Cenozoic paleogeography of the west-central United States: Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, Rocky Mountain paleogeography Symposium 3, p. 335-355.
Nelson, B. K., and DePaolo, D. J., 1985, Rapid production of continental crust 1.7 to 1.9 b. y. ago: Nd isotopic evidence from the basement of the North American mid-continent: Geological Society of America Bulletin, v. 96, p. 746-754.
Palmer, A. R., 1983, The Decade of North American Geology 1983 time scale: Geology, v. 11, p. 503-504.
Peiffer-Rangin, Francoise, 1979, Le zones isopiques du Paleozoique inferieur du nord-ouest mexicain, temoins du relais entre les Appalaches et la cordillere ouest-americaine: Academie Sciences [Paris] Compte Rendu, Series d, v. 288, p. 1517-1519.
Peirce, H. W., 1976, Elements of Paleozoic tectonics in Arizona: Arizona Geological Society Digest, v. 10, p. 37-57.
Peirce, H. W., Damon, P. E., and Shafiqullah, M., 1979, An Oligocene Colorado Plateau edge in Arizona: Tectonophysics, v. 61, p. 1-24.
Peirce, H. W., Shafiqullah, M., and Breed, W. J., 1985, Geochronology of some exotic cobbles, Chinle Formation, Arizona [abs]: Flagstaff, Museum of Northern Arizona Abstracts of the Symposium on Southwestern Geology and Paleontology 1985, p. 7.
Rea, D. K., and Bryant, D. L., 1968, Permian red chert pebble conglomerate in Earp Formation, southeastern Arizona: American Association of Petroleum Geologists Bulletin, v. 52, p. 809-819.
Rehrig, W. A., and Reynolds, S. J., 1980, Geologic and geochronologic reconnaissance of a northwest-trending zone of metamorphic core complexes in southern and western Arizona, *in* Crittenden, M. D., Jr., Coney, P. J., and Davis, G. H., eds., Cordilleran metamorphic core complexes: Geological Society of America Memoir 153, p. 131-157.
Reynolds, S. J., 1980, Geologic framework of west-central Arizona, *in* Jenney, J. P., and Stone, Claudia, eds., Studies in western Arizona: Arizona Geological Society Digest, v. 12, p. 1-16.
Reynolds, S. J., and Keith, S. B., 1982, Geochemistry and mineral potential of peraluminous granitoids: Arizona Bureau of Geology and Mineral Technology Field Notes, v. 12, no. 4, p. 4-6.
Reynolds, S. J., and Spencer, J. E., 1985, Evidence for large-scale transport on the Bullard detachment fault, west-central Arizona: Geology, v. 13, p. 353-356.
Reynolds, S. J., Welty, J. W., and Spencer, J. E., 1986, Volcanic history of Arizona: Arizona Bureau of Geology and Mineral Technology Field Notes, v. 16, no. 2, p. 1-5.
Rose, P. R., 1976, Mississippian carbonate shelf margins, western United States: United States Geological Survey Journal of Research, v. 4, p. 449-466.
Scarborough, R. B., and Peirce, H. W., 1978, Late Cenozoic basins of Arizona, *in* Callender, J. F., Wilt, J. C., and Clemons, R. E., eds., Land of Cochise: New Mexico Geological Society 29th Field Conference Guidebook, p. 253-259.
Schweickert, R. A., 1976, Shallow-level plutonic complexes in the eastern Sierra Nevada, California, and their tectonic implications: Geological Society of America Special Paper 176, 58 p.
Schweickert, R. A., 1978, Triassic and Jurassic paleogeography of the Sierra Nevada and adjacent regions, California and western Nevada, *in* Howell, D. G., and McDougall, K. A., eds., Mesozoic paleogeography of the western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 361-384.
Seager, W. R., Shafiqullah, M., Hawley, J. W., and Marvin, R. F., 1984, New K-Ar dates from basalts and the evolution of the southern Rio Grande Rift: Geological Society of America Bulletin, v. 95, p. 87-99.
Sears, J. W., and Price, R. A., 1978, The Siberian connection: a case for Precambrian separation of the North American and Siberian cratons: Geology, v. 6, p. 267-270.
Shafiqullah, M., Damon, P. E., Lynch, D. J., and Kuck, P. H., 1978, Mid-Tertiary magmatism in southeastern Arizona, *in* Callender, J. F., Wilt,, J. C., and Clemons, R. E., eds., Land of Cochise: New Mexico Geological Society 29th Field Conference Guidebook, p. 231-241.
Shafiqullah, M., Damon, P. E., Lynch, D. J., Reynolds, S. J., Rehrig, W. A., and Raymond, R. H., 1980, K-Ar geochronology and geologic history of southwestern Arizona and adjacent areas, *in* Jenney, J. P., and Stone, Claudia, eds., Studies in western Arizona: Arizona Geological Society Digest, v. 12, p. 201-260.
Shride, A. F., 1967, Younger Precambrian geology in southern Arizona: United States Geological Survey Professional Paper 566, 89 p.
Silver, L. T., 1978, Precambrian formations and Precambrian history in Cochise County, southeastern Arizona, *in* Callender, J. F., Wilt, J. C., and Clemons, R. E., eds., Land of Cochise: New Mexico Geological Society 29th Field Conference Guidebook, p. 157-163.
Silver, L. T., and Anderson, T. H., 1974, Possible left-lateral early to middle Mesozoic disruption of the southwestern North American craton margin: Geological Society of America Abstracts with Programs, v. 6, no. 7, p. 955-956.
Silver, L. T., Bickford, M. E., Van Schmus, W. R., Anderson, J. L., Anderson, T. H., and Medaris, L. G., 1977, The 1.4-1.5 b.y. transcontinental anorogenic plutonic perforation of North America: Geological Society of America Abstracts with Programs, v. 9, no. 7, p. 1176-1177.
Smith, A. G., Hurley, A. M., and Briden, J. C., 1981, Phanerozoic paleocontinental world maps: Cambridge, England, Cambridge University Press, 102 p.
Speed, R. C., 1979, Collided Paleozoic microplate in the western United States: Journal of Geology, v. 87, p. 279-292.
Spencer, J. E., 1984, Role of tectonic denudation in warping and uplift of low-angle normal faults: Geology, v. 12, p. 95-98.
Steiner, M. B., 1983, Mesozoic apparent polar wander and plate motions of North America, *in* Reynolds, M. W., and Dolly, E. D., eds., Mesozoic paleogeography of the west-central United States: Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, Rocky Mountain Paleogeography Symposium 2, p. 1-11.
Stewart, J. H., 1972, Initial deposits in Cordilleran geosyncline: evidence of a late Precambrian (850 m.y.) continental separation: Geological Society of America Bulletin, v. 83, p. 1345-1360.
Stewart, J. H., 1976, Late Precambrian evolution of North America: plate tectonics implications: Geology, v. 4, p. 11-15.
Stewart, J. H., McMenamin, M. A. S., and Morales-Ramirez, J. M., 1984, Upper Proterozoic and Cambrian rocks in the Caborca region, Sonora, Mexico—physical stratigraphy, biostratigraphy, paleocurrent studies, and regional relations: United States Geological Survey Professional paper 1309, 36 p.
Stewart, J. H., Anderson, T. H., Haxel, G. B., Silver, L. T., and Wright, J. E., 1986, Late Triassic paleogeography of the southern Cordillera: the problem of a source for voluminous volcanic detritus in the Chinle Formation of the Colorado Plateau region: Geology, v. 14, p. 567-570.
Thomas, J. J., Shuster, R. D., and Bickford, M. E., 1984, A terrane of 1,350- to 1,400-m.y.-old silicic volcanic and plutonic rocks in the buried Proterozoic of the mid-continent and in the Wet Mountains, Colorado: Geological Society of America Bulletin, v. 95, p. 1150-1157.
Titley, S. R., 1981, Geologic and geotectonic setting of porphyry copper deposits in the southern Cordillera, *in* Dickinson, W. R., and Payne, W. D., eds., Relations of tectonics to ore deposits in the southern Cordillera: Arizona Geological Society Digest, v. 14, p. 79-97.
Titley, S. R., 1982, Geologic setting of porphyry copper deposits, southeastern Arizona, *in* Titley, S. R., ed., Advances in geology of the porphyry copper deposits, southwestern North America: Tucson, University of Arizona Press, p. 37-58.
Van Schmus, W. R., and Bickford, M. E., 1981, Proterozoic chronology and evolution of the mid-continent region; North America, *in* Kroner, Alfred, ed., Precambrian plate tectonics: Amsterdam, Elsevier, p. 261-296.
Vedder, L. K., 1984, Stratigraphic relationship between the Late Jurassic Canelo Hills Volcanics and the Glance Conglomerate, southeastern Arizona: Tucson, University of Arizona, unpublished M.S. thesis, 129 p.
Walker, J. D., and Burchfiel, B. C., 1983, Permotriassic rocks in the northern Mojave Desert region and their tectonic significance: Utah Geological and Mineral Survey Special Studies 60, p. 48-50.
Wright, L. A., Troxel, B. W., Williams, E. E., Roberts, M. T., and Diehl, P. E., 1976, Precambrian sedimentary environments of the Death Valley region, eastern California: California Division of Mines and Geology Special Report 106, p. 7-15.
Young, R. A., 1979, Laramide deformation, erosion, and plutonism along the southwestern margin of the Colorado Plateau: Tectonophysics, v. 61, p. 25-47.
Zoback, M. L., Anderson, R. E., and Thompson, G. A., 1981, Cainozoic evolution of the state of stress and style of tectonism of the Basin and Range Province of the western United States: Royal Society [London] Philosophical Transactions, v. A300, p. 407-434.

Revised manuscript accepted 1986.

in Jenney, J. P., and Reynolds, S. J., 1989, Geologic evolution of Arizona: Tucson, Arizona Geological Society Digest 17, p. 17-55.

PROTEROZOIC PLATE TECTONIC EVOLUTION OF ARIZONA

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ABSTRACT

The tectonic processes that built new crust at convergent continental margins during the Proterozoic were actualistic to modern plate tectonics, but differed in detail sufficiently to warrant recognition that a unique style of plate tectonics operated during the Proterozoic era. Arizona is a key reference section for tectonic interactions and the plate tectonic processes that accreted the continental crust of the United States from 2.0 Ga to 1.3 Ga.

Arizona's Proterozoic tectonic evolution began with rifting of the Archean Wyoming craton at about 2.0 Ga and growth of Proterozoic oceanic crust throughout Arizona. Archean-derived clastics filled a shelf-slope wedge along the Wyoming craton margin that was deformed and intruded by basic dikes and possibly plutons prior to 1.8 Ga. On an upper Proterozoic oceanic crust of primitive tholeiites, layered mafic-ultramafics, spilite, keratophyre, pelagic sediments, and chert, the Prescott-Jerome belt developed as an intraoceanic island arc from 1800 Ma to 1740 Ma above a subduction zone dipping southeast under the Archean fragment detached from the Wyoming craton. Three tholeiitic volcanic provinces of the Bradshaw Mountains, Mayer, and Black Canyon Creek Groups developed sequentially across the arc to form a southeast trend of alkali enrichment that may have included a calc-alkaline province farther southeast.

A major 1750-1740-Ma change in plate motions caused subduction to flip and thereafter dip northwest under the Wyoming Archean craton. This flip shut off formative volcanism and produced plutons and batholiths with a northwest alkali-enrichment trend across the Prescott-Jerome arc. As the allochthonous arc swept toward the Wyoming craton, the intervening ocean basin was consumed to create the Antler-Valentine volcanic belt as an incipient continent-margin arc fronting the deformed clastic wedge at the continental margin. The Bagdad volcanic belt (previously formed in a spreading-center, arc, or oceanic-island setting), together with (1) fore-arc basin sediments and melange to the south that remained from the earlier subduction under the Prescott-Jerome arc, (2) the Antler-Valentine volcanic belt, and (3) the craton-margin clastic wedge to the north, resisted subduction as the ocean basin closed. Across the United States, all similar oceanic and arc elements were coalesced into a new Proterozoic crust that was accreted to the Archean margin by about 1730 Ma.

Northwest-dipping subduction produced granodiorite-tonalite batholiths and plutons across the Prescott-Jerome arc and made it an emergent continent-margin arc that was separated from the Wyoming craton by a wide, shallowly submerged, back-arc basin of accreted oceanic elements. By 1730 Ma, a chain of submarine calc-alkaline Union Hills Group volcanoes had formed along the new southeast front of the arc, turbidite graywackes were shed into intervolcanic basins, and volcaniclastic detritus was dispersed into a fore-arc basin to the southeast. From 1730 to 1720 Ma, the trench started to shift oceanward in progressive increments as a subduction complex grew to the southeast, which caused both the fore-arc basin to prograde over the melange and the subduction dip to flatten.

As subduction flattened, plutonism stepped farther inboard into the oceanic collage behind the arc, extending the alkali-enrichment trend of arc plutonism to the northwest. Continued subcrustal heating produced widespread fusion of oceanic, arc, and sedimentary crusts behind the Prescott-Jerome arc, so that by 1720 Ma the Northwest Gneiss Belt was pervaded by huge calc-alkaline batholiths as it underwent major orogeny, high-grade regional metamorphism, crustal thickening, and stabilization.

Concurrently, vertical structural readjustments in the continent-margin arc formed grabens at pluton edges for Texas Gulch clastics and felsic volcanism in the old emergent Prescott-Jerome part of the arc, and for Alder Group clastic sedimentation along the submerged front of the arc. These sedimentary events signified a major ca. 1720-Ma hiatus in evolution of the continent-margin arc, after mafic volcanism ended but before primary felsic magmas were emplaced into the arc front. This hiatus occurred as the trench stepped farther southeastward to create throughout southeast Arizona a very wide Pinal terrane of detritus shed from the arc and fore-arc basin, accreted oceanic sediments, and possibly allochthonous crustal pieces swept in from southeasterly intraoceanic settings.

Formation or accretion of the Dos Cabezas arc moved the trench out of Arizona and established the Pinal terrane as a wide inter-arc basin between the Dos Cabezas arc and the central Arizona magmatic arc. Very shallow subduction under the Pinal basin from 1705 to 1695 Ma caused crustal fusion and resulting felsic magmatism across the frontal 350 km of the continental margin. Felsic volcanics overwhelmed Alder clastics as huge felsic magmas were emplaced into the central Arizona arc to crystallize as granite batholiths beneath ignimbrite carapaces. Calc-alkaline rhyolites erupted in the Dos Cabezas and Ray-Aravaipa belts above shallow depths on the subduction zone, whereas alkali-calcic ignimbrites erupted at the front of the central Arizona magmatic arc above greater depths on the subduction zone. Felsic tephra were shed oceanward into the inter-arc Pinal basin coextensively with its younger quartz-wacke sedimentation.

Thus by 1690 Ma, the central Arizona magmatic arc became fully emergent and, together with emergent older terranes behind it, formed a newly evolved Proterozoic continental crust. As the arcs eroded, Mazatzal strata succeeded felsic volcanics as fluvial, estuarine, littoral, and shallow-marine conditions prograded back across the central Arizona arc, and as open-marine conditions in the Pinal basin and shoaling in the Dos Cabezas arc persisted to 1680 Ma.

Deformation younged southeastward across Arizona: (1) in the northwest it accompanied major 1720-Ma batholiths; (2) in the Prescott-Jerome arc it lasted from 1710 to 1695 Ma, as ignimbrites erupted to the east; (3) in the eastern central magmatic arc and the northern Pinal basin it extended from 1660 to 1650 Ma; and (4) in the southern Pinal basin and Dos Cabezas arc it lasted from 1630 to 1620 Ma, when the southeast terranes were sutured to the margin. Broadly, Proterozoic deformation progressed southeastward as a wave of orogenic disturbance that immediately followed crustal formation and initial plutonism, and that immediately preceded the peak of regional metamorphism, because the process of crustal shortening caused major heat flow in the crust.

Thus there was a fundamental order to the process of crustal growth and the timing of continental accretion at the Proterozoic convergent plate margin. Proterozoic lithospheric subduction, which first dipped southeast and later northwest under Arizona, was ultimately responsible for all events of volcanism, plutonism, deformation, and metamorphism. The tectonic events that formed and accreted new Proterozoic crust at the convergent plate margin, as summarized in this paper, define the Proterozoic plate tectonic style.

INTRODUCTION

Proterozoic orogeny was once perceived as ensialic reworking of Archean crust to form intracratonic "Proterozoic mobile belts" (Read and Watson, 1972; Sutton, 1972). Dewey and Spall (1975) and P. Anderson (1976) first suggested that *processes* more akin to those of modern plate tectonics operated during the Proterozoic to generate crust at Proterozoic accretionary continental margins, but recognized that the *features* of modern plate tectonics are radically different from those of Archean tectonics. Thus, Proterozoic plate tectonic regimes are a critical link in understanding how the Earth's processes changed with time (Dewey and Spall, 1975; P. Anderson, 1986). It is now generally accepted that some form of plate tectonics operated during the Proterozoic era, but the exact nature of such tectonic processes is not well understood because the generation of new Proterozoic crust has not been documented in detail.

Proterozoic continental margins, where crust formed and was accreted during the Proterozoic era, are as pivotal to understanding Proterozoic plate tectonics as active continental margins are to analyzing modern plate tectonics. The entire Proterozoic crust of Arizona is made up of rock packages that formed at an accreting Proterozoic continental margin; collectively it comprises the best preserved cross section of newly formed Proterozoic crust in the United States. The writer's systematic, comprehensive study of Arizona's Proterozoic crust during the past 15 years has provided insight into how this crust was generated between 2.0 and 1.6 Ga, how its history constrains the plate tectonic interactions that could have formed it, and what that history says about the processes of Proterozoic plate tectonics.

This paper outlines the plate tectonic evolution of Arizona from its inception at about 2.0 Ga to its major consolidation by 1.6 Ga. The subsequent middle and late Proterozoic history is given elsewhere (P. Anderson, n. d.). In all plate tectonic models, different plate interactions can lead to the same resultant tectonic configuration (Dewey, 1975). Where geologic relations in the Arizona Proterozoic are well known (see P. Anderson, this volume), the possibilities are tightly constrained to only a few models that are reasonable, but where relationships are less clear, such as in northwest and southeast Arizona (and may never be well known because of later tectonic overprinting), multiple tectonic alternatives are considered. This paper first summarizes the tectonic settings of Proterozoic belts that make up Arizona and then derives a plate tectonic model that accounts for the observed features.

PREVIOUS OVERVIEWS

Since the first pioneering geologic work in Arizona (Walcott, 1890; Jaggar and Palache, 1905; Van Hise and Leith, 1909), it has been recognized that three fundamentally different rock assemblages constitute the Proterozoic crust of Arizona. The dominantly volcanic makeup of central Arizona (Ransome, 1915; Lindgren, 1926; Lausen, 1930; Wilson, 1922, 1939) contrasts markedly to the highly metamorphosed sedimentary basement rocks of the Grand Canyon (Van Hise, 1892; Noble and Hunter, 1916; Darton, 1932; Hinds, 1938; Keyes, 1938) and to the weakly metamorphosed sedimentary and volcanic rocks of southeast Arizona (Ransome, 1903, 1919, 1923; Darton, 1925; Darton and others, 1924).

Despite these three very different Proterozoic crustal compositions, subsequent studies concentrated on geologic evolution within each region, rather than on the broad interrelationships between regions. Consequently, overviews of the Precambrian structure and evolution of central Arizona (C. A. Anderson, 1951, 1966, 1968; Gastil, 1958; Silver, 1964, 1967, 1968; Livingston and Damon, 1968), northwest Arizona (Aldrich and others, 1957; Lanphere and Wasserburg, 1963; Pasteels and Silver, 1966; Brown and others, 1974), and southeast Arizona (Lance, 1959; Shride, 1961; Damon, 1962; Erickson, 1968) have been presented largely independently of one another (Wilson, 1962).

Wilson (1939) and Gastil (1958) saw the correct order of evolution of central Arizona's Proterozoic crust: felsic volcanics and sediments were built upon a basement of mafic volcanics, not the opposite, as is claimed for other Proterozoic regions. Tectonic models proposed for central Arizona include ones of primary continental formation (Gastil, 1962) and a resemblance to Archean greenstone belts (C. A. Anderson and Silver, 1976); for southeast Arizona, a deep-basin geosynclinal model has been implied (Silver, 1978); and for northwest Arizona, multiple orogenic cycles have been suggested (Livingston and others, 1974).

The first plate tectonic description of Arizona's Proterozoic tectonic evolution (P. Anderson, 1976) provided a means to integrate the three different crustal

belts of Arizona into a systematic pattern of crustal accretion at an evolving Proterozoic continental margin. Since then it has become increasingly clear that a style of plate tectonics unique to the Proterozoic era operated to generate the Proterozoic crust of Arizona (P. Anderson, 1976, 1978a; P. Anderson and Guilbert, 1979; P. Anderson, 1986), and this plate tectonic model for Arizona's Proterozoic tectonic evolution has been largely adopted by others (Condie, 1982; Silver and others, 1986; DeWitt, this volume).

Tectonic studies (P. Anderson, 1976, 1978a, 1986; P. Anderson and Guilbert, 1979) suggest that Proterozoic volcanic belts were similar in many ways both to modern island arcs and Archean greenstone belts in overall size, volcanic structure, chemical makeup, and evolution, but also display distinct dissimilarities to both modern and Archean analogs, implying unique features and possibly mode of formation of Proterozoic volcanic belts. Deformation of the Arizona Proterozoic volcanic belts more closely resembles Archean than Phanerozoic deformation (P. Anderson, 1976); however, Proterozoic deformation was generally more intense than Archean deformation, and detailed comparisons show that the differences outweigh similarities (P. Anderson, this volume).

This paper is condensed from a more detailed discussion of Arizona's Proterozoic plate tectonic evolution (P. Anderson, 1986) and provides a new perspective on where the oldest crustal fragments of Arizona were conceived, how they evolved, and how they were amalgamated into the diversity of tectonic assemblages that now constitute the Proterozoic crust of Arizona. The data presented in this paper and elsewhere (P. Anderson, 1986, n. d.) also offer new insight into how plate tectonics operated during the Proterozoic era.

BROAD PROTEROZOIC STRUCTURE OF ARIZONA

MAJOR TECTONIC BELTS

This study confirms that three fundamentally different major tectonic assemblages make up the Proterozoic crust of Arizona, and new maps showing the lithologic distribution of early Proterozoic stratified rocks in Arizona (figs. 1 and 2) clearly contrast these three different terranes. Figure 1 emphasizes rocks of sedimentary derivation, contrasting those in northwest Arizona to those in southeast Arizona. Figure 2 complements this by showing the distribution of Proterozoic rocks of volcanic derivation in all belts. Together these figures define:

(1) *The Northwest Gneiss Belt*—a diversely structured region of migmatitic paragneiss of mainly sedimentary and lesser volcanic derivation, parts of which were subjected to two different plutonic, metamorphic, and deformational events;

(2) *The Central Volcanic Belt*—a unified region of dominantly volcanic and volcaniclastic rocks subjected to a single major deformational, metamorphic, and plutonic cycle, whose timing differed slightly between older and younger portions;

(3) *The Southeast Schist Belt*—An enigmatic terrane of deep-submarine shales and immature quartz wacke interspersed by small volcanic centers, and subjected to lesser plutonism, metamorphism, and deformation than the other belts. These are not formal names and are capitalized in this paper only for clarity.

Superimposed on this primary Proterozoic crustal structure are younger orogenic effects, such as 1400-Ma plutonism and Tertiary mylonitic and metamorphic reworking, that greatly complicate the clear primary picture. Such younger effects have no bearing on the early evolution of Arizona's Proterozoic crust and are not considered in this paper.

Strata in the three belts have been referred to respectively as "Vishnu Schist" (migmatitic paraschist in the Grand Canyon), "Yavapai Schist" (metavolcanic schist in Yavapai county), and "Pinal Schist" (pelitic schist and semischist in Pinal county). The term "Yavapai Schist" has now been superseded by the more specific stratigraphic term **Yavapai Supergroup,** which includes all formative volcanic and related volcaniclastic sequences in north-central Arizona (P. Anderson, this volume). Pinal Schist has not yet been superseded by a new term, such as "Pinal Supergroup," because its stratigraphy of groups and formations has not yet been defined. If the term **Pinal Supergroup** is used, it should refer only to sedimentary and related volcanic sequences in southeast Arizona between the Salt River and Willcox (figures 1 and 2), because formative strata in this crustal segment originated in a different tectonic realm than those in adjacent segments.

Similarly, the Northwest Gneiss Belt contains two sedimentary sequences, with very different ages, that may overlap stratigraphically or occupy different crustal segments. Hence, two separate stratigraphic designations will be appropriate in the future when the pre-metamorphic stratigraphies of these two sequences are resolved. "Vishnu Schist" in the Grand Canyon represents only a small part of one such sequence and is neither typical nor representative of most Proterozoic exposures in northwest Arizona. Therefore, another term will be necessary for the extensive high-grade gneiss terranes that are widely exposed throughout Mohave County and that occupy the greater part of the Northwest Gneiss Belt.

