ABSTRACT


The Gold Hill fault zone is a first-order structure within the peri-Gondwanan Carolina zone that separates Neoproterozoic metavolcanic rocks of the Charlotte and Carolina terranes. Previous work suggests that a north-south trending ductile shear zone, the Davis Branch shear zone, merges with the Gold Hill fault zone (s.s.) in the vicinity of Waxhaw, North Carolina. This region of the Carolina zone was chosen to investigate the relationship between the Davis Branch shear zone and the Gold Hill fault zone, and to delineate the structural relationship between the Gold Hill fault zone and surrounding volcaniclastic terranes. The results of this study may shed new light on the tectonic history of the southern Appalachian Carolina zone.

Geologic mapping identified low-grade volcanic sequences on each side of the Gold Hill fault zone. Northwest of the Gold Hill fault zone lies the Marvin sequence, which is composed of dacitic quartz tuffs, porphyries, and minor amounts of mafic-derived sedimentary rock and schist. Southeast of the Gold Hill fault zone lies the Twelve Mile sequence, which consists of a suite of rhyolitic volcaniclastic lapilli and lithic tuffs and an overlying sedimentary package of fine-grained siliciclastic sediments and minor conglomerate. Deformed rocks in the Twelve Mile sequence make up the bulk of the Gold Hill fault zone. Intruding the Twelve Mile sequence is granitic Waxhaw pluton that has thermally metamorphosed portions of the volcanic sequence.
Both the Marvin and Twelve Mile sequences exhibit geologic features that are similar to the other volcaniclastic sequences in the Carolina terrane. These observations question early interpretations that the Gold Hill fault zone represents a terrane boundary between the Carolina and Charlotte terranes. Moreover, detailed geologic mapping and strain analysis also indicate the Davis Branch shear zone, originally reported as a north-south trending ductile shear zone, is not present within the Waxhaw, NC study area.

Structural analysis of the Gold Hill fault zone suggests it is a steeply dipping zone of dextral wrench-dominated transpression defined by a wide footwall damage zone in the Twelve Mile sequence and narrow hanging wall damage zone in the Marvin sequence. Structures within the Gold Hill fault zone are attributed to deformation along the Gold Hill fault, which is a right lateral reverse fault with southeast vergence that thrusts the Marvin sequence over the Twelve Mile sequence. Strata in both volcanic sequences are folded into upright, northeast-plunging, folds imprinted by a steeply dipping axial planar cleavage that strikes sub-parallel to the trace of the Gold Hill fault zone. Metamorphic and microtextural data suggests that the Waxhaw granite, which intrudes the Twelve Mile sequence, is synkinematic to deformation within the fault zone, and support the interpretation that Gold Hill fault thrusts the Marvin sequence over the Twelve Mile sequence.
THE STRUCTURE AND KINEMATICS OF THE GOLD HILL FAULT ZONE
IN THE VICINITY OF WAXHAW, NORTH CAROLINA

by

JOHN STEFAN ALLEN

A thesis submitted to the Graduate Faculty of
North Carolina State University
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Master of Science

MARINE, EARTH, AND ATMOSPHERIC SCIENCES

Raleigh

2005

APPROVED BY:

[Signatures]

Chair of Advisory Committee
BIOGRAPHY

The author was born in Spartanburg, South Carolina on July 2nd, 1981 where he grew up and received his kindergarten through Grade 12 education. After eviction from the Sunnyvale trailer park by Park Supervisor Jim Lahey and the loss of his part-time job to then Weekend Assistant Supervisor Randy (a.k.a. Smokey), the author left Spartanburg for the sunny hamlet of Greenville, South Carolina to attend college at Furman University in the fall of 1999. In June of 2003, he was fortunate enough to receive a Bachelors of Science in Earth and Environmental Science from Furman University under the guide of the esteemed Dr’s William Ranson and John Garihan.

Two days later, the author began his graduate field work in the area of Waxhaw, North Carolina under the tutelage of Dr. James Hibbard at North Carolina State University in Raleigh. The author quickly developed a strong addiction to alcohol, blue-grass music, and womanizing, to the dismay of his advisor, as the result of the rapid transition to an overly stressful graduate life. However, Mr. Allen was able to overcome his addictions and after taking vows of sobriety and celibacy, he completed his graduate course work and research requirements in the fall of 2005 at North Carolina State University. He now lurks somewhere in the area of Lexington, Kentucky where he runs a moderately successful business remarking old shopping carts and barbeques.
ACKNOWLEDGEMENTS

I would like to convey my deepest gratitude to Dr. James Hibbard, chief advisor of my thesis committee and to Dr. Edward Stoddard and Dr. Ron Fodor for their participation on my advisory committee. I would especially like to thank Dr. James Hibbard for the many hours of consultation, constructive criticism, and for giving me the opportunity to be a part of this research project. I would like to convey my appreciation to my fellow graduate students for their camaraderie and discussion, and especially to Gordon Box for our many academic discussions concerning my research and the tireless effort he put forth in the editing of this thesis.

I would also like to convey my personal thanks to Dr. Irene Boland for introducing me to the study area as well as for the discussions and criticism she offered me during the course of my field work and research. Without the tireless effort and dedication she put into mapping the local geology around Rock Hill, Van Wyck, and Waxhaw, my research would not have been possible.

Finally, I would also like to extend my thanks to Doug Carter and his wife for introducing me to many of the local property owners around the study area and for the countless glasses of ice tea and conversations they offered me concerning the geology of Waxhaw, North Carolina.

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<td>Opq</td>
<td>undifferentiated opaque minerals</td>
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Introduction.

Focus of Study.

The Gold Hill shear zone (Butler and Fullagar, 1978), herein termed the Gold Hill fault zone (GHfz), is a first order structure in the southern Appalachian Carolina zone (Figure 1) that has traditionally been interpreted as the boundary between ductily deformed plutonic and volcanic rocks of the Charlotte terrane (Hibbard et al., 2002) and low-grade volcanic and epiclastic rocks of the Carolina terrane (Secor et al., 1983, Hibbard et al., 2002). Detailed work in central North Carolina has revealed the GHfz is a Late-Ordovician brittle-ductile high strain zone and that the dominant deformation within the fault zone in that region is sinistral transpression (Standard, 2003, Hibbard et al., 2003). However, structural mapping and geochronological studies completed in south-central North Carolina (Schroeder, 1987, Boland, 1996, Lavallee, 2003, Allen et al., 2005) and northern South Carolina (Lawrence and Corbett, 1999, Lawrence, 2001) indicate the GHfz is a dextral reverse-oblique fault zone that was active during the Middle Devonian, Carboniferous, and Early Permian.

