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STRATIGRAPHIC FRAMEWORK, VOLCANIC—PLUTONIC EVOLUTION, AND VERTICAL DEFORMATION OF THE PROTEROZOIC VOLCANIC BELTS OF CENTRAL ARIZONA

by
Phillip Anderson
Precambrian Research and Exploration Inc.
810 Owens Lane, Payson, Arizona 85541

ABSTRACT

This four-part paper summarizes more than a decade of intensive research into the stratigraphy, plutonism, and deformation of central Arizona's Proterozoic volcanic belts, and redefines tectonic evolution of Proterozoic volcanic belts.

PART 1 — Early Proterozoic stratigraphy and volcanic evolution of the Prescott-Jerome volcanic belts

The Prescott-Jerome volcanic belt evolved from 1800 to 1740 Ma in six major formative depositional stages and two younger successor stages, all separated by intervolcanic unconformities of regional extent and low angular discordance. Volcanic strata comprise complexly interlensed, wedge-shaped depositional units laterally interrelated by facies changes and separated from younger and older units by unconformities. Contact relations of the major rock sequences defy analysis using former concepts of a single, vertically stacked stratigraphic column.

The Prescott volcanic belt was built up in three major volcanic cycles. The *first cycle* (depositional stages 1 and 2) began on mafic basement in the west with primitive, deep-submarine, low-K, Mg-tholeiitic, bimodal basalt flow sequences and coeval gabbros. The axis of volcanism then shifted east in two major jumps, each jump initiating a new, spatially separate volcanic cycle that backfilled intervening troughs and lapped over strata of earlier deposits.

The *second volcanic cycle* (depositional stages 3 and 4) produced a thick, submarine, bimodal, Fe-rich tholeiitic basalt-rhyolite pile in the center of the belt. The quartz-tholeiite source magmas underwent extensive fractionation by crystal separation to evolve a high-K tholeiitic andesite-rhyolite suite, which was rapidly succeeded at the close of the second volcanic cycle by dacite pyroclastics that lapped unconformably back over strata of the first cycle.

The *third volcanic cycle* (depositional stages 5 and 6) extruded huge volumes of altered calc-alkaline felsic fragmentals and minor andesite from two main edifices in the eastern part of the belt. After the last major felsic outpouring, ensuing clastic deposits became successively less volcanic and more sedimentary, but submaturity was attained only in the later Texas Gulch and Mazatzal (7 and 8) stages, after plutons and batholiths invaded the volcanic belt at about 1740 Ma.

The Jerome volcanic belt evolved similarly, but from east to west and on a smaller scale. The earliest low-K tholeiitic mafic volcanics of the **Ash Creek Group,** coeval with cycle 1 of the Prescott belt, were followed by evolved high-K tholeiitic felsic magmas, then by youngest Grapevine Gulch calc-alkaline pyroclastics that were finally overlapped by pyroclastics from cycles 2 and 3 of the Prescott belt. Grapevine Gulch strata extend west of the Shylock zone without major offset, which indicates that the Shylock is not a wrench fault.

Volcanic stratigraphy of the Prescott belt is newly subdivided into three major rock groups with unique lithostratigraphic, petrologic, and chemical attributes. These three new groups, proposed here for formal adoption, exactly reflect the three major volcanic cycles through which the Prescott volcanic belt evolved: **the Bradshaw Mountains Group** represents the first, westerly cycle, the **Mayer Group** the second, central cycle, and the **Black Canyon Creek Group** the third, easterly cycle. The name Ash Creek Group is retained for the Jerome belt.

The Bradshaw Mountains, Mayer, Black Canyon Creek, and Ash Creek Groups, together with other 1800- to 1740-Ma volcanic groups elsewhere in north-central Arizona, make up the **Yavapai Supergroup,** a new term with precisely defined stratigraphic limits, proposed to replace the former "Yavapai Series."

Systematic increase in alkali content of volcanic rocks southeast across the volcanic belts implies magma generation from a subduction zone dipping southeast under the Prescott-Jerome arc prior to 1750 Ma.

PART 2 — Early Proterozoic stratigraphy and volcanic evolution of the New River—Cave Creek—Mazatzal Mountains—Diamond Butte volcanic belts

The younger central Arizona volcanic belts in the New River, Cave Creek, Mazatzal Mountains, and Diamond Butte areas complement the older Prescott-Jerome belts because they began their evolution when the older belts ceased formation 1740 Ma ago, and continued on from the last volcanic stages of the older belts towards greater stratigraphic, structural, petrologic, and geochemical maturity.

The younger belts began as isolated submarine mafic centers of polymodal basaltic andesite-andesite-dacite-rhyolite from pyroxene-plagioclase-phyric, quartz-normative magmas. The Union Hills and Mount Ord centers produced low-K calc-alkaline basaltic andesite flows, while the Cramm Mountain and East Verde River centers produced calc-alkaline andesite pyroclastics. Rhyolitic flows, fragmentals, and tuffs were derived from all centers. Andesite fragmentals, tuffs, and graywackes were shed outward from the centers into intervening deep submarine basins in sequential patterns of interleaved lateral-facies aprons. Formative volcanic rocks of the younger belts comprise five major volcanic formations, which together compose the **Union Hills Group.**

After Union Hills Group volcanism, the submarine volcanic chains were reworked, and volcaniclastic detritus was shed into intervolcanic basins. Deposition of the first purple shale-siltstone-graywackes transgressed westerly along the longitudinal axis of the belt, and progressively more felsic volcanic debris was added to the basin by volcanism derived from primary felsic magmas. These deposits are assembled into the **Alder Group,** within which three main formations interrelated by facies changes are distinguished. Diorite-granodiorite plutons, fractionated from calc-alkaline, hydrous, primary mafic magmas related to Union Hills Group volcanism, concurrently intruded deeper parts of the volcanic pile, and only thin dikes reached the upper sedimentary strata of the Alder Group.

Alder Group clastic sedimentation was finally overwhelmed at 1710 Ma by felsic volcanism, as thick volcanic conglomerate deposits heralded the start of ignimbrite eruptions. The huge **Verde River Granite batholith** breached surface on its west side via transitional subvolcanic phases of the **New River Mountains Felsic Complex** to extrude ignimbrite valley flows in the New River belt, and on its east to form the Mount Peeley ignimbrites. Red Rock then Haigler ignimbrites extruded farther east, from magmas that also produced Payson granite. The ignimbrites have an alkali-calcic chemistry shared by no earlier deposits.

As the subaerial ignimbrite sheets were eroded back to sea level, fluvial, followed successively by littoral and shallow-water marine, environments reworked the felsic detritus into the distinctive red hematitic quartzite-conglomerate deposits of the Mazatzal Group. Three major formations of the Mazatzal Group are recognized in the **Mazatzal Mountains,** and Mazatzal exposures at Sheep Basin Mountain and elsewhere are defined as facies variants or formations signifying different depositional environments in the Mazatzal fluvial-marine setting.

Union Hills Group mafic volcanics are more alkali rich than formative volcanics of the Prescott-Jerome belts, and hence may represent an easternmost part of the magmatic arc formed by the early event of southeast-dipping subduction. The Alder Group and younger deposits, however, are not part of this early magmatic trend.

PART 3—Plutonic suites and metamorphism of the Prescott region

The first plutons in the Prescott region were (1) an earliest tholeiitic *synvolcanic gabbro* suite that was subvolcanic to mafic flows of the Bradshaw Mountains Group, lower Mayer Group, and Ash Creek Group and (2) an early calcic *intervolcanic gabbro-diorite* suite diapiric into the Bradshaw Mountains Group and Grapevine Gulch Formation. These earliest suites were consanguineous with formative volcanism of the Prescott-Jerome belts, and emplacement of the gabbro-diorites produced early intervolcanic unconformities by local uplift and submarine erosion.

Latter stages of formative volcanism at 1750 Ma heralded emplacement of the first *pretectonic I-type hornblende biotite granodiorite plutons* (Government Canyon, Brady Butte, Crooks Canyon) into the west side of the Prescott belt. Major plutonic stabilization and crustal thickening occurred from 1740 to 1720 Ma when *pretectonic hornblende-biotite tonalite and granodiorite batholiths* (the huge composite Cherry Springs batholith to the east; Wilhoit and Minnehaha granodiorites to the west) pervaded the perimeters of the volcanic belts.

Beginning with the first major tectonism of the Prescott-Jerome volcanic belts after 1720 Ma, *syntectonic biotite granodiorite and monzogranite plutons* (Prescott, Johnson Flat, Longfellow Ridge, Skull Valley) were emplaced at depth along the northwest side of the Prescott belt. A stress field controlled their emplacement, shape, internal foliation, and northeast-trending extensive dike swarms that cut up through linear weak zones in the volcanic pile (e.g., Johnson Flat-Mt. Elliott dike swarm in Chaparral zone). Huge monzogranite batholiths pervaded felsic crust northwest of the Prescott belt during this same interval.

Later tectonic stages were dominated by smaller *porphyritic monzogranite plutons* (Horse Mountain and Iron Springs) in the western Prescott belt, and larger porphyritic batholiths (Yarnell) to the southwest. In the final stages of tectonism at about 1700 Ma, high-grade regional metamorphism in the southern Prescott belt fused pelitic rocks throughout much of the southern Bradshaw Mountains. Anatectite was emplaced higher in the Prescott belt as the Crazy Basin granite, causing (1) intermediate-pressure, staurolite-sillimanite metamorphism that overprinted earlier low-pressure, andalusite-cordierite metamorphic facies to the west and (2) strong vertical shear in the Black Canyon belt and in host rocks to the north that overprinted steep fabrics of the earlier Prescott regional metamorphism and deformation.

Crustal evolution and origins of the plutonic suites can be tracked by systematic chemical variations. The calcic pretectonic pluton and batholith suite and the calc-alkaline syntectonic pluton suite show remarkably systematic alkali enrichment to the northwest, which implies generation from subduction dipping northwest under the Prescott-Jerome belts. Chemical evolution from the earliest gabbros to the youngest porphyritic monzogranites shows that Proterozoic crust in the Prescott region evolved progressively with emplacement of each plutonic suite. However, persistent chemical differences in elements such as Ca and Sr among members of each suite show that the composition of each pluton closely reflects the composition of its host (i.e., source) volcanic rocks: both plutonic and volcanic rocks to the northwest are richer in Sr than those to the southeast, regardless of age.

These chemical variations require a model of magma generation in which subduction heated the base of the crust in a series of migrating melting loci that progressed down dip of the subduction zone with time, adding a vapor-rich alkali fraction to partial melts derived mostly from fusion of the basal crust beneath the volcanoplutonic arc.

PART 4—Proterozoic vertical deformation and its tectonic significance

Deformation in the central Arizona Proterozoic volcanic belts was previously explained by *fold models,* including subhorizontal fold axes and "a" slip lineations, tilting of strata and steep folding, subhorizontal fold axes rotated to vertical by polyphase folding, and initial low-angle thrusts later deformed to vertical by steep folding. Such fold models fail to explain the main features of Proterozoic deformation in the central volcanic belts. Instead, the data indicate that vertical foliation and steep to vertical linear fabrics were imposed on volcanic rocks in a single deformational event that produced mainly moderate to steep minor folds, but no regional folds or major closures of stratigraphy. Structural evidence clearly indicates that the belts were subjected neither to regional polyphase folding nor to foreland thrusting prior to deformation.

Structural work in the Proterozoic volcanic belts has produced a new model that explains how structures become oriented to vertical in a single event of deformation. A *pure strain* approach shows that vertical foliation, steep lineations, and plunging minor folds are a natural consequence of imposing a horizontally constrictive stress regime on a relatively incompetent and poorly layered segment of crust (a volcanic belt). The greater density of the volcanic belt causes it to be deformed vertically into a downward-narrowing trough between surrounding rising plutonic masses. All features indicate that material moved vertically, so the tectonic regime is described as *Proterozoic vertical deformation.* It differed from other tectonic regimes because shortening occurred in all horizontal directions, not just one.

Weakly strained parts of the volcanic belts have weak foliation, no lineations, and few gently plunging minor folds, and stratigraphy is still basically subhorizontal. In moderately to strongly strained areas, lineations first appear in moderate to steep orientations where strain was sufficiently strong to form linear fabrics. Bedding is reoriented to steeper dips by minute translations on foliation planes. Minor folds, most abundant where bedding departs from regional foliation, plunge moderately to steeply, but major regional folds are absent. Thus stratigraphic units are not repeated east-west across the volcanic belts, but originally trended northeast, parallel to their depositional basins.

In strongly to extremely strained zones (high-strain zones), bedding, foliation, lineations and folds all approach vertical as all primary features are distorted to prolate (cigar) shapes by horizontal constriction and strong vertical extension. High-strain zones are integral parts of the primary Proterozoic deformation of volcanic belts; they exist on all scales and mark high strain gradients. The largest (Shylock zone, Black Canyon belt, Gun Creek zone, and others) define the axes of highest strain in each segment of a volcanic belt that was structurally discrete during deformation; they are the central "creases" or synclinal keels into which structures ultimately converge at depth.

Distorted primary features are strain gauges that show how foliation and lineation intensity and steepness reflect strain variations throughout the belts. Such variations show that strain was governed more by original heterogeneities in the volcanic pile than by crustal depth, and because original weakness zones were vertical, high-strain zones originated with steep structures before strain penetrated adjacent competent areas. The deformational makeup of the belts is thus a direct result of their primary stratigraphic makeup. Consequently, the Proterozoic volcanic belts were fated to respond to initial crustal shortening by vertical deformation, before subhorizontal tectonic regimes could develop.

INTRODUCTION

Ever since the turn of the century when geologists first ventured into the Bradshaw Mountains south of Prescott (Jaggar and Palache, 1905), the complex Proterozoic geology of central Arizona has remained elusive. More importantly in recent years, the need for a clear understanding of Proterozoic tectonic processes has reached the forefront of geologic research, because the radically different styles of Archean and modern plate tectonics cannot be reconciled without a much better knowledge of Proterozoic tectonics (P. Anderson, 1976).

The rocks that formed at continental margins during the Proterozoic—and thus by inference the processes that generated new Proterozoic crust—are the keys to understanding Proterozoic tectonics. Nowhere in the United States are rocks that built up a Proterozoic continental margin better exposed than in central Arizona (P. Anderson, 1986). Central Arizona was the site of a deep ocean basin 1800 m.y. ago. Rock packages that formed on the thin, unstable oceanic crust of that basin were subsequently deformed, metamorphosed, tectonically sliced, and finally sutured together, and are necessarily complex, even more so than those at young continental margins, because of metamorphism and deformation.

Some of the earliest pioneering geologic work in central Arizona had remarkable insight into its broad Proterozoic structure: Wilson (1939) laid the foundations of a general stratigraphic framework that is still viable today, and Gastil (1958) perceived the essentially correct order of evolution of the Proterozoic crust—felsic volcanics then sediments were built upon a basement of primitive mafic volcanics, not the opposite as is claimed for other Proterozoic regions. Work between 1948 and 1972 concentrated on the details of the Prescott and Payson areas, without benefit of a comprehensive stratigraphic framework for the entire state. Nevertheless, each work contributed valuable information to a largely unexplored geologic region, information without which we could not build the syntheses we can today.

In fact, many U.S. Geological Survey quadrangle mapping studies of the central Arizona Proterozoic were completed just prior to two major advances in geologic understanding: (1) the new perspective of lateral interlensed stratigraphy in volcanic piles, pioneered in the Archean of Canada and the Mesozoic-Cenozoic of Japan; and (2) the new global plate tectonic perspective which causes us to question the juxtaposition of rock packages and to incisively test field relationships in ways not thought of before. Clearly, those pre-1972 studies could not perceive the Proterozoic volcanic stratigraphy and structure of central Arizona in the same complexity and detail as we can today, and consequently, one must accept that more recent geologic studies and syntheses will, indeed should, substantially revise both the stratigraphic frameworks and concepts of the former generation.

In 1973, the author started a systematic, comprehensive examination of the Arizona Proterozoic to test whether Proterozoic tectonics closely resembled Archean tectonics, modern plate tectonics, or neither. Field work was concentrated largely in the Proterozoic volcanic belts of central Arizona, specifically between Prescott and Payson, where the most vital information on the generation of new Proterozoic crust was to be found. The result of more than a decade of detailed work was summarized in a Ph.D. dissertation (P. Anderson, 1986), and this paper is condensed still further from just a part of that work.

This paper centers on a single key outcome of the study—*the new regional stratigraphic framework for the central Arizona Proterozoic.* Since Wilson's (1939) pioneering effort, no one has succeeded in developing a detailed and comprehensive stratigraphic framework that is valid for all of central Arizona and that has been continuously linked and tested across the entire region. The new stratigraphy presented here accomplishes what Wilson first set out to do: link all Early and Middle Proterozoic stratified rocks in central Arizona into a unified, internally consistent stratigraphic framework.

Although very broad isotopic age limits exist for Early Proterozoic stratified rock sequences of central Arizona, precise numerical age constraints on key parts of the new stratigraphic framework are lacking, and several major formational groups have not yet been dated. This is because isotopic studies were restricted to places in the Payson and Prescott areas where there was a history of previous work. As the present study shows, those sites were not necessarily the best places to resolve key stratigraphic relations or to date key rock units. The regional stratigraphic framework presented in this paper will ultimately lead to a much more precise isotopic chronology for Early Proterozoic rocks in central Arizona.

This study deduced a relative chronology of events for the central Arizona volcanic belts, from their inception at 1800 Ma to their stabilization at 1650 Ma, that is more detailed than the published isotopic chronology. In essence, stratigraphic relations are now known more precisely than isotopic relations, because only a few points in the total stratigraphic column have been dated. However, based on relative ages of undated formations and groups to the few dated formations, it is possible to estimate numerical ages of many undated rock units to within 5 to 15 m.y., as is done in this paper.

The new stratigraphy outlined here results from original detailed and reconnaissance mapping of more than eighty-five 7.5′ quadrangles in the central mountain region of Arizona, where Early Proterozoic volcanic and volcaniclastic rocks are widely exposed over a 22,500-km^2 region from Phoenix north to Mingus Mountain (near Jerome), and from Skull Valley (west of Prescott) to Young (east of Payson). Early and Middle Proterozoic rocks crop out over about 50 percent of this region, about half of which are Proterozoic plutonic rocks and half are stratified rocks. Through detailed structural-stratigraphic mapping of original protoliths, this study uncovered more than 25 major unconformities within the stratified sequences that are of regional importance throughout the Proterozoic volcanic belts, only a few of which had been recognized previously. By detailed analysis of the relative-age relations of rock sequences bound by each unconformity, a regionally consistent stratigraphic framework for the central Arizona Proterozoic was developed.

This new stratigraphic framework is based on either adoption of, or revisions to, existing stratigraphic nomenclature for the Arizona Proterozoic. All new names introduced here have been cleared with the U.S. Geological Survey Geologic Names Committee, and consequently the new stratigraphic framework presented in this paper is proposed for formal adoption by the geologic community. This new stratigraphic picture clarifies the singularly most difficult aspect of understanding the Arizona Proterozoic—its complex stratigraphy and structure—and provides a logical, orderly perspective concerning how the Proterozoic crust of central Arizona evolved. The new understanding provided by this paper is a springboard to launch further detailed investigations into the complex Proterozoic geology of central Arizona.

GEOLOGIC OVERVIEW

All Proterozoic volcanic belts in central Arizona are collectively referred to as the "central volcanic belt," but figure 1 shows that this belt is readily divisible into an older part to the northwest and a younger part to the southeast. The older part is dominantly more mafic and the younger part is generally more felsic, even though both parts contain mafic and felsic rocks. This usage of older and younger is not to be confused with the terms "younger Precambrian of Arizona" for the Middle Proterozoic Apache Group (Shride, 1967) and "older Precambrian of Arizona" for Proterozoic rocks predating the Apache Group by a major orogeny (Wilson, 1939). All Proterozoic rocks described in this paper are between 1800 Ma and 1400 Ma in age and predate both the major orogeny and the Apache Group.

The central volcanic belt as a whole has great stratigraphic complexity, contains diachronous volcanic-rock packages that evolved at different times and places across the belt, was subjected to pervasive deformation and metamorphism of variable intensity at various times across the belt, and includes different plutonic suites emplaced as unique igneous events. It is impossible, therefore, to treat all aspects of the belt concurrently, so the early stratigraphic evolution of the belt [Parts 1 and 2][1] is described separately from the younger plutonic, metamorphic, and structural evolution [Parts 3 and 4]. Part 1 of this paper presents the volcanic stratigraphy of the Prescott-Jerome areas and describes the 1800- to 1740-Ma formative volcanic sequences of the older northwest part of the belt. Part 2 outlines the 1740- to 1660-Ma volcanic stratigraphy of the younger, eastern part of the belt, in the New River-Cave Creek-Mazatzal Mountains-Diamond Butte areas, and also describes plutonic rocks that are intimately associated with the younger, eastern volcanic sequences. In contrast, the long plutonic history of the northwest part of the belt represents major orogenic stabilization of the older crust, and is described in Part 3. The nature of Proterozoic deformation in the central volcanic belt and its diachronous history has been the subject of much misunderstanding in the past, so Part 4 is devoted to elucidating this important aspect of the belt. Finally, all parts considered together describe the generation of new protocontinental crust during an important time in the Earth's history—the Early Proterozoic.

[1]References to other parts of this paper or to the accompanying Proterozoic tectonics paper are listed as: [see Part 1] or [see tectonics paper].

PART 1—EARLY PROTEROZOIC STRATIGRAPHY AND VOLCANIC EVOLUTION OF THE PRESCOTT-JEROME VOLCANIC BELTS

Early Proterozoic volcanic stratigraphy of the Prescott-Jerome belts is formative to Arizona's tectonic evolution because it represents the oldest crust in central Arizona (P. Anderson, 1986) [see tectonics paper]. The earliest primitive strata contrast sharply to more evolved felsic strata in the younger belts to the southeast, which further underscores the need for a comprehensive stratigraphic framework to distinguish the two. Volcanic rocks of the older belt extend over a 5,000-km^2 region between Prescott, Jerome, Mayer, and the Black Canyon belt. Those in the Prescott-Jerome region are noted here; distal exposures in the Copperopolis, Crown King, Hieroglyphic, and other areas west and south of Prescott are described elsewhere (P. Anderson, 1986).

PREVIOUS WORK

Former work (C. A. Anderson and Creasey, 1958; C. A. Anderson and Blacet, 1972a,b,c; Krieger, 1965; C. A. Anderson and others, 1971) subdivided the complex volcanic stratigraphy of the Prescott-Jerome belts into broad map packages bounded either by faults or planes parallel to foliation. More recent studies (e.g., P. Anderson and Guilbert, 1979; DeWitt, 1979; P. Anderson, 1978a) indicate the need for major modifications to this stratigraphic nomenclature, because the "fault boundaries" of many rock units are actually unconformities (P. Anderson, 1986; O'Hara and others, 1978), and true stratigraphy is more commonly discordant to foliation than parallel to it.

Isotopic data from the Prescott-Jerome volcanic belt are relatively few: the Cleopatra unit near Jerome is dated at 1800 ± 15 Ma, and the Spud Mountain unit near Prescott is dated at 1755 ± 15 Ma (C. A. Anderson and others, 1971). All U-Pb zircon ages cited in this paper were recalculated using decay constants recommended by Steiger and Jaeger (1977). Younger 1640- to 1610-Ma Rb-Sr ages from similar rocks in the Prescott area (Lanphere, 1968) indicate disequilibration of Sr isotopes during deformation and metamorphism. The youngest rock sequence in the belt—the Texas Gulch Formation—is approximately 1720 Ma (Silver, 1976), which suggests an 80-m.y. maximum for stratigraphic evolution of the belt. Fifteen intervolcanic unconformities of regional significance have now been identified within this 80-m.y. time span (P. Anderson, 1986), which implies the need for a much more detailed stratigraphic resolution of the belt.

REDEFINITION OF YAVAPAI SUPERGROUP

C. A. Anderson and others (1971) and C. A. Anderson and Silver (1976) used a time-stratigraphic term "Yavapai Series" to refer to all pre-Texas Gulch rocks in the Prescott-Jerome volcanic belts deposited between 1800 and 1750 Ma. Recent stratigraphic work shows this to be incorrect and that an accurate time-stratigraphic subdivision is not yet possible, because neither the youngest nor oldest volcanic rocks predating the Texas Gulch unconformity have yet been dated. A 1750-Ma upper boundary would arbitrarily split several contiguous and totally unified rock-stratigraphic sequences in the Prescott-Jerome belts. More isotopic work is needed before an accurate "series" designation is possible.

The term "Yavapai" is best used in a rock-stratigraphic sense, because regional mapping now exists to define the stratigraphic and areal limits of all volcanic rock sequences in central and northern Arizona. C. A. Anderson (1968b) used "Yavapai Supergroup" in just this way to encompass all formative strata in the Prescott-Jerome belts, but the term was not widely adopted. Recent mapping (P. Anderson, 1986) shows that other volcanic rock sequences in adjacent regions (e.g., Bagdad and Antler-Valentine belts) are stratigraphically comparable to those in the Prescott-Jerome belts. *Therefore, it is now proposed that* **Yavapai Supergroup** *be formally adopted as the term to refer to all volcanic and related strata in the volcanic belts of central and northern Arizona* (see P. Anderson, 1986). This term, however, does not include the metasedimentary terranes in northwest Arizona correlative to Vishnu Schist in the Grand Canyon.

Within Yavapai Supergroup, volcanic rock formations in the Prescott belt are assembled into the **Bradshaw Mountains Group, Mayer Group,** and **Black Canyon Creek Group,** as outlined in this paper. Formerly all rocks in the Prescott volcanic belt were grouped as "Big Bug Group" (C. A. Anderson and Creasey, 1958), but the diversity of stratigraphically and petrochemically distinct rock suites in the belt is too great to be encompassed by a single rock group. Recent work also shows "Big Bug Group" formational designations to be incorrect or have invalid type sections. Hence "Big Bug Group" is no longer appropriate in the new stratigraphic framework of the Prescott belt and should be dropped.

In contrast "Ash Creek Group" (C. A. Anderson and Creasey, 1958) is a valid rock group in the Jerome belt and is retained because the stratigraphic sequence it encompasses is basically correct, with some minor modifications. Similarly, volcanic sequences in other parts of north-central Arizona are assigned to other rock groups under the Yavapai Supergroup [see tectonics paper].

TIME-SPACE STRATIGRAPHIC COMPLEXITIES

Prior attempts to stratigraphically organize the oldest rock sequences in central Arizona were largely unsuccessful because both primary stratigraphic complexities and the lensoidal nature of depositional units in deep-water submarine volcanic settings were not taken into account. Such complexities are well known in Archean and

Mesozoic volcanic belts (Hutchinson and others, 1971; Goodwin, 1965; Horikoshi, 1969; Ohmoto and Skinner, 1983), but the concepts have been only recently applied to the central Arizona Proterozoic volcanic belts (P. Anderson, 1977, 1978a,b; P. Anderson and Guilbert, 1979).

Previous studies of the central volcanic belt depict all rock units as superimposed in a single column, implying wide lateral persistence of volcanic strata across the belt. However, such an interpretation is incorrect, because most strata in volcanic piles are lensoidal and discontinuous, taper to wedge-shaped terminations, and interfinger laterally with adjacent rock units. The truly "formative" rock units—those that first built up the main body of the volcanic belts—are rarely depositionally superimposed on one another, as are younger successor deposits.

Figure 2 illustrates primary stratigraphic complexities across the Prescott volcanic belt from Prescott to Mayer. If stratigraphic interlensing is not perceived, fictitious isoclinal folds, faults, and other structures must be inferred in order to reconstruct the complex stratigraphy shown in figure 2. As well as containing such internal complexity, each major stratigraphic assemblage is laterally and longitudinally diachronous on a broad scale, so any transect across the belt will find the oldest units becoming progressively younger or older (P. Anderson, 1978a). Thus the major rock-stratigraphic groups are themselves laterally discontinuous and juxtaposed laterally across the belt, not vertically superimposed on one another. This has the following important consequences: (1) It is impossible to assemble a single stratigraphic column that is correct for the entire belt, because rock successions at each locality are different; (2) thickness estimates of the major rock units, when summed perpendicular to foliation, are incorrect because they measure the width of the volcanic pile, not its true thickness; and (3) conventional concepts of horizontally layered, vertically superimposed sequences cannot yield the true stratigraphy of the central volcanic belt.

DEPOSITIONAL STAGES AND THE NEW STRATIGRAPHIC FRAMEWORK

BASIS FOR REVISION

A major outcome of mapping the Prescott-Jerome volcanic belt is the understanding that it was built up by a series of temporally and spatially discrete igneous events with unique petrologic and geochemical characteristics. These events produced seven sequential depositional stages that are represented by time-stratigraphically unique, unconformity-bounded rock units, the order of which describes the evolution of the volcanic belt. The sequence could only be accurately deciphered by studying the entire Prescott-Jerome volcanic belt.

Previous work placed boundaries of major map units where major faults were thought to be, missed many critical unconformities, and established map units with lithologic-petrologic features identical to other map units. Where former work placed faults between map units, recent work finds stratigraphic continuity across zones of high strain, not faults. Major problems arose where type localities of earlier map units were determined by recent work to consist of rocks belonging to other formations of the previous workers (e.g., Iron King mine area). Thus, so as to maintain some stratigraphic validity to the former rock sequence, critical parts of it had to be revised, in preference to disbanding it entirely. The fewest revisions are in the Jerome area, where the **Ash Creek Group** defined by C. A. Anderson and Creasey (1958) remains largely intact; the most revisions are in the Prescott volcanic belt, where many previous map units were found to contain stratigraphic units of different ages.

MAJOR ROCK GROUPS

In the Jerome belt, C. A. Anderson and Creasey (1958) grouped all but the youngest of seven formations—the Texas Gulch Formation—into "Ash Creek Group." In the Prescott belt, C. A. Anderson and others (1971) recognized three major volcanic units: Green Gulch, Iron King, and Spud Mountain volcanics. New stratigraphic mapping shows that (1) "Green Gulch volcanics" includes rocks of the Grapevine Gulch, Spud Mountain, and other formations; (2) the "Iron King volcanics" type section is made up wholly of Spud Mountain rocks; (3) a "Spud Mountain" map unit east of the Shylock zone is totally unrelated to true Spud Mountain volcanism; (4) true stratigraphy has been traced right across former map-unit boundaries; and (5) many important rock units were missed by the former work. The main part of Spud Mountain volcanics is retained as a valid formation (P. Anderson, 1986), but most other stratigraphy has required major redefinition, as outlined here.

C. A. Anderson and Creasey (1958) grouped all rocks in the Prescott belt into the "Big Bug Group." But, because of extensive stratigraphic problems in it, summarized above, because only one of its formations survives, and because it does not describe the true stratigraphy of the belt, the term "Big Bug Group" is no longer viable and should be dropped from formal stratigraphic nomenclature. The diversity in space, time, and key chemical-petrologic attributes of rocks in the Prescott belt is much too great for a single rock group, and the belt is best described by several rock groups of contrasting parameters. Consequently, rock formations in the Prescott volcanic belt are newly assembled into three major, spatially distinct rock groups, proposed here (and P. Anderson, 1986) for formal adoption: (1) the **Bradshaw Mountains Group,** occupying the western part of the belt; (2) the **Mayer Group,** occupying the central region near the town of Mayer; and (3) the **Black Canyon Creek Group,** occupying the eastern area in the Shylock zone and Black Canyon belt. These three new rock groups replace the former, now obsolete, "Big Bug Group."

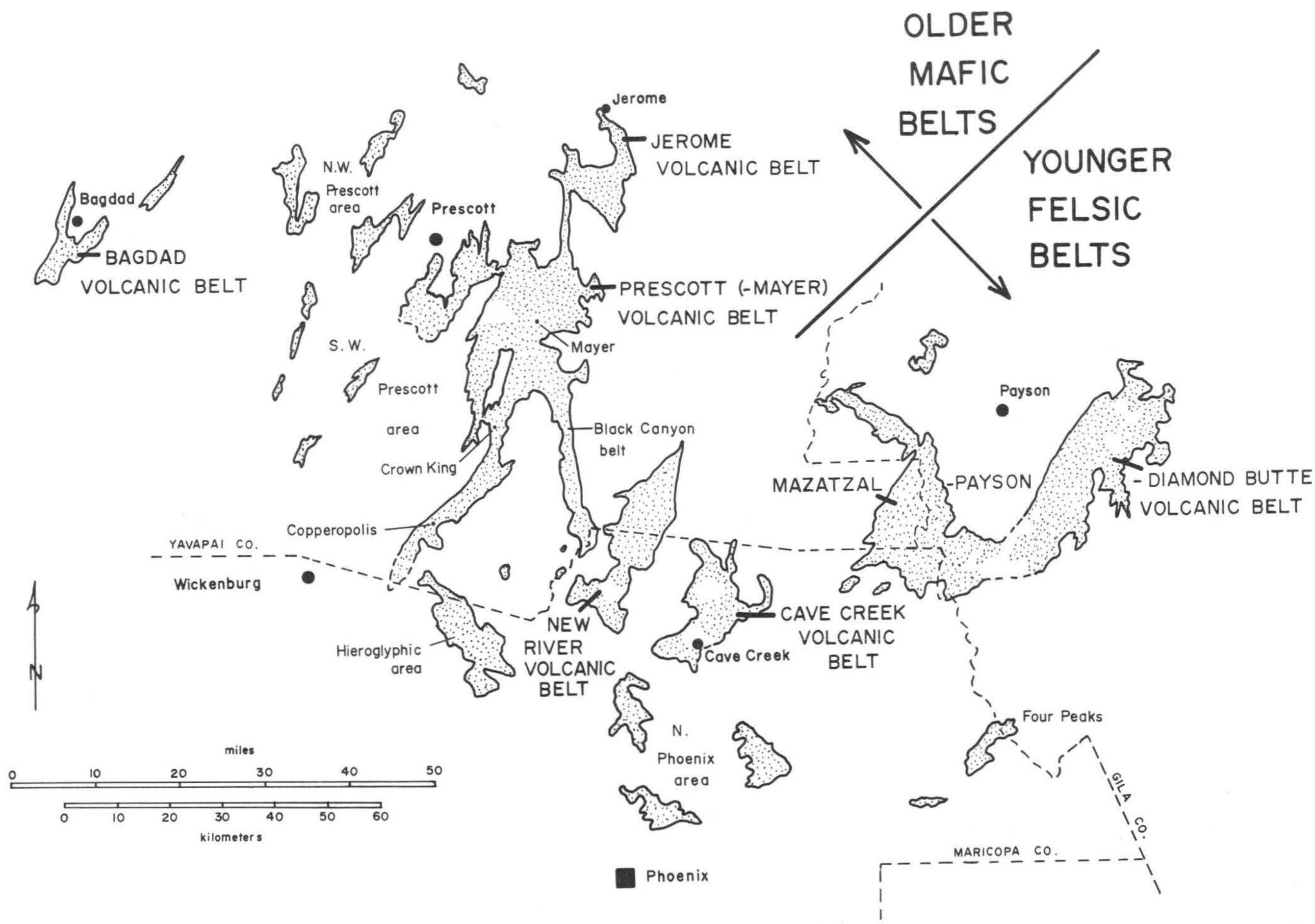


Figure 1. Location map of early Proterozoic volcanic belts in central Arizona in relation to towns and county boundaries. The older, more mafic belts to the northwest include (1) the large composite volcanic belt of the Prescott, Mayer, and Black Canyon Belt areas (collectively termed "the Prescott volcanic belt"), and the Jerome volcanic belt. Distal fragments of the Prescott belt extending northwest and southwest of Prescott and in the Copperopolis and Hieroglyphic areas are described elsewhere (P. Anderson, 1986), whereas the Bagdad belt is discussed elsewhere in this volume [see tectonics paper], along with other Proterozoic volcanic belts in northwest and southeast Arizona (not shown). The younger, more felsic volcanic belts to the southeast include exposures in the New River, Cave Creek, Mazatzal Mountains, Diamond Butte, and north Phoenix areas, as well as distal remnants to the east and southeast. These areas are stratigraphically related, but appear on the figure to comprise individual volcanic belts because they are separated by plutonic rocks and post-Precambrian cover.

DEPOSITIONAL STAGES

The Prescott volcanic belt has been divided into three rock groups not just for clarity, but because of several fundamentally important reasons: (1) Each group has a petrologic unity and distinctive geochemical signature that distinguishes it from adjacent rock groups, both in space and time; (2) from beginning to end, each group records a discrete stage in the evolution of the volcanic belt that is distinct from those of neighboring groups; and (3) there is a regular pattern to the evolution of each group, and how it laps over preceding groups, that repeats throughout the history of the belt. Thus, the most fundamental aspect of the volcanic belts is the *depositional stages* by which they evolved: these stages are closely mirrored by the three rock groups listed above, and they provide the clearest basis for understanding and describing the stratigraphic evolution of the volcanic belts.

GENERAL GEOLOGIC RELATIONSHIPS

Presented here are a new geologic map of the Prescott-Jerome volcanic belt and related exposures (fig. 3), identifying all major rock units, and a diagram (fig. 4) showing the new stratigraphic sequence for the Prescott-Jerome belt as independent stratigraphic columns. These columns cannot be superimposed in a single one, because units are areally restricted to different parts of the volcanic belts and were never laid down in a single succession (fig. 2).

The oldest rocks in the Prescott-Jerome belts are stratified mafic volcanics and hypabyssal equivalents. There is no evidence for existence of a crust significantly predating formation of the volcanic belts; felsic basement is specifically precluded because the oldest plutons postdate the formative volcanic sequences and contain no felsic inclusions. "Formative" volcanic sequences make up the oldest crust in any part of the belts, and include all rock

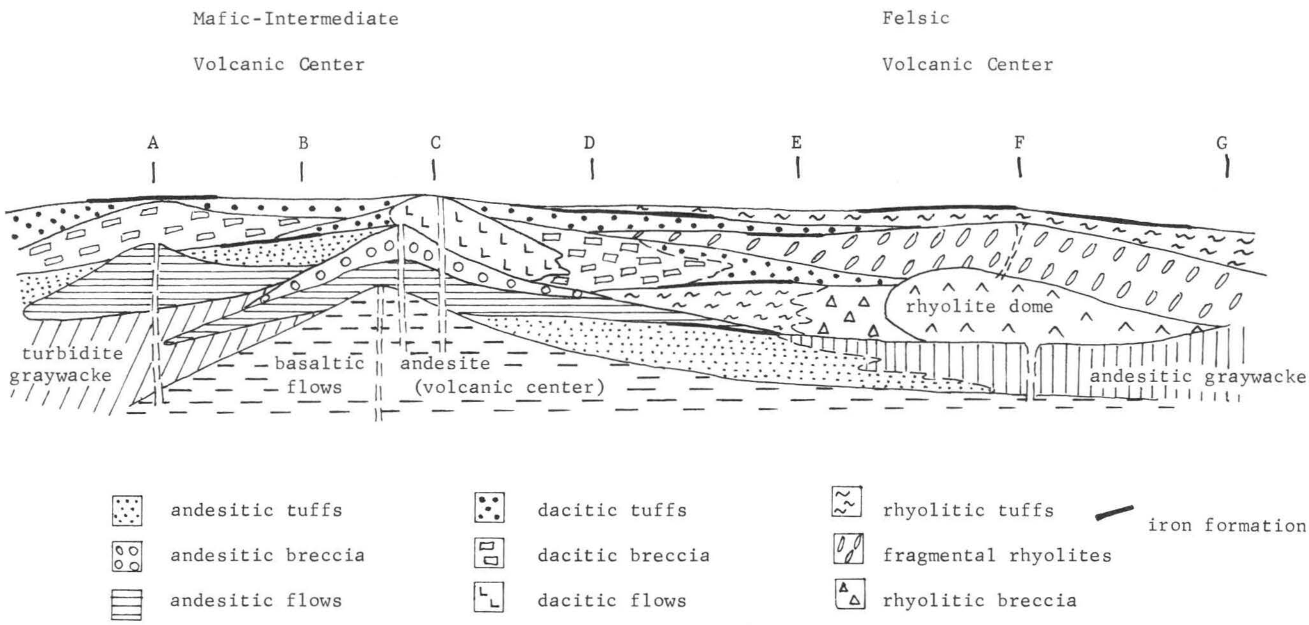


Figure 2. Diagrammatic representation of stratigraphic complexities across part of the Prescott volcanic belt used as a model for stratigraphic complexities mapped throughout the entire central volcanic belt (P. Anderson, 1986). The diagram shows a 20-km-long transect across the volcanic belt, with no vertical exaggeration intended. Letters A to G are reference points. Note how all depositional units are interrelated by facies changes and are laterally juxtaposed across the width of the belt. Also note the gentle unconformable relations between units and how they cause major stratigraphic onlap across the belt. Further note that, once the volcanic pile is tectonically sheared and facies relations are obscured, lithofacies packages such as the dacite flow-breccia-tuff suite in the middle of the diagram could be erroneously interpreted as detached isoclinal fold cores. Finally note that it is impossible to assemble the stratigraphy of the total volcanic pile into any single stratigraphic column that is representative of the entire belt.

units predating plutonism prior to Texas Gulch deposition (fig. 4). The oldest mafic volcanics of the Jerome belt occur only on the east side of Mingus Mountain, whereas the oldest mafic volcanics in the Prescott belt lie only on its western side and are absent to the east. Thus, the central Mayer area has younger strata of both belts and contains key relationships that stratigraphically link the volcanic histories of the two belts prior to plutonism.

STRATIGRAPHY OF THE JEROME VOLCANIC BELT

OLDEST MAFIC UNITS

The oldest rocks of the Jerome belt are tholeiitic basalt flows and mafic pillow units, including hypabyssal equivalents, with no exposed base (fig. 3). Basalt flows are laterally gradational into basaltic breccias, and are conformably overlain by plagioclase-rich tholeiitic andesite, dacite, and rhyodacite flows, breccias, and tuffs. Unconformably overlying the basic rocks are plagioclase-rich sodic quartz-feldspar-crystal rhyodacitic flows, breccias, and crystal tuffs. These oldest basalts, dacites, and rhyodacites are pervaded by a hypabyssal diabase complex only recently recognized as younger than the basaltic strata. The diabase grades up into thick basaltic andesite dikes that change northward into andesite-dacite towards a volcanic vent area. Intrusive contacts below the vent contrast to extrusive textures above in like rocks.

C. A. Anderson and Creasey (1958) did not recognize the stratigraphic age and singularity of the oldest mafic volcanic suite, assigning it to different map units depending on location. The southern part is in their "Gaddes basalt" and the northern part in their "Shea basalt." They believed the Gaddes to be the oldest unit in the belt, but in it they also included intermediate to felsic volcanic agglomerates and conglomerates south of Mingus Mountain that are some of the youngest rocks in the belt. Moreover, their Shea basalt includes not only mafic flows, but also the younger diabase complex that intrudes the flows.

The oldest tholeiitic basalt flow sequence is now renamed the **Shea Formation** (figs. 3 and 4) and is redefined as a basalt, andesite, and rhyodacite flow, tuff, and breccia sequence not including diabase. The younger hypabyssal diabase-basaltic andesite differentiated intrusive complex is newly named the **Silver Spring Gulch Diabase** for its type section in Silver Spring Gulch below the Shea mine; it intrudes the Shea Formation and is the hypabyssal feeder to the next volcanic event. Because C. A. Anderson and Creasey's (1958) "Gaddes basalt" includes both very old and very young rock units, it is an *assemblage* rather than a formation, and remains so on figure 4 until future work resolves it into distinct formations. The lower part of the new Shea Formation, shown on figure 3, includes the oldest mafic flow units that were once mapped as part of the Gaddes assemblage, as shown on figure 4.

MIDDLE DACITIC AND RHYODACITIC UNITS

After deposition of the Shea Formation, Silver Spring Gulch Diabase intruded as a sill-dike complex that fed to surface to form a large mafic to intermediate vent center. Basaltic andesite dikes change in composition and texture northward toward the vent, grading into feldspar-phenocrystic dacite-andesite. Beneath the vent, intrusive textures and structures predominate in a subvolcanic dacite complex, and above the vent are extrusive features in a plagioclase-phenocrystic dacite flow, breccia, and tuff-breccia sequence. Rapid change from dacite to rhyodacite up section records quick evolution to intermediate and felsic compositions. Continued felsic volcanism deposited thick rhyodacitic boulder, cobble, and pebble agglomerates in lensoidal units and caused chloritic alteration of the felsic strata. The appearance of quartz phenocrysts occurs in uppermost agglomerates, after which volcanism produced quartz-feldspar-phyric, welded rhyodacite ash flows. At the top of the ash flows near Jerome, hematitic rhyodacitic domes, iron formation, and the largest Proterozoic massive-sulfide deposit in Arizona were formed.

C. A. Anderson and Creasey (1958) included the dacites described above in "Shea basalt," distinguishing neither their separate stratigraphic position nor younger age. They termed the rhyodacitic agglomerates "Deception rhyolite" and mapped what they thought to be andesite beds in their rhyolite. They named the overlying quartz-feldspar-crystal ash-flow tuffs "Cleopatra quartz porphyry," believing the welded ash flows to be mainly a felsic intrusive body.

The vent complex of andesite-dacite flows, breccia, tuff, and feeders are now recognized as a new formation, stratigraphically and temporally distinct from the Shea, which records an important stage in evolution of the Jerome belt. The dacite-andesite suite is newly named the **Del Monte Formation** for its type section in Del Monte Gulch south of Mescal Gulch; its hypabyssal feeders cut the Shea Formation, and its extrusive facies overlie Shea rhyodacitic tuffs and conformably underlie dacitic breccia at the base of the Deception Formation.

"Deception rhyolite" is renamed **Deception Formation** and is redefined as a distinctive suite of lensoidally interstratified rhyodacite-dacite boulder agglomerate and tuff. Deception Formation has neither the rhyolite nor andesite formerly described; the rocks are rhyodacitic, and chlorite alteration zones that cut the agglomerate strata were mistaken for andesite. "Cleopatra quartz porphyry" is renamed **Cleopatra Formation** and is now known to be a lithostratigraphically distinct suite of phenocrystic agglomerates and quartz-crystal ash-flow tuffs, whose upper chloritically altered strata host the bedded United Verde massive-sulfide deposit at Jerome. Quartz porphyry feeder dikes are present in the Cleopatra Formation, but are minor or absent at the type locality.

UPPER VOLCANICLASTIC UNITS

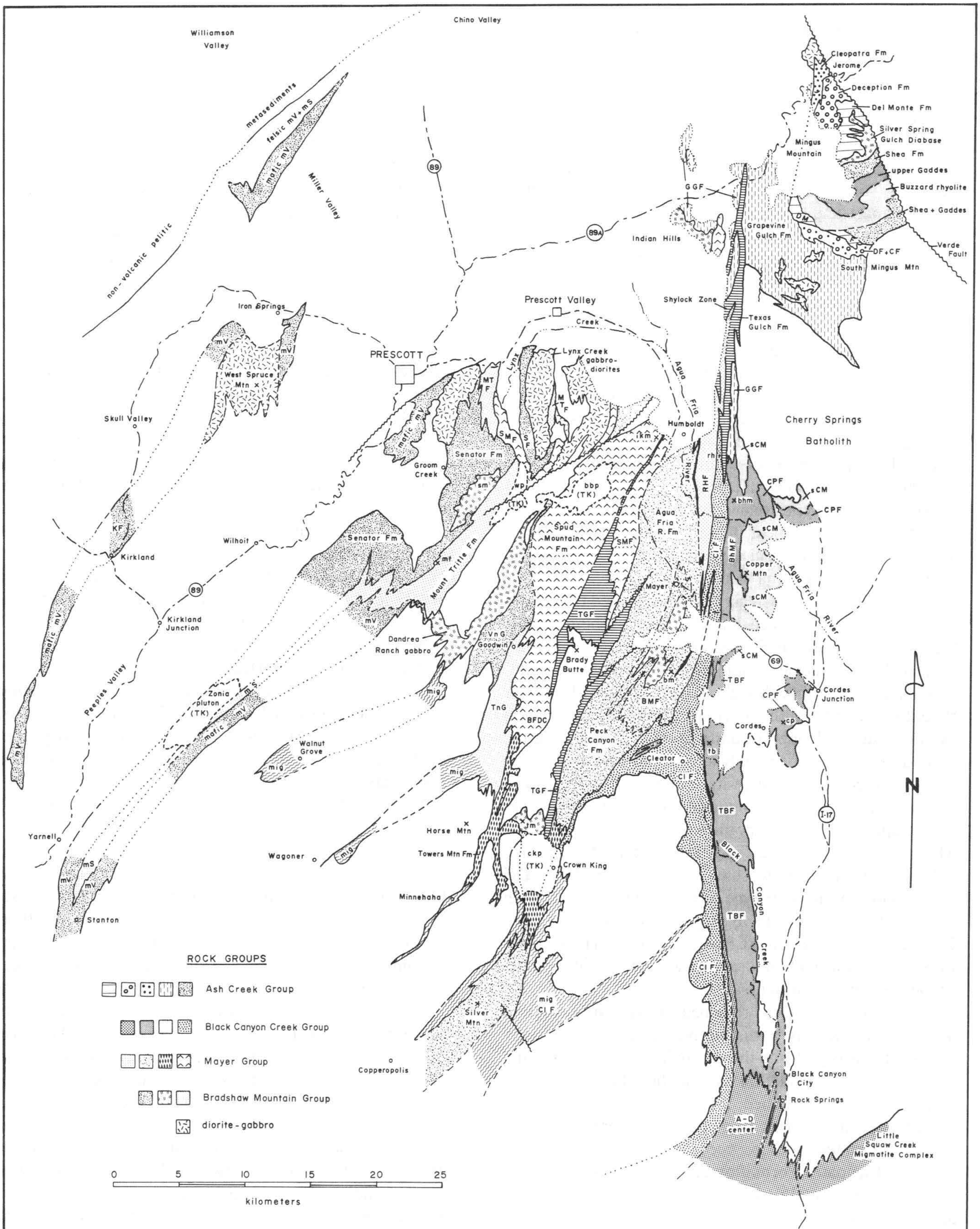
A major regional unconformity overlies the Cleopatra Formation near Jerome, but under Mingus Mountain to the southwest, it progressively cuts down section to erode the underlying Deception Formation. South of Mingus Mountain, the same unconformity cuts into southeasterly equivalents of the Deception and Cleopatra Formations at Jerome, thus demonstrating that major regional facies changes exist in the Jerome belt and mirror abundant facies changes and lensoidal units seen on a more detailed scale.

Grapevine Gulch Formation — New Regional Extent

Overlying this regional unconformity is the youngest formation of the Jerome volcanic belt: a thick, widespread suite of volcaniclastic deposits of rhyodacite to dacite composition, including bedded tuff, turbidite graywacke, agglomerate, conglomerate, and diamictic tuff. These strata form lensoidal deposits that undergo several facies changes and thicken dramatically in the southwest part of Mingus Mountain. This sequence was named **Grapevine Gulch Formation** by C. A. Anderson and Creasey (1958); it is lithologically distinctive, has withstood tests for stratigraphic uniqueness, and is therefore adopted intact.

However, prior mapping restricted Grapevine Gulch Formation to just the Mingus Mountain area east of the Shylock zone, and extended it no farther west or south (C. A. Anderson and Creasey, 1967). This confined Ash Creek Group to Mingus Mountain and thus supported C. A. Anderson's (1968b) hypothesis that the Shylock zone was a major strike-slip fault juxtaposing what he presumed to be very different rock groups of the Jerome and Prescott belts.

Now, important new relationships have been found in and west of the Shylock tectonic zone. Distinctive Grapevine Gulch volcaniclastic strata and their characteristic gabbro-diorite bodies occur both within the northern Shylock zone and in the Indian Hills west of Mingus Mountain (fig. 3). Grapevine Gulch rocks have been traced south down the Shylock zone toward Mayer where they lap out and are unconformably overlain by younger volcaniclastic (stage 5) strata of the Prescott belt, as described later. The Grapevine Gulch Formation therefore continues on its northwest trend across the Shylock zone, not substantially offset by it, although rocks within it are highly strained. Thus, it is the Ash Creek Group that makes up the Indian Hills, not other formations once thought to be part of the Prescott belt (C. A. Anderson and others, 1971). These crucial relationships show that the Ash Creek Group is not bounded by the Shylock "fault" and disprove major lateral offset along the Shylock zone.


Chino Valley
Williamson Valley
metasediments
felsic mV + mS
mafic mV
Miller Valley
non-volcanic pelitic
89
89A
Cleopatra Fm
Jerome
Deception Fm
Del Monte Fm
Silver Spring Gulch Diabase
Mingus Mountain
Shea Fm
upper Gaddes
Buzzard rhyolite
Shea + Gaddes
Verde Fault
DM
DF+CF
South Mingus Mtn
GGF
Indian Hills
Grapevine Gulch Fm
Shylock Zone
Texas Gulch Fm
Prescott Valley
Creek
Iron Springs
PRESCOTT
West Spruce Mtn
Lynx Creek gabbro-diorites
Lynx
Agua Fria
MTF
MF
SMF
SF
Skull Valley
Humboldt
Cherry Springs Batholith
Senator Fm
Groom Creek
sm
wp
(TK)
bbp (TK)
lkm
Spud Mountain Fm
Agua Fria R. Fm
River
RHF
rh
sCM
CPF
bhm
BhMF
Copper Mtn
Cl F
KF
Kirkland
Wilhoit
mt
Mount Tritle Fm
Mayer
TGF
Kirkland Junction
mV
Dandrea Ranch gabbro
VnG Goodwin
Brady Butte
69
Cordes Junction
TBF
bm
BMF
Peeples Valley
Zonia pluton (TK)
mS
mig
TnG
BFDC
Walnut Grove
Peck Canyon Fm
Cordes
cp
tb
Cleator
N
I-17
Yarnell
Horse Mtn
Towers Mtn Fm
tm
Wagoner
ckp (TK)
Crown King
Minnehaha
Stanton
Black Canyon Creek
ROCK GROUPS
Ash Creek Group
Black Canyon Creek Group
Mayer Group
Bradshaw Mountain Group
diorite-gabbro
Silver Mtn
mig Cl F
Copperopolis
Black Canyon City
Rock Springs
A-D center
Little Squaw Creek Migmatite Complex
0 5 10 15 20 25
kilometers

Diorite-Gabbro Bodies

A key feature of the Grapevine Gulch Formation is its self-contained gabbro-diorite bodies, which pervade the strata at many localities (especially southern Mingus Mountain), but are nowhere found intruding adjacent, older strata. Intrusion of the gabbro-diorites seems to have occurred just after deposition and possibly before consolidation of Grapevine Gulch strata and was apparently localized to the Grapevine Gulch depositional basin.

Younger Fragmentals

On the southeast end of the Jerome belt are thick felsic fragmentals unlike any other units in the belt. Polymictic dacitic agglomerate, limestone-fragment conglomerate, and other types of recycled rocks are overlain by a chaotic array of oxidized rhyodacite agglomerate and breccia. These shallow subaqueous and subaerial deposits are now known to be stratigraphically younger than all foregoing deeper submarine deposits in the Jerome belt. C. A. Anderson and Creasey (1958) included the lower dacitic and upper rhyodacitic parts of this young fragmental pile into their oldest "Gaddes basalt" and "Buzzard rhyolite," respectively. Although these young rocks correlate to the youngest intermediate-felsic fragmentals of the Black Canyon Creek Group in the Prescott volcanic belt (fig. 4), they are retained as upper Gaddes and Buzzard units (fig. 4) until future work resolves their complex stratigraphies.

Texas Gulch Formation

In the northern Shylock zone is a distinctive sequence of purple slate, siltstone, wacke, conglomerate, and tuff that is younger than all preceding strata of the Jerome volcanic belt: where quartz diorite of the Cherry Springs batholith intrudes formative volcaniclastic (Grapevine Gulch) strata, the purple slate sequence rests in depositional unconformity upon the quartz diorite (P. Anderson, 1986). C. A. Anderson and Creasey (1958) named this sequence **Texas Gulch Formation,** but evidently did not perceive the significance of its unconformable relation on the quartz diorite. The Texas Gulch Formation is now known to be a time-stratigraphically unique formation that postdates all other strata in the Jerome volcanic belt by an event of plutonism, uplift, and unroofing of the Cherry Springs batholith. The Texas Gulch Formation was first included in the Ash Creek Group, was later recognized as younger (C. A. Anderson and others, 1971), and is now not considered to be part of that group (fig. 4 and P. Anderson, 1986).