Rocks of strictly sedimentary origin dominate the northwestern part of the Northwest Gneiss Belt north of Kingman (fig. 1). The region southeast of Kingman (fig. 1) consists of volcanic and clastic rocks derived from the Antler-Valentine volcanic belt (fig. 2). Farther southeast is a zone dominated by orogenic plutonic rocks and little supracrustal material (fig. 1), farther southeast still is the Bagdad volcanic belt (fig. 2), and farthest southeast is a region of dominantly sedimentary rocks between the Bagdad and Prescott volcanic belts (fig. 1). Thus, the

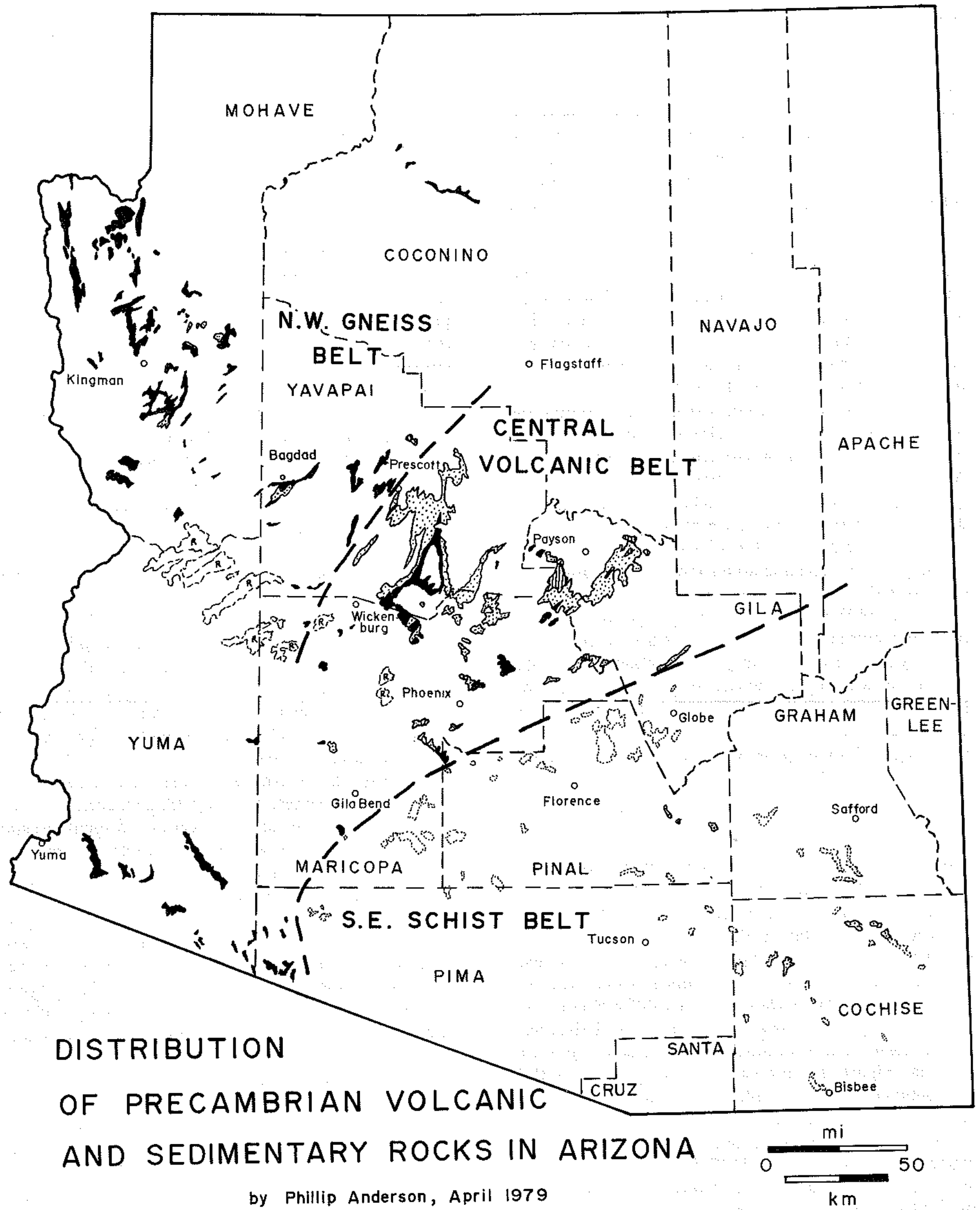

Figure 1. Distribution of Precambrian volcanic and sedimentary rocks in Arizona in relation to the three major lithologic belts—Northwest Gneiss Belt, Central Volcanic Belt, and Southeast Schist Belt—that make up the Proterozoic crust. Rocks of dominantly sedimentary origin are shown in black, rocks of volcanic origin are stippled, rocks of mainly volcanosedimentary origin in the Southeast Schist Belt are shown in dotted outlines, the Mazatzal Group is vertically ruled, and terranes remobilized by younger events are denoted "R."

Northwest Gneiss Belt consists of five different segments, based on lithologic characteristics alone. Each segment also has a tectonic history different from adjacent segments.

The Central Volcanic Belt includes the older, more mafic Prescott-Jerome volcanic belts to the northwest, and the younger, more felsic New River-Cave Creek-Mazatzal Mountains-Diamond Butte volcanic belts to the southeast (fig. 2 and P. Anderson, this volume). In addition, volcaniclastic metasedimentary rocks lie distal to all major volcanic chains of the central belt (fig. 1).

A line along the Salt River from Gila Bend to Globe (fig. 1) marks a major lithologic change in the Proterozoic makeup of Arizona: to the southeast volcanic belts are small and widely scattered (Ray-Aravaipa and Dos Cabezas belts, fig. 2), and the dominant earliest crustal constituents are well-layered quartz wacke, immature siltstone, gray shale, and other pelagic metasedimentary rocks. Except for isolated volcanic belts in this terrane and their derivative younger quartzites, which are clearly linked to evolution of the younger part of the Central Volcanic Belt, most sedimentary rocks in the Southeast Schist Belt bear little if any resemblance to those in north-central Arizona (fig. 1).

It should be borne in mind that the lithologic makeup described above refers *only to the oldest (formative) crustal components* of each Proterozoic belt in Arizona and does not take into account any later rocks, even the oldest orogenic plutonic rocks. Many such later rocks have little to do with tectonic interactions that produced the formative rock sequences, except for orogenic plutonic rocks that are essentially as old as the volcanic sequences themselves (P. Anderson, this volume), and which were ultimately generated by the same tectonic processes that gave rise to the volcanic belts.

PLUTONIC MAKEUP

The orogenic plutonic makeup of Arizona varies greatly from one belt to the next, but uniformity exists within each belt in that orogenic plutonism and deformation thickened and stabilized the crust typically 50 m.y. after its inception. This was part of a systematic cycle of new crustal arc generation, invasion by arc-type granodiorites, then deformation and crustal thickening, and is seen in all regions except possibly the oldest polycyclic gneiss terrane of far northwestern Arizona, which shows a more complex history.

Two types of orogenic plutonic rocks dominate the Northwest Gneiss Belt: an early suite of hornblende diorite-granodiorite plutons and batholiths that intruded edges of the volcanic belts at about 1740 Ma soon after formative volcanism, and a much more widespread suite of ca. 1720-Ma porphyritic granodiorite and monzogranite batholiths that pervaded the crust of northwest Arizona and effected its major crustal thickening and stabilization. These batholiths are so pervasive between Bagdad and the Antler volcanic belt (fig. 1), and again in areas north of Kingman, that few supracrustal deposits remain to track the earliest history of the Proterozoic crust in these regions.

The orogenic plutonic history of the northwest part of the Central Volcanic Belt (Prescott-Jerome belts) is similar to that in northwest Arizona, except that older (1760-1750-Ma) plutonic rocks exist in the volcanic belts, and early orogenic (1740-Ma) I-type plutons and batholiths occupy more of the crust and have more primitive compositions and chemistries in central Arizona than in northwest Arizona. Thus, the Central Volcanic Belt was thickened and stabilized mainly by 1740-Ma plutonic rocks, whereas the Northwest Gneiss Belt was thickened and stabilized later mainly by 1720-Ma porphyritic batholiths.

Southeast of the New River volcanic belt a totally different plutonic picture emerges (fig. 1). Except for dioritic plutons coeval with the oldest mafic volcanic sequences, pre-1720-Ma plutonic rocks are effectively absent: the earliest major batholiths are 1710-1700-Ma granites coeval with ignimbrites of the younger felsic complexes. These red granites are voluminous between the New River and Mazatzal Mountains and on both sides of the Diamond Butte volcanic belt (P. Anderson, this volume). South of the Central Volcanic Belt, however, only younger 1670-1650-Ma monzogranite and granodiorite batholiths and plutons prevail throughout the Southeast Schist Belt to the Mexican border (fig. 2). These younger plutonic rocks are tectonically equivalent to the 1720-Ma batholiths of northwest Arizona, in that they represented the first major orogenic plutonism and stabilization that the southern crust experienced.

Silver (1964, 1966, 1967, 1968, 1976) first recognized this plutonic boundary in central Arizona as a fundamental division in Arizona's Proterozoic tectonic evolution. The plutonic makeup, however, conveys a vastly simpler picture than the supracrustal structure (figs. 1 and 2) and by itself suggests that the tectonic histories of the Northwest Gneiss Belt and Central Volcanic Belt were much the same, which is definitely not true. Plutonic rocks become a secondary consideration in Arizona's total Proterozoic tectonic picture because, although locally more voluminous, they are less fundamental than the stratified supracrustal sequences, whose sources and origins reveal a far more intricate crustal tectonic makeup than do the plutonic rocks.

NATURE OF THE BASEMENT

Archean rocks in the western United States make up the Wyoming block, which extends from Montana to its southern limit at the Mullen Creek-Nash Fork shear zone in northern Colorado (Hills and Armstrong, 1974). Farther west in Utah, the southern Archean boundary is not clearly exposed, but Archean remnants occur in the Wasatch range of northern Utah, possibly as far south as 40°N. latitude (Hedge and Stacey, 1980). Thus, the southernmost known Archean rocks now lie 500 km north of the most northerly Proterozoic rocks exposed in Arizona. Based on Proterozoic plate reconstructions (P. Anderson, 1976), the distance to the Archean edge was probably 200 km or less during early Proterozoic time.

South of this Archean edge, specifically throughout the Proterozoic of Arizona, there is no record of Archean rocks: neither isotopic data nor field evidence suggests any rocks older than early Proterozoic (i. e., 2.2-2.0 Ga). Erosion of the Archean craton to the north almost certainly cycled some Archean detrital material into paraschists and gneisses of far northwestern Arizona, but 3.0-2.5 Ga detrital ages have not yet been found or systematically sought. Essentially a complete cross section of Arizona's original Proterozoic crust is displayed from highest supracrustal levels in central Arizona to the deeper levels in western Arizona and California (P. Anderson, 1976, 1978b). If any Archean remnants existed, they should have already been found. Thus, until geologic or isotopic evidence emerges to indicate otherwise, it must be concluded that Proterozoic rocks of Arizona did not form on Archean basement.

Evidence suggests instead that Proterozoic oceanic crust formed the basement to most of central and north-central Arizona (P. Anderson, 1986), and the volcanic belts developed both upon and through this mafic basement. Only Blacet (1966) thought he found granodioritic basement to the volcanic sequences, but later realized the granodiorite intruded volcanic rocks. Inclusions in the earliest mafic plutonic and volcanic units of the Central Volcanic Belt and most of the Northwest Gneiss Belt are almost always more mafic than the units themselves, and are locally ultramafic (i. e., pyroxenitic), which suggests both a mafic-ultramafic oceanic basement to, and oceanic source for, the volcanism.

Unfortunately, most evidence for the mafic-ultramafic basement has been eradicated by subsequent plutonism along the edges of the volcanic belts, but enough exposures exist to deduce the nature of the basement. Oldest mafic parts of the Prescott belt (e. g., Senator Formation), Jerome belt (e. g., Silver Spring Gulch diabase), and Bagdad belt (e. g., Mulholland Basin gabbro-anorthosite) all contain rock units similar to the upper portions of modern ophiolite sequences. At deep levels are differentiated gabbro-diorite bodies, local layered gabbro-anorthosite bodies, and usually medium- to fine-grained massive gabbro-diorite. At middle levels are subvolcanic gabbro, microgabbro, and diabase, locally with sheeted diabase dikes, and at upper levels are low-K tholeiitic pillowed basalt flow sequences, deep-sea pelagic sediments, mafic tuffs, and ribbon chert. Most ultramafic plutonic material at middle and upper levels is pyroxenitic rather than peridotitic and invariably comprises intrusive bodies, not tectonites.

Although stratified mafic-ultramafic plutonic-volcanic sequences such as those noted above (see P. Anderson, 1986 for details) may contain individual rock types just like those in modern ophiolites (Moores, 1976), the entire Proterozoic oceanic crustal package, considered as a whole or as a collective rock assemblage at any specific crustal level, is certainly not equivalent structurally, compositionally or chemically either to modern ophiolites or to Archean ultramafic sequences (P. Anderson, 1986). Instead, the oceanic package is the Proterozoic tectonic analog of these mafic-ultramafic crusts.

Similar rocks in the Northwest Gneiss Belt also imply that an oceanic basement much like that in the Prescott-Jerome area once underlay parts of the northwest region. Gabbro-diorite-diabase subvolcanic to tholeiitic basalt flow sequences in the Antler-Valentine volcanic belt (More, 1980; Stensrud and More, 1980), large metapyroxenite enclaves in granodiorite of the northern Hualapai Mountains, and pyroxenite enclaves northeast of Kingman (P. Anderson, 1986) are fragments of mafic-ultramafic oceanic substratum from under the volcanic belts.

The basement under much of the younger felsic portion of the Central Volcanic Belt is very different. Ultramafic rocks are rare and the most mafic inclusions in dioritic plutons are usually gabbro. The sequence low-K Mg-rich tholeiite-diabase-microgabbro capped by deep-sea pelagic sediments and chert that occurs in the Prescott-Jerome belts is typically not found; instead the mafic sequences are mainly of basaltic andesite composition. The Gibson Creek complex near Payson has abundant pyroxenitic inclusions and sheeted diabasic dikes and may locally form a mafic basement to the younger eastern part of the Central Volcanic Belt. In general, mafic basement to the younger belts is more evolved in composition and chemistry than that of the older belts.

The Southeast Schist Belt has few mafic volcanic rocks and even fewer mafic plutons. The most mafic inclusions in diorites intruding Pinal Schist are gabbroic, and high-K tholeiitic basaltic andesites in the Dos Cabezas volcanic belt to the south are underlain by a local diorite-gabbro substratum, not mafic-ultramafic rock. However, throughout the vast expanses of typical Pinal Schist (meta-quartz wacke) that occupy the bulk of the Southeast Schist Belt, no evidence for any type of basement—mafic or otherwise—is known.

In summary, the Proterozoic of Arizona lacks evidence for any Archean basement to either the Proterozoic volcanic belts or their intervening crustal segments and lacks evidence for granitoid basement of early Proterozoic age. Only arkosic rocks in far northwest Arizona suggest deposition on transitional crust at the edge of the Archean craton in a continental margin shelf-slope setting. Throughout the rest of northern and central Arizona (i. e., between Kingman and Globe, fig. 1), Proterozoic oceanic crust of mafic-ultramafic and mafic composition and close in age to the volcanic belts themselves formed the basement. Evidence from the Prescott-Jerome, Bagdad, and Antler-Valentine volcanic belts suggests that north-central Arizona was underlain by a gabbro-pyroxenite oceanic crust, whereas the younger eastern part of the Central Volcanic Belt and Dos Cabezas belt were underlain by a more chemically evolved gabbroic oceanic basement. In contrast,

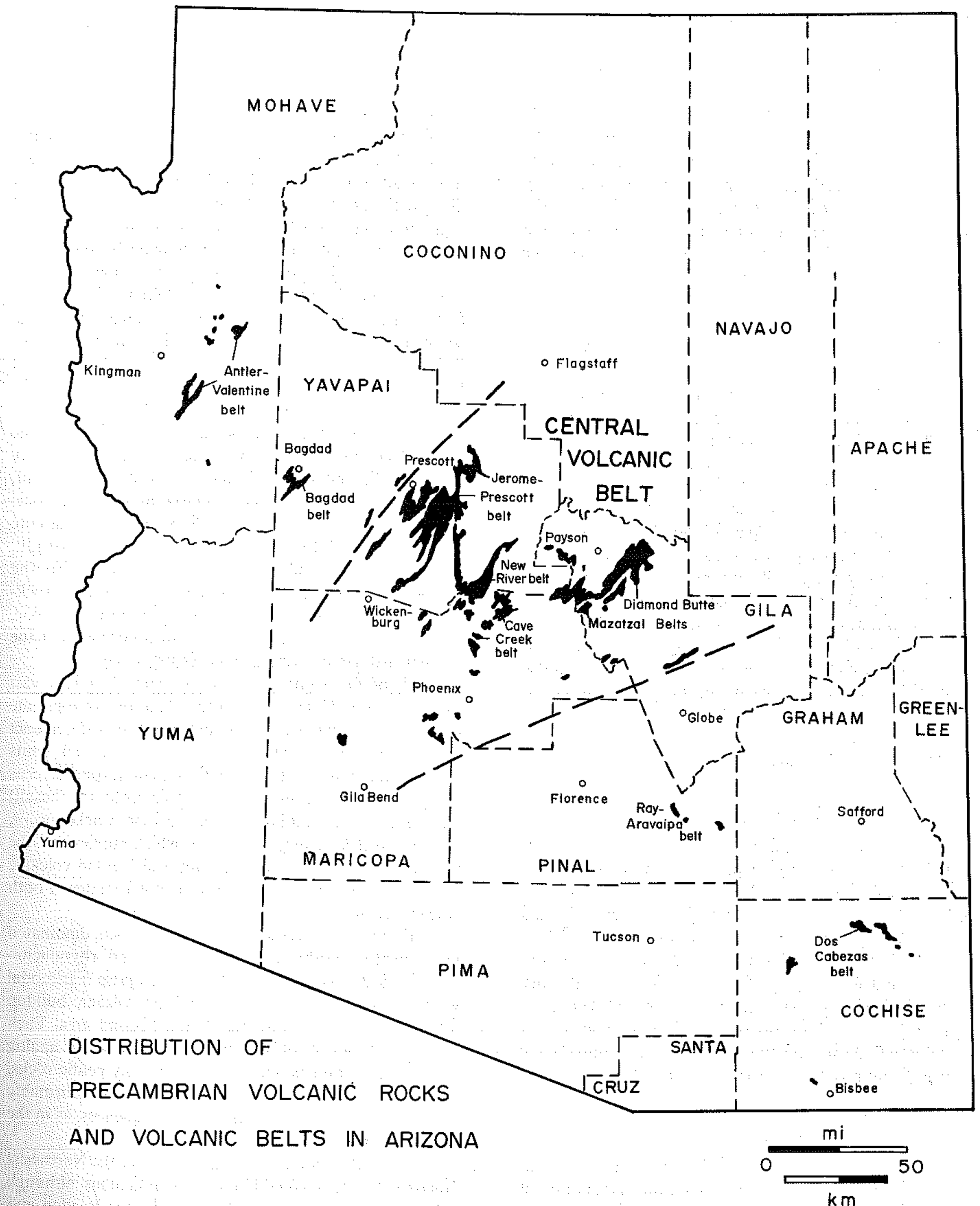

Figure 2. Distribution of Proterozoic volcanic rocks in Arizona in relation to the three major crustal belts described in this paper (only the Central Volcanic Belt is labeled).

much of the Southeast Schist Belt (Pinal Schist between Globe and Willcox, fig. 1) lacks evidence for any type of basement, suggesting a tectonic setting very different from the volcanic belts.

Consequently, it is surmised that the Proterozoic volcanic belts of Arizona formed primarily in open-ocean environments upon Proterozoic oceanic crust, in tectonic settings largely unrelated to preexisting crustal nuclei (P. Anderson, 1976). Such oceanic settings preclude all intracratonic rift environments, mobile-belt settings, or other intracontinental settings that have been commonly inferred for Proterozoic regions elsewhere in the world.

GEOLOGIC OVERVIEW OF TECTONIC BELTS

Each major Proterozoic tectonic belt of Arizona contains several crustal segments with differing lithologic, stratigraphic, petrologic, and geochemical features, so the early formative tectonic evolution of each belt is summarized here by comparing and contrasting these features between segments. Each major belt was thickened and stabilized by its own orogenic event that accreted it to the Wyoming Archean craton between 1750 and 1650 Ma to sequentially build up the diversely structured Proterozoic crust of Arizona. This sequence of events requires a particular tectonic history for Arizona's Proterozoic crustal evolution, and from that history, a plate tectonic evolution is deduced that describes the opening and closing of a Proterozoic ocean basin, accretion of oceanic and arc elements to the Proterozoic continental margin, and suturing of the entire collage during final continental accretion. This is the actualistic analog of the Wilson Cycle (Dewey and Spall, 1975) as it operated during the Proterozoic era.

NORTHWEST GNEISS BELT

Protoliths of Proterozoic rocks in the Northwest Gneiss Belt are more difficult to decipher than those in the other tectonic belts because of strong (typically amphibolite-facies) metamorphic overprints. The oldest stratified sequences commonly occur in migmatite complexes and are preserved only as thin, fragmentary metamorphic screens squeezed between huge granodiorite-monzogranite batholiths. Thus, original depositional geometries of the rock sequences are highly distorted, and many original contact relations have been obliterated. Despite these problems, the original protoliths of most rocks can be deduced.

Oldest Crustal Components

The oldest crustal components of the Northwest Gneiss Belt are all of sedimentary and volcanic derivation, except for minor mafic-ultramafic plutonic fragments that do not belong to the stratified sequences and may represent original mafic basement, as noted above. The proportion of sedimentary rocks far outweighs that of volcanic rocks in the belt, in contrast to the Central Volcanic Belt, where the opposite is true. Based on distribution of original lithologies, the belt can be divided into five segments, as described below, and from the different crustal structures and basement characters of adjacent segments, a tectonic history for the Northwest Gneiss Belt is deduced.

Overview of the Five Crustal Segments. Major variability in the dominant lithologic character of the original supracrustal succession is the critical factor that subdivides the Northwest Gneiss Belt into five different crustal segments. The most northerly segment, which lies northwest of Kingman (figs. 1 and 3), involves well-bedded arkosic and arenitic protoliths that once composed a well-stratified clastic succession of K-feldspathic arkose, feldspathic arenite, quartz arenite, siltstone, and local conglomerate, derived in large part from granitoid detritus. This is the only significant portion of the Arizona Proterozoic crust where true arkosic protoliths (i. e., K-feldspar derived from an earlier granitic source) exist.

The next lithologic belt southeast of Kingman—the Antler-Valentine volcanic belt (figs. 1 and 3)—is very different and contains a volcanic core of mafic flows, tuff, wacke, pelite, subgraywacke, quartz arenite, and local volcanic agglomerate protoliths. On both sides of this core are metasedimentary rocks of volcaniclastic origin, including pelite and volcanic conglomerate. To the northeast near Valentine is a younger suite of volcanic rock-fragment conglomerate, lithic wacke, hematitic siltstone, and purple slates identical to Texas Gulch Formation in the Prescott belt (P. Anderson, 1986; Beard, 1985).

South of the Antler-Valentine volcanic belt is a vast expanse occupied by huge zoned granodiorite-monzogranite batholiths with many different phases and complex contact relationships. Little evidence of the oldest rock units remains in this region except in thin metamorphic screens along batholith margins. Both volcanic protoliths (basalt) and nonvolcanic protoliths (pelite, siltstone, wacke) occur, and volcaniclastics may be present, which implies that the region between the Antler-Valentine and Bagdad volcanic belts was a basin receiving clastics distally from one or both volcanic centers.

Farther southeast is the Bagdad volcanic belt, dominated by tholeiitic basalts, dacitic breccias, and rhyodacite-rhyolite flows and crystal tuffs that comprise a trimodal volcanic suite. The belt signifies a brief volcanic buildup that was apparently restricted to the Bagdad area. Each volcanic component of the belt has its subvolcanic analog equivalent in composition and time. Mainly pelitic rocks, now metamorphosed as Hillside mica schist, unconformably overlie the formative Bagdad volcanic sequences, as in the Antler-Valentine volcanic belt.

