Structural/Kinematic and Metamorphic Analysis of the Mesoproterozoic Novillo Gneiss, Tamaulipas, Mexico

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Master of Science

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Structural/Kinematic and Metamorphic Analysis of the Mesoproterozoic Novillo Gneiss, Tamaulipas, Mexico

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ABSTRACT

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Structural/Kinematic and Metamorphic Analysis of the Mesoproterozoic Novillo Gneiss, Tamaulipas, Mexico (117 pp.)

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The Novillo Gneiss is one of several exposures of the Mesoproterozoic (ca. 1 Ga) basement along the backbone of Mexico that have been collectively interpreted as parts of a single crustal block termed Oaxaquia. To test this interpretation, this study presents a structural-kinematic and metamorphic analysis of the Novillo Gneiss so that its tectonic history can be compared to that of the Oaxaquia archetype, the Oaxacan Complex of southern Mexico.

Exposed within the Sierra Madre Oriental near Ciudad Victoria (Tamaulipas State) in northeast Mexico, the Novillo Gneiss comprises two major Mesoproterozoic suites. The older host suite, dated at 1235-1115 Ma, principally comprises: (1) K-feldspar megacrystic metagranites, (2) metagabbros, and (3) a calc-silicate unit. Geochemical data suggest arc/back-arc affinities. The younger suite, dated at 1035-1010 Ma, is interpreted to be an anorthosite-mangerite-charnockite-granite (AMCG) assemblage. Both suites are intruded by two sets of amphibolite dikes, the earlier of which predates deformation and metamorphism under granulite facies conditions at ca. 990±5 Ma, whereas the later set is of low grade and was emplaced at ca. 546 Ma.

Structural-kinematic and metamorphic analysis of the Novillo Gneiss is consistent with a tectonic history involving eight major events: (1) Emplacement of host rock at 1235-1115 Ma, (2) coeval migmatization, (3) intrusion of the AMCG suite at
1035-1010 Ma coeval with sheath folding in the AMCG suite and mylonitization of the host rocks, which show lineation rotation patterns consistent with intrusion into a zone of sinistral transpression with top-to-the-SW kinematics, (4) intrusion of earlier (garnet) amphibolite dikes in the interval 1010-990 Ma, (5) reburial to produce static granulite facies metamorphism and overprinting granoblastic textures at ca. 990 Ma, (6) rapid exhumation, (7) intrusion of later (porphyritic) amphibolite dikes at ca. 546 Ma, and (8) late Paleozoic dextral thrust juxtaposition against the Granjeno Schist, a polydeformed assemblage of Paleozoic metasedimentary and metavolcanic rocks, and serpentinized mafic-ultramafic units.

This tectonic history is closely comparable to that of the Oaxacan Complex, supporting the Oaxaquia model. Minor structural differences between the two suggest that the latter represents a more strongly sheared and slightly differently oriented component of the same tectonic package. The most striking difference between the two is the absence of the later porphyritic amphibolite dikes in the Oaxacan Complex, which may be linked to their relative paleogeography at the time of dike emplacement.

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INTRODUCTION

To the east of Ciudad Victoria in Tamaulipas State, northeastern Mexico, Precambrian rocks of Mesoproterozoic age (ca. 1 Ga Novillo Gneiss) outcrop in the exposed core of the Huizachal-Peregrina anticlinorium, a large northwest-trending Laramide structure located in the front ranges of the Sierra Madre Oriental (Ramírez-Ramírez, 1974, 1992; Ortega-Gutiérrez, 1978). The Novillo Gneiss represents one of four outcrops of Mesoproterozoic (Grenville-age) basement in eastern Mexico (Figure 1). Although separated by as much as 700 km, these ca. 1.0-1.2 Ga gneisses (Lawlor et al. 1999; Keppie et al., 2003; Solari et al., 2003; Cameron et al., 2004) have a number of characteristics in common, including pervasive granulite facies metamorphism, abundant anorthosite complexes, and a northwest trending structural orientation. The gneisses have been proposed to represent isolated exposures of an extensive Mesoproterozoic (ca. 1 Ga) basement underlying most of eastern Mexico (Ortega-Gutiérrez et. al., 1995), and have figured significantly in paleocontinental reconstructions for the late Precambrian and Paleozoic. Early workers considered this basement to be an extension of the ca. 1 Ga Grenville orogen of eastern Laurentia (e.g. Cserna, 1971; Shurbet and Cebull, 1987). However, paleomagnetic data (Ballard, 1989) and faunal evidence (Robison & Pantoja Alor, 1968; Rowley & Pindell, 1989; Stewart et. al., 1999) are inconsistent with this correlation and, instead, favor correlation with Gondwana.

On the basis of this and available isotopic and geochronological data, Ortega-Gutiérrez et al. (1995) proposed that this portion of the Mexican basement originated as a microcontinent termed ‘Oaxaquia’ that evolved near Gondwana in the Early Paleozoic.
Oaxaquia is thought to have been close to Laurentia by the Early Mississippian (Stewart et al., 1993), based on faunal considerations, and had firmly accreted to Laurentia by the Permian, based on stratigraphic relationships (Gursky et al., 1989) and paleomagnetic data (Torrez et al., 1993; Sanchez-Zavala et al., 1999; Keppie and Ramos, 1999).

Oaxaquia is inferred to extend from northern Mexico to Central America, a distance of some 2000 km (Manton, 1996), making it a major segment of the worldwide belt of Mesoproterozoic orogenesis (Ortega-Gutiérrez et al., 1995; Lawlor et al., 1999). However, on the basis of Pb isotope data, Ruiz et al. (1999) have suggested separate origins for the northernmost exposure of Oaxaquia (the Novillo Gneiss), which isotopically resembles gneisses of the Laurentian Grenville belt in Texas, and the Grenville-aged gneisses of southern Mexico, (Huiznopala Gneiss, Oaxacan Complex, and the Guichicovi Gneiss), which are isotopically similar to those of the Colombian Andes. This would mandate a suture between the Novillo Gneiss and the Huiznopala Gneiss and, hence, would discredit the concept of the Oaxaquia microcontinent. If substantiated, this would have important implications for paleocontinental reconstructions. More recent Pb data (Cameron et al., 2004) show the northern and southern exposures to have a similar isotopic signature, lending support to the concept of a single crustal block.

Recent paleocontinental reconstructions for the Mesoproterozoic supercontinent Rodinia have focused on possible tectonic linkages between western South America and eastern North America, (Bond et al., 1984, Hoffman, 1991; Moores 1991; Dalla Salda et al., 1992; Park 1992; Keppie, 1993; Dalziel et al., 1991, 1997; Keppie et al., 1996). Of critical importance to these reconstructions are the small continental fragments, or tectonic tracers, that lie between the more coherent cratonic cores. Oaxaquia, if it existed, would form one of
the largest of these fragments, hence the question of its existence is central to plate tectonic models for the Late Precambrian-Paleozoic.

Studies of the Mesoproterozoic complexes in southern Mexico (Herrmann et al., 1994; Ruiz et al., 1999) support their Gondwanan affinities. But while paleomagnetic, faunal, and Pb isotopic data from northern Mexico support the existence of Oaxaquia (Ballard, 1989; Stewart et al., 1999; Cameron et al., 2004), some isotopic data does not (Ruiz et al., 1999). What is lacking are the data needed to resolve this conflicting issue. This study proposes to contribute to such a database by applying modern structural-kinematic and metamorphic analysis to the Novillo Gneiss and then comparing these data with the better-known Oaxacan Complex (Solari, et al., 2003; Keppie et al., 2003) to the south, the Gondwanan affinities of which are well documented. In doing so, it is hoped the study will contribute to existing plate tectonic models for the region and perhaps shed new light on the nature of the Grenville orogeny.

Present Study

In the present study, the Novillo Gneiss was mapped in the cañon de Novillo, which cuts NE-SW through part of the Sierra Madre Occidental, west of the city of Ciudad Victoria in the state of Tamaulipas (Figure 1). Mapped exposures of the Novillo Gneiss are limited to the bed of the Novillo River and several of its tributaries, as well as sporadic exposures along road cuts that follow the canyon along much of its length. The Novillo Gneiss is bordered to the north along a major fault by the Granjeno Schist, mapped by Dowe (2004). It is bordered to the south by Mesozoic units, which overlie the gneiss unconformably.
Figure 1. Location Map. The four areas where the Grenvillian granulites are exposed in eastern and southern Mexico are Ciudad Victoria (Novillo Gneiss), Molango, The Oaxacan Complex, and La Mixtequita (Guichicovi Gneiss). (from Lawlor et. al., 1999).
Within this narrow belt, the various lithologies of the Novillo Gneiss were mapped and measurements were made of the various structural features of the gneiss including metamorphic layering, foliations, mineral lineations, folds, shear zones and faults.

In addition, a suite of oriented hand samples was collected for subsequent petrographic and structural analysis. These data were collected in order to determine the deformational-metamorphic history of the Novillo Gneiss and so provide a framework for comparison with the better-known Oaxacan Complex. The results of this study are presented in the following chapters.
MEXICAN TERRANES

Terrane analysis was first applied to Mexico by Campa and Coney (1983) on the basis of lithology, deformation, metamorphism and age. Campa and Coney (1983) grouped the terranes of Mexico into three subdivisions based on their provenance: (1) terranes of North American provenance – a northwestern zone considered to be a direct continuation southward into Mexico of autochthonous North American Precambrian basement and its Paleozoic/Mesozoic cover; (2) terranes of Gondwanan provenance – an eastern zone surrounding the Gulf of Mexico of mainly Late Paleozoic age, which while heterogeneous, was considered to have had a common origin as material accreted to North America during the latest Paleozoic Appalachian-Ouachitan-Marathon orogeny; and (3) terranes of Pacific provenance – a western zone making up the wider Pacific margin of Mexico, which is characterized by a heterogeneous assemblage of mainly submarine volcanic and sedimentary rocks of late Mesozoic age, with presently unknown basement.

Sedlock and Ortega-Gutiérrez (1994) reassessed Campa and Coney’s (1983) work, redefining some terrane boundaries and assigning new names to others (Figure 2). They identified sixteen terranes that are broadly similar to those of Campa and Coney (1983) but placed more consideration on their tectonic evolution. Among the Gondwanan terranes, the Coahuila and Sierra Madre terranes were each split into two, and the Mixteca and Oaxaca terranes were recognized as having a Gondwanan (rather than Pacific) provenance. The Alisitos and Guerrero terranes were also each split into two, and the Vizcaíno terrane was subdivided into four subterranes. In addition, whereas most tectonic models for Mexico dealt with only a few continental blocks and focused on Mesozoic events associated with the
Figure 2. Terrane map of Middle America showing the location of the Novillo Gneiss (modified after Keppie (2004).
breakup of Pangea, Sedlock and Ortega-Gutiérrez (1994) proposed a tectonic model for Mexico beginning in the Late Neoproterozoic (600 Ma).

Dickinson and Lawton (2001) subsequently subdivided the Mexican terranes into eight, coherent Permian to Cretaceous crustal blocks, but only addressed the tectonostratigraphic record following Pangea amalgamation.

These historical approaches were further advanced by Keppie (2004) with a view to clarifying the Mesoproterozoic-Permian gap in earlier tectonic models. A new terrane in Keppie’s (2004) compilation that is of particular relevance to this study is that of Oaxaquia. This name was first coined by Ortega-Gutiérrez et al. (1995) to describe a coherent ca. 1 Ga crustal block thought to underlie much of Mexico. However, Keppie’s (2004) definition of the Oaxaquia terrane is extended to include Paleozoic Gondwanan sedimentary rocks that unconformably overlie this basement and are, in turn, unconformably overlain by Carboniferous-Permian rocks and overstepped by Mesozoic-Cenozoic strata.