The purpose of this investigation is to resolve conflicting kinematic reports along the GHfz and develop an accurate deformational and metamorphic history for a region of the Carolina zone that is poorly understood. The field area around Waxhaw, North Carolina was originally chosen because there the GHfz reportedly merges with a previously mapped ductile shear zone informally termed the Davis Branch shear zone (DBsz) (Boland, 1996, Boland pers. comm., 2003), which is interpreted as an early Paleozoic sinistral shear zone. The similarity in kinematics between the DBsz and the GHfz in central North Carolina led Hibbard (2000) to
hypothesize that the DBsz was a continuation of the Late Ordovician sinistral GHfz mapped in the vicinity of Mount Pleasant and High Rock Lake, North Carolina, whereas dextral motion mapped along the GHfz in southern North Carolina and northern South Carolina was suspected to be the result of middle and/or late Paleozoic reactivation of the fault zone in those regions.

Detailed structural mapping at the 1:24,000 scale was conducted in the immediate vicinity of Waxhaw, North Carolina (Figure 2) in order to establish the nature of the DBsz and its relation to the GHfz. However, detailed mapping was unable to establish the existence of the DBsz. Therefore, the primary focus of this investigation was shifted to the GHfz in an effort to provide some insight into its structural and kinematic history in that area. This investigation is part of a collaborative study, which also involves structural analysis of the GHfz in other regions of North Carolina, high resolution U-Pb and $^{40}$Ar/$^{39}$Ar geochronology of rocks deformed within the fault zone, and Sm-Nd isotopy of volcanic strata associated with terranes in the Carolina zone. The results of this collaborative investigation could have profound implications for the geologic history of the Carolina zone and its accretion to North America (Hibbard, 2000).

Regional Geology.

The GHfz lies within the Carolina zone, which comprises the eastern most exposed crustal block in the southern Appalachian orogen. The purpose of this section is to review up-to-date regional geology of the southern Appalachians to place the GHfz in a regional context. To this end, this section will outline the geology of
the two largest exotic terranes within the Carolina zone where the study area is located. The final part of this section will review recent work completed on the GHfz in central North Carolina and northern South Carolina.

The Carolina zone is only one of four first-order divisions within the southern Appalachians (Figure 1). The other three divisions include from west to east native Laurentian rocks, the Piedmont zone, and the Goochland zone. All four zones are distinguished on the basis of constraints in stratigraphy, tectonic history, and paleocontinental affinity. The Laurentian and Carolina zones have recognizable paleocontinental affinities. Native Laurentian rocks consist of heterogeneously deformed and metamorphosed sedimentary rocks and basement that are native to the Laurentian paleocontinent, while rocks in the Carolina zone consist of volcanioclastic terranes that are of Gondwanan affinity (Secor et al., 1983; Hibbard and Samson, 1995). The Piedmont and Goochland zones, however, are of unknown continental affinity (Hatcher and Williams, 1983, Hibbard and Samson, 1995). The Piedmont zone lies directly east of the Laurentian zone and is composed of intensely deformed and metamorphosed suspect terranes that are thought to have accreted to the Laurentian margin during the Early to Middle Ordovician (e.g. Hatcher, 2002). The Goochland zone is composed of 1.0 Ga gneisses, Devonian orthogneisses, and schists of uncertain age, which have experienced tectonism that predates late Paleozoic deformation. Interpretations regarding the origin of the Goochland zone vary from possible basement of exotic volcanic arcs in the Carolina zone (e.g. Hibbard et al., 2002) to a displaced fragment of native Laurentian crust (e.g. Bartholomew and Tollo, 2004).
Figure 1. Lithotectonic map of the southern Appalachian orogen highlighting the terrane distribution in the Carolina zone and the regional setting of the Gold Hill fault zone (red line). Box represents the approximate area for Figure 4. PMB-Pine Mountain Belt, SMA-Sauratown Mountains Anticline, CPsz-Central Piedmont shear zone. Modified after Hibbard et al. (2002) and Hibbard and Samson (1995).
The Carolina zone (Hibbard and Samson, 1995) is the largest accreted constituent of the southern Appalachians, spanning more than 600 km along strike from central Virginia to northeastern Georgia. It is bounded to the east by Mesozoic and Cenozoic onlap sequences of the Atlantic coastal plain and to the west by suspect terranes of the Piedmont zone. The boundary between the Piedmont and Carolina zones was originally interpreted as a cryptic suture termed the central Piedmont suture (Hatcher and Zietz, 1980). Later investigation revealed that the central Piedmont suture is instead a late-Paleozoic thrust fault that has either tectonically buried or obliterated the original suture between the Carolina and Piedmont zones and for this reason the boundary between these two zones has been renamed the central Piedmont shear zone (Hibbard et al., 1998, West, 1998, Wortman et al., 1998).

The Carolina zone is composed of an amalgamation of infrastructural and suprastructural volcanic terranes that are predominantly Neoproterozoic in age and can be distinguished on the basis of stratigraphy, metamorphic grade, faunal assemblages, structure, and isotopic evolution (e.g. Hibbard et al., 2002). All the terranes within the Carolina zone are considered exotic to Laurentia and are most likely of peri-Gondwanan affinity for the following reasons: 1) a diverse assemblage of Acado-Baltic trilobite fauna identified within the largest terrane in the Carolina zone (Secor et al., 1983, Samson et al., 1990), 2) terranes within the Carolina zone record a gross geologic evolution and tectonic history that predate the Iapetan cycle (Glover and Sinha, 1978, Harris and Glover, 1988, Dennis and Shervais, 1991, Hibbard and Samson, 1995, Dennis and Wright, 1997, Barker et al., 1998), 3) the
position of the Carolina zone within the Appalachian orogen is spatially equivalent to
known peri-Gondwanan terranes in the northern Appalachians (Hibbard et al., 2002).