The southeasternmost crustal segment of the Northwest Gneiss Belt is distinctly different and anomalous: although it lies between two mafic volcanic belts (the Bagdad and Prescott belts), no mafic or related volcaniclastic rocks have been found in it; instead, only Al-rich felsic metapelitic protoliths, now muscovite and muscovite-andalusite schists

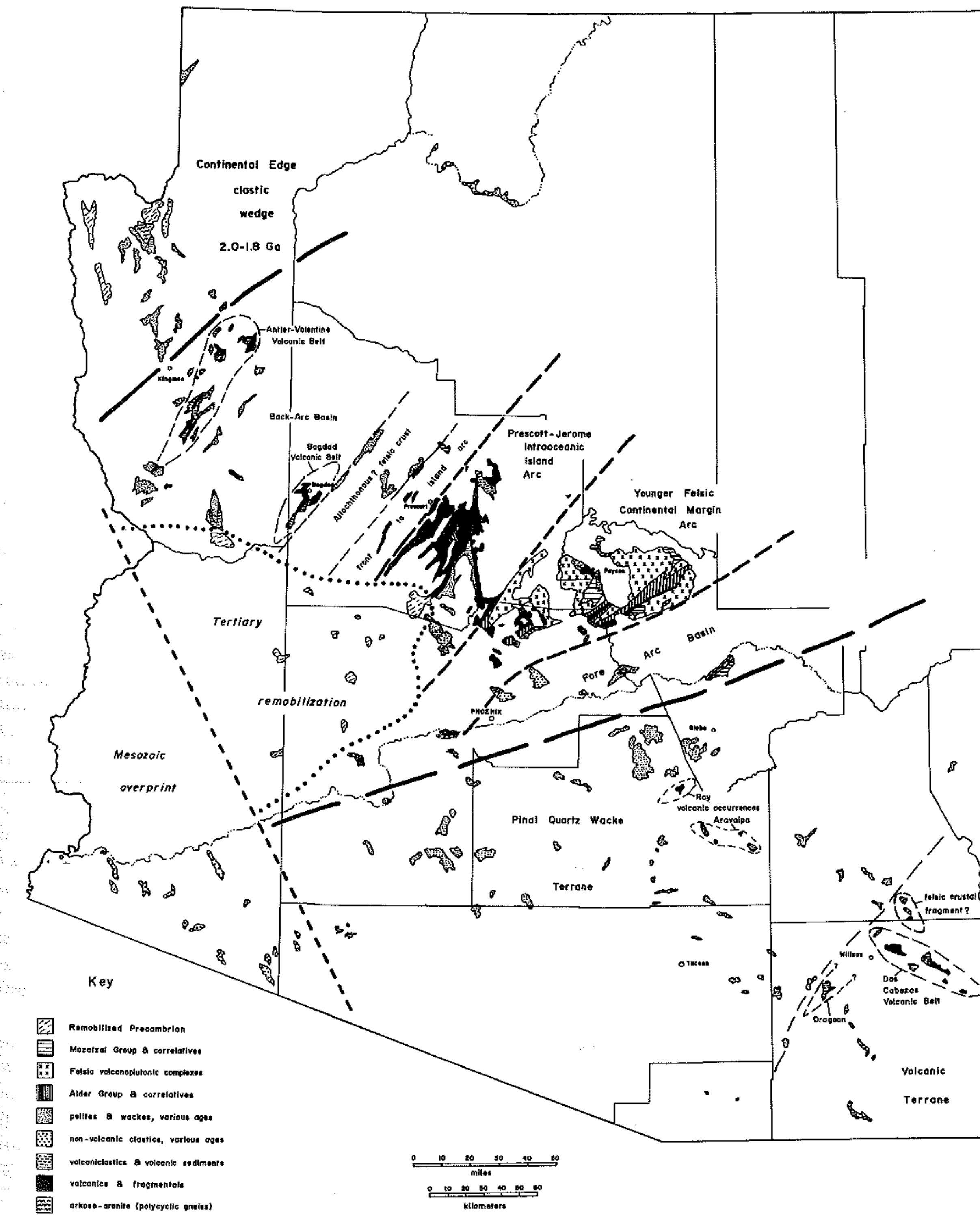

Figure 3. Geologic map of major tectonic belts that make up Arizona's Proterozoic crust, in relation to towns and county boundaries. Mountain range names cited in the text are omitted for simplicity. All data is new and from P. Anderson (1986) and revises earlier geologic maps of Arizona. The two heavy long-dashed lines are boundaries between crusts of profoundly different compositions and origins: the basement in the central region is oceanic, but unknown on either side. The intermediate long-dashed lines subdivide the Central Volcanic Belt and separate it from adjacent tectonic terranes, whereas the thin long-dashed lines delimit tectonic boundaries in the Northwest Gneiss and Southeast Schist Belts. The dotted line shows the approximate limit of extensive reworking of Proterozoic crust by Tertiary tectonic and metamorphic overprinting. Other Tertiary remobilized areas are omitted for simplicity. West of the short-dashed line, the Proterozoic crust has been extensively overprinted by Mesozoic plutonic, metamorphic, and tectonic events.

invaded by potassic granite and migmatite, are found. The pelitic metasedimentary rocks of this segment may be either in tectonic contact with volcanics northwest of Prescott (P. Anderson, this volume), or else unconformable on them and equal in age and stratigraphic position to Hillside mica schist in the Bagdad belt or Texas Gulch Formation in the Antler-Valentine and Prescott volcanic belts. This latter possibility does not explain the absence of mafic volcanic rocks and lack of evidence for mafic basement in the segment, so it is concluded that this felsic crustal segment originated and evolved in a tectonic setting different from either volcanic belt.

Two Contrasting Metasedimentary Sequences. The crustal segment northwest of Kingman contains contrasting terranes that represent two originally different stratified sedimentary rock sequences (fig. 3). One terrane includes polycyclic (i. e., polydeformed) metasedimentary gneiss with ubiquitous refolded folds, but contains foliated basalt-diabase dikes that experienced only one deformational event. The protoliths of the polycyclic gneisses were bedded arkoses, arenites, wackes, pelites, and impure quartzites and siltstones; their arkosic granitoid detrital component was almost certainly derived from part of the Archean Wyoming craton to the north.

This early polycyclic arkosic arenite terrane occupies only part of the northwesternmost crustal segment. In the same region is a younger terrane of mostly quartz wacke and pelite protoliths showing evidence of only one event of deformation, as do all other Proterozoic rocks in Arizona. Also, the deformed basic dikes are absent from the younger terrane. These relationships indicate that the older arkosic terrane was deformed, metamorphosed, and invaded by basic dikes prior to deposition of the younger quartz wacke-pelite sequence, probably unconformably on the arkoses; then both sequences were deformed and metamorphosed during major 1720-Ma Proterozoic orogeny of the Northwest Gneiss Belt. Exact stratigraphic relationships have not yet been determined because the two sequences exist in separate mountain ranges (fig. 3). However, the younger wacke-pelite sequence was most likely unconformable on the older terrane, the unconformity signifying events of deformation, uplift, and erosion.

The younger, metasedimentary gneiss sequence of immature wacke-pelite protoliths seems to grade southward through facies changes into gneisses of pelitic, wacke, and volcaniclastic protoliths along the northwest edge of the Antler-Valentine volcanic belt, the proportion of volcanic material increasing toward the volcanic belt. Unfortunately, original stratigraphic continuity is eradicated in many places by metamorphism, migmatization, and deformation. As the Antler-Valentine volcanic belt evolved, coarse volcaniclastic detritus was deposited along the flanks of the volcanic chain, and only fine material spread far from the volcanic chain. The fine-grained detritus was evidently deposited in a shallow submarine basin to the north, thus causing distal volcaniclastic facies to lap over the older polycyclic arkosic arenite terrane to the north.

In contrast, the depositional basin for similar volcaniclastics was much deeper to the southeast between the Antler-Valentine and Bagdad volcanic belts and locally included basic volcanics. Such metamorphosed mafic volcanics occur with metasedimentary gneisses in the southern Hualapai Mountains near Groom Peak, but the broader region between Wikieup and the McCracken Mountains involves mainly fine-grained wacke, siltstone, pelite, and local argillaceous quartzite protoliths. These relatively silicic (psammitic) metasedimentary protoliths contrast markedly to the argillaceous (pelitic) protoliths that make up the anomalous felsic crustal segment to the southeast, where rock sequences correlative or distally related to adjacent volcanic belts have not yet been found.

Stratigraphic Evolution of the Volcanic Belts. Of the oldest supracrustal components, volcanic belts retain the most information because of their least regenerated states. The Antler-Valentine belt contains (1) a basal suite of microgabbro-diabase, gabbro, and tholeiitic massive basalt in the Antler Mine area (the "bulge" of More, 1980; Stensrud and More, 1980) and (2) supracrustal sequences of basalt flows and tuffs, andesite-dacite tuffs, dacite to rhyolite fragmentals and tuffs, volcanic conglomerates, and other volcaniclastics in both the Antler and Valentine areas (P. Anderson, 1986). A distinctive suite of felsic tuff and wacke, now muscovite-andalusite-cordierite schists, unconformably overlies volcanic sequences along the axis of the volcanic chain, but changes facies laterally to pelitic metasedimentary rocks in adjacent parts of the Hualapai Mountains and Cottonwood Cliffs (fig. 3; P. Anderson, 1986). The ascending stratigraphic succession is thus mafic volcanics, felsics and fragmentals, then youngest felsic tuffs and clastics, except at edges of the volcanic chain where graywacke may be facies equivalent to mafic volcanics. This same overall stratigraphic sequence exists in all Proterozoic volcanic belts of Arizona, even though the belts differ in age.

The Valentine area contains a younger sequence of volcanic-rock-fragment conglomerate, lithic tuff, wacke, purple slate, and quartzite that unconformably overlies the formative volcanic sequences described above, and is equivalent in lithology, stratigraphic position and probably age to Texas Gulch Formation of the Prescott-Jerome volcanic belt (P. Anderson, this volume).

The Bagdad volcanic belt shows an analogous crustal succession of mafic and felsic plutonic and hypabyssal bodies beneath a supracrustal sequence of tholeiitic basalt flows and tuff, dacite flows and breccia, quartz-feldspar rhyodacite flows and crystal tuff, and tuffaceous sedimentary rocks. This sequence is unconformably overlain by volcanogenic sedimentary rocks (quartz wacke, lithic

wacke, volcanic siltstone, and pelite) metamorphosed as Hillside mica schist (Anderson, Scholz, and Strobell, 1955). The uppermost rhyodacitic tuffs are dated at 1740 Ma (Silver, 1966), an age consistent with a 1727-Ma Pb-Pb model age from the Bruce massive sulfide orebody (Clayton and Baker, 1973). A Pb-Pb model age for galena in the Antler mine is 1850 Ma (Stensrud and More, 1980), which implies that the Antler-Valentine belt probably began evolving before the Bagdad volcanic belt; no dates yet published refute this probability.

Plutonic-hypabyssal units emplaced at deeper levels in the Bagdad belt include: (1) a large altered granite-rhyolite body that fed felsic extrusives, (2) subvolcanic gabbro and gabbro-diorite bodies that fed mafic pillowed flows, and (3) a cordierite-anthophyllite altered dacitic center that was a source for bedded dacite fragmentals. At deep levels in the volcanic belt, the **Mulholland Basin gabbro-anorthosite** was emplaced as a stratified lopolith, with gabbro-pyroxenite, Fe-rich gabbro, leucogabbro, and anorthosite layers reflecting its Fe-rich tholeiitic differentiation trend. Beneath the gabbro-anorthosite, a noritic plutonic substratum to the Bagdad volcanic belt is intruded by younger porphyritic granodiorite of the Aquarius batholith.

Petrologic and Chemical Variations. Detailed chemistry shows that both the Antler-Valentine and Bagdad belts are distinctly trimodal (P. Anderson, 1976, 1986), but parts of the Bagdad belt are bimodal tholeiitic basalt-rhyodacite. Omitting sampling the important dacitic fragmental component of most Arizona Proterozoic volcanic belts leads to the erroneous conclusion that they are all bimodal (Condie, 1982). Both trimodal suites are low-K tholeiitic, with Na but little K enrichment, as are all volcanic belts in northern and north-central Arizona. The chemistry of mafic volcanics and gabbro-anorthosite suggests that the Bagdad belt and also probably the Antler-Valentine belt followed Fe-rich tholeiitic differentiation trends equal to stage 2 in the Prescott-Jerome belt's evolution (P. Anderson, this volume). Both stage 1 volcanics of the Prescott belt (primitive Mg-rich bimodal tholeiites) and later stages (voluminous felsic fragmentals from evolved stage 4 magmas) are lacking in the less evolved northern volcanic belts. Also, calc-alkaline and alkali-calcic volcanic sequences that dominate the younger eastern parts of the Central Volcanic Belt are absent in the northern belts.

A transect across the five crustal segments of the Northwest Gneiss Belt encounters vastly different crustal compositions and petrologic features in each region: (1) The oldest polydeformed metasediments of the northernmost belt are rich in K and Si, perhaps more so than any other crust in Arizona. (2) Immature pelitic metasediments that overlie this arkosic terrane are richer in Na and Al, and become richer in Ca, Fe and Na towards the Antler-Valentine volcanic belt, a chemical trend also reflected by the plutonic suites emplaced into the two different crustal segments. (3) Metasedimentary rocks rich in Na, Fe, Ca and Al dominate the basin between the Antler-Valentine and Bagdad volcanic belts. (4) Trimodal volcanics of the Bagdad belt are rich in Ca, Fe, and Na, like the Antler-Valentine belt, but may be richer in Mg and Al than the Antler-Valentine belt. (5) In contrast, the southernmost metapelitic segment is a K-rich crust lying between two volcanic belts that are very deficient in K. Such sharp compositional changes from region to region imply markedly different tectonic histories and origins for each crustal segment.

Plutonic Rocks

The earliest supracrustal components in the Northwest Gneiss Belt are of sedimentary and volcanic (or subvolcanic) derivation and all formed prior to 1730-Ma inception of major plutonism. The earliest pretectonic plutons include diorite, granodiorite, and local gabbro bodies that were emplaced into and along the edges of the volcanic belts soon after the close of volcanic activity, at about 1740 Ma, but prior to deposition of the slate-conglomerate sequence in the Antler-Valentine belt that correlates to ca. 1720-Ma Texas Gulch Formation in the Prescott-Jerome belt. Culminating at about 1720 Ma but continuing to 1710 Ma and perhaps 1700 Ma, huge zoned granodiorite and monzogranite batholiths were emplaced syntectonically to produce deformed margins and undeformed cores commonly pervaded by autogenous dike swarms. Coarsely porphyritic monzogranite in such batholiths is virtually indistinguishable from 1400-Ma granites in all but its textural and structural aspects. The 1720-Ma batholithic invasion of the Northwest Gneiss Belt was so profound that few early supracrustal deposits except the more coherent volcanic cores remained unaffected by their intrusion.

Pretectonic Granodiorites. The earliest pretectonic, deformed granodiorites occur in and along the margins of the Bagdad and Antler-Valentine volcanic belts, as they do in the Prescott-Jerome belt, but some in the northern Cerbat and southern Hualapai Mountains lie far from volcanic belts. In the northern Hualapai Mountains, the ca. 1750-Ma gneissic hornblende-biotite **Hualapai granodiorite** (Kessler, 1976; P. Anderson, 1986) borders the Antler belt on the north and the Valentine belt on the south (fig. 3). Similar pretectonic gneissic biotite granodiorite in the Bagdad belt intrudes 1740-Ma Hillside mica schist but is cut by 1720-Ma porphyritic monzogranite intruding the schist (both dates from Silver, 1966). Thus, deformed granodiorites in both belts are comparable in age and tectonic setting and predate major 1720-Ma deformation of the Northwest Gneiss Belt.

Pre- and Syntectonic Batholiths. Feldspar-porphyritic batholiths of zoned granodiorite-monzogranite-granite pervaded the Northwest Gneiss Belt just prior to and during its major orogeny. Such bodies include the huge Garnet

Mountain-Music Mountain batholith in the north, the Valentine body, the huge Aquarius batholith in the Aquarius Mountains, the Burch Peak batholith in the southern Hualapai Mountains, and many others in the Northwest Gneiss Belt. Most are characterized by K-feldspar-porphyritic monzogranite and granite core phases, local autogenous pegmatite dike swarms, strongly deformed feldspar-porphyritic granodiorite border phases, and wide contact metamorphic-migmatitic zones that include fine-grained granite dikes migmatizing host metasedimentary and metavolcanic strata.

Few such batholiths have been studied in detail, but available dates (Silver, 1966, 1967) indicate that emplacement ages cluster at about 1720 Ma. In contrast to the calcic chemistry of the pretectonic granodiorites, the syntectonic batholiths are calc-alkaline with locally high alkalis, but are not peralkaline. They are chemically similar to 1400-Ma porphyritic monzogranites throughout Arizona that define an alkali-rich, calc-alkaline magma series. Most 1720-Ma porphyritic batholiths in the Northwest Gneiss Belt appear to be products of lower crustal anatexis: the magmas started crystallizing in the feldspar field and underwent only limited fractionation (P. Anderson, 1986).

Felsic Plutons. Parts of the Northwest Gneiss Belt with abundant supracrustal pelitic strata underwent partial fusion during the thermal peak of regional metamorphism to produce migmatitic complexes pervaded by red granite-leucogranite anatectites. Such two-mica S-granites (muscovite > biotite due to felsic source sediments) are widespread along the edges of the Antler volcanic belt where felsic pelitic strata underwent regional anatexis, and in migmatites of the Cerbat Mountains. Because the S-granites formed during the peak of thermal metamorphism, they may either predate or postdate 1720-Ma porphyritic monzogranite batholiths in any area depending on timing of batholith emplacement and regional metamorphism. These felsic plutons are undated by U-Pb zircon methods, but are expected to cluster at 1710-1700 Ma, with some as old as 1730 Ma. Rb-Sr dates on S-granites near the Hualapai Peak granite are reset to its 1400-Ma age (Kessler, 1976).

Deformation and Metamorphism

Polycyclic metasedimentary gneisses in the northernmost part of the Northwest Gneiss Belt (fig. 3) are structurally and generically distinct from all other Proterozoic strata in Arizona because they show clear evidence of two distinctly different events of deformation. In central Arizona where polyphase folding has been hypothesized, divergent fold axes are usually coplanar and consistent with progressive strain in a single event of deformation (P. Anderson, this volume). Polycyclic gneisses in the Northwest Gneiss Belt, on the other hand, show clear evidence for superimposed regional folding: kinematic axes of the two fold events are very different, and ubiquitous mineral lineations parallel to early fold axes are deformed by the later folds.

Thus, evidence exists in far northwestern Arizona for a deformational and metamorphic event predating the one that the rest of Arizona's Proterozoic crust experienced. Also it appears that events of diabase diking, probable uplift and unroofing, and possibly an event of plutonism are all recorded in the polycyclic gneisses of northwest Arizona. These events predated formation of all Proterozoic crust to the southeast, and therefore significantly predate 1850 Ma, and may be as old as 2200 Ma (P. Anderson, 1986).

The younger Proterozoic deformational and metamorphic event (the main event that affected all rocks in the Northwest Gneiss Belt) postdates both deposition of supracrustal strata and emplacement of pretectonic deformed biotite-hornblende granodiorites. The 1720-Ma porphyritic batholiths are not strictly posttectonic bodies, even though parts appear undeformed, and do not provide an exact younger limit on this main event. The Aquarius batholith is typical: the porphyritic granodiorite core in the Aquarius Mountains appears undeformed, yet border phases in the Hualapai Mountains are strongly lineated gneisses. The batholithic bodies have strongly deformed migmatitic margins, well-foliated border phases, and weakly undeformed cores.

Such relationships show that the zoned 1720-Ma porphyritic batholiths were primarily syntectonic, because it was their emplacement into the crust and the sheer volume of their magmatic event that caused major metamorphism and deformation of the Northwest Gneiss Belt. The larger batholiths are 40 km or more in all dimensions, hence they occupied most of the thickness of the crust, so emplacement levels vary from deep to shallow with position in the batholith. The batholiths were initially emplaced into a static, nonmetamorphic setting, but the first-crystallized border phases were deformed and metamorphosed after final batholith crystallization by the peak of deformation and metamorphism that was ultimately caused by formation and emplacement of the batholiths themselves. Thus, the porphyritic 1720-Ma batholiths are both pre- and syntectonic.

The major deformational and metamorphic events of the Northwest Gneiss Belt so closely followed batholith emplacement across the belt that their isotopic ages largely coincided with batholith emplacement ages at 1720 Ma. Although deformation may have precisely matched batholith emplacement in places, the thermal metamorphic peak likely followed batholith emplacement in most places, as the metamorphic isograds rose to higher crustal levels with continued crustal heating. Thus earlier penetrative foliation was sequentially overprinted in most places with new fabrics of metamorphic recrystallization as the plutonic event swept from southeast to northwest across the entire belt.

The style of regional metamorphism in the Northwest Gneiss Belt is high-T, low-P andalusite-cordierite facies series, and the architecture of isothermal surfaces was largely governed by the array and relative emplacement levels of batholiths. With the present crustal structure (P. Anderson, 1978b) staurolite should occur at deep crustal levels in the Mohave and Hualapai Mountains but to date it has been found only in migmatites of the Hualapai Mountains. Andalusite, staurolite, and cordierite all exist in the Hualapai Mountains, which suggests a P-T regime in the range 500°-700°C and 3-5 kb.

Comparable metamorphic conditions were found in Vishnu Schist of the Grand Canyon (Brown, Babcock, and Clark, 1974). The Grand Canyon exposures represent just a small part of one of the many high-grade migmatite complexes that exist throughout the northwest part of the gneiss belt. Compared to the entire Northwest Gneiss Belt's vastness, diversity of terranes, and multitude of tectonic relationships, Vishnu Schist exposures in the Grand Canyon are neither typical nor definitive, and have received inordinately more attention than their overall tectonic importance warrants.

Tectonic Setting and Evolution

As noted previously, south of a northeast-trending line through Kingman (fig. 3), mafic-ultramafic fragments occur in batholiths and along the edges of the volcanic belts. This indicates that the southern part of the Northwest Gneiss Belt, from the north end of the Antler-Valentine volcanic belt to the south end of the Bagdad volcanic belt, was floored by mafic-ultramafic oceanic crust. No comparable mafic-ultramafic material has yet been found north of the line, but neither has preexisting granitoid basement. Basement to the oldest polycyclic gneisses northwest of Kingman probably lies deep beneath a thick early Proterozoic clastic wedge that formed along the southern edge of the Archean Wyoming craton, and may not be exposed in the Proterozoic orogen. This basement was likely transitional in character, such as that developed at the edge of a rifted Proterozoic continental margin where Archean granitoid basement interfaces with Proterozoic oceanic crust. The exact location of the basement transition is less important than its plate tectonic significance.

Crustal Structure. With open-ocean conditions throughout the Northwest Gneiss Belt in early Proterozoic time, and a basement of mafic-ultramafic Proterozoic oceanic crust south of Kingman, the Antler-Valentine and Bagdad volcanic belts must have evolved as either small oceanic islands or island arcs in the open Proterozoic ocean basin. This ocean basin shallowed to the northwest toward the Archean craton, where a previously deformed arkosic-arenitic clastic wedge constituted a local basement to volcaniclastic detritus shed distally from the volcanic axis. The basin deepened to the southeast, where clastics from the submerged Antler-Valentine and Bagdad volcanic belts were most likely deposited directly on oceanic crust. The thin supracrustal succession in this deep ocean basin included basic volcanics, tuffs, and pelagic sediments, and was part of a region of thin crust predisposed to later extensive batholithic invasion.

Neither of the volcanic belts became emergent during their volcanic evolution, only after initial plutonism and uplift, possibly prior to deposition of Texas Gulch-like recycled clastics. Formative volcanism built up the volcanic piles sufficiently that shallower water pelitic sediments succeeded volcanics across most parts of the volcanic chains, as well as across earlier deep-water flanking sediments. If the segment of felsic crust southeast of Bagdad is indeed without mafic basement, and if its pelitic sediments were not part of the later shallow-water pelitic sequences in the volcanic belts, then the segment does not belong to the deep-water intraoceanic setting of volcanic belts on oceanic crust; hence it must be exotic to that deep-ocean setting, and therefore allochthonous to the crustal region in which it now lies.

Tectonic History. The five lithologic segments of the Northwest Gneiss Belt, described by dominant lithology above, uniquely correspond to crustal segments with different tectonic histories. The history of the northernmost segment northwest of Kingman goes farther back in time than all other Proterozoic crustal regions of Arizona. It is possible that the oldest polycyclic gneisses in this segment are Archean and were tectonically detached from the Wyoming craton to the north, but this is unlikely because the Archean craton was almost certainly the source for the dominant arkosic character of the gneiss protoliths, particularly in view of the fact that similar arkosic protoliths are found nowhere else in the Arizona Proterozoic. It is also possible that the second deformational event in the gneisses is post-Precambrian, but this is also unlikely, because of the high metamorphic grades accompanying gneissic recrystallization.

Instead, it appears that an initial event of early Proterozoic rifting along the southeast edge of the Wyoming Archean craton established a continental margin shelf-slope clastic wedge that received material eroded from the craton. The southern part of this slope assemblage in northwestern Arizona was evidently deformed once, probably intruded by plutons, and also possibly locally exposed. It was then intruded by basic dikes (perhaps related to mafic volcanism to the south or to rifting preceding volcanism) and then overlain unconformably by pelites and wackes probably related to volcaniclastic detritus shed distally from the Antler-Valentine volcanic belt to the south. Finally, the early clastic sequence, dikes, and later pelite-wacke sequence were all deformed, metamorphosed to amphibolite grades, and intruded by batholiths and plutons during ca. 1720-Ma major regional deformation of the Northwest Gneiss Belt.

This history indicates that the earliest arkosic sediments postdated emergence of the Archean craton to the north and initial rifting of its margin, but predated evolution of mafic volcanic belts and their related clastics to the south. It also possibly indicates a second rifting event (with basic dikes) that preceded formation of the Antler-Valentine volcanic belt. Hence the depositional age of the arkosic sequence lies in the broad range 2300-1800 Ma, but is probably between 2100 and 1900 Ma because of interceding rifting events and broad-scale continental considerations (P. Anderson, 1976). The age of the first deformational event of these rocks predates formation of crust to the south and, therefore, lies in the range 1900 to 1800 Ma.

The age of detrital material in the sediments, however, is much older. If any detrital zircons survived the high-amphibolite metamorphism, their ages are expected to exceed 2.0 Ga and reflect the age of the Archean granitoid source. Because of repeated high-grade metamorphic events, however, most zircons are likely to be hybrid: U-Pb zircon ages near 1700 Ma would reflect the most recent metamorphism, those between 1800 and 1900 Ma the earlier metamorphic event, and pre-1900 Ma ages the detrital source material. Recent Nd-Sm isotopic work (Bennett and DePaolo, 1987) indicates the presence of a 2.0 Ga crustal component in the region, supporting the pre-1900-Ma age assignment (P. Anderson, 1986) for the oldest metasedimentary terrane of the Northwest Gneiss Belt.

The younger, immature wacke-pelite sequence was broadly coeval with volcaniclastics deposited at the edges of the Antler-Valentine volcanic belt, which may be either coeval with, or slightly younger than, formative volcanics in the belt. As the volcanic belt evolved, fine-grained detritus was shed back over the older arkosic terrane to the north, possibly with some reworking in the shallow submarine setting to the north. The deep basin to the southeast between the Antler-Valentine and Bagdad volcanic belts was, however, an interarc basin coeval with evolution of the two volcanic belts.

The brief trimodal tholeiitic basalt-dacite-rhyodacite buildups of the Antler-Valentine and Bagdad volcanic belts were broadly coeval at about 1750 Ma, and close stratigraphic, petrologic, and chemical similarities between the belts suggest they may have been exactly coeval. Pb-Pb isotopic and other geologic data favor slightly older ages for volcanic and plutonic rocks of the Antler-Valentine belt. Also, the tectonic settings of the two belts were importantly different: evidence suggests that the Antler-Valentine belt evolved along the edge of the cratonic clastic wedge first, and the Bagdad belt formed slightly later by backarc spreading in an intraoceanic setting (see later discussion). Their formation between 1750 and 1740 Ma corresponds to later stages in evolution of the Prescott-Jerome volcanic belt, when younger mafic volcanics were extruded along its eastern edge. The primitive, unevolved state of the northern volcanic belts and their lack of subsequent development implies that the Bagdad and Antler-Valentine belts formed in a different tectonic setting than did the well-developed volcanic belts of central Arizona.