The Oaxaquia Terrane

The Oaxaquia terrane, or Oaxaquia, as defined by Ortega-Gutiérrez et al. (1995) and Keppie (2004), forms the basement of the Oaxaca, Mixteca, Juarez and Sierra Madre terranes as well as major parts of the Maya and Coahuila terranes. It is bounded by: (1) the early Mesozoic Juarez terrane to the east, (2) the upper Paleozoic Juchatengo terrane to the south, and (3) the lower Paleozoic Mixteca terrane to the west (Centeno-Garcia et al., 1999). Outcrops occur in the states of Tamaulipas (at Ciudad Victoria), Hidalgo (at Molango), Oaxaca (at Oaxaca and La Mixtequita) and Chiapas, and PEMEX boreholes indicate the
presence of the basement complex from Tamanzunchale to Tampico (Ortega-Gutiérrez et al., 1995).

Since outcrops of the Oaxaquia terrane are sparse, its description is usually based on the Oaxacan Complex (e.g. Keppie, 2004). This complex is the largest exposure of Oaxaquia and consists of paragneisses (marble, calc-silicate, quartzofeldspathic gneiss and graphitic gneiss) and orthogneisses (anorthosite, charnockite, amphibolite and pegmatite) (Ortega-Gutiérrez, 1984, Keppie et al., 2003). These basement rocks are overlain nonconformably by unmetamorphosed Paleozoic siliciclastics and their Mesozoic cover.

**Basement Units**

The Oaxaca Complex is made up of three structural gneissic units that have respectively yielded U-Pb zircon protolith ages of $\geq 1350$ Ma (orthogneiss), $>1157-1130$ Ma (paragneiss and orthogneiss), and ca.$1012 \pm 12$ Ma (anorthosite-mangerite-charnockite-granite [AMCG] suite) (Solari et al., 2003; Keppie et al., 2003). Geochemical data (REE, Nd and Pb isotopes) suggests that the ca. 1.0-1.2 Ga igneous protoliths constitute a magmatic arc intruded by an intraplate AMCG suite, (Keppie 2001, 2003, 2006; Solari et al., 2003). U-Pb isotopic analysis of zircon shows the oldest unit was migmatized at 1106 ± 6 Ma and that all units were polydeformed and affected by granulite facies metamorphism at ca. 990 Ma.

In the oldest gneissic unit ($\geq 1350$ Ma El Catrin unit), the ca. 1106 Ma migmatite fabric is preserved only within a ‘low-strain window’, where it is represented by stromatic to nebulitic, cm-scale mesosome and leucosome banding. Solari et al. (2003) interpreted this banding to be the product of metamorphic differentiation and migmatization that overprinted an original, banded magmatic fabric. The leucosome consists of quartz, alkali feldspar,
plagioclase (An$_{20-45}$), secondary epidote, calcite and chloritized biotite. The mesosome contains additional augitic clinopyroxene, hypersthene, and rare hornblende and scapolite. The migmatitic fabric grades into striped/banded gneiss associated with a granulite facies metamorphic foliation. The mineralogy suggests that the protolith of the migmatite was gabbro-diorite.

The younger, ca. 1157-1130 Ma El Marquez unit contains paragneisses consisting of quartz-feldspar-garnet gneiss, two pyroxene-quartz-feldspargneiss, and mica-graphite-sillimanite-rutile gneiss, marbles and calc-silicates. Intruding the paragneiss are minor igneous bodies of amphibolite and pegmatite, as well as major charnockitic, meta-syenitic and meta-gabbro units. The generally granoblastic texture of the parageneses suggests that both the paragneiss and orthogneiss underwent granulite facies metamorphism that overprinted the original fabrics (Solari et al., 2003).

The youngest, ca. 1012 Ma Huitzo unit is an AMCG suite that consists of meta-anorthosite and intercalated Fe-metadiorite, metagabbro, mafic cumulates and garnet-bearing charnockite. The Fe-metadiorites and anorthosites are intimately interleaved as a result of intense deformation that reoriented primary magmatic contacts parallel to the foliation.

All of these units experienced a second tectonothermal event that polydeformed and metamorphosed them to the granulite facies. This event is bracketed between the 1012$^{±3}$Ma protolith age of the charnockite in the Huitzo AMCG suite and the 977$^{±2}$ Ma age of a post-tectonic blue-quartz pegmatite.

The subsequent metamorphic evolution of the Oaxaca Complex as determined by $^{40}$Ar/$^{39}$Ar analysis of hornblende, is marked by rapid cooling through $\sim$500°C by 977$^{±12}$ Ma,
implying a rapid transition from granulite to amphibolite facies (Keppie et al., 2003 Solari et al., 2003).

Cover Rocks

Unconformably overlying the Mesoproterozoic basement rocks of the Oaxaca Complex is a sequence of four unmetamorphosed Paleozoic units ranging in age from Cambrian to Permian or younger. U-Pb detrital zircon data from these rocks are more consistent with a South American provenance than with a North American one (Gillis et al., 2005).

The Uppermost Cambrian-Lower Ordovician Tiñu Formation lies directly on the basement and consists of interbedded, marine limestone and shale that grade upwards into predominantly shale and siltstone containing fauna of Gondwanan affinity (Landing et al., 2007). Tectonically overlying this formation is the Mississippian Santiago Formation comprising quartz-rich calcareous sandstone and conglomerate overlain by marine limestone, calcareous siltstone and shale. Associated fauna are of Laurentian (Mid-Continent) affinity (Centeno-Garcia and Keppie 1999). The Santiago Formation is tectonically overlain by the Lower-Middle Pennsylvanian Ixtaltepec Formation comprising shale, silt, sandstone and limestone.

Unconformably overlying the Ixtaltepec Formation is the Yododene Formation consisting of conglomerate sandstone, siltstone and minor unfossiliferous shale. This formation is of uncertain age but the presence of Lower Permian limestone clasts in the conglomerate suggests it is Permian or younger (Centeno-Garcia et al., 1999).
Unconformably overstepping the Paleozoic rocks are Lower Cretaceous marine limestone and Tertiary red beds, (Centeno-Garcia et al., 1999). The Mesozoic rocks form part of a cover succession across much of Mexico.
LOCAL GEOLOGICAL SETTING

The Novillo Gneiss of northeastern Mexico is considered to be the northernmost exposure of Oaxaquia and comprises Mesoproterozoic basement rocks overlain by a Paleozoic cover succession, (Ortega-Gutiérrez et al., 1995; Lawlor et al., 1999). The basement rocks are located just west of Ciudad Victoria in three canyons of the Sierra Madre Oriental, the cañon de la Peregrina, the cañon de Caballeros and of greatest importance to this study, the cañon de Novillo. The gneiss is exposed within the Huizachal-Peregrina anticlinorium, a large northwest-trending Laramide structure located in the front ranges of the Sierra Madre Mountains. The gneiss itself is exposed in the core of the anticlinorium and comprises a granulite facies complex of ca. 1 Ga age that is interpreted to be correlative with the Oaxacan Complex. The local geologic framework of the complex is provided by the unpublished doctoral research of Ramirez-Ramirez (1992).

The Novillo Gneiss itself occupies an area of about 35 square kilometers and comprises a succession of consistently NE-dipping units that includes an intrusive AMCG suite that is flanked to the northeast and southwest by a suite of host orthogneisses and paragneisses. Both suites are intruded by cross-cutting mafic dikes of two generations (Figure 3).

The Novillo Gneiss is tectonically juxtaposed to the west against low-grade Paleozoic rocks of the Granjeno Schist, which make up part of the Sierra Madre terrane (Figure 4). This polydeformed assemblage of metasedimentary and metavolcanic rocks, and serpentinized mafic-ultramafic units is interpreted to be an oceanic accretionary prism possibly associated
Figure 3 Geological map of the Huizachal-Peregrina anticlinorum and cross section. Pre-Mesozoic units are exposed in the core. The red rectangle indicates the present study area shown in Figure 4. The cross section is along the line labeled A-A’ line on the map. (modified from Ramirez-Ramirez, 1992).
Figure 4. Geological map of study area: Canon de Novillo.
with the Late Paleozoic closure of the Rheic Ocean (Dowe, 2004; Nance et al., 2007).

A leucogranite with a poorly constrained U-Pb zircon age of 351 ± 54 Ma (Dowe et al., 2004) has been emplaced into the fault zone separating the Granjeno Schist from the Novillo Gneiss.

To the east, the Novillo Gneiss is unconformably overlain by an unmetamorphosed Paleozoic (Silurian-Permian) sequence of marine clastics (Stewart et al., 1999). In ascending order, this sequence includes: (1) Silurian conglomerate, silicic volcanic arenite, limestone, sandstone and siltstone, the latter containing marine fauna of Gondwana affinity, (2) Lower Mississippian shallow-marine, quartz and lithic arenite, siltstone and shale containing Laurentian fauna, (3) Mississippian rhyolite, (4) an unconformably overlying Lower and Middle Pennsylvanian unit consisting of bioclastic grainstone, sandstone, and limy sandstone, and (5) a lower Permian unit consisting of turbiditic siltstone and sandstone rich in volcanic detritus. The Paleozoic succession is unconformably overlain by Mesozoic (Lower Jurassic - Cretaceous) conglomerate, limestone, siltstone and sandstone. Cretaceous-Early Cenozoic thin-skinned tectonics (Laramide orogeny) have deformed the Mesozoic sequence above the Jurassic, folding and displacing the overlying strata along northeast-directed thrusts (Zhou et al., 2006).

Basement Units of the Novillo Gneiss

U-Pb zircon dating has shown that the Novillo Gneiss is composed of two major Mesoproterozoic suites (Cameron et al., 2004). The oldest is dated at 1235-1115 Ma and principally comprises: (1) K-feldspar megacrystic metagranites, (2) metagabbros, and (3) a calc-silicate unit. Geochemical data suggest arc back-arc affinities (Lawlor et al., 1999;
Cameron et al., 2004; Keppie et al., 2006). The second suite, dated at 1035-1010 Ma, is interpreted to be an anorthosite-mangerite-charnockite-granite (AMCG) suite (Cameron et al., 2004). Both suites were intruded by a set of undated dikes prior to being polydeformed and metamorphosed under granulite facies conditions, dated at ca. 990±5 Ma on the basis of metamorphic zircon growth (Cameron et al., 2004). Peak values for metamorphic pressure and temperature have been estimated at 8.9-9.7 kbar and 730-775°C (Orzoco et al., 1991).

A second set of low-grade mafic dikes intrudes all the high-grade rocks of the Novillo Gneiss and has yielded a hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 546 ± 5 Ma that is considered to closely post-date their emplacement (Keppie et al., 2006). The geochemical signature of the dikes suggest an intraplate magmatic source that has been linked to either a mantle plume or to decompression melting during passive asthenospheric mantle upwelling associated with lithospheric extension (Keppie et al., 2006).

