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STRUCTURAL EVOLUTION AND TIMING OF DEFORMATION ALONG THE PROTEROZOIC SPRING CREEK SHEAR ZONE OF THE NORTHERN TUSAS MOUNTAINS, NEW MEXICO

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Abstract—The Spring Creek shear zone is a newly mapped structural feature that closely parallels a lithologic discontinuity in the north-central Tusas Mountains. Both strike east-west and dip steeply to the south. The lithologic discontinuity juxtaposes island arc mafic metavolcanic rocks and metasedimentary rocks of the Moppin Complex to the north to northeast with metarhyolites and quartzites of the Vadito and Hondo Groups to the south to southwest. The shear zone juxtaposes amphibolite facies rocks to the south and greenschist facies rocks to the north. The Moppin Complex is typical of Yavapai Province rocks; the southern domain is typical of the Yavapai-Mazatzal transitional domain in New Mexico. Preliminary U-Pb dates on two granites near the boundary help constrain the timing of deformation. The Tres Piedras metagranite (orthogneiss) is 1700-1693 Ma and is interpreted to truncate the lithologic discontinuity on the eastern end. It contains high-temperature solid-state to possibly magmatic flow foliation that parallels a low-temperature fabric in the host rock. Closed to isoclinal similar folds that contain little to no axial planar cleavage, as well as localized reorientation of F2 folds suggest that strain partitioned during cooling of the pluton. The 1693 Ma Tusas Mountain orthogneiss appears to be part of the Tres Piedras orthogneiss in time and composition. It intrudes greenschist-grade Moppin Series north of the Spring Creek shear zone. It lacks folds, but contains possible syn-tectonic magmatic flow structures with little evidence for solid-state deformation. D_1 and D_2 are interrupted to have occurred in the 1700-1690 Ma window during the Yavapai to Mazatzal orogenic transition and were probably a progressive event. These early fabrics were variably reactivated during later Mesoproterozoic 1450-1350 Ma tectonism, with increasing degrees of reactivation south of the Spring Creek shear zone. The reverse fault offset of D_2 structures and amphibolite to greenschist facies juxtaposition that define the Spring Creek shear zone post dates D_2 and is probably a Mesoproterozoic structure.

INTRODUCTION

Proterozoic basement rocks of the southwestern U.S. consist mostly of juvenile crustal materials that were derived from mantle reservoirs shortly before they were added to Laurentia between 1800 Ma and 1600 Ma (Fig. 1; Grambling et al., 1988; Karlstrom and Bowring, 1988; Williams, 1991; Williams et al., 1999; Whitmeyer and Karlstrom, 2008). Following this model, tectonic investigations have assumed that basement rocks were deformed by the succession of deformational events that took place during the assembly of continental crust, and hence these rocks provide an opportunity to investigate the processes and rates of formation and stabilization of continental lithosphere (Karlstrom and Bowring, 1988; Karlstrom and Williams, 1998). Metamorphic P-T paths reflect syntectonic clockwise looping P-T paths followed by periods of isobaric cooling in the middle crust (Williams and Karlstrom, 1997). Metamorphism occurred at least twice, once in the Paleoproterozoic ca. 1700-1650 Ma, and in the Mesoproterozoic ca. 1450-1350 Ma (Karlstrom and Williams, 1998: Williams et al., 1999).

Recent work has focused on finding better constraints for the timing of tectonism and characteristics that distinguish the intensity and styles of metamorphism and deformation in both the Paleoproterozoic and Mesoproterozoic events. In general, the proportion of deformation and metamorphic assemblages attributable to Mesoproterozoic (1450-1350 Ma) tectonism has increased during recent years. Some of the ca. 1400 Ma granites are now known to record significant strain in their thermal aureoles (Kirby et al. 1995; Nyman et al., 1994; Selverstone et al., 2000; Siddoway et al., 2000; Amato et al., in press). Metamorphic grade at 1400 Ma exceeded 500° C in northern New Mexico based on resetting of $^{40}Ar/^{39}Ar$ hornblende ages (Karlstrom et al., 1997; Shaw et al., 2005). Most or all of the New Mexico "triple point" metamorphic rocks are now interpreted to have reached peak conditions of 500˚C and 3.5 kbars at about 1400 Ma (Bishop et al., 1996; Williams et al., 1999). For some workers, debate has shifted to whether all of the ductile deformation and metamorphism in northern New Mexico may be 1450-1350 Ma, given the lack of strong evidence for an earlier Paleoproterozoic tectonism (Daniel and Pyle, 2006). It is therefore important to continue to find timing constraints for fold and fabric forming events, and porphyroblast growth, in the framework of both accretion events and later intracontinental deformations. Evidence that deformational structures and fabrics associated with earlier events can be reactivated by later events creates ambiguous relative timing relationships between Paleo- and Mesoproterozoic structures (Williams et al., 1999). Furthermore, the rocks may have resided at middle crustal depths for the long period of time between Paleoproterozoic and Mesoproterozoic events and thereafter.

The northernmost Tusas Mountains have been interpreted (Williams et al., 1999) to record a significantly less intense overprint of Mesoproterozoic metamorphism and deformation, and thus represent a region that can provide insight into the timing of older events. This paper presents preliminary U-Pb zircon ages for granites from the northern Tusas Mountains. Field relationships in and around these plutons allow constraints to be placed on the timing of development of folds and fabrics and on tectonism in general. These plutons are found along and just north of an important shear zone that juxtaposes a mafic metavolcanic assemblage to the north against felsic supracrustal metasedimentary rocks to