The two terranes within the Carolina zone that are most important to this
study are the Charlotte terrane (Hibbard et al., 2002) and the Carolina terrane (Secor
et al., 1983). The Charlotte terrane is composed of medium- and high-grade
metamorphic gneiss and schist (Hibbard et al., 2002). Rocks constituting the
Charlotte terrane form the largest infrastructural terrane in the Carolina zone (Figure
1), extending north from central Georgia to Winston-Salem, North Carolina; and
possibly extending into central Virginia. Rocks of the Piedmont zone bound the
Charlotte terrane to the west along the central Piedmont shear zone while to the east
the Charlotte terrane is bordered by low-grade volcanic rocks of the Carolina terrane.
The boundary between the Charlotte and Carolina terranes was originally interpreted
as a metamorphic gradient (Secor et al., 1986, Dallmeyer et al., 1986); however,
recent studies indicate it is tectonic along much of its trace (Offield et al., 1995,
Boland, 1996, Barker et al., 1998, Dennis et al., 2000, Minninger and Nunnery,
2001).

King (1955) first described the Charlotte terrane as dominated by plutons
ranging in composition from granites to gabbros which intrude metaigneous and
metapelitic suites that comprise the country rock (Figure 3). These intrusive suites
range in age from Neoproterozoic to late Paleozoic (Butler and Fullagar, 1978,
from mafic to felsic volcanic suites interlayered with minor quartzite, schist, and
marble. Limited isotopic and geochronologic studies suggest that parts of the
Figure 3. Stratigraphic chart for the Charlotte terrane and volcanic sequences in the Carolina terrane, including isotopic data as well as fossil and U-Pb detrital and xenocrystic zircon ages. Though all four known sequences in the Carolina Terrane are shown, the Virgilina and Albemarle sequences of the Carolina terrane are the only sequences considered in this investigation. Modified after Hibbard et al. (2002).
Charlotte terrane country rock where derived from juvenile crust (Fullagar et al., 1997) and some of these metavolcanics are ca. 570 Ma in age (Dennis and Wright, 1997). Geochemical studies also indicate that large Neoproterozoic metamafic complexes that comprise the country rock of the Charlotte terrane formed in a supra-subduction zone setting related to intra-arc rifting (McSween et al., 1984, Dennis and Shervais, 1991, 1996). Local occurrences of retrograde eclogites have also been recognized along the eastern flank of the Charlotte terrane in the area of Newberry, South Carolina (Shervais et al., 1997, Dennis et al., 2000, Shervais et al., 2003) indicating the Charlotte terrane has experienced an intense metamorphic history.

Structures in the Charlotte terrane are difficult to categorize due to the abundance of post-deformational plutons. Foliations observed throughout the terrane are steep-dipping with northeast strikes. Folds are tight to isoclinal with near vertical axial surfaces and plunge gently to the northeast and southwest (e.g. Butler and Secor, 1991). Many of the plutons within the Charlotte terrane are structurally massive towards their interiors and foliated around their margins (Butler and Secor, 1991). 40Ar/39Ar ages obtained from amphiboles in the deformed country rock suggest that ductile deformation within the Charlotte terrane is Silurian or older (Sutter et al., 1983). Sillimanite has been observed in pelitic and semi-pelitic schists implying amphibolite facies regional metamorphism in the Charlotte terrane (Butler, 1983); however, kyanite and andalusite have also been reported in the Charlotte terrane (Goldsmith et al., 1988).

The suprastructural Carolina terrane extends approximately 500 km along strike from eastern Georgia to central Virginia (Secor et al., 1983) and has a
maximum outcrop width of 140 km from its western boundary with the Charlotte terrane to its eastern boundary with the Atlantic coastal plain making it the largest accreted volcanic terrane in the Carolina zone (Hibbard et al., 2002) (Figure 1). Strata in the Carolina terrane are composed of metavolcanic and epiclastic rocks that are intruded by plutonic and hypabyssal suites that range from felsic to mafic in composition (Butler and Secor, 1991). Rocks within the Carolina terrane are heterogeneously deformed and metamorphosed at greenschist facies conditions (e.g. Hibbard et al., 2002). At least four distinct volcanic sequences within the Carolina terrane can be identified on the basis of stratigraphy, geochronology, geochemistry, and tectonic history (Figure 3). Rocks in the study area share similar stratigraphic and structural characteristics with at least two of the four sequences: the Virgilina sequence and the Albemarle sequence. For this reason, these two sequences shall be discussed in detail for the purposes of comparison throughout the remainder of this investigation.

The *Virgilina sequence* comprises the bulk of the metavolcanic rocks in the northern portion of the Carolina terrane along the border between North Carolina and Virginia (Figure 4). Strata in the Virgilina sequence are estimated to be at least 8 km thick (Glover and Sinha, 1973) and form three formations (Figure 5). The lowest formation is a thick sequence of felsic to intermediate volcanic and volcaniclastic rocks called the Hyco Formation, the base of which is not exposed (Glover and Sinha, 1973). The Hyco Formation is overlain by the Aaron Formation, which is composed of conglomerates and metaclastic sediments. Above the Aaron Formation is a series of metabasalts and greenstones termed the Virgilina Formation (Harris and Glover,
**Figure 4.** General geology of the Carolina zone in central North Carolina and southern Virginia depicting the Virgilina and Albemarle sequences. Box represents approximate area for Figure 6. Modified after Wortman et al. (2000) and Harris and Glover (1988).
1988). High precision U-Pb zircon ages obtained from rhyolites, dacitic tuffs, and intrusive rocks in the Virgilina sequence constrain its age to ca. 633-570 Ma (Wortman et al., 2000). Strata in the Virgilina sequence appear to have been deformed prior to the deposition of the Albemarle sequence by an event termed the Virgilina deformation (e.g. Glover and Sinha, 1973, Harris and Glover, 1988), the timing of which has been constrained to ca. 617-544 Ma on the basis of U-Pb zircon ages gathered from deformed gneisses and undeformed granites throughout the sequence (Wortman et al., 2000, Hibbard et al., 2002).

The *Albemarle sequence* extends from central North Carolina to the South Carolina state line (Figure 4) and consists predominantly of epiclastic and volcanic rocks including the Albemarle Group along with sills, dikes, and stocks of mafic and felsic magmatic rocks (e.g. Conley and Bain, 1965, Ingram, 1999). The entire sequence is estimated to have a thickness greater than 15 km (Butler and Secor, 1991) and is interpreted to unconformably overlie the Virgilina sequence (Harris and Glover, 1988). Nomenclature for the strata in the Albemarle Group has been revised several times since it was first established by Conley and Bain in 1965 (Stromquist and Sundelius, 1969, Sundelius and Stromquist, 1978, Gibson and Teeter, 1984) with the most widely accepted revision being that of Milton (1984). Figure 5 illustrates the general stratigraphy for the Albemarle and Virgilina sequences as well as the evolution of Albemarle sequence nomenclature.

