plutonic mass was later eroded and beveled, and thereafter followed a period of sedimentation. During a subsequent orogenic cycle the pluton was mobilized anew and new granitic magma was injected into the plutonic rock at the same time as it was deformed into gneiss, causing its migmatization and granitization, or paligenesis.
The polyepisodic history of the Karelide basement has been confirmed by isotopic studies (Wetherill and others, 1962; Kouvo and Tilton, 1966), which have yielded an array of discordant mineral ages. The age data were interpreted by Kouvo and Tilton to indicate crystallization of the basement complex 2800-2600 m.y. ago, followed by basement reactivation and metamorphism of the overlying sedimentary column 1900-1800 m.y. ago.

In schematic cross-section (Fig. 2-1), Eskola centered each mantled gneiss dome above an ancient pluton. He did not portray an extensive gneissic basement, because he could not imagine that such a basement "in the loci of the present domes had something in it that made it well up and caused the granitic materials to collect in domes" (p. 468). However, later experimental studies (i.e. Ramberg, 1967a) have suggested that a low density source layer will respond to its gravitational instability by forming a number of discrete domes with a characteristic spacing, subject to experimental parameters, between domes. Thus the restriction of the term "mantled gneiss dome" to cases in which each dome is centered on an intrusive granite, as maintained by Nicholson (1965, p. 161-162), seems unnecessary.

Eskola did not specify that the mantling rocks be metamorphosed, although in subsequent reports of structures which conform well to the mantled gneiss dome model, a metamorphic mantle has emerged as a universal characteristic. The mantle may, however, contain metavolcanic as well as metasedimentary rocks (see Johnson, 1968). Eskola's original model has also been enlarged to include domes in which paragneiss, rather than orthogneiss, forms the core. The Baltimore gneiss, found in Maryland, which Eskola considered to be "surprisingly similar" (p. 470) to Karelide gneisses, is now considered to be largely metasedimentary in origin (Hopson, 1964).

MECHANISMS

According to Eskola (p. 475-476):

In most cases the upheaval of the domes is accompanied by granitization\(^1\) of older granitic or dioritic intrusions, and it seems that the rising granitic magma, as a rule of ideal-granitic, potash-rich composition, has supplied the elevating power. What, then, gives the granitic magma its power to move upwards and to lift its cover? And what is the explanation of the universal concentration of granitic magma in the orogenic zones? The only answer I can find to the first question is the lesser density of the granitic magma as compared with the average crystalline rocks.

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\(^1\)Eskola regarded granitization as a process impossible without the presence of a granitic magma.
Figure 2-1
The history of a mantled gneiss dome, as diagrammed by Eskola in his classic paper (1949, p. 469).
Thus, in Eskola's mind, mantled gneiss domes were formed by the buoyant rise of a granitic melt. He in no way envisioned the completely solid-state emplacement modeled by Ramberg (1972) or Fletcher (1972).

The term "diapir" (Greek: to pierce) was first used by Mrazec (1910) to describe anticlinal folds in the Carpathians in which salt had pierced the hinge of the anticline, cutting younger strata. Wegmann (1930) subsequently applied the term to granitic rocks, which like the salt, were found piercing an anticlinal arch. In recent literature, the term "diapirism" often refers to the buoyant phenomenon which in some instances may result in piercement, rather than to piercement itself.

Thus, Eskola and most subsequent workers have regarded the emplacement of mantled gneiss domes as a consequence of some sort of "diapirism," driven by a density inversion; whether the less dense phase was in the solid or liquid state is debated, however. Those who propose solid-state diapirism must demonstrate a density contrast between core and mantle rocks as they are presently found. In this regard, Fletcher (1972) has argued that metamorphism of the mantling strata is not incidental to, but rather is necessary for, dome formation. In the Rum Jungle area of northern Australia, Stephanson and Johnson (1976) described diapiric granites (density 2.67 g/cm³) that pierced a metasedimentary complex of mean density 2.77 g/cm³. Overlying, unmetamorphosed sediments had a mean density of 2.55 g/cm³. In a similar fashion, Johnson (1968) considered the room temperature density contrast between granite (specific gravity 2.6-2.7 g/cm³) and amphibolite (s.g. 3.0-3.1 g/cm³) alone to be a sufficient mechanism for the formation of mantled gneiss domes in the Mozambique Belt of Rhodesia. He noted that such domes were absent where the granite was not overlain by the dense amphibolite. Experimentally, Ramberg (1967a) has demonstrated that if the viscosity contrast between lighter and denser layers exceeds $10^5-10^3$ poises, spindly structures resembling salt diapirs form instead of bulky domes. Since the viscosity contrast between solid rocks and granitic melt is much greater than this figure—on the order of $10^{14}$ poises—Ramberg argues for solid-state emplacement.

In addition to a dense metamorphic mantle, a considerable time span is necessary for solid-state diapirism. Stephansson and Johnson (1976) estimate that solid-state intrusion of granite could be accomplished in 10-100 m.y. By Fletcher's calculations, the viscosity of both core and mantle rocks must be reduced by at least a thousand fold by orogenic heat flux before solid-state
diapirism can be effective over geologic time (for instance, the growth of a dome 5 km in amplitude in 30 m.y.).

Even in a column of material which originally has a uniform specific gravity, density instability will occur if heat from a basal source is not distributed upward quickly enough by conduction; convective motion may thereby be induced. It has been suggested that orogenic thermal gradients may permit subsolidus convection in the crust, and that this process may be responsible for mobilization, and ultimately homogenization, of gneissic domes (Talbot, 1971; denTex, 1975). Continental crust is considered in these models to behave as a viscous Newtonian fluid. Convection will occur when a critical value of the dimensionless Rayleigh number, defined by

\[
R = \frac{g_0 \alpha_0 \beta d^4}{\gamma \nu}
\]

is exceeded (Chandrasekhar, 1961, p.10). From the above equation, it is seen that instability of the layer in question may be caused either by increasing its thickness or by lowering its viscosity; denTex (p. 71) suggests that tectonic thickening may be a prerequisite to lower crustal convection.

The most extensive experimental modeling of gravitational instability on the geologic scale has been performed by Ramberg (1963, 1967a, 1967b), with the aid of a centrifuge to intensify gravitational effects. From Ramberg's studies, the archetypal diapir emerges as mushroom-shaped (Fig. 2-2). The centrifuged models are produced without redistribution at the upper surface, in contrast to the theoretical model of Fletcher (1972), which assumes that erosion and redeposition keep the upper surface perfectly plane. A mathematical model invoking redistribution produces a simple dome without a "mushroom head" (Fletcher, p. 208). Another modification of the Ramberg model has been suggested by Talbot (1974), who predicts that if the original interface between light and dense layers is inclined rather than horizontal, an asymmetric, nappe-like structure will be produced in preference to the symmetrical mushroom structure (Fig. 2-3).

Dixon (1975) conducted similar experimental studies, but was able to quantitatively monitor finite strain by a construction method equivalent to building the model of initially square 1-mm elements. Dixon found that in the mantling rocks, the greatest strain was at all times a horizontal stretching above the rising
Figure 2-2. Geometric features typical of domes produced in the centrifuged models of Ramberg (1966, 1967, 1970). Figure depicts the marginal sink or rim syncline (arrows), the trunk region (T), and the hat (H) of the dome (from Ramberg, 1970).

Figure 2-3. Stages in the development of gravity structures. Structures were formed from originally horizontal (A) and inlined (B) interfaces between fluids with unstable density arrangements (Talbot, 1974).
diapir, averaging 4:1 at first, but reaching a value as high as 60:1 (Fig. 2-4); such an observation is compatible with the purely mathematical result obtained by Fletcher (1972, p. 206).

Ramsay (1967) presents two alternatives to gravitational instability for the mechanism of mantled gneiss dome formation. First he suggests that such domes may result from compressive strain "acting in all directions," causing the basement-cover contact, because of the viscosity contrast across it (basement assumed to have higher viscosity), to be deformed into a series of pinched synclines and more gentle anticlines, analogous to smaller-scale mullion structures. Secondly, he proposes (p. 521-524) that mantled gneiss dome terrains may represent areas of superimposed folding characterized by a Type 1 interference pattern (egg carton pattern). In either case, domes are interpreted as somewhat fortuitous by-products of one or more episodes of tangential compression rather than as sites of active vertical upwelling. Hobbs, Means, and Williams (1976, p. 430) suggest that compressive forces and processes of gravitational instability may operate synchronously. According to Kranck (1972, p. 18-19):

It must be emphasized that even if gravitational buoyancy is the principal cause of the rise of diapirs, there is often a close connection between diapirs and axial deformation. Evaporite diapirs may form along the crest of an anticline, as is the case on Axel Heiberg island and other regions (Kranck, 1963), and in particular, migmatite naps are generally formed by an interaction between tangential and gravitational forces.

SURVEY OF REGIONAL LITERATURE

This section will attempt to present a composite description of mantled gneiss domes, as compiled from local studies. In most instances, the structures included in the survey were explicitly attributed to the mantled gneiss dome model of Eskola by the local geologist; although in some cases, the structures were identified as mantled gneiss domes by authors with only a literary knowledge of the area in question. Russian geologists have developed an independent literature on gneiss domes (see Kalyayev, 1970, Pavlova, 1972, Salop, 1972); it is therefore uncertain whether structures termed "gneiss domes" by them in all cases conform to the model of Eskola, but a number are included in this review.

Geometry

Gneiss domes may be circular in plan, but are often elongated parallel to the tectonic grain of the region, so that their geometry is actually that of a doubly plunging antiform, or
Contours of the values of maximum principal extension in two model domes of different amplitudes, produced by Dixon (1975, p. 98, 101). Heavy stipple = values greater than 4.00.
"brachyanticline", to use a translated Russian term. Gneiss domes rarely occur singly, but are found in clusters or "herds", as Pavlosky (1970) described the many small domes of the Ukrainian Shield. In the Karelide zone described by Eskola there are approximately two dozen domes (Salop, 1972). Often the domes form a linear array, coinciding with a regional arch. According to Brun (1980), the Karelian mantled gneiss domes cluster along nine ridges, oriented NNE-SSW to NE-SW, and separated by a periodic spacing.

Fletcher (1972, p. 200) reported that preliminary measurements of a few tens of mantled gneiss domes in the Appalachians and Caledonides yielded an average nearest neighbor spacing between domes of 25 ± 5 km; this compares with a figure of 10 km, reported by Fletcher from the Gulf Coast salt dome field. ČenTex (1975) reported that 5 subdome structures within the Lepontine gneiss region of Switzerland were regularly spaced at approximately 25 km. Spacing between 10 domes in the Pyrenees is also roughly 25 km (Zwart, 1968). The domes of the Shuswap Complex of British Columbia occur at somewhat larger intervals of 40-50 km (Reesor, 1970). Gneiss domes range in diameter from several km to several tens of km; Salop (1972) reported that a different type of gneissic structure, termed "folded gneiss oval" by him, was typically much larger, ranging from 80 km to 600 km in diameter.

The essential geometry of a mantled gneiss dome consists of a polycyclic, crystalline core, and a stratified, metamorphosed mantle. In some structural studies, an outer zone of less-metamorphosed or less-deformed rocks ("fringe zone" of Reesor and Moore, 1971, "envelope" of Brun, 1977) may be distinguished from high-grade metamorphic rocks adjacent to the core. Core rocks may have either a sedimentary or igneous heritage, but a polyepisodic history should be demonstrable; isotopic studies may be useful in this regard. However, radiometric discordance, in the absence of field evidence for basement mobility following deposition of cover sequence, is insufficient to identify a mantled gneiss dome (see, for example, Lanphere and others, 1964).

Late-stage granitic intrusions which transgress the basement-cover unconformity are often leucocratic, and commonly display a two-mica composition (Gunpowder Granite, Hopson, 1964; Guilford dome stock, Skehan, 1961). Didier and Lameyre (1969) have suggested that such leucogranites may result from anatexis at minimum melting conditions, and it seems possible that these plutons within mantled gneiss domes represent in situ melting of the basement complex. The frequent occurrence of migmatites, attributed by Winkler (1974) to early-stage anatexis, in mantled gneiss domes indicates that P-T conditions close to those necessary for partial melting were attained in many dome cores.
Metamorphism

In many instances mantled gneiss domes coincide with thermal domes defined by metamorphic isograds. Maximum metamorphic grade varies from complex to complex, but is generally within the almandine-amphibolite facies. Sillimanite, which occurs in the high temperature sub-regions of this facies, is frequently—but not always—present.

The metamorphic assemblages of most mantled gneiss domes are suggestive of deep-seated metamorphism. Fletcher (p. 200) has suggested that 25-35 km is an appropriate thickness for mantle rocks prior to dome formation in New England. Similar figures have been proposed for the Baltimore gneiss domes (Hopson, 1964) and for the Caledonide domes of Norway (Ramberg, 1967a). denTex (1975) suggests a depth of burial for the Agout dome of France of 15-19 km, and for the Lepontine massif of Switzerland, 15-26 km.

Metamorphic grade often decreases rapidly outward from the core; isograds are frequently telescoped and the abnormally high geothermal gradients calculated from such isograd patterns (for instance, 200°-300°C/km, denTex, 1975, p. 64) are quite problematic. As denTex notes, the frequently used model in which radioactive decay is the principal heat source and conductivity the exclusive transfer mechanism fails in this instance because it predicts that the spacing of isograds will increase with depth, as the column of underlying, highly radioactive, granitic material is decreased. A number of alternative schemes have therefore been proposed to explain the juxtaposition of high-grade core rocks with low-grade mantle rocks over relatively short distances. Among them:

1) Low temperature "granitization" of the core.

2) Diachronous metamorphism. Brown (1978) iterates the viewpoint that the core of the St. Malo massif, France, was migmatized prior to the deposition of the mantling strata and that the low-grade metasediments of the Brioverian succession were later brought into contact with the high-grade core by faulting.

3) Heating by a subjacent magma body.

4) "The basement effect," Fontelles and Guitard (1968) have proposed "l'effet de socle"--the basement effect—in order to explain telescoped isograds around gneiss domes in the Pyrenees. Among other factors, this hypothesis considers that mantling strata undergoing metamorphism are subject to endothermic reactions involving
dehydration; "dry" basement is presumably immune to such reactions, and this contrast serves to steepen geothermal gradients in the vicinity of the core-mantle boundary.

5) Convective/conductive heat transfer. denTex (1975), following Talbot (1971), hypothesizes that the characteristic orientation of foliations in a mantled gneiss dome—near-horizontal in the upper portion and near-vertical in the trunk zone—may encourage a combination of convective and conductive heat transfer that concentrates heat in the upper portion of the dome.

6) Attenuation of mantling strata around a rising dome. Examination of the experimentally produced diagrams of Dixon (Fig. 2-4) suggests that mantling strata will be severely attenuated above a rising diapir. This feature of dome formation may be sufficient in itself to explain the telescoped isograds found around mantled gneiss domes.

Structure

Foliation in the core and mantle rocks of mantled gneiss domes generally dips outward from the culmination of the dome. Where the "rim syncline" of Ramberg is well developed, dips may be locally overturned. As in the structures described by Eskola, these foliations are most often concordant with each other and with the core-mantle contact, although local truncation of basement foliation is occasionally reported (Sims and Peterman, 1976). The zone of concordance is often quite small, however, and the central part of the core may be undeformed, as Eskola himself noted (p. 462), or it may exhibit an older, discordant foliation.

The mantling rocks above the crest of the dome are expected to lie in a domain of subequal extension in all directions, which may be reflected by horizontal foliation in the mantling strata, and polygonal boudinage of competent units (Fletcher, 1972, p. 209). Boudinage is indeed quite common in mantling rocks; Pavlova (1972) describes from the domes of west central Kazakhstan boudinage which has "developed on a background of folding." In some cases, the ductility contrast between different strata in the mantle may cause boudinage and lithologic differentiation on a scale larger than that of a single outcrop. Adjacent to the Baltimore gneiss domes, the incompetent Cockeysville Marble is frequently thinner and missing. According to Choquette (1960, p. 1032):

The reason for these pinch outs is uncertain, but they occur within such short distances that they may be structural rather than stratigraphic, mainly because of flowing induced
by differential movement over the gneiss domes. Direct
evidence that the carbonate rocks were extremely mobile lies
in the flowage and drag folds found at almost every outcrop
in the area.

The extensional environment is often dramatically demonstrat-
ed by deformed clasts within basal conglomerate units. In some
cases, the principal elongation of stretched pebbles is parallel
to a prominent mineral lineation in the dome (Sinitsa, 1965).

Mineral lineation may be fairly constant in orientation
within an individual dome or even from dome to dome, within a
regional cluster. Escher and Pulvertaft (1976) compared the con-
stancy of a biotite lineation in the mantling formation of domes
in West Greenland (Umanak area) to the wide dispersion of major
and minor fold axes. Brun (1980) noted a "surprisingly regular"
NE-SW trend of lineation in all the domes of the classic Karelide
region, and surmised that gravitationally-controlled deformation
had been accompanied by regional, compressive stress.

Reverse drag folds--"spruce tree folds"--are often reported
in the mantle rocks in the region of the dome flanks. They have
been described as "the most characteristic minor structure observed
in the mantling rocks of natural domes" by Fletcher (1972, p. 208),
who cites the work of Skehan (1961) on the gneiss domes of Vermont.
Reverse drag folds have also been described from domes in the East
Ural anticlinorium by Chesnokov (1966), who, like Skehan, attri-
butes their development to the predominance of vertically directed
forces over tangential compression during the diapiric rise of the
domes. Wheeler (1965, p. 19) attributed "fir-tree folds" in the
gneisses on the west side of Frenchman's Cap dome (Shuswap Complex)
to the "diapiric movement of the central part of the dome."

Escher and Pulvertaft have described (1976, p. 113), within
the strata which mantle West Greenland gneiss domes, refolded
"zig-zag" folds, "thought to be gravity-induced structures which
slid off the slopes created by the rising domes."

The most critical structural horizon in the mantled gneiss
dome is the core-mantle interface. As Watson (1967) points out,
the ancient unconformity provides an important datum plane of
known initial orientation (broadly horizontal) between rock
types of fundamentally different mechanical properties; thus the
present configuration of this surface provides insight into the
nature of "mobilization." Reesor and Moore (1971) state that the
core-mantle boundary of the Thor-Odin dome, British Columbia, was
first deformed into large folds thousands of feet in amplitude and
several miles in extent and then refolded by the diapiric rise of the core zone as a whole.

Although the ancient unconformity may demonstrate extremely ductile transformation, it is also frequently the locus of intense shearing and cataclasis. The foliation developed at the periphery of the core is regarded by several authors as cataclastic in origin. Johnson (1968, p. 250) noted that development of augen gneiss from the porphyritic granite of the Chirwa Intrusion, Rhodesia begins 500 ft from the outer contact of the granite: "At the edge of the granite, the phenocrysts become lenticular and, along thin but extensive foliae of biotite, the rocks become fissile." Biotite enrichment and well-developed gneissosity at the boundary of the core zone is also described from domes in the Mackenzie District by Frith and Leatherbarrow (1975). Mallick (1967) states that zones of intense shearing skirting the Mpande dome in Zambia appear to have a mineralogy similar to that of the core gneisses, but are strongly enriched in biotite. He describes a transition from massive core rocks to schistose gneisses, in which biotite clusters become drawn out, and quartz and plagioclase are reduced to groups of small equant grains. According to Sinitsa (1965, p. 60), in the Kuotmar dome of the Transbaikal, the core granite has experienced "marked alteration (cataclasis and mylonitization) in relatively narrow (2 to 4 km) zones around the dome, where Jursassic rocks are also foliated."

Most authors find the cataclastic margins of the domal complexes compatible with the waning stages of mobile behavior. Describing the classic Karelide zone, Salop (1972, p. 1220) states:

It should be noted that the presence of zones of blastomylonites at the contacts of crystalline basement and metamorphic beds in many domes of northern Ladoga caused some investigators (Sudovikov, 1954) to conclude that disruptive dislocations played a leading role in the formation of domes and the peculiar horst character of the structures. Our observations do not enable us to agree with this point of view, but are indicative of a generally plastic character of deformation at the dome formation. The presence of blastomylonites is most often related to locally poor mobilization of the material of the basement, and in some cases with subsequent tectonic shifts along contacts.

Several excellent studies of structural zonation within mantled gneiss domes have been published (Brun, 1977; Reesor and Moore, 1971). Brun has recognized "une opposition constriction-aplatissement," in which the structural trends within the gneiss dome reflect a regime of constriction at the center of the dome and
one of flattening at the periphery. This relationship is compatible with the experimental studies of Dixon (1975) and has also been noted by Johnson (1968) for the major deformation affecting the Chirwa dome, Rhodesia.

Some generalizations can also be made about the structural chronology of mantled gneiss domes. The first deformational event typically produces recumbent isoclinal folding and bedding-plane foliation (Brun, 1977; Mallick, 1967); bedding may be transposed. Large-scale infolding of the core-mantle boundary as described by Reesor and Moore may occur at this time. Chesnokov (1966) states that normal drag folds are characteristic of this deformational phase while reverse drag folds typically develop later. Later folds are also generally upright rather than recumbent.

Dome formation is subsequent to this first phase; it also generally postdates the climax of metamorphism, as in the case of the Ntungamo gneiss dome (Nicholson, 1965). According to Read and Watson (1975, p. 138), in the Copperbelt of Zambia, tight folds and dislocations in the cover are earlier than the rise of the domes. Cataclasism and shearing along the margins of the dome, attributed by many authors to the final phases of dome formation, are late-stage events.

The polyphase evolution of a diapiric structure, Stephans and Johnson have cautioned, may lead some workers to incorrectly assume several temporally distinct periods of regional deformation.

Mineralization

Several types of economic mineralization occur in the vicinity of mantled gneiss domes (see Chapter 4). One class of deposits is associated with the younger intrusions of the domes, and includes contact metasomatic deposits and pegmatite deposits which might develop in relation to granitic plutons, independent of the gneiss dome setting. The dolomitic skarns of the classic Pitkaranta dome, in which Eskola (1949, p. 463) reported "numerous deposits of chalcopyrite, sphalerite, galena, cassiterite and magnetite," probably belong to this class. Tourmalinization, commonly described in the strata above the core gneiss, may reflect metasomatism induced by volatiles migrating away from young granitic liquids.

Other deposits seem to be linked to recycling processes which are unique to the gneiss dome environment. The historical sequence which results in the formation of a mantled gneiss dome may in some instances effectively concentrate elements which are only slightly enriched in the original basement. The rich uranium deposits of the Alligator Rivers area of northern Australia are a possible example. Many of the major orebodies of the region are
stratabound in the Lower Proterozoic Cahill Formation, a cover sequence which has been complexly infolded between a gneiss dome and a migmatite complex (Needham and Stuart-Smith, 1976). The ultimate source of uranium in the Alligator Rivers area is thought to be the Archean basement; analyses of the Nanambu Complex gneiss have averaged 5 ppm U, slightly greater than typical values for granitic rocks. A first stage of concentration probably occurred during the deposition of the Cahill Formation under reducing conditions. The sedimentary ores thus formed were then reconstituted and further concentrated during a period of basement reactivation which occurred ~1800 m.y. ago. The role of near-surface processes in mineralization is controversial, but may have been substantial.

The Rum Jungle area to the west of Alligator Rivers seems to have a similar pattern of ore genesis. According to Stephansson and Johnson (1976, p. 184), "the diapiric emplacement of granites provided a possible energy source to remobilize and concentrate base metal, copper, and uranium ore." The copper and uranium deposits of the Zambian Copperbelt may also fall into the category of syngentic, sedimentary ores which have been redistributed during the formation of mantled gneiss domes. The proposed "consanguineous" (=syngentic) origin for the Passagem de Mariana gold deposit, Brazil, situated on the margin of the Bacao gneiss complex (Fleischer and Routhier, 1973, 1974), raises the possibility that this area too has benefitted from polycyclic concentration of ore.

Metamorphic concentration associated with gneiss dome formation can apparently operate on volcanogenic, as well as sedimentary, mineralization in the mantle. Il'ina (1977, p. 333) has described gneiss domes in central Karelia, mantled by basic metavolcanics and amphibolites: "The widely disseminated sulfide mineralization of the basic rocks in all probability, serves as the original material from which the formation of high concentrations of ores is possible under the influence of metamorphism".

The mantled gneiss dome setting may also permit the concentration of metals via anatectic melting. The primary mineralization at the Rössing uranium deposit, Namibia, may exemplify this process. The Rössing deposit is located on the southwest flank of a gneiss dome, where low-grade uranium mineralization is disseminated in alaskitic permatites which intrude the metasedimentary mantle. The protore of these deposits is most likely the core gneiss (Jacob, 1974). Anatexis to yield uraniferous melts may have involved lower sedimentary units, as well as the gneiss itself; therefore, sedimentary processing of uranium may have contributed to its present concentrations at Rössing (Nishimori and others, 1977).
Uranium mineralization in the Bancroft District, Ontario is in many ways analogous to that at Rössing, being associated with pegmatite swarms marginal to mantled gneiss domes (Little and others, 1972). The distribution of mineralization in both the Rössing and the Bancroft deposits is consistent with the observation of Nedashkovskiy (1976, p. 222), who investigated geochemical zonation in the vicinity of two Siberian gneiss domes: "The highest concentration of lithophile elements occurs in displaced granite melts above domes."

Finally, mantled gneiss domes may contain structural horizons which are favorable to later, post-tectonic mineralization. A dome in southern Primor'ye on the east coast of the Soviet Union apparently hosts a post-tectonic gold deposit (Epshteyn, 1969); gold is largely confined to the dome core, and tends to be concentrated along the core-mantle interface.

Spatial and Temporal Distribution

The mantled gneiss domes reviewed by this survey are plotted in Figure 15. Fletcher (1972, p. 197) states that gneiss domes "have joined nappes and overthrusts as important elements in the tectonics of orogenic terranes." Dixon (1955, p. 89) called mantled gneiss domes an important structural feature of the core zone of orogenic belts and, in similar fashion, Salop (1972, p. 1219) states that they are found in almost all folded complexes, although "their significance in tectonic structures of different ages is highly varied." As illustrated in Figure 2-5, mantled gneiss domes have been reported to occur on all continents.

Although early Archean structures have sometimes been referred to the mantled gneiss dome-model (Lowman, 1976, p. 21), they have generally been excluded from this compilation. The numerous granitic plutons invading Archean greenstone belts, causing the "gregarious batholith" pattern described from the Rhodesian craton by MacGregor (1951), are often considered as diapirs, but it is now doubted that they represent rejuvenation or remelting of older sialic crust (Glikson, 1972; Arth and Barker, 1976; Barker and Arth, 1976). The classic mantled gneiss domes of Eskola are of Proterozoic age, and Salop (1972) regards such structures to have been most prominent during this time span, and especially characteristic of the early Proterozoic. He believes that Phanerozoic mantled gneiss domes are of restricted occurrence and that their "dying-off" is a reflection of "sclerosis" of the earth's crust over time.
Figure 2-5

Global distribution of mantled gneiss domes reported in geologic literature. Numbers correspond with those given to individual descriptions of the domes in Chapter 2, Part II.
CONCLUDING REMARKS: MANTLED GNEISS DOMES AND

METAMORPHIC CORE COMPLEXES

Because Cordilleran metamorphic core complexes have widely varied pre-Tertiary histories, one cannot appraise, in a general way, their similarity to mantled gneiss domes. Each complex must be individually assessed on the basis of the material presented in the other reports, and in this report.

One may remark, however, that most of the mantled gneiss domes presented in Appendices I and II are linked, at least circumstantially, to a collisional tectonic setting: gneiss domes are ubiquitous near Precambrian sutures, and within the core zones of Caledonide and Hercynian orogens. In addition, none have yet been identified within the South American Cordillera. It seems appropriate to reiterate at this point den Tex's suggestion that crustal thickening is a necessary precursor to subsolidus convection. This factor, or some other, may serve to limit gneiss dome formation to collisional environments.
PART II: DESCRIPTIONS OF INDIVIDUAL AREAS

NORTH AMERICA

1. **Uchi Subprovince, Ontario.** Guided by conspicuous ovoid magnetic anomalies, Breaks and others, (1974), delineated three gneiss domes in the Uchi Subprovince of Ontario. The cores of the domes consist of foliated trondhjemite, and they are mantled by metavolcanic amphibolite intercalated with biotite quartzofeldspathic gneiss and biotite trondhjemitic gneiss. Numerous stocks and dikes of unmetamorphosed leucocratic quartz monzonite intrude the dome rocks, and, in many areas, have obliterated the original core-mantle relations. Thurston and Breaks (1978) have interpreted the core gneiss as ancient sialic basement onto which mafic lavas were extruded, in the time period 2960-2740 m.y. They believe that the resulting density inversion caused a gravity-driven deformation which was characterized by northward verging nappes, akin to the "pleurotoid" diapirs of Talbot (1974).

2. **Emile River, Northwest Territory.** Several gneiss domes were described in the Arseno Lake map area near Emile River by Frith and Leatherbarrow (1975), and a polymetamorphic evolution of the area was subsequently confirmed by the isotopic determinations of Frith and others, (1977). Slightly foliated granitic gneiss occupies the dome cores, and is encircled by migmatized sediments of the Proterozoic Snare Group. The core gneiss of the "Amoeba" Lake dome has yielded a Rb-Sr whole rock isochron of 2712 m.y., thought to represent its absolute age; and undeformed alaskitic pegmatite has yielded an isochron of 1808 m.y. The domes are within 15 km of the boundary between the Bear and Slave provinces, and are believed to have formed during a compressional event along this boundary, which culminated 1900 m.y. ago (Frith, 1978).

3. **Watersmeet, Northern Michigan.** The geochronology of the Watersmeet gneiss dome (Sims and Peterman, 1976) is quite similar to that of the Emile River domes. According to Sims and Peterman, who conducted Rb-Sr isotopic studies, the feldspathic augen gneisses and biotite quartzofeldspathic gneisses which now outcrop in the core of the dome were formed at least 2600 m.y. ago. Following deposition of the iron-bearing sediments of the Marquette Supergroup (Precambrian X), these gneisses were reactivated to form a mantled gneiss dome, approximately 1800 m.y. ago. The dome now lies within the garnet isograd, and is the center of a metamorphic node, as defined by James (1955).

4. **Bancroft District, Ontario.** Hewitt (1957) described a series of four mantled gneiss domes at the southern end of the Hastings-Haliburton Highlands. The gneiss complexes are composed of hybrid granite gneiss with migmatites, pink leucogranite gneiss
and granitic pegmatites. The mantle assemblage consists of marble, paragneiss, and amphibolites of the Grenville Group (early Heli-
kian, Stockwell and others, 1968). Pegmatite dikes occur in swarms which are concordant with the enclosing metamorphic strata, and are associated with the major uranium mineralization of the Bancroft District. The absolute age of uraninite from the dis-
trict has been estimated at 1060-1020 m.y. (Little and others,
1972), and dome formation is presumably an effect of the Grenville orogeny. Isotopic investigations, however, have thus failed to detect an ancient pre-Grenville heritage in the granitic rocks (Silver and Lumbers, 1965).

5. Adirondack Inlier, New York. Marbles, amphibolites and other metasedimentary rocks of the Grenville Series also occur in the Precambrian inlier of the Adirondack Mountains, New York, where they are associated with anorthosite, and related gneisses. The anorthosite and gneisses have traditionally been regarded as intrusive into the Grenville Series, but deWaard and Walton (1967) have argued that in some localities, these rocks occupy the cores of mantled gneiss domes and nappes. They have suggested that the anorthosite and gneisses belong to a pre-Grenville basement which was severely deformed during the Grenville orogeny.

6. El Oro, North-Central New Mexico. The El Oro gneiss dome recently defined by Budding and Cepeda (1979), is an elongated, doubly plunging structure which contains in its core locally migmatitic, mica gneiss. Budding and Cepeda consider the gneiss to be metasedi-
mentary, but do not rule out an origin from felsic volcanic rocks. The mantle, which exhibits upper amphibolite facies mineral assem-
blages, consists of mica schist, impure marble, amphibolite and quartzite. The contact between the gneiss and mica schist is grad-
ational. Structural analysis has demonstrated multiple deformations in the gneiss, but geochronological data for the area is absent. Budding and Cepeda explicitly refer the structure at El Oro to the classic mantled gneiss dome model, but their description of the El Oro dome, as it stands, falls short of convincing. A former nonconformity apparently cannot be demonstrated at the present core-mantle con-
tact, and it seems possible that the gneiss and schist represent a conformable supracrustal sequence which was recrystallized during a single metamorphic event.

7. Connecticut Valley Synclinorium, New England. Slightly west of the trough of the Connecticut Valley Synclinorium is a series of seven domes (Rodgers, 1970; Skehan, 1961); the northern six are in the state of Vermont, while the southernmost is in Connecticut. The five southern domes expose cores of paragneiss and felsic metavolcanics, rocks which are probably Precambrian in age. Surrounding these domes is the Lower Paleozoic synclino-
rial sequence, essentially complete but greatly thinned. The two northern domes do not expose rocks below a limestone-phylilit
unit in the upper one third of the sedimentary sequence, but gravity surveys suggest that gneissic cores are present in the subsurface (Bean, 1953). Granites cross-cutting the gneisses are uncommon in the Connecticut Valley domes, but a stock of muscovite-biotite granite is present in the Guilford dome, southern Vermont. A regional maximum in metamorphic grade coincides with the line of domes. The age of the principal deformation in the Connecticut Valley Synclinorium is stratigraphically constrained to a period between the Early Devonian and the Late Carboniferous; dome formation is therefore considered an Acadian event.

8. Bronson Hill Synclinorium, New England. About 20 drop-like domes are aligned along the Bronson Hill Anticlinorium of New Hampshire, Massachusetts, and Connecticut. Granitoid rocks of the Oliverian plutonic series form the cores of the domes; these are commonly, but not invariably gneissic. Adjacent to the gneiss in most domes are the Ammonoosuc metavolcanics of Ordovician age. Overlying Ordovician strata are variable, but rusty slates to schists are typical. These rocks are unconformably overain by a thick Silurian-Lower Devonian clastic sequence, with a quartzite at its base. The Oliverian granitic rocks were long considered, following Billings (1956), to be concordant Devonian intrusions. However, Naylor (1969) has in recent years described post-Ammonoosuc granite, unconformably overlain by Silurian quartzite, and dated at 440 and 450 m.y.; yet more recently, Hills and Dasch (1971) have recovered an Avalonian (616 m.y.) date from one of the core granites. It therefore seems likely that the Oliverian Series includes rocks of several origins.

The structural complexity of the Oliverian domes is formidable; the domes deform older nappe structures and have themselves mushroomed and overridden each other. Exotic, tongue-shaped lobes characterize structure sections through the domes (Thompson and others, 1968).

9. Baltimore-Washington Anticlinorium. The Baltimore Gneiss outcrops in seven domes which are localized along a regional structure, the Baltimore-Washington Anticlinorium (Broedel, 1937; Hopson, 1964). The formation includes coarse augen gneiss and granitic gneiss, as well as more extensive veined gneiss and migmatites. The late Precambrian Glenarm Series mantles the gneiss domes, and includes quartzites, feldspathic mica schist, marble and pelitic schist. Younger intrusions transgress the gneiss-metasediment contact; the two-mica Gunpowder Granite is described by Hopson (p. 47) as "a rheomorphic offshoot of the Baltimore Gneiss."
U-Pb measurements on zircon and whole-rock Rb-Sr analyses give consistent ages of 1050 m.y. for the Baltimore Gneiss. Dome formation is assumed to have been essentially complete by 425 m.y., the Rb-Sr age of post-Glenarm pegmatite swarms (Wetherill and others, 1966). The metasedimentary rocks yield a scatter of K/Ar mineral ages of 350-300 m.y., which is interpreted as a relic of gradual cooling.

10. **Shuswap Complex, British Columbia.** The Shuswap Metamorphic Complex, situated in the core zone of the Canadian Cordillera, contains three domal outcrops of gneiss on its eastern margin: (from N to S) Frenchman's Cap, Thor-Odin and Valhalla gneiss domes (Reesor, 1970). A fourth domal structure in the Pinnacle Peaks region appears to represent a stratigraphic level higher than that of the core gneiss. Much of the granitoid core gneiss of the Shuswap domes is of metasedimentary origin, and in the recent past was considered as migmatized Windermere Group (late Proterozoic). The mantling zone, which consists of quartzite, marble, and pelitic, psammitic and calc-silicate gneiss, was regarded as equivalent to a lower Cambrian sequence (Reesor and Moore, 1971). Older Precambrian basement was not recognized, and the core-mantle interface was assumed to be the boundary of a "migmatite front," which had been halted at a resistant stratigraphic horizon. High-grade metamorphism, and the rise of the migmatite front was thought to have accompanied Columbian orogeny.

However, geological interpretations of the Shuswap Complex are presently being revised. Wanless and Reesor (1975) reported 1.96 b.y. -old zircon from a granodiorite gneiss in the Thor-Odin dome; orthogneisses from Frenchman's Cap dome have yielded Rb-Sr ages of 2.1 b.y. (Brown, 1980). Paragneisses from four Frenchman's Cap localities have also produced Aphebian Rb-Sr ages. The mantling zone may therefore correlate with the Proterozoic Purcell Group (Brown, 1980; Read, 1980). The age of the main metamorphic stage in the complex is still considered Late Jurassic (Columbian), but an Eocene thermal overprint is also recognized (Medford, 1975). Read (1980) has argued that the Frenchman's Cap and Thor-Odin complexes are not domal in cross-section, but rather exhibit nappe geometries. According to Read (p. 19): "The nappe structure, lack of diapirism, and non-coincidence of thermal culminations with extensive areas of core gneiss do not support a gneiss dome concept."

**GREENLAND**

11. **Rinkian Mobile Belt, West Greenland.** Escher and Pulvertaft (1976) have distinguished a distinct tectonic province in the Precambrian terrane of West Greenland, north of Jacobstavn, the Rinkian mobile belt, in which the most obvious structures are large gneiss domes. Gneiss domes were first described in the
Umanak area by Henderson (1969). Outcropping in the dome cores is the Umanak Formation, consisting largely of biotite- or biotite hornblende-gneiss, with at least some metasedimentary horizons. Overlying the gneiss are the supracrustal rocks of the Karrat Group, which comprises a lower, dominantly quartzitic formation and an upper formation, the Nukavsak, largely semipelitic and pelitic schists. In spite of the transitional character of the gneiss-sediment boundary over most of the area, and the absence of observable discordance between the Umanak Gneiss and the Karrat Group, Henderson has convincingly argued that the gneisses have a basement-cover relationship with Karrat Group sediments. Metamorphic grade in the Karrat Group clearly increases with proximity to gneissic cores. The Umanak gneisses are considered to be Archean by Escher and Pulvertaft; biotite from the gneiss has yielded a K/Ar date of 1790 m.y., presumably a metamorphic age. Two schist samples from the Karrat Group yielded K/Ar ages of 1700 m.y., and biotite from a pegmatite in the gneiss was dated at 1690 m.y.

South of the Umanak area, Escher and Pulvertaft have recognized another large gneiss dome, the Talorssuit. Core and mantle sequences at Talorssuit are lithologically similar to those at Umanak, except that metavolcanic rocks constitute an important part of the Talorssuit mantle. A feature unique to the Talorssuit dome is a huge granitic sheet, developed along the contact between the younger supracrustal rocks and the basement rocks. The western flank of the dome is strongly overturned and the resulting nappe-like structure exhibits a maximum overlap of 12 km.

12. Central Metamorphic Belt, East Greenland Caledonides. The well known structural synthesis of the Central Metamorphic Belt by Haller (1955) involved a superstructural mantle of gently folded metasediments, an infrastructure rendered highly mobile by a rising migmatite front, and a zone of detachment between the two levels. Upwellings of the migmatitic infrastructure formed bulges which Haller classified as domes, foreheads, sheets and mushrooms. The superstructure was presumed to contain the metamorphosed equivalents of the Elsonore Bay Group (late Precambrian-Ordovician) and mobilization was considered to be a Caledonian phenomenon. Recent radiometric studies have suggested that basement of the Central Metamorphic Belt dates from the Archean (3000-2500 m.y.; Henriksen and Higgins, 1976). However, these studies also indicate that the metamorphosed supracrustal sequence is older than the Elsonore Bay Group, and that a pre-Caledonide, middle Proterozoic orogeny probably affected the area. Some of the structures formerly ascribed to a single episode of migmatitic upwelling may therefore be the result of superposed deformations of widely different ages.
SOUTH AMERICA

13. Bacao Complex, Quadrilátero Ferrífero, Minas Gerais, Brazil. The Bacao complex in the Quadrilátero Ferrífero contains in its core a weakly foliated granodiorite which has been dated at 2440 m.y. (K/Ar, biotite; Herz and others, 1961). The granodiorite is surrounded by the well-foliated Itabirito granite, dated at 1340 m.y. The Itabirito granite is in turn enclosed by metasedimentary rocks which probably belong to the Rio das Velhas Series, of uncertain age. Herz and others suggested that the Itabirito granite had formed by complete anatexis of the granodiorite, and incorporation of argillaceous sediments into the resulting melt. Fleischer and Routhier (1973, 1974), however, emphasize the concordance of the granite-metasediment contact and repeat an earlier suggestion for interpretation of the complex as a mantled gneiss dome.

EURASIA

14. Norwegian Caledonides. Ramberg (1967a) described the Namsos-Grong and Møre basal gneiss culminations in the Norwegian Caledonides, in which domal structures predominate. Gneiss of the Namsos region has yielded Rb-Sr ages of 1900-1800 m.y. (Z.W.O. Lab, 1968). A thin autochthonous cover of Eocambrian sparagmites and Cambrian schists is locally present, but was probably overestimated by Ramberg (see Roberts, 1978); most of the cover succession found within the Trondheim synclinorium to the east is allochthonous, and represents a series of nappes that were probably derived from west of the present coastline of Norway. Metamorphic fabric within the nappe piles, assumed to have been developed during the early stages of thrust-faulting, has been assigned a minimum age of 438 m.y. (Wilson and Nicholson, 1973). To the west, where basement remobilization occurred, dynamic metamorphism outlasted nappe emplacement, camouflaging the basement-allochthon contacts.

15. Eastern Finland and Southern Karelia. The establishment of the Finnish Karelides as a type area for mantled gneiss domes has proven opportune. In few other areas are the domes so numerous or free from structural complexity and subsequent orogenic overprinting. The cores of the Karelide domes contain migmatites, granitic gneiss and porphyritic granites, and are mantled by the transgressive Jatulian sedimentary sequence.

Radiometric dating is complementary to the geological chronology proposed by Eskola (1949). Biotite from granite gneiss has yielded a K/Ar age of 1740 m.y. (Wetherill and others, 1962). Zircons from the basement complex give discordant U-Pb ages which are suggestive of lead loss. Wetherill and Kouvo (1966) show that the U-Pb ratios can be interpreted according to
a model which assumes an absolute age of 2800 m.y., analogous to Saamidite basement to the east, with an episodic loss of lead 1800 m.y. ago.

16. Central Karelia, USSR. Il'ina (1977) has located 21 dome structures, many on the basis of aeromagnetic data alone, along the juncture of the Karelidite belt with the older Belomoride belt to the east. The domal cores consist of granite gneiss, which frequently grades into granite towards the center of the domes. Basic metavolcanics and amphibolite apparently form much of the mantle sequence. K-Ar ages of the gneiss do not exceed 1800 m.y., and are assumed to reflect rejuvenation of Archean basement.

17. St. Malo Complex, Massif Armorican, France. Many ages have been proposed for the St. Malo massif on the northwestern coast of France, but Brun (1977) argues that the complex developed entirely in Cadomian time, culminating in the formation of a migmatite dome ~600 m.y. ago. He has recognized three lithostructural units within the complex: a core of migmatites and anatectic granites, a gneissic mantle and an "envelope" of mica schists. The schists, he maintains, pass conformably into the low grade Brioverian (900-650 m.y.) metasediments of central Brittany.

Brown (1978, 1979), who contends that migmatization preceded the deposition of the Brioverian cover sequence, has noted the absence of critical radiometric data for the area.

18. Agout Dome, Montagne Noir, France. The geology of the Agout Dome, Montagne Noir, has been summarized by denTex (1975); a more detailed account is provided by Schuiling (1961). The formation of the dome postdates early Hercynian nappe development in the Montagne Noir; the dome itself is aligned with late Hercynian structures. The core of the dome consists of ortho- and paragneisses, with a central migmatite zone. The orthogneiss has yielded a whole rock Rb-Sr isochron of 530 m.y.; the migmatite generally produces isochrons of 475-419 m.y., although some samples indicate local homogenization at 320-280 m.y. (Hamet and Allegre, 1972, 1976). The mantle of the dome consists of Upper Brioverian to Lower Carboniferous sediments, metamorphosed to upper amphibolite grade.

19. Pyrenees. The axial zone of the Pyrenees contains a number of large, gently-arched gneiss domes, such as the Canigou and Aston-Hospitalet massifs (Rutten, 1969). These domes consist of porphyritic orthogneiss and feldspathic paragneiss, and are mantled by Cambrian-Ordovician metasedimentary rocks (Fonteilles and Guitard, 1968). Metamorphism affects progressively higher parts of the sedimentary sequence as one moves west across the axial zone (Zwart, 1968). One interpretation of these domes is
that they represent autochthonous pre-Hercynian basement, mobili-
ized during the Hercynian orogeny. Vitrac and Allegre (1975) have determined the lower limit for the age of an orthogneiss from the Canigou Massif to be 535 m.y. Rutten (p. 348) has suggested that a more or less fortuitous superposition has caused Hercynian structures to be exposed in the center of a cross-
cutting Alpine orogen.

20. Pennine Alps. Although Eskola himself (1949, p. 472) proposed that the gneiss masses of the Pennine Alps might be compatible with a mantled gneiss dome interpretation, this possi-
bility has rarely been discussed in subsequent literature. denTex (1975), however, has selected the Lepontine gneiss region of Switzerland to illustrate his theory of convective remobilization of the basement. The area constitutes a thermal dome with five subdome structures. The Lepontine gneisses are conformable with mantling Mesozoic cover rocks, but have generally yielded Hercynian Rb-Sr ages. The Alpine thermal culmination postdated nappe emplacement and has been fixed for this region at 38 m.y. (Hunziker, 1970).