However, the later evolutions of all volcanic belts in northern and central Arizona were demonstrably linked in space and time, after formative volcanism, early arc plutonism, and emergence. The distinctive purple slate-siltstone-conglomerate sequences, whether named Alder Group in the eastern part of the Central Volcanic Belt, Texas Gulch Formation in the northwestern part, or unnamed in the Valentine volcanic belt, are all alike: they all (1) succeeded formative volcanism and recycled previous volcanic deposits; (2) were restricted to structural troughs caused by differential uplift as the volcanic belts became emergent; (3) are associated with minor calc-alkaline felsic effusive volcanism; and (4) record major turning points in tectonic evolution, when the belts changed from intraoceanic to continent-margin settings.

Stratigraphic continuity between isolated occurrences of these unique Alder-Texas Gulch sequences cannot be proved, because each deposit was areally restricted to structural troughs in its own volcanic belt and never formed a blanket deposit across the entire terrane. These deposits formed only in the volcanic belts of central and northwest Arizona, during the late-stage evolution of these belts, and do not exist in the Southeast Schist Belt; they, therefore, were broadly coeval at about 1725 Ma, although probably diachronous in detail. This relation establishes a critical time link, during and after which all Proterozoic volcanic belts throughout central and northwestern Arizona shared a common stratigraphic, structural and tectonic history. Most importantly, the correlation suggests that all volcanic belts were "in place" by that time, and any major rearrangement or transcurrent dissection of the Proterozoic crust would have occurred before that time.

The anomalous terrane of pelitic metasediments between Bagdad and the west edge of the Prescott volcanic belt is enigmatic in its lack of correlation to either neighboring volcanic belt. It may be either in tectonic contact with adjacent volcanic belts, or else is partly unconformable upon them and thus equal in age and stratigraphic position to the Hillside Mica Schist. If this small belt of apparently felsic crust was originally allochthonous to the edges of one or both volcanic belts, and was subsequently tectonically emplaced, then the above reasoning indicates that this crustal segment was emplaced close to its time of formation, or no later than 1720 Ma ago.

The last major stage in crustal evolution of the Northwest Gneiss Belt involved fusion of lower crustal material and emplacement of huge batholiths into the middle-upper crust. This pervasive batholithic event occurred just after deposition of Texas-Gulch-type sediments and during regional deformation and metamorphism of the Proterozoic crust of the Northwest Gneiss Belt at about 1720 Ma. Deformation predated the earliest calcic granodiorite plutons and was coextensive with batholith emplacement, but the rise of the thermal infrastructure followed batholith crystallization in many places. Consequently, deformational fabrics became overprinted by high-grade metamorphic fabrics as the batholithic event swept from southeast to northwest across the entire Northwest Gneiss Belt.

Clearly, some broad-scale tectonic force was responsible for this major 1720-Ma orogeny of the Northwest Gneiss Belt, a force sufficiently profound to completely regenerate the full 250-km width of Proterozoic crust in northwest Arizona and consolidate it into a thickened, stable Proterozoic craton. The evidence for that force is not to be found in the Northwest Gneiss Belt itself, but lies in the Proterozoic tectonic belts of central and southeast Arizona.

SOUTHEAST SCHIST BELT

Like the Northwest Gneiss Belt, the Southeast Schist Belt consists predominantly of metasedimentary rocks, but unlike it, they are at generally low metamorphic grades, so protoliths can be more confidently interpreted. In some areas, low-grade rocks with penetrative Proterozoic fabrics are converted to high-grade gneisses by Mesozoic and Cenozoic tectonic overprinting. The interpretation of protoliths in these areas is comparable to that in migmatite complexes of the Northwest Gneiss Belt. Unlike the Gneiss Belt, the Southeast Schist Belt is remarkably uniform in composition, lithology, and structure for much of its extent. Volcanic rocks are subordinate, are confined to small, discrete belts, and are similar in age and stratigraphy to the younger eastern part of the Central Volcanic Belt, having formed in the range 1700-1650 Ma.

The youngest eastern volcanic chain of the Central Volcanic Belt did not extend south of the Mazatzal Mountain and Diamond Butte belts (figs. 2 and 3). Instead, a wide basin of turbidite volcaniclastics and graywackes, with local interspersed volcanics, lay to the south in the Hess Canyon, Four Peaks, and north Phoenix areas (P. Anderson, this volume) and extended as far south as a line following the Salt River (fig. 3). This key line defines the southern limit of the Central Volcanic Belt and the beginning of the Southeast Schist Belt (fig. 3). South of the Salt River line lies a vastly different terrane of immature, silicic, micaceous, metasedimentary rocks derived from quartz wacke protoliths; these rocks are known as Pinal Schist and are lithologically distinct from all others in the Arizona Proterozoic belts.

Oldest Crustal Components

The dominant, characteristic protolith of the Southeast Schist Belt is argillaceous quartz wacke with thin alternating silica-and clay-rich layers composing a well-bedded, but very immature, sand-silt-clay accumulation. Much material was apparently of felsic volcanic origin but was deposited far from volcanic centers and reworked in a deep-water submarine, locally turbiditic environment. Many immature deposits show evidence of having been chaotically dumped, slumped, or structurally disordered in this deep-submarine environment, but the more mature silicic deposits are better bedded.

These distinctive protoliths, where metamorphosed to greenschist-grade sericite-quartz-biotite schists, semischists, and phyllites, constitute the distinctive "Pinal Schist" lithotype (Ransome, 1923). Although felsic volcanism was coextensive with Pinal Schist deposition in parts of the Pinal basin (e. g., Ray-Aravaipa area), the larger mafic-felsic volcanic centers (e. g., Dos Cabezas belt) to the south are stratigraphically distinct from Pinal Schist and formed in very different tectonic settings. In the past, the term "Pinal Schist" has been used to refer to formative volcanic strata southeast of Willcox (see line on Fig. 3), but this usage is incorrect and should be discontinued (P. Anderson, 1986).

Pinal Schist extends from its type area of the Pinal Mountains west into Maricopa County, south and west throughout Pinal County, southeast into Graham County, and into northeast Pima and adjacent Cochise County (fig. 3). Although the main protolith is argillaceous quartz wacke, other rock types locally dominate. For example, thinly bedded argillaceous quartzite, wacke, and siltstone compose a more silicic section in the Pinal Mountains, whereas phyllite, sericitic slate, pelite, and tuffaceous metasedimentary rocks dominate the region to the west between Florence Junction, Superior, and Ray. In most other areas, Pinal Schist is remarkably uniform in lithology. Metamorphic grade is locally elevated to low-amphibolite facies near batholiths (e. g., in the Table Top and Javelina Mountains, Pinal Schist has abundant andalusite and cordierite near the Maricopa batholith). At its westernmost (at Ajo) and easternmost (Santa Teresa and Pinaleno Mountains) exposures, Pinal Schist has been overprinted by Mesozoic and Tertiary metamorphism and tectonism, but is still recognizable.

The northern Pinal Schist terrane is a relatively uniform assemblage of quartz wacke, argillaceous siltstone, and pelite (fig. 3), with felsic tuff at widely scattered localities indicating deep submarine working of volcaniclastic debris. However, southeast of a line through Cochise County between Safford and Benson (fig. 3), a major change in the oldest crustal components occurs, and everywhere south of this line, typical Pinal Schist lithologies are not found. Instead, a more heterogeneous assemblage of rocks of direct volcanic derivation, together with distal volcaniclastic units unlike the Pinal Schist lithology, dominate the southeasternmost part of Arizona. The distribution of volcanic rocks in the Southeast Schist Belt clearly shows the change (fig. 3).

Volcanic Centers and Belts

South of argillaceous and pelitic variants of Pinal Schist near Florence Junction and Ray is a small volcanic center of quartz-feldspar-phyric rhyolite flows, tuffs, and highly sericitic phyllites of distal, felsic tuff origin. The proximity of this felsic volcanic center to argillaceous facies of Pinal Schist strongly suggests that the clay-rich Pinal Schist protoliths were derived from fine felsic tuff debris shed into a deep basin peripheral to the Ray volcanic center. This setting contrasted to a shallower water environment in the Pinal Mountains 15 km to the east.

The full extent of the Ray felsic volcanic center is unknown because of cover to the south, west, and east. It may be continuous with a small felsic volcanic center in

Aravaipa Canyon to the south, where quartz-feldspar phyric rhyolite flows or subvolcanic units are similar to those near Ray. Scattered exposures of volcanic and tuffaceous strata to the south suggest that more of the belt lies under cover of San Pedro Valley. Felsic and intermediate tuff and tuffaceous siltstone in the Tortilla Mountains farther south may mark the southern end of the volcanic belt. If both volcanic centers are continuous under cover, a 30- to 50 km-long belt comparable in size to the Dos Cabezas volcanic belt could be represented in the Ray-Aravaipa area (fig. 3).

The largest exposures of volcanic rocks in the Southeast Schist Belt lie in the Dos Cabezas Mountains (figs. 2 and 3). Whereas the Ray-Aravaipa belt is a limited suite of felsic flows and tuff, the Dos Cabezas belt involves a buildup of mafic volcanics, subvolcanic phases, felsic volcanics, and derivative volcaniclastics, all of which make up a belt comparable in size to the Bagdad volcanic belt. In major petrologic and chemical aspects, however, the Dos Cabezas belt is unlike volcanic belts in northern and central Arizona.

The dominant chemical character of the Dos Cabezas belt appears to be bimodal basalt-rhyodacite, but dacites have been found, so the assemblage may prove to be trimodal. Their chemistry, however, sets them apart from the other belts: all bimodal basalt-rhyodacite or trimodal basalt-dacite-rhyodacite suites in central and northwest Arizona are of low-K tholeiitic chemistry, and formative calc-alkaline volcanism is limited to polymodal basaltic andesite-andesite-dacite-rhyolite suites (e. g., the Union Hills Group). The Dos Cabezas belt, on the other hand, has a calc-alkaline (to high-K tholeiitic) chemistry, even though its volcanics comprise a bimodal basalt-rhyodacite or trimodal basalt-dacite-rhyodacite suite (the calc-alkaline rocks are fresh and cannot be rejected as altered). The Dos Cabezas volcanic belt is therefore anomalous because of its primitive bimodal basalt-rhyodacite character and evolved calc-alkaline chemical signature.

Stratigraphic relations in the Dos Cabezas volcanic belt resemble those in the younger parts of the Central Volcanic Belt. The oldest volcanics are pyroxenitic basalts and are underlain by massive pyroxene-crystal microgabbro and basalt, which are remnants of the source magma chamber that fed overlying mafic flows. Extrusive dacites lie adjacent to and interfinger with the mafic flows and tuffs, and their subvolcanic feeders intrude the earliest mafic units. Overlying and locally interstratified with these units are quartz-feldspar-phyric rhyodacite bodies and quartz-crystal rhyodacitic tuff interlayered with finer grained felsic flows and tuffs. The upper portions are interstratified with, and perhaps locally unconformably overlain by, tuffaceous siltstones and wackes that become more predominant away from the axis of the volcanic chain.

In the southern Dos Cabezas Mountains, the tuffaceous siltstones and wackes are unconformably overlain by a sequence of white quartzite, siltstone, and shale with abundant internal folding. These quartzites have features in common with both Alder quartzites of central Arizona and distal facies of the Mazatzal Group (e. g., Hess Canyon area). Present evidence favors the likelihood that they are broadly time correlative to the Mazatzal Group, without implying stratigraphic continuity with Mazatzal strata in central Arizona.

Evolution of the Dos Cabezas belt in many ways paralleled that of the younger eastern part of the Central Volcanic Belt. The oldest mafic volcanics are comparable in stratigraphic position to the Union Hills Group in the Cave Creek and Diamond Butte belts, but Dos Cabezas basalts are chemically less evolved than Union Hills basaltic andesites. Felsic volcanics are similar in petrology to ignimbrites of the Hess Canyon, Aravaipa, and Diamond Butte areas but are subaqueous crystal tuffs in the Dos Cabezas, not subaerial ignimbrites. Finally, Dos Cabezas white quartzites appear broadly time correlative to the Mazatzal Group, but in detail will likely be found to be time transgressive from Mazatzal strata and never continuous with them. Such diachroneity is supported by a 1685-Ma age on rhyodacites in the Dos Cabezas and Johnny Lyon Hills (Silver, 1967, 1976, 1978), a slightly younger age than lithologically similar quartz-crystal rhyolitic ash flows and tuffs in central Arizona with 1705-1695-Ma ages. With quartzites immediately following on from felsic volcanism in both regions, quartzites in the Dos Cabezas belt likely postdate their analogs in central Arizona and therefore would not be of Alder age.

These relationships establish a crucial stage in evolution of the Southeast Schist Belt when the Dos Cabezas volcanic belt was stratigraphically linked, *in a broad sense not implying direct continuity of strata,* to the younger felsic volcanic belts of central Arizona; both volcanic terranes were overlapped by mature shallow-water marine and littoral sands soon after similar events of felsic volcanism ceased in each area. This situation is exactly analogous to Texas Gulch-Valentine successor clastics in central and northwest Arizona where direct stratigraphic continuity between isolated localities never existed, but where the sedimentary event marks the latest time at which all the tectonic elements of the terrane were "in place." Similarly, the Southeast Schist Belt from central Arizona to at least as far south as the Dos Cabezas belt must have constituted one contiguous crustal terrane soon after 1685 Ma.

Other Crustal Components

Volcanic and clastic rocks in the Little Dragoon Mountains to the west differ structurally from those in the Dos Cabezas belt and provide valuable constraints on the pre-1685-Ma tectonic history of the Southeast Schist Belt. In the eastern Little Dragoon Mountains the mafic volcanic sequence is a chaotic array of basaltic flows, coarse breccias, agglomerates, and tuffs, in which huge blocks of volcanic rock with various bedding orientations lie in a matrix of mafic volcanic and volcaniclastic material. The chaotic structure predates deformation, because a steeply dipping penetrative foliation is superimposed on the chaotic fabric. Even overlying mafic tuffs and iron formation that are in turn overlain by turbidite graywackes to the west are chaotically dissected.

In the western Little Dragoon Mountains, the metamorphosed graywacke-siltstone-shale suite is a classic turbidite sequence deposited in a deep-water submarine environment (Silver, 1954) and is analogous to turbidite sequences flanking younger parts of the Central Volcanic Belt (P. Anderson, 1986). Thus, if the eastern Little Dragoon volcanics form the west edge of the Dos Cabezas belt (fig. 3), the turbidite graywacke suite represents a volcaniclastic apron to the edge of the volcanic belt. If so, both the west edge of the volcanic belt and its graywacke apron were chaotically dissected and structurally imbricated prior to regional deformation of the Southeast Schist Belt.

Further tectonic complexity exists in the southern Pinaleno Mountains, where schists appear to be derived from quartz- and K-feldspar-rich arkosic and conglomeratic protoliths. The schists contain small granitoid pebbles with quartzofeldspathic intergrowths that are more granophyric than granitic. However, the exact origin of the pebbles is obscured by strong Phanerozoic metamorphic recrystallization and tectonic fabrics that affect the Pinaleno Mountains. The pebbles were most likely derived locally from erosion of quartz-feldspar phyric rhyolites and granophyres in the Dos Cabezas belt less than 20 km away. If so, protoliths of the feldspathic schists correlate to other Mazatzal-age rocks and signify only local conglomeratic facies of quartz arenite deposits. If, however, the conglomerates predate felsic volcanism in the Dos Cabezas belt, they are clear evidence for an old felsic volcanic or granitic source terrane somewhere in the vicinity. If such an older granitoid terrane did exist, it is not now exposed or recognized in southeast Arizona, it must have been originally exotic to the Proterozoic oceanic crust that occupies the rest of Arizona, and it was therefore accreted to that oceanic crust by tectonic processes. The most southerly Proterozoic rocks in Arizona lie in the Mule Mountains near Bisbee. The protoliths of the pale-green chlorite-sericite schists were volcaniclastics, sediments, and tuffs of andesitic to dacitic composition, but were mostly reworked. Such rocks were presumably derived from a coeval or older andesitic volcanic center, which could be the Dos Cabezas belt to the north, except that andesites are not present in that belt. Thus, another Proterozoic volcanic belt of andesitic composition may lie farther south in Sonora, Mexico, but its existence and tectonic significance have hitherto not been recognized.

Plutonic Rocks

The Southeast Schist Belt is notable in its relative paucity of orogenic plutons compared to other Arizona Proterozoic belts. Representatives of the oldest plutonic suite were first named Madera Diorite (Ransome, 1919), and most early mafic pretectonic plutons intruding Pinal Schist are so similar that they can be termed Madera-type diorite. The bodies are most abundant between Ray, Globe, and Superior, but also occur to the north, west, and southeast, paralleling exposures of Pinal Schist (fig. 3). Madera-type diorites are pretectonic and deformed at their margins, but less foliated granodiorite core phases have been mistaken as posttectonic (e. g., Livingston, 1969). The bodies, which include diorite, quartz diorite, granodiorite, and tonalite, contain xenoliths of Pinal Schist host rocks and gabbro-pyroxenite source material. In the Ray-Pinal Mountains area, Madera Diorite has Rb-Sr and K-Ar dates between 1720 and 1600 Ma, but other bodies intruding metarhyolite near Ray have 1650-Ma Rb-Sr whole-rock ages (Livingston, 1969; Livingston and Damon, 1967). U-Pb zircon ages show that Madera-type diorites and related bodies near the northern boundary of the Southeast Schist Belt were emplaced between 1685 and 1660 Ma (Silver, 1976).

A very distinctive feature of Madera-type diorites is their high titanium and iron contents, which exceed those of mafic pretectonic bodies in comparable tectonic settings in the Central Volcanic and Northwest Gneiss Belts (fig. 4), where much less tholeiitic granodiorites were produced by fusion of much more mafic host crusts (P. Anderson, this volume). These factors, plus petrologic evidence for derivation from tholeiitic gabbro sources, suggests that the Madera-type diorites were derived directly from a Fe-Ti-rich quartz tholeiite parent magma of subcrustal origin. The diorites cannot have been derived by anatexis of a crust as evolved as their Pinal Schist host.

Granodiorite plutons intrude the southern Southeast Schist Belt in the Dos Cabezas-Johnny Lyon Hills region and have been described as posttectonic (Silver, 1978); but, like the huge batholiths of the Northwest Gneiss Belt, the granodiorites are foliated at their margins and less deformed in the cores, and show all earmarks of being syntectonic. U-Pb zircon ages of 1625 and 1620 Ma on these southern granodiorites (Silver, 1978) significantly postdate the 1660-Ma age (Silver, 1976; Silver and others, 1986) of syntectonic Sunflower granite south of Payson, which indicates that the southernmost part of the Southeast Schist Belt was deformed later than the younger part of the Central Volcanic Belt.

Deformation and Metamorphism

Where syntectonic plutons in one crustal region postdate posttectonic plutons in an adjacent region, the two regions must have experienced different tectonic histories. Such is the case not only between the Southeast Schist Belt and the younger portion of the Central Volcanic Belt, but also between the two different terranes of the Southeast Schist Belt itself. The northern part of the belt dominated by Pinal Schist was intruded primarily by pretectonic diorites and granodiorites, and was deformed soon after or during the last stages of their intrusion at about 1650 Ma. The southern portion, including the Dos Cabezas volcanic belt, in contrast, appears to have been deformed as late as 1625-

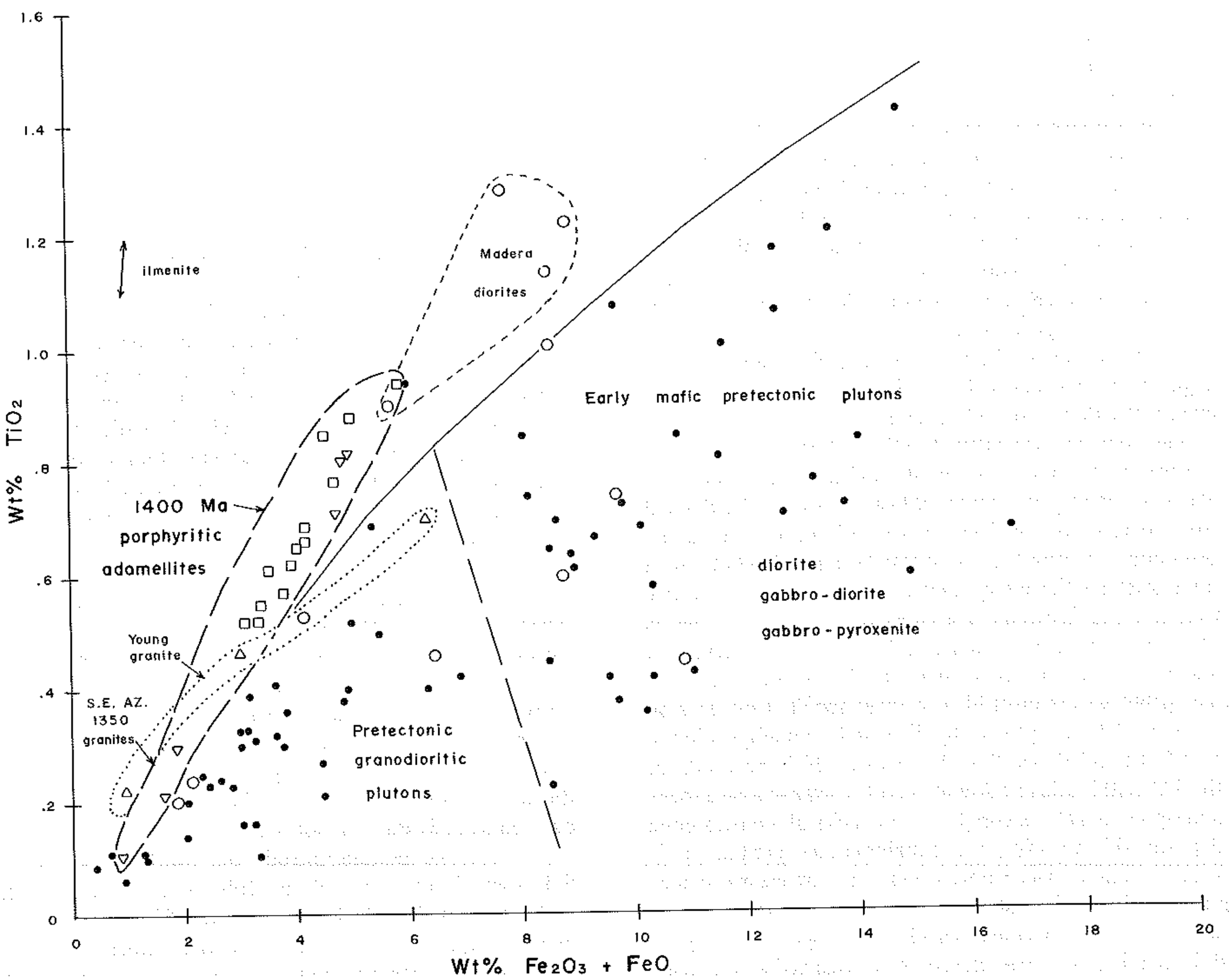

Figure 4. TiO_2 vs. total iron plot of various plutonic rocks in Arizona showing the small, well-defined fields of Madera diorites (open circles, dashed field) and 1400-Ma porphyritic granites (squares and triangles) compared to the larger fields of pretectonic plutons and batholiths in the Central Volcanic Belt. Solid circles in the pretectonic fields are analyses from the older Prescott-Jerome belt, whereas open circles are analyses of pretectonic plutons in the younger New River-Cave Creek-Mazatzal Mountain-Diamond Butte belts. The Young granite field marks a division in the 1400-Ma granite field between more felsic bodies in southeast Arizona (triangles) and the main suite of 1400-Ma granites in central Arizona (squares).

1620 Ma (Silver, 1978), and therefore must have been deformed after the northern part of the Southeast Schist Belt. These data also indicate that deformation of the Southeast Schist Belt occurred after that of the Central Volcanic Belt, and that the Southeast Schist Belt experienced a different tectonic history than central and northern Arizona.

Deformation of the Southeast Schist Belt is moderate to strong penetrative strain in incompetent Pinal Schist and volcanic rocks, but is weak in competent, massive volcanic and subvolcanic bodies. Likewise, the pretectonic diorites and granodiorites are most highly deformed at their borders, whereas syntectonic granodiorites are weakly foliated at their borders. This type of deformational structure is analogous to vertical deformation in the Central Volcanic Belt (P. Anderson, this volume) except for three important differences: in the Southeast Schist Belt (1) strain is less intense overall, and less penetrative in bodies of similar competency; (2) tectonic fabrics trend more easterly, from 060° to 110° rather than from 025° to 075° as in the Central Volcanic Belt; and (3) deformation postdates that of the Central Volcanic Belt. Such contrasts indicate that the nature of the forces was similar, but that the geometry and kinematics of deformation were different.

Structural fabrics and metamorphic grade variations in the Southeast Schist Belt are closely analogous to those in the youngest units in the Central Volcanic Belt that composed the upper crustal layers during deformation and

metamorphism, namely the Alder Group, rhyolitic ignimbrites, fragmentals of the felsic complexes, and the Mazatzal Group (P. Anderson, this volume). Lineations tend to be shallow rather than steeply plunging, and open folds are more common than tight folds. Nevertheless, high-strain zones exist throughout the Pinal Schist terrane and may also occur in the Little Dragoon Mountains where the Dos Cabezas volcanic belt bounds the Pinal Schist terrane. Identification of the key high-strain zones in the Southeast Schist Belt is rendered difficult by abundant younger 1400-Ma plutons and superimposed Phanerozoic tectonic fabrics.