STRATIGRAPHY OF THE PRESCOTT VOLCANIC BELT

Previous studies suggested little if any correlation between volcanic stratigraphies of the Prescott and Jerome belts. The above summary shows five major depositional stages in evolution of the Jerome belt, and, as seven similar stages in the Prescott belt's evolution unfold below, close links between the two belts will become clear (fig. 4). The seven depositional stages of the Prescott belt can be translated directly into one or more lithostratigraphically unique formations that were deposited during each stage. Each formation belongs to one of three major rock-stratigraphic groups: rocks of stages 1 and 2 belong primarily to the **Bradshaw Mountains Group,** rocks of stage 3 and stage 4 make up the **Mayer Group,** and rocks of stage 5 and stage 6 compose the **Black Canyon Creek Group.**

BRADSHAW MOUNTAINS GROUP—STAGE 1

The oldest stage 1 rocks in the Prescott volcanic belt lie in the northern Bradshaw Mountains south of Prescott and comprise a thick sequence of submarine aphanitic basalt flows, pillow lavas, basaltic breccia, agglomerate, and tuff, with minor intercalated, altered rhyodacite flows, dikes, and tuffs. The basalts are tholeiitic, commonly spilitic, the rhyodacites are keratophyric, and lateral facies equivalents in deep-water pelagic settings comprise a suite of tholeiitic basalt flows, agglomerate, tuff, graywacke, and ribbon chert. This deep-water oceanic suite was once extensive throughout the region west and southwest of Prescott, but now occurs only as mafic volcanic screens within and between granodiorite plutons and batholiths that subsequently pervaded the area.

Farther south in the Bradshaw Mountains, near Goodwin, an identical sequence of tholeiitic aphanitic basalt flows, pillowed units, and mafic tuffs is separated from northern exposures by several plutons, but shows all evidence of being stratigraphically equivalent to the main stage 1 tholeiitic basalt exposures to the north (fig. 3). The

Figure 3 (facing page). Geologic map of major stratigraphic sequences in the Prescott-Jerome volcanic belts. Areas between the major units are occupied by Proterozoic plutonic rocks (heavy solid lines), discussed in Part 3, or post-Proterozoic cover (thin dotted lines). Some smaller areas of cover within the major units are omitted for simplicity. The following abbreviations are used. **Ash Creek Group:** CF = Cleopatra Formation; DF = Deception Formation; DMF = Del Monte Formation; GGF = Grapevine Gulch Formation; TGF = Texas Gulch Formation. **Bradshaw Mountains Group:** SF = Senator Formation. MTF = Mount Tritle Formation; VnG = volcanics (Senator Formation) near Goodwin; TnG = tuffs (Mount Tritle Formation) near Goodwin. KF = Kirkland Formation. **Mayer Group:** RHF = Round Hill Formation; BMF = Bluebell Mine Formation; SMF = Spud Mountain Formation. **Black Canyon Creek Group:** BhMF = Binghampton mine facies; CPF = Cordes Peak facies; TBF = Townsend Butte facies. sCM = subaerial rocks of Copper Mountain volcanic center; Cl F = Cleator Formation. **Other rock types:** A-D center = Andesite-dacite center near Black Canyon City; mig = migmatite and migmatitic varieties of other formations; mV = metavolcanic rocks; mS = metasedimentary rocks. **Laramide (TK) Plutons:** ckp = Crown King pluton; bbp = Big Bug pluton; wp = Walker pluton. **Place Names:** BFDC = Battle Flat dacite center; bm = Bluebell mine; bhm = Binghampton mine; cp = Cordes Peak; ikm = Iron King mine; mt = Mount Tritle; mu = Mount Union; rh = Round Hill; sm = Spruce Mountain; tb = Townsend Butte; tm = Towers Mountain.

most significant features of the earliest tholeiitic basaltic volcanism are its primitive, magnesian, bimodal character, very deep water setting, and lack of differentiation.

New Stratigraphy

The stratigraphically oldest stage 1 mafic volcanics of the Prescott belt, including tholeiitic basalt flows, pillow lavas, breccia, tuff, agglomerate, and correlatives in the northern Bradshaw Mountains and near Goodwin, are newly and formally named the **Senator Formation** for their type section along the Senator road near the Senator mine south of Prescott. To the southwest away from the main Senator volcanic center of the northern Bradshaw Mountains, other volcanic strata interface with the distinctive Senator strata and may, in the future, warrant recognition as separate formations.

Previous Stratigraphy

The earliest volcanic units of the Prescott belt were not recognized as a time-stratigraphically unique igneous suite, and were placed in different map units depending on location: (1) the main northern exposures, together with younger rock units, were mapped as "Green Gulch volcanics," which is no longer a viable formation (P. Anderson, 1986); and (2) the southern exposures near Goodwin were included in the "Spud Mountain volcanics" map unit because they lie south of C. A. Anderson and Creasey's (1958) "Chaparral fault." The Chaparral fault was thought to be a profound fault separating major rock packages, but new mapping shows it to be a zone of high strain that mylonitized intruding plutons and caused only minor (1 km) offset of stratigraphic units as they continue across the zone (fig. 3).

Gabbroic Bodies

The oldest plutonic rocks in the Senator Formation are pyroxenitic enclaves in the Wilhoit batholith and pyroxene gabbro and microgabbro bodies that are subvolcanic to mafic flows of the Senator Formation. Large gabbro-microgabbro bodies occur on Spruce Mountain and to the south as the **Dandrea Ranch Gabbro,** which is newly named for its type section near Dandrea Ranch, north of Goodwin (fig. 3). The Dandrea Ranch gabbro may have subvolcanic or intrusive relations with the Senator Formation near Goodwin. Other stage 2 plutons of diorite-gabbro composition in the northern Bradshaw Mountains postdate the gabbroic bodies.

BRADSHAW MOUNTAINS GROUP—STAGE 2

The Senator Formation is unconformably overlain to the southeast by a diverse tuffaceous sequence consisting mainly of mafic-felsic tuff, volcanic graywacke, and tuffaceous siltstone. The unconformity is of low (2- to 10-degree) angular discordance, but over kilometers of strike, substantial stratigraphic onlap occurs (fig. 3). Stratigraphy in the tuffaceous sequence is very complex, with abundant turbidites, recycling, and local unconformities; also, major facies changes exist along strike, so the sequence may be time transgressive.

New Stratigraphy

The tuffaceous sequence is newly and formally named the **Mount Tritle Formation** for key exposures on the east flank of Mount Tritle west of Mount Union (fig. 3). Tuffs and volcanic siltstone that unconformably overlie Senator Formation mafic volcanics near Goodwin are a facies variant of the type (Mount Tritle) locality, and may later warrant distinction as a separate but coeval formation. Both exposures of Mount Tritle Formation are unconformably overlain along their eastern side by dacitic breccias and tuffs of the Spud Mountain Formation (see Mayer Group).

Abandonment of Former Stratigraphy

Both the Senator and Mount Tritle Formations were collectively mapped in the past as "Green Gulch volcanics" (C. A. Anderson and Blacet, 1972a,b,c; Krieger, 1965; C. A. Anderson and others, 1971). The "Green Gulch" name has since been abandoned because it includes strata of the Spud Mountain, Senator, and Mount Tritle Formations, all of which extend across both sides of the Chaparral zone, and none of which are limited by faults (fig. 3; see P. Anderson, 1986 for details). Instead of occurring at faults, the fundamental lithologic-petrologic-geochemical breaks in the volcanic belts are stratigraphic; one such break is between strata of the Bradshaw Mountains and Mayer Groups.

Gabbro-Diorite Bodies

Large gabbro-diorite bodies similar to those intruding the Grapevine Gulch Formation in the Jerome volcanic belt intrude both the Senator and Mount Tritle Formations, but not the younger Spud Mountain Formation. Such bodies in the northern Bradshaw Mountains are newly named the **Lynx Creek Gabbro-Diorite** and correlate to gabbro-diorite bodies in the Grapevine Gulch Formation of the Jerome belt. Whereas the earlier microgabbro bodies are texturally gradational into basalt and conformable with Senator mafic volcanics, Lynx Creek bodies are discordant plugs and small stocks diapiric into the host volcanics.

MAYER GROUP—STAGE 2

East of the Bradshaw Mountains near Humboldt is a tuffaceous sequence equivalent in age and stratigraphic position to the Mount Tritle Formation, but without an underlying basalt flow sequence comparable to the Senator Formation. Basal andesite flows and tuffs are overlain by tuffaceous siltstone, rhyolitic to intermediate tuff, and iron formation, among which many facies changes occur. A rhyolite dome with mineralized rhyolitic tuff facies at the stratigraphic top of the sequence is capped by a major

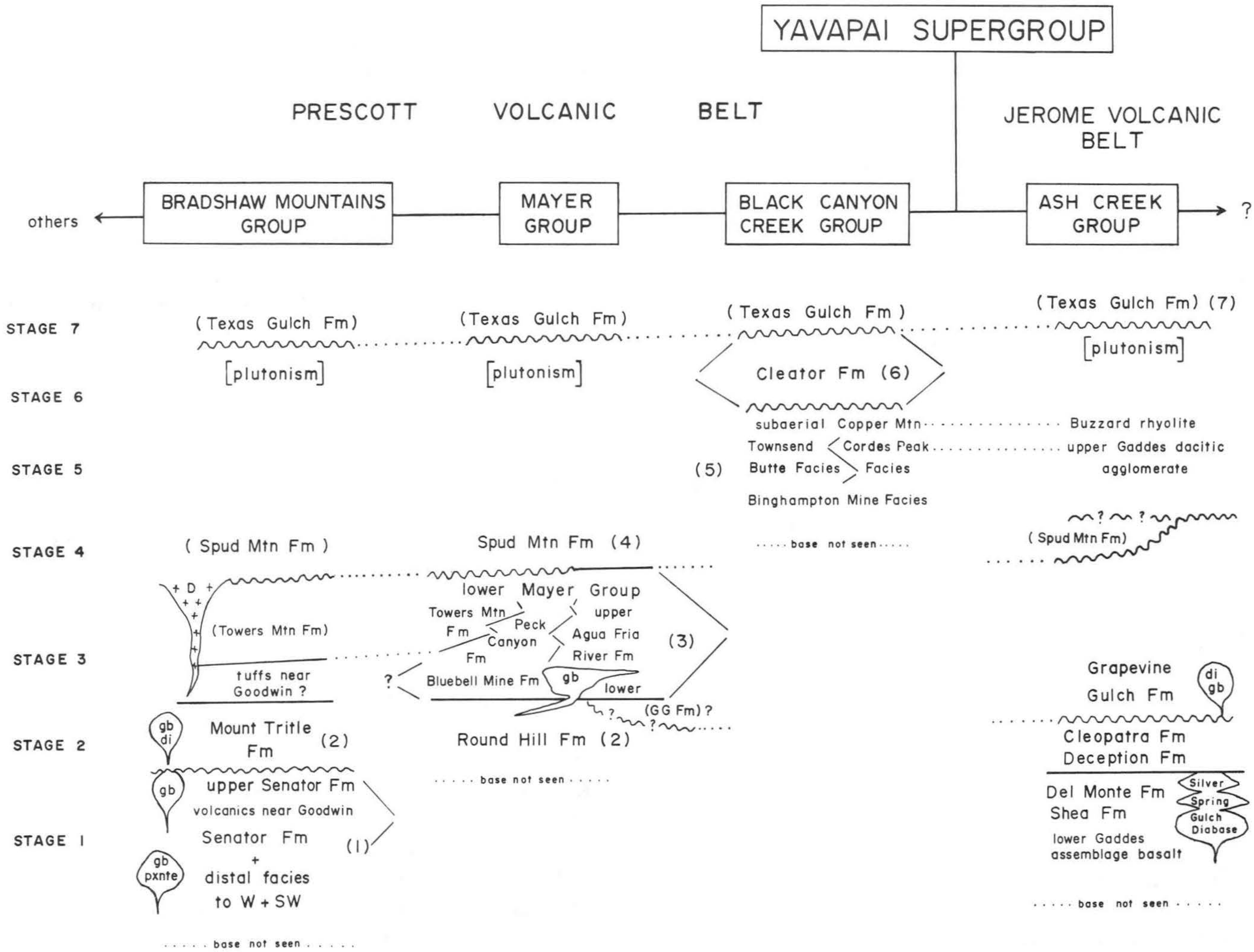


Figure 4. New stratigraphic subdivision of the Prescott-Jerome volcanic belts showing the relationship of groups and formations to depositional stages (numbers in brackets), unconformities (wavy lines), conformable contacts (straight lines), major facies changes (wedge-shaped edges), and correlations between areas (dotted lines). Departure of some unconformities from horizontal depicts the time-transgressive nature of the contacts. Formation names enclosed by brackets indicate places where the formations are not part of the major rock group beneath which they are listed (e.g., Texas Gulch Formation in the Prescott belt), but are shown in those positions only to demonstrate unconformable relations. The wedge-shaped edges of some of the major rock packages (e.g., lower Mayer Group, Cleator Formation) depict the laterally discontinuous nature of the rock units. All depositional relations have been observed in the field (P. Anderson, 1986). An early event of plutonism occurred between depositional stages 6 and 7 and is shown on the diagram only to depict unconformable relations with the Texas Gulch Formation. Abbreviations: D = dacite; di-gb = diorite-gabbro; GG Fm = Grapevine Gulch Formation; gb = gabbro; pxnte = pyroxenite.

regional horizon of iron formation (fig. 3). The tuffaceous sequence is unconformably overlain to the east by younger sedimentary rocks and conformably overlain to the west by mafic volcanics of the Mayer Group, and hence is bounded on both sides by younger rock units.

New Stratigraphy

This tuffaceous sequence of andesitic and rhyolitic tuffs, flows, and iron formation is newly formally named the **Round Hill Formation** for exposures west of Round Hill, east of Humboldt. It correlates in time and stratigraphic position to the Mount Tritle Formation, but is slightly different. Its base is not seen, although mafic flows are more abundant in the lower parts. The Round Hill Formation is unconformably overlain to the east in the Shylock zone by the southern end of the Grapevine Gulch Formation, and is overlain to the west by pillow basalts exposed along the Agua Fria River east of Humboldt (figs. 3 and 4). Former mapping did not recognize the distinct stratigraphic position of the Round Hill Formation, and the formation, along with other rocks in the Shylock zone unrelated to true Spud Mountain pyroclastic volcanism, was included by C. A. Anderson and Blacet (1972a) in their broad "Spud Mountain volcanics" unit.

MAYER GROUP—STAGE 3

Immediately following deposition of the Mount Tritle and Round Hill Formations, a huge basaltic volcanic complex evolved in the Mayer area to spread mafic rocks from north of Mayer to as far south as Crown King. The main volcanic center near the Bluebell Mine south of Mayer has a coarse-grained gabbro core enveloped by microgabbro and massive basalt (fig. 3). To the north near the Agua Fria River, thick pillowed basalt flows cored by microgabbro predominate. The mafic rocks are interbedded with rhyolitic flows, felsic and mafic tuff, and abundant hematitic chert or iron formation.

Repeated extrusions of mafic material southward from the main Bluebell mine gabbro center built up the following sequence toward the Peck Canyon area: (1) basalt flows, tuffs, and minor rhyolite flanked by a mafic agglomerate wedge; (2) a thick sequence of basalt flows, mafic and felsic tuff, mineralized rhyolite and iron formation, and mafic dikes that cut the agglomerate; and (3) more mafic agglomerates and calcareous mafic tuffs in the Crown King area. A smaller companion mafic volcanic vent at Towers Mountain west of Crown King crystallized a gabbro-diorite in its core, surrounded by microgabbro and massive basalt, and extruded distinctive coarse-fragment basalt breccia that fines outward from the center into finer breccia and calcareous mafic tuff. These calcareous mafic tuffs from the Towers Mountain volcanic neck spread widely to the south and west over older Mount Tritle rocks, and also capped coarser breccia at the mafic volcanic center.

The thick basaltic pile to the north along the Agua Fria River (fig. 3) is well fractionated into the following westward-younging sequence: (1) lowermost pillowed basalt flows and gabbro-microgabbro bodies cutting the Round Hill Formation to the east, (2) pyroxene-crystal and aphyric basalt, (3) plagioclase-crystal basaltic andesite, rhyolite, and iron formation, and (4) uppermost plagioclase-crystal andesite, dacite, mineralized rhyolite, breccia, tuff, and iron formation. This fractionated sequence is overlain to the west by plagioclase-phyric Spud Mountain dacitic breccias compositionally like the upper andesites and dacites, but separated from them by an unconformity.

New Stratigraphy

Mafic volcanic rocks extruded from several centers during stage 3 make up the Mayer Group stratigraphy, but the entire pile is too stratigraphically complex for a single formation, so it is newly subdivided into four formations. Lithostratigraphic units are zonally disposed about and young away from volcanic centers, so formations are defined as areally distinct, sequential lithofacies.

Lithofacies near the Bluebell mine mafic center—the subvolcanic basalt-gabbro core, basalt, and minor rhyolite flows and tuffs, iron formation, and mafic agglomerate—are newly named the **Bluebell Mine Formation,** which may predate other Mayer Group formations (figs. 3, 4). The **Peck Canyon Formation** is defined as including all basalt and rhyolite flows, tuff, and iron formation in Peck Canyon south of Turkey Creek and north of the Gladiator mine. The Peck Canyon Formation is overlain to the south by younger mafic agglomerate and tuff of the **Towers Mountain Formation,** which is newly defined here as including a volcanic neck of gabbro-microgabbro-basalt, zonally disposed basalt breccia and agglomerate, and widespread capping calcareous mafic tuff.

All northern exposures of Mayer Group mafic basalts are now described as the **Agua Fria River Formation.** Pillow basalt flows cored by microgabbro dominate the type section along the Agua Fria River, but unpillowed basalt flows, basalt-andesite-rhyolite tuff, and hematitic chert predominate elsewhere. Plagioclase-phyric andesite flows, with rhyolite flows, breccia, and iron formation in the upper western part of the formation just beneath Spud Mountain rocks may be the youngest part of stage 3 Mayer Group volcanism. All four Mayer Group mafic volcanic formations have a common petrogenesis and evolved in a brief igneous event, so they can be assembled into a single petrologic subgroup (P. Anderson, 1986).

Previous Stratigraphy

Mafic volcanic rocks of the Mayer Group were formerly termed "Iron King volcanics," the "type section" being Iron King Gulch east of the Iron King Mine (C. A. Anderson and Creasey, 1958). Detailed mapping has now shown this "type section" to consist entirely of sheared but distinctive feldspar-crystal dacitic tuffs and breccias of the Spud Mountain Formation (P. Anderson, 1986). Because the type section consists of rocks of another formation, it is invalid, so "Iron King volcanics" is not a valid stratigraphic unit and therefore should be dropped entirely. DeWitt (1979) reached the same conclusion from different structural-stratigraphic arguments south of Mayer. Also, a narrow belt of Texas Gulch strata, interpreted as being bounded on both sides by faults (C. A. Anderson and Blacet, 1972c) and as separating "Iron King" and Spud Mountain map units in the Iron King mine area (C. A. Anderson and Blacet, 1972a,b), is not bounded by faults (O'Hara, 1980). Instead, Texas Gulch strata lap regionally across older strata and unconformably overlie Spud Mountain breccias near the Iron King mine (P. Anderson, 1986).

MAYER GROUP—STAGE 4

Immediately following construction of the huge Mayer Group mafic volcanic pile, voluminous dacite-andesite pyroclastics were extruded into the trough between the earlier stage 1 and stage 3 mafic volcanic centers, and now continuously border Mayer Group mafic volcanics on the west (fig. 3). This pyroclastic suite originated from two coeval volcanic centers: (1) a minor dacite-andesite center in the north related to the upper Agua Fria River Formation andesites and (2) a major subvolcanic dacitic center at Battle Flat to the south, from which pyroclastics spread asymmetrically northward as a classic suite of flow, breccia, and tuff lithofacies in a single stratigraphic package (fig. 2). The upper pyroclastic units show facies variations from plagioclase-crystal dacite flows, to breccia, agglomerate, and welded pumice-fragment crystal tuffs away from the volcanic center.

Previous Stratigraphy

The dacitic pyroclastic sequence was named "Spud Mountain volcanics" by C. A. Anderson and Creasey (1958) for a type section on Spud Mountain near Humboldt. Because the type section identifies a lithostratigraphically unique rock sequence, it is valid, so the Spud Mountain term is retained, but in revised form. C. A. Anderson and Creasey (1958) and C. A. Anderson and Blacet (1972a,b) also included many other rock units in and east of the Shylock zone in their Spud Mountain map category, units now known to be lithostratigraphically distinct clastic formations unrelated to pyroclastic volcanism of the Spud Mountain Formation except for possible slight time overlap in one or two examples. Hence they are excluded from the newly defined Spud Mountain Formation.

New Stratigraphy

The newly defined **Spud Mountain Formation** is a lithostratigraphically distinct suite of plagioclase-crystal dacite flows, breccias, and crystal tuffs areally restricted to the Bradshaw Mountains west of Humboldt and parts of the Indian Hills north of Humboldt (fig. 3). It includes no other exposures of feldspathic tuffs to the east, especially in the Shylock zone, nor any pelitic sedimentary sequences to the south near Cleator. The formation has a relatively homogeneous dacite to andesite composition and is entirely of pyroclastic derivation. The Battle Flat dacite center was the source for most pyroclastics, producing mostly dacite-andesite breccias in lower parts, dacitic agglomerates in central parts, and welded dacite crystal tuffs in upper parts of the formation.

Regional Unconformable Relations and Stratigraphic Links to the Jerome Belt

The Spud Mountain Formation is nearly conformable with the underlying Agua Fria River Formation near Humboldt, but cuts down section progressively to the south. Regionally, however, the base of the Spud Mountain Formation is an unconformity transgressing all subjacent strata (fig. 3). South of Battle Flat, Spud Mountain dacite crystal tuffs unconformably overlie Towers Mountain calcareous mafic tuffs and Mount Tritle tuffs, and at Battle Flat, the massive dacite intrudes Mount Tritle tuffs. Near Goodwin, dacite breccias lap across the contact of the Mount Tritle and Senator Formations but do not reach the subjacent microgabbros. Farther north near Walker, Spud Mountain breccias and crystal tuffs again lap across the Mount Tritle-Senator Formation contact and associated gabbro-diorite bodies.

Thus, the Spud Mountain Formation unconformably overlies the Senator, Mount Tritle, Towers Mountain, and Agua Fria River Formations of the Prescott belt. In the Indian Hills to the north, however, Spud Mountain dacitic breccias also unconformably overlie the Grapevine Gulch Formation of the Jerome belt and its contained gabbro-diorites, in parallel with the overlap of gabbro-diorites in the Bradshaw Mountains Group of the Prescott belt. These unconformable relations conclusively show that by at least stage 4, the Prescott and Jerome volcanic belts were stratigraphically linked; they were not, therefore, later juxtaposed by a hypothetical wrench fault along the Shylock zone. Moreover, correlation of diorite-gabbros in the Grapevine Gulch and Senator-Mount Tritle Formations links both volcanic belts even further back in time, almost to their inception.

Nature of Intervolcanic Unconformities

These unconformable relations are crucial to a clear understanding of the primary volcanic structure of the Prescott belt and its relationship to the Jerome belt. Not only do they demonstrate stratigraphic linkage of the belts, but they refute a single-rock-column concept for the volcanic stratigraphy. If the strata of stages 1, 2, and 3 were all superimposed in a single stratigraphic column, the Spud Mountain unconformity would have to transect 20 km of section on a 30-degree discordance, necessitating uplift and orogeny halfway through the Prescott belt's volcanic evolution without its emergence from a submarine environment—clearly an impossibility. No such tectonic activity is needed if stage 4 deposits fill in a shallow trough between variably interlensed, wedge-shaped, subjacent rock units (fig. 2).

Thus, intervolcanic unconformities in the central volcanic belt are not disconformities or paraconformities, but are true unconformities because they result from the spread of new volcanic deposits discordantly across older volcanic strata, major rock-unit contacts, *and* early intrusions. One of the key earmarks of intervolcanic unconformities is their low angular discordances—5 to 10 degrees, rarely 15 degrees—in contrast to unconformities with great angular discordances caused by orogenies. The low discordances are subtle and are missed by many workers, but mapping of stratigraphy under the unconformity on a regional scale clearly shows they are truly major regional unconformities.

BLACK CANYON CREEK GROUP — STAGE 5

Closely following or slightly overlapping in time with Spud Mountain dacitic pyroclastic volcanism was inception of a very different event of felsic pyroclastic volcanism in the eastern Prescott volcanic belt. The new stage 5 deposits do not extend into the west-central Prescott belt, just as the older mafic deposits do not extend as far east as the new volcanic suite. Stage 5 rocks were laid down on mafic basement east of stage 3 rocks and do not overlie stage 4 strata. Stage 5 shows great lithologic diversity and stratigraphic complexity at inception, but remarkable lithologic uniformity at culmination, when vast quantities of rhyodacitic-rhyolitic pyroclastics were extruded.

The oldest stage 5 volcanics—phenocrystic dacite-andesite flows and breccias—are overlain by feldspar-phyric dacite breccia and agglomerate, quartz-feldspar-phyric rhyodacite and rhyolite, and feldspar-crystal graywacke and conglomerate west of Copper Mountain and near Cordes Peak (fig. 3). The earlier phenocrystic rocks are overlain in slight angularity by the voluminous aphyric dacite-rhyodacite fragmentals that were extruded south down the Black Canyon belt as submarine valley flows from the Copper Mountain volcanic center (fig. 3). These submarine felsic pyroclastics were interleaved with andesite-dacite flows and tuffs from a coeval volcanic center farther south near Black Canyon City. As the Copper Mountain center became subaerial, felsic agglomerates spread northward unconformably over the Grapevine Gulch Formation and dacitic agglomerates of the Jerome belt. The felsic event's waning stages are marked by decreased coarse debris, finer grained tuffs, and a major iron formation horizon extending the 40-km length of the Black Canyon belt. Overlying the iron formation are thin andesite flows that signal the end of stage 5 volcanism.

New Stratigraphy

Because the stage 5 felsic volcanic rocks are widely and best exposed in Black Canyon Creek south of Mayer, the new name **Black Canyon Creek Group** is formally introduced to describe the complex felsic volcanic assemblage and its associated tuffaceous sedimentary rocks. In the felsic volcanics that make up most of the group, stratigraphy is extremely complex and has not been clearly resolved to date (Jerome, 1956; C. A. Anderson and Blacet, 1972c; Winn, 1982), partly because most stratigraphic units are lensoidal, laterally discontinuous, and interrelated by facies changes. Hence, only lithofacies units that appear to be the essence of true stratigraphy are herein formally recognized at the present time.

The older feldspar-phyric strata of the Black Canyon Creek Group are well exposed near the Binghampton mine and are collectively referred to as the **Binghampton mine facies,** which includes a conglomerate that unconformably overlies Grapevine Gulch Formation and is unconformably overlain by the younger aphyric felsic fragmentals of Copper Mountain. The younger felsic fragmentals have three main facies (see P. Anderson, 1986 for details): (1) the **Cordes Peak facies,** consisting of volcanic center and vent deposits of Copper Mountain and Cordes Peak; (2) the **Townsend Butte facies,** comprising bedded felsic fragmentals in the Black Canyon belt to the south, named for Townsend Butte exposures south of Mayer; and (3) an unnamed facies representing the oxidized subaerial deposits north of Copper Mountain and on south Mingus Mountain.

The southernmost strata of the Townsend Butte facies are interbedded with basaltic andesite and dacite flows and tuffs from an unnamed mafic volcanic center southwest of Black Canyon City. This andesite-dacite sequence provides a critical link with the geology of the New River-Cave Creek volcanic belts [see Part 2]. Prominent beds of iron formation and thin andesite flows at the top of the Townsend Butte facies in the transition from volcanic to sedimentary strata effectively demarcate the upper limit of felsic volcanism in the Black Canyon Creek Group; above them are Cleator metasedimentary rocks, the youngest formation of the Black Canyon Creek Group (see stage 6 below).

Stratigraphic Links to the Jerome Belt

Reworked dacite-rhyodacite agglomerates and conglomerates comparable to the Townsend Butte facies were deposited unconformably upon Grapevine Gulch strata at the southeast end of Mingus Mountain. Thereafter, oxidized rhyolitic breccias and huge boulder agglomerates correlative to the youngest subaerial stage of Copper Mountain volcanism were deposited on these subaqueous rocks (fig. 4). Both the subaqueous and subaerial deposits were originally continuous with exposures north of Copper Mountain, but were separated later by intrusion of Cherry quartz diorite (fig. 3). Such correlations lend even further support to the earlier Spud Mountain-Grapevine Gulch link, showing that the Prescott and Jerome belts were linked throughout their volcanic evolution.

Previous Stratigraphy

Both subaqueous and subaerial agglomerates noted above in the south Mingus Mountain area are younger than the Ash Creek Group, as they overlie its youngest formation—the Grapevine Gulch (fig. 4). Despite this, C. A. Anderson and Creasey (1958) included the dacitic agglomerates in "Gaddes basalt" and the rhyolitic ones in "Buzzard rhyolite," together with much older rocks of the Ash Creek Group. Consequently, the name "Buzzard rhyolite" is no longer formally recognized, and is not representative of stage 5 subaerial felsic volcanism. Like the Gaddes, it may be considered an assemblage until future work resolves its components. Part of the "Gaddes" and part of the "Buzzard" correlate to the Townsend Butte facies and subaerial facies of Copper Mountain respectively (fig. 4). C. A. Anderson and Blacet (1972a,b,c) mapped all felsic fragmentals of the Black Canyon Creek Group as "Spud Mountain volcanics" and the Copper Mountain volcanic center as "silicified and sericitized rocks." The Spud

Mountain Formation is now known to be unrelated to the Black Canyon Creek Group, the two being petrologically and chemically distinct igneous events.

BLACK CANYON CREEK GROUP—STAGE 6

Overlying the Townsend Butte iron formation is a sedimentary sequence whose base broadly conforms to the iron formation for more than 40 km, but in detail cuts down on a 2- to 5-degree discordance to locally erode the iron formation. The base of the sedimentary sequence has abundant tuff wacke and volcaniclastic tephra reworked from stage 5 volcanic rocks. Upper sedimentary units include well-bedded subgraywacke, quartz wacke, argillaceous quartzite, and pelite derived by erosion and recycling of the underlying volcanic rocks. This sedimentary sequence marks the first stage in evolution of the Prescott volcanic belt when an entirely clastic sequence accumulated, even though the material was ultimately of volcanic origin.

On its west side, the sedimentary sequence is clearly unconformable on the earlier volcanic terrane, as the contact laps across older stratigraphic units with a 5-to 10-degree discordance. Many mafic volcanic units in the Mayer Group are transected, including calcareous mafic tuffs of the Towers Mountain Formation near Crown King, basalt flows and iron formation of the Peck Canyon Formation near the DeSoto mine, and mafic agglomerate of the Bluebell Mine Formation near the Bluebell mine, into which an erosional trough was cut (fig. 3).

Northeast of Mayer, the basal unconformity of the sedimentary sequence continues to cut down section into the older Round Hill Formation on the west, and into broadly conformable stage 5 volcaniclastic units on the east. Volcanic detritus was mostly derived by erosion of this northern area, as indicated by a southward decrease in tuffaceous component in the stage 6 sedimentary sequence. At its northernmost exposure in the Shylock zone, the stage 6 sedimentary sequence is unconformably overlain by purple slates of the Texas Gulch Formation.

New Stratigraphy

This stage 6 sedimentary sequence is the youngest formation in the Black Canyon Creek Group and is newly and formally named the **Cleator Formation** for its type section near Cleator. The Cleator Formation includes pelitic and tuffaceous sedimentary rocks extending down the Black Canyon belt, southwest to Crown King, and north up the Shylock zone (fig. 3). The Cleator Formation is nearly conformable on the Townsend Butte and Binghampton mine facies to the east, but is markedly unconformable upon the older mafic volcanic formations of the Mayer Group to the west, cutting significantly down into both the Bluebell Mine Formation south of Mayer and the Round Hill Formation north of Mayer.

Previous Stratigraphy

The Cleator Formation is younger than the Spud Mountain Formation, a fact not evident from former published maps. C. A. Anderson and Blacet (1972a,b) showed the entire Black Canyon Creek Group, including the Cleator Formation, as part of their "Spud Mountain volcanics," but as noted previously, rocks of the Black Canyon Creek Group are unrelated to the Spud Mountain Formation, each sequence representing a separate igneous event with its derivative sedimentary rocks.

TEXAS GULCH FORMATION—STAGE 7

Between depositional stages 6 and 7 is a major plutonic event (fig. 4), during which the first pretectonic granodiorite and tonalite plutons and batholiths were emplaced into the Prescott-Jerome volcanic belts [see Part 3]. Subsequent clastics of the Texas Gulch Formation accumulated in depositional troughs uniquely confined to the edges of two early plutons that intruded the volcanic belts: one trough is along the northwest edge of the Cherry Springs batholith where the Shylock zone links the Prescott and Jerome belts (fig. 3) [see Part 3]; the other trough is at the north end of the Brady Butte Granodiorite and extends down its western and eastern sides toward Crown King (fig. 3).

The northern trough contains mainly well-bedded purple slate, wacke, siltstone, and conglomerate in southward-thickening depositional lenses. These unmetamorphosed sedimentary rocks rest depositionally on quartz diorite near Cherry, which requires that Texas Gulch deposition postdated emplacement and unroofing of the Cherry Springs batholith. Deformed contacts in the northern Shylock zone show the Texas Gulch Formation resting in gentle discordance on tuffs of the Round Hill Formation, felsic volcanics of the Black Canyon Creek Group, and tuffs of both the Grapevine Gulch Formation and younger Cleator Formation.

In the southern trough, a basal conglomerate derived from Brady Butte Granodiorite and felsic volcanics lies unconformably on the Brady Butte body (Blacet, 1966). Purple slate, siltstone, and graywacke fill the western part, quartzose debris and felsic tuffs from rhyolite feeders fill the eastern part, and recycled volcanic detritus and coeval felsic tuffs fill the northern part of the trough. Rhyolite flows and dikes in the sequence are themselves overlapped by purple slate near Mayer, which indicates that the Texas Gulch Formation was transgressive northward. The basal unconformity laps northward from Agua Fria basalts onto Spud Mountain breccias to where upper purple slate beds lap out near Humboldt. Even though events of plutonism, uplift and unroofing separate stage 7 from all preceding events, the Texas Gulch basal discordance is only 3 to 7 degrees, like all previous unconformities. These low discordances indicate that all seven sequences were deposited prior to the main regional deformation.

MAZATZAL GROUP — STAGE 8

Chino Valley north of the Prescott-Jerome volcanic belt has exposures of clean, mature, quartz-pebble conglomerate and trough cross-bedded quartzite (Wilson, 1939; Krieger, 1965) that postdate all other rocks in the Prescott-Jerome belts. The quartzite-conglomerate sequence correlates to Mazatzal Group strata in the Mazatzal Mountains that were laid down in fluvial and shallow-marine environments (P. Anderson and Wirth, 1981). The Chino Valley strata gained maturity through a high-energy fluvial environment that cleaned the sediment to residual quartz, chert, jasper, and resistant rhyolite pebbles in a matrix of clean sand, silt, and heavy minerals (Wirth, 1982).

PETROLOGY AND GEOCHEMISTRY

In spite of the great lithologic and stratigraphic complexity of the Prescott-Jerome volcanic belts, petrologic and geochemical variations are so systematic that just a few key petrologic and geochemical attributes can uniquely define igneous suites and unify great diversities of rock types. Implicit is the need to have petrologic-geochemical data from only those rocks closest to their original state, unaffected by alteration or metasomatism.

Major elements best define major petrologic and chemical attributes of the rock sequences, because they are inseparably linked to major rock characteristics. Evaluation of 120 new major-element analyses (P. Anderson, 1986; n. d.) and published analyses (C. A. Anderson and Creasey, 1958; C. A. Anderson, 1968a, 1972; C. A. Anderson and Blacet, 1972c; Krieger, 1965) indicate that the key chemical trends are best displayed on alkali-silica plots of K_2O + Na_2O vs. SiO_2 and K_2O vs. SiO_2. Plots of the volcanic rocks are summarized on figures 5 and 6 as fields of related samples or linear trends of sample suites.

JEROME VOLCANIC BELT

Old Gaddes assemblage basalts are similar in petrology and chemistry to basalts of the Shea Formation and are grouped with them. The oldest mafic volcanics in the Jerome belt are relatively primitive, high-Mg-Fe, low-alkali tholeiitic basalts with a Na-enrichment trend, but no K-enrichment trend. The dominant magma character is bimodal, with sodic rhyodacites at the felsic end; however, low-alkali dacites in the lower Gaddes assemblage make it a trimodal sequence. The lower part of the Shea Formation is bimodal basalt-rhyodacite, and the upper part is basaltic andesite. The lower Gaddes-Shea igneous suite is distinctively aphyric, with only a few small clinopyroxene, andesine, and oligoclase crystals.

Ensuing basaltic andesite feeder dikes and extrusive dacite flows and tuff breccia of the Del Monte Formation fractionated from a slightly higher-K tholeiitic basalt parent to produce a trimodal basaltic andesite-dacite-rhyodacite suite along a Na $>$ K, low- to intermediate-K tholeiitic trend. Similarly, the companion Silver Spring Gulch Diabase shows differentiation internally along a typical Skaergaard-type plutonic trend of Fe-Ti enrichment.

The Deception and Cleopatra Formations have no rocks more mafic than dacite, so their fractionation trends are difficult to define, but all dacite-rhyodacite analyses fall in the tholeiitic field (fig. 5). Plagioclase, quartz, and minor K-feldspar are phenocryst phases. Lower Deception phyric dacite breccias are linked to the Del Monte Formation, but upper aphyric agglomerates were derived from a separate felsic magma fractionated from an intermediate-K tholeiitic parent. Widespread chloritization is characteristic of both Cleopatra and Deception strata and is intrinsic to the depositional units, not a younger superimposed alteration, although such alteration overprints are also present. Na-Fe-Mg enrichment appears to be distinctive of the magma type.

The Deception-Cleopatra boundary is not simply a lithologic contact, but a phenocryst phase boundary that transects agglomeratic stratigraphy, where quartz-feldspar phenocrysts become large and abundant in overlying welded ash-flow tuffs. The phase boundary tracks a stage in evolution of the source magma's crystallization when its composition encountered the quartz-plagioclase cotectic, and when extrusive deposits changed from coarse fragmentals (Deception) to finely fragmental ash-fall and ash-flow tuffs (Cleopatra). These features link upper Deception and Cleopatra units as deposits that were sequentially extruded from a common source magma undergoing progressive fractional crystallization.

In contrast to all foregoing units, the Grapevine Gulch Formation is entirely calc-alkaline in geochemistry, based on total alkali and alumina content. All analyses fall above the long dashed line of figure 5, whereas all analyses from preceding formations fall below that line. The gabbro-diorites emplaced into the Grapevine Gulch Formation provide data in the low-silica range: they are also calc-alkaline, were derived from quartz-normative source magmas, and are part of the Grapevine Gulch calc-alkaline magmatic trend. The Grapevine Gulch Formation's chemical trend on figure 6 lies below a typical calc-alkaline trend, implying a low-K calc-alkaline chemistry.

PRESCOTT VOLCANIC BELT

The Senator Formation and its coeval gabbro intrusions are the most mafic, magnesian, and alkali-depleted rocks in the Prescott volcanic belt. The pyroxenite, subvolcanic microgabbro, and lower aphyric basalt flow sequences are olivine normative and were derived directly from olivine tholeiite parent magmas. Lower volcanic portions of the formation are alkali-depleted, low-K tholeiitic basalts with K_2O contents of less than 0.3 percent, and $K_2O + Na_2O$ contents of about 2 percent. Felsic flows and dikes coeval with these basalts are alkali-poor, aphanitic sodic rhyodacites. This bimodal basalt-rhyodacite pair comprises a spilite-keratophyre suite with a low-K tholeiitic trend of virtually no alkali enrichment (figs. 5 and 6).

Pillowed basalt flows, agglomerates, and tuffs in the upper Senator Formation are somewhat more Fe and alkali enriched, as are the slightly younger intrusive gabbros (figs. 5 and 6). The more alkali-rich character of these upper units is a primary feature shared by mafic flows at Goodwin and is not an artifact of alteration. Analyses of altered rocks enriched in alkalis lie in the "altered Senator Fm" areas on figure 5; their alkali mobility resulted from sea-water alteration of pillowed sequences or subsequent metamorphic-intrusive activity. Rhyodacites richer in K_2O complete the bimodal pair in the upper Senator Formation, which indicates that the rocks have a tholeiitic trend slightly richer in alkalis than that of the lower Senator Formation.

Lower Mayer Group mafic volcanics consists of alkali-poor tholeiitic basalts that overlap on figures 5 and 6 with similar rocks from the Senator Formation. However, other subtle chemical differences such as slightly higher Fe-Ti contents place the lower Mayer Group mafic volcanics on a more normal tholeiitic trend of iron enrichment. This trend is well displayed by the felsic fractions, which are rhyolites rather than rhyodacites, are higher in K_2O, and define a tholeiitic trend of moderate alkali enrichment (figs. 5 and 6). The lower Mayer Group mafic volcanics represent a strongly bimodal basalt-rhyolite association that started as 80 percent basalt and 20 percent rhyolite, but increased in felsic component in its later stages. The Towers Mountain Formation may be marginally higher in K_2O (fig. 6).

The source for Mayer Group bimodal basalt-rhyolite volcanism was a quartz tholeiite parent magma that underwent discontinuous fractional crystallization and removal of pyroxene to generate the differentiated sequence seen in the upper Agua Fria River Formation. In contrast to the aphyric Bradshaw Mountains Group volcanics, Mayer Group volcanics are characteristically phyric. The lowest Mayer Group basalts and subvolcanic gabbros are clinopyroxene phenocrystic; intermediate levels in the Agua Fria River Formation are marked by increased size and abundance of plagioclase crystals at the expense of pyroxene crystals as compositions change from basalt to basaltic andesite; finally, the uppermost Agua Fria River andesites are dominated by plagioclase phenocrysts, with lesser quartz-crystal rhyolite flows and breccias.

This compositional fractionation of the lower Mayer Group corresponds to progressively increasing alkalis in felsic members and evolution toward a calc-alkaline magma chemistry. The lowest units are intermediate-K tholeiitic (figs. 5, 6), but upper units in the Agua Fria River Formation near the Spud Mountain contact lie along a higher-K tholeiitic trend (fig. 6). Spud Mountain rocks are of even higher K tholeiitic to calc-alkaline chemistry, apparently as an extension of the same fractionation trend (figs. 5 and 6). Thus, lower Spud Mountain rocks appear to be chemically related to final crystallization of the differentiated source magma that formed the upper Agua Fria River Formation. Lithologically, however, the Agua Fria River-Spud Mountain contact is a sharp break, above which rocks are of dacitic composition. The formational boundary is therefore the change from andesite-rhyolite flows below to dacitic breccias above.

Only the lowest part of the Spud Mountain Formation contains breccias compositionally like upper Agua Fria River Formation andesite flows (fig. 5). The rest of the formation consists of dacite breccias and dacite-rhyodacite ash flows and lithic-crystal tuffs of truly intermediate composition, as shown by analyses from the Battle Flat dacite center clustering near 63 percent SiO_2 on figure 6. Pyroclastics extruded from the center are plagioclase-quartz-crystal dacites-rhyodacites of high-K tholeiitic (calcic) to calc-alkaline chemistry.

Like Deception and Cleopatra strata in the Jerome volcanic belt, dacites and rhyodacites of the Black Canyon Creek Group are widely affected by intrinsic chlorite-sericite-silica alteration. The alteration was broadly syndepositional because, although discordant alteration dominates vent areas, chloritic alteration is stratigraphically bound in distal pyroclastic and tuff strata. Analyses of the Black Canyon Creek Group fall mainly on a high-K tholeiitic trend (figs. 5, 6) but have wide alkali-silica dispersion because of this intrinsic alteration. Samples with low K (fig. 6) but normal K + Na (fig. 5) underwent K depletion because of overall Na enrichment of the felsic pile.

An important feature of figures 5 and 6 is that felsic agglomerates from "Buzzard rhyolite" in the south Mingus Mountain area plot in an identical position to rhyodacitic agglomerates of the Black Canyon Creek Group; likewise, dacitic agglomerates of the upper Gaddes assemblage underlying "Buzzard rhyolite" on south Mingus Mountain plot very close to anomalous Si-depleted, Na-enriched dacites of the Black Canyon Creek Group. This is further evidence to strongly support correlation of "Buzzard" and upper Gaddes units on south Mingus Mountain to rocks of the Black Canyon Creek Group (fig. 4).

The youngest andesite flows of the upper Black Canyon Creek Group at the base of the Cleator Formation are distinctly more alkalic than all earlier mafic volcanics of the Prescott volcanic belt, and are clearly of calc-alkaline chemistry (fig. 5). Tuffaceous metasedimentary rocks of the Cleator Formation are also calc-alkaline, but are not shown on figure 5 because they are not strictly volcanic rocks. K-feldspar-rich rhyolite flows and dikes related to rhyolite tuffs in the Texas Gulch Formation west of Mayer are calc-alkaline in character (fig. 5), and are distinctly higher in K_2O than all preceding rocks of the Prescott volcanic belt. These relationships show that the Jerome and Prescott belts progressed with time from primitive low-K tholeiitic mafic volcanics to evolved tholeiitic and calc-alkaline rock series of higher alkali contents.

This systematic progression in alkali enrichment occurred not only in time (the oldest rocks in both belts are the least alkali enriched, whereas the youngest are the most alkali enriched) but also in space: the lowest alkali, tholeiitic rocks are confined to the western part of the Prescott volcanic belt, whereas the highest alkali, tholeiitic and calc-alkaline rocks dominate the eastern part of the belt. Therefore, there is a broad geographic sweep in alkali enrichment of volcanic rocks from west to east across the Prescott volcanic belt in parallel with a progressive younging of the formative volcanic sequences to the east (fig. 4). Thus, magmatism swept southeastward across the contiguous Prescott and Jerome volcanic belts to build a complex Proterozoic magmatic arc ranging in composition from primitive Mg tholeiites in the west and north to moderately evolved calc-alkaline volcanics in the east and south. This sweep has been interpreted to have been caused by a Proterozoic subduction zone dipping southeast under the Prescott-Jerome volcanic belt to form the volcanic cycles observed (P. Anderson, 1976, 1978a; P. Anderson and Guilbert, 1979).

VOLCANIC STRUCTURE AND EVOLUTION OF THE PRESCOTT-JEROME BELT

STRUCTURE

Both the Prescott and Jerome volcanic belts started evolution with very primitive, very low-K tholeiitic basaltic magmas that extruded in a deep-water submarine setting. Pillowed flow sequences formed extrusive caps to a stratified mafic igneous floor of massive aphyric basalt and microgabbro, and deeper, subvolcanic pyroxene gabbro. The earliest volcanic units were not deposited on a preexisting granitoid crust, but were evidently built upon a mafic-ultramafic gabbro-pyroxenite basement similar to the earliest subvolcanic rocks (P. Anderson, 1986). The mafic substratum is now so extensively intruded by plutonic rocks that its exact nature is not easily discerned (Moores, 1976).

Both the Prescott and Jerome belts have asymmetries born into them from the beginning, and subsequent volcanic growth of the belts was always asymmetrically away from their oldest portion—a rule that was never violated. The Prescott volcanic belt's oldest part is at its northwest edge near Prescott; each new volcanic center or chain of volcanic centers developed east of the preceding one, and their youngest deposits lapped back over the top of older deposits to the west. Thus the Prescott volcanic belt was built outwards asymmetrically to the east and southeast in successive stages. The Jerome volcanic belt was conceived at its northeastern edge and grew toward the Prescott belt with time, finally joining and becoming interstratified with it.

This asymmetry of the volcanic belt does not require the presence of a huge synclinorium cored with the youngest rocks. A broad synclinorium, if it exists, could influence outcrop patterns, but there is simply no evidence that the younger formative volcanic sequences were deposited on the older sequences, nor is it reasonable to assume that the volcanic belt was built up in a single, vertically stacked sequence. If so, 40 km of supracrustal volcanic deposits would exist in the Prescott belt alone above mafic basement and beneath more than 5 km of supracrustal sediments, now mostly stripped off. Even with a thin 25 km of oceanic crust as basement, the total crustal thickness would have exceeded 70 km before any plutonic crustal thickening, a figure that is unreasonably large.

Moreover, contact relationships do not permit the interpretation that the younger formative volcanic sequences were built upon a foundation of the earlier ones. First, the older units contain no dikes or subvolcanic bodies either feeding up into the younger sequences or similar to the younger sequences in petrology or chemistry; instead, subvolcanic feeders to younger extrusive deposits exist only within the younger sequences themselves. Second, the oldest units should be the most strongly migmatized and metamorphosed of all because of deepest burial and repeated igneous activity punctured through them—yet exactly the opposite is true: the oldest sequences are the least regenerated of all. Third, in a vertical succession, the base of the younger sequence should lie in contact with the top of the older sequence, but this is usually not so. Typically *only the youngest units of the younger sequences locally overlap the older ones,* a relationship impossible in a single vertical column, but not at all paradoxical in the context of figure 2.

These and other observations show conclusively that the volcanic belts are not simply homoclinally tilted and eroded wedges of a single, thick stratigraphic sequence. Their true *primary volcanic structure* is much more complex than any simple model, but is approximated by the analysis to follow of the Prescott belt's volcanic structure and evolution, as depicted in figure 7. The Prescott belt was built up in three major petrologically distinct cycles, represented by the three major rock groups defined in this paper. The Prescott belt's greater complexity and advanced evolution are absent in smaller volcanic belts such as the Jerome belt, where only a single major volcanic cycle, represented by one major rock group, is found.

EVOLUTION

The Prescott belt began evolving near Prescott, but progressed in space and time east and southeast toward Mayer and beyond. The entire belt is therefore diachronous from northwest to southeast, yet few lithostratigraphic formations and rock groups are themselves time transgressive; instead, contacts between adjacent rock groups are commonly time transgressive. The diachroneity and southeastward evolution reflect a shifting axis of volcanism

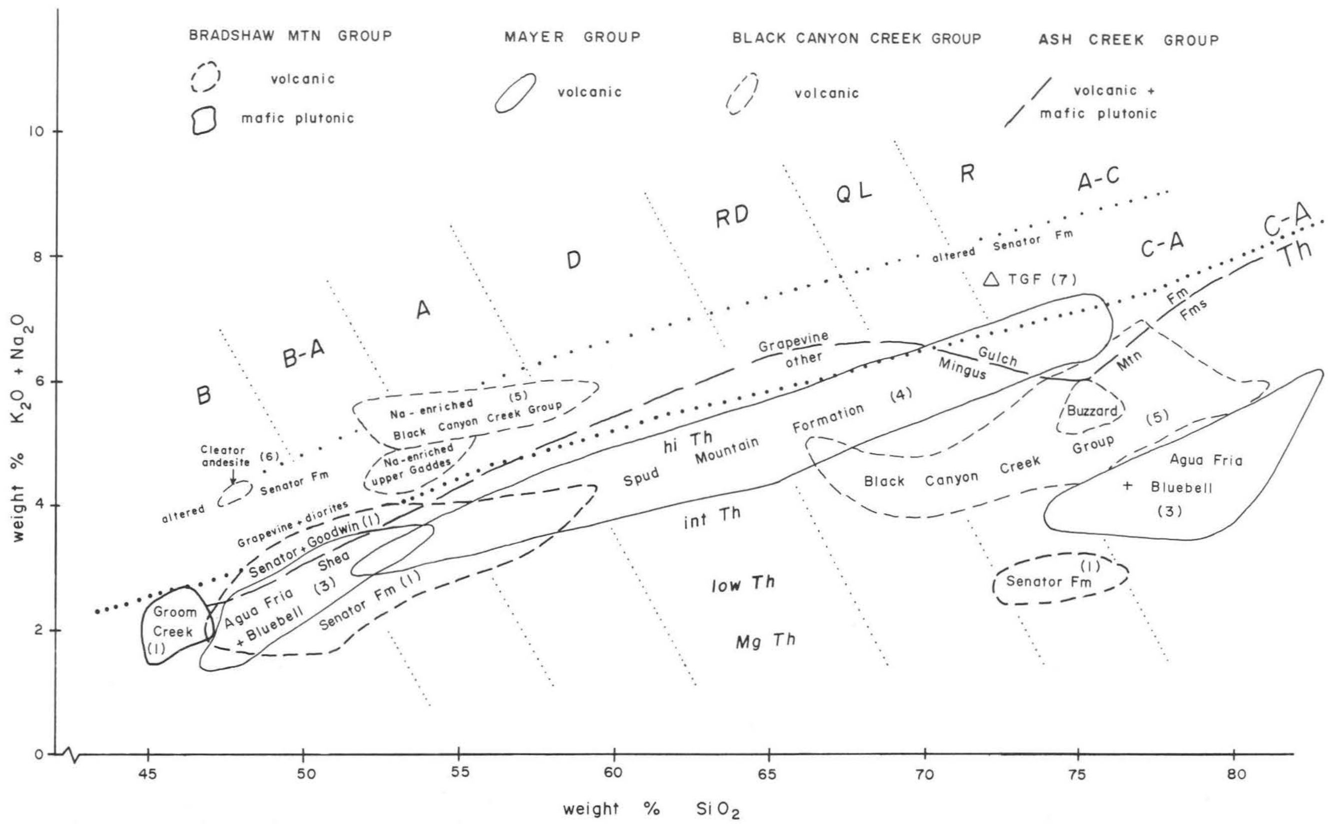


Figure 5. Alkali-silica ($K_2O + Na_2O$ vs. SiO_2) plot of volcanic rock groups and formations in the Prescott-Jerome volcanic belts. Lines enclose fields of analyses of unaltered, related rocks; analyses of altered rocks are not enclosed or are labeled as enriched. Numbers in brackets refer to the depositional stages discussed in text. Total number of analyses = 120. The long-dashed line at low silica values separates the Shea Formation below from gabbro-diorites in the Grapevine Gulch Formation above. TGF = Texas Gulch Formation rhyolite; Buzz = Buzzard rhyolite, Mingus Mountain. Dotted lines denote rock classification used by P. Anderson (1986) for the Arizona Proterozoic, which is similar to most conventional schemes in common use: B = basalt, B-A = basaltic andesite, A = andesite, D = dacite, RD = rhyodacite, QL = quartz latite, R = rhyolite. A-C = alkali-calcic, C-A = calc-alkaline, Th = tholeiitic (subdivided into: hi Th = high-alkali, int Th = intermediate-alkali, low Th = low-alkali, and Mg Th = magnesian tholeiitic).

across the belt with time, as shown in figure 7 by A, B, and C. Each letter corresponds to a major volcanic cycle in the Prescott belt's evolution. Under the new formal stratigraphic nomenclature established here, the *deposits of cycle A are the Bradshaw Mountains Group, deposits of cycle B are the Mayer Group,* and *deposits of cycle C are the Black Canyon Creek Group.*

With the major axis of volcanism at A (fig. 7), evolution of the belt began with a unique suite of mafic, magnesian, very low-K tholeiitic, aphyric basalt flows, bimodal basalt-rhyodacite, and subvolcanic masses, all derived directly or fractionally from olivine tholeiite parents. Closely following was extrusion of more aphyric bimodal basalt-rhyodacite flows and intrusion of olivine gabbro and pyroxenite masses, all of which were derived from an olivine tholeiite parent magma of slightly more K-Na-rich tholeiitic chemistry. Both mafic extrusive events comprise the Senator Formation and correlatives.

As the major axis of volcanism proceeded to shift from A to B, the mafic volcanic center at A was locally eroded on the east, causing andesitic graywackes and tuffs, felsic tuffs, and sediments to be shed into a wide, deep submarine basin to the east. The mafic and felsic tuffs originated from small andesite and rhyodacite-rhyolite centers within the basin. This depositional stage 2 is represented by the Mount Tritle Formation in the Bradshaw Mountains Group, and by the Round Hill Formation at the base of the Mayer Group; both are broadly coeval, but are somewhat different lithofacies.

Following mafic volcanism at axis A, mafic plutonism from quartz-normative hydrous magmas of high-K tholeiitic to low-K calc-alkaline chemistry ensued at A. Both the Senator and Mount Tritle Formations were intruded by diapiric gabbro-diorites of different chemistry than the earlier gabbro-pyroxenites coeval with the volcanics. Emplacement of these gabbro-diorite bodies may have been synchronous with inception of stage 3 volcanism at a new axis to the east.