21. Menderes Massif, Turkey. Augen gneiss forms the cores of four domes within the Menderes massif, Turkey (Brinkman, 1976-
von der Kaaden, 1971; Graciansky, 1966). The gneiss is mantled by a metasedimentary sequence of mica schist phyllite, meta-
quartzite, and marble. The schist, phyllite, and quartzite are thought to have been derived from Ordovician-Devonian sediments, while the marble is probably equivalent to Lower Carboniferous-
Jurassic? limestones. Whole-rock Rb-Sr analysis of the augen gneiss has produced ages of 529 m.y. and 490 m.y. (Cambrian-
Ordovician). A uraninite vein in the southern gneiss core has been dated at 268 m.y. (Permian), and an undeformed granite has yielded a whole-rock Rb-Sr isochron of 167 m.y. (Jurassic). Brinkman feels that the last metamorphism was probably Jurassic.

22. Saksagan Dome, Ukraine, USSR. Kalyayev (1970) described the Saksagan migmatite dome, some 80 km wide, east of Krivoy Rog, Ukraine, which contained in its core reconstituted, migmatitic basement and younger granites. The mantling sequence, as described by Kalyayev, consisted of apospilites and orthoschists at the base, followed by the extremely thick, heterogeneous Krivoy Rog series. Lysak and Sivoronov (1976) have since argued, however, that the apparently domal structure within the Saksagan block actually results from the juxtaposition of two tectonically dis-
tinct complexes.

23. East Ural Anticlinorium USSR. Gneiss domes have been described in the East Ural Anticlinorium by Chesnokov (1966, 1967). The core rocks of the Larino compound dome consist of granite
paragneiss; the dome has an inner mantle of gneiss-amphibolite and an outer mantle of schist and quartzite. Mantling strata belong to the early Paleozoic eugeosynclinal Larino series. The domes are presumably Uralide structures, and are regarded by Chesnokov (1966, p. 45) as "the natural result of intense geosynclinal folding and regional metamorphism in the axial zone of mobile belts."

24. Ulu-tau and Kokchetav Massifs, West-central Kazakhstan, USSR. According to Pavlova (1967), a series of arches and elongate domes occur along an anticlinorium in the core of the Ulu-tau massif. Granite occurs in the centers of these domes and is gradational into granitic gneiss near the contact with Riphean (upper Proterozoic?) strata. The cover sequence consists of a lower suite of acidic volcanics ("porphyroids") and an upper sequence of volcanogenic sediments. The morphology of the Ulu-tau domes is simple and is not suggestive of great plasticity.

The Kokchetav massif lies north of the Ulu-tau massif, within the same zone of uplifts in west-central Kazakhstan. The structure of the massif is dominated by four domes, the largest of which is 80 km in diameter (Pavlova, 1972, citing Rozen and Serykh, 1969). The Kokchetav basement consists of ortho- and paragneisses and amphibolites; the cover sequence is similar to that of the Ulu-tau region (Nalivkin, 1973). The form of the domes is again relatively simple but intrafolial folding is widespread on the dome flanks. The time of deformation is assumed to coincide with that of amphibolite grade metamorphism (~1000 m.y., Rozen and Yanitskiy, 1974).

25. Mama Region, Transbaikalia, USSR. According to Salop (1972), the Mama region of Transbaikalia contains about 20 mantled gneiss domes, many of which are elongated in a NE-SW direction. Some of the domes are bulb- or mushroom-shaped. The crystalline basement in this area, which does not outcrop in all the domes, consists of Lower Proterozoic gneissic granites. The domes are mantled by a thick Upper Proterozoic metasedimentary sequence, the base of which consists of high grade meta-arkose with orthoamphibolite horizons (Koganda Formation). Migmatization of the metasedimentary rocks is sometimes observed near the core-mantle contact.

26. Aldan Shield, USSR. Many gneiss domes have been reported within the Aldan Shield of the Soviet Union, which forms the southeastern part of the Siberian Platform. Salop (1972) has recognized, within the Aldan Shield, a number of "gneiss folded ovals, "huge structures which he considers characteristic of Archean deformation; the Lower Timpton dome (Grabkin, 1965), which attains a diameter of 170 km, is an example. Smaller gneiss domes are common within the folded ovals, and are considered by Gladkov
and Grabkin (1978) to have been superimposed on the ovals by subsequent "gneiss dome orogeny." Within the Verknealden folded oval or "amoeboid," Salop has mentioned the Suon-Tit granite gneiss dome, in which alaskite outcrops from beneath a mantle of Archean strata (Yengra and Timpton Subgroups).

27. Kodar Udokan Region, USSR. In the western Aldan Shield, numerous gneiss domes of various morphologies are found in the Kodar-Udokan region (Leytes and Fedorovskiy, 1972; Sorachev, 1974). In this area, Archean schists and gneisses of the Chara Series now occur as small inliers and remnants, the original basement having been largely reconstituted during extensive migmatization and granite intrusion in Early Proterozoic time. According to Leytes and Fedorovskiy, "granite-gneiss and migmatite domes, mushroom structures, and sharp interdomal synclines" were formed during this episode of magmatism and metamorphism. The mantle sequence of the structures consists of the Lower Proterozoic Udokan series, which is metamorphosed to amphibolite facies in the vicinity of the domes.

28. Nercha(insk) Range, Southeastern Transbaikalia, USSR. Sinitsa (1965, 1975) has described two Jurassic mantled gneiss domes in the Nercha Range of southeastern Transbaikalia: the Tsagan-Oluya dome and the Kuotmar dome. The core of the Tsagan-Oluya dome contains biotite-hornblende gneiss, which is commonly migmatitic. The gneiss is mantled by Lower- to Mid-Jurassic conglomerates and sandstones, which have been metamorphosed to amphibolite grade and extensively invaded by pegmatites in a zone within 2-5 km of the core. In the eastern part of the dome, a massive biotite granite intrudes both the core and mantle rocks. The Kuotmar dome, which is located northeast of the Tsagan-Oluya dome, contains a granitic core which is deformed only in a narrow zone (2-4 km) around the edge of the core. Jurassic deposits are foliated close to the contact with the granite; biotite and epidote occur as alteration minerals in both the granite and the sediments in the contact zone.

Sinitsa considers the cores of both domes to represent basement of probable Paleozoic age. Late Jurassic tuffs and extrusives and Early Cretaceous sediments unconformably overlie the mantle sequence.

29. Bureyan Massif, Soviet Far East. Nedashkovskiy (1976) has conducted geochemical studies on granite-gneiss domes in the Soviet Far East. The Yaurinsk dome in the southeastern part of the Bureyan massif, is formed of Proterozoic plagiogneiss with biotite granodiorite in its core; a migmatitic zone separates the granodiorite from the plagiogneiss. The overlying metamorphic rocks are in the epidote-amphibolite facies. The younger intrusive rocks of the dome may fall into an early Paleozoic magmatic suite described by Putintsev and others (1972).
30. **Central Kamchatka, USSR.** Granite gneiss domes in the metamorphic zone of central Kamchatka were described by Lebedev and others (1970). Additional information may be found in this reference, which was not available to this survey.

31. **Southern Primor'ye USSR.** In southern Primor'ye, near Vladivostok, a small (5 x 6 km) mantled gneiss dome has been described within a zone of gold mineralization (Epshteyn, 1969). The core of the dome consists of middle Paleozoic diorites which have been cataclastically deformed. Upper Permian clastic sediments, including a basal conglomerate, overlie the middle Paleozoic rocks; the sediments are transformed into phyllites and chlorite-sericite schists near the core of the dome. The metamorphic aureole is about 1.5 km in width, and its outer edges coincide with a zone of "severe dislocation" in the Permian rocks. Rocks of both the core and the metamorphic mantle are extensively intruded by numerous aplite and pegmatite dikes. Epshteyn believes that magmatization occurred in Upper Cretaceous time (125-113 m.y.). Gold mineralization is apparently a post-tectonic phenomenon, but concentration of gold within the dome structure is many times higher than that of the mineralized belt as a whole.

32. **North Korea.** Several mantled gneiss domes were reported to exist in North Korea by Salop (1972). Lower Precambrian gneissic granites, gneisses and schist occur in the cores of these domes; these rocks are mantled by upper Proterozoic sediments (Sanvon and Kuchen Formations) which have undergone epidote-amphibolite and amphibolite grade metamorphism. The first noticeable angular unconformity in the cover sequence occurs at the contact between the Phkhanan Group (Middle Carboniferous-Lower Triassic) and Upper Triassic-Lower Jurassic continental beds, where the angular discordance is sharp. Salop therefore proposes that remobilization of the basement complex occurred in Triassic time.

33. **Core Zone, Malaysia.** Richardson (1950) described the Bukit Berentin and the Bukit Ranjut complexes of Malaysia as igneous intrusions, but more recently, Hutchinson (1973a) has tentatively suggested a reinterpretation of both complexes as remobilized portions of the basement. The core of the Bukit Berentin Complex is well foliated and the structure is conformable with that of the surrounding metasedimentary rocks. The rocks surrounding the gneiss are intruded by swarms of minor granite apophyses. Gold placers in nearby streams are thought to derive from the Bukit Berentin complex.

Hutchinson has also considered the Stong Migmatite Complex and the Taku Schist terrane, north of Bukit Berentin and Bukit Ranjut, to represent infrastructural upwellings of Malaysia's
metamorphic core zone. The Stong Migmatite was apparently derived from a predominantly arenaceous sedimentary sequence that underwent large-scale anatexis. The Taku Schist, derived from mainly pelitic sediments, has yielded K/Ar dates of 215-220 m.y. Both the Stong Migmatite and the Taku Schist are presumed to be the metamorphic equivalents of Lower Paleozoic rocks.

34. Mysore State, India. Several circular to elliptical bodies of gneiss, 10-35 km in diameter, are found completely encircled by schists of the Dharwar Supergroup in Mysore State, India, and have been considered by some geologists to resemble classic mantled gneiss domes (Pichamuthu, 1967; Radhakrishna and Vasudev, 1977). The gneiss is tonalitic in composition and presumably formed the basement on which the sediments of the Dharwar Supergroup were deposited. A tonalitic cobble from a conglomerate of the Lower Dharwar Group has been dated at 3250 m.y. (Ventkatasubramanian and Narayanaswamy, 1974, p. 318). The Dharwar sediments themselves are thought to have been deposited in the period 2600-2100.

AFRICA

35. Central Karagwe-Ankolean Belt, Southern Uganda. In southwest Uganda, a number of low-lying areas, termed "arenas," are underlain by domes of non-resistant granitic rocks which are encircled by basal quartzites of the Precambrian Karagwe-Ankolean cover sequence. Local geologists have suggested as long ago as 1951 (Nicholson, 1965, p. 157) that these represent mantled gneiss domes. Nicholson (1965) presented a description of the Ntungamo gneiss dome of southern Uganda to a meeting of the Geological Society of London, and discussion participants felt that it also could be referred to the mantled gneiss dome model.

The Karagwe-Ankolean cover sequence is regionally considered to postdate 1800 m.y. Granitic gneisses from the Ntungamo dome core have yielded a whole-rock Rb-Sr isochron of 1185 m.y., (Cahen and Snelling, 1966).

The Kalahari craton of southern Africa is rimmed by mobile belts characterized by radiometric ages in the range 650-400 m.y.: the Damaran orogen on the west, the Lufilian or Zambezi belt to the north, and the Mozambique belt on the west. A number of mantle gneiss domes have been described from the circum-Kalahari region, with estimates for the age of crustal rejuvenation frequently falling around 500 m.y.
36. **Abbabis Complex, Namibia (South West Africa).** An inlier of pre-Damaran basement southwest of Karibib, Namibia—the Abbabis Inlier—has long been known to local geologists, and Smith (1965, p. 10) suggested that remetamorphosed Abbabis rocks might also be present in the cores of numerous dome structures in the area. Recently, it has been demonstrated (Jacob and others, 1978) that much of the so-called Red Granite-Gneiss, of supposed syntectonic Damaran age, actually correlates with the ancient Abbabis granite-gneiss; Jacob and others support a model of reactivation into mantled gneiss domes. Zircons from two samples of the Abbabis gneiss have yielded a U-Pb concordia intercept of 1925 m.y.

The Rössing uranium deposit (Berning and others, 1976) lies on the southwest flank of one of these gneiss domes. Low-grade uranium mineralization is disseminated within alaskitic pegmatites which intrude the rocks of the metasedimentary mantle (Nosib Formation, Damara System). Age determinations on uraninite, davidite, and biotite from the Rössing area indicate that metamorphism and the emplacement of uraniferous pegmatites occurred within a narrow time period around 510 m.y. (von Backstron, 1968).

37. **Rietfontein Inlier, Namibia.** Some 200 km east of the Rössing area, the Rietfontein Inlier (Martin, 1965, p. 12) contains granite and gneisses of the Marienhof Formation, a possible equivalent of the Abbabis Formation. In some localities, the Nosib Formation overlies the Rietfontein granite with a thick basal conglomerate, while in other places, the Nosib and overlying formations have been intruded by pegmatites and by granitic rocks which seem to pass gradationally into the gneisses of the inlier. Martin has suggested that the Marienhof Formation was refoliated and locally remobilized during the Damaran orogeny.

38. **Copperbelt, Northern Zambia.** All the Zambian copperbelt-type deposits occur in the Lower Roan Formation (Katanga Series), in proximity to granite domes and to the Kafue anticline, which may represent the coalescence of a number of domes (Garlick, 1961). Before 1940, the granites were generally considered intrusive, and the base metal deposits were thought to be epigenetic; however, subsequent detailed mapping demonstrated that the granites are all pre-Katangan in age. No pre-Katangan radiometric ages are yet available from direct analyses of the Copperbelt granites, but Snelling and others (1964) suggest that the basement of the area dates back to ~2700 m.y. Snelling and others obtained a whole-rock Rb-Sr isochron from the Nchanga red granite of 570 m.y., but argued that the high value for $\frac{87}{86}$Sr ($0.795$) in the granite indicated probable rejuvenation and isotopic homogenization during the Lufilian orogeny. The granites of the Copperbelt, in contrast to those of other mantled gneiss domes, are generally undeformed.

Uranium mineralization is also present in the Copperbelt.
region and, according to Snelling and others, occurred in two major phases: one at 620 m.y., and one at 520 m.y.

39. **Mpande Dome, Southern Zambia.** The Mpande gneiss dome of southern Zambia, as described by Mallick (1967), has a geological history similar to that of the Copperbelt domes. A core of granitic gneiss and granite is surrounded by stratified rocks of the Katanga System, and was considered as an intrusive complex by earlier workers. Mallick has concluded that the dome formed during Lufilian deformation, as ancient granitic basement swelled upward through several horizons of its Katangan cover.

40. **Fungwi and Chimanda Reserves, Rhodesia.** Talbot (1971) has mentioned four gneiss domes in Fungwi and Chimanda Reserves, Rhodesia, of which he considers the one near Marymount Mission to be the most like a typical mantled gneiss dome. The metasedimentary mantle sequence in this area belongs to the Umkondo System (deposited 2000-1650 m.y. ago). The pre-Umkondo basement here consists of a lower paragneiss and an upper acid gneiss, and it is the latter that predominates in the dome cores. Metamorphism of the Umkondo System and dome formation presumably occurred during the Zambezian (Lufilian) orogenic event.

Talbot's highly innovative paper focuses on the Fungwi mantled gneiss dome, in which 12 or 13 small-scale domes can be recognized within the larger structure. Talbot has argued that these represent "frozen" convection cells which, if they had continued to operate, would have homogenized the gneiss, creating a rock of magmatic appearance, but at subsolidus temperatures.

41. **Chirwa Intrusion, Rhodesia.** The Chirwa Intrusion, Rhodesia, occurs in an ovoid outcrop, 3 mi in diameter, and consists of sodipotassic porphyritic and non-porphyritic granite, locally gneissic around the intrusion's rim (Johnson, 1968). The granite is completely surrounded by Archean amphibolites of the Bulawayan System, which, outward from the intrusion, are in turn overlain by garnetiferous pelitic and semipelitic schists of the Umkondo System. According to Johnson, the Archean basement and its Umkondo cover were deformed and metamorphosed during the Mozambique orogeny ~500 m.y. ago. Biotite from the Chirwa granite has yielded a Rb-Sr age of 460 m.y., with a whole-rock analysis reputedly yielding a substantially older date. Johnson believes that the Chirwa Intrusion represents remobilized basement, equivalent to the Archean sodipotassic granites which outcrop over wide areas to the west and south.
42. **Rum Jungle Area, Northern Territory, Australia.** The Rum Jungle and Waterhouse complexes of northern Australia, which for some time were assumed to represent intrusive granites, actually comprise variegated assemblages of metasediments, schists, gneisses and several types of granite (Stephansson and Johnson 1976). U-Pb isotopic studies on zircons from the Rum Jungle Complex have demonstrated that at least part of it is Archean, with an interpreted age of 2550 m.y. (Richards and others, 1966). Rb-Sr analyses from the Waterhouse Complex suggest an age of ~2450 m.y. (Compston and Arriens, 1968). The gneissic basement complex of the Rum Jungle area is overlain by Lower Proterozoic sediments of the Bachelor, Goodparla, and Finnis River groups, which are now in the lower greenschist metamorphic facies, with higher grade assemblages locally present. Rb-Sr ages of various granites intrusive into Lower Proterozoic sediments of the area fall in the range 1830-1720 m.y. (Compston and Arriens, 1968). Stephansson and Johnson believe that it was the upwelling of such granites, beneath the present-day Rum Jungle and Waterhouse complexes, which caused the Archean basement and its cover to be deformed into domes.

Both uranium and base metal mineralization occurs in the Rum Jungle area, with major deposits occurring in the synclinal zone between the two basement complexes. Mineralization is most prominent within the black shale and chlorite schists of the Golden Dyke Formation (Goodparla Group).

43. **Alligator Rivers, Northern Territory, Australia.** The Alligator Rivers area displays a geology very similar to, and perhaps continuous with, that of the Rum Jungle area. The Alligator Rivers area, like the Rum Jungle area, has been the site of rich uranium mineralization (Hegge and Rowntree, 1978; Needham and Stuart-Smith, 1976). Major uranium deposits within the Lower Proterozoic Cahill Formation (Koolpin Formation equivalent), a sequence of quartzofeldspathic and pelitic sediments which was metamorphosed to amphibolite grade. The Cahill Formation is draped around the granite-gneiss-migmatite Nanambu Complex, termed by Needham and Stuart-Smith a mantled gneiss dome. To the northeast, the Proterozoic metasediments grade into the Nimbuwah migmatite complex, which may have the form of "migmatite nappe" (Smart and others, 1975). Regional metamorphism, deformation, and the intrusion of anatectic granites in the Alligator Rivers area are thought to have occurred 1800 m.y. ago.
ANTARCTICA

44. Fosdick Mountains, Marie Byrd Land. The gneisses and migmatites of the Fosdick Mountains, Marie Byrd Land, have been interpreted by Wilbanks (1972) as the exposed part of an infrastructural dome which was once mantled by pre-Cretaceous sediments, such as those presently outcropping in ranges to the south. Halpern (1972) has suggested that Rb-Sr ages of 102-92 m.y. for Fosdick samples reflect metamorphic resetting during a Cretaceous orogeny.
## PART III: TABULAR SUMMARY OF MANTLED GNEISS DOMES OF THE WORLD

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<td>Northwest Territory</td>
<td>Frith and others (1977)</td>
<td>granitic gneiss -2712 m.y. (Rb-Sr)</td>
<td>(Snare Group) -Proterozoic</td>
<td>&quot;Hudsonian&quot; orogeny</td>
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<td></td>
<td>Frith (1978)</td>
<td>Undeformed alaskitic</td>
<td>Sillimanite grade near gneiss</td>
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<td>pegmatite -1808 m.y. (Rb-Sr)</td>
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<td>3. Watersmeet northern</td>
<td>James (1955)</td>
<td>Feldspathic augen</td>
<td>Interbedded iron-formation and mafic</td>
<td>~1800 m.y.</td>
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<td>Michigan</td>
<td>Sims and Peterman (1976)</td>
<td>gneiss; biotite</td>
<td>intermediate volcanics; interbedded iron-</td>
<td>Penokean (=Hudsonian orogeny)</td>
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<td>quartzofeldspathic</td>
<td>formation, argillite and greywacke</td>
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<td>gneiss -Crystallized</td>
<td>(Marquette Group) -Precambrian X (~Lower</td>
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<td>&gt;2600 m.y</td>
<td>Proterozoic)-Garnet grade.</td>
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<td>Ontario</td>
<td>Silver and Lumbers (1965)</td>
<td>Pink leucogranite</td>
<td>(Grenville Group) -early</td>
<td>Grenville orogeny</td>
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<td>Little and others (1972)</td>
<td>gneiss -1250,</td>
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<td>1125 m.y. (U-Pb)</td>
<td>Helikian (=Middle Proterozoic). Granulite</td>
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<td>Granitic pegmatites</td>
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<td>-Uraninite associated</td>
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<td></td>
<td>with pegmatites</td>
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<td>1060-1020 m.y.</td>
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<th>Type and Age Details</th>
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<td>Skehan (1961)</td>
<td>Mica gneiss and Migmatites -Ordovician Graphitic slates Devonian Acadian Orogeny</td>
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<td></td>
<td>Doll and others (1961)</td>
<td>Quartz dioritic-granitic gneiss, massive quartz monzonite and granite ~616 m.y. (Rb-Sr, Stony Creek granite, N 450 m.y. (Rb-Sr and U-Pb, gneiss and granite of Masscoma Dome) Ammonoosuc metavolcanics -Ordovician Graphitic slates to schists; quartzite, mica schists, calcsilicates -Ord-Sil-Dev Garnet to staurolite grade.</td>
</tr>
<tr>
<td></td>
<td>Thompson (1968)</td>
<td>Quartzite, feldspathic mica schist, marble, and pelitic schist (Glenarm Series) Late Precambrian Upper amphibolite facies (kyanite near cores) Devonian Acadian Orogeny</td>
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<td></td>
<td>Rodgers (1970)</td>
<td>Veined gneiss and Migmatites; Quartzite, feldspathic mica schist, marble, and pelitic schist (Glenarm Series) Late Precambrian. Upper amphibolite facies (kyanite near cores) ~425 m.y.</td>
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### 10. Shuswap Complex

**British Columbia**

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<td>Granitic gneisses, paragneisses, migmatites -2100 m.y. (Rb-Sr)</td>
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<tr>
<td><em>d. Valhalla Dome</em></td>
<td>Reesor (1965)</td>
<td>Veined granodiorite augen gneiss, leucogranitic gneiss; Massive granitic (granodiorite to leucogranite) rocks</td>
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### 11. Rinkian Mobile Belt

**W. Greenland**

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<tr>
<td>Central Metamorphic Belt, E. Greenland</td>
<td>Haller (1955; 1971)</td>
<td>Biotite and hornblende gneisses, with amphibolite bands and pods (Flyverfjord infracrustal complex) ~3000 m.y. (Rb-Sr) Migmatites</td>
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<td>Central Karelia USSR</td>
<td>Il'ina (1977)</td>
<td>Granite gneiss, granite ~1800 m.y.</td>
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<td>Bacao Complex, Quadrilatero Ferrifero, Minas Gerais, Brazil</td>
<td>Herz and others (1961)</td>
<td>Weakly foliated granodiorite ~2440 m.y. (K/Ar biotite)</td>
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<td>Norwegian Caledonides</td>
<td>Ramberg (1967)</td>
<td>Quartzofeldspathic gneiss ~1900 - 1800 m.y. (Rb-Sr)</td>
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<td>Eastern Finland and Southern Karelia</td>
<td>Eskola (1949)</td>
<td>Granitic gneiss, porphyritic granite, migmatites ~2800 m.y.</td>
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<td>12. Central Metamorphic Belt, E. Greenland Caledonides</td>
<td>Hendriksen and Higgins (1976)</td>
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<td>16. Central Karelia USSR</td>
<td>Il'ina (1977)</td>
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<td>17. St. Malo Complex, Massif Armorican, France</td>
<td>Brun (1975, 1977), Brown (1978)</td>
<td>Paragneisses, migmatites -2600 m.y.</td>
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<td>24. West-Central Kazakhstan USSR</td>
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<td>a. Ulu-Tau Massif</td>
<td>Pavlova (1967, 1972)</td>
<td>Granite, gradational into granite gneiss</td>
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<td>25. Mama region, Transbaikalia, USSR</td>
<td>Salop (1972)</td>
<td>Gneissic granite -Early Proterozoic Porphyritic granites</td>
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<td>26. Aldan Shield</td>
<td>Grabkin (1965); Salop (1972); Gladkov and Grabkin (1978)</td>
<td>Alaskitic granite</td>
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<td>27. Kodar-Udokan region, USSR</td>
<td>Leytes and Fedorovskiy (1972); Sorachev (1974)</td>
<td>Charnockite gneisses, granites, granodiorites -3100-2800 m.y. Granite-gneisses, gneissic granites (Kuanda granites)</td>
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<td>32. North Korea</td>
<td>Salop (1972)</td>
<td>Gneissic granites, gneisses, and schist -Archean-Early Proterozoic</td>
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<td>34. Mysore State, India</td>
<td>Pichamuthu (1967)</td>
<td>Radhakrishna and Vasudev (1977)</td>
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<td>36. Abbabis Complex, Namibia (South West Africa)</td>
<td>Smith (1965)</td>
<td>Jacob (1974)</td>
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<td>37. Rietfontein Inlier, Namibia</td>
<td>Martin (1965)</td>
<td>Gneisses, granites (Marienhof Fm)</td>
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<td>38. Copperbelt, Northern Zambia</td>
<td>Garlick (1961)</td>
<td>Snelling and others (1964)</td>
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<td>39</td>
<td>Mpande Dome, Southern Zambia</td>
<td>Mallick (1967)</td>
<td>Microcline-rich orthogneiss; granite</td>
<td>Lufilian Orogeny</td>
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<td>Metarhyolites, quartzose schists, psammites, phylolite, semipelitic and pelitic schists, marble (Katanga Series) -Upper Proterozoic Staurolite grade (amphibolite facies)</td>
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<td>40</td>
<td>Fungwi and Chimanda Reserves, Rhodesia</td>
<td>Talbot (1971)</td>
<td>Lower Paragneiss, upper acid gneiss - granitic to granodioritic</td>
<td>Zambezian (Lufilian) orogeny</td>
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<td>Quartzite, semi-pelitic and pelitic schists (Umkondo System) -2000-1650 m.y.</td>
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<td>41</td>
<td>Chirwa Intrusion, Rhodesia</td>
<td>Johnson (1968)</td>
<td>Sodipottassic porphyritic and non-porphyritic granite -Archean? &gt;2650 m.y.</td>
<td>Mozambique Orogeny</td>
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<td>Amphibolites (Bulawayan System, Archean); pelitic and semi-pelitic schists with quartzite and marble horizons (Umkondo system, &gt;1600 m.y.). Garnetiferous.</td>
<td>~500 m.y.</td>
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<td>42</td>
<td>Rum Jungle area, Northern Territory, Australia</td>
<td>Compston and Arriens (1968) Stephansson and Johnson (1976)</td>
<td>Paraschists, paragneisses, volcanics, amphibolite, migmatite, several varieties of granite -2550-2450 m.y. (Rb-Sr)</td>
<td>1830-1720 m.y.</td>
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<td>Quartzite, dolomite, algal reefs, hematite sandstone, arkose, quartz greywacke, conglomerate, siltstone (Bachelor Group); siltstone-mudstone, greywacke (Goodparla-Finniss River Groups) -Lower Proterozoic Lower greenschist facies (retrograde metamorphism from amphibolite facies).</td>
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PART IV: CROSS-SECTIONS OF MANTLED GNEISS DOMES
AND CORDILLERAN METAMORPHIC CORE COMPLEXES

The structure of both mantled gneiss domes and metamorphic core complexes has been conceptualized by many authors in cross-sectional drawings; a sampling of such cross-sections is presented here to enable comparison of the two structural types, as they have been envisioned by local geologists.

Four cross-sections of mantled gneiss domes, adapted from the cited authors are included:
1) the Late Proterozoic (Baikalide) Dzhelanzhin dome of the Mama region, Transbaikalia, USSR (Salop, 1972),
2) the Hercynian Agout dome, Montagne Noire, France (Geze, 1949),
3) the Pelham dome, Massachusetts, on the Acadian Bronson Hill anticlinorium (Thompson and others, 1968), and
4) the Chester dome Vermont, in the Connecticut Valley synclinorium, formed during the Acadian orogeny (Doll, 1961).

Four cross-sections of Cordilleran metamorphic core complexes are also included. These have been modified from the sources cited on the basis of recent geological reconnaissance by S.J. Reynolds.
1) South Mountains complex, Arizona (Reynolds and Rehrig, 1980),
2) the Bitterroot frontal zone and underlying Idaho batholith, Montana and Idaho (Lindgren, 1904),
3) the Rincon Mountains and Happy Valley area, Arizona (Drewes, 1974, 1978), and
Bitterroot Complex, Idaho and Montana
RAFT RIVER COMPLEX, UTAH
CHAPTER 3

METAMORPHIC CORE COMPLEXES--STRUCTURAL CHARACTERISTICS, KINEMATIC EXPRESSION, AND RELATION TO MID-MIOCENE LISTRIC FAULTING

By

George H. Davis

STRUCTURAL CHARACTERISTICS

Introduction

Cordilleran metamorphic core complexes are deformational systems within the Basin and Range province which owe their distinctive physical expression to superimposition of mylonitic tectonite fabrics and brittle denudational faulting. In the southern part of the western Cordillera, crustal rocks softened by prodigious Laramide magmatism were subjected to a shearing process which gave rise locally to the mylonitic tectonite development. The shearing fashioned thick, regionally continuous, gently dipping zones of lineated tectonite from rocks as young as 55 m.y., but no younger than 25 m.y. (Anderson and others, 1977, 1980; Shakel and others, 1977; Keith and others, 1980; Reynolds and Rehrig, 1980). During the time that the tectonite was forming, large fault-controlled basins developed, and these received growth-faulted accumulations of continental redbeds and volcanics (Davis and Coney, 1979). Regionally extensive mid-Miocene listric normal denudational faulting was superimposed on tectonite and non-tectonite alike, and in such a way that transport along the faults was aligned strictly parallel to lineation in underlying or nearby tectonite exposures.
Comprehensive descriptions of the geologic properties of individual 'Cordilleran metamorphic core complexes' are presented in a memoir by the same name (Crittenden and others, 1980). In this chapter an attempt is made to describe the composite properties of these deformational systems, especially in the context of the regional system of Cenozoic deformation which they comprise. Defining what constitutes the characteristics of metamorphic core complexes is a scale-dependent process. Viewed as regional tectonic elements along the length of the western Cordillera from Sonora to southern Canada, the metamorphic core complexes might be considered 'small' outcrop areas of relatively high topographic relief which comprise arches of distinctively deformed and metamorphosed igneous and sedimentary rocks. The deformed crystalline rocks are separated from unmetamorphosed country rocks by discontinuities (décollement) marking strikingly sharp thermal-strain gradients. Resting on these discontinuities are 'tiny' thin plates of deformed but generally unmetamorphosed rocks. Generally the isotopic systems of core complex rocks disclose a mid-Tertiary thermal disturbance. The complexes are commonly co-spatial with Oligocene-Miocene continental sedimentary sequences and Miocene ignimbritic volcanic rocks.

At the mesoscopic scale, the array of structures in metamorphic core complexes is awesome in its diversity and pervasiveness: tectonites contain low-dipping mylonitic foliation, systematically aligned low-plunging mineral lineation, slickenside striae, normal ductile faults, abundant boudins and pinch-and-swell features, and recumbent to overturned tight pytgmatic folds in aplites and pegmatites. Zones of contact between compositionally (and mechanically) dissimilar tectonites are frequently marked by folded mylonitic schist. Tectonites of marble and quartzite show extreme thinning and attenuation. The common array of structures includes tight isoclinal recumbent to overturned intrafolial folds, axial-plane cleavage, boudins, pinch-and-swell, flattened-stretched pebble metaconglomerate, lineation, and ductile faults. The tectonite fabric is generally overprinted by pervasive fractures. A décollement zone typically marks the uppermost limit of tectonite forming a planar to curviplanar zone of microbrecciated mylonitic tectonite and cataclasite. Within the zone, original foliation and lineation are rotated, overprinted, and locally masked completely by cataclastic granulation. Above the décollement zone, there may be modest amounts of mylonitic tectonite, but generally there are unmetamorphosed but faulted cover rocks characterized by internal folding, bedding-plane cleavage, and faults and shear zones. Although in normal stratigraphic succession, formations represented in the cover rocks are markedly tectonically thinned.
Terminology

In this assessment of structural and tectonic characteristics of metamorphic core complexes, the major components of the deformed terranes are referred to as (1) mylonitic tectonites, (2) décollement zones, and (3) detachments (Fig. 3-1). The term tectonite underscores the fact that all rocks within the zone, regardless of age or composition of protolith, were affected by penetrative slip or flow on newly generated closely spaced foliation and fracture discontinuities. 'Tectonite' places emphasis on the deformed state of these rocks, not simply the metamorphosed state. Misch (1960) recognized the power of the term when he applied it to the infamous marble tectonite below the Snake Range décollement in eastern Nevada. Regional mapping of the tectonites has slowly revealed that they give way downward to country rock from which the tectonite was derived (Compton and others, 1976; Davis, 1977, 1980; Coney, 1980; Davis and Coney, 1979; G.A. Davis and others, 1979; Reynolds and Rehrig, 1978, 1980). This relationship signals the fact that the zone of tectonite is produced at the expense of rocks which are part of the normal upper-crustal geological column for the region. It also reveals that the tectonite is not part of wholesale ductile shearing of all immediately underlying upper crustal rocks. Indeed, the zones are curvitabular, hundreds to thousands of meters thick, and continuous for kilometers (Crittenden and others, 1980).

There are three main field occurrences of mylonitic tectonite. The dominant one is 'tectonite gneiss' (Fig. 3-2a), composed of protomylonitic augen gneiss of granitic to quartz monzonitic composition. In southern Basin and Range examples, the tectonite gneiss is largely derived from granitic basement (Creasey and Theodore, 1975; Banks and others, 1977; Shakel and others, 1977; Davis 1977, 1980; Rehrig and Reynolds, 1977, 1980; G.A. Davis and others, 1980; Reynolds and Rehrig, 1980). Tectonite gneiss is pervaded by a gently dipping and macroscopically smoothly curved foliation which gives expression to awesome dome-like massifs, referred to by some as gneiss domes or magmatic blisters. The Rincon Mountains east of Tucson, Arizona is a good example (Davis, 1980), but an even better expression of this is Mazatan Mountain, east of Hermosillo, Sonora, where a Laramide batholith is partly converted to tectonite gneiss (Anderson and others, 1980).

A second common field occurrence is 'tectonite schist' from phyllonitic overprinting of metavolcanic and/or metasedimentary, schistose protolith (Fig. 3-2b). This class of tectonite schist gives way downward to regionally metamorphosed, foliated, but non-overprinted rocks. Structural kinematics are more difficult to unravel in such terranes as these where two (or more) penetrative deformations have been superimposed. Such tectonites are exemplified by overprinted Jurassic (?) metasedimentary and
Figure 3-1 Schematic diagram showing structural subdivisions of metamorphic core complexes.
Figure 3-2. Chief varieties of metamorphic core complexes. Varieties distinguished on the basis of protolith for mylonitic tectonites. Mylonitic tectonite is derived at the expense of (A) Precambrian granite and Tertiary quartz monzonite, (B) Precambrian Pinal schist (or Mesozoic schist), and (C) younger Precambrian and Paleozoic sedimentary rocks.
Figure 3-3  Recumbent fold in calc-silicate and marble layers within tectonite carapace. Rincon Mountain area, Arizona
metavolcanic rocks in the Papago Indian Reservation west of Tucson (Davis, 1977, 1980) and in northern Sonora (Anderson and others, 1977, 1980). They are also represented by tectonites formed at the expense of older Precambrian Pinal Schist in places like the Rincon Mountains near Tucson (Drewes, 1978; Davis, 1980) and at South Mountain near Phoenix (Reynolds and Rehrig, 1980).

The third variety of tectonite, and perhaps the most spectacular is derived from unmetamorphosed sedimentary strata and is classically exemplified by the marble tectonite of Misch (1960). This kind of rock is here referred to as 'tectonite carapace', for it typically is expressed by a relatively thin curvitabular, strongly foliated sheet which is plated concordantly to underlying tectonite gneiss or non-tectonite protolith (Fig. 3-2c). Internally it is spectacularly folded (Fig. 3-3). Dominant lithologies are marble, quartzite, phyllite, and stretch-pebble conglomerate. The marble tectonite in the Snake Range of eastern Nevada is plated atop tectonite gneiss derived from Jurassic granite and tectonite quartzite (a part of the carapace) derived from Eocambrian-Cambrian Prospect Mountain Quartzite. The best examples in Arizona are in the Rincon and Coyote Mountains (Davis, 1975, 1977, 1980; Frost 1977; Gardulski, 1980).

Above the mylonitic tectonite there are detachments containing rocks which have not been affected by the tectonite overprint (Fig. 3-1) (Davis, 1977, 1980; Coney, 1979; Davis and Coney, 1979). These lie above the décollement zone. Rocks in the detachments range in age from Precambrian to Miocene, and they are not exotic to the country rock sequence. The upper surface of the décollement zone is a 'planar' fault (décollement). At the scale of a mountain range the décollement may be curviplanar, 'arched' in sympathy with the gross structure of the underlying tectonite foliation (Pashely, 1966; Drewes, 1975, 1978; G.A. Davis and others, 1979, 1980). Large mullion structures are locally seen at the décollement zone. Below the décollement there is characteristically a conspicuous meter-thick ledge of crushed and granulated but strongly indurated, fine to very fine-grained cataclasite. It is referred to here as microbreccia (Fig. 3-1), and its composition reflects the nature of the underlying tectonite. For example, quartzo-feldspathic microbreccia overlies tectonite gneiss derived from granitic rocks; marble microbreccias overlie tectonite marble derived from Paleozoic carbonate (Trevor, M.S. thesis, in prep)

Below the microbreccia, but still within the décollement zone, there are rocks which are excessively faulted, fractured, rotated, comminuted, and altered. These zones display crudely tabular, lens-shaped to wedge-like forms, ranging in thickness from a few meters to at least 50 meters. Until recently these
rocks have been thought of as restricted to décollement zones overlying tectonite gneiss. In such settings the rocks have been described as chlorite breccia (Rehrig and Reynolds, 1977, 1980; Davis, 1980; Davis and Coney, 1979) or, less fashionably, as 'junk-rock' (G.H. Davis and students). Here they are referred to as microbrecciated mylonitic tectonite (Fig. 3-1). The highly fractured, fine-grained, blue-green rocks of this zone weather to light brown/pale green shattered outcrops. In places the original tectonite gneiss fabric is so strongly masked that it cannot be recognized. Where mapped carefully (Reynolds and Rehrig, 1980; Davis and Gardulski, in prep.), the orientation of relict tectonite foliation is seen to be rotated orthogonally to lineation in underlying mylonitic tectonite, and with foliation dips as steep as 90°. The internal structure is suprisingly systematic, even though the rocks are the most deformed within metamorphic core complex terranes. Rocks of the décollement zone do not slowly grade downward into tectonite gneiss, but are separated from gently dipping lineated tectonite gneiss by low- to moderate dipping fault zones. The overall form and geometry suggests that these rocks are made up of imbricate detachments of mylonitic tectonite. Trevor (M.S. thesis, in prep.) has observed in the eastern Rincon Mountains that such rocks underlie an extensive ledge of marble microbreccia and consist of anomalously highly fractured, faulted, rotated, and brecciated tectonite marble. Only the deep blue-green chloritic alteration is absent.

The detachments (Fig. 3-1), in their simplest form, consist of Oligocene-Miocene strata which are rotated to moderately steep or steep dips and rest in low-angle fault contact on microbreccia of the décollement zone (Pashley, 1966; Davis, 1975, 1980; Drewes, 1975, 1978; Davis, G.A., and others, 1979). Bedding strike of the detachment strata is typically oriented at right angles to the trend of lineation in mylonitic tectonite (Rehrig and others, 1980; G.A. Davis and others, 1979). The more complicated and challenging detachments contain pervasively shattered Precambrian granite and spectacularly folded Paleozoic/Mesozoic strata, most of which is non-tectonite (Coney, 1974; Davis, 1975, 1980; Drewes, 1975, 1976, 1978). The décollement zone, then, separates rocks of sharply contrasting mechanical properties and deformational characteristics. It is the contrast between rocks above and below the décollement which has been so puzzling to explain structurally and tectonically (Thorman, 1977; Davis, 1977). Simply stated, how were the tectonite formed without affecting rocks in the detachment plates, whose distance of tectonic transport cannot be very far?? In addressing this question it must be recognized that detachments of the type described lie both within and outside of metamorphic core complex terranes in the southern Cordillera. For example, in the Black Canyon region south of Lake Mead (Anderson, 1971, 1978) and in the San Pedro Valley northeast of Tucson (Creasey, 1965; Cornwall
and Krieger, 1975; Krieger, 1974; Scarborough and Peirce, 1978),
detachments of Oligocene and Miocene strata rest in low-angle
découlrement-like fault contact on non-tectonite granitic basement,
with the closest outcrops of lineated mylonitic tectonite kilo-
meters to tens of kilometers away.

Descriptions

Tectonite Gneiss

By far the greatest volume of rocks in the metamorphic core
complexes comprise tectonite gneiss. Tectonite gneiss underlies
exposed surface areas of more than 500 km$^2$ within single complexes,
and expresses an exposed vertical relief of as much as 200 m.
Tectonite gneiss undergirds some of the highest summits in the
Basin and Range province. Mylonitic foliation in the complexes
is characteristically low-dipping and commonly defines large
upright, doubly plunging foliation arches or antiforms, half-arches,
upright synforms or troughs, and irregular amoeboid domical struc-
tures. Some of the doubly plunging arches have exceptional physio-
graphic expression (Fig. 3-4). In general, penetrative mineral
lineation within tectonite gneiss is co-axial with axes of arching;
yet there are a number of examples where arching resulted in clear
rotation of the original unidirectional penetrative lineation.

Tectonite gneiss in the southern part of the western
Cordillera
is derived in part from medium to coarse-grained Precambrian granu-
lar rocks, approximately 1.4 to 1.6 b.y. old (Creasey and Theodore,
1975; Banks and others, 1977; Shakel and others, 1977; Davis and
others, 1979). In outcrop, the tectonite gneiss thus derived is
relatively dark colored and cut by the penetrative flat to moder-
ately dipping mylonitic foliation (Fig. 3-5). Condordant aplite
and pegmatite veins serve to enhance the foliation, and such sills
and veins commonly display abundant boudinage and pinch-and-swell.
Aplies and pegmatites that cut the foliation at high angles are
almost always pytgmatically folded. These folds are axial planar
with respect to low-dipping penetrative foliation and, thus, are
typically strongly overturned to recumbent.

In addition to mylonitic tectonite derived from Precambrian
granitic rocks, abundant tectonite is derived from quartz monzonite,
garnet-diorite, granite, and alaskite of Phanerozoic age. Many
textural and compositional varieties of these rocks exist, but most
tend to be medium-grained and quartz monzonitic in composition.
Garnet-bearing two-mica granites are not uncommon. As in the case
of the coarse-grained tectonite gneisses, a distinguishing struc-
tural characteristic of these rocks is low-dipping cataclastic
foliation and penetrative lineation. Deformational fabrics with-
in these rocks range from barely recognizable to mylonitic and
schistose. Foliation is typically defined by thin laminae of quartz, parallelism of micas, and aligned augen of cataclastically deformed feldspar. The rocks are cut by abundant pegmatite, aplite, and flat joints. The striking physiographic expression of the penetrative low-dipping foliation and jointing calls attention to these core-complex tectonites from afar.

Establishing the exact age of tectonite-imprinted Phanerozoic plutons has been difficult, for the rocks are characterized by profoundly disturbed Rb/Sr isotopic systems. K/Ar and fission-track ages typically range from 30 m.y. to 24 m.y. for these tectonite gneisses and associated pegmatites and aplites. U/Pb age-determinations on zircons from tectonite gneiss in northern Sonora (Anderson and others, 1977) and southern Arizona (Shakel and others, 1977) reveal that the lineation and foliation-forming process was operative after 55 m.y.b.p. Keith and others (1980) provide a useful summary of the geochronological problems and offer some new interpretations. It is becoming increasingly well documented that the core-complex terranes contain both mid-Tertiary and pre-mid-Tertiary plutons. Recent isotopic data suggest that even some of the mid-Tertiary bodies have been affected by low-dipping foliation and penetrative lineation (Reynolds and Rehrig, 1980). The Tertiary event, whatever its dynamic nature, imposed its thermal-tectonic signature on tectonite gneiss plutons, regardless of whether they existed before the event or were emplaced during the event.

The tectonite gneisses everywhere display low-plunging mineral lineation of a mylonitic nature (Fig. 3-6). The lineation generally is penetrative on the scale of hand specimen, with a delicacy of development dependent in large part on grain size and texture of host rock. The lineation has a variety of forms, but generally is expressed in the plane of foliation of the augen gneisses by the alignment of long directions of inequant feldspar and quartz-feldspar augen, striae, elongate aggregates and streaks of crushed minerals, and crenulations on quartz ribbons. Additionally, lineation occurs on concordant or discordant low-dipping aplite and pegmatite layers, mylonitic zones, transposition foliation surfaces, and low-angle normal faults. Even in some equigranular hypidiomorphic quartz monzonitic bodies lacking penetrative lineation, the lineation locally occurs on shallow-dipping aplite and pegmatite layers which have served to localize tectonic movements.