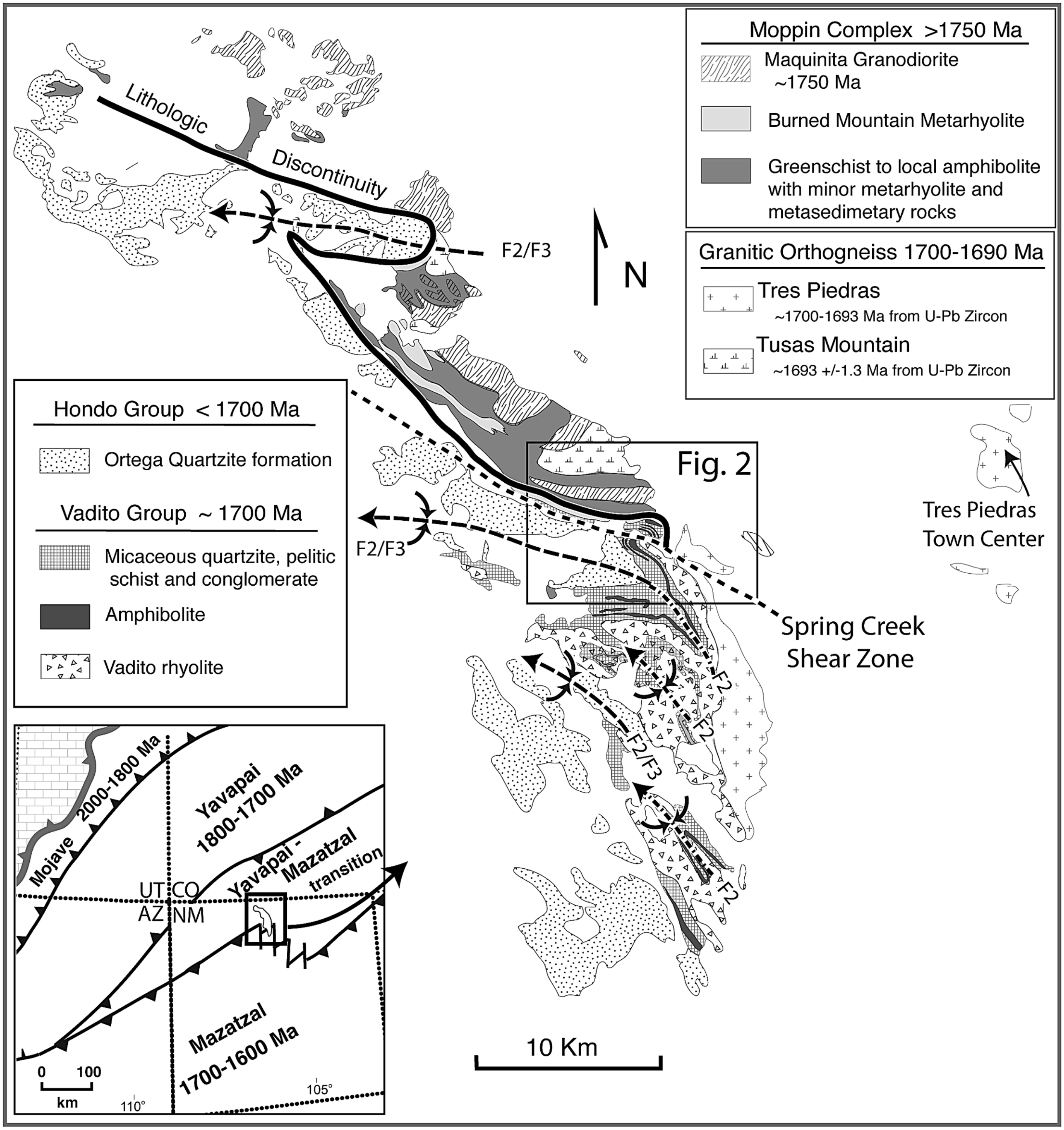

FIGURE 1. Generalized geologic map of the Tusas Mountain Range compiled from Mike Williams, 1991, Bishop et al., 1996 and the University of New Mexico field camp under the direction of Karl Karlstrom. The inset shows the location of Yavapai and Mazatzal crustal provinces and their transition modified from Karlstrom et al., (2004). Red lines mark the axial trace of regional scale F2 folds, blue lines mark the traces of F2/F3 composite folds. Black line is the lithologic discontinuity.

the south (Fig. 1; Barker, 1958; Williams, 1991). Detailed field and laboratory structural observations and geochronological data are used to produce a model that explains foliation development in this region during both 1690 Ma and 1450-1400 Ma tectonism. The earlier, possibly Paleoprotereozoic, events recorded in the Tusas Mountains have been previously correlated regionally to other Proterozoic cored uplifts (Williams et al., 1999). If so, these findings provide much needed regional context.

GEOLOGIC SETTING

The bulk of continental crust of the southwestern United States was accreted to the margin of Laurentia, possibly during the Paleo- to Mesoproterozoic (Grambling et al. 1988; Karlstrom and Bowring, 1988; Daniel and Pyle, 2006), or earlier such that this portion of the crust represents an older crust extended by rifting (Fig. 1; Bickford and Hill, 2007). These rocks now define a NE-SW trending, 1000-km-wide orogenic belt that can be divided into two domains or provinces (Fig. 1). The Yavapai province, the northern portion of this belt, typically consists of mafic to felsic metavolcanic arc rocks produced from 1800-1700 Ma that were accreted during the Yavapai orogeny ca. 1720-1700 Ma (Karlstrom and Bowring, 1988). The southern part of the belt, the Mazatzal province, consists of metavolcanic, igneous, and metasedimentary rocks that have ages of 1700-1600 Ma. These materials are thought to have been primarily deformed during the Mazatzal orogeny ca. 1650-1630 Ma (Karlstrom and Bowring, 1988; Williams et al., 1999; Amato et al., 2008; Jones et al., 2009). Mesoproterozoic granites of 1450-1350 Ma intruded both provinces and in some cases were related to middle crustal intracontinental deformation and metamorphism (Nyman et al., 1994; Karlstrom and Humphreys, 1998).

Proterozoic rocks exposed in the Tusas Mountains lie in the transition zone between the Yavapai and Mazatzal crustal provinces (Shaw and Karlstrom, 1999). The rocks include metavolcanic and metasedimentary rocks that have been intruded by mafic to felsic metaplutonic rocks (Fig 2; Williams, 1991; Bauer and Williams, 1989). The oldest rocks in the region, the Moppin Complex, a mafic sequence of amphibolite, greenschist, and immature metasedimentary rocks (Fig. 2; Williams, 1991; Bauer and Williams, 1989) are exposed in the northern part of the range. Metavolcanic rocks of this complex are older than the intrusive 1750 Ma (age recalculated using Steiger and Jager, 1978) Maquinita Granodiorite, a medium-to coarse-grained strongly foliated and lineated unit (Barker, 1958; Barker and Friedman, 1974). The Moppin Complex is overlain by the Vadito Group, a heterogeneous package of fine- to medium-grained felsic metavolcanic

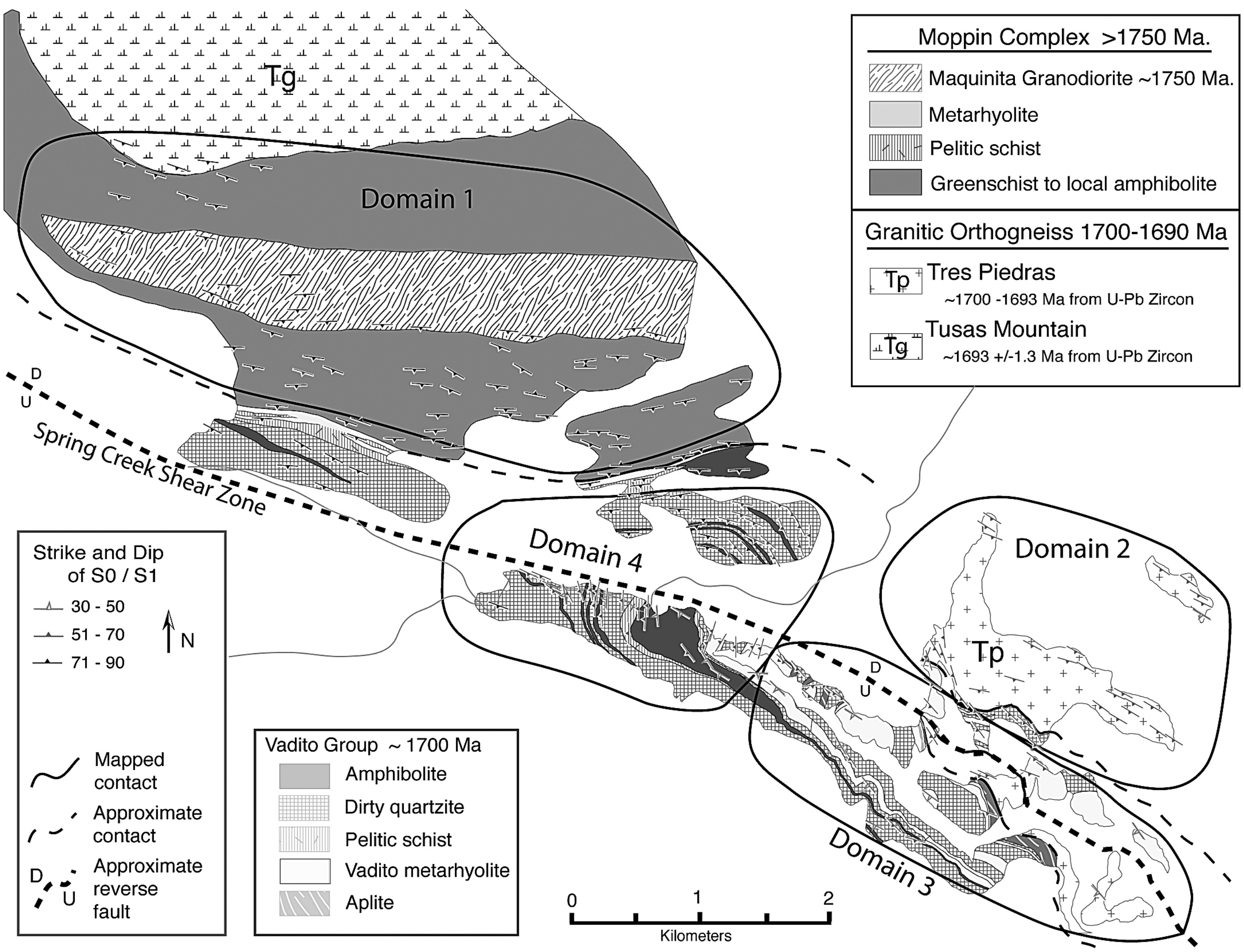

FIGURE 2. Proterozoic exposures across the Spring Creek shear zone and lithologic discontinuity. The four domains of structural data in Figure 4 are shown.