Next, the major axis of volcanism shifted far enough east of the stage 1 edifices for an entirely separate volcanic center to evolve at B during depositional stage 3 (fig. 7). Phenocrystic pillow basalt flows, tuffs, and agglomerates were extruded from gabbro centers into a deep-water submarine setting during stage 3 and were interbedded with

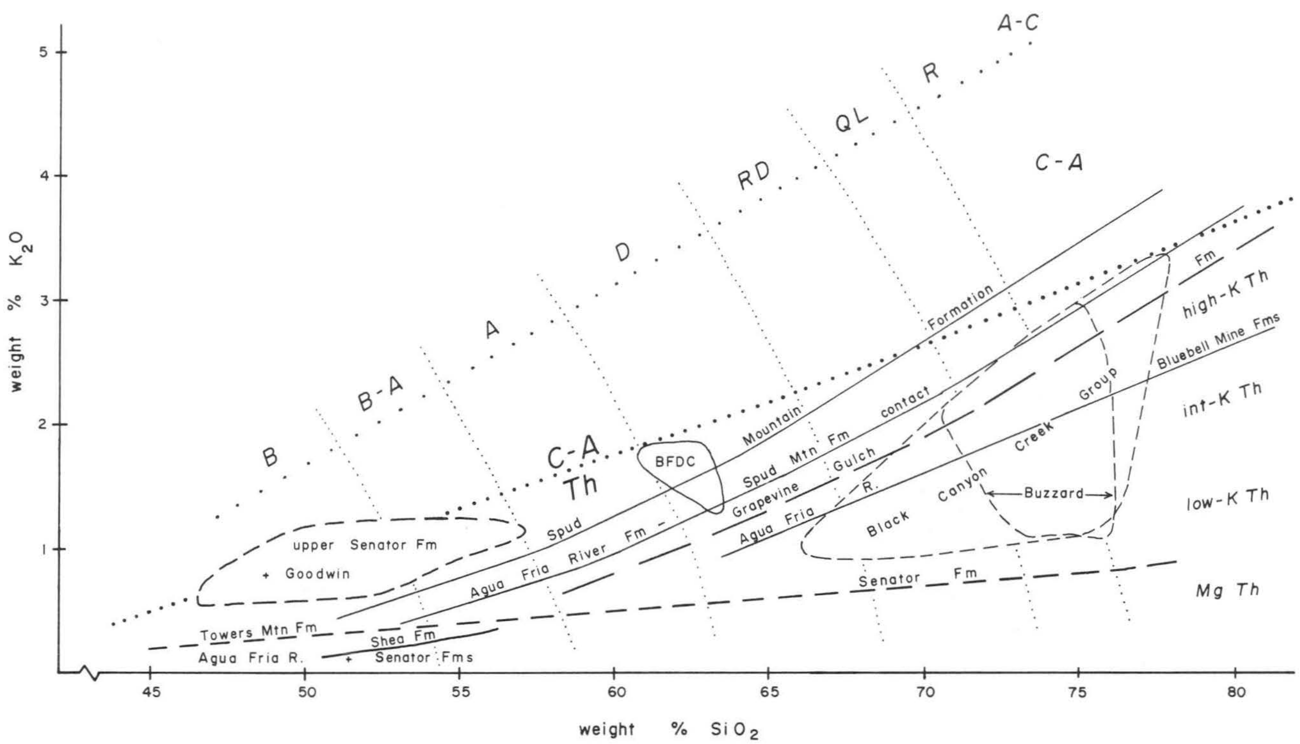


Figure 6. K_2O vs. SiO_2 plot showing K-enrichment trends of volcanic rock groups and formations in the Prescott and Jerome volcanic belts. Each line represents an average trend of related mafic and felsic volcanic rocks. BFDC = Battle Flat dacite center analyses. Rock classification scheme is the same described in figure 5 except that the tholeiitic field is subdivided into high-K tholeiitic, intermediate-K tholeiitic, and low-K tholeiitic; these terms are used in a relative sense and not meant to imply K_2O contents as high as calc-alkaline.

rhyolite flows, tuff, and iron formation. The earliest part of the mafic accumulation was the Bluebell Mine Formation, then the Peck Canyon and Towers Mountain Formations in the south, then the Agua Fria River Formation in the north. The lowest units are low-alkali tholeiitic, normal Fe-rich bimodal basalt-rhyolite derived by differentiation of quartz-tholeiite parent magmas. The tholeiitic parent underwent fractionation by removal of early pyroxene phases to produce a higher-K tholeiitic basaltic andesite source magma dominated by plagioclase. This new source underwent further fractional crystallization to produce high-K tholeiitic andesites and rhyolites in the upper Agua Fria River Formation.

At this point an important trend in the lithologic makeup of the entire volcanic belt was established. The preceding large jump in the major axis of volcanism from position A, where the Bradshaw Mountains Group developed, to site B, where the Mayer Group evolved, left a gap or deep trough in the supracrustal deposits which was only partly filled by distal strata interleaved at the edges of both major volcanic edifices. Thus, volcanism from the second major cycle at site B proceeded to backfill the trough in a transgressive manner that caused only the youngest deposits of the Mayer Group—the Spud Mountain Formation—to lap unconformably back over older sequences to the west, including Mount Tritle and Senator Formations plus their gabbro-diorites.

Plagioclase-crystal dacite-andesite breccias in lowest parts of the Spud Mountain Formation are petrochemically related to waning Agua Fria River volcanism, whereas middle and upper parts were derived from an evolved tholeiitic to calc-alkaline differentiated magma extruded from the Battle Flat dacite center. Most pyroclastics were shed northward from the Battle Flat center to fill the intervolcanic trough and lap over older rocks to the east and west, leaving the Spud Mountain Formation as a laterally discontinuous rock package with older mafic volcanics on either side.

The major volcanic axis took another large jump from positions B to C after Mayer Group volcanism ceased, and the same pattern of establishing a new volcanic center far to the east and subsequently backfilling the intervening intervolcanic trough to the west was repeated at axis C. The first new extrusives of the Black Canyon Creek Group were dacite-rhyodacite flows, tuffs, and breccia of phenocrystic character similar to Spud Mountain rocks, but with different chemistry. The first phenocrystic agglomerate and conglomerate heralded an impending major outpouring of vast quantities of aphyric felsic agglomerates and breccias across the eastern Prescott and Jerome belts, which had been stratigraphically linked long before Spud Mountain time.

As the Copper Mountain center evolved into a large subaerial felsic complex east of Mayer, the Prescott-Jerome

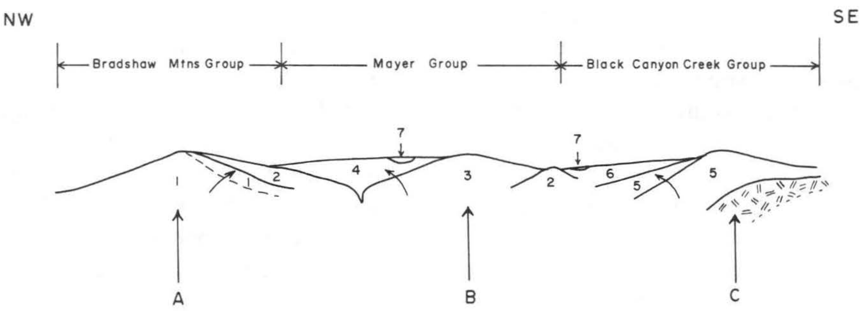


Figure 7. Highly simplified, diagrammatic representation of the original volcanic structure and evolution of the Prescott volcanic belt, as interpreted from rock sequences in their present deformed (horizontally shortened) state. The volcanic belt is presently 40 km wide, but was originally much wider. The numbers represent depositional stages in which particular rock sequences were deposited, as discussed in the text. Letters A, B, and C show sequential positions of the major axis of volcanism as it shifted east-southeast across the belt, and correspond to the three major volcanic cycles of the Prescott belt.

volcanic belt became emergent for the first time. Subaqueous valley flows of felsic fragmentals spread south down the Black Canyon belt to interface with andesites in the south, some agglomerates spread north over older strata of Mingus Mountain, and younger subaerial felsic agglomerates were shed in chaotic arrays to the north. All such felsic rocks have a distinctive Na-enriched, K-depleted geochemical pattern due to intrinsic chlorite-sericite-albite alteration characteristic of the magma series. Consequently, some rocks plot as tholeiitic, but their true chemistry is almost certainly calc-alkaline, as are coeval unaltered andesites.

Tuffaceous sediments of the younger Cleator Formation were deposited in the trough intervening between the two major volcanic edifices at positions B and C. As the trough was backfilled to the west, the sediments transgressed unconformably over the *oldest,* not the youngest, easterly mafic parts of the Mayer Group (fig. 3). Thus, like most units in the Prescott belt, the Cleator Formation is an unconformity-bounded depositional wedge, being elongated north-south parallel to its depositional basin, laterally discontinuous east-west, and *bounded on both sides by different older units. In this lensoidal and unconformable manner, the entire Prescott volcanic belt is made up of contact relations that defy analysis using concepts of a single vertically superimposed stratigraphic sequence.*

Evolution of the Prescott volcanic belt has been described in terms of a shifting axis of volcanism, because it truly was an axis and not just the locus of a single volcanic center that shifted. The axis is clearest in stage 3 Mayer Group mafic volcanism, where at least three volcanic centers, aligned on a north-south axis from north of Mayer to south of Crown King, contributed to the cycle. Similar axial trends for the other cycles are evident from a broad perspective.

The Jerome volcanic belt evolved mostly during the early stages of the Prescott belt, and the deposits bear many petrologic similarities to each other. However, the Ash Creek Group does not uniquely coincide with a single early magmatic cycle as does the Bradshaw Mountains Group, because of the way the Ash Creek Group was originally defined to include younger formations. The Jerome and Prescott belts were neighboring volcanic belts from the outset, were linked stratigraphically almost from their inception, and were not later juxtaposed by a strike-slip fault. The youngest depositional event recognized in both belts is represented by the Texas Gulch Formation, with its associated rhyolitic calc-alkaline igneous activity. The Texas Gulch Formation postdated emplacement and unroofing of calcic granodiorite plutons and tonalite batholiths, so it was a successor sequence deposited in narrow troughs at the edges of batholiths and plutons, and was not formative to the primary igneous evolution of volcanic belts. Lastly, fluvial Mazatzal strata were deposited north of the belts.

CONCLUSIONS

Several features of major importance in the volcanic structure and evolution of the Prescott belt are apparent from the foregoing analysis:

(1) There is a cyclical pattern to volcanic evolution of the belt. When volcanic activity had built up deposits of a particular geochemical-petrologic character along a major axis, new volcanic activity jumped far to the southeast to build up a new volcanic axis of a different geochemical-petrologic character. This left an intervening trough that was backfilled by younger deposits of the new cycle. The pattern repeats twice in the Prescott-Jerome belt, and may also repeat again to the south [see Part 2].

(2) Rocks formed during shifts of the volcanic axis are transitional in nature.

(3) The three major rock groups now recognized—Bradshaw Mountains, Mayer, and Black Canyon Creek Groups—uniquely coincide with subdivision of the belt into the three major igneous cycles described above. Consequently, the new Group divisions recognize profoundly important stages in evolution of the Prescott volcanic belt, as well as identify lithostratigraphically unique rock sequences, and are thus fundamental subdivisions of the Prescott belt.

(4) The belt is diachronous from west to east, with each volcanic cycle (major rock group) being staggered in sequentially younger stages from west to east.

(5) There is an overall chemical evolution from primitive tholeiitic magmas to partly evolved high-K tholeiitic to low-K calc-alkaline magmas from west to east with time, but the belt never reached highly evolved calc-alkaline and alkali-calcic stages of the younger volcanic belts in east-central Arizona [see Part 2].

(6) Irreversible petrochemical evolution of the crust occurred as the Prescott belt developed: each deposit is successively more evolved than preceding ones, with no returns to the primitive conditions represented by earlier stages.

(7) The sequential igneous stages that evolved from northwest to southeast across the Prescott belt represent one broad magmatic sweep that is readily accounted for by a single event of 1800- to 1740-Ma Proterozoic subduction of oceanic lithosphere that once dipped southeast under the Prescott and Jerome vocanic belts [see tectonics paper].

PART 2—EARLY PROTEROZOIC STRATIGRAPHY AND VOLCANIC EVOLUTION OF THE NEW RIVER—CAVE CREEK—MAZATZAL MOUNTAINS—DIAMOND BUTTE VOLCANIC BELTS

The older Prescott-Jerome volcanic belts [see Part 1] command special importance as the earliest formed crust in central Arizona, but their formative 1800- to 1740-Ma history accounts for only half of the central volcanic belt's complete evolution. The later 1740- to 1680-Ma history is best displayed in the younger New River-Cave Creek-Mazatzal Mountains-Diamond Butte belts (fig. 1); their evolution complements that of the older Prescott-Jerome belts, taking over from where the older belts left off in terms of stratigraphy, petrology, and chemistry. Contrasts between the older and younger belts provide a clear picture of how the entire Proterozoic crust of central Arizona evolved [see tectonics paper].

The southeast limit of the Prescott volcanic belt near Black Canyon City closely approaches the New River volcanic belt (figs. 1 and 3), providing critical relationships that link the older and younger volcanic sequences of central Arizona. The Cave Creek volcanic belt and Union Hills area (fig. 1) contain key volcanic assemblages of major importance to the origin of the younger volcanic belts. Yet, prior to this work, the New River and Cave Creek belts had not been studied. The Mazatzal Mountains-Diamond Butte belts have a diversity of rocks that were studied only locally in the past (Wilson, 1939; Gastil, 1958; Ludwig, 1974; Conway, 1976), but which have now been investigated over the entire region (P. Anderson and Wirth, 1981; P. Anderson, 1986).

MAJOR STRATIGRAPHIC ASSEMBLAGES

Throughout the New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts, five broadly different stratified rock sequences are present: (1) an oldest assemblage of mafic, intermediate, and felsic volcanic rocks, plus related tuffs and fragmentals, deposited around volcanic centers; (2) older clastic sediments of volcanic origin, broadly time equivalent to the oldest volcanic rocks, but deposited distally from major volcanic centers; (3) a younger purple shale and quartzite sequence with interbedded felsic tuff that unconformably overlies the older sequences; (4) a young assemblage of felsic ignimbrites and extrusive pyroclastics that unconformably overlies the older volcanic and sedimentary units, and which is related to intrusive hypabyssal felsic volcanic and plutonic facies; and (5) a youngest suite of relatively mature quartzite and conglomerate that unconformably overlies a partly eroded terrain of the older rock sequences.

Wilson (1939) named the youngest sequence the **Mazatzal quartzite** for its principal exposures in the Mazatzal Mountains. This regionally important sequence has now been elevated to **Mazatzal Group** (P. Anderson and Wirth, 1981; P. Anderson, 1986). Also in the Mazatzal Mountains, Wilson (1939) first named part of the felsic ignimbrite sequence **"Red Rock Rhyolite"** and named the younger sedimentary sequence (3 above) **"Alder series."** Both Wilson (1939) in the Mazatzal Mountains and Gastil (1958) near Diamond Butte noted the presence beneath the "Alder series" of mafic volcanics that are lithostratigraphically distinct from Alder strata, but did not name these oldest volcanic rocks.

Since then, however, the term "Alder" has been generalized to include all rocks older than felsic ignimbrites in the Mazatzal Mountains-Diamond Butte areas, including the pre-Alder mafic volcanics (Conway and Wrucke, 1986; Conway and Silver, this volume). This usage cannot be condoned because the pre-Alder volcanic rocks in the younger belts are thick, diverse, extensive deposits with major stratigraphic, petrologic, and age differences with Alder strata. Usage of "Alder Series" in a time sense (C. A. Anderson and others, 1971) conflicts with Wilson's (1939) original definition and is also not condoned. Wilson's original "Alder series" described a lithostratigraphically valid and highly distinctive sedimentary sequence, so the name is retained to describe just that sequence and its true stratigraphic correlatives. "Alder" has been elevated to group status and is now formally recognized as the **Alder Group** (P. Anderson, 1986).

AGE RELATIONSHIPS

Isotopic dates in the younger belts are limited to the Diamond Butte and Mazatzal Mountains areas, mainly in the felsic ignimbrites. Dates (adjusted with new constants of Steiger and Jaeger, 1977) range from 1710 Ma on dacites in the upper Alder Group (Ludwig, 1974) to 1630 Ma on posttectonic plutons (Silver, 1967). New dates of 1700 Ma on ignimbrites and 1703 Ma on related red granites (Silver and others, 1986) support the coevality that has been apparent from field relationships since 1958 (Gastil, 1958; P. Anderson, 1986). *Most importantly however, none of the older mafic volcanic rocks that predate the Alder Group have yet been dated in the younger volcanic belts.* Based on stratigraphic relationships, the undated volcanic rocks formed between 1740 and 1710 Ma, a period long believed to be a gap in volcanism (Silver, 1967). Detailed relative-age relations now established for central Arizona show that volcanism was more continuous than published isotopic ages indicate and that the "gap in volcanism" is mainly an artifact of the dating (P. Anderson, 1986).

The key relative-age relationships are summarized below and closely referenced to figures 8 and 9. The westerly New River-Cave Creek volcanic belts are considered first because they contain critical links to the older Prescott-Jerome volcanic belt.

NEW RIVER VOLCANIC BELT

Felsic volcanics of the Black Canyon Creek Group [see Part 1] are interbedded with andesite-dacite flows and tuffs from an andesitic volcanic center near Black Canyon City. This andesite-dacite sequence is widely exposed in the southern Black Canyon belt, west of Moore Gulch, and in the western New River area (figs. 1 and 8), where it contains Fe-Mn-rich, shaly iron-formation beds correlative to those in the Hieroglyphic Mountains. West of Moore Gulch the andesite-dacite sequence is intruded, metamorphosed, and migmatized by the southeast edge of the Cherry Springs batholith [see Part 3], forming an extensive migmatite complex that continuously borders the southeast edge of the batholith. This newly named **Little Squaw Creek Migmatite Complex** is evidence that the andesite-dacite sequence predates 1740- to 1720-Ma (Bowring, 1986) phases of the Cherry Springs batholith.

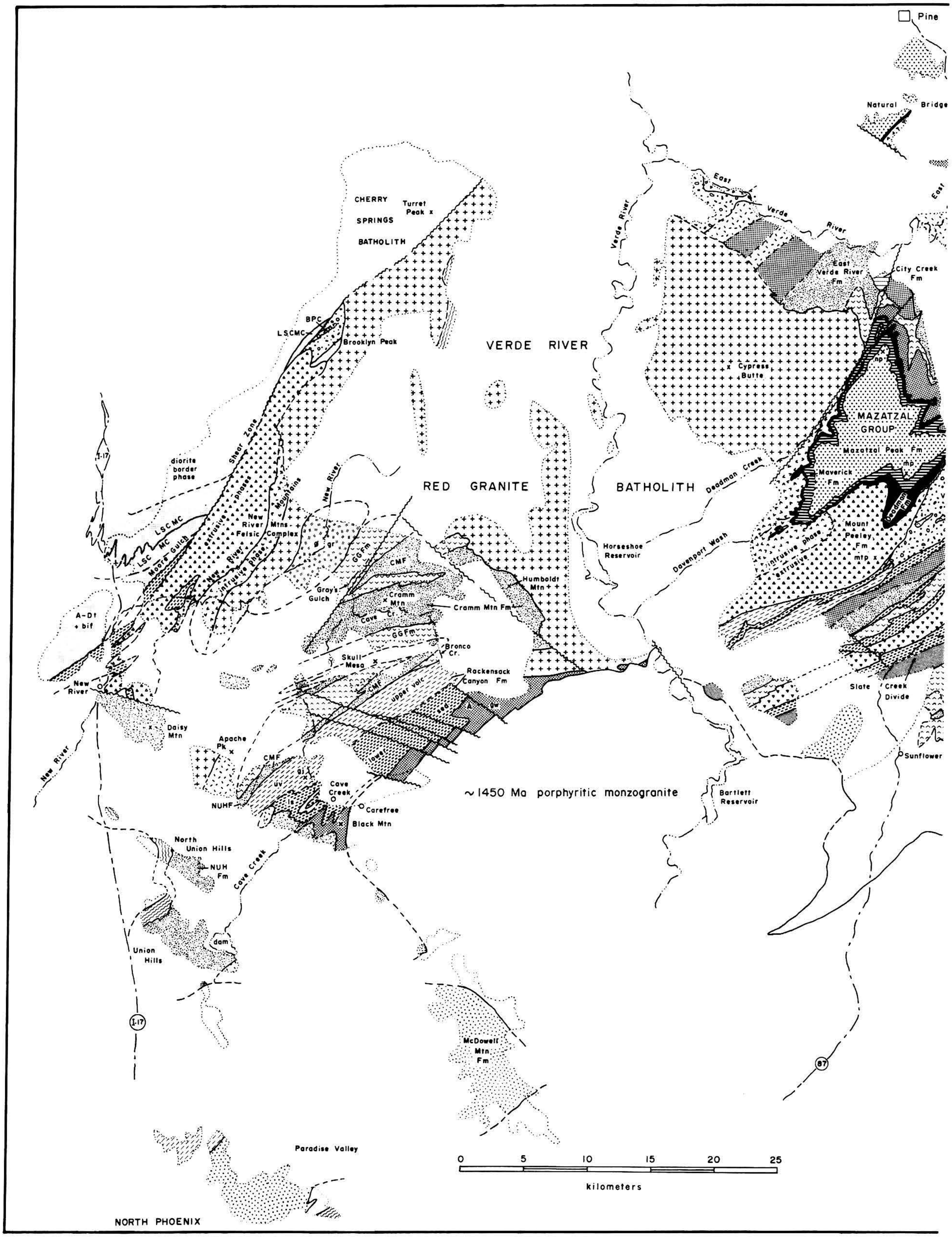
The Little Squaw Creek Migmatite Complex is overlain in Moore Gulch by an andesite-rhyolite tuff sequence, which is in turn overlain unconformably to the east by a purple slate and phyllite sequence. These two sedimentary sequences are vitally important, because they or their equivalents occur in all younger volcanic belts and delimit two major depositional hiatuses in the stratigraphic evolution of the belts. The purple slates are in turn overlain unconformably by rhyolite breccia, agglomerate, tuffs, and flows that grade up to welded ash-flow tuffs toward the top of the New River Mountains.

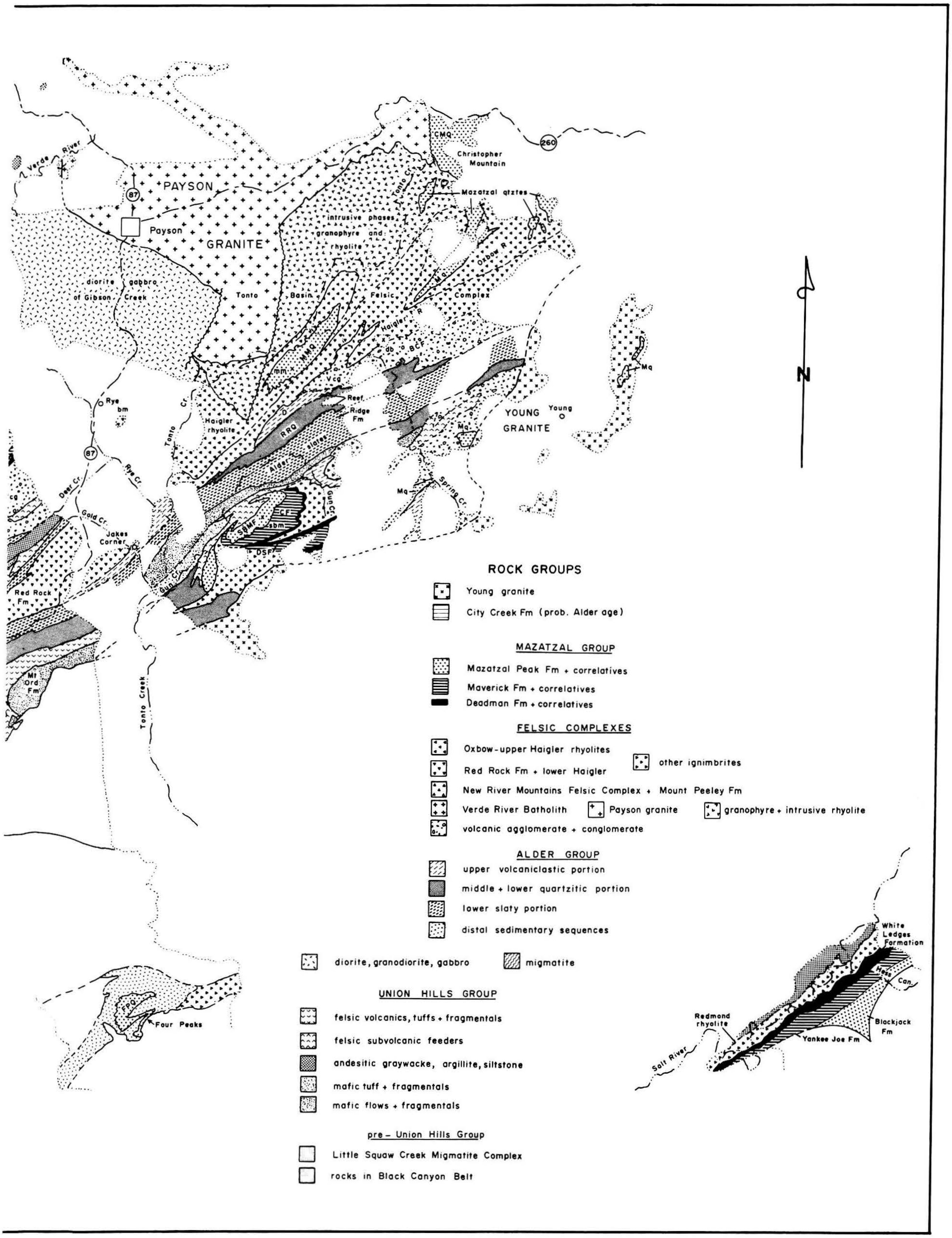
The Moore Gulch shear zone, a major tectonic element of the New River volcanic belt that extends for more than 50 km along the east side of the Cherry Springs batholith (fig. 8), obscures this unconformable relation of the rhyolites and ignimbrites on the purple slate sequence in Moore Gulch. But farther south near New River town, clear unconformable relations are preserved where the rocks are less strained. Progressing farther north in Moore Gulch, first the purple slates then the tuffaceous rocks are attenuated as strain intensifies northward in the Moore Gulch shear zone; finally rocks of the Little Squaw Creek Migmatite Complex are juxtaposed directly in fault contact with the felsic volcanics along the shear zone.

NEW RIVER MOUNTAINS FELSIC COMPLEX

These youngest felsic volcanics are exposed over a huge region east of the Moore Gulch shear zone, including all of the New River Mountains, as well as areas to the north and south. The western New River Mountains contain principally volcanic rocks—megabreccia, agglomerate, tuff, lahars, lithic-crystal ash-flow and ash-fall tuff, flows,

Figure 8. Simplified map of the Proterozoic geology of the New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts showing the distribution of the major rock groups defined in this paper. All units are discussed in the text, and intervening blank areas are either postvolcanic plutonic rocks (heavy solid lines) or middle Proterozoic Apache Group and younger cover (thin dotted lines). Note the orientations of symbols for the felsic ignimbrites: upward-pointing triangles denote rocks related to the Verde River Granite batholith; downward-pointing triangles are for the Red Rock Rhyolite and its more easterly correlatives; left-pointing triangles signify upper Haigler and Oxbow rhyolites; and right-pointing triangles include all other ignimbrites, mostly those to the east. Abbreviations are as follows: **Small letters:** bm = Black Mountain (Payson area); db = Diamond Butte; gj = Go John Mountain; mm = McDonald Mountain; mp = Mazatzal Peak; mtp = Mount Peeley; np = North Peak; sbm = Sheep Basin Mountain. **Rock types:** A-Dt + bif = andesitic-dacitic tuff and banded iron formation involved in the Little Squaw Creek Migmatite Complex; A gw = andesitic graywacke; D = dacitic units of the upper part of the Alder Group; ls = lower sed = lower sedimentary portion of the Alder Group; uv = upper volc = upper volcanic portion of the Alder Group; V cg = felsic volcanic conglomerate-agglomerate; ∅ gr = porphyritic granite-monzogranite core phase of New River Felsic Complex. **Capital letters:** BCF = Board Cabin Formation; BPC = Brooklyn Peak Conglomerate; CF = Coffeepot Canyon facies; CMF = Cramm Mountain Formation; CMQ = Mazatzal quartzite of Christopher Mountain; DSF = Del Shay facies; FPQ = Mazatzal quartzite of Four Peaks; GGF = Gray's Gulch Formation; LSC MC = Little Squaw Creek Migmatite Complex; MMQ = Mazatzal quartzite of McDonald Mountain; Mq = Mazatzal quartzite; NUHF = North Union Hills Formation; RRQ = quartzites of the Reef Ridge Formation; SBMF = Sheep Basin Mountain facies.


Pine
Natural Bridge
East Verde River
Verde River
CHERRY SPRINGS BATHOLITH
Turret Peak
BPC
LSCMC
Brooklyn Peak
VERDE RIVER
RED GRANITE
BATHOLITH
Cypress Butte
City Creek Fm
East Verde River Fm
MAZATZAL GROUP
Mazatzal Peak Fm
Maverick Fm
Deadman Creek
Deadman Fm
Mount Peeley Fm
Horseshoe Reservoir
Davenport Wash
diorite border phase
Shear Zone
New River Mountains
New River Mtns Felsic Complex
Moore Gulch
LSC MC
Humboldt Mtn
CMF
GG Fm
Gray's Gulch
Cramm Mtn
Cramm Mtn Fm
Cave Cr.
Bronco Cr.
Skull Mesa
Rackensack Canyon Fm
A-Dt + bif
New River
Daisy Mtn
Apache Pk
upper vol.
lower
Cave Creek
Carefree
Black Mtn
NUHF
North Union Hills
NUH Fm
Union Hills
dam
I-17
~1450 Ma porphyritic monzogranite
Bartlett Reservoir
Slate Creek Divide
Sunflower
McDowell Mtn Fm
87
Paradise Valley
0 5 10 15 20 25
kilometers
NORTH PHOENIX


PAYSON
GRANITE
Payson
Christopher Mountain
Mazatzal qtztes
intrusive phases granophyre and rhyolite
Tonto Basin Felsic Complex
Oxbow R
diorite of Gibson Creek
gabbro
Haigler R
YOUNG GRANITE
Young
Haigler rhyolite
Reef Ridge Fm
Alder slates
Rye
Tonto Cr
Rye Cr
Deer Cr
Gold Cr
Jakes Corner
Gun Cr
Spring Cr
Red Rock Fm
Mt Ord Fm
Tonto Creek
Four Peaks
Verde River
Salt River
Redmond rhyolite
White Ledges Formation
Blackjack Fm
Yankee Joe Fm
ROCK GROUPS
Young granite
City Creek Fm (prob. Alder age)
MAZATZAL GROUP
Mazatzal Peak Fm + correlatives
Maverick Fm + correlatives
Deadman Fm + correlatives
FELSIC COMPLEXES
Oxbow-upper Haigler rhyolites
Red Rock Fm + lower Haigler
other ignimbrites
New River Mountains Felsic Complex + Mount Peeley Fm
Verde River Batholith
Payson granite
granophyre + intrusive rhyolite
volcanic agglomerate + conglomerate
ALDER GROUP
upper volcaniclastic portion
middle + lower quartzitic portion
lower slaty portion
distal sedimentary sequences
diorite, granodiorite, gabbro
migmatite
UNION HILLS GROUP
felsic volcanics, tuffs + fragmentals
felsic subvolcanic feeders
andesitic graywacke, argillite, siltstone
mafic tuff + fragmentals
mafic flows + fragmentals
pre – Union Hills Group
Little Squaw Creek Migmatite Complex
rocks in Black Canyon Belt

and columnar-jointed ignimbrites—of quartz latite to rhyodacite composition. The New River peaks are occupied by black, vitreous, welded quartz latite ignimbrite that represents the topmost part of the felsic complex. The eastern New River Mountains, however, are dominated by hypabyssal and plutonic phases of the same felsic complex —extrusive quartz latite grades downward into massive, hypabyssal intrusive quartz latite and granophyre, and textures gradually coarsen eastward into fine-grained plutonic rock. Farther east in upper New River, the rock grades into coarse-grained, porphyritic monzogranite.

Thus, a continuous gradation from volcanic to plutonic textures in rocks of quartz latite-monzogranite composition occurs easterly across the New River Mountains. This gradation represents a cross section from the upper extrusive parts to subvolcanic hypabyssal parts to deeper plutonic parts of a huge felsic complex. All such parts are broadly coeval, even though discordant relations can be found locally in the upper bedded portions. This large felsic volcano-plutonic complex is newly and formally named the **New River Mountains Felsic Complex** and is a major component in the younger evolution of the New River volcanic belt. The complex is exposed over a length of 50 km and a width of 10 km in the New River Mountains. On a broader scale, however, the complex is just one extrusive manifestation of the vast Verde River Granite batholith, which extends for 50 km in an east-west direction between the New River and Mazatzal Mountains (fig. 8).

RELATIVE-AGE RELATIONS

To the north near Brooklyn Peak, one of the most crucial age relations in the Proterozoic of Arizona is exposed. At the eastern margin of the Cherry Springs batholith, in locally sheared depositional contact with the batholith and with amphibolites of the Little Squaw Creek Migmatite Complex, is a virtually unmetamorphosed sequence of felsic tuff, conglomerate, and volcaniclastic rocks. The **Brooklyn Peak Conglomerate** lies within this sequence and contains boulders of hornblende-biotite tonalite derived from the batholith, which requires that the Cherry Springs batholith was uplifted and unroofed before the conglomerate was deposited. The volcaniclastic sequence, including arenite and quartzite beds, is unconformably overlain by breccia, ash-flow tuff, and columnar-jointed ignimbrite of the New River Mountains Felsic Complex (fig. 9).

To the south near New River, conglomerates in stratigraphic positions similar to that of the Brooklyn Peak Conglomerate delimit local troughs filled by felsic volcanic debris prior to the main quartz latite flows and ignimbrite eruptions. South of New River near Daisy Mountain, bedded felsic breccias and ignimbrites of the New River Mountains Felsic Complex are separated from the older rocks by purple slates or faults. The older andesite-dacite-iron formation sequence of the Black Canyon Belt occurs only west of New River; to the east, an entirely different basaltic andesite-andesite-rhyolite flow, tuff, and breccia sequence, with flanking andesitic graywackes, takes its place. This mafic sequence east of New River is stratigraphically younger than the mafic volcanic sequence in the Black Canyon Belt and is widely exposed in the Cave Creek and Union Hills areas.

UNION HILLS AREA

In the Union Hills southeast of the New River volcanic belt (fig. 8) is a well-preserved suite of mafic volcanic rocks and graywackes whose importance to the Arizona Proterozoic has been previously unrecognized. The North Union Hills contains almost unmetamorphosed, finely granular basaltic andesite to andesite flows with chilled bases, microdiorite cores, glassy and variolitic silicic upper portions, and vesicular flow tops. Compositional differentiation from basaltic andesite to andesite to dacite is present. The flows are interleaved with cogenetic andesitic breccias and tuffs and are laterally interstratified with thick wedges of andesitic graywacke. Most flows are aphanitic, few are phyric, and all show evidence of original pyroxene, but no olivine, phenocrysts.

In the Union Hills, the same sequence is somewhat more metamorphosed and intruded by granitic rocks, but contains essentially similar components. Thick units of locally pyroxene-and plagioclase-phyric basaltic andesite and andesite flows are interbedded with thinner dacite, albitic rhyodacite, iron formation, andesitic tuff, and breccia units, all of which are laterally interstratified with thick wedges of crudely bedded andesitic graywacke.

Small diorite and hornblende granodiorite plutons intrude the mafic volcanic sequence of the Union Hills, New River, and Cave Creek areas, as well as the slaty sequences in these areas that are stratigraphically equivalent to the Alder Group in the Mazatzal Mountains (fig. 9). The diorite-granodiorite suite is close in age to the Alder Group and therefore significantly postdates the 1740- to 1720-Ma Cherry Springs batholith, which the Alder Group overlaps. The main basaltic andesite and graywacke sequence of the Union Hills could possibly predate the Cherry Springs batholith, but is nowhere intruded by the batholith. It is more likely that the mafic volcanic sequence postdates the earliest 1740-Ma batholith phases and predates the younger 1720-Ma phases. The Union Hills mafic volcanic rocks appear to stratigraphically overlie the Black Canyon andesite-dacite-iron formation suite, and the two volcanic suites are entirely different in lithology, petrology, chemistry, and geologic setting.

These relative-age relationships indicate that the Union Hills mafic volcanics formed after 1745-Ma volcanics of the Prescott belt, long before 1710-Ma volcanic conglomerates in the upper Alder Group, and almost certainly prior to younger 1720-Ma phases of the Cherry Springs batholith. Thus, Union Hills volcanics evolved in the period 1745 to 1720 Ma, a period formerly believed to be a gap in

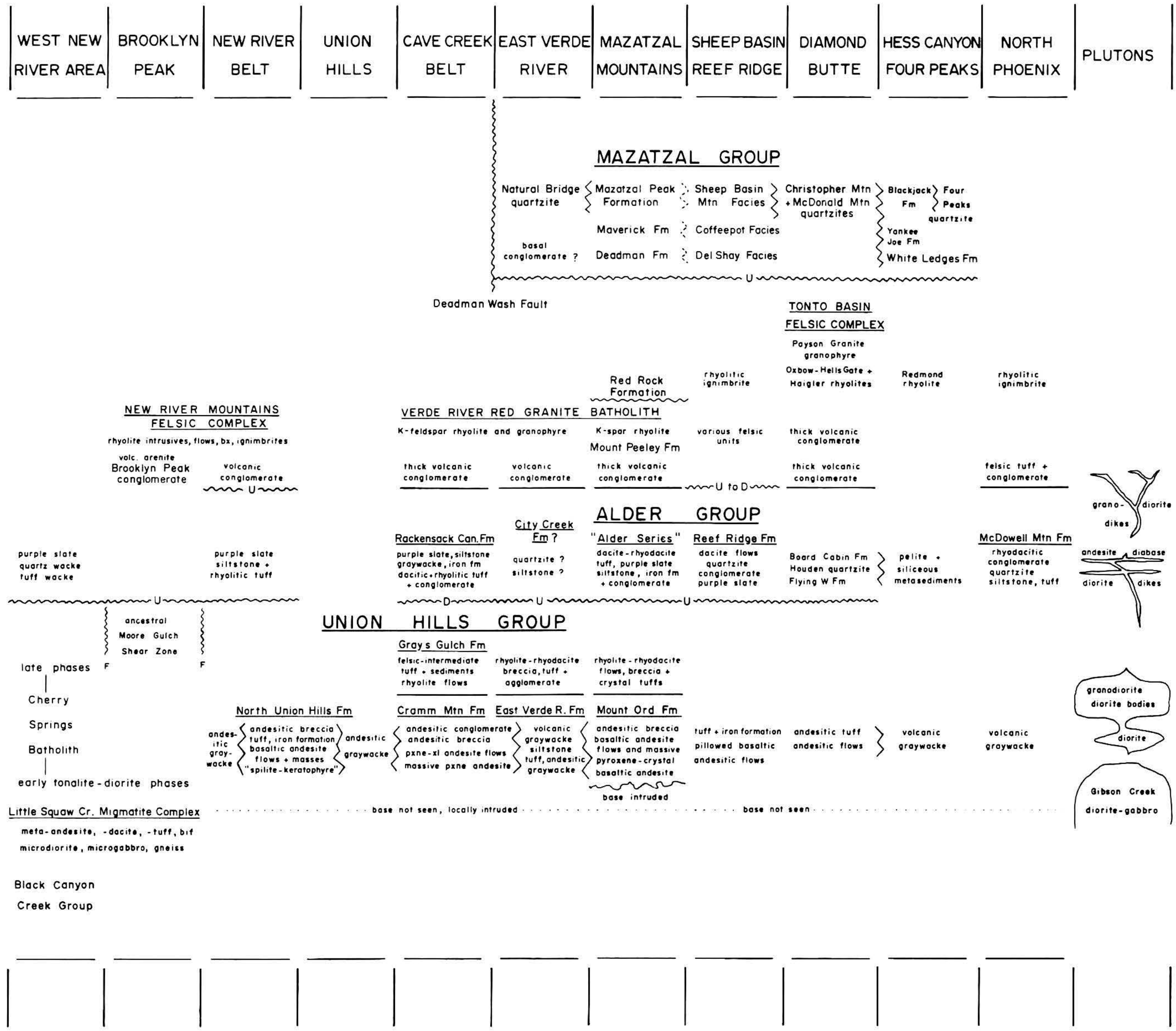


Figure 9. Tabular summary of stratigraphic relations in the younger eastern volcanic belts of central Arizona. Geologic relations are tabulated according to the areas shown at the top of the table; the positions of mafic plutons at the right correspond to the stratigraphic levels into which they intruded. The wavy lines depict either unconformities (U), disconformities (D), or structural (fault) boundaries to deposition (F), whereas straight lines depict conformable contacts. The wedge-shaped edges of lithologic packages in the Union Hills Group depict the major lithologic facies changes within and between various formations. Minor lithologic changes between formations of the Mazatzal Group and their direct easterly facies equivalents are depicted by dashed lines. Figure 10 shows facies changes in the Mazatzal Group in more detail.

Proterozoic volcanism in Arizona. Because of the importance of these volcanic rocks to a new tectonic-evolutionary understanding of central Arizona, and because they are best exposed in the Union Hills and North Union Hills, a new formal name—the **Union Hills Group**—is assigned to all mafic volcanics in the younger central Arizona volcanic belts that are lithostratigraphically correlative to those in the Union Hills and North Union Hills. The type section of the mafic volcanic suite is the North Union Hills, so the basaltic andesite and graywacke sequence in those hills is newly named the **North Union Hills Formation.**

Union Hills Group mafic volcanics are widespread throughout all younger central Arizona volcanic belts: they are the mafic volcanic rocks that Wilson (1939) recognized in the Mazatzal Mountains and Gastil (1958) recognized in the Diamond Butte area as predating Alder sedimentary strata, the same mafic volcanic units that others (e.g., Conway and Wrucke, 1986) have erroneously included in the Alder Group. The mafic volcanic rocks in the easterly volcanic belts are assigned to new formations of the Union Hills Group, but the thickest and most complete section of the mafic volcanics occurs in the Union Hills area, where the **Union Hills Group** was first formally recognized as a key stratigraphic component of the younger volcanic belts (P. Anderson, 1986).

CAVE CREEK VOLCANIC BELT

As in the Union Hills, the oldest units in the Cave Creek volcanic belt are basaltic andesite to andesite flows and breccias near volcanic centers and andesitic graywackes distal from the centers. These rocks are overlain by andesite-rhyolite flows, tuff, and iron formation at middle stratigraphic levels and by purple slate, graywacke, and conglomerate at upper stratigraphic levels. Felsic volcanic rocks of the New River Mountains Felsic Complex are absent, but plutonic phases of the complex intrude tuffs at the northern edge of the Cave Creek volcanic belt (fig. 8). Small plutons and dikes of diorite and granodiorite, equivalent in age to those noted earlier, cut some stratified units of the belt (fig. 9).

Mafic volcanic sequences dominate the central Cave Creek belt and are superbly exposed in upper Cave Creek. Cramm Mountain is a mineralized vent of a major volcanic center of pyroxene-phyric basaltic andesite and plagioclase-phyric andesite. Coarsely fragmental tephra with blocks as large as 4 m were shed from the slopes of this volcanic edifice to accumulate as lenses of autogenous breccia, polymictic agglomerate, and volcanic conglomerate in a wide region extending to the southwest and east to Humboldt Mountain. The upper Cave Creek drainage provides a classic section through the midst of an andesitic stratovolcano. These mafic volcanic deposits are newly grouped into a single formation, named the **Cramm Mountain Formation** for its Cramm Mountain type section. The Cramm Mountain Formation is a key part of the Union Hills Group, like the North Union Hills Formation to the south, and both formations appear to be broadly coeval.

The coarse mafic fragmentals fine upward to andesite tuff, which is interbedded with rhyolitic flows, rhyolitic breccia, agglomerate, conglomerate, and felsic tuff. These volcanic strata are overlain by a diversity of reworked felsic tuff and volcaniclastics, hematitic chert, iron formation, and shaly tuff, which are in turn overlain by clastic units not of direct volcanic derivation. This tuffaceous sequence is newly named the **Grays Gulch Formation** for its Grays Gulch type section, and is the uppermost formation of the Union Hills Group in the Cave Creek belt (fig. 9). The formation includes a lower silicic rhyolitic portion and an upper tuffaceous sedimentary portion; the gradual change from volcanic to reworked deposits with time implies a gradational boundary between the Cramm Mountain Formation and Grays Gulch Formation that varies in lithology from place to place. The Grays Gulch Formation occurs primarily in the north-central part of the Cave Creek belt and is minor to absent in the south.

Instead, south of Skull Mesa is a dominantly sedimentary sequence of well-bedded, graded-bedded purple slate, siltstone, and graywacke with lesser dacite flows, feldspar-crystal tuff, rhyolitic tuff, wacke, and limestone. This sequence, newly named the **Rackensack Canyon Formation** for its type section in Rackensack Canyon north of Carefree, unconformably overlies the Grays Gulch and Cramm Mountain Formations in windows along Bronco Creek (fig. 8). Its dominant sedimentary lithology and direct correlation to Alder Group strata in the Mazatzal Mountains makes it the main component of the Alder Group in the Cave Creek volcanic belt (fig. 9).

The southern Cave Creek volcanic belt west of Cave Creek town abounds with facies changes where the distal edges of many rocks units are interleaved. Graywacke, slate, volcanic wacke, conglomerate, andesite, rhyolite, iron formation, and felsic to intermediate tuff envelop and are partly interleaved with the North Union Hills Formation andesites. Overlying these rocks between Black Mountain and Go John Mountain is a thick suite of purple slate, siltstone, argillite, graywacke, and felsic volcaniclastics that are a facies equivalent of the Rackensack Canyon Formation of the Alder Group.

East of Apache Peak (fig. 8), distal parts of the Cramm Mountain and North Union Hills Formations are interbedded with and overlain by rhyolite tuff representing the thin southern extremity of the Grays Gulch Formation. This tuff completely thins out east of Apache Peak, right where overlying reworked volcanic and sedimentary units equivalent to the Rackensack Canyon Formation thicken. In this manner, the thinning of one formation allows another to attain a greater thickness, a distinctive feature of laterally interleaved rock units that is invariably seen in all the volcanic belts.

In summary, the primary distinction between the Union Hills and Alder Groups in the Cave Creek area is that volcanic and unreworked volcaniclastic deposits dominate the Union Hills Group, whereas reworked volcanic material and clastic sediments dominate the Alder Group. Because the Grays Gulch Formation is primarily of direct felsic volcanic derivation, it is included in the Union Hills Group. In the Cave Creek volcanic belt, Alder Group sedimentary strata (Rackensack Canyon Formation and facies equivalents) unconformably overlie Union Hills Group volcanic strata (North Union Hills and Grays Gulch Formations), just as in other parts of the younger volcanic belts. Some Alder Group strata in the Cave Creek belt persist laterally for tens of kilometers, are more persistent than correlative units in the Mazatzal Mountains, and could be distinguished as individual formations. The Cave Creek area, in the future, warrants recognition of many more formations than distinguished here.

VERDE RIVER GRANITE BATHOLITH

One of the largest plutonic bodies in central Arizona is a batholith of red granite named here the **Verde River Granite.** It occupies 1,200 km^2 of the lower Verde River drainage and extends from the New River-Cave Creek volcanic belts east to the Mazatzal Mountains (fig. 8). The Verde River Granite is a medium- to fine-grained granite, is red from primary dispersed hematite, and has distinctive small zoned K-feldspar phenocrysts. The batholith is compositionally very homogeneous, but is structurally zoned to include a granite core, granophyric borders, chilled feldspar-phenocrystic margins, and dikes and apophyses of feldspar-porphyritic rhyolite.

The granite intrudes all rocks in the New River-Cave Creek belts older than the New River Mountains Felsic Complex, but nowhere cuts extrusive rocks of the complex. Instead, fine-grained red granite of the batholith is identical to fine-grained red granite in the felsic complex that is gradational upward into intrusive quartz latite and extrusive equivalents. Thus, plutonic parts of the felsic complex are continuous with the Verde River Granite, so the felsic complex signifies a site where granitic magma extruded via subvolcanic phases to form an elongate ignimbrite body. The New River Mountains Felsic Complex is thus strictly coeval and comagmatic with the Verde River Granite.

Another place where the granite surfaced through gradational phases is along its east edge in the Mazatzal Mountains (fig. 8). The core phase of the batholith extends east to Deadman wash, where vertical offset along the Deadman Wash Fault terminates the granite against fine-grained porphyritic rhyolite, telescoping the gradation evident at New River. East of the fault, the same pattern of fine-grained red granite grading to intrusive rhyolite then to extrusive phases is seen from Davenport Wash southeast to Mount Peeley (fig. 8). Boundaries across which there is textural but no compositional difference exist between adjacent phases and reflect crude stratification of the plutonic-subvolcanic complex, analogous to stratification in the overlying extrusive ignimbrites. Based on relative-age relations to isotopically dated, broadly coeval volcanic rocks and to the slightly younger Payson Granite, the ages of the Verde River Granite, New River Mountains Felsic Complex, and Mount Peeley ignimbrites are about 1710 Ma (P. Anderson, 1986; Silver and others, 1986).

MAZATZAL MOUNTAIN-DIAMOND BUTTE VOLCANIC BELTS

UNION HILLS GROUP

As in the Cave Creek-Union Hills areas, the oldest rocks in the Mazatzal Mountain-Diamond Butte area are mafic volcanics of basaltic andesite and andesite composition, which extend from Mount Ord in the Mazatzal Mountains northeasterly along Gun Creek to Diamond Butte (fig. 8). Only the upper pillowed and brecciated mafic flows are exposed between Tonto and Spring Creeks, but the subvolcanic pyroxene-plagioclase phenocrystic basaltic andesite core is exposed at Mount Ord. Because the most complete section is on Mount Ord, the mafic volcanic rocks are newly named the **Mount Ord Formation.** It is the basal formation of the Union Hills Group in the Mazatzal-Diamond Butte area and correlates to similar mafic volcanic formations of the Cave Creek and Union Hills areas.

The other main locality where mafic volcanic rocks are widely exposed is at the northern edge of the Mazatzal Mountains, along the East Verde River. There, the rocks are dominantly fragmental andesites (breccia, agglomerate, tuff, and minor flows) that interface with thick wedges of andesitic graywacke, argillite, siltstone, and slate on the flanks of the volcanic center. Both mafic volcanic rocks and flanking clastics are somewhat different than those of the Mount Ord Formation: the East Verde River andesitic graywacke-argillite wedge is much thicker than that associated with the Mount Ord Formation. In the Cave Creek area, the same andesitic graywacke-argillite suite is separated from the mafic volcanic units by Alder Group (fig. 8).

As was done for the North Union Hills Formation, both mafic volcanics and derivative clastics along the East Verde River are assembled into a single formation, newly named the **East Verde River Formation.** The mafic volcanic part correlates directly to andesite breccias on Humboldt Mountain near Cave Creek, whereas the graywacke-argillite portion correlates to identical rocks northeast of Carefree (P. Anderson, 1986). Removal of the younger Verde River batholith realigns both detached suites as a single, originally continuous belt of mafic volcanic rocks enveloped by volcanogenic sedimentary rocks. Although the East Verde River Formation is now separated from the Cave Creek belt, it is directly correlative to the mafic rocks in that belt (P. Anderson, 1986).

Mostly fragmental rhyolite-rhyodacite, quartz-feldspar-crystal tuff, and minor andesite flows and tuff overlie the Mount Ord Formation to the north and are lithologically and stratigraphically equivalent to the Grays Gulch Formation in the Cave Creek belt. A similar felsic sequence of fragmental rhyodacite, felsic agglomerate, and tuff overlies volcanic rocks of the East Verde River Formation west of City Creek, and its related rhyodacite feeder dikes cut older andesitic graywacke east of North Peak. Away from both volcanic centers, felsic units become cryptically thin and are not regionally persistent. Thus, the felsic rocks are not assigned to separate formations at this time but are included with their respective mafic formations and recognized as coeval with the Grays Gulch Formation in the Cave Creek area.

The East Verde River Formation and the Mount Ord Formation represent the two earliest mafic volcanic centers that evolved in the Mazatzal Mountains-Diamond Butte areas; all subsequent volcanism and clastic sedimentation was built upon and outward from these mafic foundations and flanking volcanogenic sediments. The Mount Ord and East Verde River mafic volcanic centers appear at present to be broadly coeval—certainly their flanking clastic wedges appear to be interstratified in the Slate Creek Divide area—so both formations are included here in the Union Hills Group. These are the mafic volcanics in the Diamond Butte area that Gastil (1958) recognized as lying stratigraphically below Alder Group strata, and they are also the mafic volcanics that Wilson (1939) recognized beneath the type "Alder series" in the Mazatzal Mountains.

UNION HILLS GROUP-ALDER GROUP CONTACT

The major change from volcanism of the Union Hills Group to clastic sedimentation of the Alder Group is marked by a widespread disconformity or unconformity that represents a hiatus clearly demarcating a profound change in evolution of all younger volcanic belts in central Arizona. The change was from primary mafic volcanism to primary felsic magmatism, and accumulation of lower Alder Group strata spans the time interval of that change. This major change in the evolution of the younger volcanic belts is reflected by the two major rock-stratigraphic groups defined here: the Union Hills and Alder Groups. Throughout all younger volcanic belts of central Arizona, mafic volcanic rocks or derivative graywackes lie everywhere at the base of the section, and Alder sedimentary strata overlie this mafic basement. Consequently, stratigraphic columns not recognizing mafic volcanics or related graywackes at the base of the section beneath Alder sedimentary strata incorrectly depict the stratigraphic relationships (see P. Anderson, 1986 for details).

ALDER GROUP

Wilson's (1939) type section of "Alder series" (now Alder Group) was Alder Creek west of Slate Creek Divide, since renamed "Sycamore Creek." The original type area consists primarily of reworked volcaniclastic and sedimentary rocks of dacitic to rhyolitic derivation. Mafic volcanic rocks are typically lacking, despite statements by other workers who have lumped Union Hills Group mafic volcanics into the Alder Group (e.g., Conway and Wrucke, 1986; Conway and Silver, this volume). Unlike the lensoidal, laterally impersistent Union Hills Group units, Alder sedimentary strata extend widely within the Alder depositional basin and consequently lap across many different rock units in the subjacent Union Hills Group. Purple slate, quartzite, and coarse volcanic wacke are most distinctive of the Alder Group; in fact, purple slates are so diagnostic of Alder sedimentation that all deformed Proterozoic purple slate sequences in Arizona may correlate to the Alder Group [see tectonics paper].

Alder Group exposures at Slate Creek Divide are separated (fig. 8) by a central rhyolitic ignimbrite exposure named **Red Rock Rhyolite** (Wilson, 1939). The Red Rock Rhyolite appears to occupy the core of a syncline (Ludwig, 1974), but Alder units do not match across the axis, and differences are not due to facies changes (P. Anderson, 1986). The base of the Alder Group rests on strata of both the Mount Ord Formation to the southeast and the East Verde River Formation to the northwest. Different clastic arrays were shed from each volcanic center into the deep submarine basin that once existed in the Slate Creek Divide area. Alder Group sediments were subsequently deposited in the basin between the mafic centers, and therefore lap unconformably across a diverse array of subjacent lithologies. Alder strata were later locally eroded and were unconformably overlapped by subaerial Red Rock rhyolitic ignimbrites, which also cut across many different Alder lithologies. Thus, two major regional unconformities exist in the Slate Creek Divide area, stratigraphic discontinuities that may go unnoticed in small areas (Ludwig, 1974), but which are obvious on a regional scale (P. Anderson, 1986).

With such great stratigraphic complexity, the Slate Creek Divide area is too complex for type formations that best characterize the true sedimentary nature of Alder Group in the Mazatzal Mountains-Diamond Butte region: unusual lithologies abound, felsic volcanic debris overwhelms the sediments, and the Alder sequence is truncated unconformably at different levels by Mount Peeley and Red Rock ignimbrites. Regional study (P. Anderson, 1986) shows that Reef Ridge east of Tonto Creek is an ideal locality where a stratigraphic sequence truly representative of the Alder Group's dominant sedimentary character (and almost identical to the Alder sequence in the New River-Cave Creek area) is superbly exposed in a simple northwest-facing section. This sequence is newly named the **Reef Ridge Formation** for its type section from Gun Creek across Reef Ridge, a ridge of argillaceous quartzite diagnostic of the Alder Group.

Purple slate, siltstone, quartzite, conglomerate, wacke, and tuff at the base of the Reef Ridge Formation unconformably overlie Union Hills Group mafic volcanic rocks in Gun Creek. The lower slates are partly of felsic volcanic derivation, are interlayered with rhyolite flows and tuffs, and are intruded by rhyolite dikes. Fine-grained diorite, diabase, and andesite dikes also intrude the slates and may be readily mistaken for mafic flows. Conformably overlying the slates are thick argillaceous quartzites and local conglomerates making up the main part of the Reef Ridge Formation. This sequence of impure quartzites and volcanic-rock- and jasper-clast conglomerates interbedded with purple slates are the rocks most diagnostic of the Alder Group (P. Anderson and Wirth, 1981).