Within any given terrane of metamorphic core complexes, lineation is generally remarkably systematic in orientation. In southern Arizona and northern Sonora, with few exceptions, lineation trends N50-70E (Fig. 3-7). In eastern Nevada, western Utah, and southern Idaho, lineation trends systematically west-northwest. In Washington, lineation trends approximately east-west.
Figure 3-4  Northeast-directed, U-2 photo showing distinctive physiographic expression of the Santa Catalina and Rincon Mountains. The city of Tucson lies in the left foreground. High mountain in far-right background is the Pinalêno Mountains, part of which is also marked by metamorphic core complex deformation.
Figure 3-5  Penetrative, low-dipping foliation in mylonitic gneiss derived from Precambrian granite. White augen are feldspar porphyroclasts enveloped by laminae of quartz, biotite, and feldspar. Santa Catalina Mountains, Arizona.
Figure 3-6. Penetrative lineation in mylonitic gneiss. Rincon Mountains, Arizona.
Figure 3-7. Stereographic projections of orientations of penetrative lineation measured within metamorphic core complexes in southern Arizona. (A) Papago domain, west of Tucson. (B) Catalina domain, Tucson and northwest. (C) Pinaleno domain, east of Tucson in the Pinaleno-Jackson Mountain area.
Ductile normal faults are associated with the mylonitic, lineated, tectonite gneiss, and typically result in impressive local thinning. The ductile normal faults are always oriented at right angles to lineation, regardless of absolute orientation. Folds which naturally arise from the ductile faulting are superimposed on intrafolial folds in the mylonitic tectonite. The folds assume the form of upright antiforms and synforms marked by thinning in the zones of maximum inflection. In some cases, the hinge zones of the folds display visible offset, and ductile flexing appears to be superseded by actual normal faulting.

Contact zones between different phases of tectonite gneiss are commonly marked by strongly foliated, folded rocks referred to here as mylonitic schist. These rocks are fine-to very fine-grained and range in color from brown to steel gray to black. White broken feldspar and rock chips, irregular in size and shape, commonly 'float' in the fine grained mylonitic matrix. Deformed concordant aplite and pegmatite layers and veins accentuate the foliation. Penetrative folding is characteristic of these rocks and presumably reflects movements that helped to shape the tectonic gneiss fabric. The folds are intrafolial, tight to isoclinal, overturned to recumbent structures whose axes lie in the low-dipping foliation. Orientations of fold axes range from broadly dispersed to nearly coaxial with lineation. Foliation surfaces are not always marked by the mesoscopically recognizable mineral lineation; indeed, where found, the lineation looks overprinted.

Tectonite Schist

In many core complex terranes in the western Cordillera, the mylonitic tectonite overprint has been superposed on schists. For example, in southern Idaho, in the Albion Range, mylonitic tectonite has been derived in part from schistose rocks metamorphosed during the Mesozoic (at about 150 m.y.b.p.). In southern Arizona, tectonite schist was produced from mylonitic deformation of older Precambrian Pinal Schist (1.6 to 1.8 b.y.), and Jurassic metasedimentary and metavolcanic rocks. The resultant tectonite schist is anomalously highly folded and lineated. Original moderate to steeply inclined foliation is transposed about the new, flat to gently dipping foliation and shear surfaces. Folds are generally strongly overturned to recumbent, and often reclined. Superimposed folding is locally evident. Evaluation of these tectonite schists requires extensive comparative analysis of metamorphism and fabric regionally, in- and outside of the metamorphic core complexes. To date, this variety of tectonite has not been studied in much detail, but it will undoubtedly serve as a focus for structural and petrological analyses in the near future.
Tectonite Carapace

In southern Arizona metamorphic core complexes, tectonite carapace is derived from unmetamorphosed younger Precambrian and lower Paleozoic metasedimentary rocks. Such tectonites are commonly metamorphosed to upper greenschist and amphibolite grade, and form a relatively thin, tabular sheet of marble, quartzite and phyllite, concordantly welded or plated to underlying crystalline rocks. There are many outcrops where not even a crack separates younger Precambrian quartzite or marble of the tectonite carapace from underlying, tectonite gneiss. The tectonite carapace generally rests in low-angle contact on either (1) medium-grained quartz monzonitic tectonite gneiss of Tertiary (?) age or (2) moderately deformed coarse-grained augen gneiss derived from 1.4-1.5 b.y. porphyritic quartz monzonite (Banks, 1977). Thin tectonic slices (<10 m) of the mylonitic gneiss occur from place to place within the lowermost part of the tectonite carapace. On the mesoscopic scale, the base of the tectonite carapace is planar and strictly concordant with the foliation in the gneiss immediately below. Viewed macroscopically, the surface is smooth and gently warped into systematic upright antiforms and synforms. Rarely, the surface is deformed into surprisingly tight antiforms that project upward in almost tent-like fashion. The vertical relief on such sharp, tight structures is at least 10-20 m. At some locations, granite gneiss below the contact displays tight to isoclinal recumbent folds.

Rocks of the tectonite carapace, wherever found, display very distinct structural characteristics (Davis, 1975; Schloderer, 1974; Waag, 1968; Frost and Davis, 1976; Frost, 1977). Strongly overturned to recumbent folds are ubiquitous (Fig. 3-8). Protoliths for the phyllite, schist, marble, and quartzite have been deformed by transposition into tectonites characterized by intrafolial, commonly rootless tight to isoclinal folds. Depending on mechanical characteristics of the original rocks, the folds may form by passive flow, passive slip, and/or flexural flow. The passive folds typify homogeneous domains of quartzite on marble. The most outstanding flexural forms occur in interlayered calc-silicate and marble. Where the rocks are dominantly marble containing only thin brittle calc-silicate or quartzite struts, the 'competent' layers are typically distended, attenuated, and transposed. Fragments of the originally continuous layers form boudins (Fig. 3-9), isolated fold hinges and assorted tectonic inclusions in the marble matrix, (Fig. 3-10). In the plane of foliation and lithologic layering of calc-silicate and marble, gash fracturing and orthogonally disposed mineral lineation are locally strongly developed.
Conglomerates within the tectonite carapace are transformed into subhorizontally flattened quartzite-pebble units. The flattened pebbles are typically profoundly elongated parallel to the lineation in underlying augen gneiss. Where the pebbles are flattened but not elongate, underlying cataclastically deformed augen gneiss is foliated but not lineated. In the Tortolita Mountains, Davis and others (1975) measured the orientation and dimensions of 86 flattened elongate quartzite pebbles within the Barnes metaconglomerate and compared these to undeformed clasts at a nearby locality in the Santa Catalina Mountains. Axial ratios were computed to be 9:2:1; the plane of flattening is subhorizontal and the direction of elongation, N58°E, is identical to that of the adjoining tectonite gneiss.

Foliation and lithologic layering within the tectonite carapace are generally shallow dipping and strongly developed. Axial planes of tight to isoclinal, overturned to recumbent folds are generally parallel to foliation and layering. The folds are commonly reclined with respect to layering and foliation. Seen on the scale of large single outcrops, the foliation and layering are marked by pinch-and-swell and boudinage, with graceful gentle changes in attitude. Predictably, boudinage and pinch-and-swell are best developed where rocks of contrasting ductilities are juxtaposed.

Individual formations within the tectonite carapace are arranged in normal stratigraphic order, but they are generally tectonically thinned or locally thickened. Specific estimates of the change in thickness are very difficult to make because of the uncertainties inherent in correlating the strongly deformed lithotectonic units with southeastern Arizona stratigraphy. In many places the thinning appears to exceed 75 percent.

Thinning and thickening within the younger Precambrian and Paleozoic sequences have been achieved by passive flowing including transposition (Frost, 1977) of units rendered ductile during the deformational process. On the mesoscopic scale, the mode of thinning is explicitly displayed in the stretched/flattened pebble metaconglomerates, passively folded marble and calc-silicate rocks, transposed schists, and dismembered, attenuated quartzite layers in ductile matrix. Detailed mapping reveals interesting macroscopic adjustments to the 'thinning process,' mainly a heterogeneous distribution of lithologic units of contrasting mechanical properties. Map patterns demonstrate lateral movement and concentration of ductile materials as a response to thinning, but carried out in such a way so as always to bring young rocks over older.
Figure 3-8. Isoclinal fold in tectonite carapace. Folded layers are calc-silicate and marble within Paleozoic rocks. Rincon Mountain area, Arizona.
Figure 3-9. Boudinage showing normal-slip on lozenge-shaped blocks within metamorphosed Paleozoic rocks. Rincon Mountain area, Arizona.
Figure 3-10. Tectonite carapace showing distended and broken 'competent' layers within a matrix of marble. Protolith was Paleozoic strata. Rincon Mountain area, Arizona.
Figure 3-10
The structural fabric of the tectonite carapace is intimately coordinated with that of the underlying tectonite gneiss. Fabrics of both structural units are marked by profound flattening perpendicular to subhorizontal layering and foliation and by extension parallel to lineation as denoted by boudinage, deformed pebbles, and the ductile normal fault zones. Davis and others (1975) analysed structures in the Tortolita Mountains and showed that the lineated surfaces within fine- to medium-grained quartz monzonite augen gneiss of the tectonite core accommodated vertically directed flattening and profound east-northeast extension; and that the fold and stretched-pebble fabric in the immediately adjacent tectonite carapace formed as a response to the same deforming process. Critical to that analysis was recognition of the coordinated nature of the augen gneiss fabric to that of the tectonite carapace: the modal long axis of pebbles is strictly parallel to lineation in adjacent augen gneiss; the fold axes in the tectonite carapace parallel the lineation orientation; ductile normal faults are orthogonal to lineation. On the basis of the strict compatibility of the normal-slip ductile and brittle faults to inferred principal strain directions inferred from the stretched-pebble data, the 'stretching' of the quartzite pebbles and the development of normal-slip faults were interpreted as the early and late stages respectively of a ductile-to-brittle continuum of extensional deformation. During this deformation no significant shift in principal strain directions took place.

Fold axes in rocks of the tectonite carapace are generally difficult to evaluate in regard to slip-line direction. The fundamental problem is establishing for certain that specific asymmetric folds are indeed first-order folds within the transposed sequences. Furthermore, the fold axes in the metasedimentary rocks generally display a broad range of orientation within the plane of slip or flow. For now it is most important to emphasize that (1) within the array of variably oriented overturned to recumbent folds, reclined folds are commonly the preferred mode, (2) the axes of reclined folds tend to be co-axial with lineation in ductile rocks like marble, (3) mineral lineation in marble is essentially orthogonal to penetrative gash fracturing in mechanically suitable lithologies in the metamorphic carapace, and (4) lineation in underlying tectonite gneiss is parallel to that in marble within the tectonite carapace.

The above observations and inferences are consistent with the interpretation that both the tectonite gneiss and carapace were affected by flattening and extension, and that these processes resulted in profound thinning through flow of the ductile, originally bedded, carapace strata. Flow in the tectonite carapace was parallel to mineral lineation, and during progressive deformation fold hinges rotated partly or wholly into alignment with the
flow direction. The degree of rotation was partly related to the mechanical properties of the deforming sequence.

Décollement Zone

A décollement separates detachments above from tectonite below. The surface separates rocks of remarkably contrasting deformational styles. Most often in southern Arizona examples, the décollement marks the top of the tectonite gneiss and the base of the non-tectonite "cover"; tectonite carapace is generally absent. Below the décollement surface, a 'décollement zone' is usually present and consists of an upper ledge of strongly indurated microbreccia, below which is a thick (up to 100 meters + ) zone of microbrecciated, faulted, and rotated tectonite gneiss. Although the contact between the cataclasite and overlying detachment rocks is sharp, planar, and conspicuous, the contact between microbrecciated tectonite gneiss and underlying 'normal' tectonite gneiss appears gradational and ill-defined. The décollement zones occur only on one or two flanks of individual metamorphic core complexes.

Décollement zones commonly display striking 'younger on older' fault relations involving tens to hundreds of meters of stratigraphic separation. They typically separate tectonite gneiss derived in part from Precambrian rock, from non-tectonite Precambrian, Paleozoic, Mesozoic or Tertiary cover rocks. This array of structural and petrologic characteristics has prompted many workers to interpret them as thrust faults (Thorman, 1977; Drewes, 1978).

One of the best examples is the Catalina Fault, a décollement zone in the Santa Catalina and Rincon Mountains of southern Arizona. It crops out along a sinuous trace, tens of kilometers in length, on the south and west flanks of the complex. The décollement zone separates tectonite gneiss (below) from a variety of deformed but generally unmetamorphosed cover rocks, including Paleozoic limestone, sandstone, and shale, Mesozoic shale, dolomite, and conglomerate, and Oligocene-Miocene red beds. The dip of the décollement zone is generally less than 15° or 20°. The exposures are commonly confined to the pediment/mountain interface; at no place is the zone known to crop out at a high level on the mountain flank.
Figure 3-11. The Catalina fault: décollement as separating mylonitic gneisses (below) from unmetamorphosed, non-tectonite Precambrian granite and Paleozoic sedimentary rocks (above). Rincon Mountains, Arizona.
Viewed at the scale of the complex, the décollement zone forms a smoothly arcuate surface conforming to the macroscopic structural geometry of the tectonite gneiss. Viewed at the mesoscopic scale, the zone may be concordant or discordant to mylonitic foliation in underlying tectonite gneiss.

In outcrop, the extremely altered rocks of the décollement zone weather brown to brownish green, but on fresh surfaces they are seen to be bright blue-green chloritic, fine to medium grained microbreccias. They are pervasively overprinted by shattering along closely spaced fractures. Within the highly deformed rock suite, overprinted lineation and foliation fabrics can still be recognized; but as will be discussed later ('Kinematic Expression'), these are rotated into moderately or steeply dipping attitudes. Clearly, the microbrecciated tectonite gneiss was produced at the expense of mylonitic gneiss, and the microbreccia ledge at the expense of the microbreccia mylonitic gneiss.

The transformation from tectonite gneiss to microbreccia is remarkable. Gardulski (from Davis and Gardulski, in prep.) reports that 'normal' tectonite gneiss derived from Precambrian quartz monzonite consists of large (up to 4 cm) porphyroclasts of rounded to elliptical shapes in a matrix of mainly quartz and feldspar. The feldspar porphyroclasts show pull-apart structures with infilling by quartz and chlorite. Quartz occurs in several modes:- feldspar-free zones of recrystallization in which quartz is undulatory and very fine grained; zones of flattened, elongate, extremely undulose quartz (L:T=20:1) adjacent to feldspar porphyroclasts; and zones of polygonal quartz in 'pressure shadows' next to feldspar augen. Fracturing is not conspicuous and alteration of feldspar is minor in the 'normal' mylonitic gneisses. In contrast, the extremely microbrecciated mylonitic gneisses are marked by extreme fracturing, brecciation, microfaulting, and alteration. The foliation visible in normal gneiss is nearly obliterated by anastomosing bands of microbreccia which cut through the rock. Microfaulting has caused rotation and tilting of micro-fault blocks of up to 60°70°. Some parts of this kind of rock are so mylonitized that only a dense aphanitic mass remains. Veinlets of chlorite, epidote, biotite, and opaque minerals are abundant and lace the rock in all orientations. Grain size is greatly reduced. Feldspar chips are as large as 0.3 mm, but the average size is 0.01 mm.
Detachments

The non-tectonite detachments which overlie tectonite in the metamorphic core complexes are interesting and instructive in their gross configuration and interval structure. For the most part, the detachments directly overlie well developed décollement zones of the type already described. Cover rocks above the décollement zone (Catalina fault) in the Santa Catalina and Rincon Mountains form thin plates that include slices of Precambrian, Paleozoic, Mesozoic, and Tertiary formations. Along the front of the forerange in the Santa Catalina Mountains, the zone separates tectonite gneiss from Oligocene-Miocene Pantano Formation, a sequence of red beds, notably mudstone, siltstone, sandstone and conglomerate.

Pantano beds locally rest on the décollement zone on the south and west flanks of the Rincon Mountains; additionally Precambrian rocks and Paleozoic and Mesozoic strata lie atop the zone as well (Drewes, 1978). Completeness (incompleteness ?) of stratigraphy in these sections is highly variable. Rocks on the west side of the Rincon Mountains within Saguaro National Monument (East) consist of Precambrian granite and schist, upper Paleozoic limestone, dolomite, and shale, and Tertiary red beds. Along the southeast flank of the Rincon Mountains, Cretaceous shale and limestone with interbedded siltstone, locally lie directly on the décollement. The thickness of the sheet is less than 90 m. At the southeasternmost corner of the Rincons, a 75 m sheet of Paleozoic rocks rests on the décollement zone. Although formations from Cambrian to Permian are represented, the thickness of the sequence is less than 10% of the full Paleozoic section.

East of the Rincon Mountains, the décollement is overlain by detachments of Precambrian granite and younger Precambrian and Paleozoic sedimentary strata, in a stacking which dips homoclinaly (Drewes, 1975; 1976a). The granite is shattered by closely spaced fractures and faults, and it rests directly on tectonite marble and quartzite derived from Younger Precambrian and Paleozoic strata. Enigmatically, specific formations which are strongly deformed tectonites below the décollement crop out within 100 m (?) of their non-tectonite, unmetamorphosed protolith counterparts in the overlying detachment (s).
Structures in the non-tectonite detachments display a wide variety of physical expressions. Precambrian granitic basement rocks form tabular to wedge-shaped, shattered masses. Structures in the sedimentary detachment strata are described by Coney (1974), Davis (1975), Compton and others (1976), and Davis and Frost (1976). Overturned asymmetric folds, detached isoclinal folds, and unbroken cascades of recumbent folds are locally abundant (Fig. 3-12). Most of the folds are transitional between ideal parallel and ideal similar folds and, thus, are characterized by some hinge-zone thickening (Davis, 1975). The scarcity of axial-plane cleavage, the abundance of bedding-plane cleavage, and the obvious influence of layering on the morphology of folds indicate that the folds evolved through slippage between layers and flow within layers.

The dominant structure in the Tertiary detachment strata is moderate to steep homoclinal tilting, and in such a manner that preserves an orthogonal relation between the strike of the beds and the trend of lineation in the closest exposed tectonite. This structural symmetry will be more fully discussed under 'Kinematic Expression'.

**KINEMATIC EXPRESSION**

**Introduction**

Interpretation of the kinematic significance of metamorphic core complexes affects our understanding of all aspects of Mesozoic and Cenozoic evolution of the southern part of the western Cordillera. Historically, the physical signatures of metamorphic core complexes have been interpreted in the context of Sevier-Laramide compressional deformation, especially overthrusting (Misch, 1960; Roberts and Crittenden, 1973; Drewes, 1976, 1978; Thorman, 1977). Most recently, DeWitt (1980) has reaffirmed the opinion that Mesozoic metamorphism and Sevier-Laramide compressional deformation were fundamental in shaping the metamorphic core complexes. Yet the geochronologic facts assembled thus far in Sonora (Anderson and others, 1977, 1980) and southern Arizona (Shakel and others, 1977; Keith and others, 1980; Haxel and others, 1979) demonstrate that mylonitic tectonite of core complex affinity formed at the expense
Figure 3-12. Folded detachment strata. Marked layers are limestone beds of Pennsylvanian-Permian age. Rincon Mountains, Arizona.
of granitic plutonic rocks as young as 55 m.y., and perhaps much younger (Reynolds and Rehrig, 1980). Such deformation post-dated the cessation of Laramide compressional deformation and thrusting (Drewes, 1972a, 1972b, 1976b, 1978; Coney, 1978; Rehrig and Heidrick, 1972, 1976). The development of mylonitic tectonite had ceased before mid-Miocene time. The tectonite was at least in part overprinted by brittle fabrics during emplacement of upper-plate detachments as late as mid-Miocene (Davis and others, 1979, 1980; Anderson, 1971; Shackelford, 1977, 1980; Scarborough and Peirce, 1978). The structural facts suggest that both the formation of mylonitic tectonite and the emplacement of the detachments are expressions of extensional deformation in the Tertiary (Davis, 1977, 1980; Davis and Coney, 1979; Coney, 1979, 1980). What remains vague are the dynamic transitions which interconnect classical Laramide tectonism, the formation of mylonitic tectonite, the development of décollement zones, the emplacement of upper-plate detachments, and classical Basin and Range tectonism.

The importance of interpreting the structural geometry and tectonic implications of metamorphic core complexes has emerged recently in a very practical way. Anschutz Corporation has framed a petroleum exploration model for southern Arizona which features interpretation by major regional overthrusting. They have interpreted subhorizontal reflecting horizons at depths greater than 2-3 km to be the expression of Paleozoic and Mesozoic strata beneath what has been considered autochthonous Precambrian basement. Drilling has been initiated at a site directly above a major culmination in the reflectors. The site is within 2 km of lineated mylonitic tectonites exposed in the Picacho Mountain metamorphic core complex terrane, and it lies directly on the projection of the culmination of the Rincon-Santa Catalina metamorphic core complex. Peter Coney and I have suggested that the subhorizontal reflecting horizons might be the subsurface expression of mylonitic tectonite, not sedimentary layering. The possibility exists that the reflection profiles of the Anschutz Corporation might be among the first published records of the 'roots' of metamorphic core complexes.

Interpretation

The kinematic significance of rocks in the zone of mylonitic tectonite has been difficult to explore because so much of the rock is tectonite gneiss derived from mylonitic reduction of granitic or quartz monzonitic plutonic bodies. Where thus deformed, the tectonite usually lacks fold structures which would otherwise help to disclose slip-line paths. Fortunately, the structures in tectonite carapace are kinematically and dynamically coordinated with that of the tectonite gneiss, thus expanding the basis on which the internal movement plan can be evaluated (Davis, 1980).
To date I have emphasized that the tectonite fabric is characterized by profound flattening and extension, as evidenced by a broad array of structures, especially boudinage, pytymatic folds, stretch-pebble conglomerate gash fractures, and ductile normal faults (Davis, 1977, 1980; Davis and Coney, 1979). Such a fabric system could evolve through pure shear and/or rotational simple shear during progressive deformation. Determining systematic sense of relative slip if such exists emerges as crucial to unraveling the strain and kinematic significance of the zones of mylonitic tectonite.

Based on fold analysis in southern Arizona, systematic simple-shear displacement within zones of tectonite are recognized. In the Pinalêno Mountains near Safford, Arizona, the sense of relative slip is clearly southwest to northeast, as revealed by asymmetric overturned, intrafolial folds which verge down the dip of northeast-dipping tectonite foliation. The line of slip is N40°E, parallel to penetrative lineation in the tectonite. In the eastern Rincon Mountains within Happy Valley, folds in tectonite marble and quartzite show a broad range in axial orientation, but the slip-line direction is westerly. This was reported by Frost and Davis (1976) and by Frost (1978), yet interpretation of this movement sense has never been agreed upon in the context of existing regional tectonic models. Evidence for a southwesterly sense of simple-shear is evident in fold data collected in mylonitic tectonite of the Rincon–Catalina complex as well (see Peterson, 1968; Waag, 1969; Schloderer, 1976). In the metamorphic core complex terrane of the Coyote Mountains, west of Tucson, ductile normal faults and sparse minor folds in tectonite quartzite and marble indicate south to north simple shear. Further west, in the Sierra Blanca Mountains, tectonite schist contains abundant intrafolial folds which together reveal north-northwest to south-southeast directed simple shear. Additionally, the structural data demonstrate that the inferred slip-line and the direction of penetrative lineation are coaxial (Davis, 1977, 1980).

Above and beyond minor fold data, certain large-scale geological relationships suggest that simple-shear deformation produced the mylonitic tectonite. In the Santa Catalina Mountains north of Tucson, coarse-grained augen gneiss derived from 1.4 b.y. quartz monzonite is injected by 'laccolithic', 'lit par lit' medium-grained augen gneiss derived from early Tertiary plutonic rocks. Workers have generally assumed that this geometric geologic relationship is a preserved intrusive contact, overprinted by mylonitic deformation. An alternative interpretation is that the upper part of the Tertiary plutonic system, emplaced in Precambrian basement, was partially 'beheaded' by the southwest-directed simple shearing which seems to characterize the mylonitic tectonite of the Catalinas (Fig. 3 – 13). The kinematics of this model are identical to those described by Ramsay and Graham (1970).
Figure 3-13. Interpretive model that describes a sill-like form of deformed Tertiary quartz monzonite as a product of distributed simple-shear.
The implications of these emerging data are important. First, the mylonitic tectonite acquired its strain during progressive simple-shear rotational deformation and not by pure shear (although pure shear probably occurred locally). Second, the direction of penetrative lineation in the tectonite lies so close to most solutions of slip-line direction based on fold orientations that it seems likely that the mineral-lineation is indeed the slip-line direction. Third, dynamic models for the origin of the tectonites must be capable of explaining simple-shear couples which act in different senses and, locally, in different directions.

With respect to present attitudes, the mylonitic tectonites appear to be zones of ductile normal slip (flow). If so, these tectonites may be partial exposures of ductile normal shear zones of regional extent. Mapping the tectonites as parts of regional ductile shear zones, and mapping both the sense of simple-shear and the slip-line path are necessary to assess the tectonic significance of Cordilleran metamorphic core complexes as a whole. The plan view geometry of the system of regional ductile normal shear zones can be reconstructed in a general way by interconnecting the geographically isolated, partial exposures of tectonite. Subsurface seismic reflection data would perhaps make this job easier and more reliable.

Regional dip of individual subhorizontal zones in some cases can be evaluated through systematic assessment of which rocks within the country-rock column are converted to mylonitic tectonite. In this regard, the Rincon-Santa Catalina complex of southern Arizona is a very provocative example. All along the west and southwest margins of this complex, mylonitic tectonite is made at the expense of Precambrian basement and Tertiary(?) intrusions. The position of the ductile shear zone represented by mylonitic tectonite appears to be well below the great unconformity. Along the northwest-trending 'culmination'(?) of the complex, tectonite was largely produced through penetrative deformation of Younger Precambrian and lower Paleozoic sedimentary rocks, producing tectonite carapace. Further northeast, the zone of tectonite occupies an even higher position in the section, such that tectonite marble and quartzite produced from Paleozoic strata as young as Mississippian-Pennsylvanian rest on non-tectonite basement. Furtherest northeast, tectonite derived from Permian and Cretaceous strata rest on non-tectonite lower Paleozoic strata. This apparent migration of the shear zone up-section to the northeast may help explain what has been described as a fundamental asymmetry to metamorphic core complexes. The recognition of such regional dip of the zone of mylonitic tectonite contradicts Davis and Coney's
interpretation that the top of the zone of penetrative deformation hovers around the great unconformity. This indeed may be the case locally, but it does not appear to be the rule.

The structural characteristics which make the décollement zones so distinctive formed in the late stages of and/or after the formation of mylonitic tectonite. The microbrecciated mylonitic tectonite which makes up most of the décollement was produced during imbricate(?) faulting and rotation of the mylonitic tectonite. In parts of décollement zones where mylonitic tectonite is only modestly overprinted by microbrecciation and faulting, there is bimodality of strike orientation of relict tectonite foliation. For example, in the western Rincon Mountains, foliation is locally 'dragged' into parallelism with north-northwest striking fault and fracture zones. Where the strike of foliation is north-northwest, dip magnitudes are extraordinarily steep. In parts of the décollement zone it is radically changed in orientation. In the Rincon Mountains example, relict foliation strikes north-northwest and dips as steeply as 90°.

Faulting, rotation, and microbrecciation were carried out at a time and at a structural position such that no upper-plate detachment strata became interleaved with microbrecciated mylonitic tectonite in the décollement zone. In essence, the formation of the microbreccia capping ledge, the décollement, and the emplacement of the detachment rocks all post-dated the development of the sub-décollement detachments.

It is suggested here that the detachments in microbrecciated tectonite gneiss are the physical expression of coalescing, imbricate listric normal faults. Rotation of the mylonitic tectonite foliation is viewed as the natural response to normal displacement(s) of tectonite along curved fault surfaces. The steepest-dipping mylonitic tectonite denotes major displacements along a single fault zone and/or multiple movements along superimposed faults. At Saguaro National Monument (East) in the Rincon Mountains, certain low-dipping domains of microbrecciated mylonitic tectonite have relict foliation/lineation attitudes which suggest that the rock has been rotated during faulting perhaps by more than 90° (Davis and Gardulski, in prep.). The rotations may be described as operating around axes oriented perpendicular to lineation in underlying mylonitic tectonite. The problem which has emerged as critically important is determining the sense of fault movement within the décollement zones.
The process of imbricate listric(?) normal faulting must have had the effect of translating upper-level country rock downward and laterally toward the zone of mylonitic tectonite. This tectonic denudation process was probably in part concurrent with formation of mylonitic tectonite (Davis, 1977a). The position of the base of the décollement zone may have been thermally controlled, prescribed by the rheological characteristics of the mylonitic tectonite. Eventually the microbrecciated, rotated mylonitic tectonite was 'truncated' at a level now corresponding to the cataclasite ledge directly beneath the décollement zone. This microbreccia ledge in itself is probably a zone of coalescence of listric(?) normal faults (Shackelford, 1977), but what controlled its exact structural position is not known. Above it, the nontectonite detachment rocks are found. The stacking of structural units -- mylonitic tectonite, sub-décollement detachments, microbreccia ledge, and upper-plate detachments -- is a profound example of 'disharmonic faulting'.

MID-MIOCENE LISTRIC NORMAL FAULTING

by

G.H. Davis and J.J. Hardy, Jr.

Introduction

The emplacement of the upper-plate detachments is a composite product of (1) an early tectonic denudation associated with formation of mylonitic tectonite and of the décollement zone, and (2) a mid-Miocene low-angle normal faulting which is now regarded as an integral part of the regional tectonic strain of the southern part of the western Cordillera. To understand the denudation contribution afforded by the mid-Miocene faulting, it is useful to examine in some depth the structural characteristics and terranes in which the detachments are mainly composed of sedimentary and volcanic strata of Miocene age. The following account is directly from the work of Davis and J.J. Hardy, Jr. (submitted for publication).

Views regarding the origin of the upper-plate detachments are changing in accord with new facts derived from careful studies of the mid- to late-Tertiary record. Armstrong's (1972)
success in discriminating between Cenozoic and pre-Cenozoic low-angle faults in the Basin and Range of western Utah and eastern Nevada provided a glimpse of extensive Tertiary 'denudational' faulting which had largely gone unnoticed. Anderson's (1971, 1977, 1978) brilliant mapping and synthesis of structures in the region south of Lake Mead in Arizona and Nevada yielded salient geological details regarding one class of denudational faults, namely a listric faulting which he regarded as having accommodated thin-skin distension of the crust in an east-west direction. The fundamental relationship which he observed repeatedly is one of faulted and rotated Miocene strata separated from underlying Precambrian granitic basement by listric normal-slip faults (Fig. 3-14). The important conclusion which Anderson (1971) emphasized was that the listric normal faulting pre-dated the high-angle faulting which blocked out the present basins and ranges. According to Anderson (1971) listric normal (growth) faulting took place mainly between 15 and 11 m.y., whereas the high-angle faulting proceeded approximately between 10 and 5 m.y.

The deformational system which Anderson (1971) described is an integral and representative component of regional tectonic strain in the southern Basin and Range. G.A. Davis, E.G. Frost, J.L. Anderson, T.J. Shackelford, and their co-workers have shown this to be true in southeastern California and westernmost Arizona. P.E. Damon, S.C. Creasey, M.H. Krieger, H.W. Peirce, W.A. Rehrig, S.R. Reynolds, R.B. Scarborough, and M. Shafiqullah independently reached the same conclusion in west central and southeastern Arizona (see references which follow).

MIOCENE LISTRIC FAULTING AND METAMORPHIC CORE COMPLEXES

Evaluating the nature and origin of mid-Miocene denudational faulting in the Basin and Range is almost inseparable from trying to unravel the tectonic evolution of metamorphic core complexes (Crittenden and others, 1978, 1980; Davis and Coney, 1979). Metamorphic core complexes owe much of their physical distinctiveness to the superimposition of (1) formation of thick zones of lineated mylonitic tectonite, and (2) extensional faulting of Tertiary age, including mid-Miocene denudational listric faulting. Tectonite and non-tectonite are separated by the distinctive low-angle décollement (see Crittenden and others, 1980), and striking contrast between rocks above and below such structural discontinuities has been very puzzling to explain.

The presence of detachments (i.e., upper plate, hanging-wall rocks in low-angle fault contact on structurally deeper footwall rocks) which were emplaced in mid-Miocene are not unique to the
Figure 3-14. Structural-profile of listric-fault relationships in Black Canyon region, northwestern Arizona. Figure is modified from Anderson (1978).
metamorphic core complexes as defined by Davis and Coney (1979); rather they may be found both within and outside of them. Away from the metamorphic core complexes, the low-angle faults separate non-tectonite steeply rotated detachment strata from underlying non-tectonite basement. The basement rocks are commonly highly fractured and marked by chloritic and ferruginous alteration along the faults, but there is a compatibility of brittle behavior between structures in the hanging-wall and footwall rocks. In these field occurrences, workers have never felt compelled to introduce terms like 'décollement' to describe the contacts. 'Low-angle (normal) faulting' has adequately identified these relations. Within metamorphic core complexes, however, the décollement sharply separates underlying lineated mylonitic tectonite from overlying, mainly non-tectonite detachment strata. The mid-Miocene (approximately 18 to 14 m.y. +) denudational faulting is physically superimposed on the tectonite fabric. In décollement zones the lineated tectonite fabric is shattered and rotated by movements related to the denudational faulting. Such facts confirm that the faulting, at least in part, post-dated the development of lineated tectonite (Coney, 1974; Davis, 1975; Davis and Coney, 1979).

The paradox which confronts workers who are interested in evaluating the relationship between the mid-Miocene faulting and the evolution of metamorphic core complexes is in reconciling the following facts:

1. mid-Miocene faulting post-dates the formation of lineated tectonite
2. everywhere in the southern Basin and Range, the slip-line direction which describes extension in the detachment strata is virtually parallel to penetrative lineation in underlying or near by tectonite.

Regional Occurrences of the Listric Fault Systems

Major Geologic Characteristics

Rocks in the detachments range in age from Precambrian to middle Miocene, and they are always made up of rocks derived from the local geological setting. Some structures within the detachments evolved in those rocks prior to the tectonic denudation, but most of the conspicuously brittle deformation appears to have attended the detachment process itself. In their simplest form, the detachments contain moderately to steeply dipping Oligocene(?)-Miocene strata which rest on expansive flat to gently dipping faults. Davis and Coney (1979) suggested that the continental sediments and associated volcanics which comprise a great portion of the Oligocene-early Miocene detachment strata were deposited in
extensional basins which grew as a response to the formation of lineated tectonite at depth. These strata are presumed to be characterized by growth-fault structure/stratigraphic patterns. Such patterns would be older than the mid-Miocene faulting of interest to us here.

In California and Arizona examples, the footwall rocks on which the detachments rest are typically composed of Precambrian rocks, although any relatively old, relatively deep component of the local geological column might compose the footwall. Workers generally agree that the footwall blocks are autochthonous. Rotation of strata in the detachments was accompanied by brittle fracturing, and in all probability was accomplished by listric normal faulting and coordinated tear faulting. Bedding strike of hanging-wall strata are typically oriented at right angles to the direction of greatest finite strain in the detachments. Regionally, there are broad structural domains of homoclinal detachment strata, each of which is characterized by (1) uniform dip direction of hanging-wall strata and, thus, (2) uniform sense of bedding rotation. Rehrig and Heidrick (1976) first called attention to the regularity of dip direction of Oligocene/Miocene volcanic and sedimentary rocks in the Basin and Range of Arizona. Furthermore, they clearly recognized that the regional pattern was produced by 'thin-skinned, rotational faulting and tilting' (Rehrig and Heidrick, 1976, p. 217).

The descriptions which follow are taken from studies in which the mid-Miocene timing for faulting can be proved and/or convincingly argued. Whether this faulting is simply a continuation of the faulting which produced décollement zone fabric is not known. Treating the mid-Miocene faulting as a discrete event may or may not be valid. Readers are referred to the original references for details regarding the stratigraphic and lithologic characteristics of the detachment strata and for geochronological facts. The work by Eberly and Stanley (1978) provides a regional description and interpretation of Cenozoic stratigraphy and faulting in southern Arizona. They contributed significantly in pioneering an awareness that mid-Miocene faulting is a distinctive tectonic episode which preceded 'late Miocene block faulting'. The papers by Scarborough and Peirce (1978) and Shafiquallah and others (1978) are all landmark papers which underscore the fact that major extensional faulting of a thin-skinned and rotational nature affected the region during the Miocene, before the classical Basin and Range disturbance.

Black Canyon Region, Northwestern Arizona

The type locality for the style of structures being considered here is that mapped by Anderson (1978) south of Lake Mead in the Black Canyon region of the Colorado River (Fig. 3-15). The
Figure 3-15. Location map for geographic areas referenced in discussion of mid-Miocene faulting.
geological framework there consists of Precambrian granite overlain nonconformably by Oligocene(?)–Miocene volcanic and sedimentary rocks. As a result of mid-Miocene denudation, the Tertiary strata within a 1000 + km² area are moderately to steeply tilted eastward within a succession of imbricate, west-southwest-dipping, listric normal faults. The direction of greatest finite strain is N70°E/S70°W, as revealed by abundant striae and by the uniform north-northwest strike of detachment strata. The dip direction of the detachment strata, combined with the listric nature of the faulting, demands that the sense of movement for the fault system was S70°W. East-west-trending tear faults, referred to by Anderson (1979, pers. comm.) as marginal shear zones, served to coordinate the internal movements of structural lobes within the regional mass. Some of the marginal shear zones have been shown by Anderson to be structures where the soles of the listric faults 'shoal up' to near-vertical attitudes.

Cross sections which Anderson (1977, 1978) presents clearly show that the faults are growth faults. Older rocks within the Tertiary column tend to show greater displacements than younger ones. Hanging-wall strata for a given unit on a given fault tend to be thicker than strata of the same unit immediately below the fault. The youngest Tertiary strata affected by the faulting locally display nonconformable contact relations with underlying Precambrian granite.

Anderson has demonstrated that the listric normal faults commonly penetrate the Precambrian basement, but they are very difficult to map. Parts of the 'autochthonous' Precambrian basement may actually be allochthonous detachments. Directly below the system of faulted Tertiary strata, the Precambrian rocks are locally intensely fractured and altered. In fact, at one location along the western margin of Black Mountain, there is a décollement-like zone of resistant, iron-stained micro brecciated rock, not unlike that found along the margins of metamorphic core complexes.

According to Anderson (1979, pers. comm.), it is the vitrophyre in the Tertiary strata that took up much of the strain of accommodation to the listric faulting and marginal shearing. Matrix of ash flows in the Tertiary strata has locally been converted to tektinite. Within the bold exposures of Black Canyon, the attenuation and stretching of the Tertiary strata are clearly seen, not only in the form of brecciation and shearing, but in the coalescing of faults to produce boudin-like, lozenge-shaped fault blocks as well as local draping of strata over low-angle normal faults.

Anderson (1971) has interpreted this array of structures as the response of upper crustal rocks to extreme distension brought
about by batholithic emplacement and crustal extension at depth. Gastil (1979) has developed such a model of crustal distension in some detail.

Southeastern California-Western Arizona

To the south of the Black Canyon region, G.A. Davis and his students have mapped the internal structure and stratigraphy of Tertiary strata in the Whipple, Buckskin, and Rawhide Mountains (Fig. 3-15) and have described the physical relationship of these rocks to lineated tectonite of core-complex affinity. The deformed terrane is fully 3000 km² in area, 100 km long in a northeast-southwest direction (G.A. Davis and others, 1980). Virtually all of the detachment sheets are separated from autochthonous footwall by a regional, subhorizontal detachment fault which they call the Whipple detachment fault to the west, Buckskin detachment fault in the central part, and the Rawhide detachment fault to the east (Shackelford, 1977; G.A. Davis and others, 1979).

The upper plate detachment rocks consist of Oligocene (?) to middle Miocene sedimentary and volcanic rocks as well as Precambrian and Mesozoic (?) crystalline rocks, and Paleozoic metasedimentary and metavolcanic rocks. Footwall rocks are both non-lineated, non-mylonitic rocks and lineated quartzo-feldspathic gneisses (Shackelford, 1977; G.A. Davis and others, 1977, 1979).

Imbricate listric normal faults, and some marginal tear faults, are the major structures within the detachments (Shackelford, 1977; Davis and others, 1979; Anderson and others, 1979). The faults are growth-fault in nature (Frost, 1979). The slip-line direction for faulting is N40°E to N60°E, as deduced by striae and bedding strike, and the sense of movement is in that same direction (G.A. Davis and others, 1977, 1979; Shackelford, 1977). Strata dip southwest as a result of back-tilting along the northeast-dipping normal faults. The tear faults helped to accommodate differential upper-plate distension. The same structural patterns and relationships have been observed by Davis and others (1979) to the north in the Chemehuevi, Sacramento, Homer, and Dead Mountains (Davis and others, 1979) and to the east in the Mohave and Artillery Mountains (Frost, 1980, pers. comm).

Detachment faulting is interpreted to have taken place in mid-Miocene, (Shackelford, 1977; Davis and others, 1977). Davis and others (1979) have concluded that 'regional mylonitization' significantly predated Oligocene (?) - Miocene detachments events 'and that' field relations indicated that the mylonitic rocks, at least at levels near the basal detachment surface, were kinematically 'dead' and cold, or at least cooling, at the onset of
of detachment faulting. Davis and others (1979) explain the detachment faulting as a regional, northeast-directed gravity gliding over a non-extending lower plate.

West-Central Arizona

To the southeast of the region studied by G.A. Davis and his students, there is a well defined and extensive belt of metamorphic core complexes which trends northwest-southeast through most of Arizona (Fig. 3-15). In the Harcuvar and Harquahala Mountains, Rehrig and Reynolds (1977, 1980) report the presence of moderately to steeply tilted mid-Tertiary volcanic and sedimentary rocks which rest on a gently east-dipping dislocation surface. Beneath the dislocation surface are chloritic, mylonitic gneisses which contain N60°E-trending cataclastic lineation. Formation of the dislocation surface pre-dates 10 to 13 m.y. basalts. "Activism of the dislocation surface, deposition of coarsely clastic sedimentary wedge deposits, and rotational faulting were nearly coincident processes" (Rehrig and Reynolds, 1980). Rehrig and Reynolds (1980) note the kinematic coordination between the east-northeast direction of extension implicit in the listric-normal faulted blocks of mid-Tertiary volcanics and conglomerates. Although they are aware that the chlorite breccia and dislocation surface are imposed on already-formed mylonitic foliation and lineation, they postulate that the 'upper plate' rocks now represented by listric fault blocks of Miocene strata were distended above a thin, sub horizontal zone of extreme extension and flattening, now represented by mylonitic gneiss.

The coordination of the mylonitic fabric and the orientation of normal faults and rotated upper-plate strata are beautifully pictured by Rehrig and others (1980). They reported structural and geochronological relationships in a terrane of denudational faulting just southeast of the Harcuvar and Harquahala Mountains, in the Vulture, Big Horn, and Eagle Tail Mountains. They suggested that east-northeast/west-northwest crustal extension affected the entire area, producing rotational normal faulting of late Oligocene/early Miocene conglomerate and mid-Miocene volcanics. Faulting took place between 20 m.y. and 16 m.y. and produced "shingled prismatic crustal blocks bounded by downward flattening fault planes" (Rehrig and others, 1980). Footwall granitic and gneissic rocks became intensely brecciated. Detachment strata are oriented perpendicular to the east-northeast direction of crustal stretching. The sense of rotation of upper plate strata is homogeneous over large expanses. From the northern part of the Big Horn Mountains through the Vulture Mountains (and nearly to the Bradshaw Mountains), the upper plate strata are tilted eastward on gentle west-dipping faults (Rehrig and others, 1980). From the southern end of the Big Horn Mountains through the Eagle Tail Mountains, strata are tilted westward on east-dipping listric faults. Rehrig and others (1980)
specifically comment that in all respects, the faulting is comparable to that described by Anderson (1971).

Further southeast, on the west outskirts of Phoenix, Reynolds and Rehrig (1980) mapped and studied the geochronology and structure of South Mountains. They concluded that mylonitic lineated gneiss was derived in part from cataclastic reduction of a pluton which they regard as late Oligocene to early Miocene in age. Age determinations were based on Rb-Sr whole rock isochrons. Chloritic breccia overlying mylonitic granodiorite is well exposed there. Relict mylonitic foliation within it is rotated out of the normal northeast strike to northwest to the perpendicular direction of mineral lineation.

Southeastern Arizona

North-northeast of Tucson in San Pedro Valley (Fig. 3-15), M.H. Krieger, S.C. Creasey, and H.R. Cornwall (see references) have mapped quite a number of detachments of Oligocene(?) - Miocene strata which are separated by low-angle faults from footwall Precambrian granite. The structural properties of the system are very similar to those described by Anderson (1971). The upper-plate strata in these detachments are moderately to steeply tilted, with strike attitudes (north-northwest) orthogonal to lineation orientation (N60°E) in tectonites which lie kilometers to the west and south. The strata dip southwestward, indicating northeastward tilting along southwest-dipping listric(?) normal faults. The domain of this tilting occupies at least 2000 km². Precambrian granite, below or more rarely within the detachments, is generally shattered and altered close to the faults. Locally it is converted into a macrobreccia. A few tear-fault boundaries have been recognized. For example, Krieger (1974) mapped such a boundary along the eastern edge of the Tortilla Mountains. There, steeply tilted younger Precambrian and Miocene strata are dragged in left-slip fashion along a northeast-striking high-angle tear fault, southeast of which is (exclusively) Precambrian granite. Minimum left slip is 2 km.

Upper-plate rocks above décollement zones in metamorphic core complexes of southeastern Arizona are generally more heterogeneous and structurally complicated than detachment rocks in the San Pedro Valley examples (Davis, 1977, 1980). These detachments locally contain mid-Miocene strata, but additionally there are Precambrian, Paleozoic, Mesozoic, and early to mid-Cenozoic rocks. The largest and most continuous exposure of a décollement in southeastern Arizona is the Catalina fault which crops out along the south base of the Santa Catalina Mountains and the western and southern margins of the Rincon Mountains (Pashley, 1966; Drewes, 1975, 1978; Banks, 1976; Creasey and Theodore, 1975). It dips 15° to 25° in
a southerly or westerly direction. Chlorite breccia is commonly well developed below this décollement, and the chlorite breccia in turn is underlain by hundreds of meters of mylonitic gneiss. Recent work by Davis and Gardulski (in prep.) reveals that relict, overprinted foliation and lineation within the chlorite breccia is systematically rotated perpendicular to lineation in underlying mylonitic gneiss. Drewes (1978) has pointed out spectacular polished fault mullion on the Catalina fault which trend N60°E.