rocks with distinctive quartz phenocrysts, local mafic metavolcanic rocks, and metasedimentary rocks, ranging from pelitic schists to conglomerates. Felsic metavolcanic rocks of the Vadito Group are 1710-1700 Ma (Bauer and Williams, 1989). The stratigraphically overlying Hondo Group is dominated by the basal Ortega Quartzite, which is locally conformable with the Vadito Group (Williams, 1991; Bauer and Williams, 1989). The Ortega Quartzite is vitreous, typically greater than 1 km thick, with minor conglomerate lenses at the bottom of the unit (Williams, 1991). The Tres Piedras and Tusas Mountain orthogneiss intrude the Moppin Complex and Vadito Group. The Tres Piedras pluton, located along the eastern margin of the Tusas range, is a fine to coarse-grained strongly foliated granitic pluton that yielded a U/Pb zircon age (Maxon, 1976) of approximately 1650 Ma (Fig. 2). The contact between the Tres Piedras orthogneiss and the Vadito Group in the eastern Tusas Mountains is gradational and heavily sheared. The weakly foliated medium to coarse-grained Tusas Mountain orthogneiss crops out as a discrete stock with sharp contacts with the Moppin Complex and Maquinita Granodiorite (Fig. 2; Williams, 1991; Wobus and Hedge, 1982). The age of the Tusas Mountain orthogneiss was interpreted to be 1450 Ma (Wobus and Hedge, 1982) based on a single highly discordant and metamict zircon grain.

The map pattern distribution of older arc related rocks of the Moppin Complex to the north to northeast with supercrustal rocks of the Vadito and Hondo groups to the south to southwest creates a lithologic discontinuity located in the northern half of the Tusas range (Fig. 2). This discontinuity originates to the northwest underneath the large quartzite syncline at Jawbone Mountain, extending down to the southeast were it is truncated by the Tres Piedras pluton.

Proterozoic rocks exposed in the Tusas Mountains generally display at least three generations of folds and fabric, interpreted to have developed during regional deformational events, are referred to as D_1, D_2, and D_3. Folds associated with the first deformational event are rare, however there is a strong planar fabric (S_1) parallel to bedding (Williams, 1991). The second event produced the main tectonic grain of the Tusas Mountains with folds (F_2) ranging from millimeters to kilometers in scale (Fig. 1), and NW striking, SW dipping penetrative axial plane foliation (S_2). Folds are reclined, tight to isoclinal, and plunge moderately to the southwest. Stretching lineations (L_2) associated with these structures also plunge to the southwest, but are generally more southerly than the fold axes (Williams, 1991). Folds (F_3) associated with the third deformation are open to isoclinal upright folds with axes that plunge very shallowly to the west (Williams, 1991). The third event produced a weak to moderately strong foliation (S_3) that generally strikes to the east and is sub-vertical. S_3 is mainly developed in mica-rich schistose rocks (Fig. 1; Williams, 1991).

The metamorphic assemblages in rocks of pelitic bulk composition of the Tusas Mountains contain biotite + muscovite + chlorite + magnetite + ilmenite + quartz, but vary in the appearance of garnet, staurolite, kyanite, andalusite, and the fibrolite form of sillimanite (Williams et al., 1999; Davis, 2003). These porphyroblasts typically contain straight or curved inclusion trails of quartz and oxides that define earlier deformational fabrics. These assemblages are generally well annealed with a third deformational fabric that wraps some porphyroblasts (Barker, 1958; Williams et al., 1999; Davis, 2003). Rocks south of the lithologic discontinuity preserve the petrologic character that is typical of the P-T evolution preserved in the central and southern Tusas Mountains (Bishop et al., 1996; Williams et al., 1999; Davis, 2003). North of the discontinuity the mineral assemblage contains garnet + staurolite (or chloritoid), as well as well-annealed matrix quartz + muscovite + biotite, but lacks a later deformational fabric (Williams, 1991; Davis, 2003).

Absolute timing of D_1 was constrained to be younger than the Tres Piedras pluton, dated at 1650 Ma (Maxon, 1976; Williams et al., 1999). The age of D_2 has been controversial. The last major event (D_3) in the Tusas Mountains is interpreted to be ca. 1400 Ma based on microprobe dating of monazite grains within porphyroblasts in the southern part of the range (Bishop et al., 1996). Recent work in the Picuris Mountains suggests that D_2 and possibly D_1 events, at least in the Picuris Mountains, may have occurred at 1450-1400 Ma (Daniel and Pyle, 2006). If these fabrics can be correlated regionally, virtually all ductile deformation in northern New Mexico must be Mesoproterozoic in age (Daniel and Pyle, 2006).

TECTONOSTRATIGRAPHY

Moppin Complex

The Moppin Complex is dominated by fine-to medium-grained mafic metavolcanic rocks, with lesser felsic metasedimentary rocks, and minor feldspathic metasedimentary rocks. Lithologic and compositional layering in all units strike roughly east-west, and dip steeply. Felsic metasedimentary rocks are fine- to medium-grained clastic rocks that show primary structures such as graded beds and cross beds; the clastic rocks are typically discontinuous at the map scale, and are interlayered with mafic metavolcanic rocks. Felsic metavolcanic units are white to pink mica-rich feldspathic schists with relict phenocrysts that are found as 0-10 meter thick layers.

Vadito Group

The Vadito Group of the north-central Tusas Mountains is dominated by felsic metasedimentary, and metavolcanic rocks that are mainly exposed in the southern half of the study area (Bauer and Williams, 1989). The upper contact of the Vadito Group is concordant with the overlying Ortega Quartzite of the Hondo Group. One of the key transitional stratigraphic sections is located on the slopes of Kiowa Mountain (Bauer and Williams, 1989), which is just south of the area mapped in this study (Fig. 2). The Vadito Group can be broken into four map units: micaceous quartzite (55%), metarhyolite (30%), amphibolite (10%), and intercalated pelitic rocks (5%). The Vadito Group in other Proterozoic-cored uplifts in northern New Mexico includes abundant metarhyolite (Williams and Burr, 1994); however the micaceous quartzite is the most dominant member in this study area.

The Vadito Group metarhyolite is light pink to bright red to dark purple, fine grained, massive, and has the general mineral assemblage of potassium and plagioclase feldspar, quartz, muscovite, minor amounts of biotite and accessory minerals. Feldspar makes up most of the fine-grained matrix. Quartz in the matrix occurs as distributed fine grains and as ribbons, or defines a mineral lineation. Phenocrysts of 1-3 mm quartz and/or feldspar are generally distributed homogeneously throughout the rock. Phenocrysts locally have sigmoidal recrystallized tails that are important tectonic transport indicators (Williams and Burr, 1994). This unit correlates with the Burned Mountain metarhyolite that shares all characteristics with the metarhyolite member of the Vadito Group, but is a thin <100 meter thick unit that lies conformably on the Moppin Complex.

Bedding-parallel layers of amphibolite lie above the uppermost Vadito Group rhyolite layer in the southeastern portion of the field area and probably represent mafic lava flows or sills. The thickness of the largest amphibolite unit varies greatly because of limb thinning from 300 m in the hinge of a large map-scale fold to a few meters thick along the southern limb of the same fold.