A cooling history for the Novillo Gneiss has been estimated by Keppie et al., (2006) on the basis of a U-Pb titanite age of 928 ± 2 Ma and a $^{40}\text{Ar}/^{39}\text{Ar}$ biotite age of 697 ± 10 Ma from a paragneiss. These data suggest the Novillo Gneiss cooled from ~660°C to 325 ± 25°C in ca. 230 Ma, a cooling of rate of ~1.45°C/yr.
PETROLOGY

The Novillo Gneiss consists of both paragneiss and orthogneiss. With the exception of a calc-silicate unit, however, the Novillo Gneiss in the cañon de Novillo is almost entirely orthogneiss. The orthogneiss includes both mafic and potassic granite units, which form part of the host suite, and intrusive meta-anorthosite, metagranite and metabasite, which form part of a younger anorthosite-mangerite-charnockite-granite (AMCG) suite.

Metamorphic leucocratic and mesocratic segregated layering occurs throughout the Novillo Gneiss, consistently trending northwest and dipping northeast. However, a general distinction can usually be made between the AMCG suite, which shows continuous, well-defined, rhythmically alternating leucocratic-mesocratic layering, and the older host suite, in which layering is discontinuous and defined by large (mm-cm scale), light colored, quartzofeldspathic augen within a darker, foliated mesosome matrix. However, stretching and flattening of the augen in the host suite is often sufficient to form a distinctive flaser-like layering. Calc-silicate within the host suite is an exception and is largely massive.

In the following petrographic descriptions, the leucosome and mesosome components of the suite are described separately. However, in postulating the protolith of each unit, the whole-rock mineral assemblage is considered.
Host Suite

The older (ca. 1.1-1.2 Ga) host suite mainly comprises: (1) granite gneiss, (2) garnet K-feldspar augen gneiss, (3) amphibolite, and (4) calc-silicate.

Gneisses of the host suite generally show a flaser-mylonitic layering with large (mm-cm scale) quartzofeldspathic augen that are often extensively elongated and sheared. Leucocratic and mesocratic compositional layering in the host suite is easily discernible and generally shows sharp, clear boundaries. However, some lithological contacts show cataclastic textures with grain-size reduction and brittle deformation.

Microscopically, granoblastic grain textures and peak mineral assemblages are generally better preserved in more competent leucocratic layers, whereas the mesocratic layers are more strongly retrogressed. Retrograde overprinting of peak, high-grade mineral assemblages also increases with cataclastic deformation, to which the mesocratic layers are more susceptible.

Granite Gneiss

The granite gneiss of the host suite is a quartzofeldspathic orthogneiss that shows continuous, leucosome-mesosome segregation layering. Unlike the other host gneisses however, the leucosome-mesosome boundaries are stromatic, indistinct and contorted. The layer boundaries are also folded into tight, highly irregular, asymmetric folds on a mm-cm scale (Plate 1). In addition, these folds possess an axial planar fabric that further obscures the leucosome-mesosome layer boundaries and is at a high angle to the overall orientation of the layering.
Plate 1. Characteristic migmatic textures: (a) highlighted, irregular, wavy or ‘flame’ textures between leucocratic-mesocratic boundaries, (b) magnified.
The leucosome of the granite gneiss comprises 35% alkali-feldspar, 30% microcline, and 35% quartz, and so lies compositionally within the granite field. The preservation of possible orthopyroxene, however, suggests that some of the granite gneiss is charnockite.

The microtexture of the leucocratic layers is dominated by equigranular K-feldspar (microcline) and plagioclase laths showing an overall granoblastic fabric. The plagioclase is extensively replaced by epidote and white mica (± calcite), often along cleavage and/or twinning twins to give a ‘gefüllte’ texture. Occasional plagioclase cores are unaltered. Microcline, in contrast, is largely unaltered and shows characteristic tartan twinning. Quartz occurs both as anhedral gains and, more conspicuously, as large recrystallized ribbons with elongation ratios of up to 10:1 in thin sections cut parallel to the XZ plane of strain. Rarely, these quartz ribbons appear to have been tightly microfolded (Plates 2 and 3). Quartz displays subgrains and undulose extinction that often plays across subgrain boundaries, indicating syn- or post-subgrain deformation. The quartz also often contains rutile needles, a texture typical of high temperature crystallization under granulite facies conditions.

In addition to the mineralogy of the leucosome, the mesosome contains large clusters of euhedral garnets and minor amounts of retrograded mafic minerals. The garnet clusters average 0.3 mm-1 mm in size and are more highly concentrated along leucosome-mesosome borders. The garnets are essentially unaltered with only slight epidote/chlorite retrogression along micro-fractures. Rare chlorite shows Berlin-blue birefringence in a fanning radiating-acicular habit, presumably replacing pyroxene.
Plate 2. Photomicrograph of migmatite, note quartz ribbon across entire top of picture and granoblastic texture in the remainder showing triple point junctions. Field of view 4mm.
Plate 3. Photomicrograph of migmatite showing quartz ribbon microfold and surrounding granoblastic texture. Plagioclase is pockmarked with retrogradation (gefüllte texture). Microfold is marked by yellow line. Note quartz defining the fold is recrystallized with sharp extinction contrast between mildly strained subgrains. Field of view 4mm.
The modal composition of the mesosome is 35% garnet, 45% feldspar, 15% quartz, and 5% replaced pyroxene with trace opaques. This mineralogy lies in the diorite field and, in combination with the leucosome, suggests a whole-rock with the composition of granite or granodiorite.

The concentration of garnet along leucosome-mesosome layer boundaries is interpreted to be a residual phase of an initially migmatically segregated leucosome. The flow-like nature of the mesosome-leucosome layer boundaries is also interpreted to be a relict migmatitic feature (Plate 1). The existence of quartz-ribbons, on the other hand, indicates mylonitization, whereas the dominating granoblastic texture indicates a subsequent phase of extensive recrystallization. The retrogression of plagioclase and the replacement of pyroxenes by chlorite indicate a late low-grade greenschist metamorphism that likely coincided with the development of the quartz subgrains.

Garnet K-Feldspar Augen Gneiss

A large portion of the host suite comprises garnet K-feldspar augen gneiss. This gneiss displays a prominent ‘zebra-stripe’, flaser-mylonitic layering at outcrop scale and makes up much of the host rock to the northeast and southwest of the main intrusive anorthosite.

The leucocratic augens are white to rarely pink, elongate and tapered, often asymmetric with a common kinematic sense of shear, but just as often symmetrical (Plate 4). They are separated by relatively thin mesocratic bands. The proportion of leucocratic
Plate 4. Cut samples of augen gneiss, note alternating quartzofeldspathic, elongated augen within mesocratic matrix.
augen to mesocratic bands is typically between 60:40 and 70:30 but varies throughout the unit.

In thin section, the leucocratic layers comprise plagioclase + microcline + minor quartz and trace opaques. The mode for the leucocratic bands is 60% K-feldspar, 35% plagioclase, and 5% quartz with trace opaques. The feldspars show a granoblastic texture with little or no expression of the augen-like shape they display in outcrop. The plagioclase is somewhat retrograded to fine-grained epidote and white mica displaying gefüllte texture. Quartz is anhedral and shows undulose extinction. The mesocratic layers in thin section are defined by large garnet bands and quartz ribbons with pyroxenes (orthopyroxene?) partially retrograded to chloritic minerals and minor opaques. The garnet bands and quartz ribbons are parallel to, and partially define, the metamorphic layering. Garnet shows minor retrograde metamorphism to chlorite along micro-fractures. The quartz ribbons show elongation ratios up to 11:1, but are recrystallized, the individual quartz grains displaying moderate undulose extinction that often plays across subgrain boundaries. The mode for the mesocratic bands is 80% garnet, 15% quartz, and 5% chlorite after (?) pyroxene and trace opaques.

Taken together, normalized leucocratic and mesocratic compositions combine to give a whole-rock mode of 42% K-feldspar (microcline), 25% plagioclase, 24% garnet, 8% quartz, and 5% pyroxene. This suggests a protolith of monzonite composition.
Amphibolite

In outcrop, this unit is very thinly laminated and dark green in color. The dark green mesocratic layers contain highly elongated leucocratic feldspathic lensoids that show no obvious sense of shear. The proportion of leucocratic to mesocratic material is approximately 35:65. Folds of ~12-15cm wavelength that deform the feldspathic lensoids are not uncommon (Plate 5) and, where cut parallel to the fold profile, show hinges with secondary, pistachio green mineralization along mesocratic layers.

In thin section, extensive retrogression accompanies the folded layers. Mesocratic layers are almost completely retrograded and leucocratic layers are extensively replaced with a strong gefüllte microtexture. The microtexture of the amphibolite is mylonitic-cataclastic and extensively overprinted by multiple generations of retrograde mineralization resulting in a highly turbid, cataclastic grain mosaic. Grain size in leucocratic layers is generally much larger than in mesocratic layers.

Leucocratic lensoids consist of plagioclase feldspar but are extensively overprinted. Some large, elongated relict feldspar crystal forms are weakly visible by alternating plane light and cross-polar views. Relict plagioclase twinning is very weakly visible in a few crystals. Quartz shows strong undulose extinction, extensive subgrain development and fracturing. Elongate quartz ribbons are recrystallized and cataclastically grain-size reduced.

The mesocratic layers are completely altered and deformed, making original mineralogy difficult to determine. However, the low-grade assemblage of chlorite minerals and opaques were presumably derived from higher-grade amphibole and,
Plate 5. Amphibolite sample cut along fold profile plane showing thin mesocartic/lecocratic layering and green mineralization.
possibly, pyroxene. The opaques are battleship grey in color under plane light and crossed polars, and form a coherent layering in thin section except where they are sharply cut and displaced by small faults. Pyroxene(?) relicts are very rare and almost entirely replaced by chlorite. The replacement mineralization is very fine even under high magnification. It is bright green and yellow in color with a green and blue birefringence and likely includes chlorite and epidote. Quartz grains are often large, elongate and highly subgrained with undulose extinction, but also occur as long aggregates of undulose quartz microcrystals. This habit is probably the result of cataclastic deformation of elongated quartz crystals causing subgrain development and grain-size reduction. Occasional late calcite veinlets cut across all features.

The mineral assemblage plagioclase + pyroxene (?) + opaques + quartz in the amphibolite indicates a mafic protolith.

_Calc-silicate_

The only metasedimentary unit observed in the field area is a massive calc-silicate. It’s contact relationships with other units were not seen, but it is located among rocks of the older host unit. It is composed of approximately 65% carbonate, 15% diopside, 5% scapolite (or wollastonite) with grossular garnet coronas, and 5% quartz with trace opaques and pyroxene(?).

The texture of these minerals is generally interlobate-inequigranular, although in areas of homogenous carbonate, it is equigranular-granoblastic (Figure 6).
Plate 6. Photomicrographs of Cale-silicate highlighting 120 triple junction in yellow. Field of view 4mm.
Carbonate grains display characteristic third order birefringence. Diopside is euhedral and essentially unaltered with characteristic crystal shapes and perpendicular prismatic cleavages. Scapolite (or wollastonite), on the other hand, is highly altered and takes the form of a low relief, colorless mineral surrounded by garnet atolls and complete coronas where it is in contact with carbonate. This may reflect the reaction (Mathavan, 2001):

\[
\text{Scapolite(?) (and/or Wollastonite)} + \text{Carbonate} + \text{Quartz} = \text{Grossular (garnet)} + \text{CO}_2
\]

Garnet in calc-silicates appears at relatively high temperatures (Winkler, 1979) and, in thin section, shows some evidence of retrogression, the larger garnets being weakly altered to finely disseminated clinozoisite with pale yellow pleochroism and blue birefringence. Quartz occurs as both isolated, anhedral grains and as long clusters of small quartz grains, that are likely the recrystallized relicts of quartz ribbons, all showing moderate undulatory extinction. Opaques occur in trace amounts, disseminated throughout the thin section. The peak mineral assemblage of carbonate + diopside + scapolite + quartz suggests an impure, siliceous carbonate protolith and is consistent with granulite facies metamorphism.
Petrological Interpretation of Host Evolution

Based on its petrology, the following geological evolution is proposed for the ca. 1.0-1.2 Ga host suite:

(1) Deposition of the paragneiss protoliths and intrusion of garnet K-feldspar granite and amphibolite protoliths.