Movements on the Catalina fault are multiple and superimposed. The full structural history includes development of lineated tectonite, listric normal faulting and formation of the décollement zone. At some locations, like Happy Valley on the east side of the Rincon Mountains, the detachment strata of unmetamorphosed homoclinal Paleozoic strata rest in fault contact on tectonite with the same simplicity and geometric regularity as described for detachments in the aforementioned San Pedro Valley examples. In other cases, like Saguaro National Monument (East) or Colossal Cave (Davis, 1975; Davis and Frost, 1976; Davis and others, 1974), detachment strata contain spectacularly and sometimes complexly folded Paleozoic-Mesozoic strata, and the total deformation cannot be related to a single, simple movement plan of mid-Miocene faulting. The variety of physical structures in such systems, and the nature of progressive deformation through time, is presented in Davis (1980).

West of Tucson, in Papago country, there is a recently discovered array of core complex terranes which are unique in that lineation in the tectonites trends approximately north-south, not N60°E (Davis, 1977; 1980). In the eastern Comobabi Mountains, lineated tectonites of core-complex affinity as well as décollement-zone rocks crop out adjacent to a major outcrop area (50 km² +) of Miocene continental sediments. Noteworthy is the fact that Haxel and others (1978) have mapped an orthogonal relationship between tectonite lineation and (1) bedding strike of the moderately dipping sediments, and (2) strike of the (low-angle) normal faults which separate the sediments from underlying older, mainly crystalline bedrock.

The Eagle Pass Detachment, Southeastern Arizona

Introduction

The Eagle Pass detachment northeast of Tucson near Safford, Arizona, (Fig. 3-16) displays a structural style of deformation which both conforms and adds to our understanding of the mid-Miocene denudational systems. The Eagle Pass structure is the easternmost known detachment in southeastern Arizona. Its geological location is especially instructive for it lies on undeformed
Figure 3-16. Location of Eagle Pass study area.
Precambrian quartz monzonite, even though lineated tectonite derived from the Precambrian protolith crops out only several kilometers to the east (Davis, 1977, 1980).

The Eagle Pass detachment occupies a broad, high pediment-saddle between the imposing Pinaleno Mountains to the south and the Santa Teresa Mountains to the north. These mountains, taken together, may be thought of as a typical 'range' of the Basin and Range province. The pre-Basin and Range origin of the detachment is partly indicated by its presence within the range. Precambrian quartz monzonite crops out over most of the saddle between the Pinaleno and Santa Teresa Mountains, but toward its western edge a low semi-circular ridge demarcates upper-plate lower and middle Miocene strata which rest in fault contact on the basement. Still further west, major westward-draining canyons carve into the faulted rock system and afford special views of its anatomy.

General Geology

The Eagle Pass detachment is underlain by a fault surface which displays a curved, convex-northeast trace which is approximately 11 km in length (Fig. 3-17). The footwall is everywhere Precambrian basement, predominantly composed of the medium- to coarse-grained porphyritic quartz monzonite. The quartz monzonite is (probably) part of the regional 1.4 to 1.5 b.y. anorogenic quartz monzonite suite (Silver and others, 1977). For the most part the quartz monzonite is light tan to off-white in color. It has decomposed to the extent that much of the landscape is covered by grus. Within the quartz monzonite there are scattered small patches of older Precambrian Pinal Schist. Northeast-striking, vertical dikes of rhyolite and rhyolite porphyry comprise a swarm which intrudes the quartz monzonitic footwall rocks. The dikes do not cut the upper-plate strata. Rehrig and Reynolds (1980) report a 25 m.y. K-Ar age based on analysis of biotite from these dikes. The value '25 m.y.' has almost become a signature marking proximity to metamorphic core complex terranes (Davis and Coney, 1979).

The detachment proper occupies 25 km² and consists of Miocene volcanic and sedimentary rocks. Three major units were identified by Blacet and Miller (1978) and these were 'carried' in our mapping. The oldest is composed of numerous reddish-brown porphyritic andesite flows containing phenocrysts of randomly oriented to radiating plagioclase laths. Its maximum thickness is approximately 945 m. This rock is known to workers in southern Arizona as the 'turkey track porphyry', and it is thought of as a guide rock of uppermost Oligocene/lower Miocene. Overlying the andesite is a middle unit which consists mainly of a rhyolitic welded ash-flow tuff with thin rhyolite and rhyodacite flows. It also contains intercalated rhyolite breccias and coarse sedimentary breccia. The thickness of
Figure 3-17. Geologic map of Eagle Pass study area. Map prepared by J.J. Hardy, Jr.
Figure 3-17
this middle unit is approximately 550 m. The andesite and rhyolite, taken together, are undoubtedly part of the Galiuro Volcanics (Cooper and Silver, 1964; Simons, 1964; Creasey and Krieger, 1978) whose type area is in the next range west. The Galiuro Volcanics are nearly 1900 m thick in the central Galiuro Mountains. Creasey and Krieger (1978) report 11 K-Ar age determinations for the formation, and these range from 22 to 28 m.y. The youngest unit in the detachment is a very thick (4485 m), red-brown, well indurated, poorly stratified fanglomerate and mud-flow breccia composed of heterogeneous boulders and cobbles of angular to subrounded andesite and rhyolite. Clasts range in length from several centimeters to more than 5 m. This coarse clastic sequence may be correlative with the Hell Hole Conglomerate (Simons, 1964), the type locality of which is on the east flank of the northern Galiuro Mountains. Its age is uncertain, although in the Galiuro Mountains it depositionally overlies the Galiuro Volcanics. All of the detachment strata dip moderately to steeply to the southwest. They are overlain unconformably by very gently dipping, moderately well indurated conglomerate of Plio-Pleistocene(?) age.

Structural Geology

The trace of the Eagle Pass fault is markedly sinuous. Although the fault zone is continuous and curviplanar, it is convenient to think of it as composed of a number of segments which are distinctive in attitude. Along the northwest margin of the Eagle Pass detachment, the fault zone strikes N50°E and dips steeply southeast. In the vicinity of Underwood Canyon, the fault assumes a very low angle of dip (less than 15°) and swings in strike from northeast through southeast to south-southwest. From the vicinity of Eagle Pass to the south end of the study area, the fault strikes north-northeast and maintains a low-angle dip. When plotted stereographically, the attitudes of fault segments disclose that the fault has a conical form, reflecting a trough-like fault morphology converging northeastward. The axis of the trough plunges 12°S 58°W.

The form of the Eagle Pass fault has profoundly influenced the orientation and style of deformation of both upper-plate and footwall rocks. Bedding within the detachment strikes consistently N50°-55°W and dips moderately to steeply southwest, except along the northwest margin of the detachment. There, the volcanic rocks swing into parallelism with the fault, conforming both in strike and dip to the fault-zone attitude. This shift in bedding attitude close to the fault resulted in elevated strain as expressed by an unusually high degree of fracturing, abundant striae reflecting adjustments along bedding and faults, and the mechanical breakdown of the volcanic units into discontinuous structural lenses.
Notably, along the northwestern and northeastern parts of the fault, the rhyolitic unit is locally juxtaposed directly against basement granite, without intervening andesite.

The fanglomerate is not as sensitive as the volcanic rocks to changes in fault attitude. The fanglomerate strikes N50°-55°W and dips steeply southwest throughout the area, except within tens of meters of the northwest margin of the detachment. There, the fanglomerate is locally dragged and faulted into parallelism with the northeast-striking, steeply southeast-dipping volcanic rocks, and with the fault zone itself.

Applying J. Hoover Mackin's (1950) down-structure method for viewing geologic maps proves to be a powerful approach to grasping the structural significance of the fault bedding relationships in the Eagle Pass area. When viewed southwest down the gently dipping slope of the fault plane, the geologic map is transformed into a provocative structure section (Fig. 3-18). The upper surface of the quartz monzonite footwall is seen to be a deep trough or groove, mullion-like in terms of its macroscopic structural significance (Wright and others, 1974). The northwest wall of the trough is especially steep and has served to localize subparallel fault zones in the detachment strata. Control for the location and orientation of the steep northwest trough-wall originally may have been afforded by one of the 25 m.y. rhyolite dikes. Along the northwest wall, detachment strata are steeply dipping and attenuated. Locally, rhyolite abuts directly against the quartz monzonite. The footwall quartz monzonite is severely fractured along that margin for a distance of several hundred meters, and ferruginous alteration in the zone of intense fracturing has converted the terrain into one of rich red-orange hues. Along the base of the fault trough there is a distinctive keel below which (10 m +) the footwall quartz monzonite has been profoundly fractured, mylonitized, and altered. Alteration is both chloritic and ferruginous. Planar surfaces in the fault zone look like 'boiler plates' with alteration-derived metallic gray/black tones embellished with tectonic polish. Just southeast of the keel, the andesite unit is attenuated and faulted in such a way that rhyolite again rests directly on granite. A few meters to tens of meters below the fault, the Precambrian quartz monzonite is absolutely undeformed and unaltered.

Movement Plan and Deformational Characteristics

A number of lines of evidence indicate that the upper-plate detachment strata were faulted into position by northeast-directed translation. The translation was accompanied by rotation of strata to steep southwest dips, and during faulting the upper part of the autochthonous footwall quartz monzonite became highly fractured.
Figure 3-18. Down-plunge view of geologic relationships shown in Figure 3-17. View is southwest-directed.
Along the keel of the detachment, the footwall rocks were converted into a rind of mylonite. Although there are data which suggest that the upper-plate strata were extended during faulting, there is no evidence in the footwall rocks for such northeast-southwest elongation. Rather, the footwall rocks appear to have been stationary and rigid when they were overridden by the detachment strata. Loci of heavy alteration in the footwall quartz monzonite are positioned along the keel of the fault and along the steep northwest trough-wall. These locations were vulnerable to migration of hydrothermal solutions along the detachment strata/fault interface and (then) into fractured and mylonitized quartz monzonite.

Data which suggest that the line of fault movement was N40°E/S40°W include (1) the orientation of the axis of the trough-like form of the fault, (2) the uniform N50°W orientation of bedding strike in the tilted detachment strata, and (3) grooves and striae in deformed quartz monzonite, especially in the keel of the fault. The important logical argument (assumption) is that detachment strata were rotated and thus tilted during faulting. The axis of the fault trough and the average strike of bedding are beautifully coordinated in an orthogonal way. Grooves and striae measured mainly at the Big Spring locality are convincingly northeast-southwest in azimuth, with a great-circle distribution oriented N50°E. There is scatter in the trend of striae, and it is emphasized that striae display a dispersal which reveals a complex internal movement plan. Indeed, one gains the impression that faulting was accompanied by converging/diverging path movements which were highly dependent on local boundary conditions. The range of the striae azimuth is quite unlike the tight lineation clustering which typifies tectonites in metamorphic core complexes. Large mullion-like structures (elongate ridges, 10 to 20 m long, faceted by polished, stained, and striated surfaces) are present in quartz monzonite at the Big Spring locale, and their orientations are northeast-southwest. The form of these structures is identical to those mapped by Drewes (1978) along the Catalina fault in the Rincon Mountains. Joints were measured on 'boiler plate' outcrops in the fault zone at Big Spring, and although they cannot be used to evaluate movement plan, it may be significant that the stereographically plotted patterns are symmetrical with respect to a N40°E/S40°W line of movement.

Striated surfaces in the upper-plate detachment strata are not particularly abundant, except along the northwest faulted boundary of the plate. Thus, it is difficult to corroborate the N40°E/S40°W inferred movement plan on the basis of upper-plate minor structures. A notable exception is a series of keystone fault blocks in rhyolite in Section 27 in the north central part
of the area. Normal-slip faults mark the borders of several rotated blocks of rhyolite. These structures lie 20 m above the Eagle Pass fault. Their presence discloses that some degree of northeast-southwest elongation attended translation during faulting. Faults in the andesite and rhyolite are difficult to find because of the poor quality of many of the hill side exposures. Certainly these rocks are highly fractured. The fanglomerate is beautifully exposed, but faults are absent except along the northwest border of the detachment. Jointing in the fanglomerate is mild to almost non existent.

The main evidence that the sense of translation on the Eagle Pass fault was northeast-directed is an extraordinary zone of tear-faulting on the northwest boundary of the detachment. Evidence for strike-slip faulting is conspicuously displayed by horizontal striae on polished fault, fracture, and bedding surfaces in rhyolite. The rhyolite forms a wall of steeply dipping strata which were dragged into parallelism with the trough-wall of quartz monzonite. The andesite, a much weaker rock than the rhyolite, was severely attenuated within this zone and is generally absent. Where it does crop out, exposures are poor and striae are not generally evident.

Nowhere along the northwest tear-fault boundary of the detachment do the volcanic units display their normal N50°W strike. The volcanic rocks are plated against the footwall quartz monzonite along the fault. In contrast, the fanglomerate maintains N50°W strike attitudes until within tens of meters of the tear-fault boundary. At that point, the layers are dragged abruptly westward and/or the rock is transformed to flattened-pebble tectonite along zones of penetrative simple shear. The drag effects clearly reveal left-slip of the detachment along the fault interface. The development of tectonite at the expense of 20 m.y. + fanglomerate containing competent rhyolite clasts is shocking. The zones of simple shear are up to ten meters wide and are separated by intervening zones of approximately the same width where the fanglomerate is highly faulted and fractured, but not converted to tectonite. The tectonite fabric is instructive in that it must have formed under dry, cool, relatively shallow conditions of deformation. The strain characteristics of the simple-shear zones are identical to those discussed by Ramsay and Graham (1976). Movement of the fanglomerate detachment strata northeastward past the footwall of quartz monzonite must have been met by profound frictional resistance to movement. This is yet another indication that the quartz monzonite footwall was stationary and rigid.

One of the most peculiar structural relationships along the tear-fault boundary is the presence of a 'stratified' monolithologic breccia which is sandwiched between the highly fractured/shattered
quartz monzonite footwall and steeply dipping, northeast-striking andesite. Layering within the breccia is concordant to that of the andesite. The exposure is approximately 200 m in trace length, and the thickness of the breccia is about 30 m, maximum. Virtually all of the clasts and matrix material of the monolithologic breccia are derived from the adjacent quartz monzonite. Clasts are as large as 1/2 m in size, and the layering in the breccia is primarily due to size-sorting of fragments and matrix. Although the rock appears to be a water-lain conglomerate/breccia, two factors suggest that the rock may be a kind of tectonite derived from comminution of the footwall. First, there is no rock like this anywhere in the detachment stratigraphy outside of the tear-fault zone. Its position within the tear-fault zone is suspicious in that it lies at an abrupt bend in the major fault. The form of the bend, combined with the left-slip nature of movement on the fault, demands that the location of this breccia formation must have been a site of significant compressive stress. Secondly, the andesite and rhyolite just southeast of the monolithologic breccia are extremely highly fractured and faulted, yet most of the breccia is absolutely devoid of fractures, even joints. Only in a few places are single, through-going fractures evident, and these tend to off set breccia clasts by small displacements. Additionally, the clasts of quartz monzonite are so internally shattered that they could not possibly have been physically transported to their present sites in their present condition. These observations suggest that penetrative comminution, granulation, and dry flow superceded regular fracturing and shattering of the quartz monzonite. Thin-section study of the monolithologic breccia has not yet resolved the problem. This rock is still under study, and preliminary results suggest that monolithologic breccia may be a legitimate strain facies of detachment terranes. Krieger (1974) has mapped macrobreccias of Precambrian granite along the margins of some detachments in the San Pedro Valley.

A final slip-line indicator is a single asymmetrical overturned fold, 2 m in amplitude, found in highly deformed andesite just a meter above the Eagle Pass fault in the southeast order of the map area. The fold trends N40°W, and is unambiguously overturned northeastward.

Structural Interpretation

Rehrig and Reynolds (1980) examined in reconnaissance the Eagle Pass fault zone and reached the conclusion that deformation was achieved by a listric normal faulting which produced southwest-to northeast transport of a upper-plate strata. Only listric faulting along a northeast-dipping fault zone could explain (easily)
the moderate to steep southwest dip of the upper-plate detachment strata. We agree with their interpretation totally and regard the data and relationships presented herein as a rather definitive proof. We believe that the Eagle Pass fault is the sole of a large, regional, listric normal fault zone whose upper reaches are simply not exposed, that the fault zone was curved is an inference born simply from the reality of rotated strata. We infer that the fault was originally flat or gently east-dipping, but was rotated to its present attitude by arching of the Pinaleno Mountains and/or Basin and Range faulting.

Movement indicators certainly support the interpretation of northeastward transport. The marginal tear-fault boundary is perceived as a zone of accommodation to differential relative movement within the larger mass. The volcanic rocks and fanglomerate in the detachment are identical to Galiuro and Hell Hole stratigraphy in the Galiuro Mountains, and this too strongly supports a westward provenance for the detachment. The detachment mass carried on the system of listric faults may have been enormous, for a detachment exists 30 km south-southeast of the Eagle Pass area which appears, from its small-scale geologic map portrayal, to display macroscopic relationships identical to those discussed.

The volcanic and sedimentary rocks in the Eagle Pass detachment may have been 'plucked' by faulting from a country rock sequence characterized by the Miocene strata in horizontal and non-conformable contact with underlying Precambrian granite (Fig. 3-19A). Such an undisturbed relationship is preserved in the central part of the Galiuro Mountains. The listric normal fault zone could have cut through the Tertiary volcanic and sedimentary rocks, curving into the discontinuity marked by the nonconformity (Fig. 3-19B). Alternatively, the fault could have cut downward and through the unconformity, thus allowing Precambrian granite to occupy an upper-plate detachment position (Fig. 3-19C). We cannot distinguish with confidence between these possibilities, although after significant field effort we have eliminated in our minds the possibility that any of the shattered Precambrian quartz monzonite occupies an upper-plate detachment position at Eagle Pass. In any case, progressive faulting resulted in lowering of the sedimentary and volcanic rocks onto Precambrian quartz monzonite footwall. By the time that the andesite and rhyolite reached the level of basement, the sole of the fault was already marked by significant curvature and irregularity, notably the large, mullion-like trough. The volcanic rocks were plated (smeared) concordantly to the southeast flank of the mullion as the detachment moved in left-slip fashion past the rigid quartz monzonite footwall. Dragging of the volcanic rocks along the keel of the sole may have been responsible for the decrease of dip of the volcanic layering.
Figure 3-19. Schematic diagrams of listric faulting. (A) Listric faulting that mainly affects rocks above the great unconformity. (B) Listric faulting that penetrates the basement.
The fanglomerate exposed along the northwest margin of the Eagle Pass detachment is believed to express a stage of the faulting in which that rock first felt the effects of interference with the basement obstruction. As the fanglomerate was lowered down on the mullion, only the closest fanglomerate layers responded to the frictional resistance to steady northeastward movement. The northeast transport must have been on the order of kilometers. Thus it is not surprising that extreme granulation and mylonitization of quartz monzonite were locally achieved.

**DYNAMIC INTERPRETATIONS**

During the Laramide in the southern part of the western Cordillera, rocks were subjected to strong northeast-southwest compression. The cause of the compression has been attributed to rapid convergence of the North American plate against oceanic plates to the west (Coney, 1978; Burchfiel and Davis, 1975; Armstrong, 1968a). The effect of the compression was northeast-southwest shortening of upper crustal rocks. Shortening was achieved by monoclinal folding and associated basement-involved uplift(s) in the Colorado Plateau (Kelley, 1955). Along the western edge of the Colorado Plateau in southwestern Utah in the vicinity of the hingeline, Laramide monoclines and basement-cored uplifts were superposed on earlier, thin-skinned Sevier thrusts of a décollement type. In southeastern Arizona, the gross structural geometry of shortening is not agreed on. Drewes (1973, 1976, 1978) believes that shortening was accommodated by regional, northeast-directed overthrusting, involving translation of upper-plate rocks for distances in excess of 100 km. In the context of descriptions presented earlier, he regards the folded tectonite carapace as a product of dynamic metamorphism of 'lower-plate rocks' during overthrusting. The rocks here described as locally-derived detachment rocks, including shattered Precambrian granitic rocks and overlying non-tectonite Phanerozoic sedimentary rocks, are thought to have been brought from the southwest to their present site by overthrusting, although gravitational adjustments in the Tertiary determined their specific 'final' locations (Drewes, 1976, 1978). Thorman (1977) supports this model and regards the penetrative lineation in mylonitic tectonite as a Laramide signature recording the direction of overthrusting. By way of contrast, Davis (1979) views the Laramide structural framework of southeastern Arizona as a foreland setting comprised of Wyoming-type basement-cored uplifts, the margins of which were partly pre determined by locations of major Jurassic and perhaps Precambrian faults. Shortening in rocks along the margins of such uplifts took place by tight upright to overturned folding, thrust faulting, cleavage development, and local formation of tectonite. The Anschutz model for the Laramide framework is the most extravagant presented to date.
It portrays the well-known southeastern Arizona country rock framework as wholly allochthonous and overlying two repeated sections of Precambrian, Paleozoic, and Mesozoic strata.

The ambiguity expressed by these multiple hypotheses forms a confusing backdrop with which to evaluate the dynamic significance of metamorphic core complexes. However, most workers agree that the deformation by crustal shortening, whatever its nature, was completed by 55 m.y. (?), for plutons and dikes in the age-range 70-50 m.y. cut already-deformed strata. The presence of 'core complex fabrics' in rocks 55 m.y. old suggests that the formation of the core complexes was temporally separate from the main regional shortening event.

The effect of major plutonic invasion into the upper crust in the early Tertiary may have set the stage for the unusual style of penetrative deformation recorded in metamorphic core complex terranes. Gordon Haxel (Pers. comm., 1979) drew this conclusion during his southern Arizona mapping, and T.H. Anderson (pers. comm., 1980) has independently reached the same conclusion regarding terranes in Sonora, Mexico. The thermal input, in effect softened the country rocks, more in some places than in others. The rheology of the cooling plutons and the locally heated country rocks made the systems vulnerable to ductile flow. There are few localities where metamorphic core complexes lack any sign of known or inferred Tertiary intrusive bodies. The regional system, consisted of domains of cool rigid rocks and contrasting domains of hot, ductile rocks. Furthermore, the upper crustal rocks were not marked by lateral continuity of lithotectonic units. Instead, the combination of Jurassic and Laramide faulting led to domains of Precambrian basement juxtaposed laterally against domains of Paleozoic and/or Mesozoic layered strata; or domains of folded, thrusted strata juxtaposed against domains of homoclinal strata; or domains featuring complete sections of Precambrian-Paleozoic-Mesozoic rocks juxtaposed against domains where Cretaceous strata rests nonconformably on Precambrian basement. The combined effects of thermal and mechanical anisotropy may have profoundly influenced the location and nature of individual core complex terranes.

Given the thermal and mechanical condition of the upper crustal rocks, what regional tectonic movements gave rise to the penetrative componental movements which produced mylonitic tectonite? Based on relationships presented here and elsewhere (Davis, 1977, 1980; Davis and Coney, 1979; Davis and others, 1975), the mylonitic tectonite fabric is considered to be a flattening/extension fabric, with direction of extension parallel to penetrative mineral lineation. The fabric seems best explained by ductile normal shear within gently dipping curvitabular zones of simple-shear. The
The net effect was to produce younger-on-older separation relations. The absence of older-on-younger displacements suggests that the formation of mylonitic tectonite did not accompany regional shortening.

If it is true that the mylonitic tectonite expresses normal simple-shear, then it suggests that the regional tectonic movements stretched the upper-crustal rocks producing giant pull-apart. The cause of the stretching is not known.

As a response to stretching, the heterogeneous upper crust was partitioned into an array of rigid blocks with ductile margins. The boundaries between adjacent rigid blocks were largely determined by locations of hot, ductile crustal rocks. It was in these zones that a major percentage of the regional upper crustal stretching strain was accommodated. The megaboudiné-like rigid crustal units became marked by shoulders and necks of normal shearing characterized by profound thinning and flattening. At depth, these zones of normal shearing were marked by thick zones of mylonitic tectonite made at the expense of Precambrian basement. Upwards, these zones became thinner, converting Palaeozoic and Mesozoic strata into tectonite. Perhaps the projection of the zones to the surface was marked by distributed, imbricate normal faulting, producing growth-fault basins of sediment accumulation. Downward, the zones may project into horizontal, lithospheric to asthenospheric regimes of lamellar flow.

As cooling of the system took place, and as the 'zone' of brittle/ductile interface slowly descended, some already-formed mylonitic tectonite was disrupted, rotated, and microbrecciated by closely spaced, mesoskopically penetrative, normal-slip listric faults. This process was probably an upper-level, brittle counterpart of still-active mylonite-tectonite formation at greater depth. The kinematic coordination is perfect. While mylonitic tectonite and microbrecciated mylonitic tectonite were being formed, hanging-wall, upper-plate rocks continued to move downward and laterally, bringing rocks once far removed from the normal shear zones to positions directly above the mylonitic tectonite. This tectonic denudation largely fashioned the dramatic contrast between rocks of the mylonitic tectonite zone and rocks above.

Abrupt cooling of the mylonitic tectonite rocks of the metamorphic core complexes throughout the western Cordillera to the south of the Snake River plain is recorded in mid-Tertiary K-Ar and fission-track ages (about 25 m.y.). Most workers have ascribed this cooling event to 'uplift'. The structural implications of such uplift are not clear. A speculative interpretation is that the 25 m.y. ± cooling ages record development of the décollement and its ledge of microbreccia by major imbricate listric normal faulting. The major translation can be thought of as an 'unroofing'
in some respects, thus promoting cooling. The normal faults coalesced at or near the uppermost level of mylonitic tectonite, in some cases within, in other cases below the microbrecciated mylonitic tectonite. The positioning of the décollement zone may have been predetermined by mechanical and not thermal factors. The trigger for this listric faulting collapse may have been the brittle rifting of rigid megaboudin units and/or boundaries between units. The rifting prompted greater extensional strain in the upper crust as a whole. Ignimbritic volcanics (25 m.y.+) exploded out of the deeply rifted crust throughout the entire belt of metamorphic core complexes.

It is shocking to realize that all of the events which proceeded from 55 m.y.(?) to 25 m.y. may have accompanied and/or may have been produced by the concurrent formation of mylonitic tectonite (at some depth). However, the emplacement of detachments in post-25 m.y. time, largely mid-Miocene, must have had a cause which is not directly related to the formation of tectonite, even though the slip-line direction for translation of the detachments identically parallels the trend of penetrative lineation in nearby or underlying tectonite.

The mid-Miocene component of emplacement of the detachments may have been sensitive to and triggered by the inferred earlier normal-slip descent of hanging-wall terranes above zones of mylonitic tectonite. For the pure mid-Miocene detachments, i.e. those containing only mid-Miocene strata, were translated perfectly parallel to the direction of the penetrative lineation in the closest mylonitic tectonite, and apparently in the same sense of slip as that of simple-shear in the tectonite. The structure profiles of such mid-Miocene detachments convey the properties of slump- or toreva-block tectonics. Tempting as it is to explain the listric fault profiles as the result of stretching of brittle multilayers atop a ductile stretching medium, the geologic facts disclose that the listric fault blocks now rest on footwall which shows no sign of penetrative flow movements. Taken together, the relationships suggest that deep upper-crustal slumping took place in the mid-Miocene. In essence, the stability of rocks at the highest structural levels was weakened by the cumulative effect of normal-slip displacements during tectonite and décollement zone formation. Those rocks nearest the margins of the boudin-like lithospheric blocks 'felt' a removal (or weakening) of lateral support, and were translated along curved normal faults in the direction of the earlier stretched and descended terranes. The event represents the upper crustal brittle collapse of rocks occupying the outer margins of megaboudins. The composite symmetry of the deformational system is striking.
All of these events preceded the classic Basin and Range high-angle faulting. Such faulting was initiated 10 to 15 m.y. ago, and its dynamics of origin seems to bear no relation to the earlier Tertiary deformations. The surprising turn of events is that the Basin and Range province owes most of its distinctive and unusual tectonic characteristics to the development of metamorphic core complexes and to mid-Miocene listric normal faulting, not to the simpler imprint of high-angle normal faults.
CHAPTER 4

A CONCEPTUAL BASIS FOR THE OCCURRENCE OF URANIUM IN
CORDILLERAN METAMORPHIC CORE COMPLEXES

By

Stephen J. Reynolds

INTRODUCTION

As discussed in previous chapters, the evolution of Cordilleran metamorphic core complexes entails numerous processes, some of which operated simultaneously while others were sequential but integral components of a prolonged geologic history. In order to properly evaluate the uranium favorability of the complexes, all processes that were active within them must be examined to determine whether they are capable of concentrating uranium. Those that are should be identified and individually scrutinized to evaluate their possible role as agents of mineralization. Table 4-1 lists the most important of the processes that are known or hypothesized to have occurred during evolution of the core complexes. The oldest events typically recorded in rocks of the crystalline core of the complexes include plutonism, metamorphism, metasomatism, and anatexis. In structurally high levels of the core, these are overprinted by pervasive mylonitization. Along the margins of the core brecciation, cataclasis, chloritization, hematite deposition, and hydrothermal alteration and mineralization are all spatially associated with the dislocation surface (décollement). In rocks above the dislocation surface, listric normal faulting, tilting, and syntectonic sedimentation, volcanism, and mineralization are common.

It is instructive to consider these core-complex processes within the framework of a recent preliminary classification of uranium deposits proposed by Bendix Field Engineering Corporation (Mickle, 1978). Table 4-2 displays the classes of uranium deposits in this classification. The Bendix classification is based
TABLE 4-1

Processes that typically occurred during the evolution of Cordilleran metamorphic core complexes

Processes that occurred in rocks above the dislocation zone

- tilting and rotation
- listric-normal faulting
- various other styles of deformation
- syn-tectonic sedimentation, volcanism, and plutonism

Processes that occurred in the dislocation zone

- faulting
- brecciation
- jointing
- retrograde metamorphism
- hydrothermal alteration and mineralization
- chloritization
- hematite deposition

Processes that occurred in the crystalline core

- pluton emplacement
- magmatic differentiation
- pegmatite formation
- metamorphism
- metasomatism
- anatexis
- mylonitization
- folding
- faulting
- boudinage formation
- intrusion of dikes
- final cooling
TABLE 4-2

CLASSIFICATION OF URANIUM DEPOSITS PROPOSED BY MICKLE

Deposits in sedimentary rocks

Placer
Quartz-pebble conglomerate
Marine black shale
Phosphorite
Water
Lignite, coal, carbonaceous shale
Evaporative precipitates
Limestone
Sandstone

Deposits in and related to plutonic igneous rocks

Orthomagmatic
Pegmatitic
Magmatic hydrothermal
Contact metasomatic
Autometasomatic
Authigenic
Allogenic
Anatectic

Volcanogenic uranium deposits

Initial magmatic
Pneumatogenic
Hydroauthigenic
Hydroallogenic

Uranium deposits of uncertain genesis

Unconformity-related deposits
Monometallic
Polymetallic

Vein-type deposits in metamorphic rocks
Monometallic
Polymetallic

Vein-type deposits in sedimentary rocks

This classification is discussed in Mickle (1978).
on genesis and the nature of mineralization, thereby facilitating direct comparison of core-complex processes to those responsible for uranium mineralization. Such a comparison reveals the broad spectrum of uranium occurrences that are conceivably present in Cordilleran metamorphic core complexes.

Uranium occurrences within and adjacent to the core complexes are formed by two fundamental kinds of processes: those intrinsic or extrinsic to the evolution of the complexes. Intrinsic processes can be grouped into four categories: plutonic, metamorphic, mylonitic and dislocation-surface-related. In this chapter, the important aspects of each of these intrinsic processes are summarized, utilizing appropriate examples from various Cordilleran metamorphic core complexes. This is followed by a discussion of the relationship between each process and the geochemical behavior and geologic occurrence of uranium. Exemplary deposits of uranium that were formed or at least influenced by each process are provided, both from throughout the world and from within the core complexes. The discussion of extrinsic processes is, by necessity, less comprehensive because of the infinite variety of both processes and resulting uranium occurrences that are incidental to the core complexes. Nevertheless, an understanding of both intrinsic and extrinsic processes is essential for constructing a conceptual basis for the occurrence of uranium in Cordilleran metamorphic core complexes.

A CONCEPTUAL BASIS FOR THE OCCURRENCE OF URANIUM RELATED TO PLUTONIC PROCESSES

Plutonic Processes in Cordilleran Metamorphic Core Complexes

All Cordilleran metamorphic core complexes are characterized by plutonic rocks of one type or another. Plutons constitute a majority of exposed rocks in the crystalline cores of approximately half of the complexes. Some notable examples include the Santa Catalina, Rincon, Tortolita, Coyote and Pozo Verde Mountains of southern Arizona; the South Mountains of central Arizona; the Kern Mountains of central Alaska; the Bitterroot front of Montana and Idaho; and the Selkirk crest of Northern Idaho. Individual plutonic bodies range in size from true batholiths (for example the Santa Catalina, Bitterroot and Selkirk Mountains) to thin, laterally discontinuous dikes which may nevertheless congregate to form remarkable swarms such as those in the Tortolita, South, and Harquahala Mountains of Arizona and the western Whipple Mountains of California. Plutonic rocks in the complexes can be somewhat arbitrarily subdivided into five general groups: 1) muscovite
or two-mica granite and granodiorite; 2) biotite- and hornblende-bearing granitoid rocks of granitic to dioritic composition; 3) biotite-bearing, coarse-grained granite; 4) pegmatite and aplitic bodies; and 5) dikes of rhyolite, microdiorite, and all compositions in between.

Some of the most common plutonic rocks in the complexes are muscovite-bearing or two-mica granites and granodiorites. Granites are generally leucocratic with minor amounts of biotite and garnet, while in granodiorites the abundance of biotite is greater, ordinarily exceeding that of muscovite. All muscovite-bearing plutons in the complexes are characterized by extensive pegmatitic, aplitic and alaskitic phases, especially near their roofs where water saturation was most pronounced. The granites and granodiorites have relatively high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, indicating a significant crustal component within the parent magmas. Illustrative examples are the Eocene Wilderness granite (Santa Catalina Mountains; Shakel, 1978; Keith and others, 1980), Late Cretaceous muscovite granite of the Ruby Mountains, (Howard and others, 1979) and granites of the Mount Spokane area of the Selkirk Mountains (Weissenborn and Weiss, 1976). Within the context of this project, we have extensively studied and sampled muscovite-bearing granites of the Ruby, Harcuvar, Harquahala, and Santa Catalina - Rincon - Tortolita complexes. Results of major and minor element chemical analyses are discussed in Chapter 5 and listed in Appendix E.

In contrast to these muscovitic granites, another suite of plutonic rocks in the complexes contains hornblende, spherne, and biotite, but no muscovite or garnet. These range in composition from granite to diorite, and are in some cases markedly porphyritic. They occur as discrete plutons and have relatively diminished apical pegmatite phases, although thin aplite dikes may be scattered throughout large parts of the plutons. Excellent examples are the middle Tertiary Catalina Granite (Santa Catalina Mountains; Shakel, 1978; Keith and others, 1980), Jurassic Kitt Peak granite (near Coyote Mountains; Haxel and others, 1980), middle Tertiary Harrison Pass Granite (Ruby Mountains; Howard and others, 1979), and the Silver Point and other Eocene granites of northeastern Washington (Miller and Engels, 1975; Pearson and Obradovich, 1977). The relationship between chemistry of these plutons and their uranium content is described in Chapter 5.

Other plutonic rocks in the core complexes are simple biotite granites and granodiorites that contain neither muscovite nor hornblende. These granites are commonly porphyritic and have local pegmatitic phases. At least some of the granites are Precambrian such as the Oracle and equivalent 1.45 b.y. old granites of
Arizona, (Silver, 1978) and probably also the Laclede and Newman Lake augen gneiss of the Selkirk Mountains (Clark, 1967; Weissenborn and Weis, 1976). A Phanerozoic analog is the coarse-grained phase of what has been collectively mapped as Jurassic granite in the Ruby Mountains (Howard and others, 1979).

Exposures of pegmatitic and aplitic rocks are widespread throughout every Cordilleran metamorphic core complex. In some cases, these are demonstrably phases of one of the three types of plutonic rocks discussed above. The muscovite granites, in particular, are especially notorious for their extensive pegmatites, aplites, and alaskites. There are other pegmatities in the complexes that are not easily relateable to any individual pluton, but are instead probably general products of intense metamorphism, metasomatism, or anatexis. Included with those of uncertain genesis are some pegmatites in the forerange of the Santa Catalina Mountains (Keith and others, 1980), in the eastern Harquahala and Harcuvar Mountains (Rehrig and Reynolds, 1980), the northern Snake Range (Hose and Blake, 1976), the Angel Lake area of the East Humboldt Mountains (Snelson, 1957), and the Wildhorse Canyon region of the Pioneer Mountains, Idaho (Dover, 1969).

Dikes are present in all core complexes and are exceptionally abundant in some. In Arizona, distinctive microdiorite dikes occur in major swarms in the Tortolita, South Mountains, White Tank, and Harquahala-Harcuvar complexes. Dikes of other compositions including granodiorite and granite are also prevalent constituents of these swarms. Similar swarms are widespread along the Colorado River in California (Davis and others, 1979, 1980). Dikes such as these are much less numerous in core complexes outside of Arizona and California.

Uranium and Plutonic Processes

The distribution of radioelements in a layered earth model has been reviewed by Gabelman (1977) utilizing meteoritic abundances, models of terrestrial lithologies, and heat-flow data. He concludes that probably more than half of the earth's uranium has been transferred to the crust, with magmatic and pneumatolitic processes being the primary agents of transport. It follows that plutonism is fundamental in governing the worldwide distribution of uranium and other radioelements. Numerous papers have examined the general relationship between uranium and plutonism (Adler, 1977; Bohse and others, 1974; Gabelman, 1977; Heier and Lambert, 1965; Heier, 1973; Klepper and Wyant, 1956; Kostov, 1977; Larsen and others, 1956; Larsen and Gottfried, 1961; Lyons, 1964; Marjaniemi and Bassler, 1972; McKelvey, 1956; Neuerburg, 1956;
Within a plutonic rock, uranium and thorium may reside in a number of mineralogical and structural sites (see Neuerburg, 1956; Sorenson, 1977). Much of the uranium and thorium occurs in receptive accessory minerals (such as sphene, zircon, apatite, epidote, allanite, and monazite), microinclusions of radioactive minerals (i.e. uraninite), and concentrated along structural defects and grain boundaries of major rock-forming minerals. It has been demonstrated that much of the uranium in plutonic rocks is only weakly bound to the rock and is situated favorably to be leached by dilute acidic solutions. Therefore, plutons are excellent source rocks because they readily release uranium which can subsequently be redistributed to form various types of uranium deposits.

Due to their large ionic radii and high valence, uranium and thorium do not readily substitute isomorphously into the major rock-forming minerals during magmatic crystallization. Also, the concentration of both elements is probably too low in most magmas to permit the extensive crystallization of minerals in which either element is the principle phase. As a consequence, both uranium and thorium are preferentially partitioned into the residual melt during crystallization and are therefore relatively enriched in more differentiated rock types such as granite and syenite. In a normal magmatic differentiation series, the last rock to crystallize (i.e. the late magmatic residuum) should be the most uraniferous.

The abundances of uranium and thorium, like those of the major elements, reflect the degree of differentiation of igneous rocks. No detailed information is available regarding the relationship between uranium content and magmatic differentiation in plutonic rocks of the core complexes, except that included in Chapter 5 of this report. However, certain granitoid rocks in the complexes exhibit similarities to the Peninsular batholith of southern California, for which there is abundant whole-rock and uranium geochemical data (Larsen and Gottfried, 1961; Larsen, 1948). Figures 4-1a and 4-1b indicate the relationship between uranium content and variation in major element abundances of average plutonic rocks in the Peninsular batholith of southern California. The diagrams verify that uranium will generally be concentrated in plutonic rocks that are high in K, Si, and K+Na, and low in Ca, Mg, Fe, Al, and Ti. For those elements that vary in a linear fashion with differentiation (i.e. Si and Ca), the
Figure 4-1. Uranium - major element variation diagrams for average rock types of the Peninsular batholith of southern California. Data are from Larsen and others (1961) and Baird and others (1979).
horizontal axes of the graphs are essentially linear scales of
differentiation or fractional crystallization. Viewed in this
manner, the variation lines or "curves" for each element ex-
plicitely reflect the interrelationship of uranium abundance
and differentiation. Importantly, most of the curves suggest
that uranium content increases exponentially with differentiation
(i.e. toward the left on the calcium line). The shape of the
curves is that predicted by numerical trace-element models (Shaw,
1970; Hanson, 1978) for an element that is preferentially par-
titioned into the residual melt during progressive crystallization
or Rayleigh differentiation. Trace element models using
this data and equations contained in the papers referenced above
require that the bulk distribution coefficient \( D = \sum X_i K_d \)
where \( X_i \) is the weight fraction of a given mineral \( i \) and \( K_d \) is
the mineral/melt distribution coefficient of a trace element
for mineral \( i \) of uranium be very close to zero. In other
words, the models suggest that a very large proportion of the
uranium remains in the melt during fractional crystallization
and that because of the exponential nature of the curves, late-
stage residual melts could contain very high uranium contents.
This is a very important concept to consider in evaluating the
uranium potential of plutonic rocks both within and outside of
the core complexes. It predicts that late in the differentiation
sequence, the uranium contents of the magma will increase dramatic-
tively for relatively small degrees of fractional crystallization.
The implications of this will be evident in discussions below
that document that for even relatively "differentiated" rocks like
muscovite granites, the degree of fractionation, as revealed by
major element abundances, has a strong influence on uranium con-
tent.

Another inference that can be drawn from Figure 4-la is
that high uranium contents do not strictly correlate with high
sodium contents. This observation is congruent with others
made during this project concerning muscovite granites of the
core complexes that have very low uranium abundances but relatively
high sodium contents. The geochemistry of these peculiar Cordil-
leran muscovite granites is discussed in detail in chapter 5.

Rock type, like major element chemistry, may be an indica-
tion of the uranium favorability of plutonic rocks (Castor and
others, 1977; Nishimori and others, 1977). For example, many
riebeckite-bearing, peralkaline granites and syenites are anom-
ously uraniferous (Sorenson, 1977). On the opposite extreme of
alumina saturation are peraluminous muscovite granites, some of
which are likewise uraniferous (such as the Hercynian two-mica
granites of France; Cuney, 1978). The extent to which all
riebeckite- or muscovite-bearing granites are anomalously high
is uranium content has not been adequately discussed in the
literature. Evaluation of the favorability of muscovite granites is especially critical in assessing the uranium potential of Cordilleran metamorphic core complexes since nearly every complex contains muscovite granites. For this reason, the world literature was extensively surveyed in order to amass as much chemical, isotopic, and mineralogic data as possible on muscovite granites and their peraluminous kindred. The results of this survey clearly indicate that the uranium contents of some muscovite granites are very high (over 20 ppm) while others are exceptionally low (less than 1 ppm). The relationships between contents of uranium versus those of major elements for this world data base are depicted on Figures 4-2 to 4-7. The figures reveal that uraniferous muscovite granites have $K_2O$ and $K_2O+Na_2O$ contents that exceed critical threshold values of 4.2 and 7.5 percent, respectively. Also, a vast majority of uraniferous muscovite granites have less than 1.5 percent CaO. Although not presented here, data on other major and minor elements such as MgO, MnO, TiO$_2$, Rb, Sr, and Rb/Sr also exhibit critical threshold values on similar plots. Of the figures included here, only $Na_2O$ (Figure 4-3) does not display a clear threshold for all the data. If the low $Na_2O$ rocks from Australia are deleted from the diagram, there is no correlation of $Na_2O$ and uranium. Earlier in this section, evidence was presented concerning a similar lack of correlation for metaluminous rocks of the southern California batholith.

The diagrams also disclose that 6 ppm may be a natural division between rocks of "normal" versus anomalously high uranium abundances. Interestingly, the 6 ppm cut-off is one and a half times the 4 ppm value often quoted as average for granites.

Utilizing this 6 ppm division, Figures 4-8 through 4-15 are element-element plots of anomalous (greater 6 ppm) versus "normal" (6 ppm or less) peraluminous granitic rocks. On most of these diagrams, the anomalous granites are consistently segregated into a confined portion of the total granite field. The overlap between the uraniferous and "normal" groups is minimized on the $K_2O-CaO$, $K_2O-Na_2O$, and $K_2O-CaO-Na_2O$ plots. Potassium content is clearly the best discriminating factor between the two groups. Except for the silica plots, the remainder of the figures are also useful in distinguishing the groups. These plots represent a standard, known data base to which other muscovite granites may be compared to evaluate their uranium favorability. Muscovite granites of the Cordilleran metamorphic core complexes are compared to these graphs in Chapter 5 to provide insight on their favorability. As discussed in that chapter, many muscovite granites in the core complexes have chemical compositions that predict that they will have low uranium contents, an inference verified by radiometric and uranium analyses.
Figure 4-2. U - K$_2$O variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data. Symbols for Figures 4-2 through 4-7 are as follows: triangles - Australia, squares - Europe (Hercynian), diamonds - Himalaya, hexagons - New Hampshire, pentagons - South Africa. Data sources are included in Appendix B.
Figure 4-3. $U - \text{Na}_2\text{O}$ variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.

Figure 4-4. $U - (\text{K}_2\text{O} + \text{Na}_2\text{O})$ variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.
Figure 4-5. U - CaO variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.

Figure 4-6. U - SiO$_2$ variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.
Figure 4-7. U - alkali-lime variation diagrams for peraluminous and/or muscovite-bearing granitoids: world-wide data.
Figure 4-8. CaO - K₂O variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.