Metapelitic layers in the Vadito Group are discontinuous, are present above and below the main amphibolite, and are in contact with the Moppin Complex in the central and western portion of the field area. These pelitic units contain garnet + staurolite + biotite + muscovite + plagioclase + quartz ± kyanite.

Tres Piedras orthogneiss

The Tres Piedras orthogneiss is a deformed igneous body found on the eastern side of the field-area (Fig. 2). Excellent exposures occur in the region of the confluence of the Rio Tusas and Tusas River canyons (Fig. 3a; Barker, 1958) and in the type locality near the town of Tres Piedras (Maxon, 1976; Wobus and Hedge, 1982). The Tres Piedras orthogneiss is a pink to light orange-brown, fine to medium grained granite high in silica (~ 76%) and low in calcium with a K_2O: Na_2O ratio of 5:3 (Wobus, 1984; Barker and Freidman, 1974), containing 30-40% quartz, 20-25% microcline, 15-20% albite, 10-15% biotite, 5-10% muscovite, and accessory amounts of epidote, zircon and opaque minerals (Fig. 3b) and is relatively homogeneous in composition (Maxon, 1976; Wobus and Hedge, 1982). The orthogneiss is penetratively foliated in the west and weakly to unfoliated in the east. Compositional bands occur along 10-meter wide zones parallel to the main tectonic fabric.

The only sharp contact observed between the Tres Piedras orthogneiss and the wall rocks is exposed in the southeastern portion of the study area at the confluence of the two rivers. The contact of the Tres Piedras orthogneiss with Vadito Group metarhyolite is structurally duplicated and appears to be heavily tectonized (Fig. 2). The northern contact is approximately 1 meter wide, and is composed of a white muscovite/ sericite- rich shear zone. The southern contact appears to be a mylonite to cataclascite and is locally intruded by a pegmatite, which is 95% potassium feldspar. This contact extends further, south of the study area, along the Tusas Box canyon, but has been notoriously difficult to map and appears gradational over a 50-meter zone (Barker, 1958; Williams, 1991).

Tusas Mountain orthogneiss

The Tusas Mountain orthogneiss is an oval shaped plutonic body that is roughly 3 km in diameter, but the orthogneiss is covered by Quaternary alluvium on the northeastern margin and the extent of the pluton may be larger in the subsurface (Figs. 1, 2). The pluton is in contact with the Moppin Complex along its southern margin and the Maquinita Granodiorite on its northwestern margin (Fig. 2). This orthogneiss is pink to pale orange, porphyritic with 1-5 mm phenocrysts of quartz and microcline that account for 40 to 70% of the rock volume (Fig. 3c). The matrix consists of fine-grained quartz, microcline, muscovite, with accessory amounts of biotite, chlorite, fluorite, zircon, and epidote. Large phenocrysts of equant and undeformed microcline and quartz vary in total modal percent and make the body appear to be heterogeneous, but with relatively little mineralogical variation (Figs. 3d, e). The body is weakly foliated in the center, but strongly foliated within a few hundred meters of its contact with the Moppin Complex (Fig. 3c). Tabular bodies of amphibolite and pelite measuring 1-5 meters by 100 meters occur near the southern contact of the Tusas Mountain orthogneiss with the Moppin Complex. These bodies are coarser grained, and have less chlorite alteration than the proximal Moppin Complex wall rock. The amphibolite and pelite blocks are interpreted to be xenoliths of Moppin amphibolite and pelite.

Two ages have been published for the Tusas Mountain orthogneiss. Maxon (1976) determined a U/Pb zircon date of 1690 Ma., and Wobus and Hedge (1982) determined a U/Pb date of 1430 Ma from a highly discordant zircon. Geochronologic data from this study support the Maxon (1976) date (see below).

STRUCTURAL ANALYSIS

The Tusas Mountains of northern New Mexico preserve evidence for three fold and fabric-forming generations that vary in intensity across the area. These deformational fabrics fundamentally change in character from north to south across the lithologic transition. Structural data have been divided into four structural domains based on the orientation of bedding (S_0) and sub-parallel first tectonic cleavage (S_1) that define the dominant surfaces for mapping (Fig. 2).

Rocks in the northern portion of the study area are dominated by a single cleavage with little evidence of folding. Rocks in the central and southern portion of the field area contain a single strong cleavage that has been overprinted by two subsequent fold-and fabric-forming events. The southernmost part of the study area contains a strong early cleavage, as well as numerous second and third generation folds (F_2, F_3). D_3 folds and fabrics are preserved to the south where they clearly overprint F_2 folds in all lithologies, and form interference patterns on the outcrop scale.

Domain 1

The largest and simplest of the four domains covers the northwestern half of the field area. Domain 1 includes the Moppin Complex and the adjacent Vadito Group (Fig. 2). S_1 in this

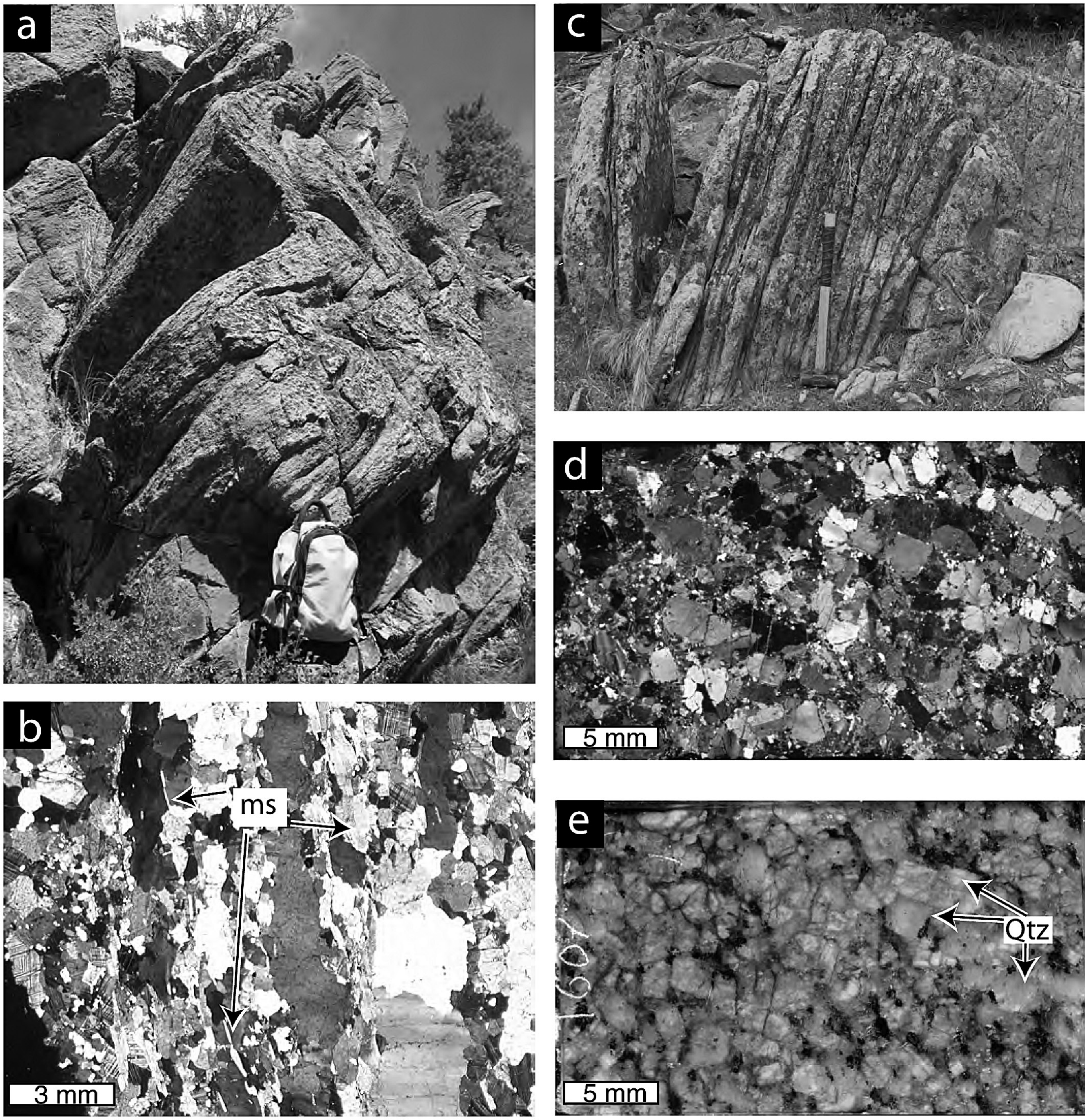

FIGURE 3. (a) Tres Piedras orthogneiss folds with a backpack for scale (b) Thin-section of a high-temperature fabric of Tres Piedras (c) A single fabric in the Tusas Mountain orthogneiss, sledgehammer is one half a meter long (d) Thin-section of magmatic fabric in the core of the pluton (e) Photomicrograph of the billet from (d) Qtz = quartz.