At the top of the Alder Group, the characteristic Alder sedimentary strata become overwhelmed by felsic volcanic conglomerate and agglomerate that immediately preceded the main ignimbrite eruptions (fig. 9). Reef Ridge quartzites are overlain by a thin unit of dacite flow, breccia, and crystal tuff, but to the east near Diamond Butte, more than a 1-km thickness of huge dacite-boulder agglomerate and conglomerate (Gastil's 1958 Flying W and Board Cabin Formations) occupies the same stratigraphic position. Still farther northeast, near the Haigler rhyolite center (Conway, 1976), Alder sedimentary units are almost totally overwhelmed by eastward-thickening agglomerate lenses. Yet, these agglomerate facies are absent at Reef Ridge only 7 km to the west, which graphically illustrates that the agglomerate wedges are lensoidal, laterally impersistent, and not typical of Alder Group stratigraphy on a regional scale. Finally at Reef Ridge, the dacites are overlain unconformably by Red Rock rhyolitic ignimbrites, whose unconformity locally cuts into the Alder section and regionally transects the Alder Group (fig. 8).

Alder Group stratigraphy in the Mazatzal Mountains is similar both to the Reef Ridge Formation, with a lower section of purple slate, quartzite, and felsic tuff overlying the Mount Ord Formation in Slate Creek Divide, and to the Alder Group in the Diamond Butte area, with a thick upper section of rhyolite, felsic volcanic agglomerate, and conglomerate that thickens eastward into Gold Creek and Deer Creek, beneath Red Rock Rhyolite. The stratigraphy of the Reef Ridge Formation is remarkably like that of the Rackensack Canyon Formation in the Cave Creek area to the west, thus demonstrating wide lateral persistence of the distinctive sedimentary lithologies within the Alder depositional basin.

The Reef Ridge Formation is not proposed to replace other Alder Group formations (Gastil, 1958; Conway, 1976), but to consolidate them, as it is the only type section truly representative of Alder sedimentary units unencumbered by excess felsic volcanic debris identified in the Mazatzal Mountains-Diamond Butte area (P. Anderson, 1986). A broad perspective shows that the Alder Group is a distinctive sedimentary sequence occupying an elongate basin extending the length of the New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts, interrupted only locally by three felsic volcanic centers (fig. 9). Ludwig (1974), following Wilson (1939) at Slate Creek Divide, and Conway (1976), following Gastil (1958) in the Diamond Butte area, coincidentally studied two of the volcanic centers, thus ingraining the concept of abundant volcanics in the Alder Group. It is vital to recognize the dominant sedimentary character of the Alder Group on a regional scale, and the Reef Ridge Formation establishes this sedimentary sequence in the eastern belts.

RHYOLITIC IGNIMBRITES

As noted previously, the Verde River Granite grades eastward, through granophyric and intrusive rhyolite phases, into rhyolitic ignimbrites of the Mazatzal Mountains that unconformably overlie the Alder Group west of Slate Creek Divide. The sequence of subaerial ignimbrites on and west of Mount Peeley, comprising stratified sheets of flow-banded and fragmental rhyolite and welded crystal tuff, is named the **Mount Peeley Formation** for the distinctive Mount Peeley section. The Mount Peeley ignimbrites warrant distinction from both Red Rock Rhyolite at Slate Creek Divide and Haigler rhyolite in the Diamond Butte area because they lie at a somewhat lower stratigraphic position: they lap unconformably across dacite tuff, felsic fragmentals, and purple slate of the Alder Group, and are separated from Red Rock Rhyolite by a thick volcanic conglomerate. The Mount Peeley and Red Rock ignimbrite sequences exist in parallel exposures for more than 10 km, do not converge or contact one another, persist as lithologically distinct units throughout their full extent, and were evidently deposited in spatially and temporally separate elongate trough.

Thus, Wilson's (1939) original Red Rock Rhyolite is recognized here as lithologically and temporally separate from both Mount Peeley ignimbrites and Haigler rhyolite near Diamond Butte (cf. Gastil, 1958). Red Rock Rhyolite is a distinctive hematite-rich formation exposed from Alder Creek to Jakes Corner, as a knob at Black Mountain, and again unconformably overlying the Reef Ridge Formation at Reef Ridge (fig. 8). At this last locality, the Fe-rich Red Rock Rhyolite is overlain by Fe-poor Haigler rhyolite of Diamond Butte (cf. Conway, 1976), and feeders to the younger Haigler rhyolite are found cutting Red Rock-type rhyolites. These relationships, plus the presence of a large additional thickness of volcanic agglomerate and conglomerate between the Alder Group and the Haigler rhyolite in the Diamond Butte area, show that Haigler rhyolite is youngest, postdating both Mount Peeley and Red Rock ignimbrites.

From these relations it is deduced (P. Anderson, 1986) that rhyolitic ignimbrites in the Mazatzal Mountains-Diamond Butte areas involve three major sequential eruptions from separate volcanic centers: Mount Peeley ignimbrites were extruded first from the 1710-Ma Verde River Granite batholith to the west; Red Rock Rhyolites were extruded at 1705 Ma from a series of Fe-rich rhyolite feeders and granophyres between Tonto Creek and the Mazatzal Mountains; finally Haigler-type and Oxbow-type rhyolites were extruded from 1702- to 1700-Ma (Silver and others, 1986) felsic centers to the east in the Diamond Butte area. This easterly progression of ignimbrite eruptions is consistent with Payson Granite being about 10 Ma younger than the Verde River batholith and being broadly coeval with Haigler rhyolite through granophyric and intrusive rhyolite phases.

Gastil's (1958) original Haigler rhyolite name was used by Conway (1976; Conway and Wrucke, 1986) as "Haigler Group" to refer to all stratified felsic extrusives in the Diamond Butte and Mazatzal Mountains area, including Wilson's (1939) Red Rock rhyolite. The name was then changed from "Haigler Group" to "Red Rock Group" (Silver and others, 1986; Conway and Silver, this volume). The foregoing analysis (see also P. Anderson, 1986) identifies the Haigler and Red Rock rhyolites as lithostratigraphically distinct formations with different relative-age relationships. Because of their stratigraphic uniqueness, both remain viable formations and should retain their original Haigler and Red Rock names; neither name should be changed to group status.

The felsic extrusive units cannot be simply grouped, because they were not deposited in a regular stratigraphic succession. Both Mount Peeley ignimbrites and the New River Mountains Felsic Complex were far removed in space and time from the Haigler felsic volcanic center and have no stratigraphic relations with rocks from that easterly center. Instead each ignimbrite deposit was areally restricted to the vicinity of its own igneous center and spread outward from this center as fans and elongate tongues within narrow valleys, not as broad sheets covering all other units. The ignimbrites started as subaqueous ash flows whose geometry was dictated by paleotopography, and progressed to subaerial ash falls only in their late stages.

Because of their geometry, the various ignimbrite deposits are not closely stratigraphically interrelated, but are much more intimately related to their source intrusive rhyolite and granophyric phases than to other ignimbrite deposits from other centers. It is suggested, therefore, that the ignimbrites be grouped with their subvolcanic source rocks, not with other ignimbrite deposits that may have no observable relations with them. It is misleading to group all ignimbrite deposits in the younger volcanic belts under a single name such as "Red Rock Group" (Conway and Silver, this volume) and imply that the various formations are stratigraphically interrelated to one another, when the evidence indicates that the ignimbrites are areally restricted deposits that never did overlap or interleave. In the case of the ignimbrites, the spatial connection to their centers is much stronger than stratigraphic connections, consequently the term "felsic complex" should be used instead of a single rock group (fig. 9). "New River Mountains Felsic Complex" accurately portrays that ignimbrites from the New River felsic center are restricted to the New River area, so the term "Tonto Basin Felsic Complex" should also be used to indicate that the diverse assemblage of ignimbrites and related subvolcanic rocks east of Payson are restricted to the Tonto Basin area. This terminology gives a true picture of the areally restricted nature of each felsic volcanic complex and its ignimbrite carapace, accurately indicating that Haigler-Red Rock strata did not extend west of the Verde River Granite, and that strata from the New River Mountains Felsic Complex did not extend east of the granite, in accord with the observed field relationships and a measurable difference in isotopic ages (Silver and others, 1986).

MAZATZAL GROUP

A distinctive sequence of clean, well-bedded red-maroon quartzite and conglomerate exposed in the central Mazatzal Mountains was divided by Wilson (1939) into a thinner basal conglomerate-quartzite named Deadman quartzite, a middle yellow pyritic shale unit named Maverick shale, and a thick upper maroon quartzite unit named Mazatzal quartzite. The Mazatzal quartzite sequence has since been recognized as a regionally important stratigraphic unit and has been correlated to other Proterozoic quartzites in the United States. In central Arizona, many isolated occurrences of Mazatzal-type quartzites exist as synclinal keels: each has its own unique lithofacies in detail, but the general lithology and depositional setting of all such occurrences are distinctly of Mazatzal quartzite affinity.

Because of the importance of the Mazatzal name, it has been elevated to group status (P. Anderson and Wirth, 1981; P. Anderson, 1986): the **Mazatzal Group** now includes the quartzite-conglomerate-shale sequence in the Mazatzal Mountains and all correlatives in central Arizona (see P. Anderson and Wirth, 1981 for details). The upper quartzite within the Mazatzal Group of the Mazatzal Mountains has been renamed the **Mazatzal Peak Formation**, for a superb type section on Mazatzal Peak, where the diagnostic changes in color, lithology, and depositional setting occur in a complete unfaulted section (P. Anderson and Wirth, 1981).

Wilson's (1939) original type sections are retained for other Mazatzal Group formations, and facies equivalents of these formations are now recognized in other areas (fig. 10). **Deadman Formation** comprises the lower specularite-rich units of basal conglomerate and quartzite, **Maverick Formation** includes the middle yellowish shale-siltstone sequence, and **Mazatzal Peak Formation** is the upper thick sequence of maroon to white quartzite and granule conglomerate in the central Mazatzal Mountains. In the Sheep Basin Mountain area 30 km to the east, the same stratigraphic sequence is exposed, but the lithologies and depositional settings of the major units are somewhat different (P. Anderson and Wirth, 1981) (fig. 10): (1) basal specularite-rich quartzite and conglomerate units exposed in Del Shay basin are the Del Shay facies of the Deadman Formation, which is like the typical Deadman Formation but has larger rhyolite fragments and a more Fe-rich character because of its proximity to the Red Rock Rhyolite source terrane; (2) the middle sequence of fine-grained red hematitic siltstone and quartzite in Coffeepot Canyon is the Coffeepot facies of the Maverick Formation; it was deposited in a shallow-water oxidizing environment, in contrast to the reduced setting of pyritic shales of the Maverick Formation in the Mazatzal Mountains; and (3) the upper unit occupying Sheep Basin Mountain is the Sheep Basin Mountain facies of the Mazatzal Peak Formation and is very similar to the upper red-maroon

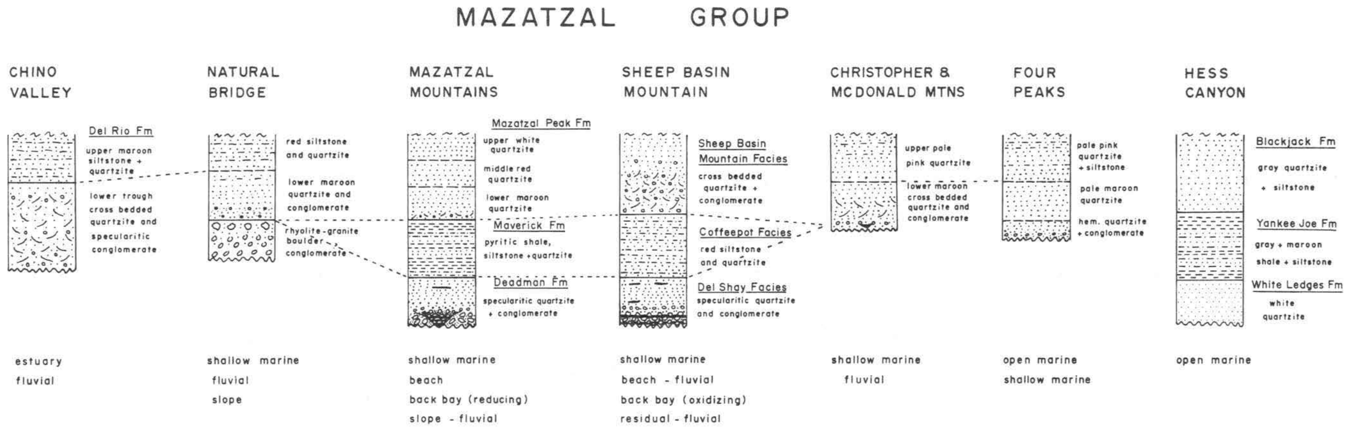


Figure 10. Stratigraphic relations in the Mazatzal Group, as shown by correlations between seven major exposed stratigraphic sections. The main lithologic types diagnostic of each unit are shown at the right of each column, and the variations in depositional environment are shown beneath each column. The columns are much simplified compared to detailed stratigraphic columns described by P. Anderson and Wirth (1981) and Wirth (1980).

quartzite sequence of the Mazatzal Mountains, except for coarser quartz-pebble conglomerate facies due to proximity of a local source terrane.

Mazatzal-type quartzites are also present farther north on McDonald Mountain, on and south of Christopher Mountain, south and north of Natural Bridge near Pine, and both north and southwest of Young. All such quartzites are fine-grained facies of the Mazatzal Peak Formation, except basal rhyolite-boulder conglomerate at Natural Bridge, which may correlate with the Deadman Formation (fig. 10). To the south at Four Peaks and in Hess Canyon, cleaner, fine-grained, less Fe-rich quartzites are part of the Mazatzal Group, but were winnowed in open-marine settings (P. Anderson and Wirth, 1981) that differed significantly from the littoral and strandline settings of the Mazatzal Peak Formation (fig. 10). Only upper quartzites correlative to the Mazatzal Peak Formation occur at Four Peaks, but in Hess Canyon, a more complete section of basal quartzite (White Ledges Formation), middle shale (Yankee Joe Formation), and thick upper quartzite (Blackjack Formation), was correlated directly with Deadman, Maverick, and Mazatzal Peak Formations by Livingston (1969). Because these direct correlations are supported (P. Anderson, 1986), Livingston's "Hess Canyon Group" is redundant, and should be replaced by Mazatzal Group.

In the Chino Valley area north of Prescott (fig. 3), quartzite and conglomerate of the Mazatzal Group are exposed (Wilson, 1939; Krieger, 1965), but do not correlate directly to formations in the Mazatzal Mountains because of a high-energy fluvial environment that deposited specularitic, coarse quartz-cobble conglomerates in all but the uppermost strata (Wirth, 1980). The sequence in Chino Valley is named the **Del Rio Formation**, after Wilson's (1939) original designation, and lies unconformably upon some of the youngest mafic volcanics of the Prescott volcanic belt.

At all locations noted above, the Mazatzal Group is unconformable on either felsic volcanic or sedimentary rocks, lying above a regional erosional surface (fig. 9). The Mazatzal Group therefore represents a stage in evolution of the younger volcanic belts when the subaerial ignimbrite fans were being eroded and the felsic volcanic rocks were being stripped back to sea level. Most fragments cycled into the Mazatzal Group are directly correlative to rock units in the underlying source terrain: fragments in basal strata are typically large rhyolite boulders and intraformational clasts localized in stream channels; fragments in strata at middle stratigraphic positions are usually rhyolite, quartzite, jasper, and chert; and only the most resistant chert, jasper, and quartz grains remain in the upper well-worked strata. Specularite and other heavy minerals derived from the underlying hematite-rich felsic volcanic sequences are strongly concentrated in the basal units, whereas finer hematite grains persist throughout all higher units except the uppermost white quartzites that were winnowed in open-marine conditions. Depositional environments of the Mazatzal Group (fig. 10) are more fully treated elsewhere (P. Anderson and Wirth, 1981; Wirth, 1980; Trevena, 1979, 1981).

CITY CREEK FORMATION

North of the Mazatzal Mountains near the East Verde River is a suite of sedimentary rocks atypical of Proterozoic rock units in central Arizona because of their almost unmetamorphosed, very well preserved condition. They were named by Wilson (1939) the "City Creek Series," and consist of hematitic mudstone, red, maroon, and brown shale, sandstone, siltstone, and other redbeds in the core of a faulted syncline in City Creek. Because of its lithostratigraphic uniqueness, this sequence is renamed here the **City Creek Formation.** Similar rocks are exposed along the East Verde River in sharp angular unconformity over volcaniclastics and diorite. The City Creek Formation postdates this diorite, which is part of the 1738 $\pm$ 4-Ma Gibson Creek complex (Conway and Wrucke, 1986; Silver and others, 1986).

The City Creek Formation lies east of the East Verde River Formation and appears to be unconformable upon the mafic volcanic rocks, although the contact is in part faulted along the western edge of the City Creek syncline. Conway and Wrucke (1986) placed the younger City Creek sequence at the base of the East Verde River Formation under the assumption that a single, extremely thick, west-facing homoclinal sequence occurs along the East Verde river. This is clearly not the case, because mafic volcanics and graywackes of the East Verde River Formation are in part lateral facies equivalents.

The City Creek Formation was originally thought to be probably younger than the Mazatzal Group (Wilson, 1939; P. Anderson, 1986), but if so, it should contain red rhyolitic debris recycled from the major 1700-Ma felsic ignimbrite event. Reconsideration of its relationships now suggests that the City Creek Formation is almost certainly older than the ignimbrites, but younger than the East Verde River Formation, which means it is comparable in stratigraphic position and age to the Alder Group. Rocks of the City Creek Formation are lithologically most like purple slates in the lower Alder Group, but the sequence differs markedly in being made up almost entirely of shaly rocks and devoid of the quartzites characteristic of the middle Alder Group. As such, the City Creek Formation could represent quiet deposition in a local restricted setting on the flank of the main Alder Group depositional basin.

DISTAL AREAS

The New River, Union Hills, Cave Creek, and Mazatzal-Diamond Butte areas represent the volcanic-dominated cores of the younger central Arizona volcanic belts; adjacent areas to the south and southeast received mainly reworked volcaniclastic detritus shed into deep-water environments distal to the main volcanic chain. Rocks deposited in such distal environments are present southwest of Slate Creek Divide, in the hills north of Phoenix, in the McDowell Mountains, near Saguaro Lake, on Four Peaks, and in Hess Canyon on the upper Salt River.

Alder Group volcaniclastics southwest of Slate Creek Divide give way southwesterly to more mature subgraywacke, quartz wacke, quartzite, siltstone, and slate. Farther southwest, these rocks are cut off by the Four Peaks Granite batholith (fig. 8), but recur in the McDowell Mountains as a thick sequence of more siliceous and mature quartzite, siltstone, quartz wacke, quartz-pebble conglomerate, subgraywacke, tuff, and iron formation metamorphosed to andalusite grade by the batholith. This silicic sedimentary sequence is sufficiently different in lithology from the bulk of Alder Group strata to warrant distinction as a separate formation, named here the **McDowell Mountains Formation**. Some strata appear to be stratigraphically correlative to the Alder Group, but others may predate the Alder Group and be the time equivalents of distal graywackes of the Union Hills Group. Still other highly quartzitic rocks may postdate the Alder Group. This same quartzite-wacke sequence extends through the hills between Paradise Valley and Phoenix, where it is intruded and metamorphosed by ca. 1650-Ma diorite, granodiorite, granite, and pegmatite. Locally the sequence contains abundant rhyolitic and crystal tuff, which suggests that some parts may be as young as the 1700-Ma ignimbrite event.

From Squaw Peak Park to North Mountain Park in Phoenix, a typical Alder Group suite of felsic tuff, rhyolitic conglomerate, intermediate tuff, tuffaceous slate, and volcaniclastics is exposed. Deformed rhyolite- and slate-fragment conglomerate and other unique lithologies are identical to rocks in the Slate Creek Divide area, which indicates that the Squaw Peak-North Mountain Park section is directly correlative along strike to (not just a distal facies equivalent of) Alder Group in the Slate Creek Divide area.

In Hess Canyon on the Salt River, rhyolitic ignimbrite (Redmond rhyolite of Livingston, 1969) is exposed in a stratigraphic position equivalent to Red Rock Rhyolite: beneath the White Ledges quartzite unconformity and above volcanic graywackes similar to those distal to mafic volcanics of the East Verde River Formation (figs. 8, 9, 10); Redmond rhyolite, however, is most probably slightly younger than Red Rock Rhyolite. Alder strata are lacking from Hess Canyon, the only metapelites being those within the Mazatzal Group above the White Ledges Formation.

On Four Peaks, specularitic quartzite correlative to Mazatzal Peak and Blackjack Formations unconformably overlies pelites metamorphosed to cordierite-andalusite hornfels by intrusion of the ca. 1450-Ma porphyritic **Four Peaks Granite batholith.** The metapelites are correlative to similar rocks in the McDowell Mountains Formation, and are a distal part of the Alder Group. Ignimbrites are absent on Four Peaks, but occur both east and west where they have been metamorphosed by the Four Peaks batholith. Thus, the main rock units and their correlations can be readily identified in distal areas flanking the central volcanic chains, but the major stratigraphic sequences are relatively discontinuous in these distal areas compared to their continuity in central parts of the volcanic belts.

PETROLOGY AND GEOCHEMISTRY

One of the most remarkable features of the central volcanic belt is that the younger belts started forming almost at the stage where the older belts ceased their evolution, not just in a time sense, but also in terms of petrology and geochemistry. The older Prescott-Jerome belts began with a primitive, magnesian low-K tholeiitic magma series and evolved to fractionated, high-K tholeiitic to low-K calc-alkaline magma series; only the very last magma suites are truly calc-alkaline in character. In contrast, the younger eastern belts began their evolution with a magma series of low-K calc-alkaline character and evolved through the calc-alkaline field to reach alkali-calcic end products in the subaerial ignimbrites. The younger belts lack rocks derived directly from olivine tholeiite parents, they lack a primitive magma series of low-K tholeiitic chemistry, and there is not even a normal Fe-rich tholeiitic trend. The felsic ignimbrites, however, are not rich enough in alkalis to be of truly alkaline character; the most evolved ones are alkali-calcic rather than alkalic.

Union Hills Group mafic volcanics were derived exclusively from quartz-normative source magmas that began crystallizing in clinopyroxene fields. Subvolcanic parts of the Cramm Mountain and Mount Ord Formations have large distinctive clinopyroxene phenocrysts; breccias in the Cramm Mountain, Humboldt Mountain, and East Verde River areas have partly reacted pyroxene phenocrysts; and mafic flows of the North Union Hills Formation contain mainly subophitic clinopyroxene and plagioclase. Nowhere in the Union Hills Group has evidence been found for primary olivine, even in reaction relation with pyroxene, and orthopyroxene is also rare to absent. These petrologic features contrast to those of the earliest mafic volcanic sequences in the Prescott-Jerome belts, which are typically aphyric, alkali-depleted, olivine tholeiites.

The most primitive Union Hills Group mafic volcanics occur in the Union Hills as a spilitic basaltic andesite suite interbedded with relatively sodic rhyolite—the closest rock to a keratophyre in the younger volcanic belts. However, the K_2O content of the sodic rhyolite is nearly four times that of sodic rhyodacite in the spilite-keratophyre sequence of the Senator Formation, the earliest unit in the older Prescott volcanic belt. Thus, the most primitive mafic volcanic suites in the younger belts are much more alkali rich than those in the older belts, as can readily be seen by comparing figures 11 and 12 with figures 5 and 6. Through similar contrasts (see P. Anderson, 1986) it is evident that all other volcanic suites in the younger belts are enriched in K_2O, Na_2O, SiO_2, and locally Al_2O_3, and are depleted in Fe_2O_3, MgO, and locally CaO, relative to analogous suites in the older Prescott-Jerome volcanic belts.

Such features demonstrate that the younger volcanic belts of central Arizona are fundamentally more evolved than the older ones. Instead of a low-K tholeiitic bimodal magma series that began the Prescott-Jerome belts, the Union Hills Group defines a magma series of mainly low-K calc-alkaline geochemistry (figs. 11 and 12). Some variability is present: the "spilite-keratophyre" suite is intermediate in chemistry between tholeiitic and calc-alkaline; some basalts and andesites are high-K tholeiitic, whereas others appear high-K calc-alkaline, in part due to slight alkali mobility in a submarine depositional environment. Na-enriched basaltic andesites in the eastern Mount Ord Formation near Spring Creek define an anomalous field of analyses (fig. 11). Compositional distribution in the Union Hills Group is distinctly *not* bimodal, but uniformly spans the basalt, basaltic andesite, andesite, dacite, and rhyolite fields (fig. 12).

In contrast, mafic rocks in the Little Squaw Creek Migmatite Complex and less metamorphosed counterparts west of New River and in the Black Canyon Belt are tholeiitic basalts with lower K, Na, Si, and higher Fe, Mg, Ti, and Ca contents than any rocks in the Union Hills Group, as expected by their clearly different petrogenesis as part of the older volcanic belts. The lower Fe, Mg, Ti, and Ca contents, higher K and Na contents, and usually higher Al contents of Union Hills Group rocks are typical earmarks of a calc-alkaline magma series. Calc-alkaline dacite, rhyodacite, and rhyolite of the volcaniclastic Grays Gulch Formation in the uppermost Union Hills Group are geochemically very similar to rhyodacite in the Union Hills and rhyolite in the East Verde River Formation (fig. 12). Including such felsic volcanic rocks in the Union Hills Group even more strongly defines its polymodal calc-alkaline nature (figs. 11 and 12).

Mafic plutons intruding the Union Hills Group in the New River, Cave Creek, Union Hills, Phoenix, Payson, and Diamond Butte areas are geochemically similar to the volcanic rocks. Most intrusive rocks are medium- to coarse-grained diorite or granodiorite, gabbro-diorite, and monzogranite. These coeval intrusions define a magma suite of low-K calc-alkaline chemistry nearly identical to the main trend of the Union Hills Group itself (figs. 11 and 12), implying that the two magma types may have evolved from similar sources, even though the plutons are younger than the volcanic rocks, subsequent clastic units, and most of the Alder Group. In contrast, the Gibson Creek diorite and correlatives near Payson are chemically different and probably predate the Union Hills Group. Alkali-poor calcic pyroxenite and gabbro, and calc-alkaline gabbro-diorite and diorite phases of the Gibson Creek complex, define a differentiation trend of strong alkali enrichment totally unlike that of the Union Hills Group (fig. 11).

Dacitic to rhyolitic flows, breccias, and crystal tuffs of the upper Alder Group define a calc-alkaline chemical trend with significant major-element mobility (figs. 11 and 12). Diorite-diabase intruding Alder strata are alkali enriched, and the intruded rocks are usually Fe, Ca, and Mg enriched. However, no Alder Group felsic volcanic units are of alkali-calcic chemistry: at Slate Creek Divide, upper Alder Group felsic agglomerates are a high-K calc-alkaline series, whereas many tuffs are alkali depleted (fig. 11). Unlike all preceding volcanic suites demonstrably linked to a mafic parent of some type, felsic volcanic units in the upper Alder Group and the ensuing Red Rock-type rhyolitic ignimbrites were derived from primary felsic magmas of quartz latite-rhyolite composition. Thus the *Union Hills Group* represents a magmatic suite extensively fractionated from a *primary mafic* parent that ended in the emplacement of differentiated plutonic bodies, whereas the vast *rhyolitic ignimbrite deposits* characteristic of the younger volcanic belts signify inception of a fundamentally different style of magmatism involving formation of *primary felsic magmas* in the crust.

The largest felsic magma formed the New River Mountains Felsic Complex and Mount Peeley ignimbrites and crystallized as the Verde River Granite. A second magma formed the Tonto Basin Felsic Complex and crystallized the Payson Granite. Plutonic, hypabyssal, and subvolcanic phases of all felsic complexes share similar feldspar zoning and crystallization histories. On figure 12, plutonic compositions of calc-alkaline granite are central to the ignimbrite fields, which vary from calc-alkaline and high-K calc-alkaline to alkali-calcic quartz-latite and rhyolite. The felsic igneous suites show little compositional fractionation (figs. 11 and 12), and no volcanic rocks of truly alkaline chemistry (with $>$ 10 percent K_2O + Na_2O at 75 percent SiO_2) have yet been analyzed. Only the Young Granite near Young, which is younger than felsic volcanism, is truly alkaline.

Rare rhyolite sills and flows occur locally in basal Mazatzal strata, and their 1695-Ma age overlaps with that of the younger ignimbrites (Silver and others, 1986). Such sills stem from later caldera resurgence at the Haigler center and postdate the main 1700-Ma ignimbrite events, which are stratigraphically older than the Mazatzal Group. Nevertheless, the timing indicates that inception of Mazatzal sedimentation followed closely on the waning stages of alkali-calcic rhyolitic volcanism in the Mazatzal Mountains-Diamond Butte area.

Thus, primary mafic magmas first intruded the younger volcanic belts, producing tholeiitic to calc-alkaline polymodal volcanics from 1740 to 1725 Ma, locally with coeval mafic plutons. The period 1725 to 1715 Ma was a time of transition from primary mafic to primary felsic magmas, when the calc-alkaline Alder Group was intruded by mafic dikes and local plutons. Primary felsic alkali-calcic magmatism swept eastward across the belts from 1710 to 1690 Ma, and calc-alkaline to alkali plutons were emplaced as late as 1650 Ma.

VOLCANIC STRUCTURE AND EVOLUTION OF THE YOUNGER CENTRAL ARIZONA VOLCANIC BELTS

STRUCTURE

The early *primary volcanic structure* of the younger volcanic belts was similar to that of the older belts: the younger belts were first conceived as isolated centers from which mafic volcanics were extruded into a deep-water submarine environment. Felsic deposits derived from fractionation of these primary mafic magmas were built sequentially around the mafic edifices, and coeval volcanic sediments were shed outward from the centers into distal deep-water basins to interface with coeval deposits from adjacent centers. Lateral facies relationships between volcanic centers and flanking coeval volcaniclastic aprons established a primary stratigraphic complexity in the Union Hills Group equal to that in the Prescott volcanic belt (fig. 2).

Lithostratigraphic units in the Union Hills Group are wedge-shaped packages laterally interleaved with lithologically different coeval rock units. Basaltic-andesite flows of the North Union Hills Formation and pyroxene-phyric andesites of the Cramm Mountain Formation represent two key mafic volcanic centers flanked laterally by coeval andesitic graywackes. Where the edges of the two centers interface west of Cave Creek (along the length of the volcanic chain), andesitic fragmentals are complexly interleaved with the cogenetic graywackes and tuffs that envelop the mafic centers.

The East Verde River center was a third mafic center of the main chain and originally lay northeast of the Cramm Mountain center, but is now separated from it by the Verde River Granite batholith. Fans of tuffs, volcaniclastics, and graywackes were shed from the East Verde River center, and dacite to rhyodacite fragmentals were built upon the mafic units. Mount Ord to the south was a fourth mafic volcanic edifice, with a core of pyroxene-crystal basaltic andesite grading longitudinally to submarine pillowed mafic flows and tuff, and with similar capping rhyodacite flows, breccia, and tuff. The volcaniclastics and graywackes shed from it differ from those originating from the East Verde River center, and both suites interface in the Slate Creek Divide area.

Felsic tuffs and volcaniclastics of the Grays Gulch Formation evolved from the waning stages of fractionated calc-alkaline magmatism that initiated Union Hills Group volcanism, so each felsic deposit is closely linked to its mafic center. The felsic units spread asymmetrically from the volcanic centers toward the intervening deep basins, only partly filling them and leaving a submarine volcanic-sedimentary topography prior to Alder-Group sedimentation that was high over the mafic volcanic centers and low over the intervolcanic basins. The Alder Group was thus destined to fill the intervolcanic basins.

In contrast to sharp lateral facies variations between wedge-shaped rock packages of the Union Hills Group, Alder Group strata were laid down as more laterally persistent facies. Alder Group strata fill a basin intervening between the two main volcanic chains of the eastern belt (the Cramm Mountain-East Verde River chain to the north and the North Union Hills-Mount Ord chain to the south). Because of this structure, Alder strata unconformably overlap distal volcaniclastics of the Union Hills Group rather than proximal deposits at the volcanic centers. Thus, Alder strata persist longitudinally for great distances along a northeast-trending depositional basin that was inherited from the paleotopography of the preexisting submarine volcanic chains.

Erosion of subjacent strata from positive areas and paraconformable deposition in the basin caused the Alder Group to transgress unconformably from east to west as fingers across the diachronous older terrain, until the westernmost edge of the younger volcanic belt was reached in the western New River area. Consistent with this westerly transgression is the fact that the Alder Group contains more volcanic material, is thicker, and is stratigraphically better developed in the eastern Mazatzal Mountains-Diamond Butte areas, whereas only thin basal purple slate and tuff wacke facies are found in the western areas. Thus, major lithofacies of the Alder Group are westward-thinning, elongate depositional lenses. Following an intervolcanic stage of arenaceous-pelitic sedimentation represented by middle strata of the Alder Group, dacite flows and thick dacitic to rhyolitic volcanic conglomerates in the upper Alder Group mark inception of felsic volcanism that culminated in ignimbrite eruptions.

The ensuing rhyolitic ignimbrite deposits inherited much of the Alder northeast-trending basinal geometry, in part filling topographic lows remaining at the close of Alder deposition, rather than forming expansive sheets uniformly blanketing all other sequences. The initial ignimbrite deposits were submarine valley flows fanning out from their respective felsic centers along generally northeast trending valleys. Later subaerial ignimbrites followed much the same structure to produce ignimbrite fans domed over their own felsic centers. Each center of compositionally uniform but texturally gradational felsic rocks represents a place where a huge felsic magma chamber, otherwise crystallizing as a red granite batholith, extruded via subvolcanic-hypabyssal phases to form breccia, agglomerate, and welded ash-flow and ash-fall tuffs. The ignimbrites were extruded from four major centers: (1) the New River Mountains Felsic Complex west of the Verde River batholith; (2) the Mount Peeley Formation on the east side of the Verde River batholith; (3) the Red Rock Rhyolite from felsic feeder systems west of Tonto Creek; and (4) the Tonto Basin Felsic Complex from granitic centers east of Tonto Creek. The northwesterly spread of ignimbrites west of the New River volcanic belt was limited by a scarp that existed along the east edge of the Prescott volcanic belt [see tectonics paper].

Erosion of the felsic centers back to sea level and local subsidence promoted westerly marine transgression and deposition of Mazatzal Group strata unconformably over the earlier volcanic terrane. First, narrow stream channels fed rhyolitic conglomeratic debris into local beach and back-bay environments, then widespread littoral and open-marine conditions prevailed to the southeast, but were limited to the northwest by the older emergent Prescott-Jerome volcanoplutonic belt, which was locally crossed by fluvial channels of Mazatzal age.

EVOLUTION

The younger central Arizona volcanic belts developed adjacent to the older Prescott-Jerome belt, beginning their evolution where the older belt had mostly ceased, both in time and in petrologic-geochemical terms. Formative volcanism of the younger belts postdates formative volcanism of the older belts and evolved from 1740 to 1725 Ma, a period formerly thought to have no volcanic expression in Arizona. Nowhere in the younger belts is there evidence for a preexisting basement, and rock sequences of the older Prescott-Jerome belts do not extend as basement underneath the younger belts, as implied by Silver and others (1986). Instead, formative volcanism of the younger belts was built upon a mafic oceanic substratum that comprised the basement to all supracrustal deposits of the younger belts [see tectonics paper]. In the southwestern New River area, the Black Canyon andesite-dacite sequence was overlapped by Alder strata and at an earlier time was also apparently overlapped by Union Hills Group strata, which implies that the younger volcanic belts are stratigraphically distinct from, but formed adjacent to, the older volcanic belts (P. Anderson, 1986).

The younger volcanic belts began with approximately contemporaneous eruption from three of four major volcanic centers of primary pyroxene-bearing low-K calc-alkaline basaltic andesite magmas derived from quartz tholeiite sources. The magmas formed aphanitic flows in the North Union Hills Formation, thick andesite breccia and agglomerate deposits above pyroxene-crystal flows in the Cramm Mountain Formation, and pillow basalt flows, tuff, and breccia around a subvolcanic pyroxene-phenocrystic core in the Mount Ord Formation.

The mafic magmas fractionated to produce calc-alkaline rhyodacite and rhyolite flows, fragmentals, and tuffs of the Grays Gulch Formation that partly capped the mafic centers and spread asymmetrically into intervolcanic basins. The Union Hills Group encompasses both mafic and derivative felsic formations plus associated feeders, which all evolved from about 1740 to 1720 Ma. Thus, soon after volcanism ceased in the older Prescott belt, new volcanism commenced in the younger belts and continued right up to the end of ignimbrite eruptions, with a major change in the character of magmatism during Alder deposition.

Important hiatuses during deposition of the Alder Group mark a major turning point in petrologic-chemical evolution of the younger volcanic belts. Prior to this change, primary calc-alkaline mafic magmas were emplaced into the upper crust and subsequently fractionated along typical calc-alkaline trends to felsic end products. After the change, however, primary unfractionated felsic magmas of high-K calc-alkaline and alkali-calcic chemistry were emplaced. At the same time, the depositional style changed from one of areally restricted, laterally interleaved, wedge-shaped rock packages laid down in deep-water submarine settings to one of laterally continuous deposition of the Alder Group in shallower marine conditions. Alder Group strata transgressed

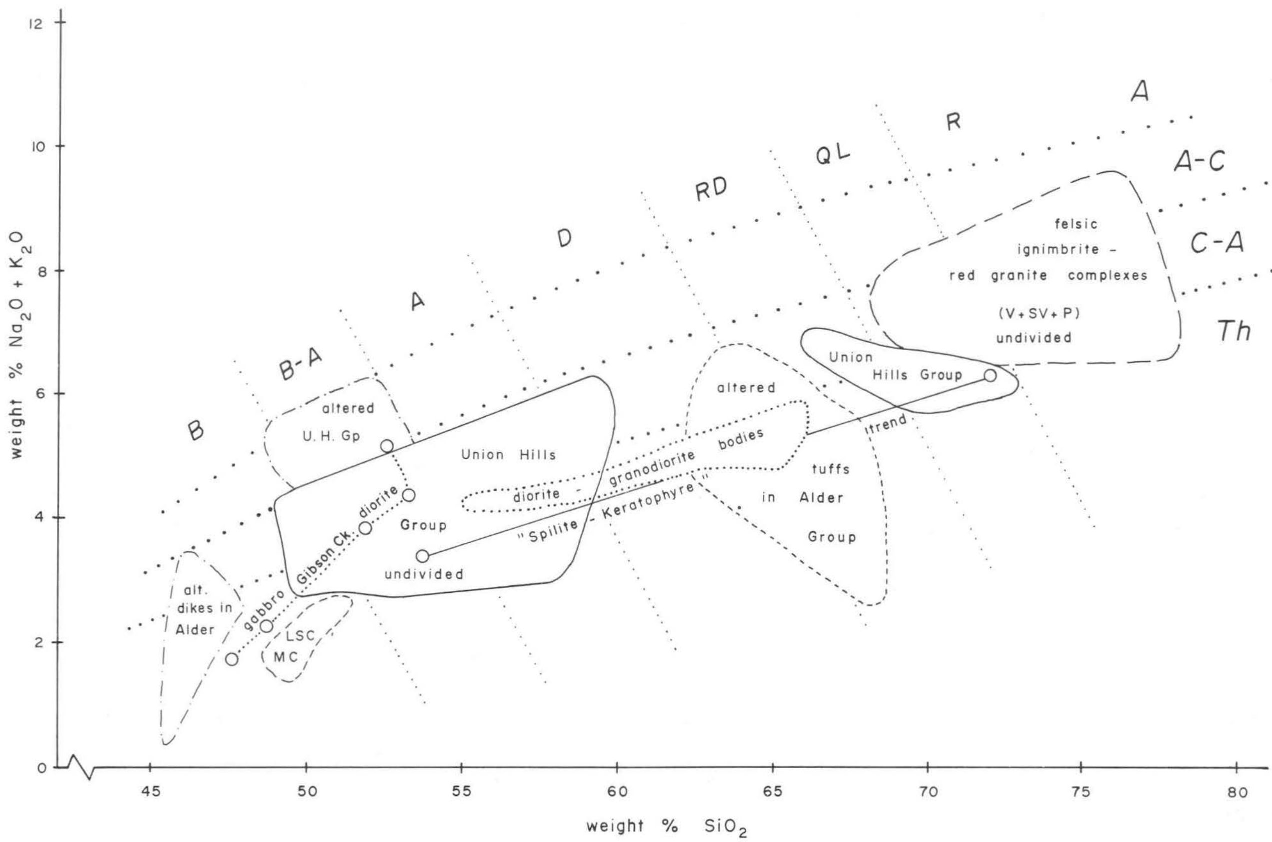


Figure 11. Alkali-silica plot for the New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts, showing fields of related analyses. Total analyses = 105. The fields of the Union Hills Group and the younger felsic complexes (including all volcanic (V), subvolcanic (S), and plutonic (P) phases) are not subdivided because analyses of their component formations overlap indistinguishably on the diagram. Alder Group dacitic-rhyodacitic tuffs are K depleted by sea-water alteration and plot in anomalously low alkali positions. The "spilite-keratophyre" trend represents probably the lowest alkali, unaltered trend in the younger volcanic belts. Mafic pillow lavas in the Union Hills Group are typically Na enriched by sea-water alteration and plot in anomalously high alkali positions (altered Union Hills Group). Mafic plutonic rocks fall into three fields: an altered low-silica field of diabase-andesite-diorite dikes in the Mazatzal-Diamond Butte area; a fractionated trend of alkali enrichment in the older Gibson Creek complex; and a diorite-granodiorite field with chemistry similar to the Union Hills Group, wherein most unaltered plutons in the New River and Cave Creek areas plot. The rock classification grid is described in figure 5 [Part 1].

along the length of the volcanic chain and lapped unconformably back across the older volcanic terrane. Deposition of the calc-alkaline Texas Gulch Formation in structural troughs of the Prescott belt most probably occurred during this same transitional period.

The Alder Group contains depositional, petrologic, and chemical trends that record the transition from dominantly mafic to dominantly felsic volcanism. The oldest Alder purple slate, volcanic wacke, and tuff extend across the entire depositional basin and are distally related to waning Union Hills Group calc-alkaline felsic volcanism. Middle quartzite-siltstone units in the Rackensack Canyon Formation of the Cave Creek area and in the Reef Ridge Formation of the Diamond Butte area are anastomosing channels of an eastward-thickening clastic wedge, which formed as shallow submarine conditions were attained for the first time in the younger easterly volcanic belts. The Reef Ridge and Rackensack Canyon Formations are important Alder Group type sections that establish the correct stratigraphic sequence of the dominantly sedimentary Alder Group, a sequence complicated in other places.

After the quartzites, Alder deposition was progressively overwhelmed by increasing amounts of volcanic material—first by dacite flows and breccias, then by rhyodacite-rhyolitic agglomerates and volcanic conglomerates—such that the sedimentary character of the upper Alder Group was lost to felsic volcanic lithologies. Ludwig (1974) thought these felsic fragmentals were indistinguishable from the overlying ignimbrites, but major geochemical, petrologic, and age differences do exist. The felsic fragmentals are fractionated cotectic low-K calc-alkaline dacites and rhyodacites (compared to the high-K calc-alkaline to alkali-calcic chemistry of overlying ignimbrites and coeval granites) and are thus petrologically distinct

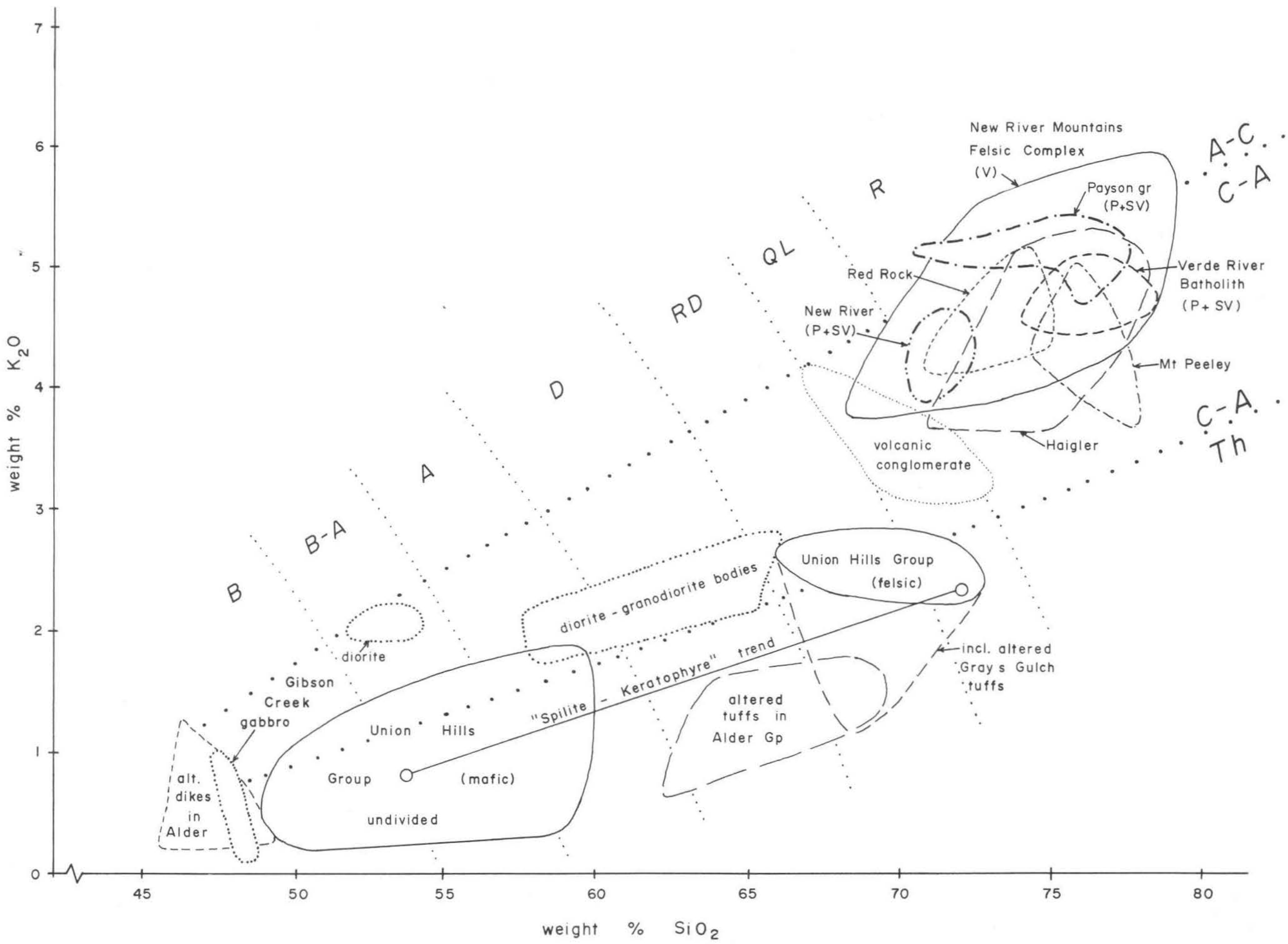


Figure 12. K_2O vs. SiO_2 plot for the New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts showing the same fields of related analyses that are shown in figure 11. The distinctly K-depleted states of dacite-rhyodacite tuffs in the Alder Group and Grays Gulch Formation are more evident on this diagram, as is the slightly more K_2O-rich nature of mafic plutons relative to Union Hills Group volcanics. In addition, volcanic conglomerates transitional from the Alder Group to the felsic complexes (see fig. 9) are shown, and various components of the felsic complexes can be resolved on this diagram. Plutonic and subvolcanic phases (P + SV) of the New River Mountains Felsic Complex, the Verde River Granite batholith in the Cave Creek and Mazatzal areas, and the Payson granite are shown, and volcanic components of the New River Mountains Felsic Complex, Red Rock Formation, Mount Peeley Formation and Haigler-Hells Gate-Ox Bow rhyolites are all distinguished, mainly to show how little compositional differentiation exists in the younger felsic complexes. The rock classification grid of this figure is described in figure 6 [Part 1].

from both unconformably overlying ignimbrites and conformably underlying Alder strata. They in fact represent a separate, widespread conglomeratic suite (fig. 9).

The 1725- to 1710-Ma time span of Alder Group deposition was an important period for mafic to intermediate plutonism. Diorite and granodiorite intruded Union Hills and Alder Group strata of the Cave Creek area, causing minor folds and local unconformities. Alder strata in the Mazatzal-Diamond Butte area, while still unconsolidated, were pervaded by diabase, diorite, and granodiorite dikes. The low-K calc-alkaline chemistry of the intrusions is identical to contemporaneous Alder Group dacites (figs. 11 and 12), and many dikes were feeders to dacite flows in the upper Alder Group. Emplacement of these hydrous mafic-intermediate magmas is consistent with the Alder transition from mafic to felsic volcanism.

In contrast to the Alder Group that transgressed from east to west, ignimbrite activity spread from west to east unconformably across the volcanic belts, mainly as valley flows and fans areally restricted to the peripheries of their generative felsic complexes. First the New River Felsic Complex west of the Verde River batholith extruded high-K calc-alkaline rhyodacite fragmentals and ignimbrites in a northeast-trending belt, then a rhyolite complex east of the batholith extruded the high-K calc-alkaline Mount Peeley ignimbrites, and finally the calc-alkaline batholith core crystallized. A similar sequence of events, repeated at more easterly locations near Tonto Creek, extruded the Red Rock ignimbrites and the more alkali-rich alkali-calcic Haigler rhyolite, and finally crystallized the calc-alkaline to alkali-calcic Payson Granite.

The ignimbrite event lasted from 1710 to 1695 Ma, whereas preceding volcanic conglomerates were deposited about 1710 Ma. Thus, from 1710 to 1690 Ma, the younger volcanic belts were markedly thickened and stabilized, and became emergent. The fluvial network that immediately ensued to effect erosion of the emergent felsic volcanic rocks resulted in deposition of the Mazatzal Group, thus accounting for its closeness in age to the ignimbrites. Most Mazatzal fluvial channels were confined to the younger volcanic belts, but at least one in Chino Valley drained part of the older Prescott-Jerome volcanic belt. As erosion continued, shallow-marine conditions encroached northwesterly over the younger volcanic belts, causing upper Mazatzal Group quartzites to be cleanest, most well worked and most mature sedimentary deposits in the Proterozoic of central Arizona.

Recent U-Pb isotopic dates (Silver and others, 1986) confirm this geologic evolution deduced previously from field relations. The eastward sweep of ignimbrites and their coeval red granite batholiths is confirmed by dates of 1709 ± 3 Ma on the Verde River Granite, 1702 ± 2 Ma on the Payson Granite, and 1703, 1705, 1700, and 1690 Ma on volcanic conglomerates, earliest ignimbrites, Haigler rhyolite, and Hell's Gate rhyolite of the Diamond Butte area, respectively. The age data do not support the idea of an instantaneous ignimbrite "flare-up" across the entire younger volcanic belt, but confirm their restricted distribution in space and time.

PART 3—PLUTONIC SUITES AND METAMORPHISM OF THE PRESCOTT REGION

Proterozoic plutonic rocks are voluminous in central Arizona and are a crucial component in the evolution of Arizona's Proterozoic crust, but have received less attention than the volcanic and sedimentary rocks. Former geologic-isotopic studies concentrated on individual plutons, leaving unstudied the broad interrelationships between plutonic rocks, their related metamorphic events, and their role in central Arizona's Proterozoic tectonic evolution. This paper (condensed from P. Anderson, 1986 and n. d.) fills those gaps by providing new petrologic-geochemical data on plutonic suites of central Arizona, by deciphering their metamorphic events, and by analyzing central Arizona's plutonic evolution in tectonic and time contexts.

GENERAL SETTING

Wilson (1939) recognized that emplacement of granitoid batholiths in central Arizona was accompanied by major Precambrian orogeny, which he thought was a single event he termed the "Mazatzal Revolution." Subsequent isotopic dating (Silver, 1968; Livingston and Damon, 1968) clearly showed two distinctly different ages of granitoid batholiths: (1) an older orogenic group emplaced in the interval 1740 to 1630 Ma; and (2) a younger posttectonic group emplaced in the interval 1425 to 1375 Ma. The older (ca. 1700-Ma) plutonic rocks are broadly orogenic in tectonic setting and either predated or accompanied major regional Proterozoic deformation and metamorphism in central Arizona; in contrast, the younger posttectonic plutons cut orogenic fabrics. This paper describes the older 1700-Ma orogenic plutonic rocks in the Prescott region that were responsible for initiating Wilson's (1939) Mazatzal orogeny. The younger 1400-Ma plutons are distinguished from older 1700-Ma ones by large perthitic K-feldspar phenocrysts with local rapakivi texture, an unfoliated habit, and an invariable granite-monzogranite composition. This simple distinction is accurate in central Arizona, but does not necessarily hold true in northwest and southeast Arizona.

Silver (1967, 1968) placed an age division in the orogenic plutonic rocks near 34° N. latitude that separated 1750- to 1710-Ma bodies to the north from 1690- to 1630-Ma bodies to the south, a boundary like that in figure 1, but trending more easterly. Orogenic plutons discussed here all lie north of the boundary; those to the south were noted in Part 2 of this paper. Previous work on plutons in central Arizona (Jerome, 1956; Blacet, 1968; C. A. Anderson and Creasey, 1958, 1967; Conway, 1976; DeWitt, 1976; O'Hara, 1980), in northwest Arizona (Kessler, 1976; More, 1980; Stensrud and More, 1980), and in southeast Arizona (Silver, 1954; Erickson, 1969; Livingston and Damon, 1968; Livingston, 1969; Swan, 1976) concentrated on individual bodies, not on consanguineous petrogenetic suites, nor on the broad interrelationships between major plutonic suites, nor on their role in Arizona's Proterozoic plutonic evolution. Such is the goal of this paper.

Plutonic rocks of the Prescott region intrude mainly the edges and deeper crustal levels of the Prescott-Jerome volcanic belts and their distal correlatives, deforming and metamorphosing the volcanosedimentary strata of those belts. Metavolcanic rocks dominate the Prescott-Jerome belts, but metasedimentary rocks dominate elsewhere, and such host-crust variations profoundly influenced compositions of the plutonic suites. The host rocks are variously metamorphosed at greenschist to anatectic grades and were deformed during and after emplacement of the orogenic plutons. Thus, deformation was largely independent of the plutons, but was partly influenced by their distribution and relative competencies. The orogenic plutons vary from highly foliated to undeformed, depending on the relative timing of pluton emplacement and regional deformation.

Some plutons discussed here already have formal names (C. A. Anderson and Creasey, 1958; Krieger, 1965; Blacet, 1966, 1968; C. A. Anderson and Blacet, 1972a,b; C. A. Anderson and others, 1971). This paper adopts such names, but in some cases with redefinition. Several batholiths, plutons, and migmatite complexes integral to central Arizona's Proterozoic evolution are newly identified and named here (and in P. Anderson, 1986), and all such names are now proposed for formal adoption. The details of type areas for the new plutonic bodies are given elsewhere (P. Anderson, 1986, n. d.). Still other plutons are assigned informal names at this time. "Batholith" is used with some formal names in this paper to indicate the size of the bodies described, and is not a requisite part of the formal name. Figure 14, indispensable to the discussion to follow, is a new geologic map of the Prescott region that shows the locations of all plutonic rocks noted here.

AGE DATA ON INDIVIDUAL PLUTONS

Isotopic ages of a few plutonic rocks in the Prescott region are given by C. A. Anderson and others (1971), Lanphere (1968), and Livingston and Damon (1968). All U-Pb ages cited here are adjusted for new U-Pb decay constants (Steiger and Jaeger, 1977) and are 20 m.y. younger than the U-Pb ages originally reported. The quartz diorite of Cherry between the Prescott and Jerome volcanic belts is dated at 1740 ± 10 Ma, whereas the Brady Butte and Government Canyon Granodiorites in the west Prescott belt are dated at 1750 ± 10 Ma and 1750 ± 15 Ma (U-Pb zircon ages, C. A. Anderson and others, 1971). The Bland quartz diorite, dated at 1719 ± 9 Ma (Bowring, 1986), and the Bumblebee-Badger Springs Granodiorites, dated at about 1740 Ma (C. A. Anderson and others, 1971), are younger and older phases respectively of the Cherry Springs batholith: lastly, the Crazy Basin Quartz Monzonite in the southern Prescott belt is dated at 1694 ± 9 Ma (U-Pb zircon ages, Bowring, 1986).

This 1750- to 1700-Ma range in U-Pb zircon ages of orogenic plutonic rocks in the Prescott region compares to 1730- to 1700-Ma ages obtained by other methods (Lanphere, 1968; Livingston and Damon, 1968). The oldest 1750- to 1740-Ma plutons are only 5 to 15 m.y. younger than their host volcanic rocks, implying broadly synchronous evolution of the formative volcanic and plutonic rocks of the Prescott-Jerome belts (C. A. Anderson and others, 1971; P. Anderson, 1978a). Large synkinematic felsic batholiths between the Prescott and Bagdad volcanic belts (fig. 1) generally postdate the predominantly pretectonic plutons emplaced strictly within the volcanic belts themselves, and have 1720-Ma U-Pb zircon ages (Silver, 1966, 1976).