Figure 4-9. MgO - K₂O variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data. Shaded circles represent samples with over 6 ppm uranium.
Figure 4-10. Na₂O - K₂O variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.

Figure 4-11. K₂O - SiO₂ variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data. Shaded circles represent samples with over 6 ppm uranium.
Figure 4-12. CaO - SiO₂ variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data.

Figure 4-13. CaO - (Na₂O + K₂O) variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data. Shaded circles represent samples with over 6 ppm uranium.
Figure 4-14. (FeO + MgO) - (Na₂O + K₂O) variation diagram for peraluminous and/or muscovite-bearing granitoids: world-wide data. Shaded circles represent samples with over 6 ppm uranium.
Figure 4-15. Ternary diagrams for peraluminous and/or muscovite-bearing granitoids: world-wide data. Shaded circles represent samples with over 6 ppm uranium.
Research on the dependence of uranium content on whole-rock chemistry has also been extended to include other types of igneous rocks. The results, as more fully discussed in Chapter 5, are analogous to those observed for peraluminous rocks. Specifically, rocks with high potassium contents or alkaline tendencies are most favorable. Several granites in the core complexes which have high uranium contents are metaluminous with high potassium contents. The chemistry of these granites contrasts strongly with that of the low-potassium, low-uranium peraluminous muscovite granites of the Arizona and California core complexes. Throughout the world, there are highly uraniferous plutons that are peralkaline (alumina undersaturated) rather than peraluminous (alumina oversaturated). One such pluton is the Bokan Mountain granite of Alaska, (Mackevett, 1963), whose chemistry plots within the fields of uraniferous muscovite granites on the figures provided earlier in this section. The tectonic setting and genesis of the Bokan Mountain granite are drastically different from those for the uraniferous muscovite granites, yet both share similar chemical attributes, except for alumina saturation. Therefore, the diagrams presented earlier may be equally applicable, with slight modification, to other types of granites. Both granite types also share another important characteristic: they represent extreme end members of the alumina saturation continuum, albeit opposite ends.

In addition to whole-rock chemistry, volatile content of magmas, especially that of water, is highly influential in determining the behavior of uranium during magmatic crystallization. Feldspar and quartz, two of the principal constituents of granitoid rocks, are stoichiometrically anhydrous, so their crystallization causes a progressive enrichment of volatiles in the melt. Eventually, the magma may become saturated in and begin exsolving volatiles (second boiling) (Burnham, 1967; Jahns and Burnham, 1969; Luth and Tuttle, 1968; Holland, 1972; Whitney, 1975). Uranium will under normal circumstances be partitioned into the exsolved hydrous fluids which are also enriched in alkali, halogen, and large-ion lithophile elements. As pointed out by Neuerburg (1956), if a magma crystallizes as a closed system (retaining any exsolved hydrous phase), then the final plutonic rock will contain most, if not all, of the magmatic uranium. Many uraniferous pegmatites in the apical, water-saturated parts of plutonic complexes (see Heinrich, 1958; Nishimori and others, 1977; Mathews, 1978b) clearly represent magmas which, upon crystallization, retained their uranium-rich, hydrous fluids. Conversely, if the uranium-rich hydrous fluids are able to escape the magma prior to crystallization of prime uranium receptors, then most of the magmatic uranium may be lost along with the fluids. Subsequent to leaving the crystallizing pluton, these escaping fluids...
may form either magmatic-hydrothermal or contact-metasomatic uranium deposits.

Uranium Occurrences Related to Plutonic Processes

Mathews (1978a, b) has proposed a classification of uranium occurrences in and related to plutonic rocks that is based on a genetic theory of the behavior of uranium in the magmatic environment (Table 4-3). He has also discussed the geologic characteristics of each class of occurrences and how these characteristics may be utilized to evaluate the uranium favorability of plutonic rocks. These favorability criteria are directly applicable to the evaluation of the extensive plutonic terranes that are present in the Cordilleran metamorphic core complexes. In order to facilitate this application, those classes of uranium occurrence that are actually related to plutonic processes are briefly summarized below with emphasis given to those favorability criteria or geologic characteristics that are most relevant to the core complexes. For each class, world-wide examples are provided and any uranium occurrences within the core complexes that are possible analogs are specified. Since the potential for the various types of occurrences is not equal within the core complexes, the comparative favorabilities of each class will be discussed. Classes of uranium occurrences described by Mathews (1978a) that are not explicitly related to plutonic processes are discussed in later, more appropriate sections of this chapter. For example, uranium occurrences that are formed in plutonic rocks by groundwater are included in the section on extrinsic processes.

Orthomagmatic Class

According to Mathews (1978a, p. 19), orthomagmatic uranium occurrences "consist of syngenetic disseminations of uranium and uranium-bearing minerals formed during the orthomagmatic stage of magmatic crystallization. They are products of late-stage magmatic differentiation and develop through liquidous crystallization of uranium and uranium-bearing minerals." Orthomagmatic-class uranium occurrences are generally found in plutons that are leucocratic, porphyritic, and that commonly contain sodic minerals such as riebeckite and aegerine (in rocks of appropriate composition). The plutons have alkaline and peralkaline affinities and may either be quartz- or feldspathoid-bearing. Uranium occurs as disseminated minerals such as uraninite, betafite, coffinite, brannerite, uranothorite, and xenotime. Albition is common both to the pluton itself and to its wall rocks. Contents of uranium in the plutons are high (over 10 ppm), as are SiO₂, Na₂O, K₂O and total alkalis. Th/U ratios are typically well within the range of estimates for the "average" plutonic rock.
TABLE 4-3

CLASSIFICATION OF URANIUM OCCURRENCES
RELATED TO PLUTONIC ROCKS

<table>
<thead>
<tr>
<th>Class</th>
<th>Mode of Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orthomagmatic Class (310)</td>
<td>Formed during the orthomagmatic stage of crystallization</td>
</tr>
<tr>
<td>Pegmatitic Class (320)</td>
<td>Formed during the pegmatitic stage of magmatic differentiation</td>
</tr>
<tr>
<td>Magmatic-Hydrothermal Class (330)</td>
<td>Formed during the hydrothermal stage of magmatic evolution</td>
</tr>
<tr>
<td>Contact-Metasomatic Class (340)</td>
<td>Formed by contact metasomatism</td>
</tr>
<tr>
<td>Authigenic Class (360)</td>
<td>Formed authigenically by post-magmatic processes</td>
</tr>
<tr>
<td>Allogenic Class (370)</td>
<td>Formed allogenically by post-magmatic processes</td>
</tr>
<tr>
<td>Anatectic Class (380)</td>
<td>Formed by Anatexis</td>
</tr>
</tbody>
</table>

This classification is discussed by Mathews (1978a).
Mathews gives Bokan Mountain, Alaska and the bostonite dikes of the Colorado Front Range as prime examples of this class of occurrence.

There are few if any peralkaline granites similar to Bokan Mountain within the confines of the Cordilleran metamorphic core complexes. Alkalic plutons near the Okanagan and Kettle complexes of Washington that were sampled by Castor and others (1977) do not have high contents of uranium. Most Eocene granites of Montana and Idaho are somewhat alkalic and exhibit approximately twice the radioactivity of the Cretaceous Idaho batholith (Bennett, 1980; Swanberg and Blackwell, 1973). Several, such as the Bighorn Crags granite of east-central Idaho, have high uranium contents and are associated with uranium mineralization. Eocene plutons skirt the western and southern fringes of the Bitterroot dome and are present both within and adjacent to the Pioneer complex. It is unknown whether any of these Eocene granites have enough uranium contents to be considered significant orthomagmatic-class uranium occurrences. Syenites and other alkalic plutons are present near the core complexes of the Papago Indian Reservation of Arizona (Haxel and others, 1980). The uranium content of every one of these Jurassic plutons is unknown but those evidently sampled by Marjaneimi and Bassler (1972) are not anomalous.

While sampling in the Ruby Mountains of Nevada, I observed that a certain monzogranite phase of what has been collectively mapped as Jurassic granite was consistently more radioactive (up to 500 cps) than any other granite measured during the remainder of the project. The granite is characterized by high alkaline contents (mostly potassium) and uranium abundances that average over 8 ppm and are as high as 13.8 ppm (by neutron activation). Petrographic examinations by James DuBois (1980, personal communication) indicate that the granite contains abundant accessory minerals such as allanite. Although no uranium mineralization is presently known to be genetically related to the granite, the presence of the granite verifies that uranium-rich, alkali-rich plutons do indeed exist within the core complexes.

Pegmatitic Class

Mathews (1978a, b) indicates that pegmatitic and aplite uranium occurrences are formed by pegmatitic fluids derived by water saturation during late-stage magmatic evolution. With magmatic crystallization, uranium becomes progressively concentrated in residual melts and upon water saturation of the magma (second boiling) will preferentially enter the hydrous fluid phase. The resulting uranium-rich fluid may induce crystallization of a uraniferous pegmatite. Factors influencing the uranium content of the pegmatite include the uranium content of the
original magma, the pre-saturation crystallization history (especially whether uranium is removed from the magma by crystallization of uranium-rich phases), and the specifics of the pegmatite fluid (i.e. the extent of pre-solidification volatile loss). Uraniferous pegmatites tend to exhibit sodic mineralogies, and contain fluorite and hematite, and logically are associated with plutons that are anomalously radioactive. Uraniferous pegmatites, like uraniferous granites, are characterized by high abundances of SiO₂, K₂O, Na₂O and total alkalis; and low contents of Fe, MgO, and CaO. However, in contrast to muscovite granites, Mathews (1978b) states that uraniferous pegmatites are generally undersaturated in aluminum (subaluminous to peralkaline), although the two have identical characteristic trace constituents (Li, Be, Mo, W, Th, U, Zn, Sn, Ta, and Nb). Albitization is commonly present both in the pegmatite and its wall rocks. Examples of pegmatite-class occurrences are in the Bokan Mountain granite of Alaska, the Bancroft district of Canada, and Crocker Well, Australia. Additional information on uranium in pegmatites is contained in Page (1950), Mawdsley (1952), Heinrich (1958), Little, (1970), Nishimori and others (1977) and Mathews (1978b).

Within Cordilleran metamorphic core complexes, pegmatitic occurrences are one of the most important classes of deposits. The best examples are the numerous uraniferous pegmatites that pervade the crystalline core of the Kettle complex (see Huntting, 1956; Weissenborn and Moen, 1974). The pegmatites locally have muscovite and garnet, indicating that they are probably peraluminous, not alumina undersaturated like the "typical" uraniferous pegmatite described by Mathews (1978b). The pegmatites are interlayered with high-grade biotite schist, biotite gneiss, amphibolite, quartzite, and a variety of metasedimentary rocks. Some mineralized pegmatites are mylonitic and others have been brecciated and weakly stained with iron oxides. Uranium mineralization is evidently in the form of uraninite and uranophane (Huntting, 1956) and is commonly confined to the pegmatite itself, although there are a significant number of localities where the adjacent metamorphic rocks are also mineralized adjacent to the contact. The pegmatites of the Kettle complex should serve as the type-example of pegmatitic-class uranium occurrences within the core complex belt. The Kettle complex is undoubtedly the most favorable core complex for a pegmatitic uranium deposit.

Besides the Kettle area, uraniferous pegmatites are also present in the City of Rocks of the Albion Range, Dawley Canyon of the Ruby Mountains, and in several localities in the Selkirk Complex. These areas are favorable for pegmatitic uranium occurrences, but much less so than the Kettle complex. In most of the remaining metamorphic core complexes, pegmatites are not
favorable for pegmatitic class uranium deposits by virtue of their profound lack of uranium occurrences, low radioactivity, and association with plutons of low uranium contents. For example, the Eocene Wilderness muscovite granite of the Santa Catalina Mountains of southern Arizona has a roof-zone of abundant pegmatites, aplites, alaskites, and garnet schlieren (Keith and others, 1980). Only the garnet schlieren have anomalous radioactivity and uranium contents, indicating the strong affinity of uranium for garnet. The rest of the pegmatites, aplites, and alaskites have characteristically low radioactivity and uranium contents (less than 4 ppm). This can be attributed to the fact that the Wilderness granite, parent pluton for the pegmatites is also very low in uranium content (frequently less than 1 ppm). Simply, the muscovite granite had too little uranium to have permitted any significant concentration in the late-stage, water-saturated, residual melt. This demonstrates the importance of the uranium content of the pluton to which the pegmatites are genetically related. Clearly, uranium-poor granites are unfavorable for pegmatitic uranium, even if such granites have extensively developed pegmatitic phases. The unique geochemical evolution of the Wilderness granite has additional implications for its uranium favorability, as discussed more fully in chapter 5.

Magmatic-Hydrothermal Class

Occurrences of this class are formed by hydrothermal solutions that are generated during the final stages of magmatic crystallization (Mathews, 1978a, b). They consist of veins that are associated with uraniferous plutons, pegmatites, and dikes. Uraninite and (or) pitchblende, the characteristic uranium minerals, are variably accompanied by quartz, base-metal sulfides, fluorite, carbonate minerals, and precious metals. Silicification, chloritization, argillization and the deposition of hematite are the usual attendant alteration features. Veins at Radium Hill, Australia and the Boulder Batholith of Montana are examples given by Mathews for this class of occurrence.

Magmatic-hydrothermal uranium occurrences are undoubtedly present within Cordilleran metamorphic core complexes. Uranium veins are found in many localities, but confidently ascribing their origin to magmatic-hydrothermal fluids is not always easy. In the Hacruvar Mountains, minor amounts of uranium accompany copper and gold mineralization at the Doland and Bonanza Mines. Both occurrences are situated near microdioritic dikes which are constituents of a regional mid-Tertiary swarm. The copper, gold, and uranium may have all been deposited by the same magmatic-hydrothermal fluids.
There are also mineralized localities in the Idaho-Montana-Washington complexes which might be magmatic-hydrothermal class occurrences. For example, on the south flank of the Bitterroot complex, uranium is present in quartz veins which might be related to Eocene intrusions of the area. In north-eastern Washington, pitchblende occurs in stringers which cut quartz veins that contain molybdenite and other sulfides at the Spokane Molybdenum Mine which lies south of the Kettle complex, near the Midnite Mine (Wiessenborn and Moen, 1974). Additional magmatic hydrothermal occurrences could be given, but they are fairly insignificant, like most deposits of this type within the core complexes.

Contact-metasomatic Class

Contact-metasomatic uranium occurrences form by the reaction between uraniferous magmatic emanations and country rocks adjacent to a uraniferous pluton (Mathews, 1978a, b). Alternatively, some uranium may be remobilized from the country rocks and concentrated near the host rock-pluton contact. This class of occurrence is generally found in high-grade metamorphic terranes proximal to silicic plutons and pegmatites. Uraninite and thorianite are the most common primary minerals and zones of albitization, sericitization, silicification, and argillization may be extensively developed. At the Mary Kathleen deposit of Australia (the clearest example of this class), uranium mineralization is associated with a garnetiferous skarn developed near granitic plutons. Another example suggested by Mathews is the Wheeler Basin area of Colorado where uranium resides in biotitic lithologies of amphibolite-grade metasedimentary rocks that were intruded by uranium-rich pegmatites and granites.

No outstanding examples of this type of occurrence are obvious within the metamorphic core complexes. Mineralization in some areas of the Kettle complex is contained in metamorphic rocks and carbonate units adjacent to uraniferous pegmatites and granites, but the origin of this mineralization is uncertain. Some of these occurrences may have been formed by uraniferous fluids that were released by the pegmatites during intrusion and crystallization. There are additional areas in the core complex belt that appear to be favorable for this type of deposit, yet have no reported occurrences. In the Ruby Mountains of Nevada, uraniferous biotite granite has intruded a series of interlayered quartzites and calc-silicate rocks. During this project, numerous outcrops of both metasedimentary rock types were measured for total gamma counts but none were anomalous. Chemical analyses of samples of quartzite, calc-silicate rocks, and interlayered metasedimentary and pegmatitic units verify that uranium contents are low, generally less than 2 ppm (Appendix E). In the Kern and
Snake complexes of Nevada, radioactive plutons such as the Skinner Canyon granite intrude calcareous, argillaceous, and quartzitic sequences, but no occurrences have been reported. Tungsten mineralization occurs in this area and also in the Pioneer complex of Idaho and the Harquahala Mountains of Arizona but was evidently not accompanied by uranium. The presence of favorable conditions for contact-metasomatism in the core complexes suggests that this class of occurrence might indeed exist, although they have not been clearly recognized or described.

**Autometasomatic Class**

Autometasomatic uranium occurrences are the product of metasomatic reactions between late-magmatic, hydrous fluids and previously crystallized plutonic rocks (Mathews, 1978a, b). The late-stage magmatic fluids may be uranium-rich or may simply remobilize uranium within the plutonic body. Uranium minerals such as uraninite, coffinite, pyrochlore and thorianite are concentrated in altered plutonic rocks adjacent to structures that acted as conduits for mineralizing fluids. Fluorite, hematite and sulfides commonly accompany the uranium mineralization. There is a strong tendency for this type of occurrence to be associated with epizonal, subaluminous to peralkaline intrusives that contain high contents of magmatic uranium. Bokan Mountain, Alaska and the Lireui complex (Jurassic ring dikes) of Nigeria are examples of this class of deposit. Both areas are characterized by highly albitized, peralkaline intrusives.

There are no obvious autometasomatic class uranium occurrences within the confines of the Cordilleran metamorphic core complexes. There are numerous granites in the core complexes that have somewhat alkalic chemistries, but few if any of these are peralkaline. There are no reported areas of extensive albitization or other type of autometasomatism in the complexes. Alkaline intrusions are present in Washington near the complexes, but these are not notably uraniferous (Castor and others, 1977). Eocene granites in Idaho such as those near the Bitterroot and Pioneer complexes are epizonal, fairly radioactive, and associated with molybdenum, tungsten, tin, and uranium mineralization (Bennett, 1980), but do not contain evidence of widespread albitization or any other autometasomatism. The Cordilleran metamorphic complexes are not highly favorable for this class of uranium occurrence.

**Anatectic Class**

This class of uranium occurrence is found in pegmatite, alaskite, and aplite that are believed to have formed by crystallization of melts that were produced by ultrametamorphism.
and anatexis (Mathews, 1978a, b). Uranium mineralization occurs as uranium-rich quartzo-feldspathic sills, dikes, and lenses within migmatite complexes, with uranium being generally in primary minerals such as uraninite and uranothorite. Secondary uranium minerals are abundant and even predominant in some localities. Alteration in adjacent high-grade metamorphic rocks is negligible, but evidence of metasomatism of the rocks is frequently present. The Rossing deposit of Namibia, Africa is cited as the example of an anatectic deposit (Mathews, 1978a). Rossing is situated on the flank of a major gneiss dome (see Chapter 2) of late Precambrian age. Mineralization occurs both as primary uraninite in alaskite and as secondary minerals in country rocks and the alaskite.

Cordilleran metamorphic core complexes locally exhibit some evidence of ultrametamorphism and anatexis. Migmatites are common in the complexes as are muscovite granites whose initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios require a crustal component in the parent magmas. However, it is doubtful that most of these major granitic plutons represent in-situ anatectites. Instead, the evidence suggests that they have risen some distance out of their source region. There are two areas in the core complexes that may contain anatectic occurrences. Uraniferous pegmatites of the Kettle complex may represent partial melts of uraniferous sedimentary rocks. Anatexis is especially suggested for those uranium occurrences that are located in small, isolated pods and lenses of granitoid rocks of minimum-melt compositions. These possible in-situ anatectites contrast strongly with the numerous uraniferous pegmatites that are more continuous and cross-cutting. Uranium occurrences in the Green Creek metamorphics of the Albion Range may partly owe their origin to anatexis. This could only be verified by detailed field work and exhaustive geochemical scrutiny. A few uranium occurrences in other metamorphic core complexes might be anatectic, but the most favorable areas for this class of deposit are those previously mentioned.

Summary of Conceptual Basis for the Occurrence of Uranium Related to Plutonic Processes

The profusion of plutonic processes that exist in Cordilleran metamorphic core complexes dictates that a wide variety of plutonic-related uranium occurrences will also be present. The most favorable plutonic rocks for any type of plutonic-related uranium occurrence are invariably highly differentiated and characterized by high alkali content—especially potassium. Favorable plutons will also have low contents of elements such as calcium and magnesium and may either be silica-saturated or unsaturated. During magmatic crystallization, uranium will
become progressively partitioned into and enriched in the residual melt or any exsolved hydrous fluids. Extremely differentiated intrusions formed by solidification of such late-stage melts may have very high uranium contents. Uraniferous hydrous fluids exsolved and emanated by these magmas are potential metasomatic and hydrothermal mineralizers. Much of the uranium in plutonic rocks may be readily leachable, creating a potential for the formation of a variety of ground water-related post-magmatic uranium deposits (discussed in later sections).

Uranium occurrences in the core complexes are generally spatially and genetically related to highly differentiated plutonic rocks. Of the classes of occurrences discussed earlier in this chapter and by Mathews (1978a, b), the core complexes are especially unfavorable for only two; orthomagmatic and autometasomatic. These two classes of deposits are characteristically associated with peralkaline rocks, few or none of which are located in the core complexes. Cordilleran metamorphic core complexes are, as a species, favorable for pegmatitic, magmatic-hydrothermal, contact-metasomatic, and anatectic uranium occurrences. Each complex may have potential for different combinations of the occurrence types, and as summarized in Chapter 6 the over-all favorability of individual complexes ranges to both ends of the possible spectrum.
Every Cordilleran metamorphic core complex contains extensive exposures of metamorphic rocks. In some complexes like the Snake Range of Nevada and the Buckskin Mountains of western Arizona, metamorphic rocks constitute a majority of the exposed crystalline core. However, in a large number of complexes, classic metamorphic rocks are subordinate to plutonic rocks which range in texture from undeformed to strongly mylonitized. These latter, plutonic-dominated complexes are best exemplified by the Santa Catalina, Rincon, Tortolita, Picacho and South Mountains of Arizona, the Bitterroot Mountains of Idaho and Montana and the Selkirk crest in Idaho. In all the complexes, plutonic rocks are at one place or another foliated and rendered into rocks that might be called gneiss or schist. Commonly, the foliation within such plutons is mylonitic in character. Mylonitization will not be discussed in this section, but is instead dealt with extensively in the next section of this chapter. The present section deals with nonmylonitic metamorphic processes in general, and specifically with regional, high-grade, progressive metamorphism. Retrogressive metamorphism is discussed in another section of this chapter that is concerned with processes that are related to the formation of the dislocation surface.

Nonmylonitic metamorphic rocks within the Cordilleran metamorphic core complexes are similar to typical metamorphic rocks found throughout the world. They are diverse with regard to present-day lithologies and to inferred original protoliths. In a majority of the complexes, gneiss is the prevalent metamorphic rock. Compositional banding of alternating melanocratic and leucocratic lithologies occurs on a wide variety of scales. For example, within the Whipple Mountains of California and the Harcuvar Mountains of Arizona, layers dominated by biotitic or amphibolitic lithologies alternate with leucocratic units on a scale of tens or hundreds of meters. A similar large-scale interlayering of varying lithologies is demonstrated by maps of the Ruby (Howard, and others 1979), Kettle (Cheney, 1980), and other complexes. Within these large-scale layers are countless, smaller-magnitude compositional bands which may mimic the larger layers, but on a scale of centimeters. The term banded gneiss, which is commonly applied to this type of rock, refers to a wide variety of lithologic sequences with a variety of origins.
So-called banded gneiss of the Santa Catalina forerange in Arizona has a dark component of mylonitic Precambrian granite interlayered with thick sills of a light-colored, Tertiary granite. In the Harcuvar Mountains of western Arizona, dark components of "banded gneiss" consist of amphibolite, biotite-rich schist, foliated metadiorite and granodiorite. Leucocratic phases of the gneiss include simple quartz-feldspar pegmatites, sills of Tertiary (?) muscovite granite and Cretaceous (?) biotite granite, and numerous types of variably foliated granitoid rocks of uncertain age and origin. In the Ruby and East Humboldt Mountains of Nevada, sills of granite alternate with quartzite and metacarbonate units. Clearly, the types of lithologies referred to as "gneiss" are numerous, both within an individual complex and between different complexes.

Schist is also a common rock type within the core complexes. In some cases, such as the Albion-Raft River and Pioneer complexes, these schists are derived from sedimentary protoliths by classic progressive metamorphism. In at least several complexes, including the Santa Catalina-Rincon-Tortolita complexes, it can be demonstrated that apparently "normal-looking" schists have been formed by mylonitization of originally granitic protoliths. Since the mode of formation of these latter mylonitic schists is much different than those that result by progressive metamorphism, they will be discussed separately in the next section. Caution is advisable in interpreting the origin and protolith of schists within the core complexes.

Mineralogy of the gneisses and schists in the core complexes is generally indicative of an amphibolite grade of metamorphism. Accordingly there is commonly evidence for the mobilization, redistribution and segregation of chemical components. Locally, there must have been significant metasomatism and partial melting (anatexis), both of which result in major geochemical changes in the rock being affected. These changes extend to minor and trace elements (including uranium) as well as major elements. Progressive metamorphism influences the abundance and mineralogical disposition of uranium in rocks and is capable of liberating uraniferous hydrous fluids which might act as metasomatic or hydrothermal mineralizers. If partial melting (anatexis) occurs, uranium partitions into the accumulating melt, thereby enriching it and transforming it into a potential source of plutonic-related uranium occurrences.
Uranium and Metamorphic Processes

Uranium Abundances of Metamorphic Rocks

Metamorphism is a dominant process within the deeper levels of the crust and is also, under the proper circumstances (such as in the core complexes) able to affect rocks at comparatively shallow crustal levels. Numerous studies of metamorphic rocks have provided insight about the lithologies and processes that characterize the subterranean crust (see for example Fahrig and Eade, 1968; Fahrig and others, 1967; Heier, 1965, 1973; Heier and Adams, 1965; Lambert, 1971; Lambert and Heier, 1967, 1968a, 1968b; Mehnert, 1968; Sighinolfi, 1971; and many additional references in Appendix B). These studies have yielded models for the vertical distribution of lithologies, structures and geochemical abundances within the crust. Almost all of the geochemical models emphasize the importance of metamorphism and anatexis for accomplishing the upward transfer of incompatible elements, in general, and radioelements (U, Th, and K) in particular. The resultant effects of these two processes are recorded in the radioelement content of metamorphic rocks of the world, including those in the core complexes. There is significant variation in these contents, both on a local scale and between different regions of the earth. For example, average uranium and thorium values for samples of the Bear Province of Canada (including both metamorphic and plutonic rocks) are 8.1 and 35.7 ppm respectively (Eade and Fahrig, 1971), while nearly all samples of a widespread plagioclase gneiss in western Greenland have less than one ppm uranium and less than 15 ppm thorium (Kalsbeek, 1976). Uranium deposits of the Rum Jungle complex of Australia lie within a region whose crystalline basement is characterized by anomalously high uranium and thorium contents as compared with typical crystalline rocks of the world (Heier and Rhodes, 1966). The average uranium and thorium contents of metamorphic sialic basement in the world are given by Rogers and Adams (1969) as 3.5 and 10.9 ppm respectively. These compare with average values of 3.6 ppm uranium and 19.0 ppm thorium for granite as given by the same authors. Regions of the earth, like the Rum Jungle complex, that are underlain by uraniferous metamorphic rocks will be more favorable for uranium deposits than regions which are not. Applying this concept to the Cordilleran metamorphic core complexes would suggest that a first-order approximation of uranium favorability is provided by uranium contents of their metamorphic rocks. As discussed in Chapter 5, this would predict that the Kettle complex is the most favorable for uranium deposits.
As would be expected, different metamorphic rock types have correspondingly different radioelement abundances. Listed below are average radioelement contents for a limited number of samples for three rock types in the Spruce Pine study area of North Carolina (Galipeau and Ragland, 1978).

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>K (%)</th>
<th>Th/U</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amphibolite gneiss</td>
<td>0.65</td>
<td>4.1</td>
<td>1.5</td>
<td>4.4</td>
</tr>
<tr>
<td>Mica gneiss</td>
<td>1.15</td>
<td>7.7</td>
<td>1.6</td>
<td>3.0</td>
</tr>
<tr>
<td>Mica schist</td>
<td>2.5</td>
<td>10.9</td>
<td>4.9</td>
<td>4.5</td>
</tr>
</tbody>
</table>

Our observations in the Cordilleran metamorphic core complexes also indicate that mica schist and amphibolite gneiss generally have higher and lower radioactivity, respectively, than typical quartz-feldspathic gneiss. Predictably, a characteristically low radioactivity has been observed for mafic gneiss in other areas around the world (Burwash, 1979; Smithson and Brown, 1977; Zagruzina and Smyslov, 1978).

Major element chemical composition is inextricably linked to lithology and therefore also partly determines the uranium and thorium content of metamorphic rocks. In one study of metamorphic rocks thorium was found to correlate positively with aluminum, ferric iron, magnesium, titanium and several trace elements; and negatively with silica, calcium and phosphorous (Aparicio and Bellido, 1976). Several other studies have documented positive correlations between the incompatible elements U, Th, K, Rb, Pb, and Ba (Lambert and Heier, 1968; Sighinoifli, 1971; Burwash, 1979).

Mineralogical and Structural Setting of Uranium in Metamorphic Rocks

Uranium resides in a variety of mineralogical and structural sites within metamorphic rocks. Many minerals found in metamorphic rocks also occur in plutonic rocks, and it is probable that the affinity of a given mineral species for uranium is similar for the two classes of rocks. Therefore, data on uranium abundances of minerals in plutonic rocks (Larsen and others, 1956; Larsen and Gottfried, 1961; Klepper and Wyant, 1956; Sorenson, 1977) can provide insight into the mineralogical distribution of uranium in metamorphic rocks. A certain amount of uranium in plutonic rocks is present in isochemical substitution positions within accessory minerals such as sphene, zircon, epidote, and garnet. These same minerals might be expected to be prime uranium receptors in metamorphic rocks. Some uranium in plutonic rocks occurs associated with major rock-forming minerals (Larsen and Gottfried, 1961), but because quartz and feldspars have such low tendencies for incorporating uranium into their crystal lattices (Dostal and Capedri, 1975), much of
this uranium is probably located in structural defects and along grain boundaries (Neuerburg, 1956). Biotite and muscovite grains in plutonic rocks generally have higher uranium contents than quartz or feldspar, but this uranium may also occur within structural defects or else as inclusions of uranium-bearing minerals such as uraninite. This tendency for uranium to be associated with the micas is also evidently applicable to metamorphic rocks, because we have observed that mica schists are commonly the most radioactive type of metamorphic rock in the core complexes.

Two important studies of the distribution of uranium within metamorphic rocks have been recently published (Wilson, 1977; Dostal and Capedri, 1978). Both indicate that uranium in amphibolite-facies rocks exhibits a very nonuniform distribution within the rock and that much of the uranium occurs along grain boundaries and in structural defects including fractures, cleavage planes, kink bands and fold hinges. Uranium shows a preference for structures within micas in one case (Wilson, 1977) and amphiboles in the other (Dostal and Capedri, 1978). Relatively high concentrations of uranium are associated with biotite, mostly either as included uranium-rich accessory minerals or as grain-boundary deposits. The data suggests that very little of the uranium in the micas is actually accommodated within crystal lattices. Opaque minerals in the metamorphic rocks commonly have high uranium contents, as does garnet. A large proportion of the uranium in the samples examined by Wilson (1977) exchanges with water, even under low temperatures and from minerals with low concentrations of uranium. In marked contrast with this, uranium in granulite facies rocks (Dostal and Capedri, 1978) is more uniformly distributed, occurs almost exclusively in refractory accessory minerals, and would probably not exchange with water. The amount of uranium in structural defects within the granulitic rocks was extremely low or not detectable. Several logical conclusions from these results are that most of the structurally controlled uranium was lost from the rock during granulite metamorphism (Dostal and Capedri, 1978) and that the loss of this uranium may account for a majority of the uranium loss for the whole rock. Another implication from these studies is that uranium in amphibolite-facies rocks, such as those in the core complexes, is apt to be readily redistributed by either metamorphic or post-metamorphic meteoric fluids.

Influence of Progressive Metamorphism on Uranium Content

Metamorphic grade has a profound influence on the abundances of U, Th and K in metamorphic rocks. In most of the world's shields, it has been documented that granulite facies rocks have
much smaller concentrations of U, Th, and K than lower grade rocks. Heier and Adams (1965) give the following average uranium and thorium contents for different grade metamorphic rocks.

<table>
<thead>
<tr>
<th>Metamorphic Grade</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Epidote amphibolite</td>
<td>3.45</td>
<td>26.48</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>1.22</td>
<td>9.39</td>
</tr>
<tr>
<td>Low-grade granulite</td>
<td>0.88</td>
<td>4.09</td>
</tr>
<tr>
<td>High-grade granulite</td>
<td>0.39</td>
<td>0.93</td>
</tr>
</tbody>
</table>

This data clearly reveals that both uranium and thorium contents decrease with increased metamorphic grade. The influence of metamorphic grade on uranium content is most pronounced and best documented for the transition between granulite and lower grades (Dostal and Capedri, 1978; Fahrig and others, 1967; Gasparini and Mantovani, 1979; Heier, 1965, 1973; Heier and Adams, 1965; Lambert, 1971; Lambert and Heier, 1967, 1968b; Smithson and Brown, 1977). The amount of uranium in granulites is extremely low, generally within the range from 0.5 to 1.0 ppm. This is significantly less than the typical uranium abundance for amphibolite and lower grade rocks, indicating that the granulites have been depleted in uranium as well as other incompatible elements. Heier and Adams (1965) show evidence that low-grade granulite facies rocks have on the average about twice as much uranium as high-grade granulite rocks. They also indicate that lower grade amphibolite facies rocks have more uranium than high-grade amphibolite facies rocks. Although it is intuitive that uranium would be lost with progressive metamorphism and dehydration, not all studies have suggested that uranium is indeed remobilized and expelled from rocks during sub-granulite grades of metamorphism (Petrov and others, 1972, Dostal and Capedri, 1978). Atherton and Brotherton (1979) interpret variations in the uranium contents of metamorphic rocks in Scotland as reflecting difference provenances rather than being the result of movement of uranium during metamorphism. However, the data of Heier and Adams (given earlier in this section) clearly demonstrate that the average uranium and thorium abundances of metamorphic rocks increase systematically from low to high grades. Russian and eastern European authors (see for example Kostov, 1977) also report lower uranium and thorium abundances in higher grade rocks, which Yermolayev (1973) attributes to the mobilization of thorium and uranium. Finally, remobilization of uranium is supported by the observation that rocks in ancient shields are on the average more depleted than those of younger orogenic regions (Lambert and Heier, 1967, 1968a), presumably because the former have been subjected to a succession of metamorphic events.
The dependence of uranium content on metamorphic grade has important implications regarding the favorability of Cordilleran metamorphic core complexes because metamorphic rocks in the core complexes are almost universally amphibolite grade. Metamorphic conditions within the complexes have probably been sufficient to mobilize and redistribute uranium, but not high enough to deplete rocks to the low uranium and thorium concentrations typical of granulites. The amphibolite facies rocks might therefore be favorable hosts for "metamorphic" or "metasomatic" uranium deposits. Attainment of granulite conditions would be extremely detrimental to the possible formation of uranium deposits via metamorphism of originally uraniferous protoliths. Rocks of amphibolite grade are also more favorable uranium source-rocks than are those of granulite grade because the uranium in the former is commonly structurally controlled and readily leached, while the relatively small amount of uranium present in granulitic rocks is largely tied up in refractory accessory minerals.

Behavior of Uranium During Progressive Metamorphism, Metasomatism and Anatexis

There are many mechanisms by which rocks undergoing progressive metamorphism might be depleted in uranium (Lambert and Heier, 1968; Smithson and Heier, 1971; Sighinolfi, 1971; Yermolayev, 1971, 1973; Wilson, 1977, Dostal and Capedri, 1978). With the rise of temperature, rocks first probably lose any uranium that occurs within hydrous pore fluids. With further temperature increase, uranium-bearing fluids will be liberated via desorption from the surfaces of rock-forming minerals. Under metamorphic conditions, dissolution of uraniferous grain-boundary deposits and of principle uranium-bearing minerals (including micas and accessory minerals) would mobilize uranium. The breakdown of biotite might be especially important because it commonly contains inclusions of uraninite and other accessory minerals. Recrystallization of minerals could liberate uranium through some type of self-purification process. Alternatively, recrystallizing or newly forming minerals such as sphene might scavenge uranium and therefore deplete interstitial fluids (Lyons, 1964; Yermolayev, 1973).

The behavior of uranium is intimately related to the presence and behavior of a fluid phase. Reactions involving dehydration are prominent in progressive metamorphism and liberate much water from the structure of hydrous minerals. Uranium has a strong affinity for hydrous fluids and will be readily partitioned into such fluid phases (McKelvey, 1956; Smithson and Heier, 1971). The hydrous fluids also tend to become enriched in alkali
elements and certain base metals, especially in the presence of chloride (Orville, 1963; Burnham, 1967; Holland, 1972; Stern and Wylie, 1973). If these fluids remain within the rock from which they were evolved, then that rock will experience minimal depletion. However, these low-density fluids will probably tend to rise away from their metamorphic parent rock, which will consequently be depleted. Such ion-charged hydrous fluids are potentially capable of later hydrothermal or metasomatic deposition of uranium, alkali elements and base metals. Metasomatism may be a particularly important process for increasing the uranium content of metamorphic rocks. Under favorable conditions and in receptive host rocks, there is likely to be significant influx and redistribution of radioelements during metasomatism (Yermolayev, 1973; Berezina and others, 1976; Grebenchikov and others, 1977). For example Berezina and others (1976) describe metasomatites that are 30 to 50 times more radioactive than associated typical metamorphic rocks. In addition to displaying anomalous radioactivity, the metasomatic rocks also exhibit a more irregular distribution of uranium and contain a larger proportion of structurally controlled uranium than adjacent metamorphic rocks. Conceivably, metasomatism within the core complexes could either directly form ore bodies, or else simply modify metamorphic rocks into more favorable sources of leachable, structurally controlled uranium.

Under high grades of metamorphism (ultrametamorphism), rocks begin to undergo partial melting or anatexis (see discussions in Brown and Fyfe, 1970; Fyfe, 1973; Lambert, 1971; McCarthy and Kable, 1978; Mehnert, 1968; Shaw, 1970, 1978; Sighinolfi, 1971; Stern and Wyllie, 1973; White and Chapple, 1977; Wyllie, 1977; Yermolayev, 1973). The first melt produced by anatexis will consist of low-temperature melting fractions and will be compositionally similar to late-stage products of magmatic differentiation. The first melt is uniformly of a granitic composition, irrespective of the starting composition of the metamorphic protolith. Uranium and other incompatible elements will be preferentially partitioned into the granitic melt, leaving a residuum that is depleted in these elements. Partial melting may, therefore, be an efficient mechanism for depleting rocks to the low abundances of uranium, thorium and other incompatible elements that are typically observed in granulites. Whatever uranium lost by the residual fraction or restite will be incorporated into the granitic melt, thus greatly enriching it. Magmas formed by anatexis of uraniferous protoliths would have high uranium contents and as such be potential hosts or mineralizing agents for a variety of plutonic-type uranium deposits. Uraniferous pegmatites of the Kettle complex may owe their origin to the partial melting, either in-situ or at depth, of uranium-rich sedimentary protoliths.
Uranium Occurrences Related to Metamorphic Processes

A large proportion of the world's uranium reserves lies within metamorphic rocks of Australia, Canada, and Africa. Several distinct types of uranium occurrences exist within metamorphic rocks including: 1) unconformity-related, 2) vein-type, 3) plutonic-related, and 4) metamorphic-related. Presently the genesis of the first two types is uncertain (Mickle, 1978; Mathews, 1978c, d), and their relationship, if any, to metamorphic processes cannot be ascertained. Due to this uncertainty, a brief characterization of the two will be included in this section, realizing that they may be related only indirectly to metamorphic processes. A discussion of the third type (plutonic-related occurrences) is contained in an earlier section of this chapter and will not be repeated here. Metamorphic-related occurrences are described below in relative detail because they are potentially very important classes of occurrences within the Cordilleran metamorphic core complexes.

Unconformity-related Uranium Deposits

Unconformity-related deposits are a recently recognized class of deposits that have received much attention, both within exploration and research circles (Kalliokoski and others, 1978; Mathews, 1978c, d; Hegge and Rowntree, 1978; Dahlkamp, 1978; Hoeve and Sibbald, 1978). As the name implies, these deposits are spatially associated with major unconformities, generally of Proterozoic age (Robertson and others, 1978). They are preferentially located in structural features such as faults, breccia zones, and fractures within metamorphic rocks and their unconformably overlying cover. The deposits have limited vertical extent on either side of the unconformity and are characterized by pitchblende, base-metal sulfides and arsenides, chlorite and hematite. They may contain very rich ore (averaging as high as several percent $U_3O_8$ in the Nabarlek deposit of the Alligator Rivers area of Australia; Minobras, 1979) and exceptionally large reserves of uranium. The Koongara and Ranger ore bodies of the Alligator Rivers area of northern Australia, and the Rabbit Lake and Key Lake deposits of northern Saskatchewan are excellent examples of unconformity-related mineralization (Mathews, 1978c, d).

No unconformity-related deposits are known to exist in the Cordilleran metamorphic core complexes and the likelihood of finding such deposits is not great. Major unconformity-related ore bodies of the world are of a restricted age range (mid-Proterozoic, 1.5 to 2.0 b.y.B.P.) and unconformities of such an age in the complexes are rare. The Okanogan is very unfavorable for this type of mineralization because it evolved in an area that
is probably not underlain by Precambrian continental crust. Several other complexes are equally unfavorable since they do not have any older Precambrian rocks exposed within them (Ruby, Kern and Snake complexes). Most complexes are unfavorable because they do not contain any known Precambrian unconformities of the proper age (White Tank, Harquahala, Harcuvar and adjacent complexes of western Arizona and southeastern California) or else are largely composed of plutonic rocks of Mesozoic and Cenozoic age (South Mountain, Coyote, and Pozo Verde complexes). It is possible that the Selkirk, Bitterroot, Albion, Raft River, and Catalina complexes have mid-Proterozoic unconformities, but whether any contain significant unconformity-related mineralization is presently unknown, but doubtful. The closest possible example within the core complex belt is the Precambrian Farmington Canyon complex which is exposed in Utah outside of the core complexes and is discussed briefly by Kalliokoski and others (1978) within the context of unconformity-related deposits. Rocks similar to the Farmington Canyon units might be present in the Raft River-Albion core complexes where older Precambrian rocks are documented. Uranium occurrences in the Green Creek metamorphics of the Albion Range could conceivably be unconformity-related. However, the metamorphic core complexes are as a species not likely to be host to significant unconformity-related occurrences of uranium.

Vein-type Uranium Deposits

Vein-type uranium deposits have been placed with deposits of uncertain genesis by Bendix (Mickle, 1978). This type of deposit occurs in structurally controlled sites adjacent to steeply dipping, major fault systems (Mathews, 1978c, d). Most of the larger veins are found in metamorphic rocks of Proterozoic age and consist of pitchblende either alone or in association with a variety of sulfides, arsenides and other minerals. Mineralization includes Co, Ni, Cu, and Ag in the deposits of the Echo Bay-Eldorado mines area of the Northwest Territories. The deposits are commonly strongly influenced by wall-rock lithology adjacent to the vein with carbonaceous, chloritic, and carbonate units being favorable hosts. Hydrothermal alteration of the wall-rocks produces chlorite, hematite, quartz, and carbonate minerals. The Beaverlodge area of Saskatchewan and the Echo Bay-Eldorado Mines area of the Northwest Territories are given by Mathews (1978c,d) as examples of this deposit type.

Within the Cordilleran metamorphic core complexes there are occurrences of uranium in veins, but it is unclear whether these are similar to those described above. In the Fools Peak area of the Rawhide Mountains, Arizona, high-grade uranium mineralization is present in veins that are associated with hematite and silicificaton. The veins discordantly cut mylonitic fabric of Tertiary(?)
age and are therefore Tertiary. Uranium mineralization is present in small veins in the Harcuvar Mountains to the south where it is accompanied by copper and gold. The mineralization is probably magmatic-hydrothermal due to its spatial and possible genetic relation with microdiorite dikes that intrude amphibolite-grade metamorphic rocks and nearby foliated granites.

Metamorphic-related Uranium Occurrences

The recent classification of uranium deposits proposed by Bendix (Mickle, 1978) has major classes of sedimentary, plutonic, and volcanic deposits, but does not contain a class that is produced by conventional metamorphism. As discussed earlier in this section, metamorphism is a process that is capable of concentrating and redistributing uranium well before the anatectic stage. This is especially emphasized by geologists from parts of the world outside of the U.S. Those uranium occurrences that owe their origin to metamorphism would reasonably be assigned to a metamorphic-related class. There are a sufficient number of uranium deposits throughout the world whose origins can be reasonably ascribed to metamorphism, that a metamorphic-related class is justified. The composite characteristics and selected examples of metamorphic-related deposits are given below. This class of occurrences is potentially very important within the Cordilleran metamorphic core complexes.