Metamorphic facies variations are almost totally governed by the array of 1400-Ma megaporphyritic granites that intrude the Southeast Schist Belt. Metamorphism by Madera-type diorite intrusions is limited to narrow contact aureoles generally devoid of aluminosilicate index minerals, but andalusite and cordierite are widespread near major 1400-Ma batholiths such as the Four Peaks, Maricopa, Ruin, and Oracle batholiths. However, such effects are postorogenic, and when just the pre- and synorogenic metamorphic structure of the Southeast Schist Belt is analyzed, it is clear that the belt is diagnostically devoid of metamorphic grades significantly above middle greenschist facies. In many places, synorogenic recrystallization was barely sufficient to convert volcanic and sedimentary protoliths to semischists and prograde depositional mineralogies to quartz-sericite-albite-chlorite.

Thus, the Southeast Schist Belt is perhaps the most enigmatic part of the Arizona Proterozoic. It is penetratively deformed, yet never underwent a major orogeny involving invasion by orogenic plutons and widespread regional metamorphism and deformation, as did other parts of Arizona. Orogenic plutons are few, small, did not regionally metamorphose or deform their host crust, and were definitely not derived by anatexis of Pinal Schist or anything remotely like it. To further complicate the picture, the lithology of Pinal Schist is remarkably uniform for such a widespread assemblage, yet stratigraphic details do not seem to correlate from area to area, and the assemblage has no known basement. Moreover, although the Dos Cabezas volcanic belt has a stratigraphic evolution similar to the Central Volcanic Belt, its chemistry is anomalous, and it was not demonstrably linked in its early history to any other volcanic belt in Arizona, or to the Pinal Schist terrane itself. However, its latest history was temporally and spatially linked to the Central Volcanic Belt. Lastly, the southern part of the Southeast Schist Belt may contain granitoid and andesitic crustal fragments exotic to the oceanic setting that existed throughout most of Arizona in Proterozoic time; these fragments must have been tectonically emplaced after their formation but prior to 1685-Ma consolidation of the entire region.

CENTRAL VOLCANIC BELT

The Proterozoic geology of the Central Volcanic Belt is summarized elsewhere in this volume (also see P. Anderson, 1986), and the following discussion proceeds directly on from that review. Here the two major components—the older Prescott-Jerome portion and the younger New River-Cave Creek-Mazatzal Mountains-Diamond Butte portion—are integrated by focusing on similarities and differences in stratigraphy, petrogenesis, and tectonic history between them. These comparisons and contrasts show how the Central Volcanic Belt developed into a contiguous volcanoplutonic arc from 1800 to 1700 Ma.

Relationship of Older and Younger Parts

The older, mafic part of the Central Volcanic Belt lies northwest of the Moore Gulch shear zone and includes the Prescott-Jerome volcanic belts, distal volcaniclastic and sedimentary strata flanking the belts to the southwest and northwest, and possibly detached pieces near Payson. The younger, more felsic part lies southeast of the Moore Gulch shear zone and includes volcanic belts in the New River, Cave Creek, Diamond Butte, and Mazatzal Mountains areas, plus distal volcaniclastic and sedimentary strata flanking them to the southeast.

This spatial separation has been taken to indicate that the Moore Gulch shear zone is a profound tectonic boundary, such as a major strike-slip fault or crustal suture, that juxtaposes two volcanic terranes of entirely different origins (Maynard, 1986). Such an hypothesis is incorrect (P. Anderson, this volume) because: (1) Alder Group purple slates lie both east and west of the shear zone and extend Arizona cross it without discernible offset near New River to the south; (2) underlying andesites and dacites west of New River extend across the zone easterly into the Daisy Mountain area without offset; and (3) to the north at Brooklyn Peak, the emergent Cherry Springs batholith west of the shear zone shed tonalite boulders into Brooklyn Peak conglomerate east of the shear zone prior to ignimbrite extrusions of New River Mountains Felsic Complex. Such clear relationships unify the volcanic evolutions of areas east and west of the shear zone from their inception to felsic ignimbrite activity, require broad-scale lateral preservation of rock units, disprove any major strike-slip offset across the shear zone, and demonstrate that the zone was not a crustal suture.

In fact, the Moore Gulch shear zone is simply a zone of high strain that formed along the edge of the Cherry Springs batholith prior to regional Proterozoic deformation, and remained a weak zone thereafter. Significant vertical movement on the order of 1 or 2 kilometers occurred after batholith emplacement, causing elevation of the older northwestern block and subsidence of the southeastern block during Alder deposition and before extrusion of the New River Mountains ignimbrite fans. Thus, spread of the ignimbrites to the west was stopped by a scarp of Cherry Springs batholith, and consequently none of the younger felsic ignimbrites are found to lap unconformably over older volcanic formations of the Prescott-Jerome belt.

The Cherry Springs batholith was emplaced from 1740 to 1720 Ma prior to ignimbrite eruptions, and intruded host rocks both to the east and west: those to the east are involved in the Little Squaw Creek Migmatite Complex; those to the west are interstratified with upper parts of the Black Canyon Creek Group near New River to the south, and appear to be unconformably overlain by Union Hills Group basaltic andesites to the east. Therefore, the Union Hills Group postdates the Black Canyon Creek Group (P. Anderson, this volume), so a clear separation in space and time exists between youngest volcanics of the Prescott belt and oldest volcanics of the New River belt. Thus, separation between older and younger parts of the Central Volcanic Belt is at a stratigraphic time line, not a fault. The time difference between the two volcanic sequences is presently undated and may be as much as 20 m.y.

The presence of 1738-Ma plutonic ages in the Gibson Creek complex near Payson led Silver and others (1986) to infer that the "Yavapai Series" forms a basement to younger Alder and Haigler Groups in the Payson area, which is not true. It has been known for decades (Wilson, 1939; Gastil, 1958) that mafic volcanic rocks stratigraphically underlie Alder Group in the Payson-Mazatzal-Diamond Butte area, and the predominance of such mafic volcanics throughout the eastern part of the Central Volcanic Belt has now been clearly demonstrated (P. Anderson, 1986; this volume). These oldest mafic volcanics beneath the Alder Group are not "Yavapai Series," but are part of the Union Hills Group. It is Union Hills Group strata that regionally underlie Alder strata throughout the New River-Cave Creek-Mazatzal Mountains-Diamond Butte areas (P. Anderson, this volume), not rocks of the original "Yavapai Series."

The above point is not merely semantic, but a crucial aspect of the Central Volcanic Belt's tectonic evolution. As shown below, a systematic southeastward progression of formative volcanism with time meant that each successive volcanic center evolved after its counterpart to the northwest had mostly ceased. Hence younger clastic sequences to the southeast (Alder Group) do not lie on older volcanics of the northwest region (e. g., Bradshaw Mountains, Mayer or Black Canyon Creek Groups) but on younger volcanic sequences that were the oldest formative components in the southeast region (i. e., Union Hills Group).

Thus, referring to formative mafic volcanics of the eastern region as "Yavapai Series" is incorrect and stems from confusing time-stratigraphic and rock-stratigraphic terminologies. Although the time-stratigraphic "Yavapai Series" term once usefully distinguished older volcanics in Yavapai county from younger Alder strata to the east (C. A. Anderson and others, 1971), stratigraphic relations are now well known (P. Anderson, 1986) and show that "Yavapai Series" no longer survives scrutiny, hence it is superseded by the rock-stratigraphic term **Yavapai Supergroup** (P. Anderson, 1986; this volume). In contrast, time limits are not well known: neither the oldest nor youngest volcanics originally included in "Yavapai Series" have been dated, so its time limits are undefined. Moreover, rocks of the original "Yavapai Series" and "Alder Series" are time transgressive across Arizona, hence the age limits of one series in one area overlaps with age limits of the other series in other areas (see P. Anderson, 1986 for detailed discussion).

Some rocks in the Gibson Creek complex near Payson may predate formative evolution of the Union Hills Group in the younger eastern belts (S. Bowring, unpub. U-Pb isotope data). Thus, early plutonic phases of the Gibson Creek complex may have been easterly mafic bodies correlative to the Cherry Springs batholith to the west, but, like the Union Hills Group, were spatially separated from the western region by later intrusion of the 1710-Ma Verde River granite (P. Anderson, this volume). Hence relationships near Payson are identical to those in the New River belt, where Union Hills Group formative volcanism of the eastern belts postdates early 1740-Ma phases of the Cherry Springs batholith.

Southeast of the Gibson Creek complex, however, pre-1738-Ma stratified rocks are lacking. Instead, Union Hills Group mafic volcanics evolved to the southeast after 1738 Ma along the eastern edge of the preceding Black Canyon Creek Group volcanic chain (and detached remnants in the Gibson Creek complex), exactly in the same manner that the eastward-stepping progression of volcanic chains built up the Prescott-Jerome belts (P. Anderson, this volume). Hence, the Union Hills Group appears to be the last stage in the eastward progression of formative volcanism that built up the entire Central Volcanic Belt, and therefore its age must lie between 1740 and 1720 Ma (P. Anderson, this volume). Dating of the Union Hills and Alder Groups will effectively close the apparent "gap in volcanism" in central Arizona that is an artifact of dating only pre-1750-Ma volcanics in the Prescott-Jerome region and only 1710-Ma and younger sequences in the eastern belts.

In summary, therefore, there is virtually no time-space overlap in detail between the older mafic volcanic part and younger felsic part of the Central Volcanic Belt. The boundary is not a major strike-slip fault or suture, but a clear stratigraphic separation between two different, adjacent rock sequences, which is locally affected by the Moore Gulch shear zone. Rocks of the Prescott-Jerome belt do not extend eastward underneath stratified sequences of the younger belts; instead, the Union Hills Group is the basal volcanic assemblage of the younger belts. Pre-1740-Ma stratified rocks exist in the younger eastern belts only as fragmentary screens or inclusions in the correlative Little Squaw Creek and Gibson Creek complexes that delimit the southeast edge of the pre-Union Hills Group volcanic terrain. Formative volcanics of the younger eastern belts developed along side the older mafic belts, so the Union Hills Group can be viewed as the youngest easternmost magmatic cycle in the total formative evolution of the Central Volcanic Belt.

Continuity of Formative Volcanism

The Central Volcanic Belt displays a continuous volcanic evolution, not in a vertical stratigraphic column, but in a diachronous manner where volcanism progressed in space and time southeastward across the width of the belt. The earliest volcanism (Bradshaw Mountains Group) involves a primitive, magnesian, low-K tholeiitic, bimodal basalt-rhyodacite suite unique to the northwest part of the belt. The next major volcanic cycle evolved huge differentiated bimodal tholeiitic basalt-rhyolite and polymodal basalt-andesite-dacite-rhyolite centers that fractionated along classic iron-rich Skaergaard tholeiitic trends. This cycle is found only in the central Prescott belt (Mayer Group) and part of the Jerome belt. A third major volcanic cycle evolved high-K tholeiitic, altered, intermediate to felsic fragmental rocks of the Black Canyon Creek Group, which are restricted to the eastern parts of the Jerome and Prescott belts. Finally, a fourth suite of calc-alkaline basaltic andesite-andesite-rhyolite volcanic centers evolved in the New River, Cave Creek, Mazatzal Mountains, and Diamond Butte areas; the calc-alkaline volcanics were formative only to the southeast part of the Central Volcanic Belt and did not spread northwest, just as earlier tholeiitic rocks to the northwest did not spread southeast.

Each volcanic cycle in the Prescott-Jerome belt evolved its own unique volcaniclastics in troughs between major volcanic edifices, both longitudinally down the axis of the volcanic chain and laterally on its flanks. When the next chain of volcanoes developed to the southeast, its oldest volcanic strata interfaced with volcaniclastics that had been shed distally southeastward from the previous volcanic cycle. Then as the new volcanic cycle evolved, its youngest deposits backfilled the trough between the two volcanic chains, ultimately lapping back over proximal units of the earlier cycle (P. Anderson, this volume). Similar volcaniclastic lenses dominate the New River, Cave Creek, Mazatzal Mountains, and Diamond Butte belts, occurring mainly between major edifices in each volcanic chain. On the flanks of the main Union Hills, North Union Hills, Cave Creek, Mount Ord, and East Verde River volcanic centers, the thick fans of andesitic graywacke, agglomerate, and conglomerate that were shed into deep troughs between edifices can be distinguished from one another.

A thick clastic wedge developed along the southeast flank of the Union Hills volcanic chain (fig. 3) and persisted as a shallow basin throughout Alder Group deposition. The well-sorted volcanic siltstone, slate, subgraywacke, and quartz wacke in this wedge are more mature than rocks of the intervolcanic basins. A smaller wedge of andesitic graywacke and siltstone was shed northward into the East Verde River area, but few clastics were deposited between the Union Hills and Black Canyon Creek Group near New River because of the proximity of the two major volcanic centers. Unlike the Prescott belt where deposits of one cycle lapped back over those of the previous cycle, there was little overlap of Union Hills Group strata onto the southeastern edge of the Prescott belt in formative volcanic stages, only during later deposition of the Alder Group.

Younger Volcaniclastics and Felsic Ignimbrites

Union Hills Group calc-alkaline basaltic andesite volcanism was soon succeeded by felsic tuffaceous volcanism in the upper Union Hills Group, and later by deposition of Alder Group fine-grained clastics, also with a felsic to intermediate volcanic component. Because the basal Alder contact is nearly conformable in many places, it is unlikely to reflect a 30-m.y. gap in erosion or nondeposition, as implied by the existing 1738 to 1710-Ma gap in isotopic ages (Silver and others, 1986). The relative chronology (P. Anderson, 1986) indicates that the Union Hills Group accumulated between about 1735 and 1725 Ma and that early Alder Group deposition, interrupted by many hiatuses, continued from about 1720 to 1710 Ma, at which time felsic volcanic conglomerates heralding ignimbrite eruptions overwhelmed Alder clastic sedimentation.

In contrast to previous lensoidal volcanic units, Alder strata formed blanketlike deposits in a shallow basin that extended the full length of the Union Hills Group volcanic chain between the Cramm Mountain-East Verde River centers to the north and the Union Hills-Mount Ord centers to the south. In this east-facing basin, westward-thinning Alder lithofacies prograded westerly along the basin to lap unconformably back over the earlier volcanic deposits and shoal against the older emergent Prescott belt (P. Anderson, this volume). Lower shales of the Alder Group near New River are comparable to Texas Gulch slates of the Prescott-Jerome volcanic belt, both in lithology and structural localization to tectonic troughs along edges of major batholiths and plutons. Thus, the downfaulted trough southeast of the Cherry Springs batholith that contained Alder Group sediments mirrored the structural trough northwest of the batholith in which the Texas Gulch Formation was deposited. Therefore both the Shylock and Moore Gulch shear zones originated as boundaries to a horst.

The Alder sedimentary interlude marks a fundamental change in the style of volcanism throughout the younger portion of the Central Volcanic Belt, from formative mafic calc-alkaline volcanism before, to felsic alkali-calcic volcanism thereafter. As calc-alkaline felsic magmas invaded younger parts of the Central Volcanic Belt, disrupting the coherency of previous volcanic units, Alder sediments were overwhelmed by felsic conglomerates, and then ignimbrite fans were sequentially extruded from three major sites where magmas breached surface, while red granite batholiths and plutons crystallized at depth. The ignimbrites spread north and south from the westerly extrusive center in the New River Mountains along the same trough as the Alder Group, their westward spread being limited by the scarp of the emergent Cherry Springs batholith. Subsequent eruption of Mount Peeley ignimbrites to the east was followed by ignimbrites from the Red Rock and Haigler centers farther east in the Mazatzal Mountains-Diamond Butte areas.

As ignimbrite activity stepped easterly across the younger felsic belt, much of the belt was left emergent. Subsequent erosion of ignimbrites back to sea level, first in fluvial then estuarine, back-bay, littoral, and finally open-marine conditions, deposited the distinctive suite of hematitic conglomerates and quartzites, shales, siltstones, and mature quartzites known as the Mazatzal Group. The initial network of Mazatzal-age fluvial channels feeding into open-marine conditions to the southeast started in Chino Valley near Prescott, where quartzose detritus from the Prescott-Jerome volcanic belts was eroded (Wirth, 1980). The fluvial network then flowed southeasterly through Natural Bridge north of Payson to join another main channel feeding hematitic detritus from Sheep Basin Mountain westerly into the Mazatzal Mountains (P. Anderson and Wirth, 1981). The main channel network then flowed south down the Mazatzal Mountains through Four Peaks and into Hess Canyon, where the strand line was located.

As erosion continued, open-marine conditions prograded back over the Mazatzal Mountain-Diamond Butte region and estuarine conditions were attained in the drowned river valley as far northwest as Chino Valley, thus spanning the entire width of the Central Volcanic Belt. Mazatzal Group deposition followed immediately upon the last stages of ignimbrite activity in the Diamond Butte area, and continued to about 1680 Ma ago. Thus, the Central Volcanic Belt was centrally drained by a major channel network that fed out to marine conditions to the southeast. The repeated southeastward progressions of first mafic then felsic volcanism across the Central Volcanic Belt and the repeated prograding of first Alder then Mazatzal strata northwestward across the belt attest to the fact that the open ocean was located southeast of the belt for its entire younger evolution, from at least 1720 (or 1730) Ma onward. By 1720 Ma, therefore, the Central Volcanic Belt had become a continent-margin magmatic belt that fronted a new Proterozoic continent.

Plutonism

In a general sense, the time span of plutonism in the Central Volcanic Belt overlapped with, but was slightly younger than, the time span of formative volcanism. In detail however, plutonism mostly followed the formative volcanic sequences in each area. The oldest gabbro-microgabbro-diabase complexes were coeval with oldest tholeiites of the Bradshaw Mountains and Ash Creek Groups, but the intervolcanic gabbro-diorite suite followed this earliest volcanism. Volcanism was thus divided into two discrete stages: an early primitive stage predating the gabbro-diorite bodies and a later, more evolved stage that overlapped them. After a new tholeiitic volcanic cycle formed the Mayer Group to the southeast, the first small granodiorite plutons intruded previously established parts of the volcanic belt to the northwest at about 1750 Ma.

As formative volcanism swept southeastward across the Central Volcanic Belt from 1755 to 1740 Ma, the belt sustained much plutonism between 1745 and 1735 Ma. Because batholiths and plutons were emplaced soon after formative volcanism in each area, the earliest plutonic phases are subvolcanic to the extrusive volcanic sequences. The west side of the Cherry Springs batholith is an example where subvolcanic K-feldspar phyric dacite is transitional between Black Canyon Creek Group rhyodacite-dacite fragmentals and feldspar-porphyritic granodiorite of the batholith, showing that early batholith phases were coeval with the Black Canyon Creek Group, but later phases intruded it. Likewise, subvolcanic dacite at Battle Flat that was coeval with Spud Mountain volcanism immediately preceded intrusion of Brady Butte granodiorite, with no major petrologic or chemical differences. Therefore, the earliest plutons closely followed or overlapped volcanism, because felsic magma chambers for the last extrusions of each volcanic cycle crystallized at depth as the earliest plutons.

After about 1740 Ma, however, the Central Volcanic Belt recorded a very different plutonic evolution. The southeastward progression of volcanism with its attendant stratiform, diapiric, subvolcanic, or small-sized plutonic bodies ceased, and the entire volcanic belt was pervaded by plutonism on a much grander scale. Peripheries of the Prescott-Jerome belt were invaded by many pretectonic granodiorite and tonalite bodies, including the Wilhoit batholith, early Cherry Springs batholith phases, Minnehaha Granodiorite, and other similar bodies (P. Anderson, this volume). The Gibson Creek diorite near Payson was genetically related to the Cherry Springs batholith and was also emplaced near this time. All such plutons and batholiths were emplaced into a pretectonic setting and together caused the initial crustal thickening of the Central Volcanic Belt. Their emplacement produced both local and widespread thermal aureoles and deformation of host strata around their edges. Most major structural troughs and synclinal keels of downwarped volcanic rock sequences, as well as plutonic blocks soon to become structurally positive, had their inception at this time, and were further developed by regional deformation at later times.

Major granodioritic plutonism continued from 1740 Ma to 1720 Ma, with emplacement of southern phases of the Cherry Springs batholith and intrusion of pre- and syntectonic plutons into the Prescott belt. Syntectonic batholiths and plutons were emplaced throughout the Northwest Gneiss Belt during its major plutonism at about 1720 Ma. In contrast to formative volcanic evolution that stepped progressively southeastward, this major orogenic event of syntectonic plutonism swept across the Central Volcanic Belt **in exactly the opposite direction,** from southeast to northwest, and extended on into the Northwest Gneiss Belt to extensively rework its various crustal segments.

By 1720 Ma all early pretectonic batholiths and plutons in the Prescott region were emplaced, and some were already uplifted by continued plutonism and unroofed by erosion. Two narrow subsidence troughs along edges of the Cherry Springs batholith and Brady Butte pluton became sites for high-K calc-alkaline effusive rhyolitic volcanism, reflecting the evolved state of the Prescott-Jerome volcanic belt at the time. Texas Gulch successor clastics were laid down in these structural troughs unconformably on the unroofed plutonic rocks.

Concurrently, a similar fault scarp developed along the southeast edge of the Cherry Springs batholith and faced open-marine conditions to the south. In this shallow-marine basin Alder Group strata accumulated between 1725 and 1710 Ma and transgressed westerly up to the scarp front. The earliest Alder strata were thus contemporaneous with the Texas Gulch Formation, but younger Alder strata were coeval with Brooklyn Peak conglomerate to the north, which received boulders from a positive area of Cherry Springs batholith to the west. The Cherry Springs batholith horst may have resulted from intrusion of the 1720-Ma Bland tonalite at depth. Bounding scarps to the horst, after intense vertical strain, developed into the Shylock high-strain zone and Moore Gulch shear zone, which are precluded from being major strike-slip faults, thrusts, or crustal sutures because of their tight paleogeographic constraints and vertical movement histories. Lastly diabase-microdiorite dikes and sills in the Mazatzal Mountains-Diamond Butte area and diorite-granodiorite plutons in the Cave Creek-New River areas intruded Alder strata just after consolidation.

By 1710 Ma, the character of plutonism in the Central Volcanic Belt changed, as primary felsic magmas intruded the crust. Between the New River Mountains and Diamond Butte, two major batholiths of red granite were emplaced, each feeding its own carapace of alkali-calcic felsic ignimbrites. While these felsic magmas intruded the eastern part of the Central Volcanic Belt at upper crustal levels, older western parts of the belt were intruded by calc-alkaline granite plutons that originated as anatectites mobilized from metasedimentary migmatite-anatectic complexes undergoing fusion at deeper crustal levels.

Timing of Deformation and Metamorphism

After emplacement of the last pretectonic plutonic suites at about 1720 Ma, the Prescott-Jerome belt experienced profound deformation from about 1710 to 1690 Ma, which was the major regional Proterozoic deformation that affected all of north-central Arizona—the early part of Wilson's (1939) "Mazatzal revolution." This deformational event was accompanied by emplacement at intermediate crustal levels of syn- and late-tectonic plutons and batholiths along the edges of the Prescott belt, and by strong vertical deformation within the belt itself. Thus, major penetrative deformation of the Prescott-Jerome belt occurred after 1720 Ma but before 1690 Ma, and probably occurred close to 1700 Ma.

Plutonism continued past 1700 Ma in a few parts of the Prescott-Jerome belt up to 1690 Ma or later. A broad anatectic region developed at depth where downwarped sedimentary rocks were partially fused to make the Southern Bradshaw Mountains Migmatite Complex; anatectite was mobilized from this complex to be emplaced higher in the crust as the 1700-Ma (Silver, 1976) or 1695-Ma (Bowring, 1986) Crazy Basin granite. This event records the peak of thermal metamorphism in deep southern roots of the Prescott belt, just after regional deformation, when early penetrative fabrics were overprinted by recrystallized metamorphic fabrics. If deformation of the rest of the belt coincided with timing in this southern region, which is most likely, and if the regional metamorphic peak just followed Proterozoic deformation of the entire region, then the Prescott-Jerome belt was most strongly and penetratively deformed between 1705 and 1695 Ma.

Essentially all major plutons and batholiths had been emplaced by 1705-1695 Ma, and responded competently to deformation, such that strain was intensified in enveloping strata. This created a deformational regime where strain (1) was concentrated in the stratified rocks, (2) was extremely heterogeneous both in intensity and aspect (flattening-elongation) ratios, and (3) was governed in overall geometry and kinematics by the regional disposition of competent plutonic masses around and within the volcanic belts (P. Anderson, this volume). Throughout most of the Prescott-Jerome belt where deformation and metamorphism were of moderate to low intensity (middle greenschist grade and moderate strain state), the deformational and metamorphic peaks coincided in time, but at deeper crustal levels where elevated P-T conditions and ductile strain states were attained, a clear temporal separation of the metamorphic and deformational peaks exists. Thus deformation and metamorphism were not exactly contemporaneous in all places, but varied with position in the belt.

Deformation was also diachronous between the Central Volcanic Belt and Northwest Gneiss Belt: to the northwest deformation coincided with emplacement of major batholiths and migmatite formation at ca. 1720 Ma, as much as 25 m.y. earlier than in the Prescott-Jerome region. Major penetrative deformation of the Prescott-Jerome belt had ended prior to crystallization of the Crazy Basin pluton, or no later than 1690 Ma. But in the younger southeast part of the Central Volcanic Belt, 1680-Ma diorite and granodiorite plutons are strictly pretectonic and are deformed by the major deformational event that affected all of the New River, Cave Creek, Mazatzal Mountains, and Diamond Butte volcanic belts. In the Mazatzal Mountains area, 1660-Ma Sunflower granite was emplaced syntectonically during this deformational event—the later part of Wilson's "Mazatzal revolution"—and only the 1630-Ma alkalic

Young granite is truly posttectonic. Therefore deformation of the New River-Cave Creek-Mazatzal Mountains-Diamond Butte belts is bracketed to the period 1670 to 1650 Ma, which is about 20 to 25 m.y. younger than deformation of Prescott-Jerome volcanic belt.