The Uwharrie Formation is the lowest stratigraphic unit in the Albemarle sequence, the base of which is not exposed, and it conformably underlies the Albemarle Group (Figure 5) (Conley and Bain, 1965, Butler and Ragland, 1969). It
consists of pyroclastic felsic volcanic tuffs interlayered with basalt and rhyolite (Harris and Glover, 1988) all of which were deposited in a suprasubduction zone setting (Ingle, 1999). Oxygen isotope studies indicate the volcanic strata deposited in the Uwharrie Formation interacted with isotopically light waters which is indicative of a sub-aerial volcanic environment (Feiss et al., 1993).

The Uwharrie Formation is overlain by a series of submarine epiclastic sedimentary rocks and bimodal volcanic strata termed the Albemarle Group (Figure 5). From stratigraphically lowest to highest, the formations of the Albemarle Group are the Tillery, Cid, Floyd Church, and Yadkin Formations. The Tillery Formation is predominantly composed of thinly laminated argillaceous mudstones and shales while sedimentary rocks that compose the Cid Formation are distributed in significantly thicker bedded sequences than the Tillery Formation. The Cid Formation also contains a larger volume of volcanic rock including the voluminous Flat Swamp member, which caps the Cid Formation and is a good marker bed within the sequence (Figure 4 and 6). Above the Cid Formation is a series of medium-bedded clastic sediments assigned to the Floyd Church Formation and atop the entire sequence are mafic-derived greywackes that belong to the Yadkin Formation (e.g. Gibson and Teeter 1984, Butler and Secor, 1991).

Zircons from the Uwharrie formation have yielded U-Pb isotopic ages of 554±15 Ma (Ingle et al., 2003) and 551±8 Ma (Ingle, 1999) indicating the formation is late-Neoproterozoic to earliest Cambrian in age. Several inherited zircons recovered from the Uwharrie Formation yielded U-Pb ages of 991.3±4 Ma and
Figure 5. Evolution of the nomenclature for the Albemarle sequence including the interpreted stratigraphic relationship between the Albemarle and Virgilina sequences. Also included are the apparent conflicts in the perceived stratigraphy with respect to available age constraints based on geochronology and fossil data. After Hibbard et al. (2002), Butler and Secor (1991), and Stromquist and Sundulius (1969).
1892±6.3 Ma, which suggests magmatism in the unit involved a component of older continental crust (Ingle et al., 2003). This interpretation is supported by a broad range in $\varepsilon_{Nd}$ values in the formation which imply multiple sources for volcanic and plutonic rocks in the sequence (Wortman et al., 2000, Ingle et al., 2003).

Early studies (e.g. Black, 1978) indicate most of the Albemarle Group is Early Cambrian in age; however, fossil evidence has challenged this view (Koeppen et al., 1995). Fossils recovered from the Tillery and Cid Formations indicate that they are no older than early Middle Ordovician and Late Cambrian, respectively. However, the relatively narrow spectrum of ages obtained from zircons in volcanic rocks from the Uwharrie, Tillery, Cid, Flat Swamp Member, and Yadkin (Ingram, 1999; Ingle, 1999; Ingle et al., 2003) indicate that the overlying Albemarle sequence was deposited over a relatively brief time span during the Neoproterozoic and Early Cambrian. If the fossil evidence is correct, then it indicates that there are unrecognized unconformable inliers of younger strata within the Ablemarle Group (e.g. Hibbard et al., 2002).

Rocks in the Albemarle sequence are mildly deformed and retain many of their primary features, thus stratigraphy within the sequence can be established. The regional metamorphic grade for the sequence is greenschist facies (Butler and Ragland, 1969). Strata in the Albemarle sequence have been deformed by a series of northeast trending, upright to slightly inclined, southeast vergent, doubly plunging macroscopic folds (Figure 4 and 6) that possess a steep dipping axial planar cleavage. $^{40}$Ar/$^{39}$Ar cooling ages obtained from white mica associated with the axial planar cleavage to regional folds suggest that folding in the Albemarle Group took place
during the Late-Ordovician (Noel et al., 1988, Offield et al., 1995, Ayuso et al., 1997).

The GHfz is a prominent first-order structure within the Carolina zone of variable width and extends approximately 200 km along strike from Greensboro, North Carolina southward into northern South Carolina (Figure 1). Along much of its strike in North Carolina, the GHfz is interpreted to separate medium- to high-grade rocks of the Charlotte terrane from low-grade rocks of the Carolina terrane. Around the North Carolina-South Carolina border, the GHfz appears to turn westward into the Charlotte terrane and is eventually truncated by the central Piedmont shear zone near Carlisle, South Carolina (Lawrence and Corbett, 1999). Near the state border in the vicinity of Waxhaw, North Carolina the GHfz reportedly splits into a smaller south-southwest trending shear zone informally named the Davis Branch shear zone (DBsz), which continues south along the Carolina-Charlotte terrane boundary (Boland, 1996). Although no outcrop of the GHfz has been reported north of Greensboro, a magnetic gradient in that area parallel to the strike of the fault zone continues north-eastward and merges with the central Piedmont shear zone, suggesting the GHfz may follow this path (Carpenter, 1982).

In central North Carolina the GHfz (Figure 6) is a 3 to 5 km wide zone of brittle-ductile shear that deforms sedimentary and volcanic strata in the Albemarle sequence. In this region it is bounded to the northwest by the Gold Hill fault and to the southeast by the Silver Hill fault. Recent work in the vicinities of High Rock
Figure 6. Regional geology of the Gold Hill fault zone and Davis Branch shear zone in central North Carolina as observed at the onset of this study. Box indicates the approximate location of the study area. GHF - Gold Hill fault, SHf - Silver Hill fault, CPsz - Central Piedmont shear zone, SVS - Silver Valley Syncline, NLS - New London Syncline, TA - Troy Anticline. After Hibbard et al. (2003).
Lake and Mt. Pleasant, North Carolina has revealed the GHfz is a sinistral reverse fault system that dips steeply to the northwest and thrusts the Charlotte terrane over the Carolina terrane (Standard et al., 2002, Hibbard et al., 2003). A wide damage zone is associated with the footwall of the GHfz in the Carolina terrane in which the volcanlastic strata in the Albemarle sequence has been heterogeneously deformed (Standard 2003).