Metamorphic uranium occurrences in the present usage, are those occurrences in which metamorphism had a significant influence on the distribution, concentration, or mineralogical constitution of uranium. Metamorphism, in some cases, was directly responsible for the redistribution and concentration of uranium while in other cases, it merely resulted in the isochemical reconstitution of uranium that was already concentrated in the protolith prior to metamorphism. Metamorphic uranium occurrences must, by definition, be disposed in metamorphic rocks and owe their origin to metamorphism, rather than subsequent processes, whether they be plutonic, hydrothermal, or meteoric. Metamorphic-related uranium occurrences are most common in amphibolite-grade schist, gneiss and migmatite, although greenschist-grade occurrences are also reported in the literature. Uranium mineralization exhibits a preference for rocks rich in biotite, and to a lesser extent muscovite, amphibole and garnet. Hematite and chlorite are locally abundant as are pyrite, chalcopyrite, and graphitic or carbonaceous material. Uranium generally occurs either alone (predominantly as crystalline uraninite or pitchblende) or in association with base metals, generally copper. Evidence of redistribution of major and minor elements is common and metasomatism and anatexis have locally taken place.
There are many examples of metamorphic-related uranium occurrences throughout the world and some excellent ones within the western U.S. In the Kulu Himalaya, uraninite mineralization in pelitic schists has been interpreted by Narayan Das and others (1971) as being the result of metamorphic processes. They propose that uranium which was originally fixed by adsorption on argillaceous clays was mobilized and concentrated in the early stages of metamorphism. With further metamorphism, uranium continued to recrystallize and be mobilized and concentrated, resulting in the segregation of uraninite along foliation planes in the schist. Chalcopyrite, galena, pyrite and quartz are associated with the mineralization. To the south in India, uranium mineralization occurs in the Singhbum shear zone (Banjeri, 1962; Law, 1970). Host rocks for the mineralization consist of migmatite, mica schist, augen gneiss, and a variety of other metamorphic rocks. There is a close relationship between structural deformation, migmatization, and ore localization, indicating that mineralization was synkinematic as well as synmetamorphic. Uraninite and copper mineralization have been attributed by Banjeri (1962) as products of migmatization within the shear zone.

In North America, Lang (1952) has described migmatites in Canada that have thin bands of uraninite-bearing pegmatite, biotitic schist, and gneiss. In Colorado, disseminated uranium mineralization is present in migmatized gneiss and mixed gneiss and pegmatites of the Wheeler Basin (Young and Hauff, 1975). Uraninite mineralization is concentrated in parts of the host rock that are rich in biotite. It has been suggested that remobilization and concentration of uranium took place during metamorphism, probably during the intrusion of an adjacent radioactive pluton of Silver Plume Granite (Young and Hauff, 1975).

There are several important deposits that are difficult to confidently either include or exclude from the metamorphic-related class. These include Mont Laurier in Quebec (Wynn-Edwards, 1969; Allen, 1971; Kish, 1975), Chalebois Lake in Saskatchewan (Mawdsley, 1952, 1955; Beck, 1969, 1970), Crocker Well in Australia (Whittle, 1954; Campana and King, 1958), and Rossing in Namibia (Von Backstron, 1970; Armstrong, 1974; Berning and others, 1976; Nishimori and others, 1977). All the above deposits are associated with metamorphic rocks and are characterized by abundant pegmatites. For these deposits, it has been proposed that mobilization of uranium and pegmatitic melts (or fluids) was a result of metamorphism, ultrametamorphism, or anatexis. The deposits could thus be classified as anatectic, pegmatitic, or metamorphic-related. However these deposits are classified, they represent examples of the type of deposits that might be found in Cordilleran metamorphic core complexes.
Within the core complexes, the archetypical metamorphic-related deposit is the so-called Graeber Lease of the northeastern Kettle complex (U and W uranium in Huntting, 1957). In this deposit, high-grade uranium mineralization is contained within a unit of amphibolite-grade, biotitic schist and gneiss. The ore horizon displays a pervasive mylonitic fabric and is locally pyritic. This deposit may represent a metamorphosed protolith that had an originally high uranium content, or else is the result of concentration of uranium by metamorphic and mylonitic processes. This deposit has profound implications for the uranium favorability of the core complexes in general, and the Kettle and Selkirk complexes in particular. The Kettle complex has a myriad of pegmatites which are anomalously radioactive. It is possible that these pegmatites were produced by the partial melting of an originally uraniferous protolith. In any event, it is clear that the Kettle complex is very favorable for metamorphic-related occurrences of uranium. In the remainder of the complexes, there are only minor occurrences of increased radioactivity that may have been formed by metamorphic processes. During our field work and sampling in the complexes, we have consistently observed higher radioactivity and uranium contents in metamorphic rocks that are rich in biotite than in adjacent quartzo-feldspathic gneiss or foliated granite. In addition, amphibolitic rocks are systematically lower in radioactivity and uranium content than other metamorphic rocks. These observations are congruent with those of geologists around the world that biotitic rocks, especially those that are also garnetiferous, are favorable hosts for metamorphic-related uranium.

Summary of Conceptual Basis for the Occurrence of Uranium Related to Metamorphic Processes

Evidence of metamorphic processes abounds in Cordilleran metamorphic core complexes, suggesting that metamorphism may be highly influential in governing the occurrence of uranium in the complexes. Uranium in metamorphic rocks resides in a number of mineralogical and structural sites, with a large proportion being concentrated in structural defects and along grain boundaries. Mica schist and garnetiferous migmatite tend to be more radioactive than amphibolite or typical quartzo-feldspathic gneiss. Uranium content of metamorphic rocks decreases with increasing grade, indicating that rocks are depleted in uranium by progressive metamorphism. Progressive metamorphism occurs largely by a series of dehydration reactions that result in the formation of an interstitial hydrous fluid. Uranium, alkali elements, and certain other chemical constituents will generally be partitioned into the fluid phase. If the evolved fluid is driven from the parent rock, it will result in a net depletion of that rock.
Fluids thus liberated could be highly uraniferous and elsewhere form metasomatic or hydrothermal uranium deposits. If with increased grades of metamorphism partial melting occurs, uranium will be concentrated in the accumulating melt to potentially form a variety of plutonic-related uranium deposits. Of the classes of uranium occurrences found in metamorphic rocks, metamorphic core complexes are, as a group, not highly favorable for unconformity or vein-type deposits. In general, the core complexes are favorable for metamorphic-related uranium deposits, with the Kettle and Albion complexes being most favorable.
A CONCEPTUAL BASIS FOR THE OCCURRENCE OF URANIUM

RELATED TO MYLONITIC PROCESSES

Mylonitic Processes in Cordilleran Metamorphic Core Complexes

Mylonitic rocks are inevitably present within Cordilleran metamorphic core complexes, in some cases comprising a majority of the exposed crystalline core. The mylonitic fabric is manifest primarily in a gently inclined mylonitic foliation that contains a conspicuous, penetrative lineation. Microscopically, mylonitic rocks in the complexes are characterized by brittlely deformed feldspars and recrystallized quartz that occurs in sutured aggregates (for further descriptions see Davis and others, 1980; Reynolds and Rehrig, 1980). These rocks are similar to photographs of hand specimens and photomicrographs of protomylonite, mylonite, mylonite gneiss, and mylonite schist as presented in the classification of Higgins (1971).

Mylonitization generally results in a decreased mean grain size via comminution of brittle minerals and recrystallization of susceptible minerals such as quartz. In several complexes, it has been observed that mylonitization was accompanied by only minor reequilibration of protolith mineral assemblages (Reynolds and Rehrig, 1980; Davis and others, 1980). In these complexes, mylonitization occurred under such conditions that few new mineral phases were formed. As a consequence, the bulk of the minerals in the mylonitic rocks are relicts from their previous plutonic or metamorphic history. Where present, synkinematic, equilibrium mineral assemblages indicate a range of pressure-temperature regimes from greenschist to amphibolite grade (Davis and others, 1980; Anderson and others, 1979).

Mylonitic rocks in the core complexes are generally accompanied by and interlayered with abundant quartz veins. The veins ordinarily consist of a simple quartz mineralogy, although locally they also contain minor feldspar, sericite, iron oxides or sulfides. In many cases, the quartz veins exhibit a mylonitic fabric similar in orientation and physical characteristics to that in adjacent mylonites. In at least two areas, it has been reported that mylonitic rocks contain numerous quartz veins while the nonmylonitic protoliths contain few, if any, such veins (Chase, 1973; Reynolds and Rehrig, 1980). Clearly, the quartz veins are a result of mylonitization, an inference supported by microscopic evidence of extensive recrystallization and remobilization of quartz.
Uranium and Mylonitic Processes

The ability of mylonitization to recrystallize, remobilize and redistribute quartz suggests that it is also capable of remobilizing major and minor elements, including uranium.

Protolith mineralogy may be extremely important in governing the mobility of uranium during mylonitization. A variable proportion of uranium in plutonic and metamorphic rocks resides in refractory accessory minerals. Some of these minerals exhibit relatively inert chemical behavior during mylonitization and would be expected to neither lose nor gain uranium. On the other hand, minerals like epidote that are more chemically reactive under mylonitic conditions (Higgins, 1971) could either lose or gain uranium, and thereby respectively enrich or deplete the uranium content of any adjacent interstitial fluid. Significant amounts of uranium could accumulate in the interstitial fluid during mylonitization if original uranium-bearing minerals were converted to minerals that are less receptive to uranium.

In many cases, a majority of the uranium in a rock is associated with grain boundaries, structural defects and intergranular fluids. This uranium would be readily scavenged by fluids liberated during mylonitization. If the rock retains these fluids, it will undergo little net change in uranium content. However, if the rock loses these fluids in large enough quantities, it will experience a significant net loss in uranium. Such liberated fluids might conceivably form metasomatic or hydrothermal uranium deposits in peripheral receptive lithologies.

The observed geochemical changes that accompany rock deformation have been discussed for several areas outside of the core complexes (Fyfe, 1976; Kerrich and others, 1977; Beach, 1976; Hanson and others, 1969). Hanson and others (1969) examined the chemical changes undergone by a shallow granite porphyry that was overprinted by a post-emplacement, low-grade, cataclastic (mylonitic) event. Their data indicates that the deformed granite experienced losses of sodium and strontium, coupled with gains of magnesium and water. Silica and aluminum contents were evidently unchanged by deformation. In contrast, Kerrich and others (1977) discovered isochemical behavior for an adamellite that had been deformed in a shear zone under high temperature (amphibolite-facies) conditions. Chase (1973) reached a similar conclusion for major elements in granitic rocks of the Bitterroot metamorphic core complex of Idaho and Montana. These studies are exemplary of the literature's discord regarding the geochemical changes produced by deformation.
The literature is equally ambivalent regarding the effects of deformation on the uranium content of rocks. Dmitriyev and others (1977) studied the behavior of uranium during mylonitization and more brittle cataclasis and discovered that certain Russian blastomylonites have the same uranium content as the protolith, with uranium being concentrated in accessory minerals and structural defects. Uranium contents evidently increased with the initiation of "brittle" deformation because cataclasites and "mylonites" have an order of magnitude more uranium than do the protoliths and blastomylonites. In the "mylonites" uranium is concentrated in melanocratic bands composed of epidote, chlorite, sphene, and leucoxene that formed by reaction of biotite. Yermolayev (1973) likewise indicates that uranium concentration increases with cataclasis.

In contrast with this, a study of some mylonitic rocks in the Owens Valley region of California found that blastomylonites have significantly greater uranium and thorium contents than do mylonites (Cupp and Mitchell, 1978). In a regional study of the Precambrian basement of western Canada, Burwash (1979) reports that mylonitic rocks have the same uranium abundances as other rock types.

As part of our evaluation of Cordilleran metamorphic core complexes, we have sampled traverses from mylonitic to nonmylonitic phases of the same rock unit. In the South Mountains of central Arizona, rubidium and strontium abundances of undeformed and mylonitic granodiorite are similar. Strontium isotopes of the two phases are also statistically indistinguishable. Scintillometer readings are identical for the two phases, indicating that mylonitization has not resulted in significant redistribution of radioactive elements. All these observations are congruous with microscopic evidence for only subtle mineralogical changes besides imposition of a mylonitic texture.

Sample traverses across mylonitic "fronts" in the Tortolita Mountains are discussed in detail in Chapter 5. They indicate that mylonitization of both granitic and metasedimentary protoliths has caused net gains of silica and sodium, with losses of other elements such as potassium, magnesium, and iron. Abundances of uranium and thorium exhibit no change between mylonitic and nonmylonitic rocks. These results are similar to those obtained for the Santa Catalina Mountains which document that uranium contents and total gamma measurements show no variation between mylonitic and nonmylonitic parts of a single muscovite granite pluton. However, in all the above cases, we were only able to sample rocks that have relatively low uranium contents. It is possible or even probable that these samples do not exhibit
behavior analogous to more uraniferous rocks. Only sampling in the Kettle, Selkirk, or other complexes which contain mylonitically deformed, uraniferous rocks will resolve this uncertainty.

Uranium Occurrences Related to Mylonitic Processes

Throughout the world, many uranium deposits are associated with mylonitic rocks. For instance, Von Backstrom (1974) discusses uranium occurrences which he classifies as "shear zone and breccia deposits". In one such deposit in India, uranium and copper mineralization occur within the Singhbum shear zone (Banjeri, 1962; Law, 1970). The zone of crushing and mylonitization cuts migmatite, mica schist, and various other metamorphic rocks. Uranium mineralization in the form of uraninite, autunite, and torbernite shows a spatial preference for rocks rich in apatite, magnetite, tourmaline, chlorite, and biotite. Banjeri (1962) indicates that a close relationship exists between structural deformation (including shearing), migmatization, and ore localization. If so, this may be an example of a deposit formed by fluids that were generated during shearing. Nishimori and others (1977) likewise indicate that uranium mineralization in the Beaverlodge area of Canada is partly the result of remobilization of uranium during mylonitization. At Les Herbiers Mine in western France, uranium mineralization (uraninite and coffinite) is present in a mylonitic zone within mica schist and Hercynian two-mica granite (Minobras, 1979). Hercynian granites of the region are characterized by anomalous contents of uranium and could easily be the source of the uranium, especially since uranium in the granites is readily leachable (Cuney, 1978). A similar situation is present in the southern Appalachian Mountains, where uranium mineralization is localized in shear zones that cut granitic and metamorphic rocks which may be the source of the uranium (Penley and others, 1978). Among other excellent examples of this type of deposit are those found in shear zones near Owens Valley, California (Cupp and Mitchell, 1978) and in the Badger Flats-Elkhorn thrust area of Colorado (Gallagher, 1976).

Within the Cordilleran metamorphic core complexes, the best examples of uranium in mylonitic rocks occur in the Kettle complex. Uranium is associated with uraniferous pegmatites and adjacent biotitic schist and gneiss. Mylonitic fabric is common in both the pegmatites and metamorphic rocks. At the Graeber Lease in the northeast margin of the complex, (listed as the U and W mine in Huntting, 1956) high-grade uranium mineralization is distributed along a mylonitic zone within pyritic, biotite schist and gneiss. This deposit might serve as the most lucid core-complex example of uranium mineralization within mylonitic host rocks. The Kettle complex is evidently the most favorable of
the Cordillera by virtue of its high-background pegmatites and locally uranium-rich metamorphic rocks. There are no other localities within the core complex belt that are as important or as favorable as those in the Kettle complex.

Summary of Conceptual Basis for the Occurrence of Uranium Related to Mylonitic Processes

Mylonitic rocks are widely exposed in Cordilleran metamorphic core complexes and have been derived from a wide variety of plutonic and metamorphic rocks. Field and microscopic observations indicate that mylonitization is accompanied by recrystallization and remobilization of quartz and other minerals. In some cases, mylonitization has resulted in the redistribution of both major and minor elements. Uranium, because of its known high mobility, is probably also redistributed by mylonitization under favorable circumstances. Mylonitic rocks derived from plutonic and metamorphic rocks are hosts to uranium mineralization throughout dispersed regions of the world. In a significant number of cases, the country rocks cut by the mylonitic zones are characterized by high radiometric signatures and anomalously high background contents of uranium. It is possible that uranium mineralization in some of these areas was deposited by uraniferous fluids which were liberated by mylonitization. In any event, the most favorable mylonitic zones should be those that transgress country rocks of anomalously high uranium contents. The Kettle complex is the most favorable of the core complex belt due to its high-background pegmatites and known uranium occurrences within mylonitic rocks.
A CONCEPTUAL BASIS FOR THE OCCURRENCE OF URANIUM RELATED TO DISLOCATION SURFACES

Dislocation Surfaces in Cordilleran Metamorphic Core Complexes

The dislocation surface (décollement) is without a doubt one of the most intriguing features of Cordilleran metamorphic core complexes. It is a remarkable planar or curvi-planar structural discontinuity that dominates the landscape over large regions. It is perhaps most dramatic in the Whipple Mountains of southeastern California where it is a low-angle surface that separates a basement of light-colored mylonitic gneiss from overlying volcanic and sedimentary rocks of middle Tertiary age (Davis and others, 1979, 1980). The dislocation surface is so conspicuous that it is clearly visible on high-altitude photographs and satellite imagery. In this area, as elsewhere in the core complex belt, the surface is planar on the scale of an outcrop, but arched or warped when viewed within the perspective of an entire mountain range. In most complexes, the actual surface is polished, slickensided and heavily stained with iron oxides. It is underlain by a resistant ledge of microbreccia which has a variable thickness that averages approximately one meter. Newly broken surfaces of the microbreccia have a distinct gray, resinous or flinty appearance. Both on outcrop and microscopic scales of observation, the microbreccia generally reveals no planar fabric. Instead, the rock is a jumble of microscopic angular fragments derived from lithologies characteristic of the footwall. Below the microbreccia ledge is a shattered rock most aptly described as chloritic breccia. As the name implies, this rock is characterized by a green color due to ubiquitous chloritic alteration, and is dissected by a myriad of anastomosing joints, fractures, and microfaults. Some fracture surfaces exhibit numerous slickensides, while others reveal none. Hematite epidote, and limonite are common associates of the breccia, and in western Arizona and adjacent California, chrysocolla, chalcopyrite, specular hematite, and pyrite are present in such abundance that mines produced copper, iron and locally gold ore from the zone. Down structural section, the brecciation and chloritic alteration gradually die out into structurally intact rocks which commonly possess a mylonitic fabric.

Above the dislocation surface rest a variety of sedimentary, volcanic, plutonic and metamorphic rocks, with the latter two types being less common than the former two. These upper plate rocks generally lack the mylonitic fabric that is so characteristic of lower plate rocks, and in many localities, the dislocation surface separates amphibolite-grade metamorphic rocks of the footwall.
from nonmylonitic, completely unmetamorphosed middle Tertiary rocks of the hanging wall. Upper plate rocks are typically cut by listric normal faults which either merge with or terminate against the basal dislocation surface. A variant of this overall theme occurs in the Snake Range of Nevada where a marble tectonite intervenes between locally shattered footwall rocks and overlying Paleozoic carbonate rocks which are unmetamorphosed but highly brecciated and faulted (Misch, 1960; Coney, 1974). Additional descriptions and photographs of representative lithologies of the dislocation zone are contained in Reynolds and Rehrig (1980), Davis and others (1980), and Davis (1980).

Uranium and Dislocation Surfaces

There is essentially no literature available that discusses the relationship between uranium and dislocation surfaces of Cordilleran metamorphic core complexes. Instead, inferences regarding uranium behavior in the dislocation zones must be formulated strictly from our currently limited geological understanding of the zones. Articles which discuss the relationship between uranium and retrograde metamorphism, brecciation, hydrothermal alteration, and other topics relevant to the core-complex dislocation surfaces are included in Appendix B.

The wide variety of processes associated with formation of dislocation surfaces in the complexes insures that the distribution of uranium will be at least partially modified during this formation. Of fundamental importance in this regard is the ubiquitous alteration or retrogressive metamorphism that is responsible for the whole-scale conversion of biotite to chlorite in brecciated rocks immediately below the dislocation surface. This chloritic breccia also generally contains abundant epidote, hematite, and other alteration products, which indicates that hydrothermal fluids pervaded the permeable breccia zone during or after its formation. The affinity of available uranium for these fluids suggests that it will be redistributed by their passage. Uranium that occurs in grain-boundary deposits and structurally controlled sites will be most easily scavenged by these fluids. The breakdown or alteration of uranium-bearing minerals will also liberate uranium into the interstitial hydrous phases. This might be especially important for minerals such as biotite that frequently contain microinclusions of primary uranium minerals (for example, uraninite). Alternatively, newly forming minerals such as epidote might expropriate uranium from the adjacent fluid.

Hydrothermal fluids within the dislocation zone were evidently rich in other chemical constituents, as indicated by the
local abundance of iron and copper minerals in the zone. The predominance of chrysocolla over chalcopyrite and hematite over pyrite either reflects a well oxygenated environment during deposition of the hydrothermal deposits, or else a very thorough secondary (supergene) modification of original sulfide ores. In any event, the hydrothermal fluids were clearly mineralizers that were also probably capable of redistributing uranium. The presence of similar mineralization within upper plate rocks indicates that such fluids were not restricted to the chloritic breccia, but were able to penetrate above the dislocation surface.

Geochronologic studies in several of the complexes suggest a possible origin for the hydrothermal fluids. Potassium-argon ages of lower plate mylonitic rocks are locally the same age as movement on the dislocation surface. This demonstrates that lower plate rocks were still hot (over 200° C) during dislocation and therefore could have been heat sources that triggered convection of hydrothermal fluids. Viewed in this perspective, hydrothermal alteration and mineralization are concentrated along the dislocation zone because it was a permeable interface between hot lower-plate rocks and cooler, potentially water-saturated, upper-plate rocks. Verification of this intuitive reasoning would require an extensive program of geological, geochemical, geochronological, and petrological research that is beyond the scope of this project. Chemical analyses for a limited number of samples of chloritic breccia and the dislocation zone are discussed in Chapter 5 and listed in Appendix E. Results of more detailed scintillometer traverses through the dislocation zone are included in Appendix D for individual core complexes.

The chloritic breccia and dislocation zone may be less important as actual ore horizons than as permeable channels for uraniferous ground waters. Meteoric waters could leach uranium from uraniferous core lithologies and subsequently be funneled down the hydraulic gradient of the shattered and permeable chloritic breccia zone. Uranium would be deposited by these oxygenated waters upon encountering reducing conditions at the water table, in pockets of sulfides in the zone itself, or most likely in favorable upper-plate lithologies that are in contact with the zone. Phenomena such as weathering and ground-water redistribution of uranium are discussed further in the section of this chapter that is concerned with processes extrinsic to the core complexes.

Movement along the dislocation surface was ordinarily accompanied by tilting of upper-plate rocks. This tilting has modified the geometry and orientation of previously formed uranium deposits, resulting in their eventual erosional
unroofing and subsequent discovery. For example, the Anderson Mine in the Date Creek Basin of western Arizona (Otten, 1977) is located in uraniferous lacustrine units that are probably underlain at depth by a dislocation surface and that owe much of their present southwest dip to syn-dislocation tilting. Nearby uraniferous lacustrine units in the Artillery Mountains are likewise situated in an area whose structural complexity and present geometry were highly influenced by tectonism that accompanied dislocation.

There is evidence that dislocation, listric normal faulting and tilting (rotation) of upper-plate rocks resulted in the formation of basins, within which were deposited locally thick sequences of syn-tectonic clastic rocks. If contemporaneous volcanism provided uranium-rich ash to the basins, or if the source of the clastic debris was uraniferous, the sedimentary rocks of the basins might host either sandstone or lacustrine uranium deposits. For instance, the timing of deposition of the Anderson Mine sequence is suspiciously close to that documented for dislocation, listric normal faulting, tilting, and potassic volcanism in the immediate vicinity. The present Date Creek Basin and Anderson Mine sequence may occupy an area which was a basin during dislocation. The locally uraniferous, Eocene Tiger Formation of the Pend Oreille River region of Washington may also have accumulated in a basin formed as a consequence of rotation of upper-plate rocks during dislocation along the Newport fault (Miller, 1972).

Uranium Occurrences Related to Dislocation Surfaces

There is a worldwide association of uranium mineralization with zones of structural disruption. For example, vein-type uranium deposits like those in the Beaverlodge district of Saskatchewan are localized in mylonitic, brecciated and fractured metamorphic and granitic rocks. Chlorite and hematite are generally present in both vein-type and unconformity-related deposits. In various kinds of uranium occurrences, copper and iron sulfides are abundant. Clearly, uranium mineralization is commonly associated with geological attributes that are exhibited by dislocation surfaces of Cordilleran metamorphic core complexes. Worldwide examples of uranium occurrences in dislocation surfaces cannot be confidently identified, because there is the possibility that dislocation surfaces (as the term is used by Reynolds and Rehrig, 1980; Rehrig and Reynolds, 1980) are features unique to the core complexes. However, there is a sufficient number of examples of occurrences related to dislocation surfaces within the core complexes, that this type of deposit may be characterized.
The Buckskin, Rawhide, and Artillery Mountains of west-central Arizona contain several excellent examples of uranium mineralization localized within the dislocation zone. At the Red Hills prospect of the Rawhide Mountains, locally high-grade uranium mineralization is present in brecciated gneiss and limestone at the dislocation zone. Uranium minerals are accompanied by chrysocolla, quartz, barite and fluorite. In the adjacent Fools Peak area uranium occurs in altered and brecciated crystalline rocks near the dislocation surface. Limonite and hematite are abundant, as they are in many exposures of the dislocation zone throughout west-central Arizona. In this entire region, including nearby parts of south-eastern California, the dislocation zone is commonly mineralized with hematite, chrysocolla, copper and iron sulfides, barite, calcite and manganese minerals.

In the Santa Catalina Mountains of southern Arizona, significant uranium mineralization occurs at the Blue Rock Mine on the eastern flank of the range. Uranophane and autunite are present in association with fluorite, hematite, limonite, and copper minerals. Several occurrences in this area occupy low-angle fault zones that separate an imbricate stack of rocks ranging in age from Precambrian to Tertiary. Mylonitic rocks are present below the lowest fault, but it is not clear whether all the features characteristic of dislocation surfaces are represented or whether this important occurrence is indeed an example of uranium mineralization related to a dislocation surface.

Other examples of uranium occurrences related to dislocation surfaces are present in Washington and northern Idaho. Weis and others (1958), report mineralization in a brecciated pegmatite in northern Idaho that is fairly near the Newport fault (dislocation surface). In addition, some exposures of chloritic breccia in this area and along the margins of the Kettle complex are of anomalous radiometric signature. These occurrences demonstrate the potential for significant uranium mineralization in the dislocation zone.
Summary of Conceptual Basis for the Occurrence of Uranium Related to Dislocation Surfaces

Dislocation surfaces are profound structural discontinuities in Cordilleran metamorphic core complexes that separate a core of plutonic, metamorphic, and mylonitic rocks from an upper plate of various nonmylonitic lithologies. The actual dislocation surface is regionally curvi-planar, highly polished, and heavily stained with iron oxides. Rocks immediately below the dislocation surface are chloritic, shattered and dissected by numerous anastomosing joints. Upper plate rocks above the dislocation surface are cut by listric normal faults and are commonly steeply tilted and locally mineralized. The dislocation zone could contribute uranium to rising hydrothermal fluids or could act as a permeable channelway for meteoric waters descending the hydraulic gradient away from the crystalline core of the complexes such as those in northeastern Washington. Supergene leaching of uraniferous core rocks could contribute uranium to these fluids transforming them into potential mineralizers. Therefore, as suggested by Cheney (1980), the margins of the core complexes may be fertile ground for a relatively new class of uranium occurrences: those related to dislocation surfaces. The Kettle, Selkirk, Albion, Rawhide, and Santa Catalina complexes are the most favorable because they contain possible occurrences of this type and because at least the first three are characterized by numerous uranium occurrences of various other types.
A CONCEPTUAL BASIS FOR THE OCCURRENCE OF URANIUM RELATED TO PROCESSES EXTRINSIC TO CORDILLERAN METAMORPHIC CORE COMPLEXES

Processes Extrinsic to Cordilleran Metamorphic Core Complexes

The processes summarized in the foregoing sections are integral components in the tectonic evolution of Cordilleran metamorphic core complexes and can be thought of as intrinsic or essential to the complexes. In contrast, there is a spectrum of processes that are unrelated to actual tectonism in the complexes. These processes are extrinsic or incidental to the core complexes, but are nevertheless perfectly capable of redistributing uranium into various types of deposits. Prior to formation of the complexes, rocks which ultimately became amalgamated into the crystalline cores experienced geological histories that entailed numerous events. Uranium concentrations formed during these earlier histories would have been drastically modified by the superimposed plutonism, metamorphism, metasomatism, and mylonitization that accompanied activity in the core complexes. Conversely, rocks which eventually became incorporated into upper-plate positions may have been influenced only by tilting that accompanied listric-normal faulting and dislocation. Original uranium distributions in these rocks may have been preserved largely intact, but reoriented by the tilting. Therefore, upper-plate rocks could conceivably contain pristine occurrences of any class of uranium deposit identified by Bendix (Mickle, 1978).

Subsequent to their tectonic evolution, the core complexes were subjected to numerous other processes such as weathering, erosion, and ground-water circulation. These near-surface processes indiscriminately affect rocks of the core, dislocation zone and upper plate. Several complexes have highly uraniferous rocks in their core that would be excellent sources of uranium for surficial redistribution. In the Basin and Range province, block faulting has created potential traps for this uranium in graben that typically flank at least one side of the core complexes. Uranium occurrences in the basins could be an important consequence of the interaction between core complexes and processes that are extrinsic to their evolution.
Uranium Occurrences Related to Processes Extrinsic to Cordilleran Metamorphic Core Complexes

Pre- and Syn-tectonic Occurrences

Cordilleran metamorphic core complexes evolved in a wide variety of regional geologic settings. Complexes in Nevada and Utah were formed within a region characterized by thick sequences of carbonate and clastic rocks that were deposited in a late Precambrian to Paleozoic continental shelf. In contrast, complexes in Arizona are situated in areas that were cratonic during the entire Paleozoic. Further divergence is exhibited by the Okanogan complex which originated among late Paleozoic-early Mesozoic "oceanic" terranes. These differences in regional geologic setting insure that there will be considerable lithologic variation in rocks affected by evolution of the core complexes. Accordingly, there are many possible types of uranium occurrences that could have existed in these lithologies prior to formation of the complexes. A brief discussion of some possible types of pre- or syn-tectonic occurrences is included below, but much additional information on each type is contained in Mickle and Mathews (1978).

Incorporated into the cores of the complexes are an assortment of igneous and metamorphic rocks. Metasedimentary rocks in the complexes have typically been derived from protoliths that are Precambrian to lower Paleozoic in age. The existence of Precambrian rocks in the cores of many complexes indicates the potential for types of uranium deposits that are age-specific to the Precambrian. Quartz-pebble conglomerate uranium deposits are restricted to lower Proterozoic rocks that contain pyrite and uraninite of probable detrital origin. Cordilleran metamorphic core complexes have a low over-all favorability for this class of deposit because they are generally deficient in lower Proterozoic quartz-pebble conglomerates. Only the Bitterroot, Raft River, and Albion Mountains have even the slightest potential for deposits of this type. The same probably applies to unconformity-related and vein-type deposits which are also evidently restricted to Precambrian rocks.

Metamorphosed Paleozoic sedimentary rocks locally accompany Precambrian rocks in the core of the complexes. More commonly, however, Paleozoic strata are unmetamorphosed and situated in upper plate positions. In either case, they could have originally contained marine black shale or phosphorite class uranium occurrences, both of which are typically Paleozoic in age throughout
the world. The Kettle, Pioneer and Ruby Mountains would be most favorable for these two classes of deposits by virtue of their proximity to the inferred Paleozoic continental shelf-slope break, which is the characteristic depositional setting for both types.

Mesozoic and Cenozoic sedimentary and volcanic rocks are only rarely incorporated into the crystalline cores of the complexes, but instead are generally situated in fault slices above the dislocation surface. These upper-plate rocks could contain a variety of uranium occurrences such as those found in sandstone, lacustrine, and volcanogenic lithologies. The Kettle, Selkirk, Albion, western Arizona, and Catalina complexes each contain occurrences of at least one of these types and are therefore somewhat favorable for additional deposits. In areas where plutonic and metamorphic rocks are present in the upper plate, any class of plutonic or metamorphic deposit could exist. Uraniferous granites that occur above the Newport fault (dislocation surface) of the Selkirk complex are illustrative of this potential.

Post-tectonic Occurrences

Subsequent to their main tectonic activity, metamorphic core complexes continued to be affected by numerous processes. After the crystalline core of the complexes were exposed by final arching, they were subjected to intense physical and chemical weathering. Much of the uranium in plutonic and amphibolite-grade metamorphic rocks is easily leached by surficial waters and could be redistributed by these ground waters internally or externally to the complexes. Authigenic uranium occurrences would be formed if uranium is leached and redeposited within the same rock unit (Mathews, 1978a, b). Authigenic occurrences are characterized by secondary uranium minerals such as autunite that fill and coat structures in the mineralized host. Mineralization, which may be very high-grade, is usually not accompanied by significant alteration or gangue minerals. The Daybreak Mine and other occurrences of the Mount Spokane area of the Selkirk metamorphic core complex are type-examples of this class. In this area, meta-autunite coats fractures and cleavages in pegmatitic and alaskitic muscovite granite. Numerous occurrences similar to these are present in other metamorphic core complexes. The Selkirk complex of northeastern Washington and northern Idaho has an exceptionally high proportion of this type and it is extremely favorable for additional deposits. Other complexes are less favorable, but they also contain fair potential for authigenic uranium occurrences. This may be one of the most important classes of uranium occurrences.
to consider when evaluating the favorability of Cordilleran metamorphic core complexes.

If uranium is leached from rocks of the crystalline core and reprecipitated away from the source rocks, allogenic uranium occurrences are formed (Mathews, 1978a, b). In these occurrences, uranium (in the form of pitchblende or secondary minerals) is either disseminated or structurally controlled. Mathews indicates that low-to medium-grade metamorphic rocks are typical hosts to this mineralization.

There are a few occurrences in metamorphic core complexes of northeastern Washington that are probably assignable to the allogenic class. Mineralization along the east flank of the Kettle complex is locally contained in chloritic, low-grade metasedimentary rocks. These rocks are topographically lower than the abundant uraniferous pegmatites of the crystalline core of the Kettle dome, and it is conceivable that the mineralization in the metasedimentary rocks is the result of groundwater redistribution of uranium from the pegmatites. Similar occurrences may be present in the Albion range where low- to medium-grade Paleozoic carbonate and clastic rocks lie on the flanks of domes that expose cores of uranium-mineralized, granitic and high-grade metamorphic rocks.

However, the most important allogenic deposits of the Pacific northwest are located outside of, rather than within the core complexes. These, of course, are the Midnite Mine and nearby occurrences of northeastern Washington. Deposits such as the Midnite Mine lie south of the Kettle complex, but in rocks similar to those exposed between the Kettle and Selkirk complexes. The margins of either one of these two complexes are favorable for these types of deposits, especially in light of the evidence reported by Castor and others (1977) for the existence of a northeast-trending belt of anomalously radioactive granites. The general geologic relationships exposed at the Midnite Mine are mimicked in numerous areas in northeastern Washington, suggesting that the whole region, including the core complexes, is favorable for additional allogenic occurrences. Other metamorphic core complexes are also favorable for allogenic occurrences, but to a lesser degree than northeastern Washington.

In addition to subsurface ground-water processes, mechanical weathering and transportation might further aid the dispersal or concentration of uranium. Placer deposits of radioactive heavy minerals are the result of such concentration. Placer occurrences are present near the Albion Range and are characteristic of a large portion of central Idaho (Savage, 1970). Placer deposits could constitute a relatively minor class of occurrences associated with the complexes.
Post-dating tectonic activity in the complexes is Basin and Range block faulting that formed basins peripheral to the complexes that are capable of trapping uranium. Evaporite deposits and modern day playas are located adjacent to some complexes, suggesting the possibility of calcrete-type uranium. Uraniferous ground waters shed from the complexes might form sandstone-type deposits in the permeable basin fill. Only drilling in the basins adjacent to uraniferous complexes can evaluate the potential for such deposits.
CHAPTER 5

GEOCHEMISTRY OF CORDILLERAN METAMORPHIC CORE COMPLEXES

By

Stanley B. Keith and Stephen J. Reynolds

INTRODUCTION

The crystalline cores or basement terranes of the regions designated as metamorphic core complexes in this report are composed of a wide variety of crystalline rocks. On many geologic maps, these crystalline rocks are depicted as a single map unit (generally "undifferentiated gneiss") and are commonly assigned a Precambrian age (for example Wilson and others, 1969; Dickey and others, 1980). More detailed work reveals that these crystalline terranes are far from homogeneous; rather, the terranes consist of a heterogeneous assemblage of plutonic, metamorphic and mylonitic rocks that are largely Mesozoic through mid-Cenozoic in age. Processes considered characteristic of metamorphic core complexes (see Chapters 1, 2 and 4) are entirely post-Paleozoic. These processes include plutonism, metamorphism, and mylonitization which together have produced a lithologically variable crystalline core or basement terrane. The crystalline rocks formed by these processes have been locally overprinted by penetrative fracturing and chloritic alteration that accompanied movement on mid- to late Cenozoic, low-angle normal faults (dislocation surfaces). We consider plutonism, metamorphism, mylonitization and subsequent dislocation to be processes intrinsic to core complexes; their presence is an essential prerequisite to use of the term 'metamorphic core complex.' Consequently, the following discussion of geochemistry of Cordilleran metamorphic core complexes focuses on these four intrinsic processes. The discussion emphasizes the geochemistry of plutonic rocks because we consider plutonism to be the most important influence on uranium geochemistry of the core complexes.
It is important to point out that the above intrinsic processes display a diversity of temporal and spatial distributions that are by no means restricted to areas designated as metamorphic core complexes. Accordingly, conclusions discussed in this chapter regarding uranium geochemistry are equally applicable to many areas in the North American Cordillera not included in core complexes. For example, statements about geochemical aspects of 'core complex' plutons and mylonitic phenomena have many direct applications to other plutonic and mylonitic terranes throughout the North America Cordillera and the rest of the world. This chapter will not discuss the geochemistry of processes extrinsic to the evolution of metamorphic core complexes because it was beyond the scope of this project.

Many of the conclusions expressed in this chapter are based on geochemical analyses of samples collected in the context of this project. A description of sampling and analytical procedures is included in Appendix E along with lists of sample locations, lithologic descriptions, and geochemical analyses.

GEOCHEMISTRY OF PLUTONIC ROCKS

Plutonic Suites

Embedded in the crystalline core of virtually every designated metamorphic core complex area is an abundance of Mesozoic and/or Cenozoic plutonic rocks. Four major suites of plutonic rocks can be delineated within or adjacent to the core complexes: 1) Jurassic alkali-calcic, biotite-bearing granitoids; 2) middle to Late Cretaceous calc-alkalic, biotite-and hornblende-bearing granitoids; 3) Late Cretaceous to middle Eocene peraluminous, muscovite-or two-mica-bearing granitoids; and 4) Eocene through Oligocene alkali-calcic, biotite-and/or hornblende-bearing granitoids. Table 5-1 summarizes the salient mineralogic and geochemical characteristics of each suite and lists the name and location of major plutons sampled during this project.
Tectonic Setting of Plutonic Suites

The Jurassic alkali-calcic granitoid rocks in core complexes are components of a magmatic arc which extended more or less continuously from north-central Washington southward through western Idaho, central Nevada, and California. From southeastern California, the arc trended southeasterly into southern Arizona. This magmatic arc is presumably the result of an east-dipping subduction zone. Jurassic plutons are known to be present in or adjacent to core complexes in northeastern Washington, Nevada (Ruby Mountains and Snake Range), and the Papago Indian Reservation of southern Arizona (for example the Comobabi and Coyote Mountains). Other core complexes most likely evolved in areas that were located east of the Jurassic arc; these complexes therefore contain no Jurassic alkali-calcic granitoids.

After 150 m.y.B.P. magmatic activity abruptly shifted westward and began forming the great coastal batholiths by approximately 135 m.y.B.P. Core complexes south of the Snake River Plain lie well to the east of the coastal batholiths of California and western Nevada. In contrast, core complexes north of the Snake River Plain lie within regions of widespread middle to Late Cretaceous plutonism.

Between 135 and 105 m.y.B.P. plutons of calcic gabbro and tonalite were emplaced along the western margin of the coastal batholiths. After 105 m.y.B.P. magmatism progressively shifted eastward and became less calcic and more alkalic (i.e. calc-alkalic). The gradual eastward migration of magmatism across Nevada and California greatly accelerated after 80 m.y.B.P. (Coney and Reynolds, 1977; Keith, 1978). Magmatism swept eastward across the southwestern United States between 80 and 45 m.y.B.P. The magmatic sweep evidently occurred earlier (approximately 100 to 80 m.y.B.P.) north of the Snake River Plain. During the magmatic sweep (both north and south of the Snake River Plain), plutonism in any given area was initially alkalic but became progressively more calcic with time. Plutons emplaced during the accelerated phase of the magmatic sweep generally belong to the second suite of plutons: the middle to late Cretaceous calc-alkalic, biotite- and hornblende-bearing granitoids. Representatives of this suite include the middle Cretaceous (90 to 100 m.y.B.P.) granodiorites of northeastern Washington, much of the Idaho Batholith (both the Atlanta and Bitterroot lobes), and Late Cretaceous calc-alkalic plutons which are widely distributed in and around core complexes of southern Arizona and southeastern California. Only the Nevada complexes seem to lack this plutonic suite.
<table>
<thead>
<tr>
<th>Suite</th>
<th>$\text{SiO}_2$ % range</th>
<th>$K_{57.5}^1$</th>
<th>($\frac{87}{86}$ Sr/ Sr $i$)</th>
<th>'AL' $^2$</th>
<th>Alkali Character</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oligocene alkali-calcic biotite hornblende granitoids</td>
<td>58-78%</td>
<td>3.5-4.0?</td>
<td>.706-.709</td>
<td>.8-.1.1</td>
<td>Alkali-calcic</td>
</tr>
<tr>
<td>Mid Cretaceous to mid Eocene peraluminous 2-mica granitoids</td>
<td>63-78%</td>
<td>Not</td>
<td>.711-.734</td>
<td>.98-.1.31</td>
<td>Whipples</td>
</tr>
<tr>
<td>68%</td>
<td></td>
<td>obtainable</td>
<td></td>
<td>.98-.1.06</td>
<td>all others</td>
</tr>
<tr>
<td>Mid to late Cretaceous calc-alkaline granitoids</td>
<td>50-77% $\text{SiO}_2$</td>
<td>1.0-2.4</td>
<td>.707-.709</td>
<td>.8-.1.08</td>
<td>Calc-alkaline</td>
</tr>
<tr>
<td>Jurassic alkali-calcic granitoids</td>
<td>68-77%</td>
<td>3.5-4.5?</td>
<td></td>
<td>1.04-.1.25</td>
<td>Alkali-calcic</td>
</tr>
</tbody>
</table>

Notes:
1) $K_{57.5} = \% K_2O @ 57.5\% \text{SiO}_2$ on a $K_2O$-$\text{SiO}_2$ Harker variation diagram.
2) 'AL' = molecular $\frac{AL_2O_3}{Na_2O + K_2O + CaO}$
3) Modified Peacock classification (Keith, 1978).
4) Number in parentheses is % of total population over 6 ppm uranium.
OF 'METAMORPHIC CORE' COMPLEX PLUTONIC ROCKS

<table>
<thead>
<tr>
<th>Mineralogical Summary</th>
<th>U mode (Figure 5)</th>
<th>% of samples &gt; 6 ppm U</th>
<th>Plutons Sampled for This Pluton Study Complex</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hornblende present in lower silica end members below 70% SiO$_2$. Biotite present in all phases but diminishes above 70% SiO$_2$. Plagioclase and apatite are common accessories. Zircon + Allante sparingly present.</td>
<td>2-2.5</td>
<td>20%</td>
<td>Reef Granite Santa Catalina Az Tortolita Tortolita, Az. Catalina Santa Catalina Az Harrison Pass Ruby, Nevada Skinner Canyon Kern, Nevada</td>
</tr>
<tr>
<td>2 micas (biotite and celadonitic muscovite are common). Biotite locally only mica in rocks below 71% SiO$<em>2$. Plagioclase is sodic (&lt; An$</em>{30}$). Garnet and rare monazite are accessories.</td>
<td>1-1.5</td>
<td>non Washington = 21%; Washington = 74%</td>
<td>Wilderness Santa Catalina Az Derrio Canyon Tortolita, Az. Sunset Pass Harquahala, Az. Tungston Pass Kern, Nevada Unnamed Ruby, Nevada</td>
</tr>
<tr>
<td>Hornblende common in rocks below 70% SiO$<em>2$. Biotite is common (up to 15%) in less silicic rocks. Plagioclase relatively calcic (&gt; An$</em>{30}$). Sphene and apatite are common accessories.</td>
<td>1.5-2.0 and 3.5-4.0</td>
<td>(5%)</td>
<td>Leatherwood Santa Catalina Az Chirreon Tortolita, Az. Granite Wash Harcuvar, Az.</td>
</tr>
<tr>
<td>Biotite is only major mafic. Sphene and apatite are common accessories. Zircon and allante sparingly present.</td>
<td>Biotite Monzogranite</td>
<td>Ruby, Nevada Coarse-grained Ruby, Nevada granite</td>
<td></td>
</tr>
</tbody>
</table>
In almost every major segment of the core complex belt, calc-alkalic plutons of the second suite are intruded by a distinctive set of younger peraluminous muscovite-bearing granitoid rocks which typically contain muscovite and garnet. For example, in northeastern Washington, the muscovite-bearing Phillips Lake Granodiorite is inferred to intrude the middle Cretaceous Starvation Flat Quartz Monzonite by Miller and Clark (1975). Similarly, muscovite-bearing quartz monzonite may intrude more mafic, calc-alkalic phases of the Idaho batholith (Hyndman and Williams, 1977). In several western Arizona complexes, Eocene (?) muscovite- and garnet-bearing pegmatite and granite intrude Late Cretaceous granite and granodiorite (calc-alkalic). Probably the best example of the superimposition of the second and third plutonic suites is in the Santa Catalina and Tortolita Mountains of southern Arizona. Keith and others (1980) have documented that Eocene muscovite-bearing granites (Wilderness and Derrio Canyon granites; both peraluminous) have intruded Late Cretaceous plutons of calc-alkalic quartz diorite and granodiorite. These and other relationships clearly indicate that the peraluminous intrusions consistently post-date the Cretaceous calc-alkalic plutons. Synthesis of published and unpublished geochronologic data from California and Arizona suggests that the Late Cretaceous to early Tertiary peraluminous, muscovite-bearing intrusions, like the slightly older calc-alkalic, biotite-hornblende intrusions, are younger to the east. We interpret these data as indicating that peraluminous plutonism followed the calc-alkalic magmatic arc as it swept eastward. We believe that the eastward sweep of both plutonic suites is a consequence of a flattening subduction zone. The earlier age (middle to Late Cretaceous) of peraluminous granitoids north of the Snake River Plain is by inference due to an earlier age of subduction-zone flattening.