In the younger felsic part of the Central Volcanic Belt, deformation was not as intense as in the older region, partly because higher crustal levels are represented. At the highest crustal levels, a tectonic regime that included thrusting dominated the Mazatzal Group, in contrast to the vertical regime at deeper crustal levels in older volcanic sequences. Although penetrative strain in the younger portion was of similar kinematics to that in the older portion, the cumulative net strain was not so intense in the younger portion. Thus, by 1670-1650 Ma when the younger portion was undergoing deformation and low-grade metamorphism, orogenic activity had ceased in the older mafic portion of the Central Volcanic Belt, and had long since ceased in the Northwest Gneiss Belt.

Mazatzal Deformation vs. Prescott Deformation

Deformation of Arizona's Proterozoic crust can be viewed as progressing southeastward with time as each crustal portion was successively thickened and stabilized by cycles of plutonism, deformation, and metamorphism. Deformation occurred in the Northwest Gneiss Belt 25 m.y. prior to that in the Central Volcanic Belt's older portion, which in turn occurred 25 m.y. prior to that in the Central Volcanic Belt's younger portion. Despite these differences, Wilson's original "Mazatzal revolution" concept is still valid as a broad term encompassing ca. 1700-Ma orogeny of central Arizona, if used to refer to the penetrative strain events that all parts of central Arizona experienced. The term can be more precisely defined as only 1670-1650-Ma deformation, if it is restricted to the deformational event that affected the Mazatzal Group in the Mazatzal Mountains, which is certainly how Wilson intended the term to be used.

Using this restricted definition of "Mazatzal revolution," and renaming it more accurately **Mazatzal deformation,** it is clear that Mazatzal deformation postdated penetrative deformation of the Prescott belt. Mazatzal deformation was too weak to cause refolded folds or large-scale interference structures in the Prescott-Jerome belt, but some of its effects can be detected. Its axial orientation was slightly different than that of the earlier penetrative strain of the Prescott belt, a difference reflected in local crenulation of foliation in highly fissile units of the Prescott area, gentle warping of Mazatzal strata in Chino Valley, and other mild effects that are weakly superimposed upon the earlier intense penetrative fabric of the older **Prescott deformation.** This mild overprinting does not amount to "polyphase folding on a regional scale," because it occurs only in very few places and only in highly susceptible units.

Thrusting in Mazatzal strata in the Mazatzal Mountains has been postulated as an early event of regional extent in the Prescott area that predated major Prescott deformation (Karlstrom and Puls, 1984; Karlstrom, 1986). Isotopic and relative-age relationships cited above prohibit such an extrapolation because Mazatzal Group thrusting is integral to Mazatzal deformation and significantly postdated major Prescott deformation. Thrusting occurred in Mazatzal Group strata because of their brittle response to strain, and the soles of thrust planes are rooted in the incompetent Maverick shale, which was the decollement for dislocation of competent upper Mazatzal Peak quartzites during regional deformation (P. Anderson, this volume). Some thrusts locally extend down into the Deadman Formation but not into underlying ignimbrites or the Alder Group. Extrapolation of Mazatzal thrust faults throughout central Arizona is therefore incorrect, not only because of time differences between the two deformational events, but also because ductile deformation, not thrusting, occurred at depth.

Thrusting in Mazatzal strata reflects high-level brittle accommodation of penetrative strain in the underlying Union Hills and Alder Groups during horizontal shortening of the eastern volcanic belts. The underlying sequences accommodated greater total net flattening and vertical extension than Mazatzal strata, and Mazatzal folding and 1- to 2-km-scale imbrication was an attempt to match this strain, not a separate regional event. At depth in the deformed volcanic pile, movement was distributed across broad zones of strain, and both the intensity of foliation and lineation and to some extent lineation steepness directly reflected strain intensity. Where strain was strongly localized into narrow preexisting weak zones in the volcanic pile, high-strain zones developed as a key feature of Proterozoic vertical deformation (P. Anderson, this volume).

Thus, younger ca. 1670-Ma Mazatzal deformation and older ca. 1700-Ma Prescott deformation were mutually exclusive in space and time throughout central Arizona, except for local spatial superimposition in a few small parts of the Prescott belt. Thrusting was the brittle upper crustal expression of Mazatzal deformation, not a separate regional event; it occurred after Prescott deformation and was precluded from the ductile vertical strain regime that dominated the volcanic belts. Deformation of the entire Southeast Schist Belt postdated Mazatzal deformation. The northernmost part of the belt near Globe was deformed at about 1650 Ma, but the southeast part was evidently deformed as late as 1625-1620 Ma, about 25 m.y. after its northern counterpart.

PROTEROZOIC PLATE TECTONIC SETTINGS

This paper has surveyed the key tectonic elements of Arizona's Proterozoic crust and the nature of basement upon which that crust was built. From this data a plate tectonic history can be deduced that describes processes of continental accretion and plate tectonics that most likely

operated to form Arizona's earliest Proterozoic crust (P. Anderson, 1986). The following summary considers the tectonic setting of the Central Volcanic Belt first because it is best known, and then the settings of the Northwest Gneiss Belt and Southeast Schist Belt.

CENTRAL VOLCANIC BELT

Tectonic Setting

From the oceanic nature of its basement and from supracrustal relationships (P. Anderson, this volume) it is clear that the earliest part of the Prescott-Jerome volcanic belt was built directly upon Proterozoic oceanic crust in a deep-submarine, open-ocean setting several hundred kilometers from any continental crust or shallow-water clastic deposits. This oceanic crust was of Proterozoic age, just predated the volcanic belts, and was of mafic-ultramafic (gabbro-pyroxenite-peridotite) composition. The upper supracrustal section of the oceanic crust consisted of low-K tholeiitic pillowed basalt flows, underlain by subvolcanic microgabbro, and overlain by basaltic tuffs, deep-sea pelagic sediments, and chert. This supracrustal oceanic section, exposed west of the Prescott-Jerome volcanic belt, was very similar to early sequences in the volcanic belt that may include an upper Proterozoic oceanic crustal section. The key difference between areas distant from the volcanic belt, where Mg-rich low-K tholeiitic basaltic sequences comprise the upper portion of an oceanic crustal section, and earliest strata of volcanic belt where a comparable section occurs, was that volcanism persisted in the Prescott belt to develop an intraoceanic volcanic chain upon that oceanic crust.

All evidence indicates that the Prescott volcanic belt was conceived as a chain of submerged volcanoes in an open-ocean, deep-submarine environment. In this intraoceanic setting, some volcanoes emerged as islands, and the chain was likely originally arc shaped; thus, the Prescott-Jerome volcanic belt was born as an intraoceanic island arc upon Proterozoic oceanic crust and is therefore the Proterozoic analog of modern intraoceanic island arcs in both structure and setting. However, it was not identical to modern intraoceanic island arcs in important aspects such as size, detailed lithologic makeup, and compositional-petrologic features. Consequently, Proterozoic intraoceanic island arcs should be understood as analogous to recent island arcs, not identical to them.

Tectonic Evolution

The Prescott arc evolved from old, chemically primitive mafic suites in the northwest through stages of increasing chemical and petrologic maturity as the volcanic axis stepped sequentially southeast (P. Anderson, this volume). The petrologic-chemical stages were: (1) Mg-rich, low-K tholeiitic bimodal basalt-rhyodacite (Bradshaw Mountains Group); (2) Fe-rich tholeiitic bimodal basalt-rhyolite (lower Mayer Group); (3) high-K tholeiitic polymodal basalt-andesite-dacite-rhyolite (middle Mayer Group) fractionally derived from the Fe-rich tholeiitic parent; (4) high-K tholeiitic dacitic pyroclastics (upper Mayer Group); (5) tholeiitic to calc-alkaline rhyodacitic pyroclastics (main Black Canyon Creek Group); and (6) low-K calc-alkaline polymodal basaltic andesite-dacite-rhyolite (upper Black Canyon Creek Group). This chemical evolutionary sequence shows that the contents of K_2O and Na_2O (and other trace and incompatible elements such as Rb that follow alkali enrichment) progressively increased as the Prescott volcanic arc evolved.

When these alkali-enrichment trends (see figs. 5 and 6 of P. Anderson, this volume) are compared to the southeasterly spatial progression of volcanism across the Prescott belt (see figs. 4 and 7 of P. Anderson, this volume), it is evident that *formative volcanism of the Prescott belt shows a clear polarity of increasing alkalies to the southeast.* Similar alkali polarities across the Japan arc first led Kuno (1959, 1966, 1968) to infer that Japan's petrologic-chemical provinces were generated by subduction of oceanic lithosphere under its arc, with the direction of increasing alkalies pointing down-dip of the subducted slab. Comparable alkali polarities across the Prescott arc imply that it was generated by subduction of Proterozoic oceanic lithosphere and that the southeast polarity of increasing alkalies points down-dip of its subducted paleoslab. Thus, volcanic provinces of the Prescott arc evolved sequentially from southeastward-dipping subduction of Proterozoic oceanic lithosphere under the arc, as an ocean basin that once lay to the northwest was consumed.

This setting explains the shifting axis of major volcanism across the Prescott arc with time (P. Anderson, this volume, fig. 7). Magmas produced by the subduction event rose to perforate Proterozoic oceanic crust and build up the arc in stages shifting progressively southeast. The Bradshaw Mountains Group formed first as the lowest K tholeiitic province to the northwest, then the Mayer Group developed next as a low-K tholeiitic volcanic province at a second central axis, then the Black Canyon Creek Group formed a third alkali-enriched tholeiitic province at a third volcanic axis to the southeast.

This southeastward sweep of volcanism likely continued into the younger part of the Central Volcanic Belt to form the calc-alkaline Union Hills Group volcanic province at a fourth major volcanic axis to the southeast. The Union Hills Group's calc-alkaline chemistry and post-Black Canyon Creek Group age support the successive southeast younging and alkali increase of all volcanic provinces produced by this early event of subduction. Unlike the Japan arc, however, where an alkalic magmatic province lies farthest inboard from the trench, the Union Hills Group calc-alkaline province appears to be the farthest inboard province of the Proterozoic suite. A suite chemically appropriate for an alkalic province—the felsic ignimbrite suite—exists in the eastern Central Volcanic Belt but relates to a different, younger event of subduction.

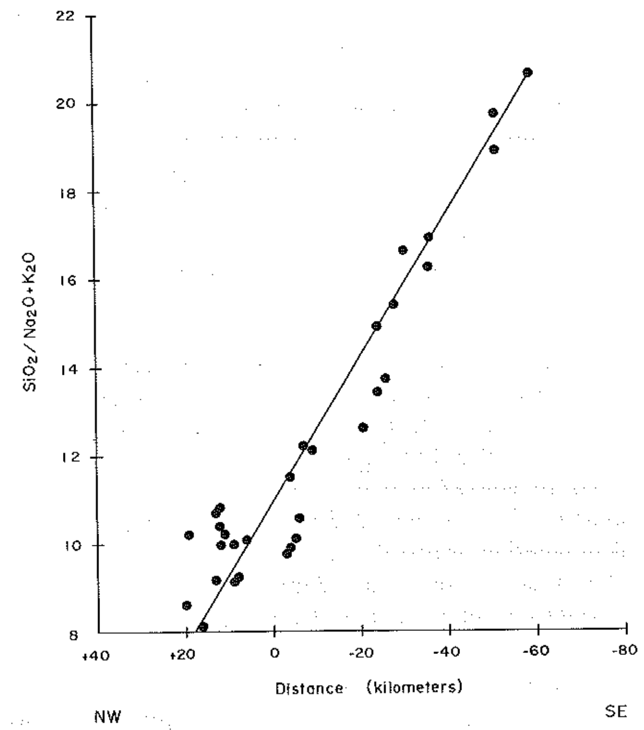

Figure 5. Alkali-enrichment plot of distance vs. silica-to-alkali ratio in the older pretectonic plutonic rocks of the Prescott-Jerome arc. The silica-to-alkali ratio is a measure of alkali enrichment across a conventional alkali-silica plot, and crosses the tholeiitic, calc-alkaline, and alkaline fields with decreasing numerical value of the ratio. Wide scatter at the low numerical end is attributed to the fact that some published analyses from the northern Bradshaw Mountains are from syntectonic, rather than strictly pretectonic, plutons. The distance axis is the distance in kilometers either northwest (positive numbers) or southeast (negative numbers) of a central reference point in the Mayer area, and locations of all analyses were projected onto a single plane trending NW-SE through the volcanic belt. An analogous diagram for the preceding formative volcanics of the Prescott-Jerome belt is the inverse of this diagram and has the opposite slope; this supports generation of the volcanics by an earlier event of southeast-dipping subduction, opposite to northwest-dipping subduction that generated the plutonic rocks. The volcanic diagram shows more scatter due to enhanced alkali mobility in the subaqueous depositional setting of the volcanic rocks, a factor not affecting plutonic systems.

The Ash Creek Group in the Jerome belt was an integral part of the Prescott intraoceanic arc and was spatially linked to the Prescott belt from inception. Old gabbros and gabbro-diorites intruded the oldest tholeiites of both belts concurrently, providing an early petrogenetic link, and volcaniclastics shed from both belts were interstratified throughout evolution of the Prescott-Jerome arc, from the oldest deposits in the northwest to the youngest in the southeast. However, the Jerome area lay east of the earliest part of the Prescott arc during subduction, and its oldest tholeiites are more alkali rich than comparable rocks in the Prescott area (figs. 5 and 6 of P. Anderson, this volume). Thus, alkali increases down-dip of the subducted slab were reflected in chemical differences even between the earliest-formed tholeiites.

Timing of Subduction

Because all formative volcanic provinces of the Prescott-Jerome arc are consistent with a single event of southeast-dipping subduction, the time span of evolution of the volcanic provinces date the duration of subduction. However, the oldest and youngest formative sequences of the Prescott-Jerome belts are undated. Although the upper Mayer and Ash Creek Groups have 1780-1760-Ma ages, a substantial history of volcanic evolution occurred before and after that interval: the Prescott-Jerome arc's early evolution occurred before 1780 Ma; the Black Canyon Creek Group formed after 1760 Ma, and the Union Hills Group province to the southeast evolved after all tholeiitic provinces to the northwest. Thus, an extensive formative volcanic evolution is recorded in the Central Volcanic Belt, including major intervolcanic unconformities signifying hiatuses of several million years. The relative chronology (P. Anderson, this volume) suggests that the Prescott-Jerome arc evolved from 1800 to 1745 Ma and the Union Hills Group formed from 1740 to 1730 Ma, but perhaps as late as 1725 Ma.

At or just prior to 1740 Ma, southeast-dipping subduction under the Prescott-Jerome intraoceanic arc ceased, and part of an ocean basin not fully subducted was preserved to the northwest between the Prescott-Jerome arc and the Archean craton edge to the north. At that time, the Prescott-Jerome arc was mostly shallowly submerged, but two volcanic edifices in the eastern Black Canyon Creek Group projected as islands above sea level. Thus, by 1740 Ma the formerly submerged Prescott-Jerome arc had developed into a partly emergent intraoceanic island arc. All formative volcanic sequences had developed, all early gabbro-diorite bodies had intruded, and small granodiorite plutons had crystallized beneath subvolcanic magma chambers related to dacitic extrusive activity. However, no major batholiths or plutonic complexes had been emplaced. Nevertheless, the timing between the last volcanics and the first major plutons is close and partly overlaps, because felsic magma chambers of some volcanic suites crystallized as pretectonic plutons.

Alternative Tectonic Setting for Younger Portion

Although the Union Hills Group is appropriate as the farthest inboard calc-alkaline province of the early southeast-dipping subduction event, another tectonic setting may be more likely, depending on the exact age range of the Union Hills Group relative to termination of southeast-dipping subduction. Because no isotopic age data exist for the Union Hills Group, the following alternative tectonic setting appears equally plausible from stratigraphic relationships.

The early intraoceanic arc that formed by southeast-dipping subduction may have been limited just to the tholeiitic volcanic provinces of the Prescott-Jerome belts. The southeastern limit of the tholeiitic province at the edge of the Cherry Springs batholith (including early volcanics in the Gibson Creek complex later split off by Verde River granite emplacement) could have been the southeast limit of the intraoceanic island arc. After the end of the southeast-dipping subduction event, the polarity of subduction flipped to dip northwest under the Central Volcanic Belt.

At inception of this new northwest-dipping subduction event, the Union Hills Group may have evolved as a chain of calc-alkaline volcanoes along the submerged southeast edge of the preexisting Prescott-Jerome arc. Unlike the preexisting arc, the new calc-alkaline volcanic chain did not form strictly in an open-ocean setting because it was built beside the already emergent Prescott-Jerome arc. Thus, the Union Hills Group volcanic chain may have formed a new southeast front to the older arc, which made the Prescott-Jerome arc a fully emergent continent-margin volcanoplutonic arc.

In this continent-margin arc tectonic setting, the Union Hills Group chain comprised a series of discrete but laterally interfingered submarine volcanic centers that shed turbidites and volcaniclastics into deep-ocean basins. The thick sequences of more mature volcaniclastics, turbiditic siltstone, graywacke, subgraywacke, and pelite that lie southeast of Union Hills Group exposures and extend throughout the Phoenix and adjacent areas (fig. 3) represent a clastic wedge that fronted the arc. By analogy to modern arc settings, this wedge comprised a forearc basin deposit.

Evidence for basement to the Union Hills Group volcanic chain supports this arc-front tectonic setting. Although later granitic rocks intrude the Union Hills Group and eradicate its basement in most places, basement remnants are mainly gabbroic. The Union Hills Group volcanic chain did not form on a basement of Prescott-Jerome volcanic rocks, because such older stratified rocks are absent throughout the younger belts—*the Union Hills Group itself is the basal volcanic assemblage.* Its basement was not an unevolved oceanic crust like that under the Prescott-Jerome arc, nor was it an evolved felsic crust. The chemistry of the gabbro-pyroxenite fragments suggest that the basement was probably a transitional mafic crust that formed at the back of the preexisting Prescott-Jerome arc and that was later punctured by Union Hills Group magmas to build the front of the new arc.

This alternative tectonic setting—that the Union Hills Group volcanic chain evolved as a continent-margin island arc just offshore from the southeast edge of the Prescott-Jerome arc—accounts for the lack of a primitive oceanic basement, because the volcanic chain was built onto the southeast front of an existing island arc. The calc-alkaline chemistry of the Union Hills Group fits the continental-margin arc setting perfectly and is rare in intraoceanic island arcs built upon oceanic crust. Only at the northwest edge of the new volcanic chain would the new volcanic deposits be expected to interface with previous ones, which is exactly what one finds: in the New River area, Union Hills Group rocks lie just east of the Black Canyon Creek Group and may overlap them; and near Payson, East Verde River volcanics lie adjacent to the Gibson Creek complex and may also overlap both it and its sheeted diabase dikes.

Arc Plutonism

After 1740 Ma, the older part of the Central Volcanic Belt near Prescott was pervaded by pretectonic and syntectonic, arc-related tonalite-granodiorite-monzogranite batholiths and plutons. Whereas the chemistry of host volcanic rocks and consanguineous early mafic plutons shows a southeast alkali-enrichment trend, the chemistry of these 1740-Ma and younger plutonic suites shows a very clear alkali-enrichment trend *exactly the opposite in polarity,* namely to the northwest (fig. 5). Thus, source magmas of 1740-Ma and younger pretectonic and syntectonic plutonic suites were generated by a Proterozoic subduction event that dipped northwest under the Prescott-Jerome arc. Therefore, a flip in subduction dip from southeast to northwest occurred after formative volcanism of the Prescott-Jerome arc but before inception of major plutonism in the arc.

Emplacement of early pretectonic plutons and batholiths into the edges of the Prescott-Jerome arc caused disruption of the volcanic pile, assimilation of basal arc material, incorporation of basement fragments into plutonic bodies, and the first substantial crustal thickening of the volcano-plutonic arc. This major tonalite-granodiorite plutonic event continued from 1740 Ma to at least 1720 Ma, during which time it transformed the Prescott-Jerome arc into a thick, stable volcanoplutonic arc that became fully emergent by 1720 Ma.

Plutonic rocks emplaced into the Central Volcanic Belt between 1740 and 1720 Ma include the pretectonic Minnehaha and Wilhoit Granodiorites, the Cherry Springs batholith, the Little Squaw Creek Migmatite Complex, and the Gibson Creek Complex. Later pretectonic bodies include younger phases of the Cherry Springs batholith and syntectonic plutons along the edge of the arc, including the Prescott, Iron Springs, Johnson Flat, Longfellow Ridge, and Horse Mountain bodies. Huge granodiorite-monzogranite batholiths were subsequently emplaced west of Prescott in the Weaver Mountains, Yarnell, and Skull Valley areas. The earlier plutons were concentrated southeast of the volcanic arc, whereas the younger syntectonic and late-tectonic bodies were concentrated northwest of it; this implies a general northwest progression of plutonism with time.

Broadly overlapping in time with emplacement of these suites in central Arizona was pervasive batholith emplacement throughout the Northwest Gneiss Belt: its major ca. 1720-Ma granodiorite-monzogranite batholiths occupied more than half the crust, converted stratified host rocks to high-grade gneisses and migmatites, and caused major deformation and metamorphism throughout the belt. While the first major batholiths were intruding the Northwest Gneiss Belt, late plutonic suites intruded the Central Volcanic Belt after the first pretectonic suites had already stabilized it. Then,

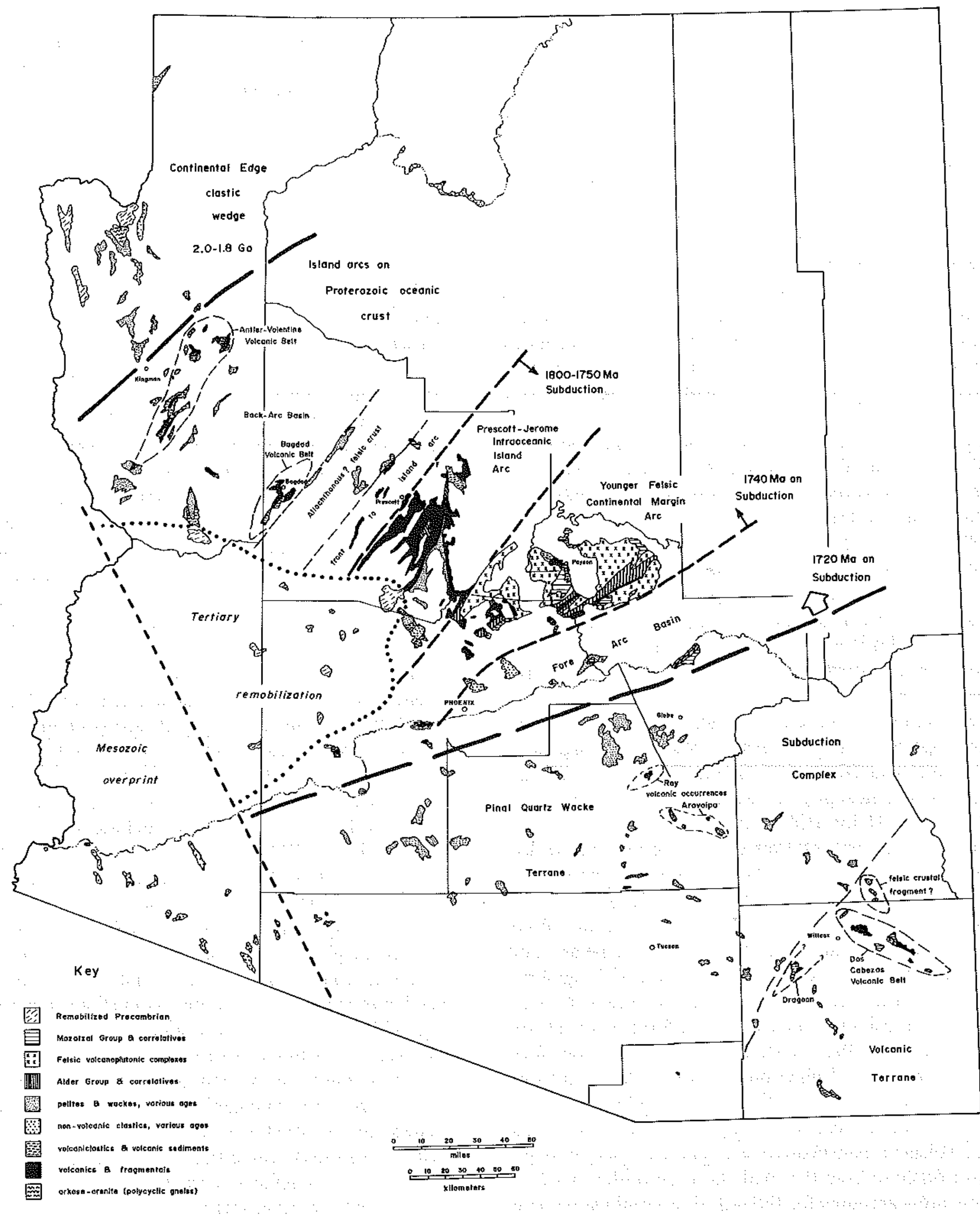

Figure 6. Proterozoic plate tectonic evolution of Arizona depicted from a map perspective. The base is the same as figure 3, but different trench positions for the southeast- and northwest-dipping subduction events are shown for the age ranges listed. The arrows point down the dip of the subducted paleo-oceanic slab for each subduction event. Proterozoic oceanic basement lay between the two heavy long-dashed lines under the volcanic belts of central and northwest Arizona, but the nature of basement on either side of this central region is unknown. 1800- to 1750-Ma southeast-dipping subduction gave rise to volcanics that formed the Prescott-Jerome intraoceanic island arc and parts of the eastern Central Volcanic Belt. From 1740 Ma on, northwest-dipping subduction produced plutonism throughout the Prescott-Jerome arc and all contemporaneous and younger volcanism in the eastern part of the Central Volcanic Belt, as well as other parts of southeast Arizona. The trench axis stepped southeast in discrete events, one recognizable where shown at about 1720 Ma or earlier, another possible transitory position is located near Willcox, and the final post-1695-Ma trench axis for northwest dipping subduction lies out of Arizona.

when plutonism had mostly ceased in central Arizona, major plutonism continued in the northwest. Thus, comparable events occurred later in the Northwest Gneiss Belt than in the Central Volcanic Belt, which suggests that the protracted plutonic event swept from southeast to northwest in a series of stages across the orogen, mirroring the northwest-dipping subduction under it. The northwest-progressing wave of batholith generation was produced by a single major event of northwest-dipping subduction under northern Arizona that culminated in the north at about 1720 Ma.