Earliest motion within the GHfz is constrained by its interaction with regional structures in the Albemarle sequence. The zone truncates and is folded by large folds within the Albemarle sequence (Figures 6) suggesting that earliest motion within the GHfz is coeval with Late-Ordovician folding in the Albemarle sequence (Hibbard et al., 2003). A recent $^{40}$Ar/$^{39}$Ar study on deformed granitoids and phyllites in the GHfz indicates the fault zone in central North Carolina was also active during the middle- and late-Paleozoic (Lavallee, 2003).

In northern South Carolina, the GHfz lies entirely within the Charlotte terrane and varies in width from 0.5 km near Rock Hill, South Carolina (Boland, 1990, 1996) to as wide as 10 km near Chester, South Carolina (Lawrence, 2001). Near the towns of Chester and Carlisle, South Carolina, the GHfz has been mapped as a ductile shear zone that predominantly consists of mylonites and lineated gneisses that were deformed and metamorphosed at amphibolite facies (Lawrence and Corbett, 1999). Kinematic indicators in the fault zone in this region reveal it to be a dextral shear zone. Outcrop of the Gold Hill and Silver Hill faults in northern South Carolina is as yet unrecognized. An $^{40}$Ar/$^{39}$Ar study performed on deformed amphibolites and phyllites mapped near and within the zone reveal the GHfz in South Carolina was
active during the middle-Paleozoic (Boland and Dallmeyer, 1997). More recent
\(^{40}\text{Ar}/^{39}\text{Ar}\) geochronologic studies in South Carolina also indicate the GHfz was active
during the late-Paleozoic (Lawrence, 2001, Lavallee, 2003).

**Previous Work**

Laney (1910) first recognized the GHfz in North Carolina, which is later
depicted on regional geologic maps compiled by Stromquist and Sundelius (1975),
Stromquist and Henderson (1985), and Boland (1996). Until recently few detailed
investigations of the fault zone have been attempted; however, there are several
interpretations concerning its significance in the southern Appalachians, including the
following: 1) a sharp metamorphic gradient between the Charlotte and Carolina Slate
belts (Tobish and Glover, 1971, Williams and Hatcher, 1982, 1983); 2) the margin of
a Neoproterozoic pull-apart basin (Moye and Stoddard, 1987); 3) a major fault zone
separating two distinct exotic terranes (Gibson and Huntsman, 1988, Schroeder and
Nance, 1988, Boland, 1996, Hibbard, 2000). The later interpretation seems most
likely; however, the significance of the GHfz and its relation to the tectonic history of
the Carolina zone has yet to be completely understood.

Previous work in the vicinity of Waxhaw, North Carolina during the last half
century was primarily limited to reconnaissance mapping related to mineralogical and
natural resource studies (Butler, 1966, Randazzo, 1972, Butler, 1977, Wagener, 1977,
Nance, 1984, Allen and Hepp, 1985). The first detailed academic investigation that
focused on the GHfz in the Waxhaw area was carried out by Schroeder (1987) at the
Howie Gold Mine in Union County, North Carolina (Figure 2). This study focused
primarily on structural controls of gold mineralization within the Howie mine.

Schroeder (1987) reported that the GHfz in this area is a middle-Paleozoic shear zone that deforms laminated phyllitic mudstones and chloritic schists distributed throughout the study area. He also reported multiple foliations within the study area. The earliest ($S_1$) is a regional cleavage that is axial planar to regional folds in the study area and is interpreted to be Taconic based on a previous report by Butler and Fullagar (1978). Schroeder (1987) reported two foliations that overprint $S_1$ (designated $S_{2a}$ and $S_{2b}$) that are axial planar to a set of variably plunging wrench and drag folds ($F_2$). $F_2$ folds and their subsequent cleavages were attributed to various stages of progressive dextral shear along the GHfz during Acadian tectonism (Schroeder and Nance, 1988).

Boland (1996) was the first to map out the extent of the GHfz in the vicinity of Waxhaw, North Carolina and Van Wyck, South Carolina. She broadly characterized the GHfz as a wide zone of recrystallized phyllites and schists that juxtapose the Carolina and Charlotte terranes and interpreted the fault zone as a dextral shear zone on the basis of limited kinematic indicators. Boland also reported the existence of the Waxhaw shear zone which was later renamed the DBsz. Boland (1996) reported the DBsz as a kilometer wide, N-S trending, ductile, sinistral shear zone, as opposed to the dextral GHfz. Boland indicated that the DBsz merged with the GHfz east of Waxhaw, North Carolina yet did not characterize the DBsz beyond the observation that it appears to be a sinistral shear zone. However, she did imply that motion along the DBsz may be coeval with the GHfz. Both the GHfz and DBsz appear to truncate earlier regional folds and foliations in the surrounding strata and
have reportedly deformed the margins of the Waxhaw pluton, where both fault zones come in contact with the granite body (Boland, 1996).

The first age reported in the study area was a Rb-Sr whole rock age of 495±21 Ma for the Waxhaw pluton (Fullagar, 1981) that was interpreted to reflect the its crystallization age. However, this age is suspect due to the highly mobile nature of Rb in deformed rock bodies. Boland and Dallmeyer (1997) obtained Devonian \(^{40}\text{Ar}/^{39}\text{Ar}\) cooling ages from amphibolites and phyllites adjacent to and within the GHfz. These ages were interpreted to reflect the timing of motion along the fault zone, which Boland and Dallmeyer (1997) constrained to c.400-368 Ma. Lavallee (2003) dated micas from sheared phyllite xenoliths in the nearby Catawba granite (Figure 7) which yielded an average age of 323±1.5 Ma. She reports similar ages from the Cotton Grove metagranodiorite located in the GHfz in the High Rock Lake, North Carolina vicinity. These later ages gathered by Lavallee (2003) and Boland and Dallmeyer (1997) indicate the GHfz was potentially active throughout the middle and late Paleozoic.