During the early Tertiary, subduction zones are postulated to have been gently inclined beneath all of the core complex belt. These gently-inclined subduction zones became steeper during the Tertiary, resulting in a dramatic rapid return-sweep westward of high potassium calc-alkalic to alkali-calcic magmatism (Coney and Reynolds, 1977; Keith, 1978). The return sweep occurred during the Eocene in Washington, Idaho, and Montana, but was mostly Oligocene to Miocene in age south of the Snake River Plain. The fourth plutonic suite (the alkali-calcic, biotite- and/or hornblende-bearing granitoids) was emplaced during this westward retrograde sweep of the magmatic arc. Accordingly, plutons of this fourth suite are Eocene north of the Snake River Plain and generally Oligocene to early Miocene south of it. The fourth suite of plutons represents the final consolidation and prograde metamorphism within the crystalline cores of the complexes. This event was commonly accompanied by final cooling of the crystalline cores to below K-Ar and Rb-Sr retention temperatures for biotite (less than 200°C). This cooling event is well-documented in the Okanogan and Selkirk complexes of northeast Washington.
Geochemical Comparison of the Plutonic Suites

The previous section has delineated four major suites of plutons in the core complexes on the basis of mineralogic character and relative or absolute age determinations. This section will indicate that each suite may be distinguished from the others by geochemical characteristics, some of which are summarized in Table 5-1.

Major Element Geochemistry

The four suites of Cordilleran granitoids may be divided fundamentally into two major series: peraluminous and metaluminous. Three of the plutonic suites have a metaluminous aspect: that is, molecular $Al_2O_3$ is less than the sum of molecular $CaO$, $Na_2O$ and $K_2O$. Metaluminous chemistry characterizes suites with wide silica ranges (45 - 75% $SiO_2$) but is especially prevalent in the lower silica ranges (generally below 65% $SiO_2$). Sphene and hornblende are typical minerals of metaluminous suites. Figure 5-1 graphically illustrates the metaluminous nature of the second (Cretaceous) and fourth (Eocene-Oligocene) biotite-hornblende suites and compares them with the mid-Cretaceous to Eocene peraluminous suite. Rocks of the second (Cretaceous) and fourth (Eocene-Oligocene suites) are entirely metaluminous below approximately 65% $SiO_2$. Both fields are strongly coincident; the Eocene-Oligocene plutons are slightly more silicic and aluminous than their Cretaceous counterparts. Samples of the Jurassic plutonic suite (not plotted) are all silicic (greater 68% $SiO_2$) and peraluminous, but the suite probably has metaluminous roots. Cretaceous-Eocene peraluminous granitoids of the third suite overlap with the peraluminous end-members of the metaluminous suites but extend into the very strongly peraluminous region of the figure (above 'Al' = 1.2 on Figure 5-1). Garnet and muscovite are the characteristic mineralogic expression of this peraluminous nature. The Whipple Mountain granitoids extensively sampled by Anderson and others (1979, 1980, written communication) may be transitional between metaluminous and peraluminous series. However, the uniformly high silica values (no samples below 65% $SiO_2$) and the presence of garnet and primary muscovite strongly suggest an affinity with the peraluminous suite.

Both metaluminous and peraluminous suites may be subdivided on the basis of their alkali (especially potassium) contents. Keith (1978) utilized a modified Peacock classification to examine the space-time distributions of southwestern North American metaluminous granitoids. The same nomenclature is used here and is summarized in Table 5-2. The nomenclature is applied to core complex granitoids in Figures 5-2 and 5-3. In this classification scheme, the Jurassic and Eocene-Oligocene granitoids fall in the alkali-calcic class with the Jurassic suite being slightly
Figure 5-1. Aluminum characteristics of selected Cordilleran granitoids. Symbols are as follows: Middle to late Cretaceous hornblende-biotite-bearing calc-alkalic granitoids (X); Cordilleran muscovite-bearing peraluminous granitoids (△); Whipple Mountains muscovite-bearing granitoids (∗); Eocene-Oligocene biotite and biotite-hornblende-bearing granitoids (○). Data sources are included in Appendix A. Whipple Mountain data was provided by Lawford Anderson (1980, written communication).
TABLE 5-2:
NOMENCLATURE AND SELECTED CHEMICAL AND MINERALOGICAL PARAMETERS FOR METALUMINOUS IGNEOUS ROCK SERIES DISCUSSED IN THIS CHAPTER (AFTER KEITH, 1978)

<table>
<thead>
<tr>
<th>Suite type</th>
<th>Depth to seismic zone in km</th>
<th>K2O slope character</th>
<th>K2O/SiO2 index*</th>
<th>θ index*</th>
<th>S index</th>
<th>K2O - Na2O</th>
<th>Na2O - K2O</th>
<th>Peacock (1953) index*</th>
<th>Iron enrichment (on AMF diagram)</th>
<th>Al2O3 content</th>
<th>Selected normative mineralogy</th>
<th>Selected modal minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calcic</td>
<td>80-120</td>
<td>0.0-3% K2O per 5% SiO2</td>
<td>0.4-1.2</td>
<td>43-58</td>
<td>0.65-1.5</td>
<td>0.25</td>
<td>2.5-4.5</td>
<td>62-68</td>
<td>strong</td>
<td>high (14-18%)</td>
<td>Moderately quartz normative</td>
<td>Hornblende and biotite</td>
</tr>
<tr>
<td>Calc-alkaline</td>
<td>120-220</td>
<td>0.3-8% K2O per 5% SiO2</td>
<td>1.2-2.4</td>
<td>39-49</td>
<td>1.3-3.0</td>
<td>0.25-0.7</td>
<td>3.5-6.5</td>
<td>58-62</td>
<td>little or none</td>
<td>high (15-19%)</td>
<td>Moderately quartz normative</td>
<td>Biotite, hornblende, and clinopyroxene</td>
</tr>
<tr>
<td>High-K calc-alkaline</td>
<td>220-260</td>
<td>0.5-8% K2O per 5% SiO2</td>
<td>2.4-3.0</td>
<td>38-44</td>
<td>2.4-3.2</td>
<td>0.6-0.9</td>
<td>4.5-7.5</td>
<td>57-59</td>
<td>little or none</td>
<td>high (15-19%)</td>
<td>Weakly quartz normative</td>
<td>Clinopyroxene and orthopyroxene; biotite</td>
</tr>
<tr>
<td>Alkaline calcite</td>
<td>260-300</td>
<td>&gt;0.8% K2O per 5% SiO2</td>
<td>3.0-4.4</td>
<td>21-43</td>
<td>1.8-6.6</td>
<td>generally</td>
<td>generally</td>
<td>4.5-9.0</td>
<td>52-58</td>
<td>generally none</td>
<td>high and variable (14.5-27%)</td>
<td>Clinopyroxene; biotite; normative</td>
</tr>
<tr>
<td>Alkaline</td>
<td>&gt;390</td>
<td>very steep; near 1.07 K2O per 5% SiO2</td>
<td>4.4-6.0</td>
<td>25-34</td>
<td>6.1-14+</td>
<td>0.3-1.0</td>
<td>6.5-12.0</td>
<td>45-52</td>
<td>moderate</td>
<td>variable (4-29%)</td>
<td>Olivine and/or normative clinopyroxene; biotite; normative clinopyroxene;</td>
<td></td>
</tr>
</tbody>
</table>

* K2O/SiO2 index = potash content at 67.5% silica in a K2O/SiO2 Harker variation diagram; θ index = SiO2 - 47 (Na2O - K2O) / Al2O3 from Sugimura (1968); S index = (Na2O + K2O) / SiO2 - 43 from Rittman (1960).
more potassic and alkalic. Both these groups are much more potassic at a given silica content than the mid- to Late-Cretaceous metaluminous granitoids which are classified here as calc-alkalic.

The muscovite-bearing peraluminous granitoid suite is probably divisible into high and low potassium groups. Data contained in Castor and others (1977) indicates that the Washington muscovite granitoids are considerably more potassic than all other Cordilleran muscovite-bearing plutons. Importantly, $K_2O/Na_2O$ ratios in available Washington data are significantly greater than 1.0. $K_2O/Na_2O$ ratios in non-Washington examples are near or less than 1.0, betraying a distinct sodic emphasis for the low-K group. Compared to the mid-to-Late Cretaceous metaluminous calc-alkalic granitoids (see Figure 5-2), the muscovite-bearing granitoids suite (exclusive of Washington) is less potassic at equivalent silica contents. Compared to world-wide peraluminous analogues, the Cordilleran low-K peraluminous granitoids exhibit significant differences (Figures 5-4, 5-5, and 5-6). On a $K_2O-SiO_2$ variation diagram (Figure 5-4), most of the Cordilleran low-K group have lower potassium contents than world-wide analogues at a comparable silica content. The more calcic nature of the Cordilleran low-K muscovite granitoids may be seen on a $CaO-K_2O$ variation diagram (Figure 5-5) where at any given $K_2O$ content, Cordilleran low-K muscovite granitoids commonly contain more calcium than muscovite-bearing granitoids elsewhere in the world. On a $Na_2O-CaO-K_2O$ ternary diagram Cordilleran muscovite granites are clearly displaced towards the $Na_2O-CaO$ join, indicating a more sodic character when compared to the global control group. An exception to this are the high-K muscovite granites of northeastern Washington which plot entirely within the global control group on all three diagrams. Significantly, the Washington high-K granites also plot within the high-uranium portion (stippled area) of the global control group. Not surprisingly, the Washington muscovite-bearing granites are highly uraniferous.

Uranium and Thorium Geochemistry

Histograms of uranium and thorium contents in samples of each plutonic suite are displayed in Figures 5-7, 5-8, and 5-9. As predicted from major-element chemistry, significant differences are evident. In the metaluminous series, uranium and thorium values closely follow potassium content; that is, the comparatively low-K calc-alkalic mid-Cretaceous suite has lower uranium and thorium values than the more alkalic Jurassic and Eocene alkali-calcic suites. In particular, the Ruby monzogranite of the Jurassic group contains the highest thorium contents of any granitoid we sampled (an average of 64 ppm thorium as compared with the 17 ppm thorium value for average granite given by Taylor, 1964).
Figure 5-2. $K_2O - SiO_2$ Harker variation diagram for selected Jurassic Cordilleran alkali-calcic granitoids. Symbols are as follows: Jurassic plutons in south-east Arizona ($\triangle$); monzogranite, Ruby Mountains, Nevada ($\circ$); granite, Ruby Mountains, Nevada ($\bigodot$); granitoid rocks of Kious Basin area, Snake Range, Nevada ($\oplus$).
Figure 5-3. $K_2O - SiO_2$ Harker variation diagram for selected Cordilleran granitoids. Symbols are as in Figure 5-1 except for the following: open triangles are Whipple Mountain data; northeast Washington muscovite-bearing peraluminous plutons are distinguished from Cordilleran analogues by a solid triangle (△). Data sources are included in Appendix A.
Figure 5-4. $K_2O - \text{SiO}_2$ Harker variation diagram for Cordilleran muscovite-bearing peraluminous granitoids as compared to world-wide analogues. Symbols are as follows: Santa Catalina Mountains, southeast Arizona ($\bigcirc$); Whipple Mountains, southeast California ($\triangle$); Pole Canyon - Can Young Canyon area, Snake Range, eastern Nevada ($\bigcirc$). World-wide analogues are represented by fields on the diagram. Stippled field contains uraniferous (greater than 6 ppm U) world-wide muscovite-bearing granitoids.
Figure 5-5. CaO – K₂O variation diagram for Cordilleran muscovite-bearing peraluminous granitoids as compared to world-wide analogues. For symbology, see Figure 5-4.
Figure 5-6a. CaO - Na$_2$O - K$_2$O ternary diagram for Cordilleran muscovite-bearing peraluminous granitoids as compared to world-wide analogues. For symbology, see Figure 5-4.

Figure 5-6b. Na$_2$O - K$_2$O variation diagram for Cordilleran muscovite-bearing peraluminous granitoids as compared to world-wide analogues. For symbology, see Figure 5-4.
The low-K muscovite-bearing granitoids have the lowest uranium and thorium contents of any of the four plutonic suites. This observation clearly invalidates the common notion that all muscovite granites are uraniferous. However, the high-K Washington muscovite-bearing plutons as a group have the highest uranium contents of any plutonic suite in the core complexes. Despite a large difference in uranium content, both the low- and high-K plutonic groups generally have low Th/U ratios (commonly less than 2); there is no significant difference in Th/U ratios between the two groups. In general, Th/U ratios appear to be lowest in the muscovite-bearing granitoids (typically less than 2), higher in the middle to Late Cretaceous calc-alkalic suite (approximately 4.5), and highest in the alkali-calcic suite (about 7). In general, the Th/U ratio increases with increasing potassium content.

Uranium Favorability of Core Complex Plutonic Rocks

We have found that uranium favorability criteria for core complex plutonic rocks closely parallels those established for world-wide peraluminous granitoids (see Chapter 4). These criteria appear to apply equally well to both Cordilleran metaluminous and peraluminous series. Chapter 4 established that for world-wide peraluminous (muscovite-bearing) granitoids, uranium and potassium content are closely related and that samples above approximately 4% K2O commonly contain greater than 6 ppm uranium. The same criteria appears to hold for granitoids in Cordilleran metamorphic core complexes (see Figure 5-10).

In Cordilleran granitoids one can observe a conspicuous inflection point in uranium variation with potassium above 4% K2O. Samples from 2-4% K2O all contain less than 6 ppm uranium while those above 4% K2O commonly contain greater than 6 ppm uranium (especially the Washington muscovite granites). This relationship holds for both Cordilleran metaluminous and peraluminous series. Significantly, as in the global control group, no discernable variation of uranium with sodium was found (see Figure 5-11). Uranium favorability for each of the four major plutonic suites is summarized on K2O - U variation diagrams (Figures 5-12, 5-13, 5-14, and 5-15) using boundaries of K2O ≥ 4% and uranium values ≥ 6 ppm to delimit favorable regions on the plots. From Table 5-1 and Figure 5-14 it can be seen that the low-K peraluminous muscovite-bearing granitoid suite is the least favorable. In fact, none of the samples are favorable if the lithologically unusual garnet schlieren leucogranite sample from the Santa Catalina complex is excluded. The calc-alkalic hornblende-biotite granitoid rocks are also unfavorable with only 5% of the samples plotting in the favorable region (see Table 5-1 and Figure 5-13).
Figure 5-7. Frequency histogram of thorium contents in selected Cordilleran peraluminous and metaluminous granitoid rocks. Class intervals are 2 ppm. Boxes marked with 'X' symbol denote muscovite-bearing plutons from northeast Washington.
Eocene to Oligocene
alkali-calcic biotite
hornblende suite

Jurassic alkali-calcic
biotite suite

Figure 5-8. Frequency histogram of uranium contents in selected Cordilleran metaluminous alkali-calcic granitoid rocks. Class intervals are 0.5 ppm.
Middle to Late Cretaceous calc-alkaline biotite-hornblende suite

Middle Cretaceous to Eocene peraluminous muscovite-bearing suite

Figure 5-9. Frequency histogram of uranium contents in selected Cordilleran metaluminous calc-alkaline and muscovite-bearing peraluminous granitoids. Boxes marked with 'X' symbol denote muscovite-bearing intrusions from northeast Washington. Stippled boxes denote uranium analyses in granitoid rocks from areas assigned 'core-complex' nomenclature.
Figure 5-10. $U-K_2O$ variation diagram for Cordilleran metaluminous and peraluminous granitoids as compared to world-wide peraluminous and/or muscovite-bearing granitoids. For symbology, see Figures 5-1 and 5-3 plus the following: selected Jurassic Cordilleran alkali-calcic granitoids are denoted by an open square ($\square$). Cordilleran field is enclosed by a solid line. World-wide muscovite-bearing granitoid field is enclosed by a dash-dot line. Uraniferous Cordilleran granitoids containing greater than 6 ppm uranium are marked by parallel lines slanted to the left. Uraniferous world-wide muscovite-bearing granitoids containing more than 6 ppm uranium are marked by parallel lines slanted to the right.
Figure 5-11. $U - Na_2O$ variation diagram for selected metaluminous and peraluminous Cordilleran granitoid rocks. For symbology, see Figures 5-1 and 5-10.
Figure 5-12. \( U - K_2O \) variation diagram for selected Cordilleran Jurassic alkali–calcic granitoids showing favorable uranium region. Open circles (○) are Jurassic granitoids from various designated core complex areas. Circles with plus sign (⊕) show the uraniferous Jurassic monzogranite pluton in the Ruby Mountains at northeast Nevada.
Figure 5-13. U - K₂O variation diagram for middle to late Cretaceous Cordilleran metaluminous calc-alkalic granitoid rocks showing favorable uranium region. Symbols are as follows: Southern California batholith (○). Core complex plutons are as follows: Leatherwood quartz diorite in the Santa Catalina complex, southeast Arizona (△); Chirreon Wash granodiorite in Tortolita complex, southeast Arizona (△); Granite wash granodiorite at the extreme western end of the Harcuvar complex, west-central Arizona (□); and assorted mid-Cretaceous plutons in northeast Washington (□).
Figure 5-14. U-K$_2$O variation diagram for Cordilleran muscovite-bearing peraluminous granitoids showing favorable uranium region. Symbology is as follows: northeast Washington plutons (△); all other Cordilleran muscovite-bearing peraluminous plutons (○).
Figure 5-15. $U - K_2O$ variation diagram for selected Eocene-Oligocene metaluminous Cordilleran alkali-calcic granitoids showing favorable uranium region. Symbology is as follows: northeast Washington plutons (△); Catalina suite plutons from Santa Catalina and Tortolita complexes, southeast Arizona (○); Harrison Pass Quartz Monzonite, Ruby Mountains, Nevada (□); and Skinner Canyon granite, Kern Mountains, east-central Nevada (△).
In contrast, the alkali-calcic suites (Table 5-1 and Figures 5-12 and 5-15) are much more favorable; 20% of the samples of the Eocene-Oligocene alkali-calcic granitoids plot in the favorable region while over 35% of the Jurassic alkali-calcic rock samples have favorable chemistry. However, the most favorable plutons of all are the northeast Washington high-$K$ peraluminous muscovite-bearing plutons (Table 5-1 and Figure 5-14) with 74% of the samples (data extracted from Castor and others, 1977) above 6 ppm uranium. Fully 50% of the Washington samples have uranium contents over 10 ppm. This compares to 15% for the Jurassic and 7% for the metaluminous Eocene-Oligocene alkali-calcic suites. It is uncertain whether the difference between the two alkali-calcic groups is statistically significant because we have only a limited number of samples from the Jurassic granitoids. However, the slightly higher amounts of potassium in the Jurassic suite should logically be accompanied by slightly higher uranium contents.

We have made a further preliminary attempt to refine uranium favorability in the metaluminous Cordilleran granitoids by comparing intersuite alkalinity with intrasuite degree of differentiation and evaluating their effect on uranium and thorium content. We accomplished this by plotting uranium and thorium contents as a function of differentiation for individual suites. This compilation used Larsen Factor $[\frac{1}{3} SiO_2 + K_2O - (CaO + MgO + FeO)]$ to measure the degree of intrasuite differentiation. Intersuite alkalinity was determined by use of the $K_{57.5}$ index (% K$_2$O at 57.5% SiO$_2$ on a Harker K$_2$O-SiO$_2$ variation diagram). The $K_{57.5}$ values are adapted to a modified Peacock nomenclature following Keith (1978). An attempt was also made to take tectonic setting into account. Generalized results of the preliminary compilation (74 igneous complexes inventoried in the literature so far) are presented in Table 5-3. Uranium and thorium values for the 74 igneous complexes at a Larsen Factor of 10 are plotted against $K_{57.5}$ in Figures 5-16 and 5-17. The compilation applies only to world-wide metaluminous rock series with enough silica variation to allow determination of the $K_{57.5}$ index. Peraluminous series are too silicic (>65% SiO$_2$) to allow determination of a $K_{57.5}$ value.

The data in Table 5-3 indicate that for a given alkalinity (magma suite column on Table 5-3), the concentrations of radioelements (U, Th, and K) within the magma suite increase with degree of differentiation. Also, with increasing intersuite alkalinity, the overall uranium and thorium contents systematically increase. Indeed a complete spectrum of increasing uranium and thorium contents with increasing intersuite alkalinity...
TABLE 5-3:
URANIUM-THORIUM VARIATION
WITH INTRASUITE DIFFERENTIATION AND INTERSUITE ALKALINITY
IN METALUMINOUS IGNEOUS ROCK SERIES

<table>
<thead>
<tr>
<th>Suite Designation</th>
<th>Number of Complexes Inventoried</th>
<th>$K_{57.5}$ Increasing Differentiation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>LARSON FACTOR</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-10 0 10 20 25</td>
</tr>
<tr>
<td>Calcic</td>
<td>19</td>
<td>.98^{3} (.25-1.2) 1.23(.87)^{2} .47(1.4)^{2} .62(2.4)^{2} 1.29(4.7)^{2} 2.14(6.9)^{2}</td>
</tr>
<tr>
<td>Calc-alkalic</td>
<td>17</td>
<td>1.69 (1.2-2.4) .36(1.8) .90(3.3) 1.96(6.2) 2.16(13.7) 4.4 (18.4)</td>
</tr>
<tr>
<td>High potassium calc-alkalic</td>
<td>10</td>
<td>2.71 (2.4-3.0) .8 (1.4)^{3} 1.53(6.2) 2.63(8.4) 3.58(12.2) 4.1 (19.2)</td>
</tr>
<tr>
<td>Alkali-calcic</td>
<td>18</td>
<td>3.68 (3.0-4.5) 1.2 (3.8) 2.0 (6.7) 2.82(9.35) 5.33(16.9) 8.6 (25.3)</td>
</tr>
<tr>
<td>Alkalic</td>
<td>10</td>
<td>5.52 (4.5-9.3) 1.6 (6.6) 2.65(13.7) 4.11(19.9) 8.3 (27.9) 11.2 (47.6)</td>
</tr>
</tbody>
</table>

Notes: 1) Number on top is the average $K_{57.5}$ value for the igneous suites inventoried; number in parentheses are range of $K_{57.5}$ values for the igneous suites inventoried.

2) Numbers in parentheses are $K_{57.5}$ values. All uranium and thorium values are in ppm.

3) i.d. = insufficient data.
exists from calcic to alkalic suites. This chemical relationship holds in a general way regardless of tectonic setting although igneous rocks in continental regions have somewhat higher thorium contents than igneous rocks in oceanic regions of the same alkalinity (Figure 5-17).

Suites that contain greater than 6 ppm uranium at some point in the differentiation sequence are not present below $K_{57.5} = 2.5$. Significantly, few of the igneous rocks in suites with $K_{57.5}$ values less than 2.5 have $K_2O$ values greater than 4.0. Average uranium abundances of 6 ppm or greater appear in highly differentiated phases of the alkali-calcic (Larsen factor equals 25) and alkalic (Larsen Factor equals 20 and 25) suites. We consider these rocks especially favorable for potential uranium mineralization.

Metaluminous plutons sampled in core complex areas (refer to list on Table 5-1) show no significant departure from the global metaluminous data base (as generalized in Table 5-3). Our data indicate that uranium and thorium content of a plutonic rock in a core complex is a function of that rock's suite type or alkalinity (particularly its potassium content). Potassium-uranium data for Cordilleran peraluminous muscovite granites plot in two distinct areas relative to their global counterparts (Figure 5-10). The Cordilleran low-K peraluminous granitoids generally contain markedly lower uranium content than global relatives of comparable $K_2O$ content. In contrast, northeast Washington peraluminous gneisitoids are clearly separated from the low-K peraluminous suite by their higher $K_2O$ and uranium contents; they plot largely within the uraniferous part of the global field.

Summary

In summary, we are strongly impressed with the influence of potassium content on uranium geochemistry of plutonic rocks in core complexes in particular and in igneous rocks throughout the world in general. Regardless of aluminum character, suite alkalinity, or degree of differentiation, uranium values of greater than 6 ppm will almost invariably not be present unless $K_2O$ content in an igneous rock exceeds 4%. To wit, a $K_2O$ threshold of greater than 4% is a necessary condition for uranium favorability of igneous rock.
Uranium (ppm)

Calc-alkalic

High potassium calc-alkalic

Alkaline-calcic

Alkaline

Figure 5-16. $K_{575}$ - Uranium variation diagram for world-wide metaluminous rock series. Calcic through alkaline nomenclature follows conventions outlined in Table 5-2.
Figure 5-17. Thorium variation diagram for world-wide metaluminous igneous rocks. Data from Keith (1980, unpublished compilation). Symbology is as follows: Volcanics erupted on oceanic basement or plutonic rocks emplaced into oceanic basement (○); Volcanics erupted onto or plutons emplaced into greater than 300 m.y. old lithosphere (□); volcanics erupted next to or plutons emplaced marginal to Precambrian aged lithosphere or volcanics and plutons emplaced onto or into 50 to 300 m.y. old lithosphere (△). Calcic through alkalic nomenclature follows conventions outlined in Table 5-2.
GEOCHEMISTRY OF METAMORPHIC AND MYLONITIC ROCKS

Introduction

Metamorphic and mylonitic rocks are widely present in Cordilleran metamorphic core complexes. As discussed in Chapter 4, it is important to consider metamorphism and mylonitization when evaluating the uranium favorability of metamorphic core complexes. The influence of metamorphism and mylonitization on the uranium content of rocks is partially discussed in the geological literature, but an insignificant proportion of this literature deals directly with core complex areas. Because of this deficiency of existing pertinent literature, we have collected geochemical samples of metamorphic and mylonitic rocks in the Ruby, Buckskin, Harcuvar, Harquahala, Tortolita, and Santa Catalina complexes (see Appendix E). In addition, we have examined the radioactivity (total gamma count via scintillometry) of these and many other complexes (see sections on uranium favorability of individual complexes in Appendix D).

In the Santa Catalina and Tortolita complexes of Arizona, we collected geochemical samples along traverses from non-metamorphosed to metamorphosed rocks and from nonmylonitic to mylonitic phases of the same rock unit. The following discussion of geochemistry of metamorphic and mylonitic rocks is largely concerned with these detailed sample traverses. The results of these traverses are most enlightening, but additional traverses in other, more uraniumiferous core complexes are needed to substantiate our conclusions and to indicate how applicable these conclusions are to the rest of the complexes.

Metamorphism and Mylonitization

Among geologists that have studied the core complexes, there has been much disagreement concerning the relative importances of metamorphic and mylonitic processes (for example, see Keith and others, 1980 for a summary of metamorphic versus mylonitic characterizations of rocks in the Santa Catalina complex, Arizona). Some workers regard foliation in "gneissic" (sensu lato) rocks as a relict of pre-metamorphic sedimentary layering. An alternative view dating back to Lindgren (1904)
in the Bitterroot complex of Idaho and Montana is that the foliation in these rocks is the result of mylonitization imposed on an originally homogenous and commonly granitic protolith. Metamorphic core complexes contain foliated rocks with both types of origins; metamorphosed sedimentary rocks predominate some complexes (for example the Albion Range of Idaho and the Buckskin Mountains of Arizona), while mylonitically deformed plutonic rocks prevail in others (such as the Bitterroot Mountains of Montana and the Santa Catalina and Tortolita Mountains of Arizona). Due to the fact that our detailed sample traverses are from the mylonitically dominated Tortolita and Santa Catalina complexes, we will be largely concerned with the effect of mylonitization on the geochemistry of granitic rocks. In this context, we have made an attempt to distinguish the geochemical changes due to mylonitization from those related to pre- or symmylonitization plutonism. To this end, we have chosen two areas that may represent end members of this probable plutonic-mylonitic continuum: 1) a traverse from virtually undeformed to highly deformed varieties of Precambrian Pinal Schist and 1.4 b.y.-old granite in the northeastern Tortolita complex; and 2) a traverse through the banded gneiss complex of the Santa Catalina forerange where 1.4 b.y. -old granite has been lit-par-lit injected by horizontal sills of Eocene Wilderness granite (Keith and others; 1980). The Tortolita traverse probably represents the mylonitic, non-plutonic case; any gains or losses are probably related to the mylonitic process alone. In contrast, the Santa Catalina forerange traverse probably represents a plutonic-dominated case where intense plutonism may have been accompanied by mylonitic deformation. We will examine the mylonitic non-plutonic traverse first.

Tortolita Mountains Traverse

Geologic Setting

We have sampled in the northeastern Tortolita Mountains because geological mapping (Banks and others, 1977; Keith and others, 1980) has revealed a complete spectrum of intensities of deformation. In this mountain range, we were able to collect samples along continuous traverses from nonmylonitic to mylonitic 1.4 b.y. -old Oracle Granite. In the same area we collected Pinal Schist that has retained its original Precambrian lithology and structure, and a schistose band of probable Pinal Schist which has been affected by intense Mesozoic–Cenozoic deformation.
Figure 5-18. Diagramatic cross-section of rocks in northeastern Tortolita Mountains, Arizona. Numbers on diagram refer to Bendix sample numbers (see Appendix E for sample descriptions and analytical data).
The traverse (see Figure 5-18) from Oracle Granite to mylonitic derivatives encountered the following gradational sequence of rocks (from north to south):

1) unfoliated, biotite-rich Oracle Granite with minor chlorite and sericite;

2) cataclastically deformed Oracle Granite that contains abundant chlorite and locally exhibits discrete, non-penetrative zones of foliated rock; and

3) muscovitic granitic gneiss derived from Oracle Granite. The major mica is probably paragonitic muscovite.

The transitions between the three rock types are completely gradational; the transition zones contain lithologies and structures of both adjoining rock types.

The second suite of samples are from:

1) Precambrian Pinal Schist just outside of the complex;

2) a schistose band of probable Pinal Schist within the complex in contact with the mylonitic Oracle Granite that we sampled (see Figure 5-18); the band experienced Mesozoic-Cenozoic deformation and metamorphism.

The contact between the muscovite granitic gneiss and the schistose band has been described as gradational by Banks (1977, 1980) who suggested the schistose rock represented a more highly comminuted mylonitic version of the granitic gneiss and Chirreon Wash granodioritic pluton to the south. Based on Rb-Sr trace element and isotopic data and sharp contacts with the mylonitic muscovite-bearing Oracle Granite, Keith and others (1980) consider the problematic schistose band to be Pinal Schist that has been deformed and recrystallized (probably during intrusion of the late Cretaceous Chirreon Wash granodiorite pluton). Chemical data generated during this study provides additional support for the Keith and others Pinal Schist postulate.

Geochemistry

Selected geochemical data for the Precambrian Oracle Granite and Precambrian Pinal Schist from the northeastern Tortolita Mountains, Arizona are plotted in Figures 5-19 and 5-20. Geochemical
data for mylonitic versus nonmylonitic Oracle Granite reveal a variety of distinct chemical changes (Figure 5-19). Silica contents of deformed Oracle Granite (chloritic and muscovitic phases) are conspicuously higher than those for undeformed Oracle Granite. Contents of Na, Mg, K, and Sr are comparable among undeformed and chloritic phases, indicating that these four elements experienced insignificant chemical changes. In contrast, Na and Sr are strongly enriched in the muscovitic phases while Mg and K are distinctly depleted in these phases. Calcium contents progressively decrease from undeformed to chloritic, and muscovitic phases. Other elements (Ce, Zn, V, Li, Y, Sc, Ni, P, T, Mn, and possibly U) are also variously depleted in the deformed muscovitic phases relative to undeformed Oracle Granite. Contents of Rb, Cr, and Th do not exhibit any major changes between the three phases. Uranium contents also show no significant gains or losses as a function of deformation and metamorphism. The uranium contents of deformed-metamorphosed Oracle Granite are comparable with those for undeformed 1.4 b.y.-old granitoids of Arizona and New Mexico (Figure 5-21). Thus, we conclude from this data, that the imposition of profound textural and mineralogical changes was not accompanied by significant depletion or enrichment of uranium in these samples.

Selected geochemical data for the older Precambrian Pinal Schist are summarized on Figures 5-20. On each diagram, in Figure 5-20, analyses of "unmodified" Pinal Schist (left side of diagrams) are compared with samples of Pinal Schist that have been modified or overprinted by core-complex deformation and metamorphism. The data reveal that "modified" Pinal Schist has lower contents of Sc, Zr, Li, V, and Zn relative to "unmodified" Pinal Schist. More limited major-element data suggest depletion of Fe, Mg, and Na during the metamorphism and deformation. These losses are compensated for by conspicuous gains in silica. Trends of other elements are less obvious. The data indicate that Pinal Schist and its deformed-metamorphosed derivatives are all characterized by high chromium contents, indicating a low mobility of chromium. The relatively high amounts of chromium seem unique to Pinal Schist and may be a useful tracer for identifying Pinal Schist protolith in other schistose metamorphic rocks which have been confused with Pinal Schist but were derived from other protoliths such as 1400 m.y. granitic rock.

Like many of the other elements, uranium and thorium values in the Pinal Schist seem little affected by 'core complex' metamorphism. Uranium and thorium data of "unmodified" schist show little systematic changes with respect to the "unmodified" Pinal
Figure 5-19. Gains and losses for selected major element oxides and trace elements in undeformed versus deformed varieties of Oracle Granite, northeastern Tortolita Mountains, Arizona.
Figure 5-20. Gains and losses for selected major element oxides and trace elements in 'unmodified' versus 'modified' varieties of Pinal Schist, northeastern Tortolita Mountains, Arizona.
Schist control sample (Figure 5-20). Indeed uranium contents of "modified" schist samples show little departure from uranium values typical of Pinal Schist phyllite lithologies throughout southeast Arizona (Figure 5-22). For these samples, we conclude that contents of uranium, thorium, and chromium, were unchanged during metamorphisms.

Santa Catalina Traverse

Geologic Setting

One of the best examples of core-complex plutonic, metamorphic and mylonitic phenomena is spectacularly exposed in the Santa Catalina Mountains of southeast Arizona. Numerous workers have studied various aspects of the geology of the mountain range (see summaries and references in Keith and others, 1980; Davis, 1980; Banks, 1980; Shakel, 1978; Creasey and others, 1977). The following discussion of the geology of the range is based on the synthesis and interpretation of Keith and others (1980).

The plutonic geology of much of the Santa Catalina Mountains (see also Appendix D) can be viewed in terms of a stacked sill complex that consists of five major lithologic pseudo-stratigraphic assemblages which have gently inclined tabular forms and boundaries, (see Figure 5-23). The five assemblages are described below from structurally lowest to highest levels.

Seven Falls Foliated Biotite Granite. A leucocratic foliated biotite granite named the Seven Falls Gneiss by Petersen (1968) is exposed at the lowest structural level of the Wilderness stacked sill complex. We hereafter refer to this unit as the Seven Falls foliated granite. Chemical analyses and modal data for the Seven Falls foliated granite are presented in Tables 5-4 and 5-5 respectively.

Forerange banded gneiss complex. A banded gneiss complex overlies the Seven Falls foliated granite throughout the Santa Catalina forerange. This gneiss complex consists of alternating light-and dark-colored lithologies which are interlayered on a wide variety of scales. The dark components are rich in biotite (20%) and are predominantly composed of mylonitically deformed Oracle Granite (1.45 b.y. B.P. emplacement age). Keith and others (1980) have interpreted most of the light-colored components of the gneiss complex as sills of Eocene Wilderness that were injected in lit-par-lit fashion into darker phases of 1.45 b.y. B.P. Oracle Granite. Petrographic data (Sherwonit, 1974; see Table 5-4) indicate that Wilderness equigranular biotite granite sills which are structurally low in the banded gneiss complex are more biotite-rich than those higher in the banded gneiss complex.

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Figure 5-21. Frequency histogram of uranium contents in Precambrian (1400 m.y. old) granitoid rocks in southern Arizona and southwestern New Mexico. Class intervals are one ppm.
Figure 5-22. Frequency histogram of uranium contents in Precambrian (1700 m.y. old) Pinal Schist of southeastern Arizona. Class intervals are one ppm.
<table>
<thead>
<tr>
<th>Reference</th>
<th>Reference</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>part of Lenmon Rock leucogranite</td>
<td>part of Lenmon Rock leucogranite</td>
<td>part of Lenmon Rock leucogranite</td>
</tr>
<tr>
<td>Control Road pegmatite</td>
<td>Control Road pegmatite</td>
<td>Control Road pegmatite</td>
</tr>
<tr>
<td>part of Lenmon Rock leucogranite</td>
<td>part of Lenmon Rock leucogranite</td>
<td>part of Lenmon Rock leucogranite</td>
</tr>
<tr>
<td>Caseco pegmatite -</td>
<td>Caseco pegmatite -</td>
<td>Caseco pegmatite -</td>
</tr>
<tr>
<td>about 30 m below upper contact of</td>
<td>about 30 m below upper contact of</td>
<td>about 30 m below upper contact of</td>
</tr>
<tr>
<td>Wilderness Granite with meta-Apache Group</td>
<td>Wilderness Granite with meta-Apache Group</td>
<td>Wilderness Granite with meta-Apache Group</td>
</tr>
<tr>
<td>Aplite - same location as</td>
<td>Aplite - same location as</td>
<td>Aplite - same location as</td>
</tr>
<tr>
<td>above</td>
<td>above</td>
<td>above</td>
</tr>
<tr>
<td>Wilderness Granite -</td>
<td>Wilderness Granite -</td>
<td>Wilderness Granite -</td>
</tr>
<tr>
<td>various locations within upper half of</td>
<td>various locations within upper half of</td>
<td>various locations within upper half of</td>
</tr>
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<td>main sill; average of 10 analyses</td>
<td>main sill; average of 10 analyses</td>
<td>main sill; average of 10 analyses</td>
</tr>
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<td><strong>East Fork gneiss -</strong></td>
<td><strong>East Fork gneiss -</strong></td>
</tr>
<tr>
<td>layer just below base of main Wilderness</td>
<td>layer just below base of main Wilderness</td>
<td>layer just below base of main Wilderness</td>
</tr>
<tr>
<td>Granite sill; average of 16 analyses</td>
<td>Granite sill; average of 16 analyses</td>
<td>Granite sill; average of 16 analyses</td>
</tr>
<tr>
<td><strong>Thimble Peak gneiss</strong></td>
<td><strong>Thimble Peak gneiss</strong></td>
<td><strong>Thimble Peak gneiss</strong></td>
</tr>
<tr>
<td>- average of 16 analyses</td>
<td>- average of 16 analyses</td>
<td>- average of 16 analyses</td>
</tr>
<tr>
<td><strong>Sabino Narrows gneiss -</strong></td>
<td><strong>Sabino Narrows gneiss -</strong></td>
<td><strong>Sabino Narrows gneiss -</strong></td>
</tr>
<tr>
<td>slight bands; average of 8 analyses</td>
<td>slight bands; average of 8 analyses</td>
<td>slight bands; average of 8 analyses</td>
</tr>
<tr>
<td><strong>Gibbons Mt. gneiss -</strong></td>
<td><strong>Gibbons Mt. gneiss -</strong></td>
<td><strong>Gibbons Mt. gneiss -</strong></td>
</tr>
<tr>
<td>average of 7 analyses</td>
<td>average of 7 analyses</td>
<td>average of 7 analyses</td>
</tr>
<tr>
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<td><strong>Soldier Canyon gneiss -</strong></td>
<td><strong>Soldier Canyon gneiss -</strong></td>
</tr>
<tr>
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<td>slight bands; average of 15 analyses</td>
<td>slight bands; average of 15 analyses</td>
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<tr>
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<td><strong>Seven Falls gneiss; average</strong></td>
<td><strong>Seven Falls gneiss; average</strong></td>
</tr>
<tr>
<td>of 7 analyses</td>
<td>of 7 analyses</td>
<td>of 7 analyses</td>
</tr>
</tbody>
</table>

- Included under "others" by Pilkington (1962).
- "An" values are from Peterson (1963).
- **Terminology for gneiss units in Santa Catalina forearange is from Peterson (1968). These units are interpreted by us as injection sheets on lower levels of the Wilderness sill complex.

m - microcline
o - orthoclase

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Figure 5-23. Cross section through the Santa Catalina forerange and main range showing major subdivisions of the Wilderness sill complex. Samples collected and analyzed for this project are shown as 800 numbers on the cross-section. CH numbers are sodium-potassium analyses in Wilderness.
TABLE 5-5:
AVERAGE CHEMICAL COMPOSITION OF COMPONENTS
WITHIN THE WILDERNESS GRANITE STACKED SILL COMPLEX

Increasing structural level

<table>
<thead>
<tr>
<th>Element Oxide</th>
<th>Seven Falls</th>
<th>Lower Portion</th>
<th>Upper Portion</th>
<th>Garnet schlieren</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Foited granite</td>
<td>Foited biotite granite</td>
<td>Two-mica granite</td>
<td>Leucogranite</td>
</tr>
<tr>
<td>SiO₂</td>
<td>64.88</td>
<td>70.6</td>
<td>74.06</td>
<td>75.2</td>
</tr>
<tr>
<td>AL₂O₃</td>
<td>16.6</td>
<td>14.5</td>
<td>14.67</td>
<td>13.1</td>
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<tr>
<td>Fe₂O₃</td>
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<td>1.68</td>
<td>1.39</td>
<td>1.95</td>
</tr>
<tr>
<td>FeO</td>
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<td>0.32</td>
<td>0.19</td>
<td>0.06</td>
</tr>
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<td>CaO</td>
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<td>1.81</td>
<td>1.3</td>
<td>0.44</td>
</tr>
<tr>
<td>Na₂O</td>
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<td>3.51</td>
<td>3.92</td>
<td>4.04</td>
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<tr>
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<td>4.23</td>
<td>3.62</td>
<td>4.48</td>
</tr>
<tr>
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<td>0.14</td>
<td>0.10</td>
<td>0.041</td>
<td>0.09</td>
</tr>
<tr>
<td>U</td>
<td>0.1</td>
<td>0.90</td>
<td>1.37</td>
<td>3.5</td>
</tr>
<tr>
<td>Th</td>
<td>6</td>
<td>6.1</td>
<td>2&lt;6.8</td>
<td>2&lt;2.5</td>
</tr>
<tr>
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<td>4</td>
<td>5</td>
<td>3.3</td>
<td>&lt;5.0</td>
</tr>
<tr>
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<td>2050</td>
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<td>110</td>
</tr>
<tr>
<td>Mn</td>
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</tr>
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<td>8.5</td>
</tr>
<tr>
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<td>&lt;25</td>
</tr>
<tr>
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<td>&lt;3.5</td>
<td>&lt;7.2</td>
</tr>
<tr>
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<td>1100</td>
<td>2050</td>
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</tr>
<tr>
<td>Li</td>
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<td>7.1</td>
<td>19.5</td>
</tr>
<tr>
<td>Ni</td>
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<td>2.1</td>
<td>7.5</td>
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<td>Sr</td>
<td>760</td>
<td>490</td>
<td>320</td>
<td>28.3</td>
</tr>
</tbody>
</table>

Notes: 1) n = number of samples
2) Less than sign (<) indicates some samples for the element in question were below the detection limit for that element.
3) Element analytical values are in ppm.
4) Element oxide values are in weight percent.
Pegmatitic Wilderness phases contain muscovite only in the upper two-thirds of the banded gneiss complex and are more biotite-rich in the lower third. We believe the forerange banded gneiss complex extends northward beneath a large higher-level sill of Wilderness granite in the Santa Catalina main range as a large, tabular, highly injected sheet (refer to Figure 5-23).

Main Range Wilderness Granite Sill. A batholithic, laccolithic sill of Eocene wilderness granite overlies the banded gneiss complex. This 2.5 km-thick intrusion exhibits an asymmetrical laccolithic geometry. The Wilderness laccolithic sill is composed of equigranular granite and interlayered pegmatite, aplite, and alaskite. Throughout most of the laccolithic sill, the equigranular granite contains biotite, muscovite, and garnet. However, near the base of the laccolith, the equigranular granite contains biotite and sparse garnet, but no muscovite. We will refer to the gradational zone between the main two-mica phase and the structurally lower, biotite-only phase as the "muscovite-in boundary". Pegmatites, aplite, and alaskite within the Wilderness main range laccolith generally contain muscovite and garnet. Modal analyses and average chemical data for the main range laccolithic sill are presented in Tables 5-4 and 5-5 respectively.

Lemmon Rock Leucogranite. Late-stage pegmatite, aplites, and leucogranites are especially abundant near the top of the main range Wilderness laccolith. In many places, alasko-pegmatitic rocks form a leucogranite "cap" named Lemmon Rock leucogranite by Shakel (1978). This "cap" grades downward into the main range laccolithic sill of equigranular two-mica granite and intrudes upward into the late Cretaceous Leatherwood quartz diorite and younger Precambrian Apache Group. Garnet is locally abundant and in places forms spectacular "railroad track" bands of garnet schlieren. Representative modes and chemical analyses of the Lemmon Rock leucogranite and the lithologically unusual garnet schlieren leucogranite are presented in Tables 5-4 and 5-5 respectively.

Sedimentary "cover" and Leatherwood Quartz Diorite. The uppermost of the five major pseudostratigraphic assemblages is composed of Precambrian Apache Group sedimentary rocks, Precambrian diabase, Paleozoic sedimentary rocks, and sill-like apophyses of Late Cretaceous Leatherwood quartz diorite which
have chiefly intruded between the Apache Group and Paleozoic sedimentary rocks. The sedimentary rocks are highly meta-morphosed directly above their contact with the underlying Wilderness granite laccolith and adjacent to the quartz diorite sills; they are less metamorphosed up structural section to the north.