(2) Migmatization and development of granite gneiss protolith. This is suggested by the convoluted nature of the layer boundaries between the leucosome and mesosome components and by the residual concentration of garnet along these boundaries. This migmatization is potentially equivalent to the 1106 ± 12 Ma Olmeacan migmatization event in the Oaxacan Complex (Solari et al., 2003).

(3) Subsequent mylonitization of the migmatite and other host units. This is indicated by the presence of recrystallized quartz ribbons and is thought to have occurred prior to the development of peak metamorphic conditions during the interval ca. 1000-980 Ma.

(4) Recrystallization of all pre-existing micro-fabrics by granoblastic textures. This is thought to coincide with peak metamorphic conditions and the paragenesis of the high-grade mineral assemblage. Peak (granulite facies) metamorphic conditions have been dated at ca. 990 ± 5 Ma on the basis of metamorphic zircon growth (Cameron et al. 2004).

(5) Rapid uplift or exhumation. This is indicated by retrogression to greenschist facies conditions in the absence of intervening amphibolite facies assemblages.
AMCG Suite

Orthogneisses of the younger AMCG suite are well layered with alternating cm-m scale leucocratic and mesocratic bands characterized, in part, by numerous, large purple-red garnets and garnet clusters (Plate 7). Cameron et al., (2004) suggested that these orthogneisses formed part of an anorthosite-mangerite-charnockite-granite (AMCG) suite, but in the study area, only layered charnokitic gneiss and anorthosite-metagabbro is exposed. Nelsonite veinlets (ilmenite and apatite) were also observed.

Mineral grains in the well-layered AMCG suite are generally polygonal and granoblastic with triple-point crystal junctions in both mesocratic and leucocratic layers. Indeed, recrystallization to a granoblastic texture is so extensive that deformational features such as folds and mineral lineations are more easily observed macroscopically than microscopically.

Charnokitic Gneiss

In outcrop, the charnokitic gneiss is well layered with alternating, coherent, leucocratic and mesocratic bands. The rock texture is equigranular-granoblastic with common 120 degree-triple point junctions, little retrograde mineralization, and little or no microdeformation. No contact relationship was observed.

The whole-rock modal mineralogy is 25%-40% plagioclase, 30% K-feldspar, 10%-15% quartz, 5%-20% garnet, >5%-15% pyroxene and 5%->10% opaques with trace rutile (as needles in quartz) and zircon (Plate 8) Ranges in the modal percentages reflect the dramatic change in mineral proportions from leucocratic to mesocratic layers. The mineral components are similar in both layers but their proportions vary greatly. Leucocratic layers are principally composed of K-feldspar + plagioclase + quartz,
whereas the mesocratic layers show a dramatic increase in the proportion of garnet, pyroxene and opaques. The normalized whole-rock modal percentage using a typical ratio of leucocratic to mesocratic layering of 60:40 results in 34% plagioclase, 30% K-feldspar, 13% quartz, 11% garnet, 9% pyroxene (orthopyroxene?) and 4% opaques. This corresponds to the composition of a monzonite, the presence of orthopyroxene making it mangeritic or charnokitic, both of which are characteristic of AMCG suites.

Plagioclase comprises large, lath-shaped grains, with characteristic plagioclase twinning. K-feldspar lacks twinning and is likely orthoclase. Quartz forms equant grains and shows occasional weak undulose extinction. Garnet is euhedral and isotropic. It is sparsely scattered throughout the felsic layers, but forms larger clusters in the mesocratic layers. The garnet is essentially unaltered but is occasionally retrogressed to chlorite along microfractures. Clinopyroxene forms subhedral grains with mutually perpendicular prismatic cleavage. The pyroxene (augite?) grains are largely preserved but often display spotty retrogression to chlorite(?). Opaques are typically amorphous and are almost always associated with garnets. They are most plentiful in the mesocratic layers.

Retrogression is more pervasive in the mesocratic layers. Feldspars, which occur largely as pristine laths in the leucocratic layers, are typically altered in mesocratic layers. Orthoclase is cloudy and plagioclase shows gefüllte texture. The minerals replacing the feldspars are too small to be optically identified but are likely to be white mica (sericite) and epidote group minerals.

_Anorthosite-Metagabbro_

In outcrop, the anorthosite-metagabbro is well layered and leucocratic. The layering is exceptionally coherent, often forming continuous layers across entire
outcrops, and alternates between white-grey and light blue-grey on a cm-m scale. The mesocratic layers contain abundant garnet. The anorthosite-metagabbro is cut by mm scale veinlets of illmenite(?) (Plate 7).

The whole rock mineralogy is approximately 75% plagioclase, 10% garnet, 5% opaques, <5% clinopyroxene, <5% quartz and <5% opaques. This corresponds to the composition of a gabbro-anorthosite. The microtexture is equigranular and granoblastic with polygonal crystals showing triple-point crystal junctions in both the leucocratic and mesocratic layers, (Plates 8 and 9).

Plagioclase shows characteristic twinning but much of it is overprinted by gefüllte texture. Garnets form large, fractured euhedral crystals and crystal clusters. Clinopyroxene and quartz are both anhedral, the quartz showing moderate undulatory extinction. The opaques are dusty(?) and commonly associated with garnet.

Retrogression of plagioclase to sericite and epidote group minerals (gefüllte texture) is more extensive in areas with greater proportions of garnet and pyroxenes, where it tends to obscure grain boundaries and plagioclase twinning. The garnets, however, are largely unaffected by retrograde metamorphism, evidence of which is limited to the presence of chlorite in micro-fractures. Clinopyroxene (augite?) is retrograded in varying degrees to a weakly green pleochroic mineral (actinolite and/or chlorite), but most crystals retain at least part of their original mineralogy.
Plate 7 AMCG outcrop picture. Note alternating meso/leucocratic layers and large fold in middle foreground. Axial surface trace is marked by yellow line.
Plate 8. Photomicrograph of Charnokite (AMCG). Note the granoblastic texture. Field of view 4mm.
Plate 9. Anorthosite photomicrograph showing granoblastic texture with 120 degree crystal junctions highlighted in blue. Note the largely pristine plagioclase on left with characteristic twinning. Field of view 4mm.
Petrological Interpretation of the AMGC Suite Evolution

Based on its petrology, the following geological evolution is proposed for the AMCG suite:

1) Emplacement of the AMGC suite protolith at ca. 1.0 Ga (Cameron et al., 2004). The absence of quartz ribbons in the AMCG suite suggests that emplacement occurred after the mylonitization event that affected the host suite. Cameron et al. (2004) used U-Pb zircon dating and cross-cutting relationships to bracket AMCG emplacement between 1035 and 1010 Ma.

2) Recrystallization of all pre-existing textures during granulite facies metamorphism. This is indicated by the pervasive granoblastic fabric and the fact that the high-grade mineral assemblage shows that the AMCG suite underwent the same granulite facies metamorphism as the host rock. Granulite facies metamorphism in the host suite is dated at ca. 990 ± 5 Ma (Cameron et al., 2004).

Mafic Dikes

Two sets of mafic dikes cut both suites of gneisses. One set comprises garnet amphibolite and was intruded prior to granulite facies metamorphism, whereas the other set comprises porphyritic amphibolite and was emplaced after this metamorphic event.

Garnet Amphibolite Dikes

The earlier set of garnet amphibolite dikes are dark in color and foliated, with large macroscopic garnet clusters. The dikes range in width on a cm-meter scale and
clearly cut the foliation of the Host and AMCG gneisses they intrude. The foliation is
defined mainly by garnet-rich bands and sporadic intrafolial/rootless folds.

The original mineral assemblage is approximately 30% plagioclase, 20%
clinopyroxene or orthopyroxene(?), 20% garnet, 20% quartz and 10% opaques. This
corresponds to a diabase protolith. Plagioclase is highly overprinted by retrograde
mineralization, which obscures the plagioclase twining. Clinopyroxene-orthopyroxene(?)
forms stubby subhedral grains with prismatic cleavage. Garnet is variably distributed,
forming a minor mineral in some areas and dominating others. It has a skeletal, anhedral
habit with numerous inclusions of quartz and opaques. Quartz is anhedral and interstitial,
and shows undulose extinction and subgrain boundaries. Anhedral opaques occur
dispersed within the large garnet masses (Plate 10). The microscopic texture of the garnet
amphibolite is granoblastic-inequigranular, except within the large garnet masses where
intra-garnet quartz and opaques occur as irregular inclusions.

Retrograde metamorphism of plagioclase has produced fine-grained epidote group
minerals, whereas clinopyroxene is altered to actinolite (uralitization) with chlorite.
Garnet appears to have been largely unaffected.

The presence of abundant garnet and the granoblastic texture displayed by the
amphibolites is characteristic of mafic rocks in the granulite facies. Garnet in particular is
a good indicator of high pressure (Barker, 1990). The presence of a foliation overprinted
by granoblastic texture indicates that this set of dikes experienced diviatoric stress prior
to granulite facies recrystallization.
Porphyritic Amphibolite Dikes

The younger set of amphibolite dikes show dm-m scale widths and clearly cut the foliation of the host gneisses. The dikes also often show chilled margins and brittle displacement along their bounding surfaces.

The modal mineralogy is 60% plagioclase (andesite: An40), 40% hornblende and >5% opaques classifying the rock as a hornblende-andesite. These porphyritic dikes show an ophitic texture of large (mm-cm scale), randomly oriented plagioclase phenocrysts in a finer grained, interstitial non-oriented hornblende matrix. In thin section, the plagioclase is fairly pristine but the hornblendes show rare clinopyroxene cores (Plate 11).

Plagioclase mainly exists as large, well-formed lath-shaped crystals with sharp plagioclase twinning and crystal faces. Hornblende comprises the majority of the matrix and shows brown pleochroism and an amphibole cleavage. However, much of the hornblende is replaced by chlorite. The opaques are anhedral and are always associated, and often encased by, hornblende. Rare clinopyroxene cores are preserved within the hornblende grains. Hence, the mineralogy records high temperature replacement of pyroxene by hornblende and later low-grade retrogression of the amphibole to chlorite.

Preservation of original igneous textures and mineralogy in thin section indicates that the younger amphibolites were intruded post-tectonically. In addition, the presence of chilled margins indicates that the host rock was substantially cooler than the dike at the time of intrusion. This is supported by a $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende cooling age of 546 ± 5Ma reported by Keppie et al. (2006), which is taken to closely post-date dike emplacement. The replacement of hornblende by chlorite, however, indicates a static retrograde metamorphic event that post-dates dike emplacement. This may coincide with the
metamorphism of the adjacent Granjeno Schist, which has yielded muscovite cooling ages of 313 ± 13 Ma and 300 ± 4 Ma (Dowe, 2004; Dowe et al., 2005).