In summary, previous work along the GHfz in the vicinity of Waxhaw, North Carolina pertaining to its timing and kinematics suggest it is a dextral shear zone that was active during the middle and late Paleozoic. This complex history may be reflected by multiple overprinting fabrics reportedly within the fault zone (Schroeder, 1987). Earlier sinistral shear in the GHfz in central North Carolina (Standard, 2003, Hibbard et al., 2003) may be manifest in the DBsz even though no attempt as of yet has been made to correlate these two structures. Understanding the relationship
between the GHfz and the DBsz as well as the nature of the multiple fabrics reportedly in the GHfz will provide a critical step in unraveling its tectonic history.

**Location and Methods**

The thesis area is located on the North Carolina-South Carolina border approximately 20 miles southeast of Charlotte, North Carolina, around the small township of Waxhaw, North Carolina. Previous mapping by Boland (1996) in the Laurens 30x60 minute quadrangle shows that the GHfz trends through the Waxhaw area and bifurcates against the Waxhaw pluton (Figure 6). The area around Waxhaw, North Carolina was chosen because it is the best locality to study the interaction between the GHfz and DBsz and it is also ideal because it offers the opportunity to examine the interaction of the GHfz with the older Waxhaw pluton as well as the volcanic and epiclastic rocks of both the Carolina and Charlotte terranes.

The field area covers approximately 178 square kilometers (68.5 sq. mi) and is contained within portions of the Catawba NE and Waxhaw 7.5 minute USGS topographic quadrangles. Topography of the study area is consistent with that of the Carolina Piedmont, being characterized by gently rolling hills and pastures with limited relief except around major drainages. The highest elevation occurs at 220 m (720 ft) along the southeastern edge of the field area while the lowest elevation is at 140 m (460 ft) and is located at the point where Twelve Mile Creek flows out of the southwest corner of the study area. Maximum topographic relief for the field area therefore is 80 m (260 ft).
Because the greatest relief is located around incised streams and creeks, a significant portion of outcrop occurs along local drainages. Major drainages within the field area include Twelve Mile Creek and its two tributaries, Tarkill Branch and Six Mile Creek. The flow of major streams is predominantly to the southwest with some smaller creeks flowing south-southeast. This pattern reflects the control of local bedrock geology on drainage patterns within the study area, with traces of many streams, including Twelve Mile creek, oriented parallel to the strike of foliations across the field area.

An extensive geologic investigation was undertaken for this particular study, the focus of which involved approximately 80 field days of detailed mapping at the 1:24,000 scale. Traverses were planned to cover key localities and take advantage of the fluctuation in relief along major drainages. The rock type, fabric, and structural nature of each outcrop were studied and documented in order to establish a local stratigraphy, characterize major geologic features, and understand the interaction between the GHfz and surrounding country rock. Accompanying field work during this study was petrographic and structural analysis of approximately 51 thin sections, detailed strain analysis on 10 oriented samples, and microprobe analyses on minerals from a metabasalt sampled from within the fault zone.

Because rocks in the study area are mildly deformed and retain many of their original sedimentary and volcanic features, the prefix “meta-“ is omitted from rock descriptions in the following chapters for brevity and in keeping with current literature (e.g. Butler and Secor, 1991). Also, sedimentary and some volcanic rock in the study area resemble slate rather than phyllite, yet they contain mica that is large
enough to produce a dull sheen on cleavage surfaces. For this reason, the term phyllitic-slate is considered a better descriptive device for deformed and partially recrystallized rock in the GHfz.
Lithostratigraphic Units

General Statement.

This chapter presents the descriptions and stratigraphic relationships of the rock types that occur within the study area. The rocks that underlie the study area are divided into two sequences that belong to the Carolina terrane (Figure 7), namely the Marvin sequence and the Twelve Mile sequence. These volcaniclastic sequences are distinguished from one another on the basis of rock composition, structural position, and interpreted ages (Figure 8). A suite of granitic bodies intrude volcanic strata within the Twelve Mile sequence and the largest of these intrusive bodies, the Waxhaw granite, has locally metamorphosed part of that sequence. Map scale bodies of gabbro occur in rocks affiliated with the Twelve Mile sequence and both volcanic sequences, as well as the Waxhaw granite, are intruded by Mesozoic diabase dikes.

Previous mapping in the vicinity of Waxhaw indicates rocks that lie in the Marvin sequence belong to the Charlotte terrane (e.g. Boland, 1996); however, evidence will be presented below that contradicts this assessment. Because the primary purpose of this investigation is to determine the structure and kinematics of the GHfz, rock units proximal to the fault zone were mapped in detail while reconnaissance mapping was completed outside of the GHfz. For this reason, general map patterns and the geographic extent of some rocks types located outside the fault zone are taken from Boland (1996). This compilation pertains to reported structural data, lithologic contacts, and fold patterns in the southeastern portion of the
Figure 7. Generalized geologic map of the Waxhaw, NC area showing the two major lithostratigraphic divisions and intrusive rocks.
Figure 8. Stratigraphic column of the rock units in the Waxhaw area.
Zms-metasediments in the Marvin sequence; Zfv-volcanic strata in the Marvin sequence; ZCfv-Twelve Mile felsic volcanic unit; ZCms-Twelve Mile sedimentary unit; Wg & Hg-Waxhaw and Hancock granites; Mgb-gabbro.
field area. However, interpretations of these same contacts, regional stratigraphic relationships, and unit designations presented here differ from previous studies completed around Waxhaw, North Carolina.

This chapter presents a discussion of the lithologic units as well as a tentative stratigraphy for volcanic and volcaniclastic strata in the Carolina terrane in the field area. The format for lithologic descriptions in this chapter will be as follows: 1) a description of the geographic extent of each unit; 2) lithologic description of each unit; 3) contact relations of the unit, and 4) age and correlations for the unit where possible. Strata in the Marvin sequence shall be described first, followed by the Twelve Mile sequence, and finally plutonic rocks that intrude the study area.