domain is a penetrative foliation defined by fine-grained muscovite and chlorite that are generally aligned parallel to bedding. S_1 is equally intense in both the Moppin Complex and Maquinita Granodiorite. S_0 and S_1 generally strike 090-110˚ and dip 70 to 90˚ to the south. Poles to S_0/S_1 in Figure 4 are clustered, giving a maximum that plunges 81˚ toward 255˚. Second-generation folds (F_2) and fabrics (S_2) are rare. Folds in the Maquinita Granodiorite are not observed. Lineations are steep to nearly vertical in this domain. Intersection lineations observed between S_1 and a weak S_2 axial planar cleavage ($L_{2/1}$), plot with scatter along a great circle that strikes 092° and dips 80°S (Fig. 4). $L_{2/1}$ are subparallel to the few observed axes of F_2 folds that are essentially vertical, plunging 86˚ and trending 097˚ (Fig. 4). Mineral lineations in Domain 1 are roughly parallel to $L_{2/1}$ intersection lineations. Elongated quartz, feldspar, and hornblende grains that define mineral lineations in the Moppin Complex and Maquinita

Granodiorite are roughly parallel to the $L_{2/1}$ intersection lineations. A dozen of these elongated grains scattered throughout the area are strongly asymmetric and suggest that tectonic transport during the second deformation was top-to-the-north.

The fabric within the Tusas Mountain orthogneiss pluton located in Domain 1 has a complex relationship to the first two deformational events. The pluton contains a moderately strong fabric along its southern flanks that decreases in intensity toward the center of the pluton. The fabric is defined by alignment of fine-grained muscovite and biotite in the fine-grained matrix, and is roughly parallel to S_1 seen in the host Moppin Complex. There are no obvious intersection lineations defined by intersecting cleavages nor folds of fabric. A weak alignment of undeformed minerals that comprise a magmatic flow fabric is recognizable near the center of the pluton, although there is very minor localized solid-state deformational fabric. The increase in intensity of a coplanar solid-state S_1 towards the margin is interpreted to indicate that S_1 was developing synchronously with pluton emplacement.

Domain 2

Domain 2 includes the main Tres Piedras orthogneiss exposure in the field area. S_1 is defined by muscovite and biotite throughout the exposure. Spacing of S_1 that ranges from 2-10 mm is controlled by the size of feldspar and quartz grains and is not a crenulation cleavage. Both quartz and feldspar are elongate and aligned in hand-sample, though quartz is found as grain aggregates and as coarse ribbon grains, whereas feldspar is generally more euhedral (Fig. 3b). Aligned muscovite and elongated feldspar and quartz that define the S_1 fabric in thin-section, also occur within larger feldspar grains that are aligned with the main fabric.

F_2 folds of the foliation are similar to isopach, reclined and asymmetric, tight to isoclinal, with moderately plunging axes generally to the southwest. Fold axes fall along a great circle with a best-fit pole that plunges 37˚ trending 267˚ (Fig. 4). Amplitudes range from centimeters to meters in scale. Most folds have one limb longer than the other with an average ratio of roughly 4:1 primarily verging to the northwest. These folds also are found in discrete layers in the Rio Tusas exposures parallel to the main fabric. Observed $L_{2/1}$ intersection lineations are shown in Figure 4 along the great circle that strikes 133˚ and plunges 52˚ to the southwest. Measured axes (b_2) of F_2 folds plot along a great circle that strikes 136˚, and plunges 47˚ to the south (Fig. 4). $L_{2/1}$ lineations plot along the b_2 plane, suggesting that the lineations are coplanar with the fold axes.

S_2 fabric, defined by the alignment of muscovite, is heterogeneous in intensity, occurring in F_2 hinges, while sub-parallel to S_1 in limbs. Poles to S_2 cluster to the northeast, sub-parallel to poles to the S_1 surface. The tight clustering suggests that S_2 is generally undisturbed by later events in this domain (Fig. 4). Mineral lineations (L_2) are defined primarily by the larger elongate quartz grains, and to a lesser extent feldspar grains. L_2 measurements fall along a great circle that strikes 142˚ and dips 40˚ to the south (Fig. 4). The pole to the great circle of S_1, and measured F_2 fold axes also lie on this average plane. Mineral lineations are strongest in folds with axes plunging to the southwest. Kinematic indicators, observed along rock faces parallel to the mineral direction, consist of sigmoidal clasts of quartz and feldspar with dominantly sigma-type asymmetric tails that suggest tectonic transport was top to the northeast. Sigmoidal tails observed on both limbs of F_2 folds indicate transport to the northeast. This suggests that folding occurred before and possibly during top-to-the-northeast tectonic transport.

There are a few folds with axes and mineral lineations that are far from parallel to these general trends. These folds are responsible for the wide distribution of mineral lineation. Mineral lineations in these folds tend to be weak, and appear at a high angle with the orientation of the majority of lineations that plunge moderately to the southwest. The axes of these folds generally plunge more shallowly, and are more open than the majority of folds. The consistency of all fold axial plane orientations and splay of

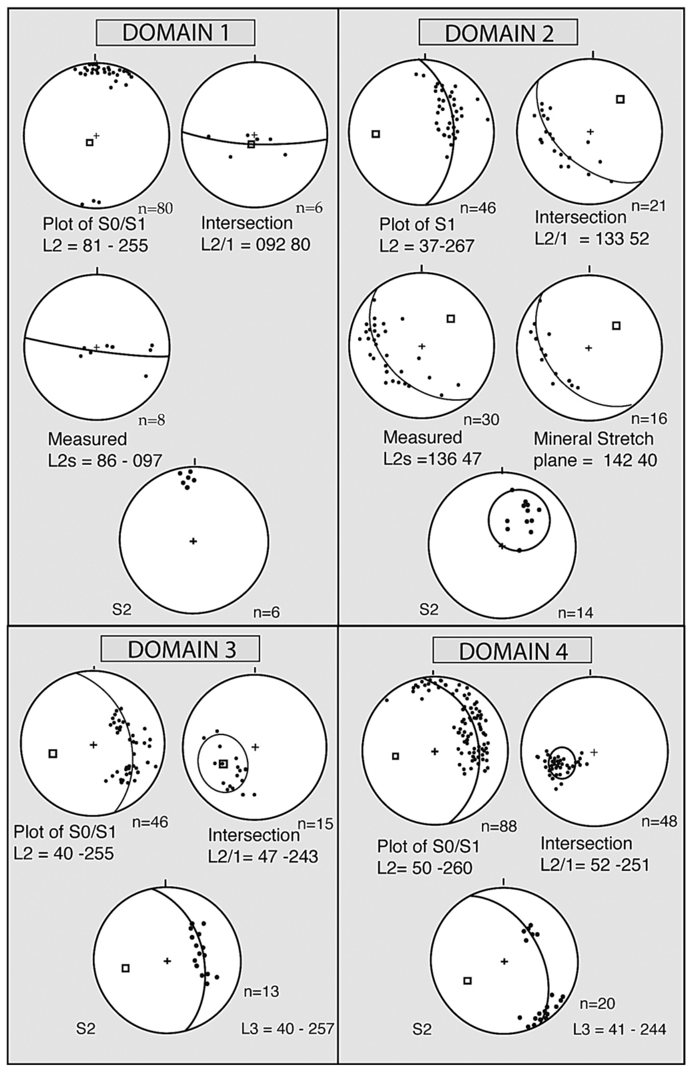

FIGURE 4. Structural data for the four domains consists of S_0 compositional layering and sub-parallel S_1 composite fabric $L_{2/1}$ intersection lineations between S_1 and S_2, measured fold axes β_2, asymmetric mineral stretching lineation, and S_2 fabric orientations.

mineral lineations along a great circle suggests that they are the product of a progressive deformational event. Mineral lineations seen on these atypical folds may have been produced prior to folding. The poor production of mineral lineations and open to closed nature of these folds suggests that they may have developed late in the D_2 generation of folding, or in a lower strain environment relative to the folds that make up the majority of the population. Reorientation by a later non-parallel event is unlikely due to the virtually undeformed nature of the S_2 surfaces in Domain 2.