Leucogranite

Emplaced between the high-grade rocks of the Novillo Gneiss and the adjacent low-grade rocks of the Granjeno Schist is a leucogranite body composed largely of K-feldspar, plagioclase, and quartz with minor interstitial muscovite and chlorite. The leucogranite has yielded a poorly defined U-Pb lower intercept age of 351 ± 54 Ma (Plate 12) (Dowe et al., 2005).

The leucogranite displays a weak fabric in outcrop defined by matrix minerals that envelope the large feldspar crystals (up to 2-3 mm). In thin section, feldspars are lath shaped, with K-feldspar showing no twinning and plagioclase showing characteristic albite twinning. The quartz is undulose with subgrain boundaries and form large crystals that wrap around the large feldspars. Quartz also occurs as aggregates of grain-size reduced microcrystals. Both forms of quartz define the weak fabric (Plate 12). Rare chlorite with purple birefringence (indicating an iron rich variety) is also aligned with the weak foliation. Muscovite occurs either as flecks within feldspars oriented along the feldspar twinning, or interstitially throughout the matrix defining the weak fabric.
Plate 10. Photomicrograph of garnet amphibolite dike showing pervasive garnet mineralization. Field of view 4mm.
Plate 11. Photomicrograph of porphyritic amphibolite dike showing randomly oriented plagioclase laths and brown hornblende. Field of view is 4mm.
Plate 12. Photomicrograph of leucogranite. Note the general foliation (orientation marked by yellow line) defined by mica and alignment of quartz. Field of view is 4mm.
STRUCTURE

Introduction

The deformation of the Novillo Gneiss has been described in several previous studies (Ramirez-Ramirez, 1974; Ramirez-Ramirez, 1992; Fernando et al., 1995; Cameron et al., 2004), but none of these have attempted to fully characterize and describe the sequence of deformational phases and their associated metamorphic textures using modern structural analysis. To remedy this, various strain features such as metamorphic layering, fold axes and axial planes, foliations, lineations (stretching and mineral), and shear sense indicators were observed and recorded, and their orientations plotted stereographically in order to characterize and separate individual phases of deformation.

In the following discussion, these structural data are individually described from: (1) the host suite, (2) the AMCG suite, (3) the garnet amphibolite dikes, (4) the porphyritic mafic dikes, and (5) the leucogranite. The succession of structures observed within each of these units is then compared and correlated in order to produce a common deformational history for the entire Novillo Gneiss.

Host Suite

The host suite of the Novillo Gneiss in the cañon de Novillo is dominated by orthogneiss and can be subdivided into: (1) granite gneiss, (2) garnet K-feldspar augen gneiss, (3) amphibolite gneiss, and (4) calc-silicate. The following sections describe the structural sequence as determined from overprinting relationships for the first three of
these units. The calc-silicate is largely massive and lacks useful structural features, and so, is not addressed in this section.

**Granite Gneiss**

\( D_{1G} \)

The earliest visible fabric within the granite gneiss is a compositional layering comprising alternating quartzofeldspathic leucosome layers and blue-grey, heavily garnet-speckled mesosome layers (Plate 13). Garnet concentrations along the borders of the leucosome layers resemble what Winkler (1979) described as migmatite ‘resisters’, which concentrate at leucosome margins in migmatites. Compositional layer boundaries also resemble those of migmatites in that they are irregular and stromatic (Plate 1). The compositional layering within the granite gneisses is therefore interpreted to be a relict migmatitic feature (Winkler, 1979; Wimmenauer et al. 2007) and is designated \( S_{1G} \). If so, development of the earliest fabric within the granite gneiss likely accompanied high-amphibolite facies metamorphism (\( M_{1G} \)). Microscopically, the pervasive equigranular-interlobate granoblastic textural overprint masks any primary relationships.

\( D_{2G} \)

Lying within the migmatitic compositional layering is a planar foliation (\( S_{2G} \)) defined by recrystallized quartz ribbons that are indicative of an episode of mylonitization (designated \( D_{2G} \)) at significantly lower temperatures than those that accompanied \( M_{1G} \) (Plate 9). The quartz ribbons show length to width ratios of up to \(~12:1\), measuring as much as \(~1\) mm wide by 12 mm long. The orientation of the current
foliation is broadly perpendicular to the general orientation of the compositional layering (S1G), even though at small scale the layering boundary varies (Plate 13).

The mylonitic foliation defined by the quartz ribbons is axial planar to small-scale (5 mm to 1 cm wavelength) folds (F2G) in the compositional layer boundaries. The folds have tight hinge zones and near-parallel limbs, but show variable geometry because the original migmatitic surface being deformed is irregular (Plate 13).

In thin section, folding of the quartz ribbons may occur in the hinge zones of F2G, (Plate 14) although the pervasive recrystallization makes it difficult to be certain. These folds appear to be isoclinal and rootless, with amplitudes of ~1-2 mm, axial planes parallel to S2G, and axes parallel to the macroscopic F2G hinge lines. Hence, the folds are considered to represent refolding during continuous mylonitization rather than a separate phase of deformation. They are therefore assigned, with F2G, to the same episode of mylonitic deformation that produced the quartz ribbons (D2G).

\(D_{3G}\)

The pervasive granoblastic texture of the granite gneiss indicates an episode of extensive recrystallization, which erased most earlier microdeformational features. This recrystallization extends to the quartz ribbons and, hence, post-dates the mylonitization. It is designated D3G. The preserved quartz ribbon structure suggests that recrystallization was static. This is further supported by common extinction patterns across the limbs of several F2G microfolds, which demonstrate that recrystallization occurred after folding.
Plate 13. Stromatic migmatic flow banding (in cut sample a & b), (S\(_{1G}\)) in Granite Gniess, yellow arrows indicate garnet resisters concentrating along leucosome-mesosome boundaries, blue line indicate quartz ribbon orientation, black arrows indicate quartz ribbon (S\(_{2G}\)).
Plate 14. a) Photomicrograph of migmatite showing recrystalized quartz ribbon, (S₂G) across entire top of picture. Note lack of undulose extinction indicating total recrystallization. Granoblastic texture dominates the remainder of the photomicrograph (note triple point crystal junctions). b) mylonitically stretched and microfolded quartz ribbon (F₂G) showing subsequent subgrain development and minor undulose extinction, surrounded by granoblastic texture resulting from granulite facies recrystallization, (D₃G). Fields of view are 4mm.
$D_{4G}$

Undulose extinction and subgrain development within the granoblastic quartz grains document an episode of deformation (designated $D_{4G}$) that post-dates recrystallization. Expressions range from simple undulose extinction patterns that play across subgrain boundaries to distinct subgrains with separate extinction patterns (Plate 14). The limited extent of undulose extinction and absence of coeval mineral paragenesis suggest a very minor deformational event or tectonic adjustment.

$K$-Feldspar Augen Gneiss

$D_{1K}$

The earliest recognizable fabric in the K-feldspar augen gneiss is a compositional banding ($S_{1K}$) interpreted to be a result of metamorphic differentiation. The banding comprises discontinuous quartzofeldspathic layers that lie within a mesocratic matrix dominated by garnet crystal clusters.

$D_{2K}$

The quartzofeldspathic layers within the K-feldspar augen gneiss are deformed into large (up to 10 cm long), elongated and flattened, lens-like augen, and the mesocratic component has a weak foliation. The layering and foliation are parallel, defining a planar deformational fabric designated $S_{2K}$.

The augen are of slightly flattened prolate shape (i.e., slightly flattened pencil-shaped with a length to width to height ratio ($X:Y:Z$) of up to 12:2:1 cm, (Figure 5 and Plate 15). The ratio shows that extension along the X-axis is extreme ($x/z =12:1$).
Only a few augen show a convincing sense of shear in the XZ plane, the majority of these being dextral. Nevertheless, they are fairly consistent in shape at outcrop scale and can be used as strain markers. The above ratio (12:2:1) can therefore be considered a measure of the strain ratio, corresponding to the X:Y:Z directions of the strain ellipsoid.

S$_{2K}$ is parallel to the flattening (XY) plane of the quartzofeldspathic augen, such that the pole to this plane, which plunges gently SW (statistically at 15°/218°), is the direction of maximum compression. S$_{2K}$ and the XY plane trend northwest and dip steeply northeast.

The long axes of the augens (X-direction) define a stretching lineation (L$_{2K}$) and the extension direction of the stress regime. The lineation plunges WNW and ENE at shallow (~10°) angles but appears to progressively steepen to as much as much as 65° towards the ESE. When plotted stereographically (Fig. 6), the lineations are distributed along a steep, NE-striking girdle. The scatter of L$_{2K}$ is consistent with rotation of the lineation within the S$_{2K}$ foliation plane about a subhorizontal NNE-trending axis. The deformational event responsible for the development of this LS-tectonic fabric is designated D$_{2K}$.

Microscopically, the quartzofeldspathic augen contain both largely undeformed, interlocking granoblastic-polygonal alkali feldspars and elongate recrystallized quartz ribbons that are indicative of mylonitization (Plate 16). Ribbon length to width ratios and orientations are similar to those of the augens that contain them, suggesting they were deformed by the same stress regime. If so, the development of both the augen and quartz
Plate 15. Quartzofeldspathic augen gneiss showing a large elongated augen, (blue lines show S_{1K} and L_{2K} orientations).

Figure 5. Flattened prolate ellipsoid.
ribbons reflects an episode of mylonitization that may have deformed an earlier compositional layering equivalent to $S_{1G}$ in the granite gneiss.

$D_{3K}$

The presence of a pervasive granoblastic texture and the absence of significant strain within the grains that define the augen and ribbon structures indicate an episode of post-$D_{2K}$ recrystallization (designated $D_{3K}$). The preservation of the quartz ribbon structures during $D_{3K}$ suggests that this event was static.

$D_{4K}$

The presence of undulose extinction and subgrains within the granoblastic quartz grains indicates a subsequent solid state, brittle deformation. This deformation is designated $D_{4K}$. 
Figure 6. Stereographic projection of K-feldspar Augen Gneiss foliation poles (solid squares), average foliation, (S2K great circle), and augen c-axis/x-axis (L2K circles).
Plate 16. Photomicrograph of K-feldspar Augen Gniess showing: (a) Quartz ribbon (D_{2K}), (b) 120 degree triple junction intersections characteristic of granoblastic texture that resulted from recrystallization during D_{3K}. Undulose extinction and subgrain, development in both figures record D_{4K}. Field of view is 4mm.
Amphibolite Gneiss

$D_{1A}$

The amphibolite gneiss is in contact with, and intruded by, the AMCG suite. The earliest recognizable structure within the amphibolite gneiss is a high-grade metamorphically segregated mesocratic/leucocratic banding, striking NW-SE and dipping NE. This banding is designated $S_{1A}$ (Plates 17 and 18). However, the amphibolite is extensively retrograded and cataclastically deformed making identification and measurement of early microstructures difficult.

$D_{2A}$

The leucocratic layers comprise discontinuous, highly sheared and cataclastically deformed feldspars and barely recognizable relict quartz ribbons that suggest $D_{1A}$ was either accompanied or followed by an episode of somewhat lower-grade mylonitization assigned to $D_{2A}$. However, the original mineralogy is almost completely replaced by low-grade phases and/or has been cataclastically deformed. Only the relict metamorphic layering is preserved more-or-less intact, often showing sharp boundaries and high contrast between leucocratic and mesocratic layers. The mesocratic layers show almost complete retrograde replacement that preserves only a relict compositional layering (Plate 17). $S_{2A}$ is axial planar to microscopic isoclinal intrafolial folds in $S_{1A}$ that are likely associated with $D_{2A}$ and are designated $F_{2A}$. $D_{2A}$ likely corresponds to $D_{2G}$ in the granite gneiss. $F_{2A}$ hinge zones are greatly thickened and often greatly drawn out into $S_{1A}$ at their apex. Thinning of the fold limbs is also significant with limbs being quickly incorporated into $S_{1A}$ to form rootless or near rootless folds. The shape of the profile is close to ideal.
Class 2 (similar) style fold geometry suggesting an important component of oblique shear (flow) in the fold mechanism. This, in turn, suggests that folding occurred at elevated temperatures during high-grade metamorphism.