Although figure 5 shows a northwest alkali enrichment of plutonic rocks only across the Prescott-Jerome arc, a similar northwestward increase in alkali contents of plutonic rocks exists across the Northwest Gneiss Belt but is less well defined because of the markedly different host-source crusts in various segments of the belt. Alkali contents increase from those values found in the northwest Prescott region, but reach a maximum, beyond which further alkali enrichment was buffered by proportionately more partial melting. Hence the steep slope of figure 5 flattens in northwest Arizona where 50 percent or more of the preexisting crust evidently was melted to form the batholiths.

Magma Generation and Subduction Inferences

The northwest-dipping subduction event under the Prescott-Jerome arc and contiguous terranes to the northwest spanned plutonic evolution, stabilization, and "cratonization" of central and northwest Arizona's Proterozoic crust from 1740 Ma on. The southeast limit of 1740-1720-Ma plutonic rocks, however, marks only the depth on the subduction zone where hydrous magmas were generated, not the location of the trench. Partial-melting considerations show that a zone of volcanic magma generation lies between the plutonic zone and the trench, a zone occupied by the Union Hills Group, whose distinctive calc-alkaline signature, age limits, and setting at the submerged front of the volcanoplutonic arc strongly suggest coevality with the 1740-1720-Ma plutonic suites.

The dip and depth to the Benioff zone and hence a paleotrench position at the submarine site of subduction, can be inferred from the alkali contents of magmatic rocks, (Dickinson, 1975; Hatherton and Dickinson, 1969). Benioff-zone depth and K_2O-SiO_2 indices for modern arcs can be used (e. g., P. Anderson, 1986) to suggest that magmas may have been generated at 100-200 km depths on a paleosubduction zone dipping 50-60° northwest under central Arizona, and that the Salt River line south of Phoenix would mark the paleotrench site (fig. 3). Such calculations, however, assume that Proterozoic subduction systems were chemically identical to modern ones, which is definitely not the case. Subtle chemical differences between Proterozoic and modern oceanic crusts mean great differences in magma chemistry for particular depths of generation, because fractional melting enhances differences in alkalis and related trace elements. Moreover, the thermal structure of Proterozoic subduction zones would have been different from modern ones because of subtle chemical differences in oceanic lithosphere. Hence an exact analogy between Proterozoic and recent arcs is not correct. Instead, the paleotrench position for 1740-1720-Ma subduction can be better deduced from the supracrustal geologic record.

Forearc Basin, Trench Position, and Southeastward Trench Evolution

Neither Union Hills Group arc volcanics nor pretectonic plutons related to 1740-Ma northwest-dipping subduction occur southeast of a line from Phoenix to Young. Instead, the oldest rocks are distally related to and younger than Union Hills Group arc volcanism; they comprise a thick clastic sequence that evolved from turbidites to immature quartzites, and which was locally overlain at Four Peaks and Hess Canyon by felsic ignimbrites and Mazatzal strata. The sequence is devoid of formative mafic volcanics and early plutons; all diorite-granodiorite bodies are younger. This thick clastic wedge represents a forearc basin fronting the 1740-Ma magmatic arc, receiving clastics shed oceanward from the arc. The clastic wedge extends no more than 20 km southeast of the Salt River, so the forearc basin ended south of Lake Roosevelt and Hess Canyon in the east, and along the Salt River near Gila Bend in the west (figs. 1 and 3).

Because the front of a forearc basin is effectively the trench site, the Proterozoic paleotrench is inferred to have been positioned 15 km south of the Salt River during early evolution of the forearc basin. The northern limit of Pinal Schist supports this trench position: everywhere southeast of that Salt River line lies Pinal Schist (fig. 6), whose formative sedimentary units bear almost no lithologic or tectonic similarity to sequences in the Central Volcanic Belt. Pinal Schist was originally unrelated to the 1740-1720-Ma contiguous terrane of central and northwest Arizona; only the post-1700-Ma deposits in central and southeast Arizona were linked.

Stratigraphic evidence indicates that most sediments in the forearc basin correlate to younger evolutionary stages of the Union Hills Group, when upper felsic volcaniclastics and Alder Group strata were deposited. However, some turbidites in the forearc basin are overlain by and thus predate Alder strata. Therefore the trench was likely positioned along the heavy dashed line in figure 6 after the earliest Union Hills Group volcanics formed, and during Alder deposition. Hence the forearc basin evolved from about 1730 to 1710 Ma.

In the early stages of 1740- to 1720-Ma northwest-dipping subduction the forearc basin was incipient, so the trench was likely much closer to the arc front (bounding the Union Hills Group along the short dashed line in figure 6). By 1720 Ma, however, the forearc basin was well established, so the trench had moved southeast of the basin (fig. 6), and by that time, some Pinal Schist may have accumulated as trench-fill melange. After 1720 Ma the rest of the Pinal Schist terrane and other rocks in southeast Arizona were accreted to the forearc basin, and the trench moved far to the southeast. Therefore at least three distinctly different trench positions existed during the history of northwest-dipping subduction: each position was part of a southeast stepping of subduction as various crustal segments were accreted to the edge of the continent-margin arc.

Younger Tectonic Setting

After 1720 Ma, the Central Volcanic Belt assumed a tectonic setting entirely different from its earlier, dominantly intraoceanic one. By 1720 Ma, all formative volcanism had ceased, pretectonic batholiths and plutons had been emplaced, some early plutonic rocks had been uplifted and their volcanic cover locally stripped off, and the Northwest Gneiss Belt had experienced its major batholith emplacement. Thus, both belts had been thickened and stabilized by major plutonism, and the entire Proterozoic crust from the front of the Central Arizona volcanic arc to the Archean craton far to the north was partly or mostly emergent. The Central Volcanic Belt had become a truly evolved continent-margin arc physically attached to the Archean craton by semicontinuous land masses.

Sedimentation at this time was limited to narrow structurally downwarped or downfaulted troughs along edges of major pretectonic plutons and batholiths, in which felsic volcaniclastics distinctively akin to one another accumulated, even though depositional troughs were far from each other. The trough in the Antler-Valentine volcanic belt received similar but coarser clastics than troughs of Texas Gulch strata in the Prescott-Jerome belt. The calc-alkaline nature of felsic effusive volcanism in the troughs reflected the evolved nature of the crust at 1720 Ma and later. The troughlike nature of these deposits on the emergent arc contrasted to the shallow-water, open-marine setting of the Alder Group, which accumulated at the front of the continent-margin arc during the same time interval. A shallow basin in the middle of the Union Hills Group volcanic chain received the southwesterly transgressive Alder suite, which began with purple shale and evolved to submature quartzites before it was overwhelmed with felsic debris heralding the impending ignimbrite eruptions.

Alkali-calcic ignimbrites erupted only along the shallowly submerged front of the continent-margin arc; their northwesterly spread was stopped by a major scarp at the edge of the Cherry Springs batholith that was produced by structural subsidence of the arc front and uplift of the batholith just prior to progradation of the Alder Group into the structural trough. Just like granodiorite batholiths that earlier thickened and stabilized the crust of the Prescott-Jerome arc, emplacement of huge granitic magmas into the arc front at 1710-1700 Ma thickened and stabilized it as alkali-calcic granite batholiths crystallized beneath their coeval, comagmatic ignimbrite carapaces. By 1700 Ma, after all ignimbrite outpourings, the continent-margin arc became fully emergent, and only the forearc basin to the southeast remained submerged.

Mazatzal Group deposits formed in immediate response to erosion of the subaerial ignimbrite fans back to sea level. The main Mazatzal fluvial channel extended from its head in the older Prescott-Jerome arc down to beach, back-bay, and littoral environments in the Mazatzal Mountains that had transgressed over earlier Mazatzal fluvial settings. Shallow open-marine conditions persisted in the forearc basin to the southeast, and finally transgressed back over the arc front as it was eroded back to sea level. No post-Mazatzal depositional record is found in central Arizona, which implies continued gradual uplift and erosion of the arc front after Mazatzal deposition, and full emergence of the entire arc, as orogeny continued in southeast Arizona to as late as 1620 Ma.

TECTONIC SETTINGS OF THE NORTHWEST GNEISS BELT

The Northwest Gneiss Belt is composed of many different crustal segments with differing tectonic histories, so no single tectonic setting can adequately account for the entire belt. The most likely tectonic settings for each of the five crustal segments are considered in sequence from northwest to southeast.

Craton-Margin Clastic Wedge

Polycyclic gneisses with arkosic protoliths in far northwest Arizona appear unrelated to volcanism in adjacent belts to the south and were derived by erosion of the southwest edge of the Wyoming Archean craton in the north. Because no granitoid crust of Archean or Proterozoic age existed southeast of the craton at the time the arkoses were deposited, the tectonic setting of the arkoses was a craton-margin clastic wedge that tapered toward the craton edge. The clastic wedge formed initially by rifting of an Archean proto-continent: one part remained as the Wyoming craton and the other part separated from North America. As the rift grew, the clastic wedge enlarged by prograding both back over the subsiding craton margin and outward over early Proterozoic oceanic crust that formed the basement to the rift.

This tectonic setting indicates that Proterozoic oceanic crust underlay the oceanward portion of the craton-margin clastic wedge and existed everywhere to the southeast after proto-continental rifting. Field evidence for basement confirms this tectonic setting: no mafic-ultramafic remnants are found in the northern arkosic paragneiss region, some occur at the boundary zone near and northeast of Kingman, and evidence for oceanic crust is abundant southeast of Kingman. Thus, the boundary (fig. 6) is by nature gradational because arkosic sediments lapped over the edge of transitional crust at the rifted margin. The craton-margin clastic wedge necessarily formed before island arcs evolved to the southeast, because it was related to initial rifting that opened the ocean basin and that must have preceded convergent tectonics in that basin.

Antler-Valentine Volcanic Belt

The Antler-Valentine volcanic belt is a small, unevolved, relatively primitive belt of minimally fractionated tholeiites whose trimodal sequence is more K rich than primitive, Mg-rich bimodal tholeiites in oldest parts of the Prescott belt, or Fe-rich bimodal tholeiites of the Bagdad belt. The Antler-Valentine chemistry compares more closely to later tholeiitic stages of the Prescott belt that had a more evolved tectonic setting. Unlike the Bagdad and Prescott belts, the Antler-Valentine belt contains only a single tholeiitic volcanic cycle, which indicates that volcanic magma generation under the belt was relatively short lived or prematurely aborted.

Three tectonic settings are possible for the Antler-Valentine belt: (1) It may have been conceived as an intraoceanic arc that was later accreted to the craton-margin clastic wedge northwest of Kingman (fig. 6). This model requires a suture between polycyclic gneisses of the clastic wedge and volcaniclastics distally related to the volcanic belt, which is unlikely because the contact is probably depositional. (2) The Antler-Valentine and Bagdad belts may have been originally joined and later detached by plutonism, but this is also unlikely because of different volcanic evolutions, structures, chemistries, and also possibly ages of the belts, plus the presence of volcaniclastics between them. (3) The tectonic setting most appropriate for the Antler-Valentine belt is an incipient island arc that formed in place along the edge of the craton-margin clastic wedge by a brief event of subduction that dipped northwest under the cratonic wedge, before being terminated by changing plate motions. This model accounts for the presence of (a) clastics shed distally from the volcanic belt into a retroarc basin to overlap earlier deposits of the craton-margin clastic wedge; (b) polycyclic deformation in the clastic wedge; and (c) a whole suite of early plutonic rocks possibly intruding the polycyclic gneisses. Such plutons have not yet been singled out in far northwest Arizona nor recognized as products of a subduction event that preceded the one under central Arizona.

The tectonic setting of crust between the Antler-Valentine and Bagdad volcanic belts is poorly known because it now exists only as fragments between huge batholiths. However, its general character is oceanic: deep-ocean-floor tholeiites occur with thin pelagic sediments to comprise what is most likely the upper part of Proterozoic oceanic crust. A thicker clastic section to the southwest probably signifies volcaniclastic detritus shed laterally from the Antler-Valentine belt into the deep ocean basin (fig. 6). With Proterozoic oceanic crust lying south of the autochthonous Antler-Valentine belt, the Bagdad belt formed in an intraoceanic setting allochthonous to the edge of the Archean craton, and was later accreted to the Proterozoic margin.

Bagdad Volcanic Belt

The Bagdad belt is similar to the Antler-Valentine belt but differs in chemical details and tectonic setting. The belt represents a magma series that differentiated along a true Fe-rich tholeiitic trend, both in its extrusive deposits and its large, layered gabbro-anorthosite body. The belt is geochemically a single major magma series but can be subdivided into an early primitive cycle and a later felsic and fragmental cycle. The layered gabbro-anorthosite complex, which lies near the base of the belt, may be a lopolith subvolcanic to the early mafic cycle or part of the upper stratified and differentiated section of oceanic crust that once floored the volcanic belt.

Because of its unique tholeiitic differentiation history, the Bagdad belt may have initially formed at an ocean-ridge spreading center and later evolved as oceanic islands in a deep intraoceanic setting after moving away from the spreading center. The Bagdad belt is less likely to represent an island arc formed above a subduction zone that lasted for any significant time because of its small size, lack of major geologic-geochemical evolution from an original tholeiitic parentage, lack of a clear chemical polarity, and absence of major fans of volcaniclastic detritus shed from the volcanic belt. It is also unlikely that the Bagdad belt was once linked to the Antler-Valentine belt, because the two belts have different tectonic settings, are separated by intervening crust, and are probably of slightly different ages.

Other Crustal Segments

Between the Bagdad belt and the west edge of the Prescott volcanic belt is a felsic crustal segment apparently devoid of mafic volcanic rocks, lacking all evidence for basement, and seemingly made up of only fine-grained K-Na-Al-rich pelite-wacke protoliths. Rocks of this segment are generally in sharp, locally tectonic, contact with mafic volcanics of adjacent volcanic belts and are unconformable upon distal volcanics of the Prescott belt only along parts of the eastern contact. The entire sedimentary assemblage, however, does not unconformably overlie such mafic volcanics: mafic volcanic basement is not exposed throughout the crustal segment nor found as pendants in plutonic bodies that pervade the region. Enclaves and pluton chemistry reflect the pelite and wacke sedimentary crust of the felsic segment. Tectonic translation of an allochthonous sedimentary crust into position by strike-slip faults is not a reasonable explanation for this segment because local unconformable relations and no structural regeneration at contacts require some autochthoneity, and the absence of basement in the segment remains unexplained.

The most likely origin for this pelitic crustal segment is revealed in the tectonic evolution of the Prescott-Jerome arc itself. The initial event of southeast-dipping subduction under the Prescott-Jerome intraoceanic arc that generated its formative volcanic sequences left mafic volcanics west of Prescott ("front to island arc" on fig. 6) as the remains of ocean-floor tholeiites upon which the Prescott arc was built. Consequently, oceanic crust that once lay northwest of the Prescott-Jerome arc must have been consumed by subduction, and hence a paleotrench site must have existed northwest of Prescott. Although the trench site may not be preserved, the subduction complex of supracrustal material should be, and it is exactly this material that most likely makes up the pelitic crustal segment between the two volcanic belts. The supracrustal assemblage could represent oceanic sediments decoupled from a subducted oceanic slab, deep-water turbidites that filled the trench, or both, and would include forearc basin sediments that prograded over both the subduction melange and the trench site, thereby covering the trench position at upper crustal levels.

The key aspect of a subduction-complex origin is that it is the only one to explain the lack of basement in the segment, because subduction melanges are usually not floored by igneous basement. The model also explains the observed unconformable eastern contact, because youngest deposits of a forearc basin typically overlap both volcanics at the arc front and subduction-complex melange. The fine-grained immature clastics of pelite-wacke composition are lithologically and chemically appropriate for a subduction complex and forearc basin, and the narrow width of the pelitic crustal segment (fig. 6) accords well with an inferred steep southeast dip of the subduction zone under the Prescott-Jerome arc. This southeast-dipping subduction event was terminated by collision of the Bagdad volcanic belt with the subduction complex.

TECTONIC SETTING OF THE SOUTHEAST SCHIST BELT

The Southeast Schist Belt also contains different crustal segments with different tectonic settings and origins, not all of which are fully understood because of the present lack of detailed data, age constraints, and the isolated nature of exposures in the belt. Each major crustal component is discussed here in sequence from northwest to southeast.

Pinal Schist Terrane

In contrast to earlier concepts that all Proterozoic rocks in southeast Arizona belong to Pinal Schist, the Pinal Schist terrane is redefined here as a lithologically distinctive assemblage of quartz wacke, pelite, and tuffaceous siltstone that extends from the Pinal Mountains southeast to about Willcox, but does not extend farther southeast where rocks of more direct volcanic origin predominate. Thus, a major boundary of profound tectonic significance to the Southeast Schist Belt exists in southeast Arizona between true Pinal Schist and more southerly volcanic and volcaniclastic assemblages (fig. 6).

The distinctive Pinal Schist lithology is remarkably similar to that of the felsic crustal segment between the Prescott and Bagdad volcanic belts; prior to metamorphism, the fine-grained protoliths of pelite, wacke, and felsic tuffaceous siltstone origin in both regions may have been in all key respects identical. The main contrast is that Pinal Schist is more Si rich and K rich because it was derived from more evolved felsic tuffaceous sources. Another key similarity is that the Pinal Schist terrane is also devoid of evidence for a basement, either of mafic oceanic or felsic granitoid character, and this lack of basement must be accounted for in its tectonic setting.

The most likely tectonic setting for Pinal Schist is revealed in the tectonic evolution of the adjacent Central Volcanic Belt. The main event of northwest-dipping subduction under central Arizona started at about 1740 Ma and lasted for at least 60 Ma, shifting location from its inception near the front of the Central Volcanic Belt oceanward to include a forearc basin by no later than 1720 Ma (fig. 6).

After the 1720-Ma trench position, four main tectonic settings are possible: (1) the trench axis took a major jump to the southeast out of Arizona and into Mexico, and the Pinal Schist terrane evolved as a very wide forearc basin deposit between the old forearc basin and the new trench position far to the south; (2) the trench axis took a similar jump, but the Pinal Schist terrane evolved as an interarc basin between the original continent-margin arc of central Arizona and a new arc in southeast Arizona; (3) the trench axis jumped southeast as an allochthonous Pinal Schist terrane was accreted to the edge of the forearc basin in a single event; or (4) packages of Pinal Schist were gradually accreted in the form of a growing subduction-complex melange as the position of the trench axis stepped incrementally southeastward.

The main time constraint on these tectonic settings is that Proterozoic rocks throughout southeast Arizona, including the most southerly volcanic belts in the Dos Cabezas and related areas, presumably must have been in place by 1695 Ma, because at this time quartz-phenocrystic rhyolites were extruded in the Dos Cabezas and Ray-Aravaipa volcanic belts more or less contemporaneously with 1695-Ma ignimbrites in the younger portion of the Central Volcanic Belt. Overlying quartzites of Mazatzal age also demonstrate that continuous crust existed between the Central Volcanic Belt and the Dos Cabezas volcanic belt.

The period between the 1720-Ma trench position (fig. 6) and the 1695-Ma time when all parts of southeast Arizona were amalgamated under a single northwest-dipping subduction zone is 25 Ma, adequate for either incremental or instantaneous growth of an accretion complex, especially considering that the dip of a subduction is more important than duration of subduction in governing the volume of supracrustal sediments stacked into a subduction complex. As the trench axis stepped southeastward, whether abruptly or incrementally, the dip of subduction flattened, and volumetrically more supracrustal oceanic sediments became decoupled from the subducting oceanic slab to be accreted with trench sediments at the front of the growing subduction complex.

A tectonic setting of the Pinal Schist terrane as a subduction complex accounts for many otherwise enigmatic aspects of its geology: it explains (1) why no basement to Pinal Schist has been found over such a vast region: oceanic basement is decoupled, and its supracrustal sediments are laterally accreted into the subduction melange with other allochthonous crustal fragments; (2) why Pinal Schist is so lithologically uniform over such a huge region, and why no well-layered sequences or stratigraphic order have been identified: stratigraphy in subduction complexes is chaotic, imbricated, and sheared on a broad scale; and (3) why Pinal Schist and adjacent volcanogenic sediments of the Central Volcanic Belt are clearly different in lithology, yet essentially similar in felsic, quartz-rich, micaceous, volcanic material: detritus shed oceanward from a continent-margin arc would have been incorporated into the subduction complex that formed at the trench axis.

Ray-Aravaipa Volcanic Belt

The main feature that a subduction-complex setting for Pinal Schist does not explain is the Ray-Aravaipa volcanic center in the midst of the complex (fig 6). If volcanic rocks of this center had entirely tectonic contacts with adjacent Pinal Schist units, the accretion of the volcanic center as a unit into the subduction complex along with adjacent melangelike material would be reasonable. However, Ray-Aravaipa volcanics seem autochthonous with their adjacent Pinal Schist lithologies and show the usual decrease in volcanic component and gradual increase of clastic material away from the center. But boundaries are poorly exposed in the region, so future detailed work may show that the Ray-Aravaipa center contains volcanic slices tectonically imbricated within Pinal Schist, thus supporting a subduction origin for Pinal Schist.

Another key feature is that quartz-phenocrystic rhyolite flows in the Ray-Aravaipa volcanic center closely resemble similar rhyolites in the Mazatzal Mountains and Dos Cabezas volcanic belts and appear to be broadly coeval with them at about 1695 Ma. It is possible that the volcanic center was punctured through the subduction complex after accretion, but this is unlikely if distal tuffs of the felsic center are interbedded with Pinal Schist lithologies. Such a model necessitates two sedimentary sequences of discernibly different ages in the Pinal Schist terrane: an older sequence accreted as a subduction complex, and a younger sequence interleaved with 1695-Ma felsic volcanics. Although such a division has not yet been found, it certainly is a possibility in the poorly mapped, lithologically monotonous Pinal Schist.

Tectonic Settings

Because of apparent stratigraphic links between the Ray-Aravaipa belt and Pinal Schist, it seems unlikely that all of the Pinal Schist terrane was gradually accreted to the margin (tectonic setting 4, presented earlier). Tectonic settings (1) and (2), with Pinal Schist as a wide forearc or interarc basin, support original stratigraphic continuity in the Pinal terrane and avoid the problem of having diverse crustal fragments later unified by felsic volcanism and sedimentation, but both require an oceanic basement to the Pinal Schist basin, of which there is no evidence in the Pinal Schist terrane. In the forearc basin model, most of the Pinal Schist terrane could be a wide subduction complex without oceanic basement, but this setting conflicts with apparent stratigraphic continuity in parts of the terrane. Tectonic setting (3), that the Pinal terrane was accreted to the continent margin in a single event as a coherent crustal fragment, seems to avoid these problems but evades the basement problem: such a wide expanse of supracrustal sediments must have had some form of basement, and given their composition, it would have been oceanic. Detailed analysis shows that this "singular accretion" model carries the same problems as the other models.

With present data, the tectonic setting preferred for the Pinal Schist terrane is a combination of settings (1), (2), and (4) above. During its early history, the Pinal Schist terrane was a combination of a subduction complex growing at depth and an oceanward-spreading forearc basin deposit that lapped over the subduction complex. However, when the Dos Cabezas volcanic belt either formed along the continent margin to the southeast, or was accreted to it, the Pinal terrane became a retroarc basin with respect to this new southeastern volcanic arc. In a strict sense, however, the Pinal basin was an interarc basin, because volcanic activity continued at the front of the Central Volcanic Belt during the younger evolution of the Pinal basin. Thus, the younger felsic volcanism and subsequent clastic sedimentation occurred across the entire basin after all tectonic elements had been consolidated.

The rapid outward growth of the Pinal terrane and formation of the Dos Cabezas volcanic belt to the southeast significantly flattened the subduction-zone dip, at least at upper lithospheric depths, and thus promoted growth of a wide forearc-interarc basin in which fine-grained Pinal clastics accumulated. Intermediate to felsic volcanism was coextensive with sedimentation where the basin was locally pierced by subduction-generated magmas. These volcanics, plus most Pinal Schist sediments, would have formed mainly in the interval from 1710 to 1690 Ma, so the subduction-zone jump most likely occurred between 1710 and 1700 Ma, which initiated ignimbrite activity in the central Arizona arc.

Near the southeast limit of Pinal Schist in southeast Arizona, a sharp lithologic change occurs between typical Pinal Schist of the Johnny Lyon Hills and a very different mafic volcanic-graywacke assemblage in the Little Dragoon Mountains immediately to the southeast (fig. 6). This line may have been an intermediate trench position for early evolution of the Pinal basin, a model supported by local evidence for tectonic imbrication of turbidites and mafic volcanics. The line, however, more likely represents a tectonic boundary where the edge of the Dos Cabezas volcanic belt was either tectonically joined to, or structurally decoupled from, the Pinal Schist terrane. Everywhere southeast of that line, rocks of more direct mafic-felsic volcanic derivation predominate the oldest supracrustal strata; distal sediments are of obvious volcanic derivation in most places, are unlike typical Pinal Schist, and apparently only come into tectonic contact with Pinal Schist along that line.