The third deformation did not form F_3 folds or an S_3 cleavage in the body of the Tres Piedras pluton. The Vadito Group near the southern contact with the Tres Piedras orthogneiss does contain minor F_3 folds, which are grouped with F_3 folds of Domain 3.

Domain 3

Domain 3 includes the southeastern exposures of the Vadito Group in contact with the Tres Piedras pluton. S_1 in this region is bedding parallel, and is defined by the alignment of muscovite and local biotite. F_2 folds range in amplitude from about 1cm to 10 cm; the folds are inclined, recumbent, and tight to isoclinal. These folds are similar in style, but fewer in number to those seen in Domain 2. F_2 axes are slightly separated into two maxima along a great circle that strikes roughly 350˚, and dips moderately to the west. The distribution of the $L_{2/1}$ intersections in Figure 5 also shows a clustering of F_2 fold axes plunging moderately to the southwest.

S_2 is also defined primarily by muscovite, but is weaker in intensity than S_1. Poles to S_2 planes (Fig. 4) are slightly distributed along a great circle similar to the great circle for S_0/S_1 (Fig. 4). Mineral lineations in this domain are defined by elongate quartz and feldspar, and are best developed along the contact with the Tres Piedras orthogneiss in the north. These lineations generally plunge 40 - 50˚ to the southwest, which is sub-parallel to the axes of F_2 folds.

F_3 folds are generally small (1-10 cm), upright, and open to closed, with axes plunging shallowly to the west. Locally, the overprint of F_3 folds produces an interference pattern in mica-rich layers and quartz-rich layers in Domain 3 and 4. The intersection angle between the two axes is between 15 and 30˚. S_3 is the strongest in this domain of any in the field area, although it is the weakest of the cleavages. Poles to S_3 planes cluster and are interpreted to be undisturbed by later deformation.

Tres Piedras orthogneiss and Vadito Group strata are apparently repeated in Domain 3 (Fig. 2). This boundary between repeated strata is found to dip approximately 60 - 70˚ to the south crosscutting the 40 - 50˚ dipping Vadito Group and Tres Piedras orthogneiss rocks. This repetition is possibly related to a large scale isoclinal F_2 fold; however, the increase in the amount of micacous quartzite up section in the Vadito Group sequence favors a thrust or reverse imbrication geometry. In a reverse fault scenario, the southern portion of repeated strata would have come from a lower structural level. Projection of the structure in crossection suggests that approximately 3 kilometers of movement would be required to realign the units along this structure (Davis, 2003). Since this structure offsets the map-scale F_2 fold, it is interpreted to be a post D_2 feature.

Domain 4

Domain 4 describes the western portion of the Vadito Group exposures in the field area centered on a map scale fold defined by the amphibolite member of the Vadito Group. Aside from this large fold, Domain 4 shares many characteristics with Domain 3.

S_1 is bedding parallel and is defined by the alignment of muscovite, biotite, and hornblende. F_2 folds of S_0/S_1 are closed to tight, and have a similar orientation to those seen in Domains 2 and 3. Poles to S_0/S_1 planes are distributed along a great circle similar to that of Domains 2 and 3 (Fig. 4).

S_2 planes in Domain 4 have been folded similarly to planes in Domain 3, the planes separate into two maxima along a great circle that strikes northwest, and dips moderately to the east (Fig. 4). This great circle is parallel to that of Domain 3, suggesting that Domain 4 was similarly overprinted by a third deformational event. The clustering of $L_{2/1}$ intersection lineations to the southwest is tight, plunging 52˚ toward 251˚ (Fig. 4).

Mineral lineations in Domain 4 are defined by elongate quartz and feldspar. Pebbly conglomerate layers in the upper portions of the micaceous quartzite member of the Vadito Group clearly show a kinematic sense of transport of southwest over northeast. These indicators are again parallel to the axes of the F_2 fold population. There are few mineral lineations recognized in this domain, probably because of the generally fine-grained nature of a majority of the Vadito Group rocks.

Similar to Domain 3, F_3 folds are rare and typically 1-10 cm, and the S_3 fabric is weakly developed. F_3 folds are upright, open to closed, with fold axes that plunge at a shallow angle to the west. F_3 folds in this domain are best developed in pelitic units and form interference patterns. S_3 is weakly developed in this portion of the field area and is generally undeformed. All poles to S_3 from the study area are horizontal and cluster to the north-northeast and south-southwest.

Correlative Vadito Group units are offset from north to south across Domain 4 (Fig. 2). This offset is in line with a possible fault plane defined by doubled Tres Piedras orthogneiss and Vadito Group strata seen in Domain 3 and is interpreted to continue directly west. If the amphibolite members of the Vadito Group of Domain 4 are correlative from north to south through the large F_2 fold, they give an apparent dextral separation of approximately 1 km. Three kilometers of vertical offset along the repetition of layers seen in Domain 3 can account for the apparent dextral motion if the plunging nature of the paired anticline to syncline structures is taken into account.

GEOCHRONOLOGY

Previous geochronologic studies have produced a basic framework on which the Proterozoic tectonic history of the Tusas Mountains has been interpreted (Maxon, 1976; Wobus and Hedge, 1982; Bauer and Williams, 1989; Aleinikoff et al., 1993).

A summary of previous dates is provided in Table 1. Five new samples from this study were dated using single crystal ID-TIMS U/Pb zircon analysis at the Massachusetts Institute of Technology under the direction of S.A. Bowring. Locations of these samples can be found on Figure 5. These preliminary results are listed in Table 2.

These five samples include previously dated Tres Piedras orthogneiss and Tusas Mountain orthogneiss, as well as a pegmatitic variant of the Tres Piedras orthogneiss and a fine-grained cross cutting metagranitic dike that cuts the main foliation of the Tres Piedras orthogneiss.

Sample TP74 of the Tres Piedras orthogneiss pluton collected 1.5 km from the Vadito contact provided a date of 1700 ± 9 Ma, which is significantly older than the 1654 Ma date published by Maxon (1976) (Fig. 5). This new date for the Tres Piedras orthogneiss is close to the date published for the Vadito Rhyolite (Bauer and Williams, 1989), which opens the possibility that the two rocks are petrogenetically related.

Sample TP85 of the Tres Piedras orthogneiss yields a date of 1693 ± 11 Ma (Fig. 5). This date is based on a collection of discordant grains, the most concordant of which, zircon 5, is reversely discordant. This sample, based on grain size and texture, is interpreted to be a pegmatite related to the Tres Piedras Pluton composed of approximately 60% potassium feldspar, 35% quartz and 5% muscovite, and cuts the Vadito Group on the map scale.