\(D_{3A}\)

To be consistent with the evolution of the other components of the host suite, a granoblastic overprint would be expected to follow \(D_{2A}\) although no obvious relicts are preserved. This probably reflects the intensity of the subsequent retrograde metamorphism and cataclastic deformation, which likely destroyed any evidence of this earlier recrystallization.

\(D_4\)

Both the metamorphic layering (\(S_{1A}\)) and isoclinal microfolds (\(F_{2A}\)) are folded by \(F_{3A}\), the latest, most conspicuous folds within of the amphibolite gneiss. \(F_{3A}\) are open folds (approximately 25 cm in wavelength) with axes that plunge moderately to steeply SE and axial planes that trend NW-SE and dip steeply NE (see Figure 7 and Plate 18).

Folding is due to displacement along axial planar fractures (\(S_{4A}\)) that are highly oblique to the compositional layering. These axial planar fractures deflect and displace \(S_{1A}\) and \(S_{2A}\) such that the layers in the limbs ‘step up’ towards the fold hinge in a pattern that alternates between dextral and sinistral across axial planes. This indicates oblique shear folding in which passive rotation of a layer occurs under simple shear directed obliquely to the layer (Plate 18). In this style of folding, the line of movement is perpendicular to the fold axis and lies within the axial plane. When the fold data are
plotted stereographically, the orientation of this line plunges moderately northwest
(36/340) (Figure 8). The vergence inferred from the asymmetry of the folds is NE over
SW, or top-to-the SW. In combination with the line of movement, this indicates oblique
shear with dextral and reverse components (dextral thrusting).

Within the axial planar displacement planes are grain-size reduced feldspars,
some showing cataclastic deformation such as grain-size reduced, ‘birds eye’ feldspars
and extensive chloritic mineralization. Evidence of subordinate flexural slip during
folding is indicated by cataclastic deformation between the compositional layers. This
includes the development of subgrains around quartz ribbons often to the point where the
contrast between the ribbon edge and the finer-grained matrix is lost, as well as
cataclastically deformed feldspar grains.

Cut samples clearly show that the low-grade chloritic mineralization that can be
seen macroscopically and microscopically replacing mesocratic layers also occurs in the
axial planar displacement fractures (S4A). Chloritic seams both cross-cut and are
truncated by the displacement planes, suggesting that mineral growth was syn/post-
tectonic with respect to D4A.

Finally, all features are cut by late calcite veins and brittle fractures.
Plate 17. Photomicrographs of Amphibolite showing S1A-2, F2A(?), (highlighted in green) comprising of retrograded quartzofeldspathic remanents and S4A (highlighted in blue) displacement planes containing significant cataclastic deformation. Fields of view 4mm.
Plate 18. Amphibolite sample showing NE over SW vergence, $S_{1-2A}$ (green arrow), $S_{4A}$ (blue lines).

Figure 7. Oblique shear model showing how folds are formed by displacement oblique to layering.
Figure 8. Amphibolite: S_{1-2A} (blue great circle), S_{4A} (green great circle), F_{4A} (cross), orientation of transport (black dot).
Host Rock Deformational History

$D_{1HOST}$

The earliest tectonothermal event ($D_1/M_{1HOST}$) recognized in the host site is the migmatitic flow banding ($S_{1G}$) in the metagranite. Such migmatization would have necessarily accompanied high-grade metamorphism and is considered to have been coeval with the development of the metamorphic compositional banding observed in the K-feldspar augen gneiss ($S_{1K}$) and the amphibolite gneiss ($S_{1A}$). Hence these fabrics collectively constitute $S_1$ (Table1).

$D_{2HOST}$

The second tectonothermal event ($D_2/M_{2HOST}$) in the host suite is a mylonitization event that produced and then folded quartz ribbons in the metagranite ($S_{2G}$, $F_{2G}$) and the amphibolite gneiss ($S_{2A}$, $F_{2A}$). In the K-feldspar augen gneiss the event produced a stretched quartz lineation ($L_{2K}$) that was then rotated into the orientation of transport. Mylonitization that produced ductile quartz ribbons in association with brittle K-feldspar porphyroclasts suggests temperature conditions between the brittle-ductile transition in quartz (ca. 300°C) and that in K-feldspar (ca. 550°C). This, in turn, suggests that mylonitization occurred at a significantly lower metamorphic grade than that associated with $D_1$ (migmatization). It is therefore designated a separate event labeled $D_2/M_2$. Hence, the mylonitic fabrics in all three host rock units collectively constitute $S_2$. 
$D_3/M_{\text{HOST}}$

A static granulite facies metamorphic event ($M_{\text{HOST}}$) overprints all previous microtextural features erasing many of them and replacing them with a granoblastic texture. This recrystallization event and its accompanying metamorphism is designated $D_3/M_{\text{HOST}}$.

$D_{4\text{HOST}}$

Common to all host units are undulose quartz grains, subgrains, microfracturing, coeval retrograde mineralization, and major faulting, all of which indicate a later, relatively brittle, low-grade deformational event. However, only the amphibolite shows associated brittle oblique shear/passive folding ($F_{4\text{A}}$), the orientation of which is consistent with a dextral, SW-verging thrust regime.

The majority of fault displacements, slickensides and foliation/banding deflections indicate a dextral shear sense (Table 1).
Table 1. Correlation chart for deformation in the Host units.

<table>
<thead>
<tr>
<th>HOST DEFORMATIONAL HISTORY (thus far)</th>
<th>Metagranite</th>
<th>K-Feldspar Augen Gneiss</th>
<th>Meta-amphibolite</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>D1/M1</strong></td>
<td>S₁g migmatization</td>
<td>S₁Kaug - metamorphic layering</td>
<td>S₁Amp Metamorphic layering</td>
</tr>
<tr>
<td><strong>D2/M2</strong></td>
<td>S₂g,F₂g - Mylonitization-quartz ribbons</td>
<td>S₂Kaug Mylonitization-quartz ribbons, (‘flattened pencil’/elongated augen)</td>
<td>S₂Kaug Mylonitization of quartz ribbons.</td>
</tr>
<tr>
<td><strong>D3/M3</strong></td>
<td>Granulite metamorphism, granoblastic recrystallization</td>
<td>Granulite metamorphism, granoblastic recrystallization (erased by retrograde metamorphism)</td>
<td>Granulite metamorphism, granoblastic recrystallization’</td>
</tr>
<tr>
<td><strong>D4/M4</strong></td>
<td>Retrograde mineralization</td>
<td>Retrograde mineralization</td>
<td>F₄Amp - Oblique shear folding, retrograde mineralization -dextral thrusting, SW vergence</td>
</tr>
</tbody>
</table>
AMCG Suite

The anorthosite-gabbro and charnockitic orthogneiss of the AMCG suite are intrusive into the host rock. This leucocratic suite is widely exposed and provides well-defined deformational structures as a result of its sharp, continuous, rhythmic layering and macroscopic deformational features. Both the charnockite and anorthosite-gabbro show the same structural sequence, and are so considered together in the following description. At least four phases of penetrative deformation are recorded in the two orthogneisses and are designated $D_{1AMCG}$-$D_{4AMCG}$ (Table 2).

$D_{1AMCG}$

The earliest recognizable phase of penetrative deformation in the AMCG orthogneiss is a gneissic banding ($S_{1AMCG}$) that is interpreted to have formed by metamorphic differentiation. The banding comprises continuous leucocratic and mesocratic layers, rhythmically alternating on a cm-dm scale. Plotted stereographically, poles to this gneissic banding fall on a great circle, the pole to which plunges gently SE (Fig 9).

$S_{1AMCG}$ is folded about tight to isoclinal intrafolial folds with an axial planar fabric. This fabric is parallel to $S_{1AMCG}$ and defined by the same high-grade minerals, suggesting that the development and folding of the metamorphic banding occurred during a single progressive developmental event. The axial planar fabric is therefore designated $S_{1bAMCG}$. $S_{1bAMCG}$ is defined by oriented minerals within the mesocratic layers, but is accentuated in hinge zones by parasitic folds in the leucocratic banding (Plate 19). The folds ($F_{1bAMCG}$) show Class 1C to 2 fold profile geometries with extensive layer
thickening in the hinges. Only a few $F_{1bAMCG}$ folds were observed and none could be sampled. But all AMCG thin sections show granoblastic recrystallization that overprints these earlier textures.

\[D_{2AMCG}\]

$S_{1AMCG}$, $S_{1bAMCG}$ and $F_{1bAMCG}$ are folded about tight to isoclinal, occasionally rootless, asymmetric folds ($F_{2AMCG}$). $F_{2AMCG}$ fold profiles vary from Class 1C to 3 geometry with significant thickening at fold hinges (Plates 20 and 21). $F_{2AMCG}$ folds lack an axial planar foliation, as a result of which the layer definition of the leucocratic banding is generally much sharper than that in $F_{1bAMCG}$.

Plotted stereographically, $F_{2AMCG}$ axes are distributed about a great circle, the pole to which plunges gently SW (Figure 10). An associated mafic mineral lineation ($L_{2AMCG}$), defined by pyroxenes, plunges moderate-steeply ESE. The distribution of fold asymmetries (i.e., S and Z folds) is consistent with sheath folding, which is thought to be a result of layer-parallel shear (Cobbold and Quinquis, 1980; Price et al., 1990). From exposed sheath fold patterns, the ‘noses’ of which can be used to indicate the direction of transport, $D_{2AMCG}$ was the result of sinistral thrusting towards the WNW, parallel to the $L_{2s}$ mineral lineation (Figure 11).
Figure 9. Stereographic projection of AMCG suite foliation poles (black squares) showing predominant NE dip of metamorphic banding (great circle).
Plate 19. $S_{1bAMCG}$ in AMCG suite (orientation shown by blue line) and, $F_{1bAMCG}$ (outlined in yellow).
Plate 20. (a) $S_{1AMCG}$ and $F_{1AMCG}$ folded by $F_{2AMCG}$ and (b) $F_{2AMCG}$ and shear z-fold in AMCG suite.
Plate 21. (a) F$_{2AMCG}$ sheath fold pattern looking down the ‘nose’ (highlighted by the black ellipse, and (b) large F$_{2AMCG}$ fold, in AMCG suite.
Figure 10. Stereographic projection of $F_{2AMCG}$ fold axis (+) and $L_{2AMCG}$ mineral lineations(♦) in the AMCG suite. Fold asymmetries shown by s and z symbols indicating symmetry and curved arrows show the associated direction of rotation.
Figure 11. Sheath fold formation: Sheath folds form during progressive shear wherein folds form originally at highly oblique angles to the direction of shear but become rotated into the direction of transport effectively 'stretching out' the fold into a conical 'nose' shape while a coeval mineral lineation forms parallel to the transportation direction. Also shown here is the progression between the geometry and the stereoplot pattern where 1 shows the lowest amount of shear and 3 the greatest. The shear plane is the shear direction; the open circle the shear direction; small dots fold axis.
$D_{3AMCG}$

Microscopically, all AMCG suite fabrics have been overprinted by a granoblastic texture indicative of recrystallization during a high-grade thermal event. As a result of this granoblastic texture, no earlier microdeformational features are visible, although macrostructures are well preserved. This recrystallization event is designated $D_{3AMCG}$ (Plate 22).