Dos Cabezas Volcanic Belt

The Dos Cabezas volcanic belt and correlatives extend from the Little Dragoon and Dos Cabezas Mountains southeast into Sonora. The eastern Little Dragoon Mountains contain basaltic units disrupted into large blocks that seem to be tectonically imbricated with turbidite graywackes to the west; this is the best evidence known for a tectonic contact between the Dos Cabezas volcanic belt and the Pinal Schist terrane. The southern Pinaleno Mountains contain a felsic sedimentary terrane with granophyric or granitoid detritus that may also be in tectonic contact with Pinal Schist to the north, but younger reworking in this area obscures a clear picture of original Precambrian relationships.

The Dos Cabezas belt is chemically peculiar among Arizona Proterozoic volcanic belts because of its high-K tholeiitic to low-K calc-alkaline bimodal or trimodal character. The Dos Cabezas belt most likely did not form in an intraoceanic setting on mafic-ultramafic oceanic crust, as did other Arizona Proterozoic belts, but formed instead in a chemically more evolved setting, such as along a continental margin or upon crust like that of the Pinal Schist terrane itself. Alternatively, the belt may have been conceived distant from the Pinal continental margin and later accreted to it after formative mafic volcanism but prior to felsic volcanism. Because of its chemical and stratigraphic uniqueness, the Dos Cabezas belt is unlikely to have originally been part of the Central Volcanic Belt that was rifted away and later accreted to the Pinal margin.

Thus, possible tectonic models for the Dos Cabezas volcanic belt are: (1) generation of early mafic volcanism by subduction in a setting distant from the Pinal margin, accretion to that margin, and subsequent felsic volcanism and quartzose sedimentation; (2) total volcanic development in an allochthonous position, followed by accretion to the margin and quartzose sedimentation; (3) autochthonous development at the front of a broad Pinal forearc and interarc basin; (4) formation as part of the younger part of the Central Volcanic Belt, subsequent rifting away from central Arizona, and later accretion to the Pinal margin; (5) tectonic imbrication of the belt in a huge subduction complex; or (6) formation as an island arc just offshore from the continent margin, with the interarc Pinal basin behind it, a forearc basin fronting it, and the new trench position lying southeast of Arizona in Sonora; later partial collapse of the interarc Pinal basin and tectonic detachment of the arc from the retroarc basin would cause stratigraphic relations to be disrupted between the Pinal Schist terrane and the volcanic terrane, but not within the volcanic arc itself. This last model is considered the most likely at present and is applicable regardless of whether the Pinal Schist terrane was originally a subduction complex, a forearc basin, or both a forearc then retroarc basin.

Rhyolites in the Dos Cabezas and Ray-Aravaipa belts and 1700-Ma felsic ignimbrites of the Central Volcanic Belt all appear to have been generated by the one event of subduction that dipped shallowly northwest under southeast and central Arizona. Shallow convergence explains 1695-Ma ignimbrite extrusions at the front of the central magmatic arc 350 km inboard from the trench, as well as contemporaneous felsic volcanism in the intervening Pinal basin. Continued shallow convergence from 1695 to 1620 Ma accounts for 1695-1660-Ma pretectonic diorite-granodiorites in the northern Pinal terrane, the 1660-1650-Ma Sunflower granite syntectonic to Mazatzal deformation in central Arizona, and 1625-1620-Ma emplacement of granodiorites syntectonic to younger deformation in the southern Pinal terrane. This southeastward plutonic progression is consistent with southeastward crustal growth, alkali increase down-dip of the subducted slab, and lower crustal heating throughout the orogen to produce deformation and metamorphism that progressed southward with time.

PROTEROZOIC PLATE TECTONIC HISTORY

The Proterozoic tectonic evolution and various plate tectonic settings of the many different Proterozoic crustal belts of Arizona have been summarized above using modern plate tectonic terms. Despite the genetic connotations of such terms, a detailed analysis of the fundamental components of Proterozoic plate tectonics, such as island arcs, oceanic crust, and other features, show that they were significantly different than their counterparts in either modern or Archean systems (P. Anderson, 1976, 1986, n. d.).

The foregoing analysis shows that only a few Proterozoic plate tectonic configurations were possible for Arizona between 1900 and 1650 Ma. This leads to establishment of a plate tectonic history for the Arizona Proterozoic that is most likely at present, given the less precise constraints for southeast Arizona. This analysis is actualistic to modern plate tectonics, meaning that similar, not the same, processes and features are implied, and it should be viewed in light of the similarities and differences noted at the end of this paper. The following summary is referenced to figure 6, which shows the timing of Proterozoic subduction in Arizona, and to figure 7, which schematically shows plate

tectonic configurations most applicable to the Arizona Proterozoic at seven key times between 1900 and 1650 Ma.

The Proterozoic tectonic evolution of Arizona began about 1900 Ma ago with rifting of the southeast edge of the Archean Wyoming craton and separation of part of the original Archean continent from North America (fig. 7). Arkosic detritus from the Wyoming craton was shed southeastward into a southeast-facing shelf-slope clastic wedge along the craton margin. Open-ocean conditions were established throughout the rest of Arizona as Proterozoic oceanic crust grew in a widening ocean basin that accommodated the crustal rifting process. The upper part of this oceanic crust consisted of low-K tholeiitic pillowed basalt flows, deep-sea pelagic sediments, and chert, locally with Proterozoic analogs of spilite and keratophyre, and differentiated layered mafic-ultramafic intrusions.

In this open ocean basin distant from the Wyoming Archean craton, a primitive intraoceanic island arc (the early Prescott-Jerome arc) was conceived on oceanic crust at about 1800 Ma above a subduction zone that for 60 m.y. dipped steeply southeast, probably under the crustal fragment earlier detached from the Archean Wyoming block. The Prescott-Jerome intraoceanic arc evolved southeastward from bimodal tholeiitic basalt-rhyodacite, to trimodal Fe-rich tholeiites and polymodal low-K calc-alkaline volcanics and volcaniclastics, as olivine tholeiite followed by quartz tholeiite parent magmas generated by subduction punctured the oceanic crust in stages progressively shifting to the southeast. The petrologic variations and southeastward alkali enrichment of the volcanic suites clearly define the arc's polarity, hence the southeast dip of the subduction zone. A youngest calc-alkaline volcanic province may also have been formed by the subduction event in the southeasternmost part of the arc.

A change in plate motions between 1750 and 1740 Ma caused a flip in the subduction zone, which thereafter (for the ensuing 100 Ma of Proterozoic history) dipped northwest under the Archean Wyoming craton. This major event had many profound ramifications: (1) Formative volcanism ceased in the Prescott-Jerome arc, and arc magmatism changed from volcanism to hydrous plutonism and batholith formation. (2) The Prescott-Jerome arc had been born in an intraoceanic setting allochthonous to North America, but this major 1740-Ma subduction flip resulted in the Prescott-Jerome arc being swept into approximately its present position. (3) In the process, the ocean basin that once lay between the Wyoming Archean craton and the Prescott-Jerome arc began to be consumed, including plate tectonic elements previously and concurrently formed in that ocean basin. (4) As the Prescott-Jerome arc swept toward North America, all ocean islands and protoarcs that existed in that basin were coalesced into a single broad collage of oceanic elements, as follows.

The Bagdad volcanic belt was originally conceived at about 1760 Ma at the mid-ocean spreading center that was the source for oceanic crust forming the original Proterozoic ocean basin between the Prescott-Jerome arc and North America. As this ocean basin started to close from 1750 to 1740 Ma, the Bagdad belt completed its development as an oceanic island or arc only for the time taken to collapse and subduct the thin oceanic crust that lay between the Bagdad and Prescott-Jerome arcs. At the same time, the other half of the ocean basin between the Bagdad belt and the Archean craton-margin clastic wedge began to collapse, and subduction of this thin oceanic crust underneath the Wyoming Archean craton gave rise to the short-lived Antler-Valentine volcanic belt as an incipient island arc at the edge of the craton-margin clastic wedge.

It may have been the 1740-Ma event of initial northwestward convergence and collapsing of the ocean basin that first deformed the arkosic clastic wedge in far northwestern Arizona and opened it to mafic dike intrusion. As both halves of the oceanic crust were finally consumed and the ocean basin was closed, only the thick tectonic elements in that basin resisted subduction, namely the Antler-Valentine belt, the Bagdad belt, the craton-margin clastic wedge, the subduction complex and forearc basin deposit between the Bagdad and Prescott-Jerome arcs that was relict from the previous southeast-dipping subduction event, and also the Prescott-Jerome arc itself. The final closing of the ocean basin resulted in collision of all oceanic elements together, possibly the initial deformation of the clastic wedge against the Archean craton or another stage of deformation in the wedge, and suturing of all supracrustal remnants into a broad collage that now comprises the Northwest Gneiss Belt of Arizona.

This interrelated series of events after 1750 Ma profoundly changed the tectonic history of North America's entire southeastern Proterozoic convergent continental margin, as all loosely assembled oceanic elements across the full length of the Proterozoic orogen in the United States were coalesced into a new Proterozoic crust generated wholly by convergent Proterozoic plate tectonics.

The Prescott-Jerome arc was now established as an offshore island arc separated from the Wyoming Archean craton by a wide shallowly submerged backarc basin of accreted oceanic elements, and from 1740 to 1720 Ma the Prescott-Jerome arc evolved from its earlier submerged state into an emergent *continent-margin arc,* as pretectonic calcic granodiorite plutons intruded the arc and pretectonic calcic granodiorite-tonalite batholiths pervaded the peripheries of the arc, preserving very few remnants of adjacent original oceanic crust.

During this interval of pretectonic plutonic emplacement, the trench axis to the southeast started shifting oceanward in progressive increments as the subduction dip began to flatten. First, a submerged chain of calc-alkaline volcanoes evolved along the new southeast front of the continent-

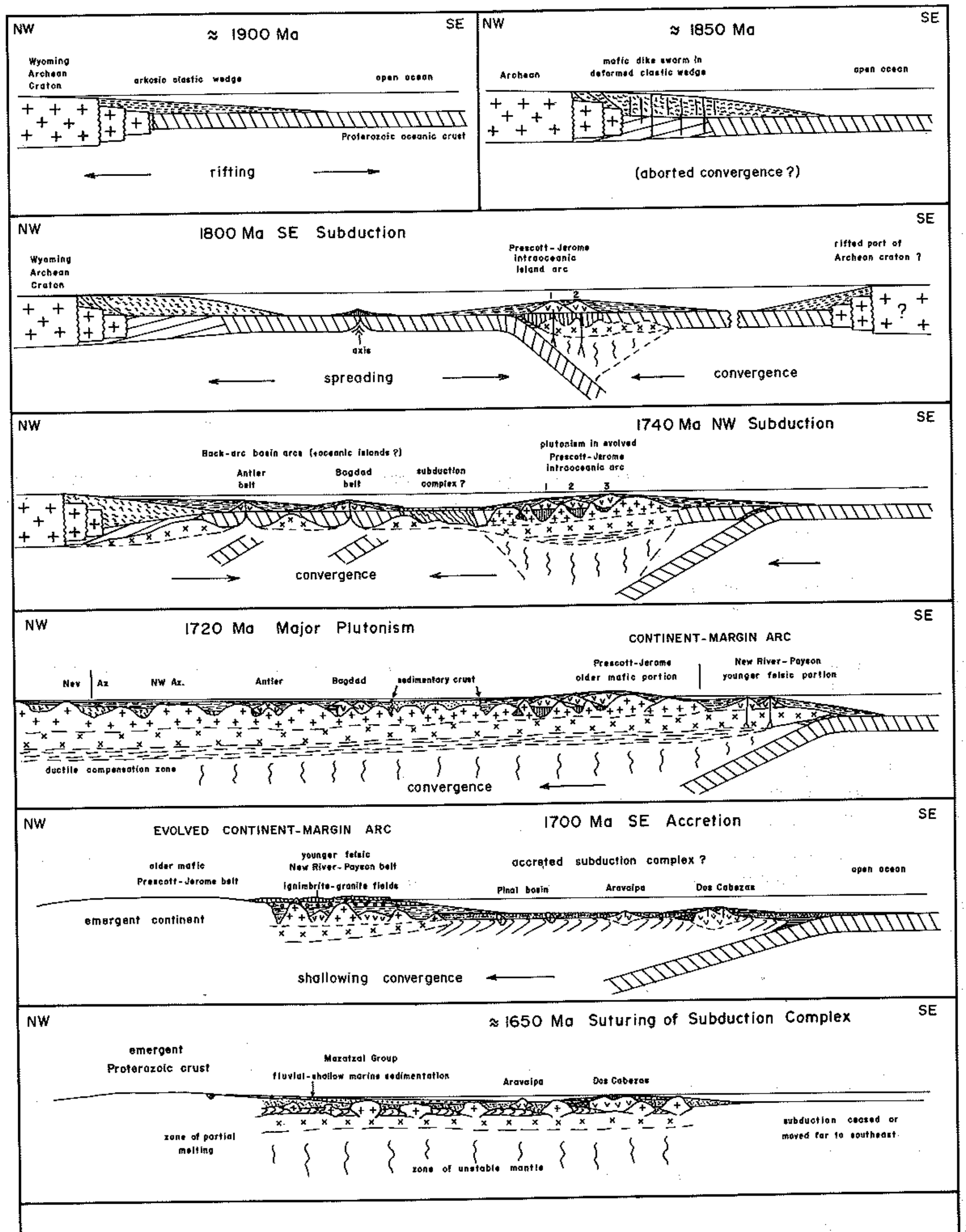

Figure 7. Proterozoic plate tectonic evolution of Arizona as depicted from a diagrammatic cross-sectional viewpoint, at seven key times between 1900 Ma and 1650 Ma. The 1900-Ma diagram depicts initial rifting of the Archean proto-continent and subsequent development of a clastic wedge at the rifted margin. The 1850-Ma diagram depicts the possibility of a brief aborted convergent event to deform the clastic wedge and thicken its subjacent transitional crust. The 1800-Ma diagram shows initial formation of the Prescott-Jerome intraoceanic arc above a southeast-dipping subduction zone at a position allochthonous to North America, and incipient development of the Bagdad belt at a spreading axis. The 1740-Ma diagram shows collapsing of the ocean basin in the northwest due to convergence, preservation of the Antler-Valentine and Bagdad belts, plutonism across the Prescott-Jerome arc, and formation of volcanics at the arc front, all produced by northwest-dipping subduction and ocean-basin closure. The 1720-Ma diagram depicts major plutonism and orogeny across the Central Volcanic and Northwest Gneiss Belts, and formation of the new arc front to the southeast. The 1700-Ma and 1650-Ma diagrams are shifted southeasterly relative to previous ones to show relationships in southeast Arizona. The 1700-Ma diagram depicts shallow subduction producing felsic volcanics in the evolved continent-margin arc and in parts of southeast Arizona. The 1650-Ma diagram depicts plutonism and crustal thickening of southeast Arizona and suturing of the entire accretionary complex to the Proterozoic continental margin.

margin arc, and turbidite graywackes were shed laterally from the volcanic centers into deep intervolcanic basins. A forearc basin developed southeast of these Union Hills Group volcanics when the trench was located near the Salt River line, and distal volcaniclastic detritus from the arc front was fed oceanward into the forearc basin and trench from at least 1720 Ma onward.

By 1720 Ma, subduction had continued under this new Proterozoic crust of central and northwest Arizona for sufficient time that the most deeply buried parts of the arc, oceanic and sedimentary crusts between the Prescott-Jerome arc and the Archean craton, started to undergo widespread partial melting as the thermal infrastructure rose through the crust from continued subcrustal heating. This pervasive lower crustal partial fusion resulted in emplacement of huge zoned calc-alkaline granodiorite-granite batholiths beneath thin supracrustal carapaces of gneissic metavolcanic and metasedimentary rocks, causing broad migmatite terranes to form throughout the Northwest Gneiss Belt. Ductile deformation and high-T regional metamorphism of all supracrustal strata in the belt generally accompanied this major 1720-Ma plutonic event, which markedly thickened and stabilized the crust of the Northwest Gneiss Belt.

As the Proterozoic crust of the Northwest Gneiss Belt experienced major orogeny, parts of it and the continent-margin arc underwent vertical structural readjustment at upper crustal levels to create narrow structural troughs along edges of pretectonic plutons and batholiths. These grabens became the sites of Texas Gulch-type successor clastic sedimentation coextensive with calc-alkaline effusive felsic volcanism. At the shallowly submerged front of the continent-margin arc, a similar clastic suite (Alder purple shale, quartzite, wacke, and tuff) accumulated in a longitudinal trough and prograded back to the batholith scarp that delimited the shoreline. These sedimentary events signified a major 1730-1710-Ma intervolcanic hiatus in evolution of the continent-margin arc, after mafic volcanism ended but before primary felsic magmas were emplaced into the arc front. Volcanic conglomerate finally overwhelmed the clastics as huge felsic magma chambers were emplaced and began to erupt across the arc front.

During the intervolcanic hiatus, subduction stepped either incrementally or in jumps southeast to create a widening forearc basin and accretion complex across southeast Arizona, which established the Pinal basin as a wide interarc basin between the Dos Cabezas arc and the central magmatic arc. This accretion complex included melange and also possibly small primitive island arcs swept in from intraoceanic settings, as well as other allochthonous crustal pieces. At least the younger felsic volcanism in the Dos Cabezas and Ray-Aravaipa belts, however, was endemic to the interarc Pinal basin. Deep-sea argillaceous and wacke supracrustal material from oceanic environments may have contributed to growth of the subduction complex, especially if the Dos Cabezas arc developed offshore from the Pinal basin and was later tectonically accreted to the basin margin. By no later than 1695 Ma, the trench for northwest-dipping subduction had stepped out of Arizona into Sonora, and silicic rhyolitic volcanism erupted in the Dos Cabezas and Ray-Aravaipa belts and in younger parts of the Central Volcanic Belt.

By 1700 Ma, the northwest-dipping subduction zone under the Pinal basin had flattened substantially, making the front 350 km of the continental margin susceptible to emplacement of subduction-generated felsic magmas. Calc-alkaline rhyolites were erupted in the Dos Cabezas and Ray-Aravaipa belts above shallow parts of the subduction zone, while alkali-calcic ignimbrites were erupted at the front of the central magmatic arc above deeper parts of the subduction zone. Red granite batholiths crystallized beneath the carapaces of rhyolitic ignimbrites, and felsic tephra were shed oceanward into the interarc Pinal basin coextensively with its younger quartz-wacke sedimentation.

This event finally made the entire central Arizona magmatic arc fully emergent by 1690 Ma. Thus, together with an already emergent older mafic part of the arc and the metamorphic plutonic terrane behind it, a new Proterozoic crust had emerged, making the arc of continental (Andean) type. Ignimbrites at the arc front immediately began to be eroded back to sea level, which resulted in deposition of successor Mazatzal Group quartzite and conglomerate as fluvial, estuarine, littoral, and shallow open-marine conditions prograded back across the central magmatic arc. Open-marine conditions persisted in the Pinal interarc basin and around the Dos Cabezas arc to the southeast, where Mazatzal-type quartzites were deposited at local shoaling positions 1690-1680 Ma ago.

The northwesternmost Arizona Proterozoic crust was deformed essentially synchronously with major batholith emplacement at about 1720 Ma. Deformation of the older (Prescott-Jerome) part of the central Arizona magmatic arc began about 1710-1700 Ma ago, during ignimbrite activity in the younger felsic part, and was mostly complete at the peak of regional metamorphism, when anatectites at middle crustal levels rose to crystallize at higher levels. Deformation of the younger southeast part of the arc was less intense, varying from penetrative strain to surficial thrusting depending on crustal level, and occurred much later, after Mazatzal strata had been deposited by 1690-1680 Ma on the arc and parts of the accretion complex to the southeast. Deformation of this younger portion of the central magmatic arc occurred between 1660 and 1650 Ma, during syntectonic pluton emplacement, and was contemporaneous with deformation in the northern part of the Pinal basin. Still later, at 1630-1620 Ma, the southern Pinal basin and Dos Cabezas arc were deformed during their syntectonic pluton emplacement, and the Dos Cabezas arc was finally sutured to the Pinal basin.

Deformation of the Arizona Proterozoic crust can therefore be viewed simplistically as a wave of orogenic disturbance that progressed through the crust from northwest to southeast over the same time span as the evolution of that crust: deformation always followed establishment of formative crustal components and their earliest plutonism, coincided with major plutonism and crustal thickening, and usually just preceded metamorphism, because the processes of deformation and crustal shortening caused major heat flow through the crust. Each newly accreted portion of crust was thus subject to a single major event of deformation, incurred few effects from deformation of adjacent regions, and was not redeformed by the younger events, except for polydeformed paragneiss in far northwest Arizona. In detail, however, deformation of each major crustal region was separated in time from that of adjacent regions by as much as 25 Ma, the same time difference between the formative volcanism of adjacent, major crustal segments. This systematic time difference implies a fundamental order to the processes of new crustal generation at Proterozoic convergent plate margins.

Proterozoic plate tectonic processes produced a pattern of continental growth that built up Arizona's Proterozoic crust by accretion of tectonic belts in stages younging successively to the southeast. Lithospheric subduction, which first dipped southeast and later northwest under the Arizona Proterozoic, was ultimately responsible for all events of volcanism, plutonism, deformation, and metamorphism. These tectonic events of new Proterozoic crustal formation and accretion at a Proterozoic convergent continental margin, with all their compositional, petrologic, chemical, and structural intricacies, define the Proterozoic plate tectonic style.

THE PROTEROZOIC PLATE TECTONIC STYLE

This paper has shown that the essential components of new Proterozoic crust were formed by processes actualistically akin to modern plate tectonics. This means that Proterozoic features and processes were similar, not identical, to their modern counterparts, and attempts to precisely equate such Proterozoic features as oceanic crustal composition and depths of magma generation to those of other eras represent the point where actualistic analogies are carried too far. To summarize how plate tectonics in the Proterozoic era were importantly different from either the Archean or modern eras, this paper concludes with a brief overview which contrasts the major characteristics of Proterozoic plate tectonics to those of modern plate tectonics and Archean tectonics. Detailed analysis of these topics is the subject of another work (P. Anderson, n. d.).

Proterozoic Island Arcs

A large body of chemical and petrologic data demonstrate that Proterozoic volcanic belts in Arizona were island arcs, and subtle differences in the data have permitted distinction of tectonic settings for specific arcs, as well as the evolution of each arc, in ways analogous to modern arc settings. However, Proterozoic arcs differed from both Archean and modern ones in several key aspects such as size, composition, differentiation trends, plutonism, structure, and detailed petrologic-chemical attributes of specific tectonic settings.

Proterozoic Oceanic Crust

Neither modern ophiolite complexes nor Archean ultramafic sequences formed during the Proterozoic, and this is one of the most significant features that distinguishes the Proterozoic from other eras. Peridotitic komatiite flows and layered ultramafics characteristic of Archean mantle material are absent from Proterozoic arcs and oceanic crust. The upper part of Proterozoic oceanic crust and basal portions of early primitive intraoceanic Proterozoic arcs are somewhat similar to upper parts of modern ophiolites, but Mg-rich peridotites, ultramafic tectonites, and related rocks found in modern lower oceanic crust are absent in the Proterozoic. Thus, Proterozoic oceanic crust in detail was very different from both Archean and modern oceanic crust, and it has generally not been widely recognized as oceanic crust because of these differences.

Subduction Complexes

Proterozoic subduction complexes seem to be devoid of blueschists, even though many studies have searched for them. This implies that the P-T regime, hence the mechanisms, of Proterozoic subduction were fundamentally different in detail from those of modern subduction. The original tectonic imbrication in a melange complex is obscured by later penetrative deformation that has affected most Proterozoic crustal regions, and so such assemblages have remained largely unrecognized. The closest known Proterozoic analog to a modern melange may exist in penetratively deformed parts of the Southeast Schist Belt of Arizona.

Deformation

The style of deformation affecting Proterozoic crust and island arcs is of intense penetrative foliation and steep lineation, with strong vertical extension. Structures in many Archean greenstone belts are similar, but they are more variable in plunge and strain state, and are sufficiently different to draw detailed contrasts. Deformation in deformed modern arcs is generally less intense and typically does not show the same intense vertical extension found in Proterozoic arcs; in fact, Phanerozoic deformational regimes tend to be characteristically ones of horizontal tectonic transport, which contrasts greatly to the dominantly vertical regime in Proterozoic and Archean systems.

Compositions

Certain rock compositions in specific tectonic settings are unique to Archean or modern tectonic systems and simply do not exist in Proterozoic systems. For example, Archean graywackes are very distinct from Proterozoic ones, and modern successor-basin molasse deposits are not found in regions of new Proterozoic crustal formation such as

Arizona. In fact, even though the same basic plate tectonic elements (such as island arcs and oceanic crust) are present in Phanerozoic, Proterozoic, and Archean systems, the composition of any one specific element in any one particular setting is different in detail than its counterparts in the other eras.

Such contrasts imply that tectonic processes broadly similar to modern ones operated during the Proterozoic, but produced tectonic features that are importantly different to modern ones in many essential details. Thus, a style of plate tectonics unique to Proterozoic convergent margins operated during the Proterozoic era. This Proterozoic plate tectonic style closely matched neither the Phanerozoic nor the Archean styles but was in many respects transitional in its tectonic features. The Proterozoic geology of Arizona is a key reference section for crustal formation and accretion at a continental margin during the Proterozoic era. The features of Proterozoic subduction, island-arc formation, and types of crustal compositions in each tectonic setting are sufficiently dissimilar to both Archean and modern tectonic styles as to warrant clear distinction from them, and to merit recognition that a unique style of **Proterozoic plate tectonics** operated during the Proterozoic era.

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Note: References for this paper are included with those listed on page 144.

in Jenney, J. P., and Reynolds, S. J., 1989, Geologic evolution of Arizona: Tucson, Arizona Geological Society Digest 17, p. 57-147.

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

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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.