The Carolina Terrane

Volcanic and volcaniclastic strata assigned to the Carolina terrane comprise the bulk of the geology in the study area. There are two separate packages of volcaniclastic strata within the Carolina terrane in the study area that are informally named the Marvin and Twelve Mile sequences in this study. Volcanic and sedimentary rocks of the Marvin sequence outcrop to the north and west of the GHf and are interpreted as the older of the two sequences (Figure 8). The younger Twelve Mile sequence lies south and east of the GHf and is separated into a felsic volcanic unit (ZCfv) and a sedimentary unit (ZCms). The following section describes the volcanic and epiclastic strata in these two sequences starting with the older Marvin sequence.
Marvin Volcanic Sequence (Zfv and Zms)

Distribution

The Marvin volcanic sequence incorporates all pyroclastic and volcaniclastic rocks that underlie the field area north of the GHf in the hanging wall of the GHfz (Figure 7). The sequence is informally named after the town of Marvin, North Carolina which is situated approximately five miles north of Waxhaw near the center of the volcanic sequence in the study area. The Marvin sequence covers approximately 45.5 sq. km of the northwestern section of the field area and extends beyond the field area to the northwest. It has a maximum outcrop width of 4720 m.

Rocks in the Marvin sequence consist of unseparated felsic tuffs and porphyries (≥85%), volcaniclastic and sedimentary phyllitic-slates (~15%), and chlorite schist (≤5%). Regional cleavages are poorly developed throughout most of the sequence except where the sequence lies near the GHfz. Cleavages are steep dipping and have an average strike of N60°E. Because only a small portion of the Marvin sequence outcrops within the GHfz, the majority of the field work completed in the sequence consisted of highway reconnaissance and only a few detailed traverses were completed. Therefore, little is known of the field relations and intra-sequence contacts within the Marvin sequence. The author was able to recognize and separate several rock types within the Marvin sequence, hence differing rock types are described below. However, because less field work was completed in the Marvin sequence than the Twelve Mile sequence, rock types in the Marvin sequence are not divided into sub-units.
Rock Types

_Felsic Volcanic Rocks (Zfv)_

Felsic tuffs and porphyries are the most abundant rock type within the Marvin sequence, making up approximately 85% of the sequence. Of these, the most common rock types are felsic tuff and lapilli tuff. Tuffs and lapilli tuffs in the Marvin sequence are quartzofeldspathic, fine-grained, light gray to powdery white, and contain variable concentrations of quartz and feldspar phenocrysts (~2-10%) and pumice lapilli. Distinctive fine and medium grained, gray and blue-gray rounded quartz is the most abundant phenocryst in lapilli tuffs and porphyries. Plagioclase laths are less abundant and range in length from 1 to 3 mm. Pumice lapilli in the felsic tuffs have an oblate shape, are chalky white in color, and range in length from 5 to 20 mm.

Quartz porphyries are interspersed with felsic tuffs in the Marvin sequence. These porphyries are mostly concentrated along the western edge of the field area and are less common in the central and eastern portions of the field area. Phenocryst concentration in the quartz porphyries ranges from 50% to 80% of the total rock volume. Distinctive medium-fine, round, blue-gray quartz is the most common phenocryst. Feldspar phenocrysts are euhedral, range in length from 1 to 5 mm and are typically fractured (Figures 9). Dipyramidal β-quartz was observed in several hand specimens of quartz porphyry which, along with the fractured feldspar phenocrysts, suggests a pyroclastic origin for these rocks.

Felsic tuffs and porphyries in the Marvin sequence have an igneous assemblage of Qtz + Olg (An28-30) + Bio and a metamorphic assemblage of Wm ±
Figure 9.  

A) Typical exposure of felsic crystal tuff in the Marvin Sequence. 
B) Same outcrop of felsic crystal lapilli tuff with fractured feldspar crystals and round blue-gray quartz. Unnamed creek in Weddington Chase housing development.
Chl ± Ep ± Grt ± Opq. In thin section, groundmass is composed of minerals that are coarser grained and more quartz rich than groundmass in tuffs from the Twelve Mile sequence. Plagioclase laths are euhedral and exhibit Carlsbad and Albite twinning. K-feldspar was not observed in thin section. The An-content of plagioclase in Marvin tuffs and porphyries along with the absence of K-feldspar in thin section indicates that felsic volcanics in the Marvin sequence have the composition of dacite.

**Sedimentary Rocks (Zms)**

Gritty siltstones and poorly sorted volcaniclastic sediments make up a sedimentary package within the Marvin sequence. The most abundant sedimentary rock in the sequence is massive to finely laminated siltstone. Marvin sequence siltstone is tan to greenish-brown with a gritty texture that sets it apart from siltstone in the Twelve Mile sequence. Outcrops of siltstone are massive to finely bedded. Bedding consists of rhythmically alternating light and dark laminations that range in width from 2 to 5 mm. Laminations typically occur as alternating light and dark brown layers; however, in several outcrops laminations are yellow, green, and red as well as brown. Siltstone occurs as lenses and small pods in the mostly volcanic Marvin sequence and forms a marker unit useful for identifying small folds and brittle faults within the sequence.

In one outcrop near the GHf, beds of poorly sorted conglomerate and volcanic breccia occur in the Marvin sequence (Figure 10). These volcaniclastic sediments consist of 5 or 6 continuous beds interlayered within siltstone and range in thickness from 0.4 to 2 m. Volcanic conglomerates are matrix supported and clast size varies
from pebbles to small boulders. Volcanic clasts within these beds are dominantly felsic and chalky white in color; however, intermediate and mafic clasts also occur in these volcanioclastic layers. Another noteworthy rock type that occurs near this same outcrop is a continuous layer of mafic, black and purple, very fine-grained phyllite. The mafic phyllite has the appearance of a graphitic phyllite with a strong phyllitic sheen on its cleavage surface and is folded by numerous kink bands. This particular rock type is commonly found adjacent to the GHf on the hanging wall side of the fault zone in central North Carolina thus suggesting that it forms a reliable marker unit in the hanging wall of the GHfz (Hibbard pers. comm., 2004).