Broadly orogenic plutonic rocks in the younger volcanic belts between Payson and the Mazatzal Mountains have 1700- to 1650-Ma U-Pb zircon ages (Silver, 1964, 1968) that are significantly younger than plutonic rocks of comparable tectonic setting in the Prescott-Jerome belts. Unlike the Prescott-Jerome area, ages of plutonic rocks in the Payson-Mazatzal area are not typically 20-30 m.y. younger than the 1740- to 1700-Ma volcanic rocks they intrude [see Part 2], with the exception of the 1740-Ma Gibson Creek complex (Silver and others, 1986), which is as old as the oldest volcanic units. The younger postorogenic megaporphyritic granites are 1425 to 1355 Ma in the eastern belts (Silver, 1968; Livingston and Damon, 1968) and 1395 Ma in the Prescott belt (Dells Granite north of Prescott; Silver and Woodhead, 1983).

TYPES OF PLUTONIC SUITES

The orogenic plutonic rocks are readily divisible in greater detail into suites of different *relative* ages, based on petrologic, chemical, structural, and metamorphic characteristics. The relative-age divisions used in this paper constitute a much more detailed subdivision of plutonic suites than is possible with existing isotopic data. First, a simple geographic subdivision between 1750- to 1650-Ma orogenic plutons of central Arizona can be made (figs. 1 and 13): batholiths of deformed gray granodiorite, tonalite, and monzogranite to the northwest contrast strongly with unfoliated batholiths of red granite to the southeast. Plutonic rocks to the northwest experienced a major regional 1700-Ma event of deformation and metamorphism that the southeastern ones did not; hence preorogenic and synorogenic bodies predominate to the northwest, whereas synorogenic and postorogenic bodies predominate to the southeast. Orogenic timing also differs either side of the boundary [see tectonics paper].

A more detailed subdivision into plutonic suites, however, leads to a more precise relative chronology of tectonic stages that describes the plutonic evolution of the crust. The orogenic plutons and batholiths can be divided into several plutonic suites (fig. 13; tables 1 and 2, p. 115 and 116) on the basis of compositional, petrologic, and chemical attributes, structural fabrics, and relative emplacement ages. Different structural and internal fabrics distinguish syn- and late-tectonic bodies from pretectonic ones: pretectonic bodies crystallized in an unstressed environment with isotropic fabrics; later deformation strongly affected the margins, but internal core phases remained weakly strained or undeformed. In contrast, syntectonic bodies were emplaced into a stressed environment during regional deformation and are generally foliated throughout. Late-tectonic bodies generally have wider metamorphic aureoles and lie in high-grade regional

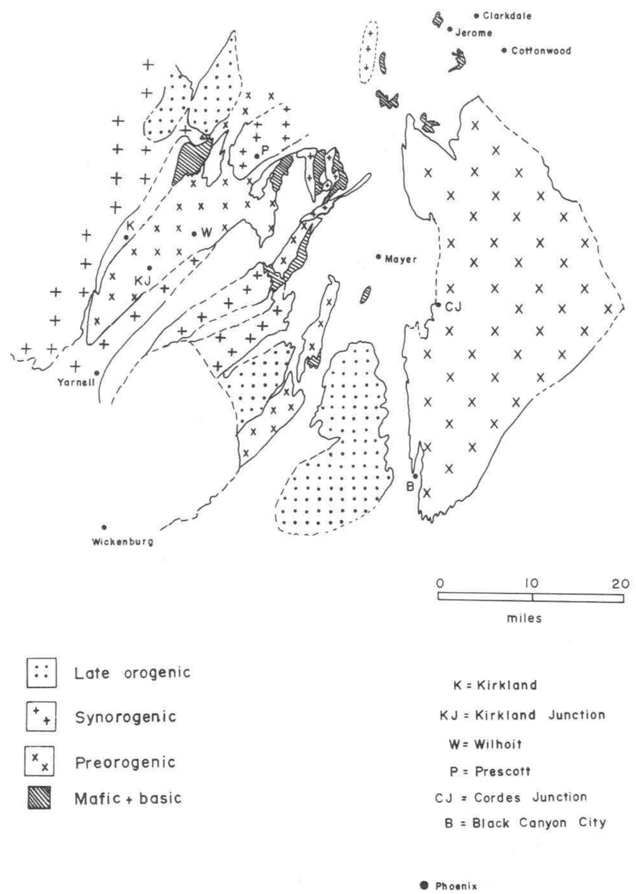


Figure 13. Location map of older plutonic rock suites in the Prescott region.

metamorphic terranes with migmatite-anatectic complexes, because they were emplaced during the thermal culmination of orogenic activity. They usually strongly deformed their host strata, but are not strongly deformed themselves, except at their contacts. Truly posttectonic plutons are undeformed and cut earlier deformed fabrics. Many key petrologic features support such distinctions: for example, late-tectonic plutons are typically K-feldspar phenocrystic whereas pretectonic and early syntectonic plutons are not.

GEOLOGY AND GEOCHEMISTRY OF PLUTONIC SUITES

Figure 13 shows the general distribution of orogenic plutonic bodies in the Prescott region. Figure 14, a more detailed map of the Bradshaw Mountain-Prescott area compiled from new 7.5′ quadrangle mapping (P. Anderson, 1986), shows the locations of all batholiths and plutons discussed here. The order in which the bodies are described here is similar to that in tables 1 and 2, p. 115 and 116. Table 1 subdivides plutonic rocks in the Prescott region solely by relative-age relationships observed in the field. This is the primary basis for a more detailed subdivision into plutonic suites that are genetically related by major petrologic and geochemical attributes (table 2). Figures 15, 16 and 17 summarize key major- and trace-element chemical data for the plutonic bodies discussed in this paper.

An increase in alkali content with time, particularly K_2O, of plutonic bodies in the Prescott region is manifested as two separate trends, both of which were caused by a single tectonic process: (1) in any one area, the K_2O and total alkali contents of plutonic rocks emplaced generally increase with time; and (2) the K_2O and total alkali contents of plutonic rocks as a whole increase to the northwest across the region, in parallel with a northwest-younging of emplacement ages (P. Anderson, 1976, 1978a). The tectonic significance of the northwest alkali enrichment trend, opposite to the southeast alkali enrichment in host volcanic rocks of the Prescott volcanic belt, is discussed elsewhere [see tectonics paper].

The key field relationships and most important lithologic, petrologic, and geochemical features of all major plutons and batholiths in the Prescott region are summarized below, from generally oldest to youngest. The oldest bodies are typically not the most strongly deformed, because many lie in regions of low strain, but they are intruded and diked by younger syn- and late-tectonic ones, thus offering clear evidence as to their relative emplacement ages. In the following discussion, these relative ages are referenced, using pre-, syn-, and posttectonic terms, to the main 1700-Ma event of deformation (tables 1 and 2). A numerical emplacement chronology, referenced to the few isotopically dated bodies, is given at the end of this paper.

GROUP 1: EARLY SYNVOLCANIC AND INTERVOLCANIC MAFIC PLUTONS

Earliest Synvolcanic Gabbro Plutons

The earliest mafic volcanic extrusions of the Prescott-Jerome belts were fed by hypabyssal magmas that crystallized as fine-grained gabbros. One major hypabyssal feeder to early mafic volcanic rocks of the Jerome volcanic belt was the Silver Spring Gulch Diabase, which consists of an upper fine-grained diabase and intrusive basaltic andesite, and a lower differentiated diabase, microgabbro, diorite, granodiorite, and monzogranite. Geochemically the diabase is a low-K_2O, calcic, tholeiitic diabase on a true iron-rich differentiation trend, like that of the host bimodal tholeiites of the Shea and Del Monte Formations (figs. 14, 15, and 16).

In the Prescott volcanic belt, the oldest plutons are mafic Mg-rich subvolcanic gabbro feeders. Both the Dandrea Ranch gabbro and the gabbro of Spruce Mountain in the northern Bradshaw Mountains (figs. 3 and 14) are calcic-series microgabbros that intrude bimodal Mg-tholeiitic basalt and sodic rhyodacites of the Senator Formation, but that predate mafic volcanic units of the upper Senator and Mount Tritle Formations. Gabbros in the Bluebell Mine and Towers Mountain Formations south of Mayer are coeval with tholeiites of the lower Mayer Group. These gabbros are slightly more alkali rich than those in the

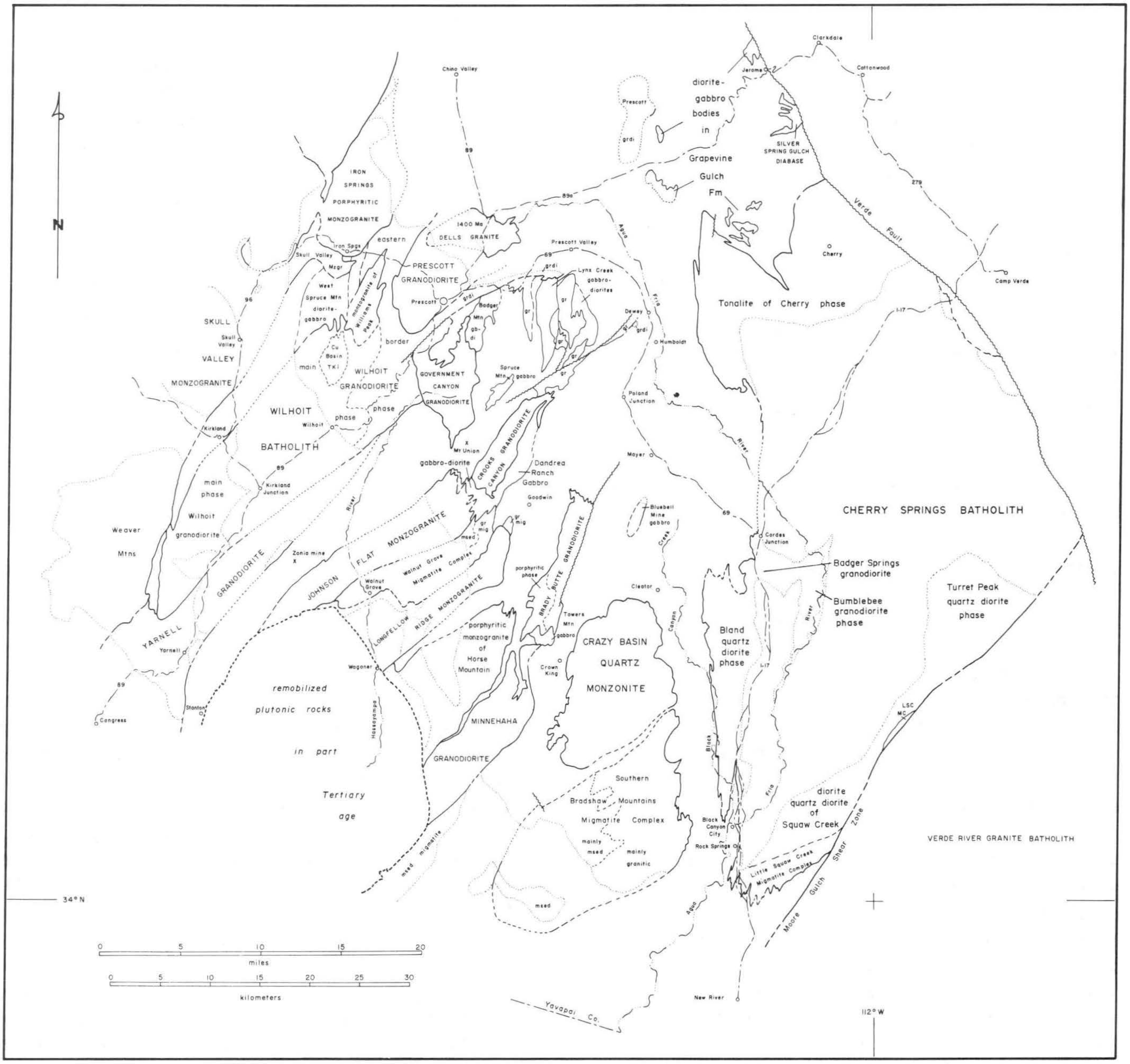


Figure 14. Geologic map of the Bradshaw Mountains-Prescott region showing the major plutonic bodies. The generalized outcrop limits of the plutonic rocks are shown by thin dotted lines where covered by post-Proterozoic rocks, and by short dashed lines where cut by post-Proterozoic intrusive rocks. The contacts of the plutonic bodies with Proterozoic metavolcanic and metasedimentary rocks are shown by solid and dashed lines, but are dotted where projected under cover. Figure 3 is the inverse of this figure, in that the blank areas on this figure are differentiated into the main Proterozoic volcanic and sedimentary rock units of the Prescott region, whereas the blank areas on figure 3 are differentiated here as the main plutonic bodies of the Prescott region.

Senator Formation and are part of the same iron-rich tholeiitic differentiation trend that characterizes the host trimodal basalt-dacite-rhyolite suite of the lower Mayer Group (figs. 5, 6, 15, 16).

Mafic gabbros near Prescott and at Cordes near Mayer are even more Mg rich than Bluebell-type calcic gabbros, with a magnesian-series magma chemistry (figs. 15, 16) and olivine from their olivine-tholeiite parents. Similar Mg-rich gabbros occur with gabbro-diorites in the Grapevine Gulch Formation of the Indian Hills northeast of Prescott. The Mg-rich gabbros appear to be coeval with the earliest calcic gabbros, and together show that the earliest plutons of the Prescott-Jerome volcanic belts were derived from very mafic, magnesian, low-alkali, calcic-series magmas. This earliest gabbro suite is petrochemically important to understanding the evolution of central Arizona's earliest crust (P. Anderson, 1978a; P. Anderson and Guilbert, 1979), but its significance has generally been overlooked in the past.

The earliest synvolcanic gabbros form elongate tabular bodies broadly conformable to and confined by host strata. The younger gabbro-diorites described below, in contrast, are oblate equidimensional bodies that noticeably crosscut stratigraphy. Whereas the gabbro compositions are almost identical to their enclosing tholeiitic basalts, the gabbro-diorite bodies are more evolved and are not tholeiitic. Chemically the difference is between a Mg-Fe-rich, low-K tholeiite and an Fe-rich, normal-alkali calcic magma series. Moreover, new regional mapping shows that a volcanic interval occurred between emplacement of the two mafic plutonic suites.

Younger Intervolcanic Gabbro-Diorite Plutons

A distinctive suite of small gabbro-diorite plutons and plugs cut the oldest stratified rock sequences of the Prescott-Jerome belts, but not younger strata, and are confined to the Grapevine Gulch Formation of the Jerome belt and to the oldest Senator and Mount Tritle Formations of the Prescott volcanic belt (fig. 14). Similar gabbro-diorite bodies also exist in the Cherry Springs batholith. Because the gabbro-diorites were emplaced between major volcanic cycles, they are an intervolcanic suite.

The intervolcanic gabbro-diorites are coarser grained than preceding fine-grained synvolcanic gabbros and are distinguished by ubiquitous compositional-textural variation and in situ differentiation from gabbro to diorite, and a mineralogy rich in Ca-plagioclase. The gabbro-diorites are richer in Ca and Al, and gabbroic portions are richer in Fe and Mg and alkalis and poorer in Ti than the earlier microgabbros. Thus, the gabbro-diorites are a calcic-series, iron-rich, high-Ca-Al magma suite derived from quartz-normative parents.

Gabbro-diorites in westerly Grapevine Gulch Formation exposures (fig. 14) are finer grained than the coarse-grained well-differentiated feldspar-rich, quartz-bearing, two-pyroxene gabbros near the United Verde mine, in which hypersthene and augite ± quartz dominated early cotectic crystallization. Plagioclase dominated later eutectic crystallization in most bodies to produce graphically intergrown quartz, magnetite-ilmenite, and ophitic pyroxene. Parent magmas for these rocks were clearly not olivine normative.

Gabbro-diorites in the northern Bradshaw Mountains, newly named here the **Lynx Creek gabbro-diorites,** include the Badger Mountain body near Prescott and others between Lynx Creek and Green Gulch that show in situ differentiation from gabbro-diorite to late granodiorite phases. Other similar gabbro-diorites occur in the Indian Hills and as an extensive suite on West Spruce Mountain west of Prescott (fig. 14). The Lynx Creek-type gabbro-diorites are chemically most like those in the Grapevine Gulch Formation, with lower Mg and higher Ca-Al-Na (i.e., plagioclase) than earlier microgabbros.

Similar gabbro-diorites occur in westerly border phases of the Cherry Springs batholith and in the Bumblebee and Badger Springs Granodiorite central phases. Some appear to be early mafic borders of the batholith, but others are clearly older and were intruded and metamorphosed by the batholith. In the Wilhoit batholith between Copper Basin and Skull Valley are several diorites, gabbros, and microgabbros that are also thought to be correlative to group 1 gabbros or gabbro-diorites.

GROUP 2: EARLY PRETECTONIC TONALITE AND GRANODIORITE PLUTONS AND BATHOLITHS

The earliest pretectonic tonalites and granodiorites were the first major plutons and batholiths emplaced into the Proterozoic crust of central Arizona. These bodies are larger than preceding mafic bodies, and in some places plutonic phases are so closely related in space, time, and petrogenesis as to be properly described as batholiths (e.g., Cherry Springs batholith). The smaller plutons of the pretectonic tonalite and granodiorite suite closely resemble "I-type" granitoids in younger island arcs (Chappell and White, 1974; White and Chappell, 1977) and were the predecessors to the larger granodiorite and tonalite batholiths emplaced soon thereafter into the flanks of the Prescott-Jerome volcanic belts.

Group 2 plutons and batholiths are more evolved than group 1 gabbros and gabbro-diorites, because the formative evolution of the volcanic belts was mostly complete before group 2 granodiorite-tonalite plutons and batholiths were emplaced. At least three depositional stages in volcanic evolution of the belts intervened between groups 1 and 2. All group 2 plutons and batholiths were emplaced, and some were even uplifted and unroofed, before the ca. 1720-Ma Texas Gulch Formation was deposited in successor basins along the edges of the Brady Butte pluton and Cherry Springs batholith. Group 2 bodies were the first to substantially thicken and stabilize the crust of the volcanic belts. Both the smaller plutons (described first) and the larger batholiths (described second) of this suite are shown on figure 14.

Gneissic Biotite Granodiorites

Brady Butte Granodiorite (Blacet, 1966). The calc-alkaline Brady Butte Granodiorite is one of the oldest dated 1750-Ma plutons in central Arizona. It intrudes Spud Mountain Formation dacites and is unconformably overlain, with a basal conglomerate of granodiorite boulders, by the Texas Gulch Formation [see Part 1]. The Texas Gulch Formation is folded, deformed, and metamorphosed, which shows that the pluton is pretectonic. Deformation of the Brady Butte body imparted a pervasive foliation or gneissosity, composed of finely recrystallized biotite trains on widely spaced foliation planes, a distinctive structure seen in no other granodiorite of the region except the Crooks Canyon. The Brady Butte Granodiorite does not extend south of Towers Mountain and is entirely

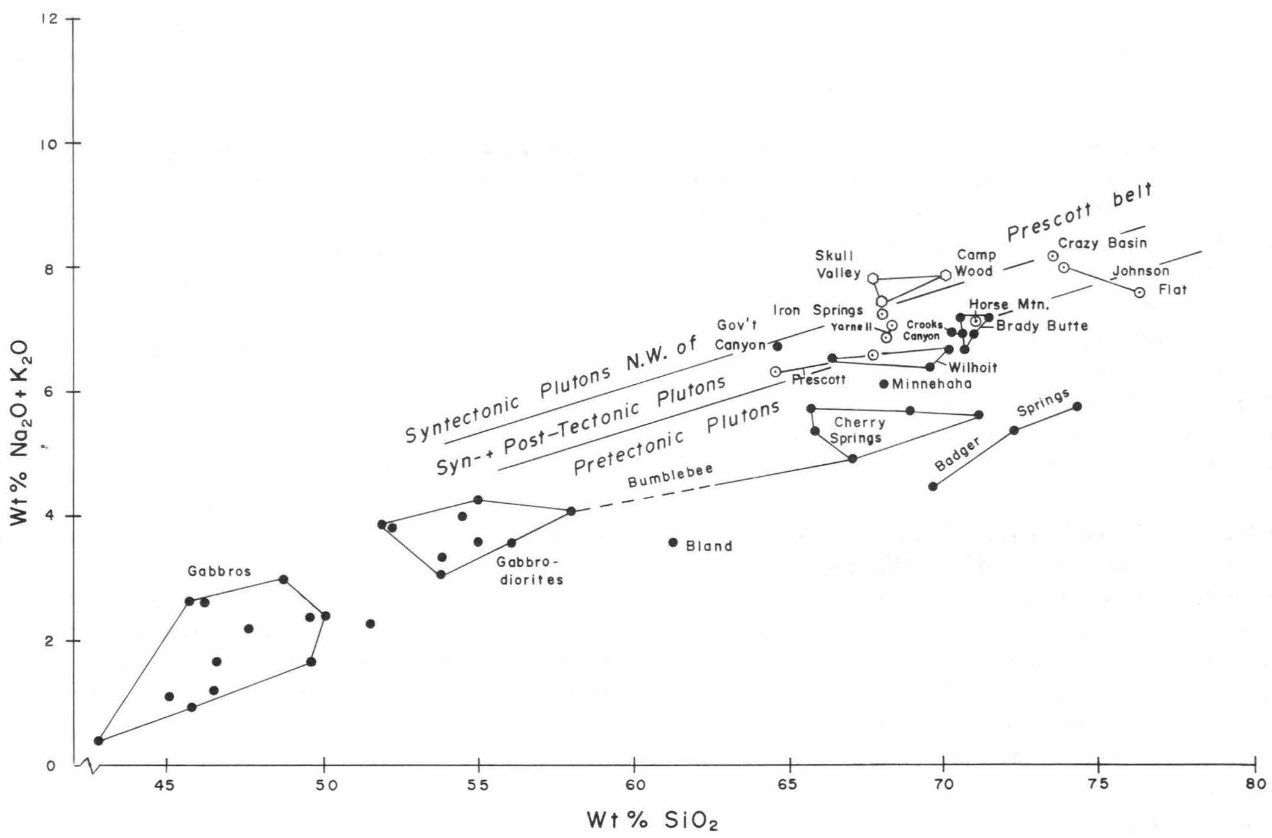


Figure 15. Alkali-silica diagram for plutonic rocks in the Prescott region. Only representative analyses from each plutonic body are shown, and the general fields of pre-, syn-, and posttectonic plutons are derived from a plot of all plutonic chemical data from central and northwest Arizona, totaling over 200 analyses. Analyses labeled as "Johnson Flat" are from related dikes, which may not be the same composition as the main Johnson Flat body. The terms "calcic, calc-alkaline, alkali-calcic" and "low-K, intermediate-K and high-K" both here and on the alkali-silica diagrams are relative, not absolute, and are taken from the classification scheme of P. Anderson (1986), which is similar to Keith's (1978) and most others in common use.

contained within the west-central part of the Prescott volcanic belt. The Minnehaha Granodiorite southwest of Crown King is nongneissic, and is more mafic than and very different from the Brady Butte Granodiorite.

The Brady Butte is a leucocratic granodiorite with resorbed hornblende and minor biotite, magnetite, chlorite and muscovite from subsolidus biotite reaction. Chemically, the pluton is calc-alkaline and is higher in Si, K, and Na, and lower in Fe, Mg, Ti, and Ca than other plutons of comparable age (e.g., Government Canyon Granodiorite). The southern porphyritic quartz monzonite (monzogranite) phase near Tuscumbia Mountain (C. A. Anderson and Blacet, 1972a,b,c) has abundant trachytic K-feldspar phenocrysts, is almost chemically identical to the main granodiorite phase, and has the same distinctive gneissosity.

Crooks Canyon Granodiorite (C. A. Anderson and Blacet, 1972a,b,c). The map unit originally shown by C. A. Anderson and Blacet (1972b) as Crooks Canyon Granodiorite is not a single pluton like the Brady Butte, but a migmatite complex in which swarms of younger fine-grained granitic dikes invade a coarse-grained, gneissic granodiorite and its host metavolcanic-metasedimentary strata. The older gneissic granodiorite is the most important part of this complex and is the rock C. A. Anderson and Blacet (1972b) originally intended to bear that name, so the term "Crooks Canyon Granodiorite" is redefined as only the older coarse-grained gneissic granodiorite of the complex, not any other rocks (P. Anderson, 1986). C. A. Anderson and Blacet's (1972b) original "Crooks Canyon" map unit is true Crooks Canyon gneissic biotite granodiorite only in some places; elsewhere their map units includes migmatite and paragneiss, younger fine-grained red granite locally deformed to ribbon mylonite in the Chaparral high-strain zone, and fine-grained granite and monzogranite dikes. These younger granitic rocks stem from younger plutons to the south—the Johnson Flat and Longfellow Ridge bodies—which belong to a later plutonic suite that caused high-grade metamorphism and migmatization in the complex.

True Crooks Canyon Granodiorite is a leucocratic biotite granodiorite with a low mafic content and a pervasive gneissic fabric. Thus, it is much like the Brady Butte body, but shows no porphyritic character, muscovite, or hornblende reactions. The major-element chemistries of both granodiorites are very similar, except that the Crooks Canyon is slightly higher in Ca and Na and is thus metaluminous, whereas the Brady Butte is weakly peraluminous.

These data indicate that the Crooks Canyon was derived from a somewhat more basic, primitive source magma than was the Brady Butte, one that underwent fractional crystallization by discontinuous reaction. Their markedly different Sr contents at constant FeO supports this contention of different geneses (fig. 17). Because the Crooks Canyon body was emplaced into a more basic, primitive volcanic host crust than was the Brady Butte (the difference between the low-K tholeiitic Senator Formation and the more evolved Spud Mountain Formation), a correlation between the host crust and magma chemistry is implied.

I-type Hornblende-Biotite Granodiorite and Tonalite

Government Canyon Granodiorite (Krieger, 1965). The Government Canyon Granodiorite south of Prescott intrudes low-K Senator Formation tholeiitic basalts, the Mount Tritle Formation, and gabbros in the most mafic, primitive part of the Prescott volcanic belt. The Government Canyon's 1750 ± 15-Ma U-Pb zircon age is comparable with the 1750 ± 10-Ma Brady Butte age (C. A. Anderson and others, 1971), but relative-age relations suggest that the Government Canyon Granodiorite may be older than the Brady Butte pluton (P. Anderson, 1986). The Government Canyon pluton is less foliated than the Brady Butte body because it lies in a less strained part of the Prescott belt than the Brady Butte body; its less foliated habit has nothing to do with age, and is not valid evidence for emplacement near the end of regional deformation (cf. C. A. Anderson and others, 1971). The Government Canyon is a typical pretectonic pluton with a massive core and foliated margins, and its weak foliation is comparable to that of its host volcanic rocks.

The Government Canyon pluton is an equigranular hornblende biotite granodiorite typical of I-type granitoids in many orogenic regions of the world and has abundant mafic, dioritic, and gabbroic xenoliths representative of the igneous source material from which the magma was generated (Chappell and White, 1974, 1976; White and Chappell, 1977). Strong plagioclase zoning in the body is due to discontinuous reactions between hornblende, magnetite, and Fe-biotite during crystallization. The pluton is also zoned, with younger felsic phases cutting slightly more mafic older phases.

The Government Canyon body has normal calc-alkaline chemistry, but, like the volcanic rocks it intrudes, is richer in Fe, Mg and Ti and lower in Si than most granodiorites in the Prescott region. Petrologically and chemically, the Government Canyon is the most definitive of all group 2 pretectonic I-type granodiorites, and many other bodies can be described just by analogy to it. Similar rocks exist in the Cherry Springs batholith to the east, between the Wilhoit and Yarnell batholiths to the southwest, and in parts of the Wilhoit and Minnehaha Granodiorites. Despite these diverse locations, petrologic and chemical similarities of such rocks to the Government Canyon body are remarkable.

Minnehaha Granodiorite (new formal name; P. Anderson, 1986). The Minnehaha Granodiorite is exposed southwest of Crown King in the southern Bradshaw Mountains, and is named for exposures around the locality of Minnehaha on the Minnehaha 7.5′ quadrangle. The pluton has a weakly foliated to undeformed massive core phase (the type area) and moderately to strongly foliated margins. The north edge of the Minnehaha Granodiorite is separated from Brady Butte Granodiorite and Towers Mountain gabbro by a continuous screen of metavolcanic-metasedimentary rocks (fig. 14). The pluton's biotite-hornblende-plagioclase-rich granodiorite mineralogy, petrology, and chemistry are all markedly distinct from the gneissic, leucocratic character of the earlier Brady Butte pluton.

To the southeast, diorite and hornblende granodiorite border phases migmatize mafic metavolcanic and felsic metasedimentary rocks. Along its northwest border, Minnehaha Granodiorite is separated from monzogranite of Horse Mountain by a continuous screen of migmatized metavolcanic-metasedimentary rocks, but to the south, porphyritic dikes from the Horse Mountain pluton intrude the Minnehaha body (fig. 14). Farther south, the Minnehaha Granodiorite is cut by a swarm of leucogranite dikes and by subhorizontal cataclastic foliation, two features that are diagnostic of Mesozoic-Cenozoic tectonic overprinting.

The Minnehaha body is a distinctive biotite>hornblende granodiorite, with finely recrystallized biotite aggregates making up large biotite "flakes" that define a foliation. Although the pluton's petrology and structure is similar to other pretectonic biotite-hornblende granodiorites, its chemistry is not. The Minnehaha pluton is a metaluminous, low-K calc-alkaline granodiorite with lower Fe-Mg and alkalis, higher silica, and much lower Sr than either the Government Canyon or Wilhoit Granodiorites (figs. 15, 16 and 17). With less Si and K, and more Fe and Ca than the Brady Butte or Crooks Canyon, the Minnehaha Granodiorite is petrologically and chemically much like central (Bumblebee and Badger Springs) phases of the Cherry Springs batholith. Its magma appears to have been derived from relatively low-Sr source rocks and underwent little fractional crystallization or differentiation during emplacement.

Wilhoit Granodiorite batholith (new formal name; P. Anderson, 1986). Rimming the northwest end of the Prescott volcanic belt (fig. 14) is a large expanse of foliated to gneissic, coarse-grained, biotite-hornblende granodiorite, whose integrity as a 300-km^2 batholith is largely obscured by extensive swarms of fine-grained granitic dikes and by small plutons and abundant mafic xenoliths. Wilhoit Granodiorite is the oldest plutonic rock in the region, comparable in age to the Government Canyon Granodiorite.

The younger fine-grained granitic dike swarms are related to subsequent emplacement of the suite 3 Williams Peak and Prescott plutons.

Southwest of Prescott near Wilhoit, the batholith can be resolved into several phases zoned about a core phase. The batholith is named for this core phase near Wilhoit, which consists of relatively massive, weakly foliated biotite granodiorite with distinctive euhedral biotite flakes. The core phase grades into a foliated biotite > hornblende granodiorite phase, then into a hornblende > biotite granodiorite border phase. The core phase extends south to Kirkland Junction under cover and north to Copper Basin, where it is cut by the Williams Peak body. A narrow western border phase extends from Kirkland to northwest of Copper Basin, intrudes the gabbro-diorite of West Spruce Mountain, and is cut by the Skull Valley and Yarnell Granodiorites (fig. 14).

The eastern mafic border phase extends from Wilhoit to Prescott, where it cuts the gabbro-diorite of Badger Mountain and is intruded by the Prescott Granodiorite. Faulting along Highway 89 complicates the contact of the eastern border phase with Government Canyon Granodiorite, but in several places the Wilhoit mafic border phase cuts off a massive core phase of the Government Canyon body. The Wilhoit and Government Canyon Granodiorites are mappably distinct but are similar in petrology and chemistry, which suggests similar emplacement ages and origins. Although both bodies are pretectonic, foliated, biotite-hornblende granodiorites, the Wilhoit batholith, especially its core phase, contains more Si, Al, Na, and K and less Fe, Mg, and Ca than the Government Canyon pluton. Both are metaluminous low-K calc-alkaline granodiorites, but the Wilhoit is distinctively lower in Sr than the Government Canyon (fig. 17). This lower Sr content of the Wilhoit Granodiorite parallels the lower Sr contents of its host rocks.

Cherry Springs batholith (new name; P. Anderson, 1986). The Cherry Springs batholith extends over a 1,650-km^2 area and is the largest pretectonic batholith in central Arizona. Previous workers mapped various phases of the batholith as separate plutonic bodies. The northern phase was mapped as "quartz diorite near Cherry" (C. A. Anderson and Creasey, 1958) and dated at 1740 ± 10 Ma (C. A. Anderson and others, 1971), and it is after such analyzed and dated exposures that the batholith is named the Cherry Springs batholith.

In total, at least ten mappably different phases can be discerned in the Cherry Springs batholith, only some of which are mentioned here. The Bumblebee and Badger Springs Granodiorites (C. A. Anderson and Blacet, 1972a) are central phases, and the Bland quartz diorite (Jerome, 1956) is a southern phase, but much of the batholith's central and south-central parts are concealed by Tertiary cover (fig. 14). The eastern phase is the largest in the batholith, extending from Moore Gulch in the south to Turret Peak and possibly Squaw Peak in the north. The batholith consists mainly of hornblende-biotite tonalite and quartz diorite, with diorite borders and biotite-hornblende granodiorite core phases.

All other plutons of the Prescott region are enveloped by screens of metamorphic rock that separate them from adjacent plutons and precisely demarcate their spatial, temporal, and genetic singularity. In the Cherry Springs batholith, the various plutonic phases mapped by previous workers directly contact one another and are not separated by such screens of metamorphic rock. Because the criterion for grouping plutonic phases as a single batholith is the presence of continuous plutonic rock over a large enough area, all Cherry Springs phases batholith can be described collectively a batholith, even if some are of different ages and origins.

General Structure and Petrology. All phases in the Cherry Springs batholith are pretectonic and were emplaced before major regional deformation. The northernmost 1740-Ma Cherry tonalite phase was unroofed prior to Texas Gulch deposition, and dioritic border phases in Little Squaw Creek to the southeast were uplifted before Alder Group deposition. The central Badger Springs and Bumblebee phases appear to be contemporaneous with the northern 1740-Ma quartz diorite of Cherry phase (S. Bowring, 1986). The southwestern Bland phase appears to be 1720 Ma (Bowring, 1986) or 20 m.y. younger, but the isotopic age of the southeastern Little Squaw Creek phase is not yet known.

In general, most phases are coarse grained, northern core phases are almost unfoliated, peripheral phases are weakly foliated, and southern border phases are well foliated to intensely deformed. This broad structure of foliated peripheries enveloping a more massive core is typical of a pretectonic batholith deformed by later tectonism. Strain intensity also varies spatially and is not proportional to age: the older northern phases are least deformed because they lie in regions of lower strain; the younger Bland phase to the southwest is highly deformed because it lies in the Shylock high-strain zone.

In a broad sense, the batholith is zoned from more primitive, mafic border phases to more evolved, felsic core phases. Hornblende-biotite tonalite, trondhjemite, and gabbro dominate in the north, biotite-hornblende granodiorite and tonalite dominate to the east, hornblende tonalite and diorite dominate in the southeast, and biotite-hornblende granodiorite dominates central phases. Western border phases contain enclaves of group 1 gabbro-pyroxenite and gabbro-diorite bodies.

Diorite, mafic tonalite, leucogabbro, and trondhjemite border phases are closest in composition and petrology to the batholith's source material, with high Al+Ca/Na+K, low Si, and high Fe/Ti typical of an anorthite-rich, two-pyroxene mineralogy. In later granodiorite and quartz diorite phases, pyroxene is reacted to hornblende, but clinopyroxene survives in some tonalites. Hornblende predominates over biotite in all but core phases. The Cherry Springs batholith has a metaluminous, low-K calcic-series chemistry with alkali enrichment on a sodic trend (fig. 15), in contrast to the calc-alkaline character of other group 2 plutons.

The Cherry Springs batholith is closer to an original gabbro parent than all other group 2 pretectonic plutons or batholiths. This parentage is clear from figure 17, on which gabbro, diorite, and mafic tonalite phases cluster at the Fe-rich side of the diagram. From that mafic field, felsic phases extend to the left along one of two continuous trends: the lower trend represents no Sr enrichment, hence no strong plagioclase fractionation into a residual liquid, and all central and southern phases (Bumblebee, Badger Springs, Bland, Little Squaw Creek) plot along this trend. The other trend is of strong Sr enrichment, hence strong plagioclase fractionation into a late residual liquid, and all northern phases in the "quartz diorite of Cherry" plot along this trend.

These two related trends of figure 17 strongly suggest that southern and northern parts of the batholith were derived by fusion of similar gabbroic parent material, but that the source for the north part was richer in Ca and Sr than that of the southern one. Late phases crystallizing in the core area even further enhanced generic differences through fractional crystallization. It is highly significant that the host rocks in both settings show the same contrast: primitive Mg-Ca-Sr-rich mafic volcanic rocks make up the host crust to the north, whereas less primitive, Mg-Ca-Sr-poor sedimentary rocks make up the crust to the south. *This correlation to host crust appears to be independent of age,* because the Bland is geochemically like the Badger Springs and Bumblebee phases and intrudes the same crust, but is 20 m.y. younger.

Four of the better known phases of the Cherry Springs batholith are as follows: (1) the northern quartz diorite of Cherry, (2) the central Bumblebee and Badger Springs Granodiorites, (3) the southwest Bland quartz diorite, and (4) the southeast tonalite-diorite of Little Squaw Creek.

Quartz diorite (tonalite) of Cherry (C. A. Anderson and Creasey, 1958). The quartz diorite (tonalite) of Cherry intrudes and metamorphoses Grapevine Gulch volcaniclastics, and is unconformably overlain by deformed but unmetamorphosed sedimentary rocks of the Texas Gulch Formation. The Texas Gulch contact at the north end of the batholith is a deformed unconformity (P. Anderson, 1986), not a fault (C. A. Anderson and Creasey, 1958). This requires that emplacement and unroofing of at least the northern part of the Cherry Springs batholith predated ca. 1720-Ma deposition of the Texas Gulch Formation and subsequent regional deformation. The quartz diorite of Cherry is a metaluminous calcic-series tonalite with high Sr content just like the Government Canyon and Prescott Granodiorites (fig. 17), which all intrude Sr-rich volcanic host rocks.

Bumblebee and Badger Springs Granodiorites (C. A. Anderson and Blacet, 1972b). The Bumblebee Granodiorite is a metaluminous, calcic-series, biotite-hornblende granodiorite similar to Jerome's (1956) Bland quartz diorite, but varies from tonalite to granodiorite. It intrudes rocks of the Black Canyon Creek Group along its western margin (fig. 14), where it is well foliated and clearly pretectonic. The Bumblebee is quartz diorite in places and has a primitive-element profile similar to group 1 gabbros (table 2), which it intrudes and to which it may be related. Both the Badger Springs and Bumblebee Granodiorites appear to have the same 1740-Ma isotopic age as the tonalite of Cherry to the north (S. Bowring, 1986), and contact relations suggest that all three bodies are phases of the one consanguineous igneous event.

The metaluminous calcic to calc-alkaline, locally porphyritic biotite granodiorite of Badger Springs intrudes the more deformed Bumblebee border phase, but gradations in lithology and strain imply little if any relative-age difference between the two phases. Consanguinity of both phases is supported by major- and trace-element similarities (figs. 15, 16, 17): Bumblebee and Badger Springs Granodiorites both overlap the gabbro field (fig. 17), and both appear to have been fractionally derived from quartz-normative basaltic source magmas. Contact relations suggest that the Badger Springs represents a deuterically altered core phase of the northern part of the Cherry Springs batholith and that the Bumblebee Granodiorite represents a border phase in intrusive contact with Jerome's Bland quartz diorite.

Bland quartz diorite (tonalite) (Jerome, 1956). The strongly foliated and deformed Bland quartz diorite borders the southwest edge of the Cherry Springs batholith from Bland Hill (near Bumblebee) continuously south to the batholith's southern terminus near Black Canyon City (fig. 14). At its southern end, the Bland tonalite is the most highly sheared plutonic rock in central Arizona: rocks near the contact are so tightly interdeformed with volcanic host rocks and so highly sheared as to be scarcely recognizable as originally plutonic rocks. The least deformed, unaltered parts are of metaluminous low-K calcic chemistry, with lower Si, Al, and K, and higher Ca and Mg than more northerly granodiorite phases. The chemistry and petrology of the Bland is similar to tonalites along the east edge of the batholith. On Bland Hill, the Bumblebee Granodiorite locally seems to intrude Bland tonalite, in apparent conflict with the Bland's younger 1720-Ma isotopic, age (Bowring, 1986).

Diorite-tonalite of Little Squaw Creek (informal; P. Anderson, 1986). East of Bland quartz diorite is the Little Squaw Creek Migmatite Complex [see Part 2], which represents the contact zone between the southeastern mafic plutonic border of the Cherry Springs batholith and metavolcanic host rocks. Diorite dominates migmatites near the contact, but hornblende-rich granodiorite, biotite-hornblende tonalite, and monzodiorite dominate to the

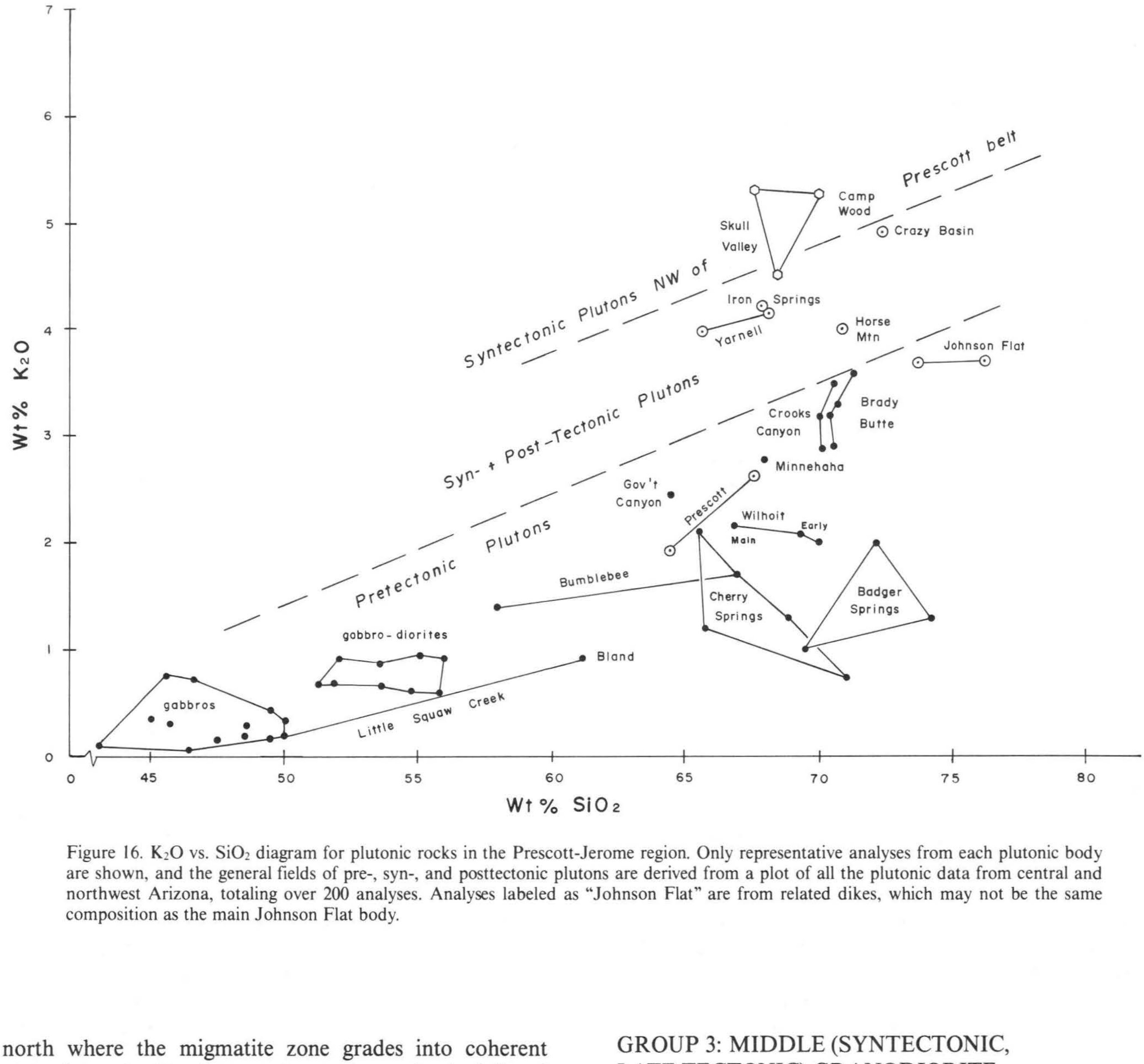


Figure 16. K_2O vs. SiO_2 diagram for plutonic rocks in the Prescott-Jerome region. Only representative analyses from each plutonic body are shown, and the general fields of pre-, syn-, and posttectonic plutons are derived from a plot of all the plutonic data from central and northwest Arizona, totaling over 200 analyses. Analyses labeled as "Johnson Flat" are from related dikes, which may not be the same composition as the main Johnson Flat body.

north where the migmatite zone grades into coherent plutonic rock. Farther north, coarse-grained, foliated hornblende-biotite tonalite grades into granodiorite towards Turret Peak. All of these plutonic phases are metaluminous and of calcic chemistry, with compositions spanning all previous fields, including gabbros at the west edge of the batholith.

The less foliated, hornblende-biotite granodiorite and quartz diorite exposed between Turret Peak and Brooklyn Peak is intruded to the west by Badger Springs Granodiorite in a zoned relationship similar to that of Badger Springs Granodiorite intruding Bumblebee Granodiorite to the west. The quartz diorite-granodiorite phase of Turret Peak and all southerly diorite, quartz diorite, and granodiorite phases of Little Squaw Creek likely predate Bland quartz diorite, and may be 1740 Ma, especially since Alder strata to the south are not metamorphosed by them.

GROUP 3: MIDDLE (SYNTECTONIC, LATE-TECTONIC) GRANODIORITE-MONZOGRANITE PLUTONS

Group 3 plutons belong to a later stage in the evolution of central Arizona, after the major batholiths and plutons of group 2 had largely stabilized and thickened the volcanic belts. They are distinctively more leucocratic, finer grained, and more evolved in petrology and chemistry than preceding plutons and batholiths, and in all cases intrude the pretectonic group 2 granodiorite plutons and batholiths.

Prescott Granodiorite, the monzogranite of Williams Peak, and Skull Valley Monzogranite all intrude the Wilhoit batholith, sending swarms of fine-grained dikes into the batholith. The Johnson Flat and Longfellow Ridge Monzogranites intrude Crooks Canyon Granodiorite and its host strata, and likewise pervade them with swarms of fine-grained dikes. Such relationships suggest a common event, process, or stage in evolution that produced group 3 plutons, an event that was synchronous across the entire evolving volcanoplutonic belt. All group 3 plutons are distinctively fine grained and nonporphyritic, unlike all later plutons and batholiths that are porphyritic. Where contacts between porphyritic and nonporphyritic plutons occur, porphyritic ones are younger, so a similar relation is assumed for bodies not in contact with one another.

Prescott Granodiorite (Krieger, 1965). Unfoliated Prescott Granodiorite discordantly intrudes the foliated Wilhoit batholith, which implies that Prescott Granodiorite may postdate some early deformation. The comparable monzogranite of Williams Peak west of Prescott is cut by late-tectonic Iron Springs Porphyritic Monzogranite, so the Prescott Granodiorite was emplaced after 1740-Ma pretectonic bodies but prior to 1700-Ma deformation. Exposures of Prescott Granodiorite in the Indian Hills west of Mingus Mountain (fig. 14) indicate that Prescott Granodiorite may be of batholithic proportions under alluvial cover of Chino Valley. In Prescott, both the Prescott Granodiorite and its swarms of fine-grained monzogranite dikes intrude and are less deformed than coarse-grained Wilhoit Granodiorite.

Chemically, Prescott Granodiorite is richer in Al and Na and is poorer in Fe and Ti than the Wilhoit batholith, and is even more enriched in Al, Ca, and Na in the Indian Hills. Prescott Granodiorite is distinguished from earlier coarse-grained I-type granodiorites by: (1) medium- to fine-grained texture, (2) very finely crystalline nature of its biotite, (3) zoned plagioclase, (4) hornblende-biotite mineral reactions, and (5) eutectic quartzofeldspathic intergrowth. The Prescott Granodiorite has a moderate to high Sr content (fig. 17), as do the mafic volcanic and metasedimentary rocks of the Senator Formation that contact it (fig. 14).

Monzogranite of Williams Peak (new informal name; P. Anderson, 1986). On Williams Peak near Prescott, a medium- to fine-grained, white leucocratic monzogranite pluton with finely crystalline biotite is texturally identical to Prescott Granodiorite, but is more silicic and closer to monzogranite in composition, and has no red iron stain distinctive of Prescott Granodiorite. Fine-grained monzogranite dikes from this Williams Peak body are indistinguishable from Prescott Granodiorite dikes and invade the Wilhoit batholith and its host rocks. The Williams Peak body intrudes gabbro-diorite of West Spruce Mountain and Wilhoit Granodiorite and its host rocks; it is intruded by Iron Springs Porphyritic Monzogranite, but nowhere cuts or is cut by Prescott Granodiorite. These relationships show that the Williams Peak is broadly if not exactly coeval with Prescott Granodiorite and that both bodies produced the intense dike swarms that extend from Prescott to as far south as Wilhoit. Based on mineralogy, the Williams Peak is a weakly peraluminous, high-K calc-alkaline rock like Prescott Granodiorite.

Johnson Flat Monzogranite (new formal name; P. Anderson, 1986). Extending from Johnson Flat, its type area in the Bradshaw Mountains, southwest under cover to recur near Walnut Grove and the Zonia mine (fig. 14), is a leucocratic, white, medium- to fine-grained monzogranite with small euhedral biotite flakes and minor reacted hornblende—a rock macroscopically almost identical to Prescott Granodiorite. Like the Prescott and Williams Peak bodies, the Johnson Flat is strongly elongate northeast parallel to regional tectonic foliation, is weakly foliated throughout but well foliated at its borders, and is associated with an intense swarm of fine-grained granitic dikes that extends the 25-km length of the body and for another 10 km northeast through Crooks Canyon Granodiorite. Near Walnut Grove, the dikes intrude host metavolcanic and metasedimentary rocks to form a wide migmatite complex, named here the **Walnut Grove Migmatite Complex.**

The Walnut Grove Migmatite Complex is vital to deciphering the geology of the area. Host volcanic rocks, originally of the Senator, Mount Tritle, and Spud Mountain Formations, are rendered into gneisses as metamorphic grade increases southward from upper greenschist near Mount Union to middle amphibolite facies near Walnut Grove. Fine-grained granitic dikes in the migmatite complex, from both the Johnson Flat and adjacent Longfellow Ridge bodies, were responsible for metamorphism and migmatization in the complex. The dike swarms continue northward through the Crooks Canyon Granodiorite, causing migmatization throughout an area that was inadvertently included by C. A. Anderson and Blacet (1972b) in their Crooks Canyon map unit. True coarse-grained, foliated, pretectonic Crooks Canyon Granodiorite is cut off by the Johnson Flat body and is absent to the south (cf. DeWitt, this volume).

The Johnson Flat Monzogranite body extends only to the southern limit of Crooks Canyon Granodiorite, where it and its dikes intrude both the Dandrea Ranch gabbro and the Crooks Canyon Granodiorite (fig. 14). North of Mount Union, the dike swarms merge back into the fine-grained biotite monzogranite of Mount Elliot, which extends north through the Chaparral high-strain zone into a low-strain region farther north. In these high-strain zones ("Chaparral Fault" of C. A. Anderson and Blacet, 1972b), monzogranite is locally deformed to ribbon mylonite by combined vertical strain and 1 km or less of strike-slip motion.

Thus, Johnson Flat Monzogranite and its related plutons and dikes can be traced for 45 km across the Bradshaw Mountains from deep crustal levels and concordant migmatites in the southwest to higher crustal levels in the northeast. All such fine-grained granitic rocks belong to a single event of syntectonic emplacement. Although the Johnson Flat body is a monzogranite and the Prescott body is a granodiorite, they are chemically quite similar (fig. 17).

Longfellow Ridge Monzogranite (new formal name; P. Anderson, 1986). The northeast-trending Longfellow Ridge body parallels the Johnson Flat body from Walnut Grove to its type section on Longfellow Ridge in the Bradshaw Mountains, where it ends (fig. 14). It is nearly identical to

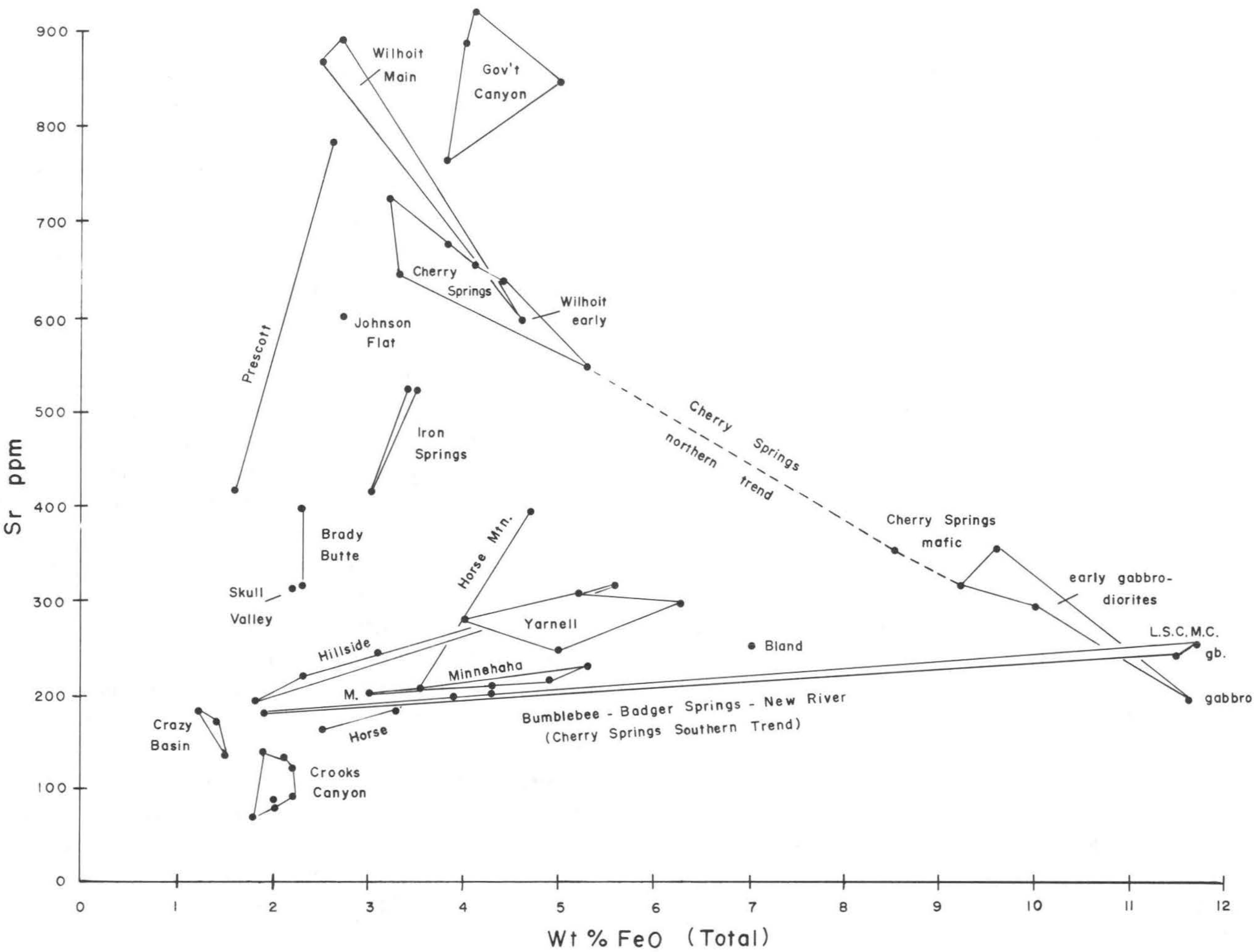


Figure 17. Sr vs. total Fe expressed as FeO for plutonic rocks in the Prescott-Jerome region. L.S.C.M.C = Little Squaw Creek Migmatite Complex. All Sr analyses were performed by the U.S. Geological Survey and the data has been generously supplied by Ed DeWitt.

the Johnson Flat Monzogranite, but is more strongly foliated, altered, and iron stained at its borders. Swarms of fine-grained red granitic dikes from the body intrude the Crooks Canyon Granodiorite and Dandrea Ranch gabbro, forming another migmatite analogous to the Walnut Grove Migmatite Complex (fig. 14).