Mylonitic Deformation

Rocks in five major pseudostratigraphic assemblages locally exhibit a gently inclined mylonitic foliation. The banded gneiss complex of the forerange, and the lower half of the Wilderness granite laccolith are generally mylonitic; rocks higher in the mountain range are generally not mylonitic (see Figure 5-23). In the context of this project, James DuBois has examined many thin-sections of mylonitic and undeformed phases of Wilderness granite, (see Appendix E), and has established the following generalized deformation sequence in muscovite-bearing, equigranular phases of the Wilderness pluton:

1) quartz recrystallizes, becoming amoeboid;
2) quartz becomes finer-grained and feldspars become somewhat rounded;
3) quartz forms streaked lenses, and coarse green muscovite begins to break down to fine-grained, white muscovite; fine-grained biotite also develops; feldspars become cracked both parallel to and perpendicular to the foliation with quartz filling the fractures;
4) biotite forms foliated laminae through the rock which are wrapped around feldspar grains; feldspars become more rounded; pressure solution deposits of quartz form adjacent to feldspars, forming small augen; and
5) total rounding of feldspars takes place; biotite becomes very fine-grained and layers of biotite form continuous, millimeter-scale bands through the rock; magnetite and sphene appear intergrown with the biotite; chlorite develops on the biotite.

This general sequence mainly applies to the muscovite-stable, equigranular phases of Wilderness granite above the "muscovite-in" boundary in Figure 5-23.
Figure 5-24. Selected major element oxides and trace elements plotted as a function of structural position in the Wilderness granite stacked sill complex, Santa Catalina Mountains, Arizona. Lower horizontal line represents the transition between muscovite- and biotite-bearing pegmatitic phases. Middle horizontal line represents the transition between biotite-bearing and muscovite-biotite-bearing equigranular phases. Upper horizontal line represents the transition between mylonitic and nonmylonitic structural levels of the complex. Equigranular phases are depicted by open symbols and pegmatitic phases are depicted by shaded symbols. Thin horizontal lines connect data points of equigranular and pegmatitic phases that were collected at the same locality.
Figure 5-24 (continued)

[Diagram showing various data points and lines indicating concentration levels in different samples.]
Mylonitically deformed Oracle Granite in the Santa Catalina forerange banded gneiss complex comprises the distinctive mesocratic augen gneisses (informally referred to as 'dark bands'). Where the coarse-grained augen gneiss is concordantly interlayered with the Wilderness leucocratic component, contacts are commonly marked by a zone of mylonite schist. Some of the leucocratic sills or light bands are completely encased in a sheath of mylonitic schist. Zones of mylonitic schist may vary from one to several meters in thickness; away from the contact, they progressively grade into coarse-grained augen gneiss. In thicker exposures of augen gneiss, there are comparatively non-rounded, rectangular 'box car' crystals of potassium feldspar; these are typical of undeformed Oracle Granite. Biotite is the only mica present in the dark component of the forerange banded gneiss. However, muscovite becomes common and locally is the dominant mica in areas of mylonitic Oracle Granite that are intruded by equigranular muscovite-bearing Wilderness phases. K-feldspar augen in the deformed Oracle Granite are enveloped by a fine-grained matrix of feldspar and quartz. Foliation surfaces are undulatory and contain aligned aggregates of biotite and recrystallized quartz. Elongate aggregates of quartz with sutured boundaries or with ribbon textures are common within the plane of foliation and are expressed as a rod-like lineation which resembles "hot" slickensides.

Geochemistry

Our detailed sampling in the Santa Catalina Mountains was designed to evaluate several unresolved geochemical questions. In the forerange banded gneiss complex, we were interested in how the geochemistry of the Oracle Granite was affected by mylonitization and/or lit-par-lit intrusion of Wilderness Granite sills. Our sampling within phases of the Wilderness Granite was meant to evaluate the geochemical effects of three processes: 1) magmatic differentiation; 2) partitioning of elements between co-existing equigranular granite and late stage, water-rich, pegmatitic phases; and 3) mylonitization. We also collected samples of metasedimentary rocks (see Appendix E) near the top of the Wilderness laccolith to examine how (or if) they were geochemically modified by the adjacent intrusion.

On Figure 5-24, selected geochemical data are plotted as a function of structural position in the Wilderness Granite stacked sill complex. Sample localities are shown in Figure 5-23. The data indicate that distinct geochemical changes occur vertically in the pluton. Rocks in lower structural levels have low contents of SiO₂, K₂O, and MnO, and high contents of CaO, FeO, Fe₂O₃, V, TiO₂, and Sr relative to rocks in upper structural levels. Intriguingly, variation patterns of Na₂O and K₂O are virtually mirror
Figure 5-25. Gains and losses for selected major element oxides and trace elements for undeformed (nonmylonitic) samples of Oracle Granite versus mylonitic counterparts in the Santa Catalina forerange, Arizona.
images of one another; that is, parts of the pluton which are relatively sodium-rich are potassium-poor, and vice-versa. Uranium contents show a tendency to increase upwards towards the top of the pluton. These variation patterns are interpreted by us to reflect progressive differentiation of the Wilderness intrusion.

Comparison of analyses of co-existing equigranular granite and coarse-grained pegmatites indicate how different elements were partitioned between the magma (which ultimately crystallized to form equigranular granite) and its water-saturated phases (which solidified into aplite and pegmatite). Pegmatic phases at any given structural level in the Wilderness pluton are consistently enriched in Mn and Nb, and depleted in Si and K as compared to co-existing equigranular counterparts. Conversely, equigranular phases are consistently enriched in Ti, V, Li, Ba, Sr, Mg, Ca, Ce, Zn, and Fe relative to cospatial pegmatites. Sodium contents reveal no systematic partitioning between equigranular and pegmatitic phases. Uranium is strongly enriched in the pegmatites relative to equigranular phases; this affirms uranium's affinity for hydrous fluids. In contrast, thorium is relatively enriched in the equigranular phases.

Geochemistry of rocks in the forerange banded gneiss complex reflects the combined results of plutonism, metamorphism, contact metasomatism, and mylonitization. In order to evaluate the relative effects of these processes, we have plotted geochemical analyses of light-colored (Wilderness Granite) components of the forerange banded gneiss within the overall differentiation framework of the entire Wilderness stacked sill complex (Figure 5-24). When compared to the overall differentiation trend, Wilderness components of the forerange banded gneiss commonly exhibit a distinct departure. In particular, Wilderness phases of the banded gneiss are strongly depleted in Na, Ti, Zn, V, Sr, Ca, Fe, Mg, and Al relative to the overall differentiation trend. These phases are strongly enriched in SiO₂ and K₂O relative to the predicted differentiation trend. No systematic depletions or enrichments are exhibited by analyses of Cu, Ba, Y, and importantly, U or Th. Dark components of the forerange banded gneiss (mylonitic Oracle Granite) are markedly depleted in K₂O and enriched in Na₂O compared to nonmylonitic control data from outside of the Santa Catalina forerange (See Figure 5-25). In contrast, contents of uranium, thorium and most other elements are similar for samples of
nonmylonitic and mylonitic Oracle Granite; these data indicate that only the alkali elements (Na and K) were sufficiently mobile during mylonitization to undergo detectable geochemical changes (See Figure 5-25).

Mylonitic samples of Wilderness Granite from the main range laccolithic sill are not markedly displaced from the inferred, overall differentiation trend (See Figure 5-24). This documents that mylonitization of the main range Wilderness laccolithic sill did not result in any major geochemical changes.

Geochemical Comparison of Mylonitic Rocks From the Tortolita and Santa Catalina Complexes

It is instructive to compare the geochemical changes that accompanied mylonitization in the Tortolita Mountains against those in the Santa Catalina Mountains. Table 5-6 summarizes the gains and losses undergone by rocks during mylonitization in both mountain ranges. In general, mylonitization was accompanied by gains of SiO₂ and losses of Li, Fe, P₂O₅, V, Mn, and possibly Ti. Contents of U, Th, Rb, Ba, Cr, and Y were unchanged by mylonitization. However, the interrelationship between potassium and sodium contents is provocative. In the northeast Tortolita Mountains, loss of potassium in mylonitic Oracle Granite is accompanied by probable potassium gain in the adjacent recrystallized Pinal Schist band. Gain of sodium in the muscovite-bearing, mylonitic Oracle Granite is matched by sodium loss in the adjacent recrystallized Pinal Schist. Similarly, in the Santa Catalina forerange, gains of sodium and losses of potassium in mylonitic Oracle Granite are complemented respectively by sodium loss and potassium gain in the leucocratic injection sheets of Wilderness Granite.

Conclusions and Speculations

Based on detailed sampling of mylonitic and metamorphic rocks in the Tortolita and Santa Catalina complexes, we offer some general conclusions regarding how mylonitization and metamorphism has affected protolith geochemistry. Geochemical data from both complexes suggests that intensive mylonitization of various interlayered lithologies was accompanied by the addition of silica. This may indicate that mylonitization occurred in the presence of a silica-rich fluid. Mylonitization evidently resulted in depletion of Fe, Ti, P₂O₅, V, and Mn. Perhaps the silica rich fluid scavanged the deforming rock for these elements. The data may indicate that Na and K were exchanged between adjacent lithologies during
**TABLE 5-6:**  
COMPARISON OF GAINS AND LOSSES IN MYLONITIC ROCKS  
FROM THE SANTA CATALINA AND TORTOLITA COMPLEXES

<table>
<thead>
<tr>
<th>Element or Recrystallized Element Oxide</th>
<th>Chlorite</th>
<th>Muscovite</th>
<th>Chlorite</th>
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<tbody>
<tr>
<td></td>
<td>Pinal Schist</td>
<td>Oracle</td>
<td>Granite</td>
<td>Oracle</td>
<td>Granite</td>
<td>Granite</td>
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<tr>
<td>SiO₂</td>
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<td>+++¹</td>
<td>++</td>
<td>+++¹</td>
<td>0²</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>0</td>
</tr>
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<td>K₂O</td>
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<td>0</td>
<td>--</td>
<td>0</td>
<td>++</td>
<td>0</td>
</tr>
<tr>
<td>P₂O₅</td>
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<td>0</td>
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<td>0</td>
</tr>
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<td>Ba</td>
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</tr>
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<td>Cu</td>
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<td>0</td>
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<td>0</td>
</tr>
<tr>
<td>Rb</td>
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<td>0</td>
<td>0</td>
<td>0</td>
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<td>0</td>
</tr>
<tr>
<td>Sr</td>
<td>0</td>
<td>0</td>
<td>++</td>
<td>0</td>
<td>--</td>
<td>0</td>
</tr>
</tbody>
</table>

**Note:** ¹ Symbols in column are defined as follows: +++ = large gain; ++ = gain; + = slight gain; 0 = no change; -- = slight loss; --- = loss; and ---- = large loss.
intense mylonitization.

In contrast, areas where a single rock type experienced only moderate to weak mylonitization, the data reveal no systematic geochemical changes due to mylonitization. Thus, in these areas (such as the main range sill of Wilderness Granite in the Santa Catalina main-range), mylonitization was fundamentally isochemical. Local ultramylonite bands in the Wilderness pluton, however, contain numerous prominent bands of ribbon quartz. Thus, zones of strong mylonitization in homogeneous granitic rocks may have undergone silica addition and iron, phosphorous, vanadium, and manganese depletion. At present, we have no chemical data to confirm this speculation.

In all types, uranium and thorium contents were not affected by mylonitization, possibly because both elements are largely tied up in refractory accessory minerals that resist mylonitization. A partial test of this hypothesis is provided by comparison of fluorometric and neutron-activation uranium analyses for some moderately to highly uraniferous rocks (Table 5-7). The table indicates that uranium values determined by neutron activation are consistently higher than those determined by fluorimetry. This is probably due to the fact that some uranium was not taken into solution during the acid-leach phase of fluorimetry. In effect, fluorimetric analyses may reflect only the loosely held or acid-leachable uranium; it is this uranium that would most likely be redistributed by mylonitization. Uranium which is detected by neutron activation but not fluorimetry may largely reside in refractory accessory minerals; this uranium would be relatively immobile during mylonitization. The high ratio of neutron activation/fluorimetric analyses for many of the Tortolita and Santa Catalina samples may indicate that most uranium is indeed tied up in refractory accessory minerals. Therefore, mylonitization in the Santa Catalina and Tortolita complexes had little effect on uranium contents. However, for increased credibility, this preliminary conclusion must await testing by detailed sampling of mylonitic and nonmylonitic phases of highly uraniferous protoliths.
### Table 5-7: Comparison of Uranium Contents in Selected Rock Types from the Santa Catalina and Tortolita Complexes by Analytical Method

<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Sample Number</th>
<th>Uranium By Neutron Activation (ppm)</th>
<th>Uranium By Fluorometric (ppm)</th>
</tr>
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<tr>
<td><strong>Eocene-Oligocene alkali-calcic suite</strong></td>
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<td>Catalina quartz monzonite</td>
<td>155882</td>
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<td>Tortolita quartz monzonite</td>
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<td>0.76</td>
</tr>
<tr>
<td>Tortolita quartz monzonite</td>
<td>155885</td>
<td>2.7</td>
<td>0.86</td>
</tr>
<tr>
<td>Aplite in Tortolita quartz</td>
<td>155886</td>
<td>6.7</td>
<td>3.5</td>
</tr>
<tr>
<td>Reef Granite</td>
<td>155857</td>
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<td>0.33</td>
</tr>
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<td>Reef Granite</td>
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<td>0.25</td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>Derrio Canyon granite</td>
<td>155912</td>
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<td>0.46</td>
</tr>
<tr>
<td>Derrio Canyon granite</td>
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<tr>
<td>Wilderness granite sill complex</td>
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<td>Seven Falls granitic gneiss</td>
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<td>Biotite granite (lower main-range sill)</td>
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</tr>
<tr>
<td>2-mica granite (upper main-range sill)</td>
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<td>Garnet schlieren in Lemmon</td>
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<td>28.50</td>
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<tr>
<td>Rock leucogranite</td>
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<td></td>
<td></td>
</tr>
<tr>
<td><strong>Mid-Cretaceous to late Cretaceous calc-alkalic suite</strong></td>
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<td>Leatherwood quartz diorite</td>
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<tr>
<td>(mylonitic)</td>
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<tr>
<td><strong>1400 m.y. granitoids</strong></td>
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<td><strong>1700 m.y. Pinal Schist</strong></td>
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<tr>
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<td>Pinal Schist (modified)</td>
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GEOCHEMISTRY OF DISLOCATION ZONES

Introduction

At the outermost margins of many metamorphic core complexes is a zone of brecciation, jointing, and faulting that ranges in thickness from ten to nearly a hundred meters. The zone of highly fractured and brecciated rocks grades progressively downward into structurally intact (non-brecciated) rocks that comprise the crystalline core of the complex. The zone is capped by a one to two meter thick ledge of microbreccia. The top of the microbreccia is a polished and slickensided, low-angle fault surface. Above the fault or dislocation surface rest a wide variety of non-mylonitic Precambrian through mid-Miocene rocks.

The zone of brecciation is inevitably accompanied by extensive retrograde metamorphism and/or hydrothermal alteration and mineralization. Extensive chloritization is the most common mineralogical consequence and lends the highly fractured rock a punky green appearance that has prompted the frequent use of the term 'chloritic breccia.' Less commonly, hydrothermal fluids have deposited hematite, pyrite, chalcopyrite, chrysocolla, gold, and locally uranium in the dislocation zone. The passage of hydrothermal or "retrogressive" fluids through structurally disrupted and therefore permeable dislocation zones must result in some geochemical changes. The widespread occurrence of iron and copper mineralization along the dislocation zones of west-central Arizona attests to the potential economic significance of such geochemical changes.

Field relationships and timing constraints indicate that the chloritic breccia phenomena postdate much, if not all, mylonitization. In Arizona complexes, the formation of the chloritic breccia may have occurred from several to over 40 m.y. later than mylonitization. Also, some chloritic breccias clearly overlie non-mylonitic crystalline basement as, for example, the chloritic breccias in the western Whipple Mountains of southeast California described by Davis and other (1980).
It is also becoming apparent that the chloritic breccia phenomena is almost exclusively associated in space and time with mid-Tertiary low-angle normal fault phenomena. These faults have commonly rotated large areas of mid-Tertiary volcanic and sedimentary sections on low-angle normal faults that merge downward into the regional low-angle surface that contains the chlorite breccia and microbreccia phenomena discussed above. This overall process of low-angle listric faulting, wholesale block rotation, chloritic brecciation, and micro-brecciation we refer to as dislocation. We refer to the regional low-angle fault surface formed or reactivated during this period of faulting as a dislocation surface. We restrict the term dislocation to movements associated only with the early to mid-Tertiary age low-angle normal fault motions and not to any earlier pre-Tertiary movements that may have occurred on these low-angle surfaces.

Geochemistry

There is only minor pertinent data available on the geochemistry of dislocation zones. Petrographic analysis of samples collected on traverses from biotite-stable mylonitic rocks into chloritic breccia and microbreccia document pronounced mineralogical changes (Reynolds and Rehrig, 1980; Reynolds and Sanders in Davis, 1978), but do not reveal whether the bulk geochemistry of the rocks was changed by formation of the chloritic breccia. Geochemical samples of chloritic breccia collected from the Buckskin Mountains in west-central Arizona during this project reveal that significant geochemical changes have occurred.

Buckskin Mountains Traverse

In the northeastern Buckskin Mountains, near Alamo Dam, there are excellent exposures of the Whipple-Buckskin-Rawhide Mountain dislocation surface and underlying chloritic breccia and mylonitic gneiss (Figure 5-26). Here, the dislocation surface is offset by a northwest-striking, northeast dipping reverse fault. Highly deformed and metamorphosed Paleozoic rocks are present in small fault slivers along the dislocation surface.
EXPLANATION

- Chloritic Breccia
- Deformed and metamorphosed Paleozoic sedimentary rocks
- Tilted Mid-Miocene sedimentary rocks
- Whipple-Buckskin-Rawhide dislocation surface

Figure 5-26. Diagramatic cross-section of dislocation zone geology in the northeastern Buckskin Mountains west-central Arizona
We sampled various lithogies in the underlying Buckskin mylonitic complex and compared their geochemistry with mylonitic rocks that were affected by the chloritic breccia and dislocation surface (Figure 5-27). We also sampled some Paleozoic calc-silicate sedimentary rocks to obtain additional perspective. Compared to the underlying mylonites, samples of the chloritic breccia, dislocation surface, and Paleozoic calc-silicate rocks are relatively enriched in V, Zn, Li, Cr, Ni, Zr, U, Th, and Cu. In particular, the uranium contents of the chloritic breccia, dislocation surface, and Paleozoic calc-silicate rocks are respectively 2, 3.5, and 12 times those of underlying mylonitic rocks. If sample No. 153809 is omitted, Ba, P, Mn, Fe, Mg, Y, Ce, Sc, and T are also depleted in the mylonitic basement compared the chloritic breccia, dislocation surface, and calc-silicate samples. The samples in the dislocation zone are strongly depleted in sodium and probably depleted in aluminum and strontium. No systematic pattern of relative change was observed for potassium. Thus, in the Buckskin Mountains (at Alamo Dam) we conclude that the dislocation process is a mineralizing process; U, Th, Cu, Zn, Fe, V, Li, and Cr, have evidently been introduced into the dislocation zone.

Limited major element data on the microbreccia of the regional Whipple-Buckskin-Rawhide dislocation zone are available (T.L. Heidrick and J. Wilkins, 1980, personal communication). The data indicate that the major element composition of the flinty microbreccia is probably unrelated to lithologic variations in upper plate crystalline rocks above the dislocation surface. Relative to enclosing rocks, formation of the microbreccia was accompanied by minor losses of Al₂O₃ and Na₂O with corresponding gains in K₂O and total iron contents. Heidrick and Wilkins believe the overall composition of the microbreccia largely reflects the calc-alkalic composition (sensu lato) of presumed original calc-alkalic parent. The gains and losses found by Heidrick and Wilkins agree with most of our findings (see above). We suggest that relatively uniform composition of the microbreccia may alternatively reflect compositional and tectonic homogenization of a once more variable crystalline basement terrane. The final development of this chemically homogeneous microbreccia would have occurred during mid-Miocene dislocation. We interpret gains and losses reported above as a reflection of hydrothermal circulation within the dislocation zone during its formation in mid-Miocene time.
Radioactivity

We have completed several scintillometer traverses from mylonitic basement terrane rocks across overlying chloritic breccias and dislocational surfaces. Scintillometer data from the Rincon, Buckskin, and Picacho complexes of Arizona indicate that total counts in mylonitic basement are approximately equal to those from chloritic breccias and microbreccias within the dislocation zones. The data, however, do not permit us to evaluate whether uranium has been significantly remobilized by the dislocational process. Our limited geochemical data in the Buckskin Mountains of west-central Arizona however, suggest that the uranium fraction of the scintillometer data may be significantly different in the dislocational zones as compared to their underlying crystalline basement. Thus, scintillometry alone cannot document differences in uranium distribution between the crystalline basement and the overlying dislocation zones; therefore, scintillometry across any of these transitions from crystalline basement to dislocation zone must be accompanied by uranium geochemistry.

Conclusions and Speculations

Our geochemical data in the Buckskin complex is limited but may have significant implications for uranium distribution in core complex regions because it suggests that in at least one area, the dislocation process is a favorable uranium concentrating process. It is important that elements that are present in anomalous amounts (U, Cu, Fe, and Zn) occur in economic quantities along the dislocation surface elsewhere in the Buckskin, Rawhide and Whipple Mountains of west-central Arizona, and southeasternmost California. For example, copper and gold have been extracted from low-dipping, tabular bodies along the dislocation surface at the Copper Penney, Planet, and Swansea Mines in the Buckskin Mountains and at Copper Basin in the Whipple Mountains. Anomalous concentrations of uranium in the Buckskin-Rawhide-Whipple dislocation surface are present in the Fools Peak and Red Hills areas of the Rawhide Mountains (see Appendix D).

Interestingly, of the complexes we sampled, elements comparatively depleted in the underlying mylonites (Fe, Mg, P, Mn, Li, V, Ti, Zn, Cu, Ni, and especially U and Th) were all found enriched in the chloritic breccia and closely associated
Figure 5-27. Gains and losses for selected major and minor elements in mylonitic basement overprinted by dislocation zone processes. Structural position of the rocks sampled increases to the right of the figure. Sample points located to the right of the solid black line represent rocks that occur within the dislocation zone.
Figure 5-27 (Continued)
dislocation surface. This pattern may reflect the effect of a hydrothermal fluid that scavanged the above mentioned elements from the underlying mylonitic and crystalline rocks in the underlying basement and then redeposited them in anomalous amounts at or near the dislocation surface. Generation of these hydrothermal fluids may have been triggered by thermal contrasts that existed during rapid uplift of the region in mid-Miocene time. In this concept, hydrothermal fluids are self-generated by convection that occurred in response to thermal contrasts induced by rapid relative uplift in mid-Miocene time (see Chapter 4). Rapid uplift would also cause the widespread resetting of radiometric K-Ar and fission-track ages in crystalline basement underneath the dislocation zone.

Alternatively, the heat source for the hydrothermal fluids may have been mid-Miocene alkali-calcic magmatism that was widespread throughout the region. The magmatism may have generated metalliferous fluids that scavanged other elements from the crystalline basement as they circulated towards the structurally inviting dislocation zone. The regional metallogenesis copper-gold (and minor uranium) character of the mineral deposits is consistent with an association with alkali-calcic and alkalic magmatism elsewhere (for example, mid-Miocene mineralization in the Vulture, Kofa, Oatman, and El Dorado Districts of western Arizona and southernmost Nevada).

In this view the widespread resetting of radiometric K-Ar and fission-track ages in crystalline basement beneath the dislocation zone results from the introduction and circulation of these hot fluids through the crystalline basement beneath the dislocation zone. Perhaps, generation of the hydrothermal fluids was a combination of the above mentioned processes. In any case, we are convinced that the dislocation process in the Whipple-Rawhide-Buckskin Mountains of west-central Arizona and southeasternmost California was associated with regional circulation of hydrothermal fluids that produced a regional scale chloritic alteration and local metallization in areas at or near the regional dislocation surface. The presence of similar highly chloritized breccias in other core-complex areas suggests that the dislocation process and associated hydrothermal fluid flow was a major Cordilleran-wide, Tertiary process. Our limited data in one area suggest that this process is capable of mobilizing and redistributing uranium in potentially economic amounts. This process might be particularly effective in areas characterized by a uraniferous crystalline basement (such as the Kettle complex of Washington). Clearly, more research into the dislocation process and its affect on uranium distribution may have profound economic implications.
CHAPTER 6

URANIUM FAVORABILITY OF CORDILLERAN METAMORPHIC CORE COMPLEXES: A SUMMARY

By

Stephen J. Reynolds

INTRODUCTION

Cordilleran metamorphic core complexes have been recently recognized as unique centers of plutonism, metamorphism, and deformation. The uranium favorability of the complexes has not been previously described in the literature. The favorability of the complexes for various types of uranium occurrences is discussed in Chapter 4. Geochemical parameters indicative of uranium favorability are presented in Chapter 5. The specific uranium favorability of each complex is outlined in Appendix D. The present Chapter briefly summarizes the more detailed conclusions contained within Chapters 4 and 5 and Appendix D, and recommends future studies that will further clarify the uranium favorability of Cordilleran metamorphic core complexes.

URANIUM FAVORABILITY

Regional Considerations

As recognized by many geologists, the most favorable areas for the discovery of uranium deposits are generally situated in uranium provinces that contain numerous known occurrences or are underlain by rocks which exhibit high background abundances of uranium. Cordilleran metamorphic core complexes are distributed over such a large and geologically diverse area that regional considerations, such as those utilized in defining uranium provinces, provide a useful first-order indication of favorability of the complexes. Studies on the regional variations in radioelement content of plutonic rocks of the western United States
Figure 6-1. Density of uranium occurrences per unit area for the western United States. Diagram is modified from Gabelman (1976). Letters indicate the approximate positions of the following metamorphic core complexes: O - Okanogan, K - Kettle, S - Selkirk, B - Bitterroot, P - Pioneer, A - Albion, RR - Raft River, G - Grouse Creek, R - Ruby, KN - Kern, SN - Snake, CH - Chemehuevi, W - Whipple, R - Rawhide, BK - Buckskin, H - Harcuvar-Harquahala, WS - White Tank-South Mountains, PC - Picacho, SR - Santa Catalina-Rincon-Tortolita, PS - Pinaleno-Santa Teresa, PR - Papago Indian Reservation area (Sierra Blanca, Comobabi, and Coyote complexes), and PV - Pozo Verde. The letter D indicates the location of the Death Valley region.
(Larsen and Gottfreid, 1961; Marjaniemi and Robbins, 1972; Swanberg and Blackwell, 1973; Munroe and others, 1975; Castor and others, 1977) are extremely pertinent since plutons in the crystalline cores of the complexes might be expected to conform with established regional patterns. Results of these studies and sampling during this project indicate that compared with the rest of the core complex belt, the Kettle, Selkirk, Albion-Raft River, and Ruby complexes are located in areas that have plutons with relatively high radioelement contents. Data on the uranium abundances of Precambrian basement of the western U.S. (Malan and Sterling, 1969) can be similarly used as an indication of regional favorability.

Uranium provinces are characterized by a high concentration of uranium occurrences as indicated on Figure 6-1 for the Colorado Plateau, Wyoming basins, and the Colorado Front Range. This figure also documents that the Kettle, Selkirk, Albion-Raft River, western Arizona, and Santa Catalina complexes are located in areas of relatively abundant uranium occurrences as compared with the remainder of the core complex belt. In order to further evaluate the distribution of uranium occurrences both within and adjacent to the core complexes, we have compiled and plotted on million-scale maps the locations of all uranium occurrences (See Appendix C and accompanying maps). These maps provide a visual comparison of the density of uranium occurrences between individual complexes and between core complexes and their surrounding areas. In general, the core complexes have approximately the same density of uranium occurrences as the rest of the Cordillera. Exceptions are the Kettle and Selkirk complexes which have an exceptionally large number of occurrences both within and adjacent to their crystalline cores. Some areas such as the Albion, Rawhide and Santa Catalina complexes have an intermediate number of occurrences, while others including the South Mountains, White Tank, Picacho, and Comobabi complexes of Arizona contain few or none. This variation in the number of occurrences per complex is probably the best over-all favorability index. Figure 6-2 is derived from the aforementioned maps and emphasizes the density of uranium occurrences per one degree quadrilateral for the core complex belt. As revealed before, certain areas of the belt such as northeastern Washington are characterized by a profusion of occurrences. These areas are clearly the most favorable for possessing significant uranium deposits and reserves.

Occurrence of Uranium in Cordilleran Metamorphic Core Complexes

As discussed in Chapter 4, numerous processes were operative during the prolonged geologic evolution of Cordilleran metamorphic core complexes. These processes could have conceivably produced a wide variety of uranium occurrences. However, the core complexes are most favorable for those types of occurrences which
DENSITY OF URANIUM OCCURRENCES PER ONE DEGREE QUAD IN THE METAMORPHIC CORE COMPLEX BELT
1) are actually known to be present in the complexes; or 2) are not documented in the complexes, but can be reasonably postulated to exist based on the concepts outlined in Chapter 4. Listed below are ten uranium occurrences from the core complex belt that are significant either because of past production or because they are illustrative of a potentially important type of deposit.

1) Kettle Pegmatites: These granitic pegmatites are interlayered with amphibolite-grade metamorphic rocks and are characterized by high uranium contents. Some of the pegmatites may owe their origin to in-situ anatexis of uraniferous sedimentary rocks, whereas others are clearly cross-cutting and have been introduced from depth. Biotitic metamorphic rocks adjacent to the pegmatites are locally mineralized, possibly by contact metasomatism.

2) Graeber Lease, Kettle complex: High-grade uranium mineralization occurs in biotitic metamorphics that exhibit a mylonitic fabric. The deposit is located on the northeastern flank of the complex and may have been formed by a combination of metamorphic and mylonitic processes.

3) Mount Spokane, Selkirk complex: Spectacular meta-autunite coats fractures in weathered, but otherwise unaltered muscovite-bearing alaskite and pegmatite. Mineralization was strongly controlled by several types of structures and was probably deposited by ground waters which had leached uranium from the granitic rocks.

4) Tiger Formation, adjacent to Selkirk complex: Uranium mineralization is associated with carbonaceous material in Tertiary sedimentary rocks. The sedimentary rocks are in the upper plate of the Newport fault (dislocation surface) but are topographically lower than some uraniferous granites of the lower plate. The source of the uranium might be the plutons or syn-sedimentation volcanism.

5) Core of the Albion complex: Uranium is present in association with pegmatites of the Tertiary Almo muscovite granite and as veins in older Precambrian basement rocks. The ultimate origin of the uranium is uncertain for both types of occurrences.

6) Goose Creek area, west of the Albion complex: Mineralization occurs in late Tertiary carbonaceous sediments. Anomalously uraniferous tuffs are reported by Mapel and Hail (1969) who suggest that the ash is the source of the uranium. Nevertheless, similar favorable lithologies such as lignites and carbonaceous shales might be present down the hydraulic gradient from uraniferous rocks exposed in the core of the Albion Range.
7) Ruby granite and marshes: This project has documented the existence of a uraniferous biotite granite phase of what has been collectively mapped as Jurassic granite (Howard and others, 1979). Any uranium leached from the granite could be trapped in adjacent sediment-filled basins which locally contain modern day playas and marshes.

8) Rawhide dislocation surface: Several uranium occurrences in this western Arizona complex are localized along the dislocation surface between lower-plate mylonitic gneisses and upper-plate Paleozoic, Mesozoic, and middle Cenozoic sedimentary and volcanic rocks. Uranium is accompanied by a distinct copper and iron mineralization which has a known regional association with the dislocation surface. The mineralization must be middle or late Tertiary and may have been deposited concurrent with movement on the dislocation surface.

9) Anderson Mine - Artillery Mountains - Black Butte area, west-central Arizona: These three occurrence areas have much in common although separated by significant distances. Uranium mineralization is present in lacustrine units of middle Tertiary age that lie up-section (northeast) of exposures of the dislocation surface. Uranium mineralization may be largely the result of alteration of volcanic ash, but the present orientation of bedding in all three areas (a pronounced southwest dip) is probably a function of tilting that accompanied movement on the dislocation surface.

10) Blue Rock Mine area, eastern Santa Catalina complex: Uranium is present in fractures and along faults that separate a variety of sedimentary, metamorphic and plutonic rocks. It is uncertain whether the fault system is a dislocation surface similar to those exposed on the western margin of the complex or Laramide low-angle faults with large amounts of displacement. Uranium mineralization may owe its origin to groundwater leaching of adjacent granitic or overlying uraniferous sedimentary rocks. Alternatively, it may have been formed by hydrothermal fluids which permeated the fault system during or after its movement.

The ten areas described above are exemplary of the various types of uranium occurrences present within and adjacent to metamorphic core complexes and are depicted in an idealized cross-section of a metamorphic core complex in Figure 6-3. The types of occurrences represented by these examples will probably be the most common in the complexes. It follows that Cordilleran metamorphic core complexes will be most favorable for additional deposits that are similar to those outlined above.
Figure 6-3. Schematic cross-section of a typical metamorphic core complex showing the geologic setting of important uranium occurrences. Circled numbers indicate the positions of the following uranium deposits discussed in this chapter: 1) Kettle Pegmatites; 2) Graeber Lease, Kettle complex; 3) Mount Spokane, Selkirk complex; 4) Tiger Formation, adjacent to Selkirk complex; 5) Core of the Albion complex; 6) Goose Creek area, west of the Albion complex; 7) Ruby granite and marshes; 8) Rawhide dislocation surface; 9) Anderson Mine - Artillery Mountains - Black Butte area, west-central Arizona; and 10) Blue Rock mine area, eastern Santa Catalina complex.
Conclusions

The uranium favorability of metamorphic core complexes is as varied as their geologic settings and tectonic histories. A majority of the complexes are very unfavorable for uranium deposits. On the opposite extreme of the favorability spectrum are complexes such as the Kettle Mountains which are as favorable for uranium deposits as any crystalline areas of the western U.S. All the complexes are grouped below according to their relative uranium favorability.

High favorability - Kettle, Selkirk, Albion

Moderate to low favorability - Okanogan, Bitterroot, Raft River, Ruby, Rawhide, Buckskin, Santa Catalina-Rincon, Pinaleno-Santa Teresa

Very low favorability - Pioneer, Grouse Creek, Snake, Kern, Death Valley, Homer, Dead, Sacramento, Chemehuevi, Whipple, Harcuvar, Harquahala, White Tank, South Mountains, Picacho, Tortolita, Sierra Blanca, Comobabi, Coyote, Alvarez, Kup, Pozo Verde.

Each complex will have different favorability for the various types of uranium occurrences discussed in Chapter 4 and earlier in this section. Therefore, the geology of each complex must be evaluated on an individual basis to predict the types of uranium occurrences most likely to be present. Cordilleran metamorphic core complexes, as a group, are probably most favorable for pegmatitic, hydrothermal, metamorphic, authigenic, allogenic, and peripheral lacustrine occurrences. The dislocation zone on the flanks of the core complexes is conceivably an important permeable channel or depositional site for ascending or descending uraniferous fluids. The uranium potential of such zones may not have been properly evaluated by traditional exploration in the complexes. Uranium mineralization in such zones might be largely concentrated below the water table, out of view of surface observation.

CRITERIA FOR EVALUATION OF FAVORABILITY

As might be expected, the criteria for evaluation of uranium favorability vary from complex to complex. Numerous criteria are dispersed throughout the text of this chapter as well as chapters 4 and 5. The most favorable complexes will be those that are situated within uraniferous provinces and contain abundant occurrences or a record of significant past production. Delineation of regional zones of uraniferous plutons or mineral deposits might help reveal favorable areas.
Within each complex, favorability criteria are governed by the geology of that specific complex. The presence of highly uraniferous plutonic rocks suggests potential for a variety of plutonic uranium deposits and provides a uranium source for subsequent redistribution by metamorphic, mylonitic, hydrothermal, and near-surface processes. High uranium content will commonly be matched by high potassium and thorium enabling the detection of such uraniferous granites by standard ground or airborne radiometric techniques. As discussed more fully in Chapters 4 and 5, the chemistry of igneous rocks is a very powerful indicator of uranium favorability. High alkalinity of the magma series is as important as a thorough degree of differentiation. Igneous rocks with high $K_2O$ and $K_2O + Na_2O$, in conjunction with low CaO, MgO and FeO contents, are those most likely to be uraniferous. Uranium-rich igneous rocks may also have high abundances of trace elements such as fluorine, lithium, beryllium, tin, tungsten and molybdenum. Models for the behavior of uranium and these elements in igneous environments provide other important favorability criteria (see Pilcher, 1978; Mathews, 1978b; Files, 1978).

Metamorphic rocks are most favorable for uranium deposits if they exhibit high background uranium contents and radioactivity. Schist, gneiss, and migmatite that are rich in biotite, muscovite, garnet, and sulfides are the typical host-rocks of metamorphic uranium deposits. Structures within metamorphic and plutonic rocks will commonly be accompanied by chlorite, hematite and mineralization such as copper, iron, and gold.

Favorability criteria for sedimentary uranium deposits will not be reviewed here, but are outlined in numerous Bendix open-file reports such as Marjaniemi and Robbins (1975), Wopat and others (1977) and Mickle and Mathews (1978). Metamorphic core complexes may be favorable for peripheral sedimentary uranium deposits if their plutonic and metamorphic rocks contain abundant leachable uranium. Zones of brecciated and permeable rocks that flank the complexes would be efficient channels for distributing uranium leached from the complexes into adjacent receptive sedimentary lithologies.

RECOMMENDATIONS FOR FUTURE STUDY

There is much information that, if available, would provide insight into additional aspects of the uranium favorability of Cordilleran metamorphic core complexes. More extensive sampling of the plutonic and metamorphic rocks of each complex will provide more data on background uranium abundances, thereby more specifically verifying the favorability of complexes not sampled during this project.
There is clearly much to be learned regarding uranium deposits produced by metamorphic and mylonitic processes. Detailed geochemical sampling programs in conjunction with petrologic study would better document the behavior of uranium during progressive metamorphism and mylonitization. Such sampling must be done in rocks that exhibit high background uranium contents, such as those present in the Kettle complex. The Kettle complex would also be ideal for examining the effect on uranium and other elements by processes that accompany formation of the dislocation surface. Understanding the origin of copper, iron, gold, and uranium mineralization associated with the dislocation surfaces, such as that in western Arizona, is critical in evaluating the general favorability of all core complexes.

The flanks of the complexes should be evaluated for subsurface uranium deposits either in the dislocation zone or in nearby upper-plate sedimentary units. This could be accomplished only by drilling adjacent to uraniferous core rocks such as those present in the Kettle, Selkirk, and Albion complexes.

Finally, the relationship, if any, between classic mantled gneiss domes and Cordilleran metamorphic core complexes must be examined in more detail. Some mantled gneiss domes including those near Rossing and the Rum Jungle regions are very favorable for uranium deposits. A large number of mantled gneiss domes are located in uranium provinces or areas otherwise characterized by abundant uranium and large-ion-lithophile element mineralization. Several uraniferous gneiss domes should be inspected to document their similarities and differences with Cordilleran metamorphic core complexes. Such a comparison might reveal why gneiss domes are generally favorable for uranium while metamorphic core complexes, as a tectonic species, are not.
National Uranium Resource Evaluation

CORDILLERAN METAMORPHIC CORE COMPLEXES 
AND THEIR URANIUM FAVORABILITY 

MAPS 

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with contributions by 
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November 1980

PREPARED FOR U.S. DEPARTMENT OF ENERGY 
Assistant Secretary for Resource Applications 
Grand Junction Office, Colorado
This report is a result of work performed by the Laboratory of Geotectonics, Department of Geosciences, University of Arizona, through a Bendix Field Engineering Corporation Subcontract, as part of the National Uranium Resource Evaluation. NURE is a program of the U.S. Department of Energy’s Grand Junction, Colorado, Office to acquire and compile geologic and other information with which to assess the magnitude and distribution of uranium resources and to determine areas favorable for the occurrence of uranium in the United States.

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Map C-1
URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT:
~ARIZONA~
Map C2

URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT: CALIFORNIA
Map C-3

URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT: IDAHO
Map C-4
URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT:
~ MONTANA ~
Map C-5  URANIUM OCCURRENCE IN
THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT:
~NEVADA~
URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT: ~UTAH~
Map C-7

URANIUM OCCURRENCE IN THE CORDILLERAN METAMORPHIC CORE COMPLEX BELT:

WASHINGTON
TECTONIC MAP OF THE BITTERROOT METAMORPHIC CORE COMPLEX, IDAHO AND MONTANA

Map D2

EXPLANATION

MAP UNITS:

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<th>Code</th>
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<td>Early Tertiary (Eocene) volcanic rocks</td>
</tr>
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<td>T2</td>
<td>Early Tertiary (Eocene) granitic rocks</td>
</tr>
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<td>Late Cretaceous-Early Tertiary mylonitic rocks</td>
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</tr>
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<td>PMS</td>
<td>Precambrian sedimentary rocks (P&amp;M where metamorphosed)</td>
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<tr>
<td>PEM</td>
<td>Precambrian metamorphic rocks unaltered</td>
</tr>
</tbody>
</table>

SYMBOLS:

- Contact
- Fault, dashed where approximately located
- Dislocation surface, dashed where approximately located or covered
- Area of major fault
- Overturned syncline
- Overturned anticline
- Strike and dip of foliation showing trend of brecciation
- Uranium occurrences

Scale 1:250,000

Kilometers
TECTONIC MAP OF THE PIONEER METAMORPHIC CORE COMPLEX, IDAHO

Map D-3
EXPLANATION

MAP UNITS

- TQR: Late Cenozoic Quaternary surficial deposits
- b: tectonic breccia
- Tsv: Tertiary sedimentary and volcanic rocks
- Tg: Tertiary granitic rocks (igneous where micasite bearing)
- Jg: Jurassic granitic rocks (igneous where micasite bearing)
- Pz: Paleozoic sedimentary rocks (fmed where metamorphosed, fits where converted to marble, not shown)
- ZCC: Late Precambrian-Cambrian clastic rocks (26cm where metamorphosed)

SYMBOLS

- Contact, dashed where approximately located
- Fault
- Low-angle normal fault
- Dislocation surface, dashed where covered or approximately located
- Tie, leader
- Uranium occurrence

SCALE 1:250,000

TECTONIC MAP OF THE SNAKE AND KERN METAMORPHIC CORE COMPLEXES, NEVADA AND UTAH

Map D-6
TECTONIC MAP OF THE DEAD, SACRAMENTO, AND CHEMENUEVI METAMORPHIC CORE COMPLEXES, CALIFORNIA

Map D-7
TECTORIC MAP OF THE SOUTH MOUNTAIN AND WHITE TANK METAMORPHIC CORE COMPLEXES, ARIZONA

Map D-9

EXPLANATION

MAP UNITS

<table>
<thead>
<tr>
<th>Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>TQs</td>
<td>Late Tertiary-Quaternary surficial deposits</td>
</tr>
<tr>
<td>Tv</td>
<td>Tertiary volcanic rocks</td>
</tr>
<tr>
<td>Tc</td>
<td>Tertiary chloritic breccia</td>
</tr>
<tr>
<td>Tg</td>
<td>Tertiary granitic rocks *</td>
</tr>
<tr>
<td>MzTg</td>
<td>Mesozoic-Tertiary granitic rocks *</td>
</tr>
<tr>
<td>PCg</td>
<td>Precambrian granitic rocks *</td>
</tr>
<tr>
<td>PCm</td>
<td>Precambrian metamorphic rocks *</td>
</tr>
<tr>
<td>*</td>
<td>Asterisk is added to end of symbol where rock is mylonitic</td>
</tr>
</tbody>
</table>

SYMBOLS

- Major foliation arch
- Strike and dip of foliation with trend of lineation
- Dike

Scale: 1:250,000
TECTONIC MAP OF THE SOUTH MOUNTAIN AND WHITE TANK METAMORPHIC CORE COMPLEXES, ARIZONA

Map D-9

EXPLANATION

MAP UNITS

- **TQs**: Late Tertiary-Quaternary surficial deposits
- **Tv**: Tertiary volcanic rocks
- **Tc**: Tertiary chloritic breccia
- **Tg**: Tertiary granitic rocks *
- **MrTg**: Mesozoic-Tertiary granitic rocks *
- **PcG**: Precambrian granitic rocks *
- **Pcm**: Precambrian metamorphic rocks *

* Asterisk is added to end of symbol where rock is mylonitic

SYMBOLS

- **major foliation arch**
- **strike and dip of foliation with trend of lineation**
- **dike**

SCALE 1:250,000

5 10 15 20 KILOMETERS
Map D-II

TECTONIC MAP OF THE POZO VERDE, COYOTE, COMOBABI, AND SIERRA BLANCA METAMORPHIC CORE COMPLEXES, ARIZONA
EXPLANATION

MAP UNITS:

<table>
<thead>
<tr>
<th>Code</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>TQa</td>
<td>Tertiary-Quaternary surficial deposits</td>
</tr>
<tr>
<td>Tuv</td>
<td>Upper Miocene through Pliocene sedimentary and volcanic rocks</td>
</tr>
<tr>
<td>Tuv</td>
<td>Oligocene through Pliocene sedimentary and volcanic rocks</td>
</tr>
<tr>
<td>Tp</td>
<td>Middle Tertiary granitic rocks</td>
</tr>
<tr>
<td>Jmd</td>
<td>Jurassic diorite of Black Mountains</td>
</tr>
<tr>
<td>Meq</td>
<td>Lower Jurassic through Cretaceous granitic rocks</td>
</tr>
<tr>
<td>Meq</td>
<td>Triassic and Jurassic volcanic and sedimentary rocks</td>
</tr>
<tr>
<td>Fm</td>
<td>Middle Cambrian through Famian carbonate and clastics</td>
</tr>
<tr>
<td>Zb</td>
<td>Latest Precambrian through middle Cambrian clastics</td>
</tr>
<tr>
<td>Phm</td>
<td>Palnump Group and Cambrian rocks (metamorphosed)</td>
</tr>
<tr>
<td>Tp</td>
<td>Palnump Group</td>
</tr>
<tr>
<td>pcb</td>
<td>Precambrian igneous and metamorphic basement</td>
</tr>
</tbody>
</table>

SYMBOLS:

- Contact
- Fault
- Low-angle normal fault
- Dislocation surface
- Tie line
- Breccia

SCALE: 1:250,000

TECTONIC MAP OF THE DEATH VALLEY REGION, CALIFORNIA
Map D-12