Sample TG26, taken from near the southern contact of the Tusas Mountain orthogneiss with the Moppin Complex, yields an age of 1693 ±1.3 Ma (Fig. 5). Three zircon grains were analyzed from this sample, and make a very tight discordia line that establishes the reported date. This date of 1693 ± 1.3 Ma is significantly different than the age of 1430 Ma assigned to the Tusas orthogneiss by Wobus and Hedge (1982), but similar to the age of 1660-1690 Ma reported by Maxon (1976).

Sample TP210 was collected outside the immediate study area a few hundred meters north of the town of Las Tablas on NM Highway 519. This rock is a dike in the Tres Piedras orthogneiss, approximately 10 meters long, and 5 –10 centimeters wide. Five zircon grains dated from TP210 were discordant yielding an intercept at 1631 ± 28 Ma. One concordant zircon grain yielded an age of 1633 ± 3 Ma (Fig. 5). This dike cuts a pervasive deformational fabric in the Tres Piedras orthogneiss host rock (Fig. 6). However, the dike itself contains a mineral alignment parallel to the walls of the dike that is not parallel to any fabric in the host rock and appears to be due to magmatic flow. This dike is planar in the field. Figure 6 shows a possible connection point between two sub-coplanar fractures.

In summary, Tres Piedras orthogneiss and pegmatite and Tusas Mountain orthogneiss all give crystallization ages of 1693 Ma. This suggests Tusas Mountain orthogneiss is related to or is part of the larger Tres Piedras pluton. As discussed below, these are

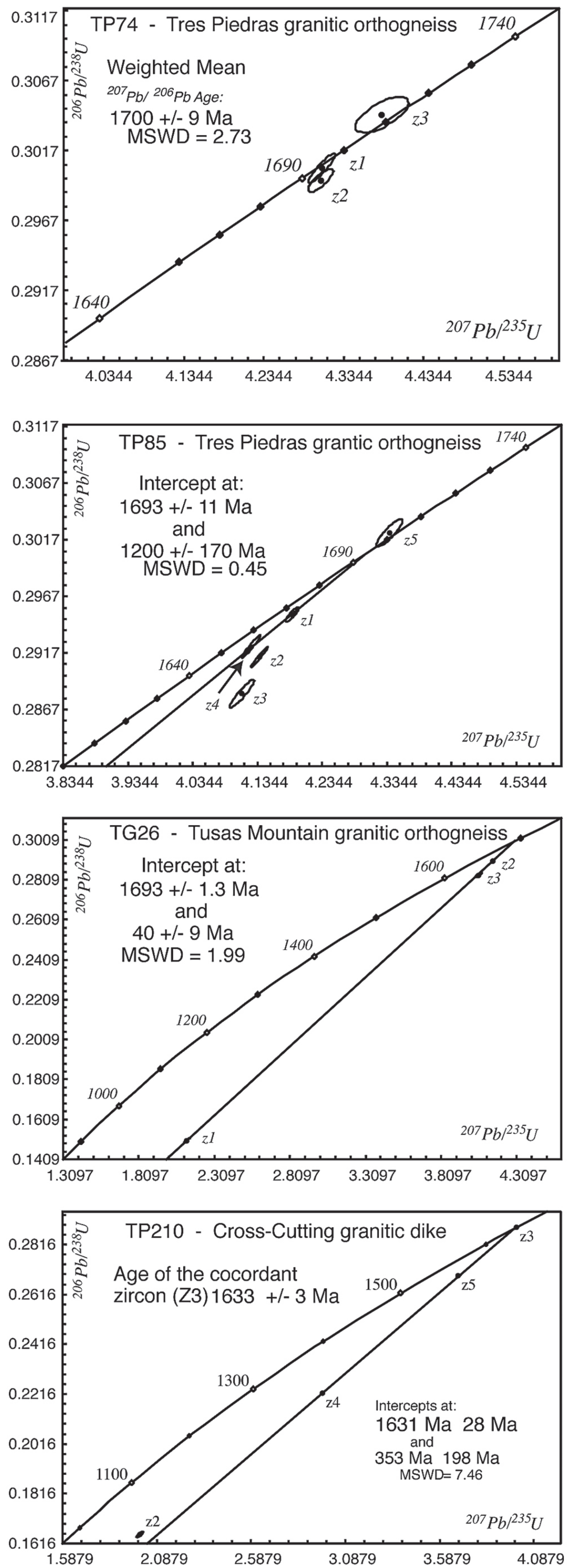

FIGURE 5. Concordia diagrams of U/Pb in Zircon for samples a) TP74, b) TP85, c) TG26 and d) TP210 show an emplacement age of approximately 1700-1693 Ma for both the Tres Piedras and Tusas Mountain granitic orthogneiss.

TABLE 1. Previously published igneous geochronologic data

Rock Unit	Published Age	System used	References
Maquinita Granodiorite	1750 ± 25 Ma	U-Pb in zircon	date by L.T. Silver (Reed et al., 1987)
Tres Piedras orthogneiss	1654 Ma 1490 ± 21 Ma 1500 ± 44 Ma	U-Pb in zircon Rb/Sr whole rock Rb/Sr whole rock	Maxon, 1976 (Type Locality) (Rio Tusas Canyon)
Vadito metarhyolite and Burned Mountain rhyolite	~1700 Ma	U-Pb in zircon	(Bauer and Williams, 1989)
Ortega Quartzite	< 1710 Ma	U-Pb in zircon	Aleinikoff et al., 1993
Tusas Mountain orthogneiss #	1660 Ma 1690 Ma	Rb/Sr whole rock U/Pb in zircon	Maxon, 1976 #
Tusas Mountain orthogneiss	1550 ± 40 Ma 1449 Ma 1421 Ma	Rb/Sr whole rock U-Pb in zircon	Wobus and Hedge, 1982

Maxon, (1976) called this body the Tres Piedras granite. It was renamed the Tusas Mountain granite by Wobus and Hedge (1982).

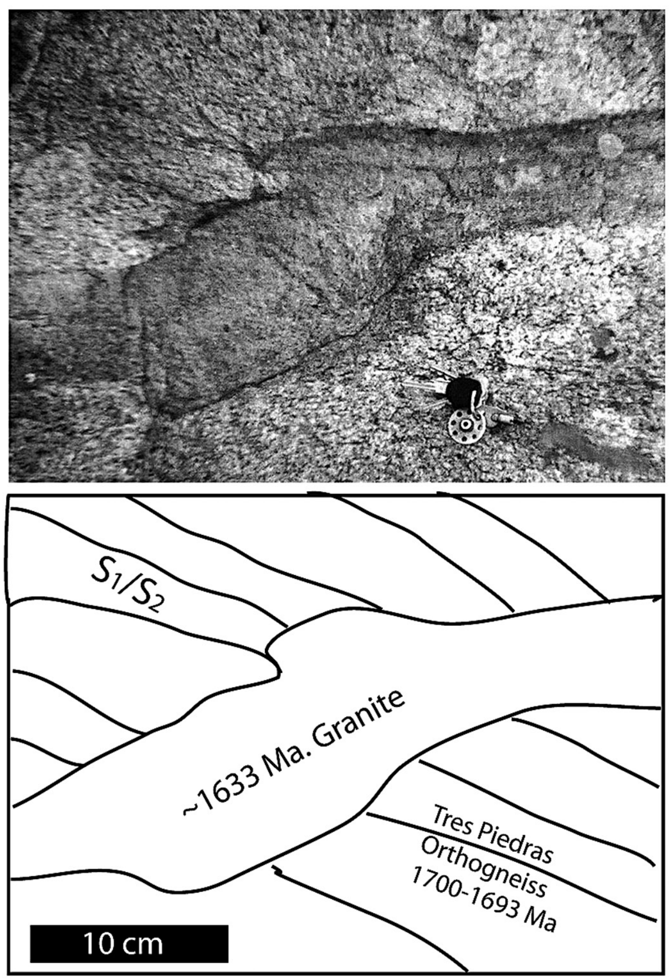

FIGURE 6. A photo and summary drawing of the sample collection site for TP210. The S1 /S2 crosscutting nature can be seen above and below the dike.

interpreted to be syntectonically emplaced with D_2. The 1633 Ma cross-cutting dike is interpreted to be post-D_2.