The Longfellow Ridge and Johnson Flat bodies are chemically similar, but a southward progression through group 3 plutons encounters progressively more felsic compositions, such that each successive body is less Ca-Sr rich and more Si-Al-K rich than its northern neighbor. This gradual chemical change closely parallels the change in chemistry of host volcanic units across the southwest flank of the Bradshaw Mountains. Thus, the Longfellow Ridge is a weakly peraluminous, high-K calc-alkaline monzogranite with lower Ca, Mg, Fe, and Sr than Prescott Granodiorite, and it intrudes a host crust equally depleted in these elements, compared to the more primitive host crust of the Prescott body to the north.

Skull Valley Monzogranite (new formal name; P. Anderson, 1986). The Skull Valley Monzogranite is an eastern border phase of a huge batholith that extends west through Hillside towards Bagdad. The entire batholith appears homogeneous and seems to consist of coarse-grained, massive, bouldery weathering monzogranite with small K-feldspar phenocrysts. The batholith differs from nonporphyritic group 3 plutons in its weakly porphyritic texture, but the Skull Valley Monzogranite is included in group 3 because of similarities in texture, composition, and age relations to other group 3 plutons. The Skull Valley Monzogranite is weakly foliated and intrudes the pretectonic Wilhoit batholith; however, it is cut by the Iron Springs Porphyritic Monzogranite (fig. 14), so is close in age to the Williams Peak body, which is likewise deformed where intruded by unfoliated Iron Springs Porphyritic Monzogranite. Moreover, swarms of nonporphyritic monzogranite dikes that extend northeast from the Skull Valley body are indistinguishable from dikes of the Williams Peak and Prescott bodies.

Skull Valley Monzogranite is separated from the Wilhoit batholith by a persistent metavolcanic screen that extends from Kirkland north to West Spruce Mountain (fig. 14), and only farther north does it intrude the Wilhoit Granodiorite. In the type area at Skull Valley, the rock is a medium- to coarse-grained, weakly porphyritic, K-feldspar-rich monzogranite with abundant biotite and minor muscovite. Phenocryst size increases to the west, but decreases easterly towards its contact.

Chemically, the Skull Valley body is higher in K and Ti, and lower in Ca, Al, Na, Mg, and Fe than Prescott Granodiorite. Skull Valley Monzogranite is similar to, but has higher K/Na than, the monzogranite of Horse Mountain: both bodies plot in similar positions on figure 17 and are notably lower in Sr than all other group 3 plutons, as are their sedimentary host rocks. However, the Skull Valley Monzogranite and its related batholith to the west are clearly peraluminous, and were emplaced into a much more felsic, nonvolcanic host crust than were plutons to the east. This sedimentary crust appears to be unrelated to the Prescott-Jerome belts [see Part 1], and the difference in crustal composition is clearly reflected in the more peraluminous, evolved nature of plutonic rocks to the west.

GROUP 4: LATE (LATE-TECTONIC) PORPHYRITIC MONZOGRANITE-GRANITE PLUTONS AND BATHOLITHS

Coarsely Porphyritic

The late-tectonic group 4 plutons are distinctive in their coarsely porphyritic habit and weakly foliated to unfoliated character. They typically intrude other plutons that were deformed either before or during their emplacement. It cannot be shown from relative-age relations alone whether all members of this group were emplaced concurrently, mainly because no pluton comes in contact with another. Strict temporal concurrence is unlikely, however, because of age differences in their associated metamorphic events, as noted later. In particular, migmatization related to the Crazy Basin Quartz Monzonite postdated emplacement of the Horse Mountain body and probably that of all other porphyritic plutons as well.

Yarnell Granodiorite (new formal name; P. Anderson, 1986). The Yarnell Granodiorite is named for exposures of foliated, coarse-grained porphyritic granodiorite to monzogranite just north of Yarnell. More coarsely porphyritic monzogranite comprises a vast batholith in the Weaver and the Date Creek Mountains to the southwest (fig. 14), but this batholith is largely unstudied and may involve 1400-Ma plutonic rocks. The Yarnell Granodiorite is therefore discussed separately from the batholith, even though their chemistries are similar (fig. 17). The Yarnell Granodiorite follows the northwest edge of the Stanton-Octave metavolcanic-metasedimentary screen to as far north as Wilhoit (fig. 14), where dikes of unfoliated Yarnell Granodiorite intrude foliated granodiorite of the Wilhoit batholith.

The Yarnell Granodiorite is distinctively coarse grained and weakly foliated, with large pinkish-tan K-feldspar phenocrysts in an equigranular matrix with biotite, plagioclase, uncommon hornblende, and abundant sphene. Macroscopically it resembles no other rock in the Prescott region; it is similar to 1400-Ma granite in phenocryst habit, but the main Yarnell body is deformed and not a 1400-Ma body. The main K-feldspar is twinned microcline, a mineral not found in most preceding plutonic rocks. Where the rock is foliated, oligoclase, perthitic microcline, quartz, and biotite are recrystallized to fine aggregates of sericite, quartz, feldspar, biotite, chlorite, and magnetite. Yarnell Granodiorite is diagnostically rich in sphene.

Chemically the Yarnell body is a metaluminous, high-K calc-alkaline, high-Fe-Ti and high-total-alkali rock of monzogranite to granodiorite composition. It appears to have been derived by fractional crystallization that started in the plagioclase-hornblende field, but it largely crystallized before reaching the eutectic, because microcline and quartz are interstitial to orthoclase and plagioclase. In biotite granodiorite compositions, hornblende survives as only a partly reacted constituent, but in porphyritic monzogranite compositions it is almost totally reacted to biotite, chlorite, and magnetite.

Porphyritic monzogranite of Horse Mountain (new informal name; P. Anderson, 1986). In the southern Bradshaw Mountains is the distinctive porphyritic monzogranite of Horse Mountain, which intrudes the north edge of the Minnehaha Granodiorite (fig. 14). The Horse Mountain monzogranite is distinguished by its large, tabular, trachytic K-feldspar crystals in a foliated biotite-rich matrix. Pegmatite dike swarms pervade central and contact regions of the body, extend north to cut Longfellow Ridge Monzogranite, form a narrow migmatite zone along the southeast border of the pluton, extend southwest to intrude Minnehaha Granodiorite, and extensively inject the body itself.

The porphyritic monzogranite has distinctive fresh 1-cm-long, Carlsbad-twinned K-feldspar phenocrysts in a finer matrix of quartz, plagioclase, K-feldspar, and biotite flakes, which locally show subsolidus reaction to muscovite, even though the rock is not chemically peraluminous. In fact, low Al content and high Si, K, and Na contents are the rock's main chemical features. Minor hornblende occurs in more strongly foliated rocks to the south. Andalusite-cordierite metamorphic mineral facies along the east side of Horse Mountain body are overprinted farther to the east by higher pressure mineral facies related to the younger Crazy Basin Quartz Monzonite (see below).

The monzogranite of Horse Mountain has a trace-element profile like Yarnell Granodiorite, and was derived from a magma chemically similar to that of the Yarnell or Iron Springs plutons. In addition, overlap of analyses from the Horse Mountain and Minnehaha plutons (fig. 17) is significant because both bodies were emplaced into the same crustal region and have similar trace-element contents, even though they are of different ages.

Iron Springs Porphyritic Monzogranite (new formal name; P. Anderson, 1986). This distinctive coarsely feldspar-porphyritic biotite monzogranite, named for exposures at Iron Springs northwest of Prescott, is a comparatively massive pluton undeformed in its type area, but weakly to moderately foliated in peripheral areas near Granite Mountain and Williamson Valley (fig. 14). The Iron Springs Monzogranite intrudes strongly foliated Wilhoit Granodiorite and moderately foliated Skull Valley Monzogranite, and its contact truncates dikes from the Williams Peak body. Thus at least some deformation in Skull Valley Monzogranite occurred before or during emplacement of the Iron Springs body.

Of all plutons in the Prescott region, the Iron Springs Monzogranite is most like a 1400-Ma porphyritic granite, except for its different microcline phenocrysts. Its major-element chemistry is comparable to that of 1400-Ma granites, but its distinct phenocrysts, higher Sr content, and deformed character all earmark the Iron Springs Monzogranite as an orogenic pluton. The Iron Springs body is richer in Sr, lying in a more Sr-rich region of crust, than either the Yarnell or Horse Mountain plutons, but may have been derived from a similar source magma by feldspar fractionation (fig. 17).

Partly porphyritic, migmatitic

Plutonic rocks of the latest tectonic type are chemically similar to coarsely porphyritic plutons, with high Si and K, high K/Na, and low Mg and Ca contents, but are even richer in SiO_2 (73 percent) and poorer in Mg and Ca than the coarsely porphyritic plutons. This latest suite was emplaced in a late-tectonic setting when regional deformation was mostly complete, and evolved very differently than did the coarsely porphyritic plutons. The main body, the Crazy Basin Quartz Monzonite, is involved in a high-grade metamorphic terrane where metasedimentary rocks underwent partial melting over broad areas, and where foliated plutonic rocks are so intimately intermixed with partially fused metamorphic rock that it is difficult to consider them separately.

Crazy Basin Quartz Monzonite (C. A. Anderson and Blacet, 1972a,c). The Crazy Basin Quartz Monzonite was named by C. A. Anderson and Blacet (1972a) for the north end of what is now known to be a much larger composite granite batholith. New mapping extends the body widely through the southern Bradshaw Mountains (fig. 14) and links it with the genetically related **Southern Bradshaw Mountains Migmatite Complex** (P. Anderson, 1986).

The Crazy Basin body consists mainly of highly silicic, foliated to gneissic, medium- to coarse-grained, biotite-muscovite granite with small, distinctive, trachytic feldspar laths. Near its contact are screens of biotite-feldspar paragneiss compositionally like the granite; farther inward, coarse-grained two-mica granite predominates; and the core is extensively silicified by abundant pegmatite dikes and quartz-feldspar veins. The Crazy Basin batholith is characterized by marked changes in texture and grain size, and abounds with patches of silicic gneissic leucogranite that are identical to leucogranitic gneiss in the Southern Bradshaw Mountains Migmatite Complex. Such rocks in the complex represent in situ and partly remobilized anatectites no older than the Crazy Basin granite; their gneissic character does not imply greater age (cf. DeWitt, this volume). A large sill of such leucogranite occurs on Lane Mountain in the batholith.

Analyses of the Crazy Basin granite plot invariably at the ternary eutectic minimum: most of the body is a peraluminous, high-K calc-alkaline to alkali-calcic granite with high Si (73.5 percent) and high alkalis ($K_2O + Na_2O > 8$ percent). The granite has a complex crystallization history related to metamorphic grade of host rocks and to high volatile content of the magma. Granitic compositions vary slightly depending on the type of nearby host metamorphic rocks. The Crazy Basin intrusion coincided with the peak of high-grade regional metamorphism and anatexis in the southern Bradshaw Mountains, which just postdated the peak of regional deformation in the Prescott volcanic belt. The Crazy Basin granite is dated at 1700 Ma (Silver, 1976) and 1694 ± 9 Ma (Bowring, 1986), and postdates the Horse Mountain pluton.

Southern Bradshaw Mountains Migmatite Complex (new formal name, P. Anderson, 1986). The contact between the Crazy Basin granite and the Southern Bradshaw Mountains Migmatite Complex is gradational, not intrusive, and its placement is somewhat arbitrary. A semicontinuous screen of highly metamorphosed migmatized paragneiss forms a convenient map boundary between dominantly plutonic rock to the north and dominantly migmatite to the south (fig. 14). Field distinction is less clear, because abundant plutonic rock equivalent to Crazy Basin granite exists in the Southern Bradshaw Mountains Migmatite Complex, just as abundant migmatite occurs in the dominantly plutonic Crazy Basin body to the north.

The Southern Bradshaw Mountains Migmatite Complex consists of medium- to fine-grained gneissic granite and granitic gneiss mixed with metasedimentary gneiss, migmatitic paragneiss, schist, pegmatite-aplite, tourmaline pegmatite, and biotite amphibolite in an extremely intricate, practically unmappable fashion. Most rocks in the complex can be termed either muscovite-biotite gneissic granite or granitic gneiss because they have attributes of both plutonic and metamorphic rocks. Coarser grained gneissic granite and monzogranite occur near the Crazy Basin granite, but grade into finer grained rocks in the southern region. Granitic rock gives way progressively to high-grade metamorphic rock west of Columbia and in the northern Hieroglyphic Mountains, where the western boundary of the complex is defined as the limit of partial fusion and migmatization (fig. 14).

TECTONIC AND METAMORPHIC EFFECTS OF PLUTONIC SUITES

The plutonic suites of the Prescott region were emplaced into the evolving central Arizona Proterozoic crust at various times and crustal levels and under differing pressure-temperature regimes. This produced a complex, locally overlapping array of thermal aureoles in host metavolcanic and metasedimentary rocks. The early suites caused smaller, lower temperature aureoles at high crustal levels, whereas the later suites caused wide, higher temperature, higher pressure aureoles at deeper levels.

GROUP 1 SYNVOLCANIC AND INTERVOLCANIC MAFIC PLUTONS

The early synvolcanic gabbro and intervolcanic gabbro-diorite suites were integral to their host volcanic sequences and caused little metamorphic effects, except for local baking of host rocks and mild mineral recrystallization at contacts. Their influence was more chemical than thermal, in that they caused assimilation, alteration, and especially Fe-Mg metasomatism (e.g., chloritic alteration around the Silver Spring Gulch Diabase in the Jerome area persists for 100 m into felsic volcanic rocks). Emplacement of the gabbro-diorite suite caused local doming, warping, and disruption of stratigraphic continuity, such that parts of the volcanic belt were raised and the first major intervolcanic unconformities were developed [see Part 1].

EARLY GROUP 2 PRETECTONIC PLUTONS

Early group 2 small plutons, such as the Brady Butte, Crooks Canyon, and Government Canyon bodies, were emplaced at high crustal levels and caused only narrow contact-metamorphic effects. The Government Canyon contact aureole may be up to 200 m wide in mid-greenschist-grade Mount Tritle tuffs, but Senator Formation mafic flows at its contact are no more metamorphosed than those 2 km away. Because the Brady Butte Granodiorite is rimmed by younger Texas Gulch Formation, evidence for its metamorphic effects occurs only to the south where the Tuscumbia Mountain porphyritic phase intrudes Towers Mountain Formation calcareous mafic tuffs. However, the tuffs are more highly metamorphosed near the Horse Mountain monzogranite to the west (fig. 14), so metamorphic grade actually decreases toward the Brady Butte body. The metamorphic aureoles of the Brady Butte and Government Canyon plutons are comparable and each about 200 m wide.

Such restricted aureoles contrast to strong migmatization northwest of the Crooks Canyon Granodiorite, but, as noted previously, that migmatization is related to later granitic dike swarms from the Johnson Flat and Longfellow Ridge bodies, not to emplacement of Crooks Canyon Granodiorite. To the north away from the dikes, Crooks Canyon Granodiorite intrudes Spud Mountain dacitic tuffs, causing only minor metamorphic effects.

After formative volcanism ceased, emplacement of suite 2 plutons within the Prescott belt and of the Wilhoit and Cherry Springs batholiths along its edges caused the first major tectonism of the volcanic belts. This resulted in domal uplifts around plutonic crests, downwarps and local fault troughs on the pluton margins, and subsequent uplift, erosion, and deposition of the Texas Gulch Formation in the successor basins.

MAJOR GROUP 2 PRETECTONIC BATHOLITHS

All major group 2 batholiths have metamorphic aureoles much larger than preceding plutons, but still of limited extent compared to the vast size of the batholiths. Along the margins of the batholiths are mafic migmatite zones that vary from narrow to wide, but which differ greatly in composition, structure, and metamorphic grade from the migmatites related to the later group 3 syntectonic granodiorites emplaced into more felsic host crusts.

Cherry Springs Batholith

Contact zones of the Cherry Springs batholith vary with location, but the western side has generally much narrower aureoles than the eastern side. The northern contact with volcanic rocks of the Jerome belt is sharp, with minor migmatization, mild recrystallization within 200-400 m of the contact, and metavolcanic enclaves prograded only to upper greenschist grade. The western contact in the Shylock zone has a narrow aureole only 200 m wide, partly due to strong flattening in the Shylock zone. Northeast of Mayer, thermally recrystallized greenschist-grade metavolcanic rocks lie up to 600 m from the contacts of quartz diorite and other early hypabyssal granodiorite phases.

In the northern Black Canyon Belt where Bumblebee Granodiorite and Bland quartz diorite broadly interdigitate with host metavolcanic and hypabyssal rocks, contact metamorphic effects are limited to within about 100-300 m of the contact. In the southern Black Canyon belt, vertical elongation and tectonic flattening against the batholith produced a planar north-south contact with a narrow greenschist-grade metamorphic aureole. At the batholith's southern terminus near Black Canyon City, granodiorite is intersheared on a tightly isoclinal scale with host volcanic units retrograded to sericite schists during deformation.

Little Squaw Creek Migmatite Complex

Along the southeast edge of the Cherry Springs batholith near the New River volcanic belt is the widest migmatite zone of the batholith—the Little Squaw Creek Migmatite Complex. This 4-km-wide zone involves early diorite, granodiorite, and late granitic dikes injecting host mafic volcanic rocks, hypabyssal microdiorite, and andesitic tuffs. Foliated dioritic rocks with septa of meta-andesite and microdiorite dominate the northern 1.5 km of the zone, strongly lineated granitic gneiss interlayered with meta-andesite and tuff make up the central 1 km, and granodioritic gneiss injecting meta-andesitic tuffs dominate the southern part of the zone. Metamorphism throughout the Little Squaw Creek Migmatite Complex is of lower amphibolite facies, but the southeast edge of the complex demarcates a steep metamorphic gradient, south of which lower greenschist grade tuffs and schistose metasedimentary rocks occur in Moore Gulch.

Farther north along the eastern Cherry Springs batholith contact near Brooklyn Peak, the same complex exists at higher amphibolite metamorphic grades but is of more limited extent, being partly cut off by the Moore Gulch shear zone. Consequently, the widest migmatite zone of the Cherry Springs batholith is associated with diorite and quartz diorite border phases on its east side, not with the main granodiorite-tonalite phase to the west.

Minnehaha Granodiorite

The Minnehaha Granodiorite, like the Cherry Springs batholith, has a wide contact migmatite zone along the southeast edge and narrow aureole to the northwest, except for later migmatization to the north by pegmatite dikes from the Horse Mountain pluton. The eastern side of the Minnehaha Granodiorite contains many amphibolite and biotite gneiss enclaves from a screen of metavolcanic and metasedimentary rocks that separates it from the Horse Mountain and Brady Butte plutons. Metamorphism in this screen is discussed in the section about the Horse Mountain pluton.

The wide southeastern contact of the Minnehaha pluton involves a 1-km-wide zone of granodiorite with mafic metavolcanic septa, bordered by a 2-km-wide migmatitic zone of paraschist and gneiss (cordierite-muscovite schists and muscovite-tourmaline hornfels) injected by granodiorite and pegmatite. This amphibolite-grade metamorphic aureole merges southward into pervasive amphibolite (cordierite-garnet-andalusite) metamorphic conditions as the regional thermal gradient increases southwesterly into the Morgan Butte-Red Picacho area near Wickenburg.

Wilhoit Batholith

The southeast edge of the Wilhoit batholith is a 2-km-wide migmatite zone near Wilhoit that includes amphibolite-grade mafic metavolcanic and metasedimentary rocks and partly assimilated gabbroic septa representing fragments of the west edge of the Prescott volcanic belt. These mafic migmatites extend to just northeast of Prescott, past which Mount Tritle Formation tuffs predominate in the migmatite zone in Prescott Valley (fig. 14). The Wilhoit Granodiorite thus skirts the entire northwest edge of the Prescott belt, incorporating whatever host rocks exist at each locality into its contact migmatite zone.

The western contact of the Wilhoit batholith appears sharper but is intruded by Skull Valley Monzogranite along much of its length. Three screens of metavolcanic-metasedimentary rock near Kirkland mark the boundary between Skull Valley Monzogranite and the Yarnell and Wilhoit Granodiorites (fig. 14). Farther north near Copper Basin, Wilhoit Granodiorite complexly intrudes, metamorphoses, and migmatizes gabbro-diorite of West Spruce Mountain and its host mafic and felsic volcanosedimentary rocks, forming a migmatite like that at the batholith's eastern contact, except for the different host rocks.

A key feature of contact-metamorphic effects around all plutons and batholiths discussed so far is the paucity of aluminosilicate index minerals or diagnostic assemblages in metapelitic rocks. Metapelites in the metamorphic aureoles of all later plutonic suites have such index minerals, and so P-T differences can be analyzed more accurately between the later syntectonic plutonic suites.

GROUP 3 SYNTECTONIC INTERMEDIATE PLUTONS

Whereas contact zones of pretectonic plutons and batholiths involved passive assimilation in a static environment, the syntectonic plutons were intruded into a dynamically stressed environment, so their emplacement was forceful and accompanied by dike swarms and formation of stress-controlled, injection-migmatite complexes. These dike swarms parallel northeast-trending pluton contacts, are most intense at pluton margins, and extend away from the ends of plutons like fingers from a hand, to gradually die out away from their source. Thermal metamorphism of host rocks is linked to the size and extent of the dike swarms, and is regional only where dike swarms are extensive. Away from pluton margins and dike swarms, metamorphic grades drop sharply from amphibolite to greenschist facies.

The Prescott Granodiorite and the monzogranites of Williams Peak, Johnson Flat, Longfellow Ridge, and Skull Valley all produced dike swarms and zones of injection migmatization at their borders and outward from their ends. The Johnson Flat swarm migmatizes Mount Tritle volcaniclastics throughout a 15-km-long, northeast-trending zone that was also a locus for the Chaparral shear zone. Dike swarms between the Johnson Flat and Longfellow Ridge bodies produced the Walnut Grove Migmatite Complex from strata of the Spud Mountain, Senator, and Mount Tritle Formations (fig. 14). Dikes from the Skull Valley body produced a long north-trending migmatite zone truncated by the later Iron Springs pluton. Emplacement of all group 3 syntectonic plutons, dikes, and injection migmatites caused deformation that produced foliation and lineated gneissic fabrics in host strata.

GROUP 4 LATE-TECTONIC PORPHYRITIC PLUTONS

Porphyritic plutons of late-tectonic group 4 produced various contact effects depending their host rocks. The Iron

Springs body near Prescott has sharp contacts with few dikes and a contact metamorphic aureole less than 1 km wide, involving retrograding of earlier amphibolite-grade metamorphic assemblages and K-feldspar metasomatism. The Yarnell Granodiorite produced a 100-m-wide cordierite-biotite hornfels aureole in dacite flows and breccias in the Stanton-Octave screen, and weak thermal effects up to 250 m from the contact. Yarnell Granodiorite is locally separated from Hillside and Skull Valley Monzogranites to the west by screens of amphibolite-grade metavolcanic and metasedimentary rocks (fig. 14).

A thin, persistent metamorphic screen, which separates the Horse Mountain monzogranite from Minnehaha and Brady Butte Granodiorites (fig. 14), shows a background lower greenschist grade increasing to amphibolite facies toward the Horse Mountain pluton. Amphibolite facies persist for up to 1 km east of the Horse Mountain pluton, as evidenced by metapelites with porphyroblastic cordierite-andalusite, muscovite, and biotite, and by mafic tuffs of hornblende-epidote-plagioclase mineralogy. Rotated inclusion trains in cordierite porphyroblasts and strong matrix foliation indicate that deformation outlasted recrystallization in this metamorphic screen. These amphibolite-grade assemblages define a low-pressure facies series, and their timing shows that the Horse Mountain monzogranite was emplaced in a syntectonic or late-tectonic setting where deformation outlasted cooling below the amphibolite isograd. Exactly the opposite is true in the metamorphic aureole of the Crazy Basin body less than 10 km to the east, where recrystallization clearly outlasted deformation, and metamorphic assemblages are indicative of a higher temperature, higher pressure facies.

Such contrasts indicate that the late-tectonic Crazy Basin granite clearly postdated the Horse Mountain pluton and the other porphyritic plutons by an important time interval, after which significantly higher pressure and higher temperature metamorphic facies developed within a broad region of crustal anatexis in the southern Bradshaw Mountains.

GROUP 4 LATE-TECTONIC PLUTONS AND ANATECTIC COMPLEXES

Latest group 4 felsic plutons and batholiths are characterized by widespread migmatite-pegmatite complexes and high-grade regional metamorphic terranes of much greater extent than those of earlier plutonic suites because they were produced by high heat flow over broad crustal regions.

Pegmatite dike swarms pervade Crazy Basin granite and the Southern Bradshaw Mountains Migmatite Complex to the south, but are minor to the north near Cleator, even though metamorphic grades are high and amphibolite facies extends throughout the Crown King-Turkey Creek region. The first amphibolite isograd lies 7 km north of the Crazy Basin contact, the second lies nearly 5 km north, and staurolite first appears 3.5 km north of the contact. Within 100-200 m of the Crazy Basin contact, sillimanite is prograded from andalusite in lineated pelitic gneisses. The absence of cordierite and the presence of staurolite throughout the Cleator metasediments indicate that the Crazy Basin metamorphism was an intermediate-pressure facies (DeWitt, 1976).

The Crazy Basin staurolite-andalusite metamorphic suite is a higher pressure facies than the low-pressure, cordierite-andalusite facies developed around the Horse Mountain monzogranite to the west (Anderson, 1986). Because the plutons are relatively close neighbors, this pressure difference cannot reflect emplacement at different crustal depths at the same time, and therefore must indicate two separate metamorphic events at different times: the lower pressure Horse Mountain series was first and was later overprinted by the higher pressure Crazy Basin series.

This timing is confirmed by textural differences in the metamorphic aureoles of each pluton: the Horse Mountain pluton produced synkinematic textures, whereas the Crazy Basin body produced postkinematic overgrowths of hornblende on earlier deformed fabrics in its northern aureole. Such late-kinematic and postkinematic textures convincingly demonstrate that emplacement of the Crazy Basin body and growth of its thermal aureole occurred after the main events of regional deformation and low-grade metamorphism of the Prescott volcanic belt. Nearly synkinematic metamorphic fabrics closer to the Crazy Basin contact show that continued deformation was localized near the body in response to its rise and thermal metamorphic peak [see Part 4]. Thus, emplacement of the Crazy Basin granite closely followed the major regional deformation of the Prescott volcanic belt.

Metapelites in the Southern Bradshaw Mountains Migmatite Complex were metamorphosed under the same high-temperature, intermediate-pressure facies series as the Crazy Basin granite. Sillimanite-bearing assemblages, locally with andalusite or staurolite, occur in the Silver Mountain region to the south, in contrast to cordierite-bearing assemblages related to the earlier Minnehaha Granodiorite.

An important difference between the southern and northern areas is that upper amphibolite facies metamorphism was of much wider regional extent to the south. The first sillimanite isograd in the north lies at the Crazy Basin contact, but in the south lies up to 3 km from the edge of the migmatite complex. The second sillimanite isograd (muscovite-K-feldspar) is absent in the north, but occurs near the edge of the migmatite complex to the south, indicating that anatectic conditions must have been widespread throughout the Southern Bradshaw Mountains Migmatite Complex. This last conclusion explains the abundance throughout the complex both of granite with eutectic minimum compositions and of partly fused migmatitic paragneiss closely resembling granite.

In summary, all pretectonic and early syntectonic plutonic bodies in the Prescott region (groups 1, 2, and 3) were emplaced under low-pressure, moderate-temperature

Table 1. Field classification of plutonic rocks in the Bradshaw Mountains-Prescott region based solely upon their crosscutting relations, petrographic similarities, and differences in deformational state, as observed in the field (P. Anderson, 1986). Petrologic descriptions of the same plutonic groups are given in table 2.

Earliest (synvolcanic and intervolcanic) gabbro and gabbro-diorite plutons		
Gabbros		**Gabbro-diorites**
gabbro-pyroxenites of Groom Creek		Lynx Creek gabbro-diorites
Dandrea Ranch gabbro		gabbro-diorite of Badger Mountain
gabbro-pyroxenites near Cordes		gabbro-diorites near Cordes
gabbro of Spruce Mountain		gabbro-diorite of West Spruce Mountain
Bluebell Mine gabbro		gabbro-diorites in Indian Hills
Towers Mountain gabbro		gabbro-diorites northeast of Mayer
Silver Spring Gulch Diabase		gabbro-diorites in Grapevine Gulch Fm
Early (pretectonic) coarse-grained tonalite-granodiorite plutons and batholiths		
(a) strongly deformed	— Cherry Springs batholith, southern parts (Bumblebee granodiorite, diorite-tonalite of Little Squaw Creek);	
	— Brady Butte Granodiorite	
(b) weakly deformed	— Cherry Springs batholith, northern and central parts (tonalite of Cherry, Badger Springs Granodiorite, quartz diorite near Turret Peak)	
	— Government Canyon Granodiorite	
(c) deformed and diked	— Wilhoit Granodiorite batholith	
	— Minnehaha Granodiorite	
	— Crooks Canyon Granodiorite	
Middle (syntectonic, late-tectonic) fine-grained granodiorite-monzogranite plutons		
Nonporphyritic	— Prescott Granodiorite (intrudes Wilhoit Batholith)	
(with dike swarms	— monzogranite of Williams Peak (intrudes Wilhoit)	
and migmatite)	— Johnson Flat Monzogranite (intrudes Crooks Canyon)	
	— Longfellow Ridge Monzogranite (intrudes Crooks Canyon)	
Porphyritic in part	— Skull Valley Monzogranite (intruded by Iron Springs)	
Late (late-tectonic) felsic (mainly monzogranite) plutons and batholiths		
(a) **Porphyritic**		
Yarnell Granodiorite and porphyritic monzogranite (intrudes Wilhoit)		
porphyritic monzogranite of Horse Mountain (intrudes Minnehaha)		
Iron Springs Porphyritic Monzogranite (intrudes Skull Valley Monzogranite)		
(b) **Nonporphyritic, migmatitic**		
Crazy Basin Quartz Monzonite (actually granite)		
Southern Bradshaw Mountains Migmatite Complex		
Posttectonic 1400-Ma felsic plutons		
(a) **Nonporphyritic**		
Dells Granite (intrudes Prescott Granodiorite and Iron Springs body)		

cordierite-andalusite facies metamorphic conditions, whereas the late-tectonic plutons, anatectites, and related migmatite complexes (group 4) formed under higher P-T conditions of an intermediate-pressure, high-temperature, staurolite-andalusite-sillimanite facies. The key factors controlling such differences are the depth of plutonic emplacement and the timing of emplacement relative to the major regional deformation and metamorphism.

PROTEROZOIC PLUTONIC EVOLUTION

The order in which plutonic suites have been described and are shown in tables 1 and 2 is close to their emplacement order. Field evidence indicates that members of some suites were emplaced so closely in time that their isotopic ages will probably be indistinguishable, but isotopic-age differences between other suites can be distinguished. The Cherry Springs batholith is an exception, because it contains plutonic phases closely related in petrogenesis, but differing in age by as much as 20 m.y. (Bowring, 1986). The total emplacement interval of all plutonic rocks in the Prescott region was approximately 100 m.y. Because only a few key plutonic bodies have been isotopically dated, it is not yet possible to accurately describe Proterozoic plutonic evolution of the region with existing isotopic data. Consequently, the sequence of plutonic, metamorphic, and tectonic events presented herein is based primarily on the detailed relative chronology of P. Anderson (1986) and this paper. The few available isotopic ages of plutons are used as reference points with which to place this relative chronology into an absolute time framework.

1800- to 1740-Ma PRETECTONIC PLUTONIC ROCKS

The first primitive magnesian basaltic flow sequences were extruded onto the deep ocean floor from low-K bimodal tholeiitic volcanic centers that evolved in an

Table 2. Petrologic classification of plutonic rocks in the Bradshaw Mountains-Prescott region using petrologic and major-element chemical criteria. "P-C" denotes petrology and chemistry and "P-G" denotes petrogenesis. Examples of each granitic type are found in table 1 and are listed in this table for all but the Group 1 category.

GROUP 1: Earliest, synvolcanic, mafic plutons and subvolcanic feeders

(a) **older gabbroic bodies:**

P-C: magnesian pyroxenite; magnesian to calcic gabbro and microgabbro; low-K to intermediate-K tholeiitic diabase; minor high-Na calcic gabbro and gabbroic anorthosite: on magnesian and iron-rich tholeiitic differentiation trends.

P-G: Derived directly from olivine tholeiite source magmas without fractionation or liquid-crystal separation, and from quartz tholeiite source magmas fractionated from olivine tholeiite parents.

(b) **younger gabbro-diorite bodies:**

P-C: calcic, ophitic gabbro-diabase; low-K calc-alkaline gabbro-diorite; differentiated gabbro-diorite-granodiorite; in situ plagioclase-pyroxene segregation related to vapor activity: early pyroxene-dominated cotectic crystallization and late plagioclase-dominated eutectic crystallization; on true iron-rich tholeiitic and low-K calc-alkaline differentiation trends.

P-G: Derived from quartz tholeiite source magmas or pyroxene-gabbro parent magmas with extensive in situ fractional crystallization from volatile activity towards leucogranodiorite end product.

GROUP 2: Early, pretectonic biotite-hornblende tonalite and granodiorite plutons and batholiths

(a) **gneissic biotite granodiorite**

examples: Brady Butte granodiorite, Crooks Canyon granodiorite

P-C: gneissic, relatively leucocratic, calc-alkaline, biotite granodiorite, hornblende-deficient, minor muscovite from subsolidus biotite reaction; sub-pyroxene feldspar-dominant cotectic crystallization.

P-G: only in volcanic belt and genetically linked to its evolution; derived from differentiated hydrous calc-alkaline dacitic magmas.

(b) **I-type hornblende-biotite granodiorite and tonalite**

examples: Government Canyon granodiorite, Wilhoit and Minnehaha granodiorites, Cherry Springs batholith

P-C: typical non-porphyritic, I-type hornblende-biotite tonalite and granodiorite with mafic diorite-quartz diorite border phases and mafic xenoliths of igneous source material: on calcic and low-K calc-alkaline differentiation trends, plagioclase zoning during fractional crystallization; batholiths and plutons internally zoned with differentiated felsic phases in cores, early mafic border phases.

P-G: hydrous quartz tholeiite source magmas derived from fusion of earlier igneous crustal material, fractionated during emplacement and upward rise.

GROUP 3: Middle (syn- and late-tectonic) leucocratic granodiorite and monzogranite plutons and dike swarms

examples: Prescott, Johnson Flat, Longfellow Ridge, Skull Valley, and Williams Peak granodiorite-monzogranites

P-C: biotite (± muscovite), relatively leucocratic fine-grained granodiorite and monzogranite; metaluminous high-K calc-alkaline, feldspar cotectic crystallization; all in the high-grade-metamorphic roots of the Prescott volcanic belt now exposed on the southwest edge of the Bradshaw Mountains.

P-G: Felsic magmas derived from anatexis of lower crustal volcanic and sedimentary rocks; generically related dike swarms governed by controlling stress field during emplacement.

GROUP 4: Late (late-tectonic) porphyritic monzogranite-granite plutons, anatectic complexes, and dike swarms

(a) **Porphyritic**

examples: Iron Springs Monzogranite, Yarnell Granodiorite-monzogranite, Horse Mountain Porphyritic Monzogranite

P-C: High-K calc-alkaline, locally alkali-calcic, porphyritic monzogranite and granite; K-feldspar-dominant cotectic crystallization from near minimum-melt compositions; late dike swarms in hydrous core phases.

P-G: mobilized from ultrametamorphic regions of high heat flow, separated and fractionally crystallized as feldspar-crystal-rich zoned magma chambers.

(b) **Non-Porphyritic**

examples: Crazy Basin Quartz Monzonite, Southern Bradshaw Mountains Migmatite Complex

P-C: High-K calc-alkaline granite-monzogranite with gneissic leucocratic and pegmatitic late phases; eutectic-dominant crystallization from minimum-melt compositions.

P-G: Derived by fusion of paragneiss in zones of high heat flow; crystallized in situ or mobilized from anatectic migmatite complexes.

intraoceanic setting at 1800 to 1780 Ma. Concurrently, the first mafic gabbro suite was emplaced primarily in a hypabyssal environment beneath the volcanic edifices. The Proterozoic crust was geochemically primitive at this early stage, being rich in Mg, Fe, Ca, and Na, and depleted in Si, Al, and K, and the gabbro bodies reflect that primitive chemistry. The subvolcanic gabbro lopoliths caused almost no contact metamorphism and were not diapiric, but they caused extensive Mg-Fe metasomatism of felsic volcanic host rocks.

Continued volcanic activity between 1780 and 1770 Ma thickened the tholeiitic pile to a point where mafic gabbro-diorite plutons rose diapirically into the upper volcanic and clastic units just below the seawater interface. The calcic-series gabbro-diorite bodies caused minor contact-metamorphic effects at deeper levels, but only metasomatism and assimilation at higher levels.

The interval 1770 to 1755 Ma was dominated by depositional hiatuses in places and extrusion of thick trimodal basalt-dacite-rhyolite sequences of Fe-rich calcic chemistry in the center of the volcanic belt. Many mafic centers are cored by gabbroic masses that are the crystallized remains of volcanic feeder necks and subvolcanic magma chambers. These gabbro feeders postdate the earliest gabbros, are coeval with their 1765-Ma volcanic host rocks, and are close in age to the gabbro-diorite bodies to the west, but are included on table 1 with synvolcanic gabbros because of similarities in petrology and tectonic setting.

By 1750 Ma the volcanic belts had thickened sufficiently to sustain the first suite of calc-alkaline granodioritic plutons. The western Government Canyon body was emplaced at about 1750 Ma into a mafic tholeiitic crust, so its chemistry is most like an I-type granitoid. The Crooks Canyon and Brady Butte bodies were emplaced at about 1750 Ma into a more evolved felsic crust to the east, so are more leucocratic in composition, lying on higher K, calcic to calc-alkaline trends. Metamorphic effects of the plutons were restricted to narrow thermal aureoles not exceeding greenschist facies, except in enclaves. Their emplacement caused local deformation of host rocks near contacts, as well as gentle warping of the volcanic pile, which in turn gave rise to differential erosion and the subsequent development of intervolcanic unconformities.

About 10 m.y. after emplacement of the first granodiorites, after the Prescott-Jerome volcanic belt had built itself southeastward, sequential intrusion of several plutonic phases into this southeast edge culminated in emplacement of the huge Cherry Springs batholith at 1740 Ma. Earliest phases were pyroxene-hornblende tonalite, diorite, quartz diorite, and gabbro, and later phases were more evolved hornblende-biotite granodiorite and tonalite. The quartz diorite of Cherry and other intrusions east of Mayer appear to have been emplaced first, followed by the central Bumblebee-Badger Springs Granodiorites and quartz diorite near Turret Peak, and then by southern mafic phases, such as the tonalite-diorite of Little Squaw Creek. Significantly later, the Bland tonalite and possibly other southern phases were emplaced at about 1720 Ma, closer to the time of regional deformation.

Southeast border diorite phases of the Cherry Springs batholith formed extensive contact-metamorphic and migmatitic zones and caused local anatexis of host rocks in the Little Squaw Creek Migmatite Complex. Northern granodiorite-quartz diorite phases of the batholith, as well as the later Bland phase, intruded tholeiitic to calc-alkaline mainly felsic volcanic-subvolcanic units, producing sharp nonmigmatitic contacts and causing narrow metamorphic aureoles that were later attenuated in the Black Canyon and Shylock tectonic zones.

Emplacement of the Cherry Springs, the first major batholith in the Proterozoic crust of central Arizona, indicates that by 1740 Ma (only 50 m.y. after inception) formative volcanism had ceased and the volcanoplutonic belt was thick enough to sustain major batholiths. Batholithic emplacement caused local deformation and metamorphism of host rocks, including uplift and stripping off of cover material, but major deformation of the belts was yet to come.

Southwest of the Prescott volcanic belt, a similar plutonic evolution is recorded. The earliest Wilhoit Granodiorite batholith was emplaced at about 1740 Ma and was fractionally derived from sources like that of the Government Canyon Granodiorite. The Minnehaha Granodiorite was also emplaced at about the same time, but does not show the compositional and structural zoning typical of the Wilhoit and Cherry Springs batholiths. All these early granodiorite plutons and batholiths are chemically much alike: except for slight alkali enrichment, they are normal metaluminous, subalkaline calcic hornblende-biotite I-type granodiorites with enclaves of lower crustal, mafic dioritic source material.

1730- to 1710-Ma SYNTECTONIC PLUTONIC ROCKS

Plutonism in the central Arizona Proterozoic crust prior to 1730 Ma occurred in a static pretectonic environment, but after 1730 Ma the structural setting of the entire volcano-plutonic arc changed toward intensified deformation as later batholiths and plutons were emplaced. All post-1730-Ma plutons are thus syntectonic or late tectonic, depending on their emplacement timing. During this same period, Texas Gulch Formation successor clastics were laid down in downwarped troughs, and incorporated eroded remnants of pretectonic plutonic rocks. The time span from the first signs of deformation to the last in this part of central Arizona was almost 30 m.y., but deformation was intermittent or spasmodic during that period, locally amplified in each area by pluton emplacement.

Post-1740-Ma syntectonic plutonic rocks were influenced both in external shape and internal fabrics by a stress field dominated by vertical tectonic transport sympathetic to northwest-southeast compression. As a result of such stress, intruding plutons became elongate northeast-southwest and assumed a pervasive northeast-trending foliation of variable intensity and structure. Not only did these syntectonic plutons mirror the deformational field, but the locations and trends of dike swarms they produced were also dictated by the stress field as the dikes invaded and migmatized the deforming host rocks.

From 1730 to 1710 Ma, peaking at 1720 Ma, syntectonic granodiorite-monzogranite plutons intruded the Prescott-Jerome volcanic belt at depth, and are now exposed along its west and southwest sides. The Johnson Flat and Longfellow Ridge bodies are enveloped by screens of deformed, metavolcanic-metasedimentary rocks migmatized by dikes from the plutons. The Prescott Granodiorite, monzogranite of Williams Peak, and their dike swarms intruded mainly the Wilhoit batholith and its host rocks. Unlike the earlier I-type Wilhoit and Minnehaha bodies, the felsic, low- to moderate-K calc-alkaline chemistries of these syntectonic plutons, with crystallization starting in the feldspar fields near eutectic compositions, suggest an origin by lower crustal anatexis of volcanoplutonic arc or oceanic crust.

A later plutonic event near 1710 Ma produced the first plutons of truly K-rich calc-alkaline character, with crystallization originating in the K-feldspar field. These plutons are weakly peraluminous K-feldspar porphyritic monzogranites with local K-rich calc-alkaline granodiorite borders, and include the Yarnell, Iron Springs, and Horse Mountain bodies. The high fluid content of the Horse Mountain body produced pegmatite dike swarms and a wide amphibolite aureole whose low-pressure, moderate-temperature, cordierite-andalusite facies series is diagnostic of the syntectonic plutons and their rather shallow crustal emplacement levels.

1700-Ma LATE-TECTONIC PLUTONIC ROCKS

By 1700 Ma, most intervolcanic crust outside the main core of the Prescott-Jerome volcanic belt had been pervaded, assimilated, migmatized, or replaced by various calc-alkaline plutons, and all volcanic, sedimentary, and plutonic rocks had been deformed by penetrative foliation and lineation. In places, such as near Prescott, the total strain state of the volcanoplutonic crust was mild, but in others such as in the Shylock high-strain zone, total strain was extreme. Only the most massive, competent volcanic and plutonic rocks, or those shielded from strain, escaped penetrative deformation.

At about 1700 Ma, middle crustal levels now exposed southwest and south of the Prescott belt underwent high-grade metamorphism of truly regional (> 300-km^2) extent as the thermal infrastructure of the crust rose in response to increased heat flow. Most of the southern Bradshaw Mountains became a region of large-scale crustal fusion, where intermediate-pressure, high-temperature, staurolite-andalusite-sillimanite metamorphic facies series prevailed, reflecting P-T conditions of 10- to 15-km lithostatic depth and temperatures exceeding granite-minimum anatexis. The Southern Bradshaw Mountains Migmatite Complex was the source region for in situ, anatectic formation of granite in the complex.

Some anatectite diapirically rose to be emplaced to the north as the high-silica, K-rich, calc-alkaline Crazy Basin granite and related leucogranites. Earlier tectonic fabrics in host rocks were thermally overprinted, deformation and vertical extension were intensified at the Crazy Basin's top in response to its rise, and high fluid flow around the body retrograded many metamorphic assemblages as the thermal complex cooled. So ended mid-Proterozoic plutonic orogeny in this part of central Arizona at approximately 1700 Ma.

The 1700-Ma Crazy Basin-Southern Bradshaw Mountains Migmatite Complex anatectic event occurred at depth as felsic volcanic and sedimentary rocks were being laid down at surficial crustal levels in the younger felsic volcanic belts to the southeast (New River-Cave Creek-Mazatzal Mountain-Diamond Butte areas). Because this anatectic event was the last stage in deformation and metamorphism of the Prescott-Jerome volcanic belt, and because its metamorphic fabrics were overprinted on tectonic fabrics of the main Prescott-Jerome deformation, it is clear that major deformation of the Prescott-Jerome belt must have occurred before deformation of the younger felsic volcanic belts, because the youngest 1700-Ma Mazatzal Group strata in those younger belts are deformed by the later deformation. Thus, the formation of the Crazy Basin-Southern Bradshaw Mountains migmatite zone demonstrates that two separate deformational events occurred in central Arizona [see tectonics paper].

ORIGINS AND SOURCES OF THE PLUTONIC SUITES

Data cited throughout this paper indicate that there is a close correlation between the chemistries of plutonic rocks and the general composition of their host crusts. This correlation clearly does not imply that plutonic rocks were derived from their immediately adjacent host rocks, because such rocks are hosts only at present emplacement levels, not in the source regions. However, the correlation does imply that the broad distribution of crustal compositions throughout central Arizona profoundly influenced, directly or indirectly, the plutonic compositions that were derived from or emplaced into each host crust.

All formative volcanic sequences of the Prescott-Jerome belts and distal sediments had been deposited before granodioritic plutonism invaded the volcanic belts and adjacent regions (only synvolcanic gabbros and intervolcanic gabbro-diorites were coeval with formative volcanic units). The lithologic makeup of the early volcanoplutonic crust was highly variable, depending on the distribution of primitive volcanic sequences versus more evolved crustal components. This original variability profoundly influenced the nature of all subsequent plutonism, as is clearly shown in figure 17, a diagram that maps crustal variability and evolution as the volcanic belt was transformed into a thickened, stabilized volcanoplutonic arc.

The oldest, most primitive part of the Prescott volcanic belt is its northwest edge near Prescott, but mafic components also exist along its east edge near the Jerome volcanic belt and as remnants in the Cherry Springs batholith. The earliest plutons emplaced into that mafic crust to the west had the highest Sr and Fe contents of all: both the Government Canyon Granodiorite and the Wilhoit batholith are the most primitive of all granodiorite-tonalite plutons, with the highest Sr contents (and lowest initial Sr ratios), highest Sr-Fe ratios, and highest values in other basic elements such as Ca and Mg. The Cherry Springs batholith was emplaced into a slightly more evolved crust, and shows exactly that difference on figure 17. Differentiation within the early bodies such as the Wilhoit batholith was along typical iron-rich tholeiitic trends (fig. 17).

Other early small plutons, such as the Brady Butte and Crooks Canyon, were derived from and emplaced into a more geochemically evolved felsic crust than were the

Government Canyon, Wilhoit and Cherry Springs bodies, and this difference is precisely reflected in their lower Sr and Fe contents (fig. 17). Similarly, the character of host crust in the Black Canyon belt southeast of the Prescott belt's main volcanic core is more felsic and evolved than in the north near Mingus Mountain, and this difference is reflected not only in the lower Sr and Fe contents of south-central Cherry Springs batholith phases (Bumblebee, Badger Springs and Little Squaw Creek) versus northern (Cherry quartz diorite and related) phases, but also in the different differentiation trends of little or no Sr enrichment in the south versus strong Sr enrichment in the north (fig. 17).

One of the most important aspects of crustal evolution (fig. 17) is that the younger plutons subsequently emplaced into these diverse crusts in syntectonic and late-tectonic settings do not have chemistries unique to the later events. Rather, their chemistries reflect the same source composition as do previous plutons emplaced into the same region (fig. 17). The Prescott Granodiorite was derived from and emplaced into the same Sr-rich crust as were the Wilhoit and Government Canyon bodies, and shares their Sr-rich source on figure 17, but with a typical younger calc-alkaline fractionation trend of no Fe enrichment. The Yarnell, Horse Mountain, and Skull Valley bodies are endemic to evolved felsic crusts with little mafic volcanic component, and are similar in composition to older plutonic rocks emplaced into the same low-Sr, low-Fe crust. The Bland tonalite is chemically like other southern phases of the Cherry Springs batholith, lying along the same trend of no Sr enrichment (fig. 17), even though it is 20 m.y. younger.

The Iron Springs and Johnson Flat plutons were emplaced at the edges of mafic volcanic regions and consequently have geochemical characteristics transitional between the other two populations. Even the latest Crazy Basin body, derived from anatexis of an essentially sedimentary crust south of the Bradshaw Mountains, plots in the lowest Fe-Sr part of figure 17, which suggests that its chemistry also directly reflects the composition of its sedimentary source crust. Other major and trace elements can be used as a basis for comparisons similar to those made here, and the same overall conclusion is apparent: *the chemical composition of the host (or source) crust governs the composition of all plutonic rocks emplaced into that crust, regardless of age of the plutonic rocks.*

Because host rocks at pluton-emplacement levels cannot be the sources of the plutonic bodies, only one possible conclusion remains. The general character of the underlying crust must have governed the compositions of the plutons, and therefore these plutons must have been derived by partial melting of that lower crust. Thus, a model of partial melting of the base of the volcanic-arc crust must be endorsed as most viable to account for the remarkably close correlation between pluton and host crust compositions in the older volcanic belts and adjacent region of central Arizona. The minor- and trace-element profiles of the plutonic rocks precisely mirrored the composition of the lower crust of the volcanic arc, even if some major fractional elements such as K and Na do not.

This evidence for lower crustal anatexis is inconsistent with models that assume that the material input to an arc came from either subducted oceanic crust or the mantle immediately subjacent to the arc crust. In the central volcanic belt, such models may be viable for generation of island-arc volcanic rocks, but they are quite inapplicable to generation of the plutonic components. Those plutonic components must have involved recycling of the earlier volcanic crust and its oceanic substratum, especially those portions that were subject to the deepest burial and most extensive fusion. In contrast, part of the alkali and vapor components of fusion must have been contributed by the subduction process, because the plutonic suites of the Prescott region have alkali contents that generally increase to the northwest, in part irrespective of alkali contents of their host crusts. This northwesterly increase in alkalis corresponds to the northwest dip of a subducted paleoslab, whose thermal structure was largely responsible for generating plutonic rocks of the Prescott region [see tectonics paper].

PART 4—PROTEROZOIC VERTICAL DEFORMATION AND ITS TECTONIC SIGNIFICANCE

Proterozoic volcanic belts throughout the world exhibit a distinctive style of deformation characterized by (1) pervasive vertical to steeply dipping foliation, (2) vertical and steeply plunging lineations and minor folds, (3) great lateral variations in strain, and (4) a lack of polyphase deformation at low metamorphic grades. Although Archean deformation appears similar, parts of Archean greenstone belts are pervaded by shallow-plunging lineations, obvious polyphase deformation, and other features distinctively *not* Proterozoic. Deformation of Phanerozoic island arcs is dominated by features of horizontal transport (major overthrust stacking, subhorizontal axes and lineations), which are even more unlike Proterozoic features. Consequently, Proterozoic deformation needs elucidation in its own right, not by analogy to the Phanerozoic or Archean, as such analogies have caused much misunderstanding of Proterozoic deformation.

Nowhere in the United States is the distinctive deformational signature of Proterozoic volcanic belts better displayed at low metamorphic grades than in central Arizona. Most Proterozoic volcanic belts in Colorado and New Mexico have been reworked by secondary deformation and metamorphism that obscures the primary deformation. In northern New Mexico for example, major thrust faults juxtapose Proterozoic crustal levels that differ greatly in deformational and metamorphic features, making it difficult to resolve the makeup of the primary Proterozoic deformation. Such overprinting is absent from the central volcanic belt of Arizona, and thus the primary Proterozoic deformation is well preserved.

Distinction between primary and secondary deformation is important because the secondary deformational histories of Proterozoic volcanic belts are all different—only features of *primary deformation* show a commonality among all Proterozoic volcanic belts worldwide. In all cases, the primary deformation occurred just after the end of formative volcanism and during initial plutonic stabilization and thickening of the volcanic belts. This part of this paper is concerned solely with elucidating the nature and kinematics of that primary deformation, which is inseparably linked to formative tectonic evolution of all Proterozoic volcanic belts. Once the early structural development of the belts is understood, the tectonic significance of such features as complex map patterns of rock units, highly variable fold plunges, and similar features that have confused former workers, will become clear.

Gastil (1953) was the first to grasp the significance of the apparent geometric dilemma regarding Proterozoic deformation that today still plagues most geologists working in the Arizona Proterozoic: *How can stratigraphy appear to be oriented toward vertical across an entire volcanic belt when the evidence is overwhelmingly for only a single major event of deformation?*

Previous models of deformation in central Arizona centered on deducing stress axes to account for the orientation of observed folds. The earliest of these fold models inferred northwest-southeast-directed principal horizontal compression to produce upright flexural folds with subhorizontal or shallowly dipping axes and lineations (C. A. Anderson, 1951, 1966, 1968b,c, 1972; C. A. Anderson and Creasey, 1958; C. A. Anderson and Blacet, 1972c; C. A. Anderson and others, 1955). This "horizontal" model explains vertical lineations as "a" slip lines and can account for vertically plunging folds only by transcurrent shear, later rotation, or by refolding of originally subhorizontal structures. The writer (and Brook, 1974) tested this model in 1974 and found it inapplicable to the central Arizona Proterozoic volcanic belts because of the overwhelming predominance of steep to vertical structures throughout the region.

Alternative models advanced to explain steep fold axes and lineations include homoclinal tilting and steep deformation (DeWitt, 1976, 1978, 1979, 1980) and widespread polyphase folding (O'Hara, 1980, 1986; Lindberg, 1986). Appalachian-style imbricate thrusting prior to the major deformation has also been postulated (Karlstrom, 1986), but it is typically only the well-layered Proterozoic clastic sequences at Archean craton edges that are regionally involved in imbricate thrusting. The volcanic belts, in contrast, accommodated crustal shortening by deforming vertically and are pervaded by steep primary fabrics, not subhorizontal ones.

Although some previously proposed models may be locally applicable to small areas, none come to grips with the overwhelming evidence that steep to vertical structures formed across a broad region of central Arizona in a single major event (not several superimposed events) of Proterozoic vertical deformation. Previous models used fold geometry to deduce stress orientation, but this paper shows that fold geometries vary greatly with strain variations, so such *fold models* cannot accurately describe Proterozoic deformation. This paper presents a *strain model of deformation* that shows how the diversity of structural features that, in the past, have been ascribed to many different types or events of folding, all occurred in a single major event of vertical *deformation.* Hence the term *Proterozoic vertical deformation* is used throughout this paper to impart the dominant character of that deformation.

The main features of Proterozoic deformation in the central volcanic belt are: (1) near-vertical foliation is most prominent and overwhelms other rock features in many places; (2) lineation is next most prominent and plunges steeply to vertically in most places, but in some regions dips less than 30 degrees; (3) foliation and lineation vary greatly in intensity throughout the belts and locally undergo extreme changes over very short distances; (4) folds are least prominent, and the presence of minor folds does not necessitate the existence of major fold closures; (5) volcanic stratigraphy does not trend northwest-southeast, perpendicular to foliation, as it would if repeated across the belts by northeast-trending, plunging regional folds; (6) major polyphase interference structures do not exist throughout the belts; and (7) virtually all rocks display evidence for only a single major event of strong vertical strain.

Minor folds with disparate attitudes and rare superimposed minor folds do exist at a few sites where changing strain regimes are to be expected, such as in high-strain zones of continued movement, or near pluton contacts where strain conditions changed markedly during deformation. However, such structures occur in a fraction of one percent of the region, and attempts to extrapolate them throughout the belts (and hence postulate regional polyphase folding) conflict radically with existing stratigraphic and structural data. Not only are large superimposed folds nonexistent throughout the belts, but the distribution of major stratigraphic units on a broad scale (figs. 3 and 8) and stratigraphy in detail (P. Anderson, 1986) clearly refute the existence of major polyphase interference structures—stratigraphy simply does not repeat across the volcanic belts in interference patterns on any scale.

Moreover, steep homoclinal tilting of strata prior to deformation cannot be invoked to explain steep lineation plunges, because of the relatively gentle (5° to 10°) nature of all unconformities (even the Texas Gulch unconformity) prior to deformation [see Part 1]. Furthermore, if lineations and folds were once subhorizontal and later folded to vertical by a superimposed event of deformation, areas must exist in the belts where fold axes and lineations retain their original subhorizontal attitude (e.g., at crests of later folds), and plunges of early lineations would vary systematically with position around the later structures (Ramsay, 1967). Such is not the case in the volcanic belts—both lineation and fold plunges vary in complex ways unrelated to later folding or rotation.