$D_{4AMCG}$

A late phase of brittle deformation ($D_{4AMCG}$) cross-cuts all previous structures. This deformation takes the form of low-grade shearing and faulting (Table 2).

Table 2. Deformation sequence in AMCG Suite

<table>
<thead>
<tr>
<th>AMCG (D$<em>{1a}$-D$</em>{4o}$)</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>D$_{1AMCG}$</strong></td>
<td>S$_{1AMCG}$ - Gneissic banding</td>
</tr>
<tr>
<td></td>
<td>F$<em>{1AMCG}$, S$</em>{2A0}$ (axial planar foliation)</td>
</tr>
<tr>
<td><strong>D$_{2AMCG}$</strong></td>
<td>F$_{2bAMCG}$, Sheath formation, s-z folds (class 1C-3??), curvilinear Fold Axis, steep ESE mineral lineation Sinistral thrusting</td>
</tr>
<tr>
<td><strong>M$_{3AMCG}$</strong></td>
<td>Granulite Metamorphism- granoblastic texture</td>
</tr>
<tr>
<td><strong>D$_{3AMCG}$</strong></td>
<td>Low grade deformation, faulting shearing</td>
</tr>
</tbody>
</table>
Plate 22. AMCG suite (a) Fold picture of F{}_{2AMCG} and, (b) Photomicrograph of same sample showing granoblastic overprint.
Garnet Amphibolite Dikes

The garnet amphibolite dikes cut the foliation ($S_{1AMCG}$) of the AMCG suite. However, at one locality a garnet amphibolite dike is deformed mesoscopically about a fold with a sub-vertical axial plane striking NW-SE (~100 degrees), and a shallow SE-plunging axis (~19 SE) consistent with NE-vergent, sinistral thrusting. Both the orientation and asymmetry of the fold are similar to mesoscopic $F_{2AMCG}$ folds in the AMCG suite and are therefore thought to be of the same generation (Plate 23).

Porphyritic Amphibolite Dikes

The porphyritic amphibolite dikes display no microdeformational strain features but, instead, preserve original ophitic igneous textures with in-situ greenschist retrograde mineralization indicating that their intrusion was shallow and post-dated all major ductile deformation. Evidence of chilled contacts also indicates that both the Host and AMCG suites were at low temperatures when the dikes were intruded. However, some brittle movement has occurred along dike contacts, which are also locally offset by minor brittle faults, the majority showing dextral sense.
Plate 23. Folded garnet amphibolite dike (line of dike highlighted in yellow). Sledgehammer in foreground for scale.
Leucogranite

The leucogranite is located between the Novillo Gneiss and the Granjeno Schist. Adjacent to the leucogranite in outcrop, the Novillo gneiss is tectonically disturbed, showing foliation/layering deflection and brittle cataclastic shearing clearly associated with dextral movement along the contact.

A very weak foliation occurs in the leucogranite defined microscopically by the orientation of parallel, interstitial muscovite and cataclastic microshears in quartz clusters/aggregates. This fabric dips NE at moderate to steep angles, broadly parallel to the contact between the Novillo Gneiss and Granjeno Schist.
PETROLOGIC AND STRUCTURAL CORRELATION BETWEEN HOST ROCKS, AMCG SUITE, DIKES AND LEUCOGRANITE

Prior to granulite facies metamorphism, which is common to both the host rocks and the AMCG suite, two phases of deformation are recorded in each of these assemblages. In the host rocks, a migmatitic fabric (D\textsubscript{1\text{Host}}) was subsequently mylonitized and microfolded (D\textsubscript{2\text{Host}}). The AMCG suite, on the other hand, shows no evidence of the quartz ribbons used to identify the mylonitic event in the host rocks and, instead, shows a gneissic banding and associated folding (D\textsubscript{1\text{AMCG}}) overprinted by an episode of ductile shear (D\textsubscript{2\text{AMCG}}) that produced sheath folds and a mineral stretching lineation. Given that the emplacement of the AMCG suite (at ca. 1Ga) post-dates that of the host rocks (at ca. 1.2 Ga), the simplest explanation of these structural relationships is that D\textsubscript{2\text{Host}} corresponds to both D\textsubscript{1\text{AMCG}} and D\textsubscript{2\text{AMCG}}. If so, D\textsubscript{1\text{AMCG}} with D\textsubscript{2\text{AMCG}} were likely the product of a single progressive deformation event, rather than two separate events. Furthermore, given the absence of evidence of mylonitization in the AMCG suite, emplacement of this suite likely coincided with the mylonitization of the host rocks. Therefore, the following tectonic history is proposed for the Novillo Gneiss (Figure 14).

D\textsubscript{1}/M\textsubscript{1}

The earliest tectonothermal event (D\textsubscript{1\text{Host}}) was an episode of migmatization that affected only the host rocks. During this event, high temperatures produced migmatitic flow banding in the granite gneiss and metamorphic differentiation banding in the
remainder of the host. Although the metamorphic mineral assemblages that must have coincided with this migmatization have been subsequently overprinted by later tectonothermal events, a significant metamorphic event must have accompanied migmatization and is here designated M₁.

\[ D₂/M₂ \]

The second tectonothermal event \((D₂/M₂)\) was likely coeval with the intrusion of the AMCG suite and produced mylonitization \((S₂\text{Host})\), micro-folding \((F₂\text{Host})\), and a stretching lineation \((L₂\text{Host})\) in the host suite. Once again, however, subsequent metamorphism has overprinted any metamorphic mineral assemblage that may have accompanied this event.

It is proposed here the AMCG suite was at too high a temperature during \(D₂/M₂\) to record a mylonitic fabric and, instead, developed metamorphic banding \((S₂\text{AMCG})\) and intrafolial folds \((F₁\text{AMCG})\) with an axial planar foliation \((S₁\text{AMCG})\). Intrusion of the AMCG suite into the host rocks while the latter were undergoing ductile shear would also account for the rotation of the host lineation \((L₂\text{Host})\) and the development of sheath folds \((F₂\text{AMCG})\) with an associated mineral lineation \((L₂\text{AMCG})\) in the AMCG suite.

Correlative structural orientations within the host rocks and AMCG suite are consistent with an environment of near layer-parallel, progressive shear common to shear zones in high-grade metamorphic rocks, (Cobbold and Quinquis, 1980; Price et al., 1990). Metamorphic banding in both assemblages consistently dips northeast throughout the field area, and the foliations in both assemblages are parallel to the banding. Although the lack of host fold axis data prevents direct comparison of fold orientations, matching fold geometries and banding-parallel axial planes suggest a correlation. Similarly,
lineations in both assemblages occur within S1-2, although those in the host suite vary in orientation from a shallow to moderate ESE-plunging, whereas the lineation in the AMCG suite consistently plunges moderately ESE.

Correlation of structures in the garnet amphibolite dikes with those of the host rock and AMCG suite is well constrained by direct structural-metamorphic relationships. The dikes were emplaced prior to the granulite facies metamorphism recorded in their high-grade mineral assemblage and granoblastic microtexture. Emplacement occurred after the intrusion of the AMCG suite, which they cross-cut, although they also record some of the same deformation.

Similarities in geometry, symmetry and orientation suggest a correlation between the folds in the dikes and those of F2AMCG in the AMCG suite. If so, dike emplacement must have occurred prior to this phase of deformation in the AMCG suite. In addition, field observation of highly deformed AMCG xenoliths within, and truncated by, the foliated but significantly less deformed garnet amphibolite dikes show that the dikes have experienced a significantly lower degree of strain. This is consistent with emplacement of the dikes during the later stages of D2.

Mineral assemblages and microtextural evidence in the host rock, the AMCG suite and the garnet amphibolite dikes are indicative of a granulite facies tectonothermal event (D3/M3). Recrystallization of all pre-existing microfabrics to granoblastic textures is considered to coincide with peak metamorphic conditions and the formation of orthopyroxene(?), clinopyroxene and garnet. The time of peak metamorphism has been dated at 990 ± 5 Ma on the basis of metamorphic zircon growth (Cameron et al., 2004).
Following granulite facies metamorphism, all rock units experienced dextral shear under low-grade (greenschist facies) conditions (\(D_4/M_4\)). Gefüllte textures were developed in plagioclase, and retrograde chlorite partially to entirely replaced portions of the mafic banding in all units, most notably within the brittle displacements that produced \(F_4\) folds in the meta-amphibolite, which show dextral thrust geometry. Chloritic mineralization is also associated with foliation deflection, slickensides and displacements that are most pronounced at contacts, the majority of which are dextral.

This episode of retrogression shown in the 546 ± 5 Ma porphyritic dikes (Keppie et al., 2006) must post-date their intrusion and is likely dated by the 313 ± 7 Ma age of the foliation in the leucogranite, which is thought to record juxtapositioning of the Novillo Gneiss and the Granjeno Schist (Dowe et al., 2004) (Table 3).
Table 3. Correlation chart for deformational and metamorphic events in the Host, AMCG Suite, intrusives and Leucogranite.

<table>
<thead>
<tr>
<th>Event Type</th>
<th>HOST</th>
<th>AMCG</th>
<th>Garnet Dikes</th>
<th>Porphyritic Dikes</th>
<th>Leucogranite</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>Migmatization/ Banding- S1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>D2</td>
<td><strong>Mylonitic event</strong>&lt;br&gt;S\textsubscript{2} quartz ribbons&lt;br&gt;F\textsubscript{2} microfolds&lt;br&gt;L\textsubscript{2} lineation, c-axis orientation varies</td>
<td><strong>Intrusion-high grade/ progressive shear</strong>&lt;br&gt;S\textsubscript{2a} Gneissic banding&lt;br&gt;F\textsubscript{2a} Folds w axial planar foliation&lt;br&gt;S\textsubscript{2a} Axial planar foliation high grade&lt;br&gt;S\textsubscript{2b} Sheath Folds (S-Z folds)&lt;br&gt;F\textsubscript{2b} Mineral Lineation&lt;br&gt;Sheath pattern&lt;br&gt;- Sinistral thrusting</td>
<td></td>
<td>-----------&lt;br&gt;Intrusion -late F2</td>
<td></td>
</tr>
<tr>
<td>M3 990Ma</td>
<td>Granulite facies metamorphism granoblastic texture</td>
<td>Granulite facies metamorphism and granoblastic texture</td>
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<td></td>
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<tr>
<td>546Ma</td>
<td></td>
<td></td>
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<td>Intrusion- 546±5Ma Igneous Microtexture</td>
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<tr>
<td>D4</td>
<td>Low grade deform, dextral shear&lt;br&gt;F\textsubscript{4} oblique Shear folding</td>
<td></td>
<td>Retrograde mineralization</td>
<td>Retrograde mineralization</td>
<td>Tectonic emplacement 313Ma, muscovite foliation</td>
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TECTONIC HISTORY OF THE NOVILLO GNEISS

The tectonic history of the Novillo Gneiss can be divided into eight stages, each marking important events in its evolution. Theses stages are summarized as follows:

1) Age data and geochemical signatures show the host rocks to have been emplaced between ca. 1235 and 1115 Ma (Cameron et al., 2004) in an arc or back-arc setting (Keppie 2001, 2003, 2006; Solari et al., 2003) proximal to an active continental margin.