DISCUSSION

The lithologic discontinuity defined by the regional contact between the Moppin Complex and Vadito Group that parallels Spring Creek contains a regional transition in the orientation of D2 structures and increasing intensity of D3 structures from north to south. The 3 km reverse offset of the large F_2 fold in Domains 3 and 4 that parallels the lithologic discontinuity defines the Spring Creek shear zone (Davis, 2003). This structure also possibly duplicates a portion of the overlying Ortega Quartzite to the west of the study area (Williams, 1991). Variability of the macro- and microstructural expression of strain within the Tres Piedras and Tusas Mountain orthogneiss plutons, as well as new geochronologic data suggests that the plutons provide important age constraints to the Proterozoic tectonism preserved in this range.

D1

The first fabric and folding generation has important cross-cutting relationships with the Tres Piedras and Tusas Mountain orthogneisss. S_1 is interpreted to be correlative across the study area because of its bedding parallel orientation, its similar low metamorphic grade (chlorite-biotite), and its lack of early folds. This fabric is likely a compaction or early horizontal translation of thrust related fabric. The strong S_1 in the Tres Piedras orthogneiss, which is also expressed as S_1 aligned mica inclusions within large S_1 aligned euhedral feldspar phenocrysts, suggests that this body was being deformed at very high sub-solidus conditions to possibly magmatic conditions in portions of the pluton following the criteria of Patterson et al., 1989. This dichotomy between a low-grade fabric in the host, but high-temperature fabric in the

TABLE 2. Samples from this study dated with U/Pb in zircon

Sample #	Rock Unit	Date Obtained
TP74	Tres Piedras orthogneiss	1700 +/- 9 Ma
TP85	Tres Piedras pegmatite	1693 +/- 11 Ma
TG26	Tusas Mountain orthogneiss	1693 +/- 1.3 Ma
TP210	Cross-cutting dike in Tres Piedras	1633 +/- 3 Ma

Tres Piedras pluton suggests that the pluton was syntectonically emplaced with respect to D_1, perhaps at a relatively shallow level.

D2

The second deformation produced folds, planar fabrics, and stretching lineations that differ a bit in strike across the four domains. F_2 folds, rare and oriented vertically in Domain 1, are consistent with regional orientations published by other workers in Domains 3 and 4 (Williams et al., 1999). L_2 mineral lineations, interpreted to be parallel to the local finite stretching direction, can be thought of as stretching lineations and are also consistently oriented sub-parallel to F_2 folds in these domains. Kinematic indicators throughout the study area suggest that tectonic transport was from the southwest to the northeast during D_2. Axes of a portion of the F_2 fold population in the Tres Piedras pluton of Domain 2 are distributed along the regional S_2 planar orientation, which is atypical of F_2 folds in the Tusas Range. L_2 orientations on these atypical folds are also variably oriented, some wrapping F_2 folds. These distributed folds are found primarily within a kilometer of the southern contact with the Vadito Group.

At least four possible interpretations are consistent with these observations. In the first model, the field area may have been folded by two early events that were coplanar. F_2 folds in Domain 2 have undergone less finite strain than Domains 3 and 4, such that the very consistent F_2 folds and L_2 orientations in Domains 3 and 4 are the product of considerable progressive reorientation during the second folding event. The less consistent F_2 and L_2 in portions of Domain 2 would then represent structures in the process of reorientation during a progressive D_2 event. This model however does not explain why F_2 folds in Domains 3 and 4 also have consistent L_2 orientations. If F_2 folds were progressively reoriented, conceivably L_2 lineations would be distributed. In a second model, the overall strain geometry may have varied during D_2 spatially across the boundary, which resulted in a variety of F_2 and L_2 orientations that occurred contemporaneously. The Moppin Complex may have caused this by acting as a rigid body distorting a regional strain field. A third model is a slight modification to the first one in which the few distributed folds in Domain 2 are the product of a dominantly flattening fabric, which accounts for the wide distribution of lineation and fold axis orientation. While this model explains the variability in the discrete zones of Domain 2, finite strain in Domains 3 and 4 show plane strain. A fourth model suggests that the pluton cooled during deformation, producing a localization of strain near the margins of the pluton as its viscosity increased. This model is favored because the volume of the Tres Piedras pluton over a several kilometer area contains F_2 folds aligned in the regional orientation. The distributed folds lie near the southern margin of the pluton.

The timing of D_2 is best constrained by three factors relating to the Tusas Mountain orthogneiss: the interpreted contact aureole of the Tusas Mountain orthogneiss, the possible magmatic flow nature of S_1 in the core of the pluton, and the lack of F_2 folds within the pluton. The fabric that the pluton contains is interpreted to be the regional S_2, but S_1 for the Tusas Mountain orthogneiss. The contact aureole around the southern flanks of that pluton contains little fabric suggesting that it post-dates a majority of the fabric forming tectonism recorded in the wall rocks. The ductility of F_2 folds and possible strain partitioning in the Tres Piedras orthogneiss suggest that D_2 occurred while the pluton cooled. These collectively suggest that D_1 and D_2 were a progressive event in time.

D3

The third generation of folds and fabrics, as well as shear zones, are most significantly developed in the Vadito Group south of the discontinuity. To the north, D_3 effects are restricted to small shear zones that fall along the discontinuity. To the south, F_3 folds are found to form individual folds, and interference patterns with a weak to locally moderately developed S_3. These factors suggest that D_3 had a diminishing affect from south to north across the field area.

Timing of D_3 is poorly constrained within this study area. D_3 structures are very similar to structures observed and dated at ca. 1400 Ma in the southern Tusas Mountains (Bishop et al., 1996) and are therefore interpreted to be Mesoproterozoic in age.

A TECTONIC MODEL

In this model, the continental crust in this location was assembled and tectonized very quickly to stable lithosphere. Deposition of the Vadito Group, Burned Mountain Metarhyolite and Ortega Quartzite occurred ~ 1700 Ma during an extensional phase of the Yavapai orogeny. The change of thickness between the Burned Mountain and Vadito metarhyolites from 100 meters to several kilometers is consistent with the Jones et al. (2009) model of localized sedimentary basins. The ~1693 Ma Tres Piedras orthogneiss is interpreted to be syn-tectonically emplaced to D_1 and D_2. Convergent deformation suggests that basin extension had inverted and that the study area was buried by thrusting or reverse fault motion to the depth of several kilometers within a few million years. The Tres Piedras orthogneiss could be interpreted to be part of the overall episode of Vadito Group volcanism or an additional igneous event that is slightly younger that was emplaced along a sub-horizontal thrust surface during D_1 - D_2 contraction. The crosscutting Tusas Mountain orthogneiss was a late stage intrusive episode of the Tres Piedras pluton near the transition between the end of Yavapai orogeny and the early stages of the Mazatzal orogeny. This date for deformation in the Tusas Mountains is older than those determined by previous workers. D_2 fabrics were then crosscut by a granitic dike ca. 1633 ±28 Ma.

Deformation and metamorphism at 1450-1350 Ma overprinted and reactivated early fabrics in most areas in the southern Tusas Mountains and other Proterozoic uplifts to the south, including the Picuris Mountains. This overprinting characteristic becomes minimal across the lithologic discontinuity and sub-parallel Spring Creek shear zone. The discrete offset of D_2 structures and apparent offset of metamorphic assemblages suggests that the Spring Creek shear zone shares several characteristics with the ~1400 Ma shear zones in Colorado described by Shaw and Karlstrom (1999). This feature is probably Mesoproterozoic, however the age is uncertain at this time. If the Spring Creek shear zone is ca. 1400 Ma in age, this structure, which is isolated from exposed 1400 Ma plutons, would be the most southerly identified zones of localized high strain formed during Mesoproterozoic intracontinental tectonism.

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