More than a decade of testing such models unsuccessfully in the Arizona Proterozoic by the writer has led to a new understanding of Proterozoic deformation that differs from all previous models. The model of Proterozoic vertical deformation presented in this paper results from persistent, detailed stratigraphic and structural work in the central Arizona Proterozoic, and accounts for the seven major features of deformation listed above. First the characteristic fabrics, major structural features, and strain variations are described below, then the model of Proterozoic vertical deformation is presented.

STRUCTURAL FABRICS AND FEATURES

Folds dominate deformed Phanerozoic sequences because they are well-layered, but in Proterozoic volcanic piles with subdued layering, folds are subordinate to foliation and lineation and are abundant only in weakly strained, well-layered strata. Foliation or cleavage predominates at lower metamorphic and deformational states, but lineation may predominate at the higher states.

FOLIATION AND CLEAVAGE

The most prominent structural feature of Proterozoic rocks in central Arizona is a vertical to near-vertical foliation that in many places overrides all other features, including bedding, folds, and lineations. In fact, steeply dipping foliation is so diagnostic of 1800- to 1650-Ma Arizona Proterozoic strata that it serves to distinguish them from rocks of all other ages (Lindgren, 1926).

The intensity of foliation or cleavage varies greatly from place to place, but not in a predictable manner. In mildly strained parts of the belts, foliation is weak to almost imperceptible and dips moderately (30° to 55°) to steeply (55° to 90°), whereas in highly strained areas, foliation is invariably vertical to near-vertical (70° to 90°). Foliation, however, does not simply steepen with increased strain: subhorizontal foliation rarely occurs in weakly deformed areas, and foliation does not become rotated to vertical with increased strain. Instead, foliation starts out in steep orientations as it becomes perceptible and simply increases in intensity as the strain increases. Shallow-dipping (< 30°) foliation occurs in atypical parts of the volcanic belts, such as near plutons, where superimposed folds are locally present, and in highly metamorphosed or migmatized areas. Some shallow foliation observed at surface results from recent down-slope bending of steep foliation on hill slopes.

The character of foliation varies greatly in several other key ways: (1) sharp lateral variations in foliation intensity exist across the volcanic belts; (2) foliation commonly converges and intensifies in zones of high strain; (3) the orientation and intensity of foliation varies between broad regions; (4) the physical makeup of foliation varies with crustal position; and (5) the presence of abundant foliation does not necessarily imply the existence of major fold closures, nor that foliation need be axial planar to fold closures.

One of the most striking features of foliation is its sharp variation in intensity from place to place. The variation exists at all scales, from the broadest differences between major parts of the belts, to the outcrop scale. In the Prescott belt, for example, foliation may vary from a weak cleavage (1 to 2 surfaces per cm) to moderate (3 to 5 per cm) to strong (5 to 10 per cm) in a single outcrop, but sharp changes to intensely (2 to 5 per mm) and extremely (>5 per mm) foliated zones can occur in only 50 to 100 meters across foliation.

Such schistose zones, in which foliation is so intense as to render rocks like paper schists (or rodded schist at higher metamorphic grades), are described here as **high-strain zones,** because a direct relationship exists between the intensity of foliation and the total strain state of the rocks. Clearly not all high-strain zones in the volcanic belts can be itemized, for they are innumerable on a detailed scale. However, only a few major high-strain zones dominate the volcanic belts on a broad scale, and are described here because of their structural and tectonic importance. Miniature equivalents of the major high-strain zones such as the Shylock exist throughout the volcanic belts: their presence on all scales is one of the distinctive features of Proterozoic vertical deformation.

Foliation in the central volcanic belt trends generally northeast, but it becomes reoriented as it converges into major (and minor) high-strain zones. This convergence of foliation is not a superimposed feature caused by later bending of an earlier fabric, but represents a continuous gradation from northeast-trending foliation of moderate intensity to foliation of high to extreme intensity that most commonly trends more northerly. The gradation occurs in all major high-strain zones (e.g., Moore Gulch and Shylock zones), and stereographic plots of structural data from such zones are not consistent with reorientation of foliation by later folding; instead, the data support field observations that foliation in the high-strain zone was oriented into parallelism with the zone when the foliation formed.

Contrasts in intensity, steepness, and orientation of foliation between broad regions is most obvious between the older Prescott-Jerome belts and the younger Cave Creek-Mazatzal Mountains-Diamond Butte belts (fig. 1). Foliation and bedding in the Prescott-Jerome belt (fig. 3) average a more northerly 020° to 045° trend than the 040° to 070° trend of foliation in the younger belts (fig. 8), a difference that persists on an even broader scale: foliation throughout southeast Arizona, on the average, trends more easterly than it does in north-central and northwest Arizona. Also, foliation in the younger belts commonly dips more shallowly than in the older belts, even though some parts of the younger belts are pervaded by steep foliation. Foliation tends to be more intense in the older belts, especially in rocks at the highest stratigraphic levels. But, because of extreme variability, some parts of the younger belts are as highly foliated as much of the older belts.

The physical makeup of foliation shows little variation in detail, but great variation on a broad scale, especially with metamorphic grade and crustal level, which are themselves in part inseparably linked. At the highest crustal levels and lowest metamorphic grades, such as in the Mazatzal Group, foliation is a widely spaced cleavage, with little metamorphic mineral growth. In unlayered volcanic units, foliation starts as acute sets of conjugate fracturelike surfaces that form cleavage-intersection lineations; this progresses to more nearly parallel sets of spaced (fracture) cleavage planes as strain increases slightly. At moderate to high stratigraphic and structural levels, which includes most stratigraphy in the central volcanic belts where greenschist metamorphic grades prevailed, foliation is defined mainly by mica, clay, and amphibole, to form ultrafissile foliation (schistosity) planes. Under such mica-forming conditions, foliation attains its greatest intensity.

At middle stratigraphic levels and middle to lower amphibolite grades, recrystallization predominates over mica cleavage, and foliation gives way progressively to a gneissic fabric. At deepest levels in the volcanic pile, corresponding to middle crustal levels after plutonism, a pervasive northeast-trending, vertical gneissosity, involving total or near-total recrystallization of all minerals and primary fabrics, becomes the rule. Where the volcanic belts project southwesterly into west-central and southwest Arizona, this vertical gneissosity is in places the only feature that distinguishes Proterozoic rocks from gneisses of other ages, and is characteristic of rocks that make up the deepest crustal remnants of the central volcanic belt. Such steeply dipping gneissosity is the tectonic equivalent, at elevated metamorphic grades and deeper crustal levels, of vertical foliation that pervades the central volcanic belt.

Although structures in plutonic rocks are not specifically described in detail here, many of the syntectonic and especially pretectonic plutons and batholiths possess a vertical foliation or gneissosity similar to that in the stratified rocks. Foliated or gneissose plutonic fabrics are best developed at the edges of plutonic bodies, especially where they border major high-strain zones (e.g., Bland quartz diorite). At deep crustal levels and high metamorphic grades, foliated plutonic fabrics give way to gneissic recrystallization and strongly developed, vertically lineated fabrics.

LINEATION

Lineations show variations that are closely analogous to variations in foliation, with two major exceptions: (1) at the lowest strain states where foliation is moderate to weak and dips moderately, lineation is usually absent; and (2) where recrystallized fabrics dominate at high metamorphic grades and strain states, lineation may become more prominent than foliation. Thus, the least strained parts of the volcanic belts—stratigraphically highest Mazatzal and Alder strata in the younger eastern belts, parts of the older Senator Formation that remained at upper crustal levels in the Prescott belt, and parts of the Jerome belt that were shielded from strain—are all devoid of significant lineation.

In contrast, central parts of the volcanic belts (figs. 3 and 8) are dominated by steeply plunging lineations that increase in intensity and usually steepen as strain increases. Mineral, cleavage-cleavage- and cleavage-bedding-intersection lineations plunge steeply (50° to 80°) northeast throughout much of the Prescott belt, but plunge less than 45 degrees to the north in strata that remained at high structural levels. In the south near the Crazy Basin body, *extended strain* from diapiric pluton rise distorted lineations past vertical to plunge steeply south throughout much of the Cleator Formation, Black Canyon belt, and Southern Bradshaw Mountains Migmatite Complex. The Jerome belt, in contrast, has mostly weak lineation or no lineation, but a few higher strain areas in the north have near-vertical mineral lineations.

Older rock units of the New River volcanic belt are pervaded by steeply northeast- and southwest-plunging mineral lineation resulting from high strain in the Moore Gulch shear zone, or local elevated metamorphic conditions. Younger rocks of the New River Mountains Felsic Complex, in comparison, are mildly strained, weakly foliated, and generally not lineated. Near-vertical fabrics and steeply plunging mineral lineations pervade the northern Cave Creek volcanic belt, affecting both Union Hills Group mafic volcanics and some Alder sedimentary strata; but strain is low to the south, so the type areas of the Union Hills Group and Alder strata near Cave Creek are weakly strained and unlineated.

Similar variations in strain and in intensity and plunge of lineation exist throughout the Mazatzal Mountains-

Diamond Butte volcanic belts, with the steepest and most intense lineations likewise occurring in the central, highest strain parts of the belts. On the flanks of the volcanic belts, near plutons, in uppermost ignimbrites of the felsic complexes, and in Mazatzal Group strata, lineations are much less intense or absent in many places. However, neither Mazatzal strata nor the ignimbrites everywhere escaped high strain, because both units are strongly deformed and lineated in the Gun Creek shear zone.

Thus, lineation intensity correlates directly to strain intensity, but the character and steepness of lineation plunge do not necessarily show the same correlation. Metamorphic conditions during deformation determine the physical character of lineation more than strain: at low metamorphic grades, lineations are typically cleavage intersections and micaceous mineral alignment; higher strain at low metamorphic grades does not necessarily form stronger mineral lineation, because micas become more highly sheared out and shredded. In fact, high strain at low metamorphic grades promotes planar fabrics—foliation and planar mica alignment—rather than lineated fabrics. Mainly under elevated metamorphic conditions are mineral recrystallization lineations well developed, as in gneisses in the southwestern parts of the central volcanic belt.

The steepness in lineation plunge correlates only partly to strain intensity, in so far as (1) low-strain regions are unlineated, (2) high-strain zones show steeply plunging lineations, (3) many areas show steeper lineations with increased strain, and (4) moderate- to low-strain regions have moderate- to shallow-plunging lineations. However, a 1:1 correlation does not exist between strain intensity and lineation steepness, because subhorizontal lineations do not pervade the least strained areas, except at highest stratigraphic levels. Herein lies a crucially important feature of Proterozoic vertical deformation: where rocks become sufficiently foliated and strained to show lineated fabrics, lineations typically first appear with *moderate to steep* plunges (40° to 55°), *not* shallow plunges. Then with higher strain, lineations usually maintain 55° to 70° plunges until the highest strain states are reached, where lineations plunge within 5° of vertical. These relations are vitally important to revealing the process of Proterozoic vertical deformation.

Instead of correlating exactly to strain intensity, lineation steepness correlates more to position in the volcanic belt, such that lineations in any one part of the belt all have similar plunges, within a range of variability. At lower strain states, lineations are most highly variable in plunge, but at higher metamorphic grades and higher strain states, greater consistency in plunge is achieved. Because high-strain zones such as the Shylock-Black Canyon zone are so impressive in their consistent orientation and steep plunge of lineations, the broader regions of moderate to low strain, which make up by far the greatest portion of the volcanic belts, tend to go unnoticed. Such lower strain regions are characterized by *wide scatter* in orientation and plunge of both lineations and minor fold axes, as well as very sinuous axial traces.

The most important point is that lineations did not start out with subhorizontal attitudes and become progressively rotated toward vertical as the rocks attained higher strain states. Instead, evidence shows that at the point where rocks became sufficiently strained to develop lineations, the lineations started out with moderately plunging orientations, and largely retained their original attitude throughout the history of strain. This means that lineations originated with shallow to moderate plunges in some places and with steeper plunges in others, depending on the geometry of strain that was imposed on the rocks at each particular locality. Consequently, the orientation, plunge, and intensity of lineation can be used to directly map strain variations throughout the volcanic belts, and all parameters together describe the strain state.

DISTORTED PRIMARY FEATURES

The above comparisons and contrasts depend upon evaluating the strain state of the rocks independently of either foliation or lineation. This task is aided in volcanic belts by the abundance of originally equant or ovoid primary features such as bombs, lapilli, tephra, various other fragments, and ubiquitous phenocrysts, all of which make good strain gauges. Flattening ratios of such fragments during deposition and compaction are known and largely dependent upon fragment and flow type; any further distortion beyond these ratios is a measure of the tectonic strain at that locality. By examining axial ratios of distorted fragments and crystals at hundreds of sites, an accurate picture of the total strain state of the volcanic belt and its variations has been built. Once the strain state is known, variations in foliation and lineation intensity and steepness can be independently correlated to strain variations.

In many weakly to unstrained parts of the volcanic belts, structures such as pillows, fragments, and crystals are almost undistorted, but in highly strained regions, distortion is so severe that original structures are scarcely recognizable. At extreme strain states, white chalky streaks signify original feldspar crystals degraded to micas and distorted to axial ratios of 1:3:20 or more, and strongly elongated vertical rods with subtly different texture to the matrix represent fragments distorted to axial ratios of 1:5:30 or more. In less severely strained areas, axial ratios are typically 1:3:5, but a complete spectrum exists in the volcanic belts from almost unstrained to ultrastrained features. Distortion is markedly triaxial in most areas, whereas plane-strain ellipsoids are found mainly in certain strain environments (fig. 20).

FOLD STRUCTURES

The diversity of fold forms in the Arizona Proterozoic is as great as the variety of structural settings and crustal levels, and ranges from the most open concentric flexural folds in uppermost competent stratigraphic units (e.g., Mazatzal quartzite) to passive isoclinal folds in deeper migmatized roots of the volcanic belts. Despite this broad range, a particular fold type is most prevalent in the volcanic belts: it is a *tight, similar fold* with an interlimb angle of typically 40 degrees. Competent layers such as chert beds deform differently to adjacent fissile strata by breaking, buckling in axial cores, and becoming separated from their matrix. Thus, passive slip and flow folds in ductile beds are commonly disharmonic to folds in adjacent competent beds. Fold shapes are symmetric and amplitudes equal wavelengths in areas of irrotational strain, but are asymmetric under rotational strain, where wavelengths are greater than amplitudes.

With increased strain, fold forms close, amplitudes increase, and many dislocation planes develop that translate folds along the predominant direction of slip, disrupting the coherency of both folds and stratigraphy. Greater strain dismembers the folds along ubiquitous slip lines as the fissile matrix passively accommodates most shear. Finally at high-strain states, all minor folds become transformed into disarrayed rootless axial clusters, the net slip on individual dislocation planes typically exceeds original fold amplitudes by an order of magnitude, and the rocks are transposed into chaos on an outcrop scale. Remarkably, however, major stratigraphic relations in such areas are always preserved on a broader scale.

Minor folds are not obvious in some areas because layering is not as well developed in volcanic piles as in bedded sedimentary sequences. But this only partly explains the paucity of minor folds, because even the well-bedded sequences such as the Alder Group lack obvious minor folds in many places. Regardless of host-rock lithology, minor folds are ubiquitous where foliation diverges significantly from bedding (e.g., around plutons). Folds are therefore best expressed where the following conditions are met: (1) where foliation and bedding have significantly different orientations; (2) where bedding is well developed (as in tuffaceous strata); and (3) where strain is moderate to low.

Foliation is subconcordant to bedding strike in much of the volcanic belts, especially where strain is high, and this low angular difference decreases the incidence of minor folds. Such regions are conventionally thought of as on the limbs of major regional fold closures, but this is not true in central Arizona—these regions are where foliation formed close in orientation to bedding, or where bedding was sheared into parallelism with foliation without folding. The development of minor folds is favored at lower strain states, not at the higher strain states, as might be expected. Increased strain transposes folds into axial rods that bear little resemblance to original folds. Thus, moderately to weakly strained regions display the greatest abundance of minor folds.

Unlike lineations, which correlate to strain state, the axial plunges of minor folds are much more variable, even though the associated axial planes are usually coplanar with regional foliation. In areas where mineral lineations have uniform steep plunges, fold plunges typically vary from steep to moderate. Shallow-plunging fold axes tend to dominate the higher stratigraphic levels and steeply plunging axes the deeper levels, but there are numerous exceptions.

Near-vertically plunging folds cannot be described as anticlines or synclines, only as folds closing in directions such as north or south, or as north-facing or south-facing folds, if tops are known. Facing directions of beds determined from a few folds, however, cannot be extrapolated to decide regional facing of stratigraphic sequences across the entire region. The dislocation planes noted previously are everpresent and cut almost invisibly through the strata to juxtapose north-facing and south-facing fold closures in direct opposition to one another. The presence of such structures on a broad scale invalidates all inferences about regional facing directions perpendicular to foliation.

It is incorrect to imply that minor folds seen in outcrop must be parasitic to some major structure or must show the direction in which the major structure should close. Consistent fold asymmetries can be traced right across the belts without ever encountering major closures that repeat rock units on the scale of 10 to 20 km. Such inferred "regional folds" simply do not exist in the volcanic belts, as evidenced by the distribution of major stratigraphic units (figs. 3 and 8). The absence of folds at this broadest scale contrasts markedly to their presence within major formations on a 1- to 4-km scale (just visible on figs. 3 and 8), their common occurrence at detailed scales, and their greatest abundance on an outcrop scale. Clearly this indicates a special interaction between stratigraphy and strain.

FOLD-FOLIATION-BEDDING RELATIONSHIPS

Well-layered sedimentary sequences have many horizontal competency contrasts that become structurally dominant during deformation, largely control the style and geometry of folds, and govern how foliation can be imposed on the sequence. Thick volcanic piles, in contrast, especially those retrograded to micaceous assemblages by greenschist metamorphism, are structurally incompetent masses with subdued competency contrasts. Bedding in such rocks acts passively during deformation, so foliation can be superimposed at any angle across the volcanic pile largely in disregard to bedding.

This is exactly what one sees in the field. Foliation appears to be "stamped" uniformly across the volcanic terrane as if bedding did not exist. Major changes in ductility, such as from subvolcanic bodies and massive flows to tuffs and volcaniclastic metasedimentary rocks, affect the intensity and orientation of foliation, but do not deflect it more than 10° to 15° from its regional trend. Bedding, in contrast, undergoes great variability on a detailed scale, some of which can be attributed to folding, but much is due to primary stratigraphic interlensing (fig. 2). Because of such interlensing and facies changes, the angular relation between bedding and foliation, in detail, changes constantly along strike. Thus, it is not coincidental that the greatest abundance of minor folds occurs in areas of regional facies changes.

Despite the complex interplay between bedding and foliation in detail, regional bedding trends maintain small but consistent acute angles to regional foliation trends (e.g., 5° to 15° more easterly or more northerly than foliation) over surprisingly large areas of the volcanic belts without ever culminating in major fold closures. Such persistent angular discordances result from foliation being superimposed obliquely across the axis of the volcanic belt, and in no way implies structural "facing directions" relative to regional fold closures. In fact, strata in the central volcanic belt are not necessarily folded about foliation, and major rock units do not repeat about inferred axes of regional closure. *The foliation is an expression of regional strain, not of folding.*

The independence of bedding and foliation in deformed volcanic piles is a key factor in understanding Proterozoic vertical deformation, because it means that the existence of foliation implies absolutely nothing about regional fold closures. In fact, the concept that foliation must always be axial planar to major closures, with minor folds parasitic to the same closures, is one of the greatest hindrances to clearly understanding regional structure of the belts.

Each volcanic belt has many examples of the lack of major fold closures. Bedding trends more easterly than foliation and minor folds are consistently z-asymmetric throughout much of the Bluebell Mine and Peck Canyon Formations (fig. 3), a geometry inferred by other workers to lie between a major north-closing fold to the northeast and a major south-closing fold to the southwest (DeWitt, 1976; O'Hara, 1980, 1986). Neither fold exists: strain increases in the Shylock zone to the northeast and mafic volcanic formations of the Mayer Group do not extend into it or beyond it (fig. 3); iron formation to the southwest does not close in the regional fold of DeWitt (1976). Recent work shows that different iron formations merely converge as strain increases near the Crazy Basin body, and entirely different sequences exist either side of the mafic volcanics.

Steeply plunging minor folds in rhyolite and iron formation east of the Iron King mine are s-asymmetric where foliation trends more easterly than bedding. Many have inferred (e.g., O'Hara, 1980) that the narrow prong of Texas Gulch Formation (fig. 3) cores a north-closing, south-facing regional fold, but once again, stratigraphic units do not repeat either side of the inferred axis; instead, Spud Mountain strata lap northeasterly over Agua Fria basalts into the Indian Hills, and Texas Gulch strata lap unconformably over both formations. Likewise, Spud Mountain strata lap over many units to the south [see Part 1], in relationships that are inconsistent with the presence of major fold closures.

Detailed (1:2,400) mapping of the entire northeast end of the Jerome volcanic belt (P. Anderson, 1986) shows that stratigraphy comprises a north-facing section with no major closures or major stratigraphic repeats west of the Verde fault (figs. 3 and 4). Although this detailed mapping endeavored to substantiate the regional folds inferred by others (Anderson and Creasey, 1958; Lindberg, 1986) from less detailed mapping in the same area, the stratigraphy that was traced out in intricate detail showed clearly that the regional folds inferred earlier from either map patterns or alteration zones are not valid.

The Cave Creek volcanic belt, like the Prescott belt, contains folds up to 4 km in amplitude within formations (fig. 8), but major closures that repeat stratigraphy on a 10- to 20-km scale also appear to be lacking. One of the largest structures is a 3-km s-fold within the Cramm Mountain Formation near Grays Gulch (fig. 8), which repeats neither Grays Gulch strata to the south nor Cramm Mountain volcanics to the north, so it is not a major closure. In the Mazatzal Mountains, Red Rock Rhyolite might appear to core a major shallow-plunging regional fold, but stratigraphic units do not repeat on either side [see Part 2], and Alder strata on both sides are unconformably overlapped by the Red Rock Rhyolite. In the Diamond Butte area, Sheep Basin Mountain might also appear to core a north-closing, south-facing regional fold, but once again, major units such as the Alder Group are not repeated on either side (fig. 8).

An exhaustive number of such examples could be cited to show that the major stratigraphic units are not repeated systematically across the belts perpendicular to foliation by major regional folds. Instead, some rock units seem to occupy the cores of isolated large folds, but adjacent stratigraphic units do not follow the same fold patterns or wrap around the "core" to define a major closure. This fold geometry therefore suggests a style of deformation that took advantage of the heterogeneity of the volcanic pile on a very broad scale, whereby domal volcanic centers and lensoidal stratigraphic units were deformed into their own isolated fold forms. Thus, regional folds that repeat strata across the belts are not part of the Proterozoic structure of central Arizona.

HIGH-STRAIN ZONES

High-strain zones have not been previously recognized as integral to Proterozoic vertical deformation in central Arizona, nor as a key component in Arizona's tectonic evolution. Consequently, there exists no accurate description of these features and their tectonic and structural importance; previous studies described them as faults and attributed large lateral offsets to them. The zones and the nature of their movements are examined here, starting with the largest and most misunderstood—the Shylock zone in the Prescott-Jerome volcanic belt.

SHYLOCK AND BLACK CANYON ZONES

Structures and Fabrics

The Shylock zone is the northern part of a 75-km-long, 1.5-to 3-km-wide structure that trends due north across the northeast Proterozoic fabric of central Arizona (fig. 3), dominates the structure of the Prescott-Jerome belt, and is the most profound high-strain zone in the Arizona Proterozoic. A narrow covered part near Mayer is a convenient division between the northern segment, called the Shylock zone, and the southern segment, called the Black Canyon zone or belt (P. Anderson, 1986). This distinction is appropriate for detailed discussion of the zone because of different strain environments and histories in the northern and southern segments, but the entire 75 km-long feature can be termed the Shylock zone.

The Shylock zone was originally defined and mapped as a fault (C. A. Anderson and Creasey, 1958), and later viewed as a major wrench fault (C. A. Anderson, 1968b). The justification was that volcanic rocks in the Indian Hills were assumed to be part of the Prescott belt, and previous mapping showed no correlation between rocks west of the Shylock zone and those on Mingus Mountain to the east (Anderson and Creasey, 1958; C. A. Anderson and others, 1971; Anderson and Blacet, 1972b). It is now known however (P. Anderson, 1986) [see Part 1] that the distinctive Grapevine Gulch Formation and its gabbro-diorite bodies extend across the Shylock zone with no major apparent offset (fig. 3), so the wrench-fault hypothesis is no longer valid. What then is the Shylock zone?

The first indication comes from what happens to Grapevine Gulch strata in the Shylock zone. The southern limit of Grapevine Gulch Formation projects linearly from southern Mingus Mountain northwest into the Indian Hills (fig. 3), but in the Shylock zone it is extended southward almost 15 km toward the Binghampton mine. Grapevine Gulch strata in the zone are highly sheared and flattened, and their distinctive rhyolite chips are drawn out into 1:3:12 vertically elongated white streaks. Development of such prolate ellipsoidal (cigar) shapes is distinctive of high-strain zones and suggests kinematics vastly different than a simple fault offset.

The possibility of major vertical offset was tested in the most severely strained portion of the Shylock zone near the Binghampton mine, where volcanic rocks are reduced to papery clay-mica schists. Detailed mapping both by Brook (1974) and the writer traced phenocrystic marker units through isoclinal folds continuously across the zone of maximum strain and into a lower strain area to the east, thus establishing stratigraphic continuity across the highest strain part of the Shylock zone. Northwest of the Binghampton mine, the preservation of key stratigraphic relations—the lapping of Binghampton mine strata over Grapevine Gulch strata, the lapping of the Cleator Formation over the Round Hill Formation, and the lapping of Texas Gulch strata over many units—all support stratigraphic continuity across the western part of the zone as well.

Such stratigraphic continuity prohibits any major horizontal or vertical offset across the Shylock zone. Instead, the highly attenuated nature of strata in the zone demonstrates exactly what the Shylock is—*a classic zone of high strain,* with strong horizontal shortening and vertical extension, but no major offset. Rock fabrics are dominated by intense vertical foliation, vertical to steeply south plunging lineations, rods with 1:3:12 elongation, and steeply plunging tight to isoclinal or rootless folds. Shallowly north plunging fold axes coplanar with the dominant steep fold axes occur in a few outcrops (Brook, 1974; P. Anderson, 1986), as is always the case in high-strain zones with long, complex movement histories.

The Black Canyon belt to the south, in contrast, does involve vertical shear offset. In this more recrystallized, southern environment, strong mineral lineations with consistent 70° to 80° south plunges and tectonic rods of 1:4:20 elongation overwhelm other rock fabrics, fold-axis plunges are uniformly steep, and extreme vertical strain produced dislocation of some stratigraphic units.

Strain Variations

Rather than a single offset plane (fault) or a series of planes with net offset (shear zone), the Shylock is a zone of *distributed strain* in which rocks are horizontally shortened and vertically extended (contacts are telescoped) without necessarily any offset. Because strain was distributed over its width, the Shylock zone has no sharp boundaries. Instead, the major rock sequences themselves largely influence how the Shylock zone is defined at any place, because strain intensity varies with competencies of the rock units, and because rock-unit boundaries typically parallel foliation in the zone. Hence, the greatest strain gradients commonly occur at major lithologic boundaries. Foliation and lineation in the zone developed at the same time that regional foliation formed.

Strain in the Shylock zone varies from moderate to extreme over short distances, but rocks near its central axis are typically the most intensely strained. In the north, the zone is only 0.5 km wide where it involves Texas Gulch strata, but widens to 1 to 1.5 km where it includes the Grapevine Gulch and Round Hill Formations. This narrow northern segment borders the Cherry Springs batholith (fig. 3), but the zone widens to 2 km to the south and its boundaries become diffuse where only sheared volcanic strata are involved. Southeast of Mayer the zone includes 1 km of Black Canyon Creek Group felsic volcanics and 1.3

km of Cleator metasedimentary rocks (fig. 3). Thus, strain dissipates over a progressively wider segment of stratigraphy to the south as the Shylock zone broadens, but narrow zones of locally extreme strain still occur within it.

The Black Canyon belt to the south is a continuation of this widening and strain dispersal, but strain appears to be more evenly distributed to the south, rather than being concentrated in a central axis. The high-strain zone that makes up the Black Canyon belt can be viewed as either including the full 3-km width of the Townsend Butte and Cleator Formations, or only parts of them near the main felsic volcanic-sedimentary interface (cf. Winn, 1982). Fabrics of amphibolite-grade metamorphic recrystallization in metasedimentary rocks of the western Cleator Formation near the Crazy Basin body and Southern Bradshaw Mountains Migmatite Complex overprint fabrics of vertical strain in the Black Canyon belt. This indicates that emplacement of the 1700-Ma Crazy Basin body and its regional metamorphism postdated development of the Black Canyon-Shylock zone, which was an integral part of the main Proterozoic deformation.

Kinematics

These and other relationships demonstrate that the Shylock zone is not a fault with discrete offset, nor does it have a significant amount of transcurrent (wrench-fault) movement. The zone does contain some small, postdeformational, brittle faults that are unrelated to formation of the zone and to regional Proterozoic deformation. Movement during regional deformation was almost exclusively vertical and distributed over the full width of the zone, as is the case with all major high-strain zones. The southern Black Canyon segment, in addition, developed significant vertical shear offset later in its history, juxtaposing different crustal levels.

Field evidence shows that rocks in the Shylock zone underwent horizontal shortening and vertical extension to deform originally spherical shapes into vertically elongate cigar shapes. For each increment of flattening in both horizontal directions, proportionately more elongation takes place in the vertical direction. This kinematic mechanism is described as pure shear (Ramsay, 1967) and is used in this paper without necessarily implying that either intermediate axes of the strain ellipsoid or volume remained unchanged. As defined, pure shear reorients subhorizontal or shallowly plunging features (e.g., axes of an ovoid fragment) toward vertical purely by *internal distortion—no rotation is involved.* In contrast, *simple shear* (sliding one side of a deck of cards up and the other side down) kinematically involves microscopic offset on each slip surface, macroscopically reorients features toward vertical by *rotation,* and involves a consistent sense of shear in a direction perpendicular to foliation. There is also evidence for simple shear in parts of the zone, as will be noted.

Rocks in the Shylock zone are so intensely foliated that they appear to have sustained large amounts of net offset (simple shear movement); however, a diagnostic feature of the Shylock and all other high-strain zones is that rock units can be mapped into them, through tight folds within them, then out the other side with no significant net offset on a broad scale. This requires that the kinematics of strain within such zones must have been largely *pure shear.* In plan, rock contacts and widths are strongly telescoped in high-strain zones (fig. 18), and in vertical section, rock units are drawn down to great depths in the zones but remain in comparable positions on either side (fig. 18).

Thus, high-strain zones contain rocks that are *highly strained*; they are not zones of distributed shear with major offset across them, and should therefore not be termed "shear zones," a term that implies offset. *High-strain zone* precisely describes not only their physical and mappable attributes, but also their kinematics and evolution.

Another crucial difference is that shear zones primarily cut across, attenuate, and offset earlier formed fabrics, as is common in high-grade gneiss terranes (Ramsay and Graham, 1970; Tchalenko, 1970). High-strain zones are quite different: they formed during initial deformation of the volcanic belts and are an integral part of that deformation: their foliation, lineation and fold fabrics are identical to those in the adjacent deformed regions except for the higher total strain manifested in the zones. Gradual changes in foliation orientation and lineation plunge on entering high-strain zones (such as the Shylock) thus do not reflect later shearing of an earlier fabric, but rather directly map the changing total strain state as rock units enter the zones, and consequently provide an accurate picture of strain gradients in the zones.

History

It is not coincidental that the largest and oldest high-strain zone in central Arizona lies along the western edge of the largest early Proterozoic batholith in central Arizona. The Shylock zone lies between the Cherry Springs batholith to the east (fig. 14) and Prescott and Wilhoit Granodiorites under cover to the west [see Part 3], and the Black Canyon belt borders the southwestern side of the Cherry Springs batholith. The position and emplacement of these major pretectonic batholiths not only governed the location of the Shylock zone and Black Canyon belt, but also largely influenced their movement histories.

The Cherry Springs batholith was emplaced before regional deformation, and its north-trending western edge was one of the most profound weakness zones in central Arizona, spanning the full thickness of the early Proterozoic crust. This crustal weakness evolved into the Shylock-Black Canyon zone. The earliest demonstrable movement on the zone was vertical, either as a normal fault or a zone of distributed shear: 1740-Ma batholith phases were uplifted and a narrow trough formed along their western side in which Texas Gulch strata accumulated at about 1720 Ma, as the Prescott-Jerome volcanic belt became mostly emergent [see Part 2]. The graben may have formed by an early event of vertical strain, which at depth could have been in response to emplacement of the 1720-Ma Bland tonalite, but movement near the surface was on a steep fault. This Texas Gulch graben was the ancestor to subsequent vertical movement on the Shylock zone.

As the Prescott-Jerome belt underwent primary vertical deformation prior to 1700 Ma, the Shylock zone became the natural locus for intensified strain, because it not only had a previous history of vertical movement, but was also where thinly bedded tuffaceous strata existed between major volcanic centers (fig. 18). The huge Agua Fria volcanic center lay to the west, and the younger Copper Mountain volcanic edifice lay to the east. In between were thin tuffaceous sequences of the Round Hill, Cleator, and Texas Gulch Formations, and distal Grapevine Gulch and Binghampton mine volcaniclastics. The coincidence of so many tuffaceous units in a single zone implies that the Shylock had an ancestry as old as inception of the Prescott-Jerome volcanic belt.

During deformation, rocks in the Shylock zone were subject to extreme horizontal shortening and vertical extension. This produced intense vertical foliation and lineation, steeply plunging minor folds, and larger isoclinal folds in felsic volcaniclastic strata west of the subaerial Copper Mountain center. Fissile felsic tuffs near the Binghampton mine in the axis of highest strain were almost totally converted to lineated papery schists. During later stages of vertical deformation, ultrafissile planar rock fabrics became highly susceptible to kinking and folding as strain progressed, and earlier vertical fabrics were folded about shallower coplanar folds in some places. After the main deformation, small faults of less than 1-km offset developed in the zone with minor shear folds, brecciation, and clay-gouge development.

The Shylock zone appears most like a fault in the north where strain is resolved into a narrow zone of stratigraphically highest Texas Gulch strata, but even this highest level, rocks shows vertical elongation, not transcurrent offset. Southward progression sees a slight increase in metamorphic grade and a more ductile response to deformation, as strain is more widely distributed throughout the strata. Finally, in the southernmost segment of the Black Canyon belt near Black Canyon City, strain is penetrative throughout both volcanic and plutonic rocks that were much more deeply buried than the Texas Gulch Formation. Such relationships reveal another key feature of the Shylock zone—a traverse from north to south down its length is a transect toward deeper crustal levels, so the changing aspect of the zone, such as gradual distribution of strain over greater widths, reflects the changing structural response with crustal depth.

Had the Southern Bradshaw Mountains-Crazy Basin plutonic rocks formed prior to regional deformation, the Black Canyon zone would have been as tightly sandwiched between two major batholiths and as narrow as the northern Shylock zone. Instead, the Southern Bradshaw Mountains anatectic event at deep crustal levels and emplacement of the Crazy Basin body at higher levels postdated the main deformation, so higher grade metamorphic and shear fabrics that formed around the body in response to its rise overprinted earlier fabrics of the main deformational event. Because both fabrics are near vertical and formed close in time to one another, they can be viewed as manifestations of a continuum of regional vertical deformation that produced the Shylock and Black Canyon zones.

CHAPARRAL SHEAR ZONE

Structures and Fabrics

The Chaparral zone (figs. 3, 14) is a major shear zone in the northern Prescott belt with similarities to the Shylock zone, but a different kinematic history. First, rather than paralleling rock contacts, the Chaparral zone cuts obliquely across them, so offset is measurable. Second, instead of distributed strain, the Chaparral is a series of anastomosing shear zones that diverge in competent plutonic rock but converge in ductile volcanic strata. Third, strain was not just vertical: mylonitic granites have shallow lineations where offset is greatest. Fourth, metamorphic conditions of mylonitization exceed those in other Proterozoic high-strain zones in the central volcanic belt. Fifth, the Chaparral is largely a shear zone, as it clearly has lateral offset across the zone and a simple-shear kinematic history. Finally, much deformation occurred in the late stages of regional deformation of the volcanic belt. Because of these differences, the Chaparral is termed a shear zone, not a high-strain zone.

The Chaparral shear zone is intruded by the Laramide Walker pluton near Mount Union, but extends 15 km northeasterly toward the Iron King mine, where Spud Mountain dacites are intersheared with granite-rhyolite dikes and bodies over a 1-km-wide zone; past this it is covered (fig. 3). It is well defined in the northeast, cutting Agua Fria volcanics and Spud Mountain dacitic breccias, which are sheared and lineated up to 500 m north of the zone, and cutting gabbro-diorite-granodiorite bodies, which are foliated and lineated in and near the shear zone. The zone is best defined near Walker where it cuts the Mt. Elliott red granite, producing fine-grained feldspar-porphyroclastic mylonite, ribbon mylonite, and gneiss up to 100 m from the main shear zone. However, southwest of Walker, no clearly defined shear zone seems to exist, and strain is spread throughout metavolcanic-sedimentary strata of the Senator and Mount Tritle Formations. The

Chaparral zone thus dissipates in distributed movement on foliation planes parallel to bedding and is effectively lost to the southwest.

Kinematics

Offset on the Chaparral shear zone is given by the net displacement of the Mount Elliott granite across the zone. Its contacts are sheared dextrally by 500 to 1,000 m near each of three main shears (fig. 14), but two show about 200 m apparent sinistral offset opposite to the shear sense. Hence dextral shear displacement may have originally been more than 2 km, but later sinistral fault offset effectively reduces net dextral displacement on the Chaparral zone to 1.5 km; also, the zone is oblique to the granite, so true offset may be even less.

The presence of (1) steeply plunging lineations (elongate fragments and streaked feldspar crystals) in volcanic rocks, (2) shallow mineral lineations in granitic mylonites, (3) consistent dextral shear of granite on approaching the shear zones, and yet (4) late apparent sinistral offset on shears, together show that the Chaparral zone had a complex movement history. Either vertical extension exceeded the horizontal shear component in an oblique-slip geometry, or else two or more distinct periods of movement caused the different shear and fault offsets. If two periods of movement, the first was near-vertical strain in a high-strain zone, and the latter was mylonitization and sinistral offset of granites; if three, the later event had two components: first dextral shear just after emplacement of granite, then oblique offset on ductile faults.

History

The Chaparral zone probably originated soon after conception of the Prescott volcanic belt and evolved throughout and after major deformation of the volcanic rocks. The zone lay at an original tuffaceous interface between two major edifices of the evolving Prescott belt: the huge Senator mafic center of the Bradshaw Mountains Group lay to the northwest, and the mafic volcanic chain of the Mayer Group to the southeast had grown westerly to shed Spud Mountain volcaniclastics into a deep trough between the two centers [see Part 1]. But Spud Mountain strata only partly filled the trough: a region of very thin crust capped by Mount Tritle and Spud Mountain tuffs still lay to the west.

This weak zone was first intruded by pretectonic gabbro-diorites, then developed into a high-strain zone, and was later pervaded by granite during syntectonic emplacement of Johnson Flat and Longfellow Ridge plutons deeper in the volcanic pile. Vertical strain likely dominated the Chaparral zone prior to granite emplacement, but mylonitization and ductile shear at elevated metamorphic grades followed granite emplacement. The granite of Mt. Elliott may have been emplaced slightly later than the granite dikes, causing ductile oblique faults opposite to the preceding oblique shear. Thus, the Chaparral zone was probably conceived as a high-strain zone, but the strain environment changed during syntectonic granite emplacement to oblique shear and faulting.

The Chaparral zone was mapped previously as a simple fault separating what Anderson and Blacet (1972b) thought were two different volcanic formations: Green Gulch volcanics were confined to the north of the fault, and Spud Mountain volcanics were restricted to the south. Recent mapping by the writer has shown this two-fold division to be inaccurate, because Spud Mountain breccias extend north of the Chaparral zone, and parts of the Bradshaw Mountains Group extend to the south [see Part 1 and fig. 3]. Nevertheless, Anderson and Blacet (1972c) located, named, and recognized the Chaparral zone as a fundamental structure.

MOORE GULCH SHEAR ZONE

General Structure

The protracted deformational history of the Shylock zone is typical of Proterozoic high-strain zones, but the width and great extent of the zone (exceeding the thickness of the crust) surpass most other zones in central Arizona. The other high-strain zone of comparable extent is the Moore Gulch shear zone in the New River volcanic belt (fig. 8). Both the Shylock and Moore Gulch zones originated as boundaries to the huge horst block of Cherry Springs batholith that developed just prior to Texas Gulch sedimentation. The structural anomaly represented by this batholith and its bounding scarps not only penetrated the entire crust, but its position between the older and younger portions of the central volcanic belt (fig. 1) may track a zone of crustal weakness back to inception of the volcanic belts [see tectonics paper].

In many respects, the Moore Gulch Shear zone is a mirror image of the Shylock zone and appears to have been largely coeval with it: (1) both zones lie on opposite sides of the Cherry Springs batholith; (2) both originated as scarps to the batholith horst; (3) both marked the graben edges where Texas Gulch and Alder Group strata accumulated in narrow troughs against the scarp; (4) both were conceived as weak zones in areas of originally thin supracrustal deposits between major volcanic centers; and (5) both are high-strain zones dominated by strong vertical extension. The Moore Gulch zone, however, shows much more complex strain variations and a more variable strain history than either the Shylock or Chaparral zones.

Extending from New River northeast to Turret Peak and beyond, the 50-km-long Moore Gulch shear zone is the main structural feature of the New River volcanic belt, separating the younger New River Mountains Felsic Complex to the east from older rocks to the west (fig. 8).

Unlike the Shylock, the Moore Gulch shear zone involves faults and offset of major rock contacts for much of its length, but the faults are younger features superimposed on earlier events of oblique shear and vertical deformation in the same zone. Because the end result is primarily one of shear offset, the Moore Gulch is termed a shear zone.

As noted in Part 2, several rock units are progressively attenuated to the north by the zone: first purple slates and phyllites of the Alder Group, then tuffs and mafic volcanics, and lastly rocks of the Little Squaw Creek Migmatite Complex (fig. 8). This oblique attenuation of rock units causes diorite and tonalite of the Cherry Springs batholith to be juxtaposed directly in fault or shear-zone contact with unmetamorphosed ignimbrites of the New River Mountains Felsic Complex to the north. Thus, fault or shear offset postdates ca. 1710-Ma formation of the ignimbrites and hence is younger than vertical deformation (high strain) in either the Shylock or Moore Gulch zones. Farther north, rocks of the Cherry Springs batholith come into contact first with unmetamorphosed felsic tuffs, then Brooklyn Peak conglomerate with boulders of Cherry Springs tonalite, and lastly red granite of the Verde River batholith. Most of the juxtaposition to the north is along a steep shear zone rather than a fault.

In southern Moore Gulch, the main fault splays into two anastomosing faults separated by horsetail faults and sheared slices of tuffaceous strata and Alder Group slates. At the south end of Moore Gulch, Alder slates persist to the limit of outcrop in fault contact with older tuffs and metasedimentary rocks to the west and younger rhyolites to the east. This section is somewhat horizontally shortened and attenuated at the faults, but is otherwise a normal stratigraphic sequence. This and many other features indicate that the faults are not strike-slip features, but have mainly vertical offset and bring rocks of lower stratigraphic levels to the west into alignment with rocks of higher stratigraphic levels to the east. Still farther south near New River town, the Moore Gulch shear zone dissipates into distributed movement within Alder slates and felsic tuffs distal to the ignimbrites. In this area, mafic tuffs west of the fault directly contact ignimbrites to the east, with only normal regional foliation in between. This sheared depositional contact shows that the Moore Gulch shear zone can be neither a major thrust zone nor a crustal suture.

Fabrics

Faults in the Moore Gulch shear zone are 10 to 50 m wide and involve clay-sericite alteration retrograded from original greenschist mineral assemblages. Detailed structural-petrofabric analysis shows that the faults were superimposed on earlier fabrics of strong foliation and steep-plunging mineral lineations. Such early fabrics constitute a 500-m-wide zone of high strain along the east side of Moore Gulch, a much wider zone than that of later shears and faults.

As the zone of high strain is approached from the west, 050°-trending foliation gradually changes to 030°-trending foliation, and variously plunging lineations steepen. This change in orientation of foliation and lineation is not later bending of older fabrics, but a primary realignment of all fabrics as strain intensifies in the high-strain zone; also, lineations are not rotated by later bending, but assume steeper orientations in the high-strain zone because of increased distortion of the strain ellipse.

The geometry of fabrics in the old high-strain part of the Moore Gulch shear zone prior to faulting and shearing involves steep west- and east-dipping foliation and steeply north- to northwest-plunging lineations. Lineations west of the high-strain zone plunge moderately west, but migmatites have shallow west-plunging lineations and steep northeast-plunging fold axes. Slates to the southwest have steep southerly to shallow westerly lineation and fold plunges, and felsic volcanics east of the shear zone have moderately south- and steeply southeast-plunging lineations and shallowly south-plunging to horizontal axes.

These data involve two main populations: (1) an early set of steeply plunging streaked and elongate-fragment lineations with 1:3:14 axial ratios and steeply plunging minor fold axes; and (2) variously plunging shallower mineral lineations and fold axes partly superimposed on the earlier vertical fabrics. This superimposition of different fold and lineation plunges is more obvious than in the Shylock zone and apparently results from a later event of oblique shear superimposed on an early event of vertical high strain in the shear zone.

Kinematics

Stereographic analysis of lineation and foliation data in the Moore Gulch shear zone by the writer in 1975 showed that no simple single or multiple deformation hypothesis could explain the huge foliation and lineation scatter and divergence between fold axes and lineations, even with features of younger faulting separated out of the data. Only a model of variable strain, where the geometry of foliation, lineation, axial planes, and fold axes all vary with the degree of strain, could account for the relationships. This model also explains realignment of regional foliation and lineation on nearing the zone, because vertical fabrics in the shear zones and those of regional deformation outside the zones are alike except for different orientations and intensity of strain.

Another key aspect of the Moore Gulch zone is its changing kinematic character along strike. In Proterozoic terranes, high-strain zones can develop into shear zones where major changes in rock competencies are encountered along strike. In fissile strata, strain is distributed over a significant width, so major dislocation or offset does not occur on any one plane. In granitic rocks, however, the same net strain is confined to a much narrower zone, so major dislocation commonly occurs. This situation well accounts for changes in the Moore Gulch shear zone along strike: to the south it is a high-strain zone where strain is distributed in metavolcanic-metasedimentary strata, but to the north where the Cherry Springs and Verde River batholiths contact one another, strain was so intensely localized in a narrow zone that it became a shear zone in which movement was facilitated by a thin plating of fissile paraschist.

History

The Moore Gulch shear zone involves (1) an early event of vertical deformation and steeply plunging lineations in a high-strain zone to the south; (2) strain being focused into a ductile shear zone to the north where two major granitoid batholiths are juxtaposed; and (3) a later event of west-side-up faulting. The Moore Gulch shear zone developed along the southeast edge of the Cherry Springs batholith largely during regional deformation, but its first horst-graben vertical movement occurred just prior to deposition of the first Alder Group strata at about 1725 Ma. It later became a high-strain zone during regional deformation just after deposition of the 1710-Ma ignimbrites, and most shearing and vertical offset to the north occurred during and after the high strain. The net result was to elevate the western block relative to the east. Finally, much later in its history, probably in Tertiary time, additional normal faulting displaced the western block down relative to the eastern block.

The final result of this complex movement history is that both lateral and vertical offset exist on the Moore Gulch shear zone. Net lateral offset is likely less than 1 km, but vertical offset may be 2 km or more, based on the alignment of lower stratigraphic levels west of the fault with higher levels to the east. Some of this alignment, however, preceded vertical strain: the zone's ancestry goes back to inception of the volcanic belt when it represented a zone of thin supracrustal deposits between two major mafic volcanic edifices of Black Canyon City and the Union Hills. Upon emplacement of the Cherry Springs batholith, rocks west of the zone were uplifted and remained elevated thereafter, forming a scarp that shed tonalite-clast conglomerate to the east and that limited both the Alder depositional basin and westward spread of ignimbrites from the New River Mountains Felsic Complex.

DEADMAN WASH FAULT

The Deadman Wash fault is a major northeast-trending fault in the Mazatzal Mountains that delimits the east side of the Verde River red granite batholith (fig. 8). In the south the Mazatzal Group is in fault contact with granite, but to the north in metavolcanic-metasedimentary rocks, the fault zone breaks into two main splays that bound a central block downdropped less than 100 m relative to the eastern Mazatzal block (fig. 8). The fault plane is vertical, and the western block is uplifted relative to the eastern, bringing Verde River Granite up level with the Mazatzal Group, whereas it would normally underlie granophyre and ignimbrites beneath the Mazatzal Group. Vertical offset on the Deadman Wash fault thus spans the top of a felsic complex.

The southernmost exposure of the fault in Deadman Wash subtly separates red granite to the west from intrusive red rhyolite to the east that grades up into granophyre and Mount Peeley ignimbrites [see Part 2]. Thus, offset may be small in the south because the full sequence of granite, granophyre, intrusive rhyolite, and ignimbrite is virtually intact. But Verde River Granite in fault contact with Mazatzal strata implies larger vertical offset and possible scissor movement on the fault to the north. However, the Mazatzal Group was deposited unconformably across a terrane that varied from intrusive rhyolite in the south to ignimbrite and older volcanosedimentary rocks in the north (the same changes as along the fault), so no rotational offset is required.

Instead, the ancestral Deadman Wash fault formed a prominent scarp by 1700 Ma that appears to have limited spread of the Mazatzal Group to the west. Anderson and Wirth (1981) documented the paleogeography of the Mazatzal Group and showed how the main channel of hematite-boulder conglomerate in the Deadman Formation extended from Sheep Basin Mountain westward under the Mazatzal Mountains to the scarp of the Deadman Wash fault, at which point it turned south. The fault or its precursor therefore predated deposition of the Mazatzal Group and may be as old as the Moore Gulch shear zone. In fact, the west-side-up vertical offset on the Deadman Wash fault parallels that of the Moore Gulch shear zone, as does the fault's orientation and geometry.

Thus, both faults may have either been coeval, or the relationship of the Deadman Wash fault to the Mazatzal Group (a bounding scarp limiting its westerly spread) is analogous to the relationship between the Moore Gulch shear zone and the Alder Group, hence the Deadman Wash fault would be 10 to 20 m.y. younger. Because the Mazatzal Group represents the uppermost stratigraphic levels during regional deformation, the Deadman Wash fault could well be an analog of the Moore Gulch shear zone at a higher crustal level. If so, then it can be assumed that distributed vertical strain at depth in a high-strain zone was expressed as finite offset on vertical faults near the surface.

GUN CREEK ZONE

Structure and Fabrics

The Gun Creek zone extends northeast along Gun Creek from Tonto Creek to Spring Creek in the Diamond Butte volcanic belt, and passes just north of Sheep Basin Mountain (fig. 8). It is a narrow zone of high to extreme vertical strain, across which significant stratigraphic dislocation occurs. To the west is a complete west-facing stratigraphic section: Mount Ord mafic volcanics (Union Hills Group) are overlain by the Reef Ridge Formation (Alder Group) and rhyolitic ignimbrites. To the east, however, the Sheep Basin Mountain quartzite sequence (Mazatzal Group) and subjacent ignimbrites are sheared into proximity with Mount Ord mafic volcanics along a narrow zone of extremely high strain.

Despite apparent vertical shear offset, present in most high-strain zones at extreme strain states, the dominant aspect of the Gun Creek zone is of high strain, where rock fabrics become steeply elongated and locally attenuated as they wrap into the high-strain zone. Incompetent Alder Group purple slates and chloritic Mount Ord volcanic strata, because of their ductility, host the axis of the high-strain zone. Structures in the Mazatzal Group near the zone are detailed below and in figure 19 because they clearly reveal the kinematics of Proterozoic vertical deformation in high-strain zones.

Nature of Strain Gradient

The Sheep Basin Mazatzal section is a classic view of the magnitude of strain variations seen where competent strata plunge into a zone of high strain from an adjacent low-strain region. A key marker unit is the Del Shay facies of the Deadman Formation, the basal conglomerate-quartzite of the Sheep Basin Mountain section. East of Coffeepot Canyon it is 150 to 200 m thick and moderately strained: conglomerate pebbles have flattened 1:3:4 aspect ratios, and moderate lineations plunge nearly parallel to fold axes that plunge 50° SW within foliation that strikes 055° and dips 70° SE. As the conglomerate extends west into the zone, it is tectonically thinned to 30 m or less as strain increases sharply: pebbles become vertically elongated with ovoid 1:3:7 axial ratios, lineations intensify and near vertical, and minor folds die out as the unit is drawn down to lower elevations into the high-strain zone. In a zone of intense foliation before the axis of highest strain is reached, the unit is severely vertically extended (1:3:15) and is finally attenuated, as pebbles transmute to vertically rodded lineations.

Coffeepot Canyon quartzites (facies of the Maverick Formation) lack good internal strain gauges, but show an even more dramatic thinning over the same strain gradient, from about 600 m in the east down to 60 m or less as they are severely horizontally shortened, vertically extended, and also drawn down to lower elevations into the zone of high strain (fig. 19).

The Sheep Basin Mountain quartzite-conglomerate facies of the Mazatzal Peak Formation, the uppermost and most competent unit in the area, also becomes strongly foliated and vertically lineated as it enters the high-strain zone in the north, and is deformed into what appears in map view to be a steeply plunging, north-closing near-isoclinal fold (fig. 19). Likewise, the southern end of the quartzite, on entering the high-strain zone from the south, is also deformed into what appears to be a mirror-image, south-closing isoclinal fold about the same axial trace as the northern fold and with a similar steep axis. This produces two hooks of intensely foliated and lineated Sheep Basin Mountain quartzite that appear in plan view to point toward each other (fig. 19). Yet outcrops of the same quartzites only 1 km to the east are nearly undeformed.

Thus in 1 km across strike, the strain gradient approaching a high-strain zone can be so great that as much as 2 km of competent rock section can be vertically extended and thinned to almost nothing. But not all rock units are attenuated: some exist in the zone as ultrathinned, grossly extended schist packages of severe internal strain, so distorted as to be scarcely recognizable. Vertical strain in the Gun Creek zone produced the classic "horseshoe fold" of Sheep Basin Mountain (fig. 19), which is not the refolded fold it appears to be from plan view, nor a normal fold. Yet in the conceptual framework of high-strain zones outlined here, the significance of the opposing hooks is elegantly simple: in three dimensions (fig. 19) they do not define folds at all, but are "tails" or "roots" that *both point the way vertically down to the axis or keel of the high-strain zone in the direction of maximum principal elongation of the strain ellipse—simply, the direction of greatest strain—vertical.*

Simplistically, therefore, the Sheep Basin Mountain Mazatzal section can be viewed as perched on the edge of a structural chasm: the bulk of the section stayed on the edge during regional deformation, but the west end of the section that was situated over where the chasm formed was dragged down into it via two vertically extended prongs (the mechanism of deformation was not "dragging," but extreme vertical extension consequent upon strong horizontal shortening across the zone). These two prongs of the Sheep Basin Mountain "horseshoe fold" and the way they define in three dimensions the axis of strain inflection (fig. 19), illustrates the asymmetric manner in which the quartzite mass was deformed in a single event of highly variable strain and shows the tremendous strain gradients that are integral to this distinctive Proterozoic vertical deformational regime.

STRAIN VARIATIONS THROUGHOUT THE CENTRAL VOLCANIC BELTS

As noted previously, a combination of foliation and lineation intensity and to some extent lineation and foliation steepness indicate strain intensity in the volcanic belts. Consequently, variations in foliation and lineation can be used to map strain variations throughout the volcanic belts.

PRESCOTT-JEROME VOLCANIC BELT

In the Prescott-Jerome belt, strain intensity is not proportional to age of the rocks: the oldest Senator Formation is one of the least deformed in Arizona, and most of the Bradshaw Mountains Group is weakly foliated and largely unlineated, except near the Chaparral zone. The Senator Formation resisted strain because of its massive, competent primitive mafic flow sequences and was also shielded from deformation by the Government Canyon, Wilhoit, and Lynx Creek plutonic bodies. Furthermore, it lay near the edge of the volcanic belt, where it was not subject to as high a strain as central parts of the belt.

Similarly, most volcanic strata in the Jerome volcanic belt show very little strain, especially south of Mingus Mountain where the wide northern end of the Cherry Springs batholith shielded them from deformation (fig. 3). Also, younger agglomerates in the south were some of the stratigraphically highest units in the Jerome belt and were never buried to significant depths to be strongly deformed. Strata in the northern part of the Jerome belt near the United Verde mine are the most strongly strained in the belt (Lindberg, 1986).