2) Emplacement of the host was accompanied by migmatization, which is common in areas of high-grade metamorphism; stromatic migmatites require temperatures of at least ~700-720°C (Winkler, 1979). With a 30°C/km geothermal gradient typical of active continental margins, such migmatization required depths of at least 25-30 km.

3) Intrusion of the AMCG suite occurred between 1035 and 1010 Ma (Cameron et al., 2004), and is thought to have coincided with mylonitization of the host rocks. Progressive rotation of the host lineation from subhorizontal orientations to the moderately ESE-plunging trend characteristic of the lineation in the AMCG suite towards its contact with the host rocks suggests that the AMCG suite was emplaced into a zone of transpressional sigmoidal shear (Fig 12). Such zones can accommodate a wide variance in the orientation of an associated mylonitic lineation. Intrusions can be transferred upwards through the crust via mid-crustal shear zones (e.g., D'Lemos et al., 1992). Continuous displacement
along a ductile sigmoidal shear zone will produce extensional jogs providing space for the underlying accent of magma. As shearing progresses, such jogs eventually become zones of compression, effectively ‘pinching off’ the ascending magma from its source.

The ‘pinching up’ movement of the intrusion would, in turn, accompany the development of a positive flower structure in which the mineral lineation would vary from strike-slip at the margins to dip-slip in the interior (Fig 13) (e.g., Berthe et al., 1979). Such lineation patterns have been documented in zones of transpression (e.g., Schreckengost et al., 1996) and would account for the lineation geometry in the Novillo Gneiss as well as the contrasting patterns of deformation in the host rock and AMCG suite prior to granulite facies metamorphism.

4) Following emplacement of the AMCG suite, but prior to the end of the accompanying deformation, mafic dikes (now garnet amphibolites) were emplaced into both the host rocks and the AMCG between 1010-990 Ma.

5) Reburial must have occurred after emplacement of the garnet amphibolite dikes, but prior to ca. 990 Ma in order to produce the granulite facies metamorphism that completely overprinted the dikes and both units of the Novillo Gneiss to produce the granoblastic textures and high-grade metamorphic mineralogies seen in these rocks. Preservation of preexisting mylonitic structures (ribbon quartz) implies static recrystallization during metamorphism. The P-T conditions of this event have been estimated to be 8.9-9.7 kbar and 730-775°C (Orozco, 1991). The tectonic setting that
Figure 12. Mechanism for upward transport of intrusion in a sigmoidal, strike-slip regime. Movement of intrusion (black arrows) into a zone of transtension and away from a zone of transpression.
(D’Lemos et al., 1992)
Figure 13. Positive flower structure in a transpressional regime where strike-slip dominates the margins (large arrows) while dip-slip dominates the interior.
produced the required ~30 km of burial is likely to have been that of the roots of an orogenic belt during subduction or collision (Jamieson et al., 1998).

6) Rapid exhumation must have followed granulite facies metamorphism in order to preserve the granoblastic textures and some of the peak metamorphic mineralogy. Keppie et al. (2006) showed that the Novillo Gneiss cooled from ~660°C (closure temperature for U-Pb in titanite; Frost et al., 2000) at 928 ± 2 Ma to 325 ± 25°C (closure temperature for 40Ar/39Ar in biotite; Harrison et al., 1985) by 697 ± 7 Ma. This corresponds to a cooling rate of 1.45°C/my (Figure 14).

7) Following exhumation, a suite of porphyritic dikes was intruded at ca. 550 Ma (Keppie et al., 2006). Preserved igneous textures and chilled margins indicate that these dikes were emplaced into a relatively brittle, cold host rock. Extrapolation of the exhumation rate suggests that the Novillo Gneiss would have reached the surface (30°C) by ~497 Ma (50 my after dike emplacement) (Keppie et al., 2006) (Fig. 14).

8) The final significant tectonic adjustment was the tectonic juxtapositioning of the Novillo Gneiss against the adjacent Granjeno Schist. The Granjeno Schist has been interpreted to be a Paleozoic oceanic accretionary wedge (Ortega-Gutiérrez, 1993) marking ocean closure. Juxtaposition is recorded in the Novillo Gneiss by greenschist facies retrogradation and in the leucogranite emplaced into the contact zone by an 40Ar/39Ar age of 313 Ma on muscovite, recording cooling through 350°C (Dowe et al., 2005).
Figure 14. Cooling curve for the Novillo Gneiss plotted on a temperature vs. age graph (Keppie et al., 2006).
Similarities between the Mesoproterozoic (ca. 1 Ga) basement exposures in Mexico were instrumental in the proposal by Ortega-Gutiérrez et al. (1995) for the existence of the Oaxaquia microcontinent as the basement ‘backbone of Mexico’. The largest of these exposures, the Oaxacan Complex, is generally regarded as the Oaxaquia archetype (e.g., Keppie, 2004) and is used here to test the level of comparison with the Novillo Gneiss.

The Novillo Gneiss and Oaxacan Complex share many similarities. Both comprise similar-aged host units (Novillo Gneiss: 1235-1115 Ma, Cameron et al., 2004; Oaxacan Complex: ≥1350–≥1140 Ma, Solari et al., 2003) with arc/back-arc geochemical signatures. The Oaxacan Complex also recorded a migmatitic event dated at 1106 ± 6Ma (U-Pb zircon; Solari et al., 2003), which had no correlative in Mexico until this study. The migmatitic event in the Novillo Gneiss is similar in characteristics (stromatic habit) and tectonic sequence, but has not been directly dated. However it is interpreted to have accompanied emplacement of the host during the interval 1235-1115 Ma.

The host rocks of both the Novillo Gneiss and Oaxaca Complex are also intruded by AMCG suites of similar ages (Novillo Gneiss: 1035-1010Ma, Cameron et al., 2004; Oaxacan Complex: 1012 ± 12 Ma, Solari et al., 2004). In addition, both were reburied and underwent granulite facies metamorphism at similar times (Novillo Gneiss: 990 Ma; Oaxacan Complex: 1004-978 ± 3 Ma), following which they were rapidly exhumed and underwent low-grade deformation. Finally both complexes are tectonically juxtaposed against similar sequences (Granjeno Schist and Acatlán Complex, respectively)
comprising metasedimentary and metavolcanic rocks interpreted as Paleozoic continental slope or accretionary prism deposits. Finally, both complexes are overlain by undeformed Lower Paleozoic strata of Gondwanan affinity.

As a result of this study, similarities in the deformational histories of the Novillo Gneiss and Oaxacan Complex are also evident. For example, in both complexes the alternating leucocratic-mesocratic metamorphic banding strikes NW-SE and records a similar deformational sequence including a banding-parallel foliation that consists of high-grade minerals axial planar to intrafolial folds, which are, in turn, refolded by locally rootless isoclinal folds with variable symmetry and no axial planar fabric.

Most fold axes in the Oaxacan Complex plunge to the NW, whereas those in the Novillo Gneiss vary in orientation as a result of progressive rotation from their original attitude towards a common SE plunge.

Both complexes also contain quartz ribbons but in the Oaxacan Complex they occur in all units and their plunges cluster toward the NW, whereas in the Novillo Gneiss they occur only in the host rock and, like the fold axes, have been rotated towards a common SE-plunging orientation. High-grade mineral lineation orientations in both complexes are parallel to the quartz stretching lineations and delineate the line of shear – NW in the Oaxacan Complex and SE in the Novillo Gneiss. Further analysis in the Novillo Gneiss has identified the sense of shear – sinistral thrusting NE over SW.

The most notable difference in orientational data between the Novillo Gneiss and Oaxacan Complex is, therefore, the variable orientations of the quartz stretching lineations and fold axes in the former compared to the uniform NW-plunge of these features in the latter. This likely reflects the degree to which these features have been
reoriented by simple shear. If so, the Oaxacan Complex simply records a more progressed state of rotation of linear features into the shear direction during the mylonitization that accompanied the intrusion of the AMCG suite. This would make the recognition of sheath folding more difficult in the Oaxacan Complex, precluding the analysis used in the Novillo Gneiss to determine the sense of shear (Figure 15).

It is also interesting to note that in both the Oaxacan Complex and Novillo Gneiss, the relative orientation of the shear direction and the best-fit plane (profile plane) to the foliation poles are similar. Consequently, if the moderately SE-dipping profile plane in the Oaxacan Complex is rotated into an orientation similar to the moderately to steeply NW-dipping profile plane in the Novillo Gneiss about the pole to their similarly oriented (steeply NE-dipping) axial planes, the shear direction in the Oaxacan Complex changes from a shallow W plunge to a SE plunge, similar to that in the Novillo Gneiss. Thus, differences in orientational data between the Novillo Gneiss and Oaxacan Complex may simply reflect local differences in the general orientation of the same tectonic package (Figure 16).

Like the Novillo Gneiss, the Oaxacan Complex also contains mafic dikes that predate granulite facies metamorphism. However, it contains no post-granulite facies dikes equivalent to the porphyritic amphibolites in the Novillo Gneiss. Hence, the Novillo Gneiss records an extensional event that is not observed in the Oaxacan Complex. Keppie et al. (2006) suggested that the porphyritic amphibolites were plume related, and linked to the Ediacaran-Cambrian separation of Avalonia from Gondwana. If so, the absence of equivalent dikes in the Oaxacan Complex suggests that this complex occupied a more interior position with respect to the developing continental margin.
Early/least shear Novillo Gneiss; sheath fold pattern

Oaxacan Complex; all features rotated into shear orientation

Figure 15. Change stereonet pattern as shear progresses with 1 showing early stages of shear, 2 being analogous to the Novillo Gneiss with varying fold axis and clustered lineations and 3 the Oaxacan complex where all features have been rotated into the shear direction.
Figure 16. Stereographic projection showing reorientation of the Oaxacan complex profile plane, along the axial plane ~40-50 degrees causes the Oaxacan complex line of shear to parallel the Novillo Gneiss.
CONCLUSIONS

Correlation between the Novillo Gneiss and Oaxacan Complex based on similarities in the host and intrusive lithologies, their age of emplacement and cover rocks has been used in previous studies to support the existence of the Oaxaquia microcontinent (e.g., Ortega-Gutiérrez et al., 1995; Keppie, 2004).

This study supports such a correlation and has identified a previously unknown early migmatitic event in the Novillo Gneiss that is probably correlative to the migmatitic tectonothermal event in the Oaxacan Complex. The second major tectonothermal event in both complexes, involving granulite facies metamorphism, also correlates on the basis of available age data (Solari et al., 2003; Cameron et al., 2004).

In addition, this study has provided a detailed structural-kinematic history of the major events in the Novillo Gneiss and has showed that the style and sequence of deformational features are very similar to those of the Oaxacan Complex and developed in very similar tectonic settings. These data further support a correlation between the two complexes.

Structural studies of the Huiznopala Gneiss, another exposure of Mesoproterozoic basement in Mexico, have only been of a reconnaissance nature, but nonetheless describe an overall NW-trending, E-dipping foliation, the poles to which define a great circle striking NW and dipping E, and rare NW-plunging fold hinges and lineations (Lawlor et al., 1999). All of these orientations are consistent with those in the Novillo Gneiss and even more so with those of the Oaxacan Complex. This further strengthens the structural/deformational evidence for the existence of Oaxaquia and highlights the need
for more detailed structural-deformational characterization of the Huiznopala Gneiss
and other ca. 1-1.3 Ga basement exposures in Mexico.
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