Strain in the Prescott belt generally intensifies and lineations steepen toward central and southeast parts of the belt. Spud Mountain strata are more highly strained than Bradshaw Mountains Group strata, and lower Mayer Group mafic volcanics, despite their massive character, are more highly strained than most Spud Mountain strata west of them. The Cleator Formation and Black Canyon Creek Group felsic volcanics tend to be the most strongly strained of all in the Prescott belt, especially in the Shylock-Black Canyon zones, which were part of regional deformation and a more intense manifestation of the same strain regime that affected the entire belt. The general intensification of strain toward the Shylock zone is a key feature of vertical deformation of the Prescott belt.

Lineations plunge more shallowly in the northern Prescott belt than in the south. Far to the north in Chino Valley they are subhorizontal and plunge shallowly north or south ($< 30°$) in northern Texas Gulch exposures. Lineations and fold axes in Texas Gulch strata at the nose of the Brady Butte pluton plunge shallowly but steepen markedly on its flanks. Moderate to shallow lineations and widely scattered fold axes in the Jerome and northern Prescott belts give way southward to steeper lineations and steeper, more consistent fold plunges. Lineations reach vertical in the Mayer and Black Canyon Creek Groups and plunge steeply south around the Crazy Basin body (DeWitt, 1976). Because this body was emplaced at the peak of regional metamorphism but late in regional deformation, recrystallized mineral fabrics overprint early steep lineations to the south.

The southerly increase in lineation uniformity and steepness correlates to increasing crustal depth to the southwest, where the deep plutonic and high-grade metamorphic roots of the Prescott belt are exposed [see Part 3]. In contrast, foliation and lineation intensity, which more closely reflect strain intensity, increase toward the Shylock and Black Canyon zones. Thus, although rocks in the southwest part of the belt have recrystallized foliated and gneissic fabrics, they are not as strongly lineated as rocks deformed at higher structural levels in the Shylock zone. The strain difference is reflected in very different rock fabrics: the higher grade gneisses have predominantly steeply dipping, planar fabrics, whereas rocks in high-strain zones have predominantly steeply plunging linear fabrics. Thus, strong vertical deformational in the Prescott belt is not related to crustal depth but to lateral strain variations within the belt.

Such strain variations were largely dictated by the array of plutons and batholiths that intruded the Prescott belt prior to deformation and that formed competent borders and buttresses against which the incompetent volcanic strata were ductilely deformed. Strata at the margins of the belt, especially where draped over batholith and pluton edges, were much less deformed than in central parts of the belt. The southern Jerome belt, northwest and west parts of the Prescott belt, and the Copper Mountain and Cordes areas were all partly shielded from high strain by batholith edges or interdigitated plutons. Such regions were subject to variable strain, depending on attitudes of subjacent or adjacent plutonic borders, and folds and lineations followed "topography" of these borders in three dimensions. Strata in the central Prescott belt had no such supports or constraints and deformed freely as their own competencies allowed. Strain was therefore maximized in the east-central part of the belt where rocks were most ductile. This central axis became the Shylock zone.

However, the positions of batholiths severely constrained the Shylock zone's orientation and its deformational geometry. The northern segment was tightly sandwiched between the Cherry Springs batholith and Prescott and Wilhoit Granodiorites, so the thin septum of strata in the zone was forced into strong vertical extension during crustal shortening. The central segment near Mayer was less constricted by plutonic rocks, so a wider zone of highly strained tuffaceous strata was able to retain essential stratigraphic coherence during deformation. Although strain was spread over a broad zone in the Black Canyon segment to the south, vertical extension was equally strong, and pelitic rocks were drawn down to deeper crustal levels and subject to partial melting.

The Shylock zone thus formed a central structural trough or keel to the Prescott belt (fig. 18), into which all linear structures ultimately converged toward vertical. The final north-south orientation of the high-strain zone was totally constrained by the positions of adjacent pretectonic and later plutonic bodies, and the Shylock zone probably acted on a very broad scale as a zone of major adjustment that allowed the batholiths to rise diapirically or jostle for relative position to accommodate the greatest amount of horizontal shortening.

YOUNGER EASTERN VOLCANIC BELTS

Strain variations in the younger eastern volcanic belts partly correlate to stratigraphic position, in that the

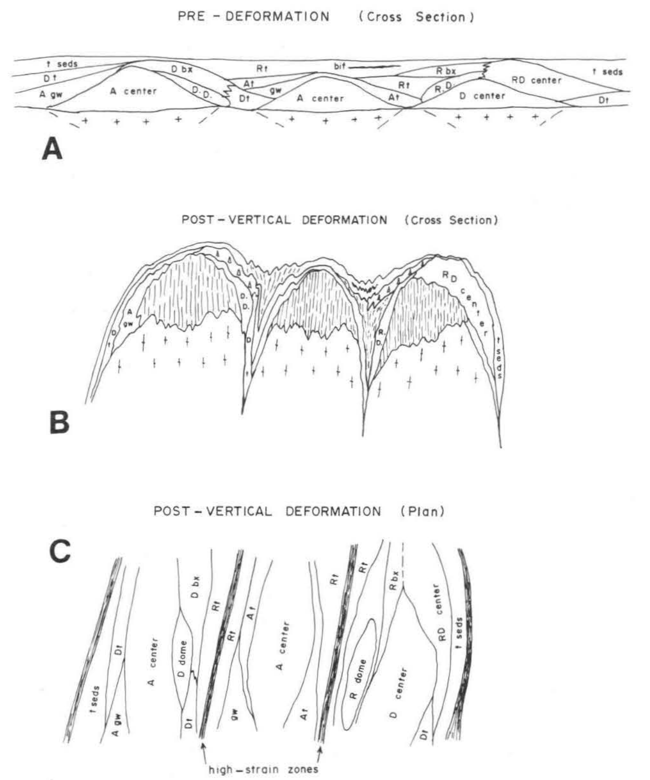


Figure 18. Diagrammatic cross-section of the central volcanic belts showing how the volcanic pile responded to Proterozoic vertical deformation. Part A shows the original volcanic (predeformational) structure, part B is a cross section after deformation, and part C is a plan of the postdeformational cross section. Part A shows the compositions and primary structure of the volcanic centers and their intervening tuff deposits. During deformation, the domal volcanic centers deformed internally (vertical foliation lines) but generally did not fold, whereas incompetent tuffs underwent strong vertical deformation and internal folding (B), and became the loci of high-strain zones (C). This diagram shows how original competency contrasts in the volcanic pile governed the final array of deformational features, including high-strain zones in areas of originally thin, weak crust, as well as relatively undeformed volcanic centers (shown) and batholiths (not shown) in areas of thicker crust. **Rock compositions:** A = andesite, D = dacite, RD = rhyodacite, R = rhyolite; **Rock type:** bif = banded iron formation, bx = breccia, gw = graywacke, t seds = sedimentary rocks, t = tuff, D.D. = dacite dome, R.D. = rhyolite dome.

youngest and stratigraphically highest Mazatzal Group in general was less deformed than underlying ignimbrites, which were less strained than the Alder Group, which was less strained that the Union Hills Group. Nonetheless, great strain variations throughout the belts render the relation between strain and stratigraphic position invalid in many places. For example, the East Verde River and North Union Hills Formations are less strained than the Alder Group, and the Sheep Basin Mountain Mazatzal section is more highly strained than most ignimbrites of the felsic complexes. Therefore, stratigraphic position is not the main factor governing strain variations.

The broad structure of the younger eastern volcanic belts is complex because they were disconnected by plutonism: both the huge Verde River Granite batholith and its felsic complexes, and Payson Granite and its felsic complex, comprise vast regions of structural competency between older volcanic and clastic strata. The felsic masses predated deformation, so the present disposition of belts is much as it was during deformation; hence the New River-Cave Creek and Mazatzal Mountains-Diamond Butte belts deformed as structurally independent units.

In the New River-Cave Creek belt, the North Union Hills Formation and correlatives on Humboldt Mountain were shielded from high strain by granitic masses at the edges of the volcanic belts. Competent quartzites in the Alder Group near Carefree (fig. 8) were folded but not subject to high strain. The central Cave Creek area contains only the oldest strata in the belt and is the most strongly deformed portion, with strong foliation and steeply northeast-plunging mineral lineations related to a large fold in Cramm Mountain and Grays Gulch strata. The New River Felsic Complex to the west was a structurally competent mass resistant to high strain, and only the upper bedded ignimbrite-breccia portions near the Moore Gulch shear zone are affected by moderate to strong foliation and moderately to steeply south-plunging lineations.

In the Mazatzal Mountains, the East Verde River Formation was shielded by the northeast end of the Verde River batholith (fig. 8), and Mount Ord mafic volcanics and Tonto Basin ignimbrites were shielded by granites, so they are all weakly strained. East of the Deadman Wash fault, strain increases and moderate to steep lineations occur in the older strata. Union Hills and Alder Group strata between Slate Creek Divide and Diamond Butte occupy a central zone of strongest deformation in the eastern volcanic belt (fig. 8), with moderate to steep lineations and variably plunging fold axes. Because the younger belts are moderately strained in most places, fold axes and lineation plunges are highly variable, and shallow fold axes commonly occur with steeper lineations.

Many northern exposures of Mazatzal quartzite in the younger volcanic belts represent shallow synclinal keels (P. Anderson and Wirth, 1981) related to gently plunging open folds in subjacent rhyolites (Gastil, 1958; Conway, 1976). Foliation and lineation intensifies, and fold and lineation plunges steepen, as strain intensifies to the south toward the Gun Creek shear zone, which is the central axis of highest strain in the younger eastern volcanic belts (fig. 19).

MAZATZAL GROUP DEFORMATION

Shallow-plunging folds typify uppermost strata of the Mazatzal Group (Mazatzal Peak Formation), but steeper lineations and folds occur throughout highly strained basal strata of the group (Deadman Formation). A key structural feature of Mazatzal strata in the Mazatzal Mountains is imbrication of upper quartzite units on south-dipping thrust faults (Wilson, 1939). The east side of the Mazatzal Mountains shows several such faults where lower red quartzites of the Mazatzal Peak Formation are thrust over the uppermost white quartzites. This imbrication apparently thickens the section by 30 percent in places. Ten or more such thrust planes can be identified: most dip 10° to 20° S, but steepen as they cut up section into quartzites and flatten down section into the shale-dominated Maverick Formation.

Detailed mapping (P. Anderson and Wirth, 1981) shows that virtually all thrust planes are rooted in the fissile shales of the Maverick Formation. The radically different strain geometries in the quartzite-dominated Mazatzal Peak and Deadman Formations require that the incompetent shales of the Maverick Formation be a zone of strain compensation and a décollement for the thrust planes. The Deadman Formation is deformed about steeply southwest- and northeast-plunging folds, and is well foliated and steeply lineated. Axial ratios of conglomerate pebbles are distorted by as much as 1:4:7, and the unit is stacked upon itself where fold axes were detached. Thus the Deadman Formation experienced moderate to strong strain and has a steep to vertical fabric in much of the Mazatzal Mountains. In contrast, Mazatzal Peak Formation quartzites are deformed by upright, open flexural and similar folds (Wilson, 1939) with shallow axes, and foliation and lineation are weak to absent in most places. The Mazatzal Peak Formation clearly experienced much less internal strain than the Deadman Formation, and the Maverick Formation, which is highly folded and contorted into sliced, imbricate rock packages, spans that gradient in strain. Thus, the thrusts are rooted in a series of dislocation planes in Maverick Formation that diverge as they cut up through the Mazatzal Peak Formation.

Deadman Formation on Cactus Ridge at the south end of Mazatzal exposures in the Mazatzal Mountains (fig. 8) includes very thick basal conglomerate units derived from a local rhyolite paleo-hill slope (P. Anderson and Wirth, 1981). The basal conglomerate wedges taper northward and lens out, but the main 100- to 200-m-thick upper Deadman Formation quartzite remains relatively constant throughout most of the Mazatzal Mountains (P. Anderson and Wirth, 1981). The Cactus Ridge Deadman section occupies the core of a gently 10° to 20° north-plunging syncline through Mazatzal Peak that opposes a gently 5° south-plunging open syncline in upper Mazatzal strata on North Peak, defining a gently doubly plunging synclinal structure.

The different thicknesses between south and central Deadman exposures, opposing dips of the synclines, and Maverick Formation exposures at high elevations in the Mazatzal Mountains have all been mustered as evidence for large-scale displacements on thrusts in upper Mazatzal strata (e.g., Puls and Karlstrom's (1986) "roof thrust" supposedly moved 15 km from a hypothetical starting point to the southeast). However, the Deadman Formation quartzite has been mapped continuously without major offset around the entire perimeter of the Mazatzal Group in the Mazatzal Mountains (P. Anderson and Wirth, 1981), which disproves such large-scale displacements on any thrusts cutting the Deadman Formation. Instead, the magnitude of offset, as superbly exposed in cliff faces on the east side of the central Mazatzal peaks, is small (typically 200 to 600 m) and just sufficient to imbricate upper units of the Mazatzal Peak Formation, and involve some Maverick Formation, but not the basal Deadman Formation. Thus it is evident that thrusting was internal to the Mazatzal Group and did not extend into subjacent rock packages to constitute a "foreland fold and thrust belt" of the type or scale imagined by Karlstrom (1986).

The reason such thrusts dissect the upper Mazatzal Group strata but not subjacent units is that strain could be distributed on many closely spaced foliation planes in subjacent, ductile Alder and volcanic strata. The thin Deadman Formation quartzite was able to accommodate most of the strain in subjacent strata by deforming into contorted, stacked folds or by breaking into pieces (P. Anderson and Wirth, 1981; P. Anderson, unpub. mapping). The overlying thick Mazatzal Peak quartzite section, however, was much too competent and massive to take up foliation that could distribute strain. It accommodated some primary northwest-southeast shortening first by flexing into broad northeast-trending folds, and attempted to compensate for secondary northeast-southwest shortening by bending of the fold axes into a synclinal keel. Any further shortening could only be accommodated by brittle deformation, so the massive quartzites became sliced by a series of low-angle imbricate faults. The Maverick Formation deformed so as to effectively compensate for the disharmonic structural responses of Deadman and Mazatzal Peak quartzites, and did so via complex folds, faults, and staked structural dislocations in which the soles of the thrust planes were rooted.

The net result was a folded Mazatzal quartzite sequence that "sagged" into a central keel as subjacent strata underwent greater horizontal shortening and vertical extension at depth. The gently plunging axes of the major folds point toward the keel of the synclinal depression and the locus of greatest vertical extension in subjacent strata. Superimposed folding (Karlstrom, 1986; Karlstrom and others, 1987) is neither required nor indicated, because folding, imbricate thrusting, and central sagging of the major fold axes all occurred in a single event of Proterozoic vertical deformation; various types of structures and attitudes simply reflect different structural responses to the deformation.

Deformation of the Mazatzal Group is thus an upper crustal expression of more profound horizontal shortening and vertical extension at deeper levels in the volcanic pile. Competent Mazatzal quartzites could comply only by first folding then breaking into imbricated fault blocks. The magnitude of thrust shortening is not huge (1 to 2 km or so), just sufficient to take up constrictive strain in the horizontal plane as subjacent strata were vertically extended.

Gently doubly plunging synclinal keels such as the Mazatzal Group in the Mazatzal Mountains, where strata are drawn toward the bottom of the keel by strong vertical strain, are common elsewhere in the Arizona Proterozoic and include most isolated exposures of Mazatzal quartzite in the eastern volcanic belts (e.g., McDonald, Christopher, and Sheep Basin Mountains, and Four Peaks). This indicates that during uplift and erosion, only the synclinal keels of the uppermost supracrustal deposits were preserved.

CONCEPTUAL APPROACHES TO PROTEROZOIC DEFORMATION

The following analysis continues on from the description of structural fabrics and features earlier in this paper to develop a model that most accurately reflects Proterozoic vertical deformation in central Arizona.

CONFLICTING FOLD MODELS

From exposures in the northern Prescott belt with shallowly plunging minor folds and steep lineations nearby, Anderson and Creasey (1958) deduced a deformational model of upright flexural folds, with lineations parallel to the "a" slip direction and orthogonal to subhorizontal fold axes. From exposures in the southern Prescott belt with steeply plunging minor folds and lineations, DeWitt (1976) deduced a deformational model of steeply plunging similar folds and "b" mineral lineations parallel to fold axes. These two models are almost exact opposites and are in radical conflict (Creasey, 1980; DeWitt, 1980). Upright flexural folds with subhorizontal axes occur in upper quartzites of the Mazatzal Mountains, but lineations and folds plunge steeply in nearby Deadman Formation quartzite. Where mineral lineations in the Jerome volcanic belt are steep, fold plunges vary from 10 to 60 degrees as fold axes snake down plunge. High-strain zones like the Shylock have horizontal and vertical fold plunges where lineations are vertical. Even though the *lineations have the same structural significance in all cases,* they have been assigned genetically different labels, with the implication of different origins, because of different fold orientations.

Such features as variations in fold orientations and plunges (O'Hara, 1980, 1986), ovoid shapes of many map units (Lindberg, 1986), and speculations that fabrics in one zone may predate those elsewhere (Karlstrom, 1986; Karlstrom and others, 1987) have been the bases for postulating regional polyphase folding throughout the central volcanic belts. Polydeformational studies show, however, that much more rigorous criteria, such as abundant earlier linear and planar features and their systematic deformation stereographically on great or small circles, are diagnostic of polydeformed regions (Ramsay, 1967). Refolding of a deformed terrane further compresses many early folds, so interference structures are invariably obvious in most outcrops throughout the terrane. Such rigorous evidence pervades all polycyclic Precambrian terranes elsewhere in the world, including refolded gneisses of northwest Arizona, so its absence in central Arizona is the strongest evidence against regional polyphase deformation.

The central volcanic belt is replete with areas where fold and lineation plunges are disparate and highly variable, but polyphase folding was not responsible for the variability. The genetic significance of lineation relative to folds can only be deduced after the total strain picture is known.

INDICATORS OF STRAIN

As noted earlier, abundant distorted primary features in volcanic rocks provide gauges that measure the degree of strain at any point; hence strain variations can be quantitatively mapped throughout the belts fully independently of foliation, lineation, or folds. Using strain state and its variations as benchmarks, it has been determined how foliation, lineation, folds, and their variations correlate to the finite (i.e., resultant) strain state of the rocks. This strain analysis reached the following conclusions: (1) the intensity of foliation development (i.e., the number of foliation planes per cm) is directly proportional to the intensity of strain; (2) the intensity and steepness in plunge of lineations (not either one by itself) together are a direct measure of the strain state; (3) the angular relation of lineation to foliation gives the orientation of the strain ellipse, but not its magnitude or axial ratios; (4) only axial ratios of distorted features precisely measure the relative magnitudes of the strain axes; and (5) no other structures or orientations are of great importance in determining the strain state of the rocks.

The consequences of these conclusions are far reaching to an analysis of Proterozoic vertical deformation. First, they mean that, in the absence of distorted features, the degree of foliation and lineation development in the field closely approximates strain intensity: in general, areas of more intense foliation and lineation are more highly strained, areas of weaker foliation and lineation are less strained, and foliated areas devoid of lineation are least strained. Where such changes occur over very short distances, it is possible to see in small areas how foliation and lineation change with varying strain. Second, the finite strain geometry at any locality can be precisely determined by simply observing three features: foliation, lineation, and distorted primary features. At higher metamorphic grades or greater crustal depths, the above relationships must be adjusted for recrystallization effects, as noted later.

The negative consequences of these conclusions are equally important. First, foliation and lineation trends are

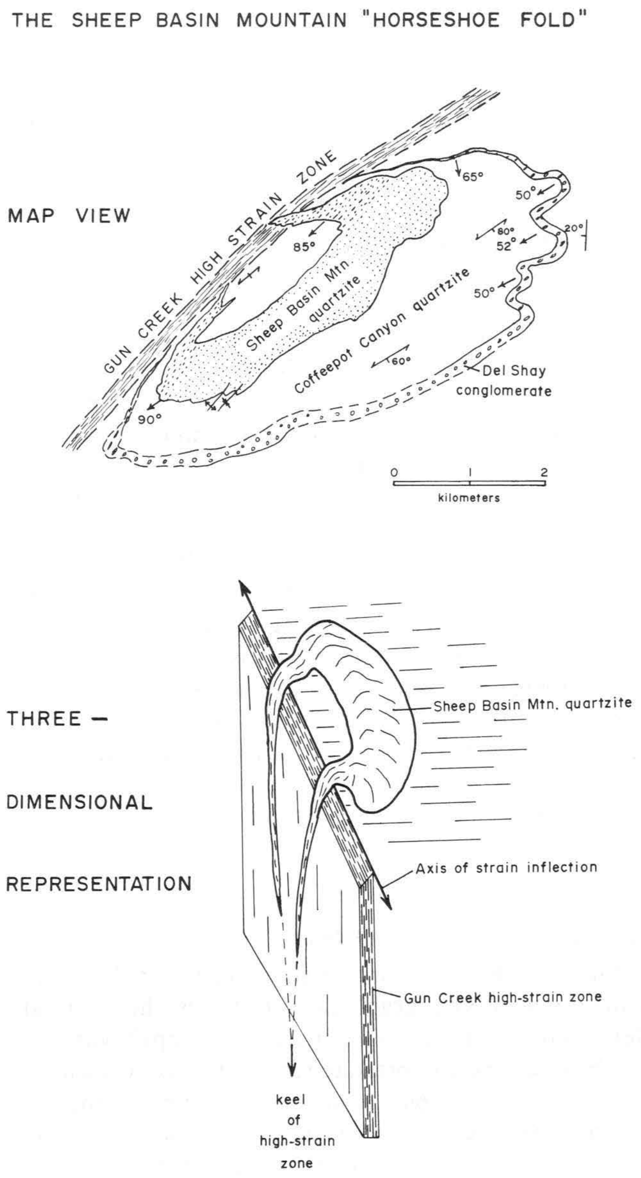


Figure 19. Structure of the Sheep Basin Mountain "horseshoe fold" in plan and three-dimensional cross section. The plan is reduced from the writer's unpublished 1:24,000-scale quadrangle mapping, and shows all components of the Mazatzal Group in the Sheep Basin Mountain area: the Del Shay facies of the Deadman Formation, the Coffeepot Canyon facies of the Maverick Formation, and the Sheep Basin Mountain facies of the Mazatzal Peak Formation [see Part 2]. For simplicity, the cross section shows only the Sheep Basin Mountain quartzite and how its ends that overlapped the Gun Creek zone underwent extreme vertical extension toward the axis or keel of the high-strain zone. The horseshoe structure might be classically interpreted as a refolded fold, but only one set of deformed fabrics (steep foliation and steeply plunging lineations) is evident in the field. This figure shows that, in three dimensions, the structure is not strictly a fold but a product of highly variable strain at the edge of a high-strain zone. The axis of strain inflection signifies where the degree of horizontal flattening and vertical extension become extreme.

of no consequence to the strain state, and both the steepness of foliation dip and lineation plunge are only partly dependent upon strain state. As a result, a statistical structural approach of amassing and stereographically analyzing numerous orientation readings does not accurately assess the total strain picture. Second, the types of minor folds, their orientations, and asymmetries provide little information about the total strain picture and do not demonstrate the existence of regional fold closures.

SIGNIFICANCE OF FOLDS

Compared to ubiquitous evidence for deformation in the volcanic belts (foliation, lineation, distortion), evidence as to how the rocks were folded is much less obvious. The style of folding cannot necessarily be inferred from the character of other deformed features, and more importantly, *the manner in which the belts were deformed cannot be deduced from the geometry of folds.* The approach of extrapolating fold patterns in a few small areas to infer a strain picture for the entire belt has lead to conflicting or irreconcilably different structural interpretations in the past, and has caused much misunderstanding of Proterozoic deformation in central Arizona.

In almost all areas where fold and lineation attitudes differ, folds are more widely scattered and shallower than lineations. In mildly strained parts of the Jerome and Cave Creek belts, for example, axes of crudely formed minor folds diverge from lineations by up to 40 degrees in a single outcrop, but the axes follow sinuous paths down plunge to cause scatter and are clearly not refolded. Such poorly developed folds typify low to moderate strain states, in contrast to areas of higher strain, where fold axes more uniformly parallel steep mineral lineations. Because lineations accurately reflect the strain state, it must be concluded that fold plunges do not always accurately reflect strain intensity.

Foliation was not produced by regional folding of the volcanic belts, and is therefore not axial planar to regional folds. Instead, regional strain produced foliation approximately parallel to the major northeast-trending basins that made up the volcanic belt. In detail, foliation was superimposed on the rocks largely independently of bedding, so the interplay of bedding and foliation changes across and along strike as foliation transects the interlensed volcanic pile (figs. 2 and 18). Where bedding diverged significantly from foliation strike, such as at facies changes, abundant minor folds formed, but where it paralleled foliation, fewer folds could form. Bedding maintains small acute angles to foliation for kilometers along strike without major closures being involved or approached; consistent minor fold asymmetries result from a regional angular relation of bedding to foliation, not from position on the limb of a major fold.

The fold style was dominantly passive slip and flow, where flattening and minute translation on foliation planes summed up to form folds in outcrop. Had foliation been imposed perpendicular to original bedding trends, much slip on each foliation plane would be needed to form tight to isoclinal minor folds, but because foliation was imposed at only small angles to the regional trend of bedding, little slip was needed to produce minor folds. In detail, competencies of the rocks governed the style of folding. Fissile volcanic rocks deformed passively at all strain states, while thin competent chert beds folded at low strain states, boudinaged at higher strain states, and were dismembered into tectonic pods at the highest strain states, as internal structural coherency in the volcanic pile was lost.

MODEL OF PROTEROZOIC VERTICAL DEFORMATION

Analysis of the original structural and stratigraphic makeup of the volcanic belts prior to deformation reveals many very fundamental aspects of Proterozoic vertical deformation, because original heterogeneities in the volcanic pile must have influenced the way it would subsequently deform.

DEFORMATION OF THE VOLCANIC PILE

In volcanic piles, the greatest competency contrasts lie not between horizontally bedded layers, but between domal volcanic centers of massive flows and breccias, and the thinly bedded fissile tuffaceous strata that flank them. In some cases, such as near the end of major volcanic cycles, these tuffaceous strata comprise subhorizontal rock packages, but more commonly they fill basins of thin crust between or flanking thick competent volcanic centers, especially in formative stages of a volcanic belt's evolution. Thus, wedge-shaped packages of the structurally least competent strata lie between major volcanic centers, so vertical weakness zones existed in the volcanic belts from their inception, zones that became prone to high strain during Proterozoic vertical deformation.

Figure 18 shows what happens when a laterally interstratified volcanic pile typical of Proterozoic volcanic belts is deformed. Dome-shaped volcanic centers of massive flows and breccias are separated by thinly bedded tuffs, so they behave structurally as competent pods in a fissile matrix. No matter how great the strain, competent domes cannot be folded or bent into arcuate shapes around adjacent ductile strata, because strain is always preferentially taken up by flow of fissile enveloping tuffs. The ends of competent units are commonly sheared and folded where they interface with enveloping tuffs, but massive core regions are not. This leads to a style of deformation where originally discrete volcanic centers, "lubricated" by fissile matrix strata, remain structurally discrete during deformation and deform independently of adjacent centers.

This does not mean that competent rock masses remained undeformed; they accommodated much strain internally, without being folded. However, the strain states of competent and incompetent rock masses were radically different, and strain gradients between them depended on competency contrasts and distribution of strain in the volcanic belt. Consequently, the volcanic pile deformed into a collage of ovoid or elongate rock units that strained internally but behaved structurally independent of adjacent units (fig. 18). This deformational style produced abundant evidence for strain internal to major units (e.g., minor folds) but little evidence external to or between units (e.g., major regional folds), which is exactly what one sees in the field.

The reason why major closures or regional folds did not develop in the volcanic belts is that the ductile units preferentially absorbed strain before competent lensoidal rock masses could be folded, and they continued to absorb increases in strain. Thus, the ductile strata became the zones where movement between neighboring competent masses was concentrated. These zones, in fact, are the high-strain zones ubiquitous throughout the Arizona Proterozoic, and as figure 18 shows, they always favor zones of tuffaceous strata or original weaknesses along the edges of the most competent masses—major batholiths. As inferred from their strongly vertically elongated fabrics, the high-strain zones sustained the greatest vertical extension, and as a result, supracrustal rocks were drawn down to unusually great crustal depths (fig. 18). High-strain zones can also develop into shear zones that vertically offset adjacent parts of the volcanic belts.

DEFORMATIONAL MODEL

Figure 20 shows the range of strain regimes in the central volcanic belt and graphically outlines the vertical-deformation model. The increasing depth variable combines inseparable pressure-temperature components of increasing metamorphic grade and confining (lithostatic) pressure, whereas the horizontal axis shows spatial position only and is *not* to be interpreted as a time progression; it shows finite (final) strain states at various positions in the volcanic belt, and does not imply that strain progressed from one regime to the next, nor that regimes to the right evolved through the stages to the left.

Figure 20 could represent, for example, a broad east-west transect of the Prescott belt, from least strained rocks of the Senator Formation near Prescott to the most highly strained rocks in the Shylock zone. In detail, it could depict increasing deformation of volcanic stratigraphy as a high-strain zone is approached. Spatial variations in strain can be observed in the field at all scales, but it cannot be proved that rocks now at high strain states ever went through the stages of lower strain states, nor do observational features suggest that they did, as discussed in more detail in the kinematics section to follow. The Proterozoic vertical deformational model is outlined in ten points below, in reference to figure 20.

(1) The stratigraphy of each volcanic belt was fully evolved and had been entirely deposited prior to deformation.

Time lines in the stratigraphy were horizontal at inception of deformation, but lithologic packages of greatly differing competencies between the time lines dipped by as much as 15 degrees because of complex stratigraphic interlensing, doming of strata around early pretectonic plutons, and local uplift to produce intervolcanic unconformities. Also, zones of vertical stratigraphic weakness and thinner crust existed between the major volcanic centers, zones that would ultimately become the loci of high strain.

(2) Although strata were deformed into vertical orientations throughout many parts of the volcanic belts, not only did major rock sequences keep their proper context, but stratigraphy on a broad statewide scale remained basically subhorizontal after deformation. For example, even though Mazatzal strata are deformed locally into vertical orientations in Arizona, New Mexico, and Colorado, they still exist at comparable elevations in all states, and can be viewed as composing a broadly undulating surface. Deformation therefore did not destroy the primary stratigraphic makeup of the volcanic belts, as is the case in major overthrust belts dominated by horizontal tectonic transport. Instead, original stratigraphic relations remained basically as conceived, even though steep tectonic fabrics were imposed and the rocks were vertically deformed (fig. 18).

(3) Thus, on a regional scale, stratigraphic units still follow their original northeast-trending depositional basins, the disposition of these basins remains basically as it was, and relationships between major rock packages were preserved during deformation. These extremely important concepts (also endorsed by Lindberg, 1986) indicates that: (a) stratigraphy was not turned on end prior to deformation; (b) thick stratigraphic sections are not "homoclinal sequences" that all face in a single direction such as east or west; and (c) stratigraphy did not originally trend northwest prior to deformation. In essence, the *original up* direction in the volcanic belts is *still up* in most places.

(4) The general lack of strong competency contrasts parallel to bedding had the following consequences: (a) strain varied laterally across the belts more than between stratigraphic units; (b) hence strain became concentrated in preexisting vertical weakness zones; and (c) foliation and lineation were superimposed across the volcanic belt with little regard for stratigraphy. Thus, vertical foliation could penetrate the rocks without deflection, and lineations could initially assume steep attitudes because of their independence from bedding. Folds, however, were constrained by bedding attitudes and formed in greatest abundance where bedding diverged from regional foliation.

(5) Consequently, at low-strain states (fig. 20) weak foliation was the first structural feature to be superimposed on the rocks, before most folds and lineations formed. Foliation development and steepening of bedding were associated with prolate or oblate vertical distortion, pressure solution, and interplanar differential movement. Also, because foliation intensity was proportional to vertical pure shear distortion and to bedding steepness, foliation development could distort bedding to vertical without rotation or macroscopic offset (Ramsay, 1967). Thus at mild strain states bedding had moderate dips, at moderate strain states it had steeper dips, and at high to extreme strain states it reached vertical dips (fig. 20 graphically shows such steepening of bedding without implying macroscopic offset). Therefore, moderate and strong strain could orient bedding parallel to foliation in plan view and in cross section respectively (figs. 18 and 20).

(6) Lineations became prominent only where foliation was well developed and bedding was steepened, at and above moderate strain states. At moderate strain states, rock contacts had been sufficiently distorted from subhorizontal to develop folds and shallow- to moderate-plunging lineations (fig. 20). At this point, strain attained a maximum heterogeneity, forming folds with various axial orientations, plunges, curved axes, and axial surfaces, and also forming variably oriented lineations in areas where strain variability was greatest. Competent beds such as iron formation broke up at this strain state and deformed chaotically into unsystematic shapes.

(7) Low to moderate strain at deeper crustal levels produced similar structures but with important differences (fig. 20). Strain was heterogeneously distributed in the more competent units, where inhomogeneous distortion, curved axial traces, and warped lineations are common. In the less competent strata, moderate strain produced schistosity under conditions of flattened constrictive strain; flattened discoid shapes were common, and planar fabrics, such as foliation, dominated over weak lineations. This is the environment in which most strongly planar rocks (schists and gneisses) formed. In contrast, *all higher degrees of vertical strain favored linear rather than planar fabrics, and are characteristic regimes of high strain.*

(8) At high strain states, strong foliation and lineation began to eradicate original rock textures and structures. A major strain threshold was surpassed, vertical elongation began to take over from flattening, and the axial ratios of distorted fragments extended vertically to form long prolate shapes. Steeply plunging lineations increased in intensity and uniformity of plunge, and as bedding reached vertical, fold plunges more uniformly paralleled vertical lineations. Marker beds such as iron formation were deformed into systematic, though highly attenuated, recognizable fold patterns, and vertical passive slip folds became a dominant feature of thinly bedded fissile units.

(9) At extreme strain states, vertical extension dominated, and all rock textures were overprinted by strong vertical lineations. Bedding and folds both became attenuated, and axial portions of folds were converted into vertical rods (fig. 20). At this intensity of strain, stratigraphic relations lost coherency and marker beds such as iron formation became completely disconnected. Ultrastrain states are dominated by extreme foliation and extreme vertical lineation that eradicated virtually all traces of original textures and other

structures. Such rocks became ultralineated rodded or papery schists in which the ratio of horizontal flattening to vertical extension was typically 1:30 (fig. 20).

(10) At high and extreme strain states but at greater crustal depths, mineral recrystallization and plastic strain (Ramsay, 1967) predominated. Gneissosity and vertical mineral lineations prevailed, but increased vertical strain vertically attenuated gneissosity and caused rodded mineral lineations to overwhelm all other features of the rocks. The upper limit of vertical strain regimes formed by pure shear (fig. 20) was surpassed in parts of central Arizona where simple shear dominated high-strain zones, or where plutons rose diapirically during deformation to vertically extend all fabrics even further by ductile extension. This is the "extended strain" regime of figure 20, where fabrics became physically extended past limits of normal deformational regimes.

KINEMATICS AND PROCESSES OF DEFORMATION

Previous fold models viewed deformation in central Arizona as a result of horizontal crustal shortening mainly in one direction (the maximum principal stress far exceeds the other principal stresses). This stress regime may apply to upper crustal fold-and-thrust belts dominated by horizontal tectonic transport (e.g., Karlstrom, 1986), but is quite inaccurate for deformation of Proterozoic volcanic belts, which were crustally shortened in all horizontal directions (compressive stresses along both principal horizontal stress axes were similar in magnitude). Such a stress regime is constrictive in all horizontal directions, so material is forced to move vertically, being resisted at depth only by lithostatic load. However, deformation of volcanic belts was far from static, because the denser volcanic rocks sank to compensate for the diapiric rise of less dense plutonic masses flanking the belts. *Consequently, the volcanic belts were dominated by a vertical tectonic regime where material moved primarily downward.* Such regimes are experimentally well defined (Ramberg, 1981) and are commonly used to model Archean greenstone belts (Schwerdtner and others, 1979).

The rise of flanking plutonic bodies and sinking of rocks in the volcanic belts during deformation provides for an almost limitless amount of net vertical extension, as the entire volcanic belt is subject to shear that, in an ideal situation, draws it into a central axis or trough (fig. 18). In reality, the central trough may be displaced to one side of the belt (e.g., Shylock zone), depending on subsurface topography of competent plutonic masses and location of the main original weakness zone in the volcanic belt. The central axis becomes the main high-strain zone of the belt: in vertical section this axis is trough shaped (fig. 18), but in three dimensions, the trough commonly converges to a central keel or series of keels whose axial regions underwent the maximum vertical elongation in the belts, toward which all linear structures ultimately converge at depth. This explains why so many synclinal keels of supracrustal sediments (e.g., Mazatzal strata) exist in the volcanic belts without anticlinal complements: downward movement in the trough promotes formation of synclinal keels, but causes structural dislocation in place of anticlinal culminations, which are unstable in the trough and occur mainly near the plutonic flanks.

Pure Shear

In a subsiding trough, although edges of the volcanic belt experienced strong vertical simple shear from diapiric rise of plutonic masses, deformation in central parts of the belt can be closely modeled by a pure shear mechanism. This means that as the incompetent volcanic rocks underwent shortening in both horizontal dimensions, they extended purely vertically without rotation. This model predicts that zones of high strain in the volcanic belts should show no net offset across them, only strong vertical elongation of fabrics within them. This is exactly what the Shylock zone shows: on a broad scale, the Grapevine Gulch contact is not offset across the zone, although rocks are drawn down into it (figs. 3 and 18), and on a more detailed scale, stratigraphic units pass right across the zone through strongly vertical structures without significant offset. Similar features occur at all scales in the volcanic belts to suggest that the mechanism of deformation in high-strain zones was primarily pure shear.

Fabrics and structures throughout the remainder of the volcanic belts can also be explained by pure-shear deformation, because deformation in high-strain zones is just a more intense expression of Proterozoic vertical deformation. Clearly, the abundant heterogeneities in volcanic piles mean that there was usually some simple-shear component to the deformation, but the idea that the vertical fabrics of Proterozoic deformation formed largely by pure shear has profound implications for the kinematics of deformation and for the way structures observed in the field must ultimately be interpreted.

Progressive strain explains reorientation of features by sequential superimposition of a multitude of infinitesimally small strains (Ramsay, 1967). Thus, features such as lineations and fold axes can be *rotated* progressively to vertical by imposing simple-shear strain on a preexisting finite strain. Such a rotational model of progressive strain, extrapolated to no strain, implies that all features started horizontal prior to rotation, but field evidence clearly shows that structures such as lineation and folds did not start horizontal, but formed in steep orientations at the outset. A model of progressive strain by continuous pure shear, however, produces *no rotation:* features continue to shorten horizontally and extend vertically by pure shear distortion, creating the *appearance* that they are rotated to vertical, when in fact *no true rotation is involved* (e.g., bedding in fig. 20 is distorted, not rotated, to vertical).

This pure shear model explains how features of Proterozoic vertical deformation—foliation, lineation and

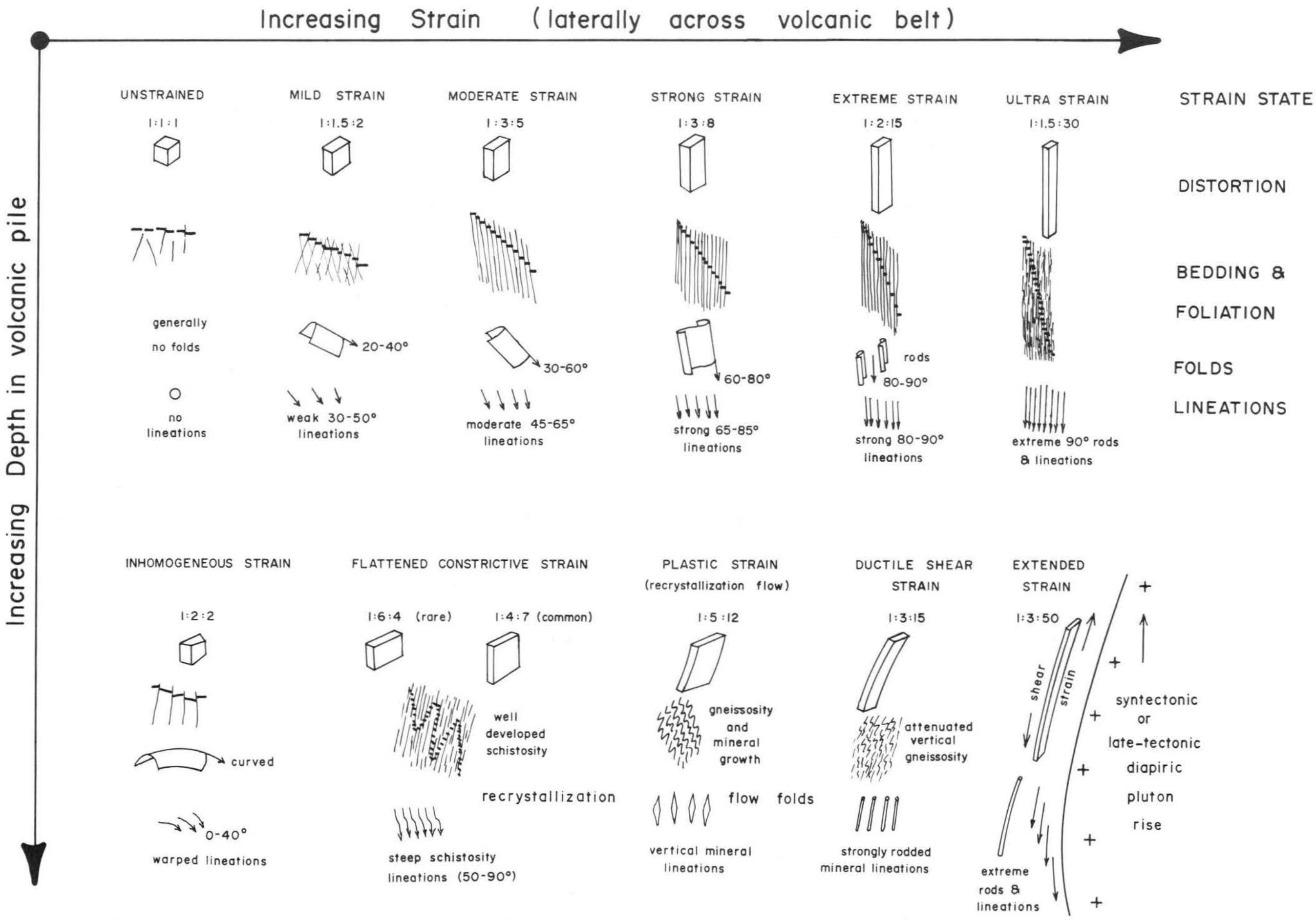


Figure 20. Proterozoic vertical deformational model for central Arizona in terms of increasing strain and increasing depth. The increasing strain axis depicts spatial change only, *not progression with time* (see explanation in text). The increasing depth axis depicts increasing hydrostatic stress (confining pressure) and increasing metamorphic grade (temperature). Names assigned to each strain state match those in the text. Axial ratios of a typical finite strain ellipse for each strain state are quoted, but a cube is used to graphically depict the strain ellipse more clearly than an ellipsoid. The relationship of foliation to bedding is shown as light vertical lines (foliation) offsetting heavy lines (bedding) to produce steeper bedding dips at higher strain states. The heavy lines only approximate bedding in the field, which dips more steeply than can be shown on the diagram. The figure shows the types of folds typical of each strain state, the lineation intensities and plunges, and the typical ranges of fold and lineation plunges. All deformational regimes on the diagram except one can be accounted for by a combination of mainly *pure-shear* deformation with lesser *simple-shear* deformation. The exception is the regime of extended strain in the region of diapiric pluton rise, where simple shear overwhelms pure shear as the diapiric body adds a shear couple to the high-strain regime. The term *extended strain* thus has a specific meaning: it is a regime where fabrics (lineations, rods, folds, and distorted features) underwent simple vertical shear far beyond the normal axial ratios of high strain. Zones of high strain would typically include the strong, extreme and ultrastrain regimes. Where the extended strain regime occurred in a high-strain zone at high crustal levels, the zone would become a fault or shear zone, but at greater depth, linear and planar features deform plastically and become wrapped around the surface of the diapiric pluton.

folds—started with steep attitudes and are not commonly found with subhorizontal attitudes in most parts of the belts. The rocks had to first undergo sufficient pure shear distortion before strain was intense enough to produce each feature: (1) 30 to 40 percent distortion could produce folds, so they started with shallower plunges than lineations in most areas; (2) more than 50 percent distortion was required before lineations became well developed, so they typically started out with steeper plunges; (3) cleavage began on shear surfaces of the pure-shear finite-strain ellipse at low-strain states, forming acute intersecting cleavage sets; but (4) at higher strain states, foliation paralleled the principal elongation axes of the strain ellipse, which remained vertical throughout the history of pure-shear strain. Under progressive simple shear, even foliation must rotate.

Simple Shear

Despite the predominance of pure shear throughout the volcanic belts, simple shear kinematics and its attendant offsets are well displayed in certain places, including: (1) shear zones and ductile faults; (2) segments of high-strain zones with significant displacement; (3) areas around plutons that rose diapirically during deformation; and (4) the borders of the volcanic belts, especially near the edges of major batholiths. Offsets were oblique along the Moore Gulch and Chaparral shear zones and dominantly vertical along the Black Canyon zone, Deadman Wash fault, and Gun Creek zone. Where high strain and shear offsets exist in the same zones (Moore Gulch, Gun Creek and Chaparral zones), field evidence shows that shear offset postdated vertical strain, but was part of the same deformation regime as the main event of vertical high strain. Thus as deformation progressed, simple shear took over as a key kinematic mechanism in some high-strain zones or in regions of ductile behavior and extended strain (fig. 20). The result was significant offset across particular zones, but no general shear couple or offset across the width of the volcanic belts.

Different structural responses of the Shylock and Black Canyon zones to progressive strain is an ideal example. Rocks were drawn down into the Shylock and Black Canyon zones primarily by vertical pure shear: in the Shylock zone, later simple shear was minor and readjusted crustal blocks either side of the zone and produced minor folds with axes orthogonal to those of earlier vertical deformation; in the Black Canyon zone, however, strata were drawn down to great crustal depths, sufficient to partially melt pelitic rocks over a wide region of the southern Bradshaw Mountains. Granitic anatectite rose from the partly fused migmatite terrane to be emplaced higher in the crust as the Crazy Basin body. Its rise created a regime of ductile simple shear in surrounding strata that caused rock fabrics to become physically extended vertically (fig. 20), as major simple-shear motion occurred across the Black Canyon belt. This resulted in kilometers of net vertical offset across the width of the zone, bringing up deeper crustal levels west of the zone to higher levels.

The timing of the Crazy Basin body's rise and thermal metamorphism—in the closing stages of deformation when earlier fabrics were overprinted by intensely elongate, recrystallized vertical fabrics—is exactly predicted by the model of simple shear following vertical pure shear in high-strain zones. Simple shear offset across the zones was a response to rising heat flow during progressive deformation and metamorphism, and its ductile nature allowed strong vertical shear and crustal offset to occur in some parts of the zones (Black Canyon belt) without requiring comparable offset in other parts (Shylock zone). Progressive deformation in the Black Canyon belt and its changing strain regime with time models other high-strain or shear zones in central Arizona with clear evidence of simple shear offset. Such motions were part of the vertical crustal readjustment of the entire central volcanic belt during Proterozoic accretion.

Strain Progression with Time

From all foregoing discussions, it can now be deduced how strain progressed through the volcanic belt with time. All areas did not start out weakly strained and progress (across fig. 20) to higher strain with time, nor did high-strain geometries evolve from low-strain geometries. Because primary competency changes in the volcanic pile were mostly vertical, vertical weakness zones destined to become high-strain zones took up strain first, long before strain penetrated competent regions. Furthermore, rocks in such zones were subjected to strong strain at the outset (i.e., large contrasts in magnitude of the strain axes), so their structures did not evolve from those of weak strain states (i.e., left to right on fig. 20), but instead assumed orientations at the outset commensurate with higher strain states. Where the magnitude of stress in both horizontal axes exceeded that of the vertical axis, neither lineations nor folds formed with shallow to subhorizontal attitudes, but assumed steeper orientations from the outset. *As pure shear progressed to distort structures in these zones to vertical, so strain penetrated into adjacent more competent regions of the volcanic belt, lastly to affect the most massive units of all.*

The above picture of strain progression with time is an *instantaneous strain model,* which says that at any instant of time, the infinitesimal strain picture of the volcanic belt closely reflects the finite (i.e., final) strain picture. This model is clearly distinct from ones assuming that higher strain states evolved from lower strain geometries—such models require that strain progression rotated subhorizontal features to vertical. *Rotational models* of progressive strain say that the infinitesimal picture changes continually and does not closely reflect the final picture (Ramsay, 1967).

There is good evidence that the instantaneous model accurately depicts progressive strain in the volcanic belts, especially where vertical pure shear dominated. In such places, the lack of rotation of features supports the model and suggests that structures assumed orientations at the outset that directly reflected the magnitude of finite strain caused by the orientation and relative magnitudes of the stress axes. In other places, however, such as at edges of the volcanic belts or near pluton margins where simple shear predominated, the rotational model is more accurate and can succinctly explain unusual features, such as the local refolding of minor folds near the Brady Butte pluton (O'Hara, 1980). Progressive migration of strata off a pluton flank during its diapiric rise and their descent into the volcanic trough causes earlier cascading folds to be refolded as the regime changes from flattening strain to horizontally constrictive strain; exactly these types of folds occur near the Brady Butte body.

To detail all such variants of Proterozoic vertical deformation and itemize the regimes where strain remained roughly constant versus those where it changed continuously with time is beyond the limits of this paper. Elegant experimental support for the writer's model of Proterozoic vertical deformation came from Dixon and Summers (1983), whose vertical tectonic models remarkably closely match deformation of the central volcanic belt, except for a flat bottom to their trough, which is absent in Proterozoic belts (fig. 18). They call high-strain zones "zones of strong vertical extension," their models support the constrictive nature of strain, support local refolding as material migrated into the trough, and confirm the predominance of synclinal keels in the volcanic troughs without complementary anticlines. The interested reader is referred to this classic paper for further information on vertical deformation.

CONCLUSIONS

Evidence overwhelmingly indicates that steep to vertical structures in central Arizona Proterozoic volcanic belts were produced by a single major event of deformation without first forming subhorizontal fabrics and rotating them to vertical. This does not require deformation in all parts of central Arizona to be exactly synchronous, because there is clear evidence that deformation in the Prescott region predated that in eastern volcanic belts [see tectonics paper], nor does it preclude local superimposed folding at sites where the strain regime changed substantially during progressive deformation. Formation of vertical fabrics in one event *does* mean that vertical deformation proceeded under a single stress regime, of highly variable strain, and that the terrane was not first deformed under one stress regime, then later redeformed under a different stress regime, as in polydeformed regions throughout the world.

Because of their primary lensoidal makeup, Proterozoic volcanic piles became *dominantly deformed terranes, not folded terranes,* so major closures of volcanic stratigraphy around regional folds do not exist, and bedding did not originally trend northwest perpendicular to foliation. Instead, the original northeast-trending depositional troughs helped guide the orientation of foliation. Folds internal to major units were best developed at moderate to low strain, in well-bedded strata, and where bedding locally diverged from regional foliation.

This study defines and describes in detail *high-strain zones* that developed in the volcanic belts at all scales, the largest marking the axes of strongest vertical extension in them. High-strain zones are integral parts of vertical deformation and represent zones where foliation and lineation are much more intensely developed than in adjacent areas. In places (e.g., Sheep Basin Mountain "horseshoe fold") it can be documented how rock units were drawn down toward keels of the zones by extreme vertical extension to produce structures resembling refolded folds in plan view. Intense foliation was not synonymous with offset: in fact a key aspect of high-strain zones is a dominance of pure shear vertical extension over simple shear; hence they are different from shear zones with major offset. Distributed vertical strain may change to shear offset where high-strain zones cut through granitic masses along strike (e.g., Moore Gulch shear zone). Changing simple shear component along strike is compensated for by ductile behavior and does not imply scissor or rotational motion.

Another key feature of Proterozoic high-strain zones is their widening and strain distribution with depth. At surface they resemble vertical faults (Deadman Wash), at high crustal levels they are narrow (Shylock zone), but at deeper crustal levels strain is more widely distributed (Black Canyon belt). This is the opposite of many shear zones and thrusts that narrow and become cryptic with depth; competency changes govern where high-strain zones become cryptic. Distributed vertical strain at depth was thus manifested as offset on vertical faults as zones surfaced. Finally, all high-strain zones in central Arizona show complex histories of typically an early event of vertical pure shear, later superimposed by simple shear offset as deformation progressed to more ductile conditions, and finally either oblique slip or brittle vertical normal faulting after deformation.

Stratigraphy of the central volcanic belt was never oriented to vertical prior to deformation and large overturned or recumbent sections are not known. In most parts of the belts, the original way up is still up, both regionally and in detail. Thus, stratigraphy kept its proper context during deformation, and broad-scale geologic relationships of the belts are preserved. The original question posed at the beginning of this paper—"how was the stratigraphy oriented to vertical in the first place?"—is therefore reduced to the mechanistic question of "how, kinematically, can vertical fabrics be imposed upon a horizontal sequence?." This paper has answered that question.

The strain analysis presented in this paper shows how the structures of Proterozoic vertical deformation started in moderate to steep attitudes as strain reaches certain threshold levels that allow each feature to be expressed, then increased pure shear strain beyond those levels progressively distorted all features, including foliation, bedding, lineations, primary features, and folds, to vertical at maximum degrees of horizontal shortening and vertical extension. The analysis also shows that the axial ratios of distorted features, foliation intensity, and both intensity and steepness of lineations are the best measures of strain and its variations throughout the volcanic belts. Fold models and stereographic analyses fail to develop a similarly accurate strain picture.

Deformation penetrated vertical zones of weakness in the volcanic pile first, causing high strain in them before adjacent, more competent units took up significant strain. It is entirely possible that the earliest fabrics in these high-strain zones originated in near-vertical attitudes from the outset, because of extremely fissile response to strain and initially large contrasts in magnitude of strain imposed on them. Therefore, regardless of whether the total net strain was weak or strong, *the geometry of features such as lineation and folds that formed in response to strain remained approximately constant throughout the history of strain.* This means that lineations originated with shallow to moderate plunges in some places, and with steep plunges in others, depending on the kinematics of strain at each locality. In other areas near the edges of the volcanic belt, simple-shear strain clearly rotated structures toward vertical during progressive deformation, and where strain progressed to extreme degrees in some high-strain zones, rock fabrics became extended or "drawn out" by simple shear, and major vertical dislocations resulted.

This presents essentially an instantaneous strain model for deformation of Proterozoic volcanic belts. Changes in plate motions are geologically instantaneous events that can turn on or off the forces that are the essential driving mechanism behind horizontal shortening of the crust and the deformation that results from it. Hence it is entirely realistic to predict instantaneous strain response to stress. The reason this paper considered only strain in the volcanic belts and not causative stresses is because stress is imposed on the crust over a scale much broader than the volcanic belts, and is ultimately linked to plate tectonic interactions at the accreting Proterozoic continental margin [see tectonics paper]. On this very broad scale, *the volcanic belts passively absorb strain* as the competent batholithic masses around them shear and readjust to produce the most efficient crustal shortening possible. Much adjustment involves density compensation, whereby lighter plutonic rocks rise as denser rocks of the volcanic belts deform downward into synclinal troughs and keels. The volcanic belts thus invariably shorten in all horizontal directions, not just one direction, and as a result, the direction of material transport in the belts is vertical. *Proterozoic vertical deformation* accurately describes this vertical motion.

Possibly the most profound aspect of Proterozoic vertical deformation is that the structural fates of the volcanic belts were sealed before they were ever deformed. Ultimately it was the original stratigraphic makeup of the volcanic belts—their thick domal volcanic edifices and intervening zones of crustal weakness, the disposition of pretectonic batholiths around them, and ancestral bounding fault scarps delimiting original depositional basins—that was to govern how the volcanic belts were to deform. These features alone controlled the great variations of strain within the belts, *variations to which the stress field was completely blind.* Consequently, Proterozoic vertical deformation is not only a unique structural style, it was a necessary part of the tectonic evolution of Proterozoic volcanic belts worldwide, no less fundamental than the primary volcanic makeup of the belts, to which it is inseparably linked in the final analysis. This suggests that perhaps the most sound, enduring structural analysis of Proterozoic volcanic belts should start with a very detailed understanding of their *stratigraphic makeup,* rather than a compass, stereonet, and many, many structural readings.

ACKNOWLEDGMENT

Both the writer and these papers are greatly indebted to Steve Reynolds, whose unfailing confidence and untiring persistence became the cornerstone of their completion.

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Revised manuscript accepted 1989.