**Chlorite Schist**

Narrow bands of chlorite schist and chloritic conglomerates occur within the Marvin sequence near the GHf. These schists make up approximately 5% of the Marvin sequence and occur as small layers and lenses in the felsic volcanic rocks of the sequence. These schists are dark green, composed of medium grained chlorite and epidote, corroded feldspar laths, and poorly sorted and rounded clasts of volcanic tuff and pumice and sedimentary rock. The mineral assemblage in these schists consists of Chl + Ep + Wm + Plag + Qtz ± Bio ± Sph ± Opq. Chlorite flakes range in size from 5 to 10 mm. Feldspar laths are subhedral and highly saussuritized. Volcanic clasts are dominantly felsic, range from 1-7 mm in length, and are rounded and flattened parallel to schistosity.
Figure 10. Volcanic breccia in sedimentary rocks in the Marvin Sequence at the Van Wyck Brick Company quarry. Hammer for scale.
Contacts

The southeastern boundary of the Marvin sequence is interpreted as the GHf, which juxtaposes the Marvin sequence against felsic volcanic and sedimentary strata of the Twelve Mile sequence. The Marvin sequence continues to the northwest where it is reportedly intruded by hypabyssal rocks and metagranites associated with the Charlotte terrane (Boland 1996). Contacts between volcanic and sedimentary strata in the Marvin sequence were not observed in outcrop and the nature of the contact between sediments and felsic volcanic tuffs in the Marvin sequence remains unclear. However, the contact between siltstones and volcanic tuffs within the Marvin sequence is fairly sharp because several outcrops of felsic volcanic tuff and sedimentary rock were observed within a distance of 30 m perpendicular to strike along Long Branch and Tarkill Branch.

Age and Correlation

Rocks in the Marvin sequence were previously mapped as an unnamed sequence of predominantly mafic volcanic rocks and meta-granitoid that were was correlated with the Charlotte terrane (Boland 1996). The primary basis for this interpretation was that the Marvin sequence lies to the west of the GHf (Schroeder, 1987, Boland, 1990, Boland, 1996). However, field work in the Marvin sequence during this study was unable to identify mafic rocks that have been ductily deformed and metamorphosed at amphibolite facies. Instead, the Marvin sequence is comprised of dominantly felsic volcanic tuffs and quartz-porphyries with some mafic-derived sedimentary rocks and schist. Furthermore, rocks in the Marvin sequence are only
mildly deformed except where they come in contact with the GHf. Based on these observations, the Marvin sequence appears to more strongly correlate with elements of the Carolina terrane rather than the dominantly plutonic and ductily deformed Charlotte terrane.

The occurrence of low-grade volcanic rocks west of the GHfz is well documented in the High Rock Lake and Mount Pleasant areas, North Carolina (Hibbard et al., 2003). In those areas, dacitic tuffs and porphyries that bear distinctive blue-gray quartz phenocrysts compose most of the country rock in the hangingwall of the GHfz. These tuffs and porphyries along with granitic plutons have yielded high precision U-Pb zircon ages between 620 and 615 Ma (Miller et al., 2003, Hibbard pers. comm., 2004), which are characteristic ages of the Virgilinia sequence in the Carolina terrane. Because felsic tuffs and porphyries in the Marvin sequence occupy a similar structural position as well as resemble volcanic rocks in the High Rock Lake and Mount Pleasant areas, they are assumed to be of a similar age. If low-grade volcanic strata belonging to the Carolina terrane occur on both sides of the GHfz, it brings into question the interpretation that the GHfz is the tectonic boundary between the Carolina and Charlotte terranes.

**Twelve Mile Sequence (ZCfv and ZCms)**

The Twelve Mile sequence comprises all volcanic and sedimentary rocks that lie southeast of the GHf. It is named after Twelve Mile Creek, which trends through much of the sequence. Because most of the field work during this study was completed in the Twelve Mile sequence, more is known about rock types and unit
relationships than in the Marvin sequence. Therefore, the Twelve Mile sequence has been split into two separate units: the Twelve Mile volcanic unit (ZCfv) and the Twelve Mile sedimentary unit (ZCms). The following section will discuss the distribution of the rock types in the Twelve Mile sequence, its contacts with surrounding sequences as well as intra-sequence contacts, and the age and correlation of the sequence. Rock descriptions have been broken into two separate sections that correspond to the two units in the Twelve Mile sequence.

Distribution

All volcanic strata that lie southeast of the GHf in the footwall of the GHfz in the study area form the volcanic unit of the Twelve Mile sequence. Volcanic rock in the Twelve Mile sequence is distinct from the Marvin sequence on the basis of morphology, mineral composition, and inferred age. Crystal tuffs and porphyries of the volcanic unit are exposed in the cores of large regional antiforms within the GHfz and appear to lie structurally beneath sedimentary strata of the Twelve Mile sequence (Figure 2). A small map scale lens of the volcanic tuff also occurs in the same general location as the Howie Gold Mine within the sedimentary unit (Figure 7).

The volcanic unit has an outcrop width that ranges from ~7500 m to less than 850 m in regional folds and extends beyond the study area to the west and southeast. Volcanic rocks in the Twelve Mile sequence are composed of unseparated rhyolitic and rhyodacitic tuffs interlayered with minor intermediate and mafic tuffs and locally developed phyllites. The dominant rock types in the Twelve Mile volcanic unit are
unseparated felsic pyroclastic rocks that can be sub-divided into massive crystal tuffs (60%), crystal lapilli tuffs (30%), and coarse crystal lithic tuffs (≤10%).

The Twelve Mile sedimentary unit is interpreted to lie structurally above the volcanic unit on the basis of stratigraphic younging indicators and the relative structural position of the sedimentary unit with respect to the volcanic unit (Figure 2). Strata in the unit consist of unseparated mudstones (≥85%), graded sandstone (~10%), and local conglomerate (≤5%). Sedimentary rocks assigned to the Twelve Mile sequence are most extensively exposed along the eastern boundary of the study area and lie in the cores of large northeast plunging synforms that extend southwestward across the central portion of the study area.

The Twelve Mile sedimentary unit has a maximum estimated outcrop width of 4,877 m and tapers to less than 600 m in the southwestern hinge areas of regional folds. The sedimentary unit discontinuously extends beyond the field area to the south and east. Previous mapping indicates the southern edge of the GHfz truncates the sedimentary unit and that all rock within the fault zone was either obliterated by deformation in the GHfz or recrystallized into phyllite with the exception of a small lens of sedimentary rock around the Howie Gold Mine (Boland, 1996). Instead, this study found that the sedimentary unit maintains recognizable sedimentary features into the GHfz.
Rock Types: The Twelve Mile Volcanic Unit (ZCfv)

_Felsic crystal tuffs_

The most extensive rock type exposed in the Twelve Mile volcanic unit is a monotonous array of featureless felsic crystal tuffs. Fresh exposures of crystal tuff are light-gray to greenish-gray and weather to a tan or chalky white. The best exposures of felsic crystal tuff are located 0.5 mi north of Twelve Mile creek on Hwy. 521 and along Missouri Branch just east of S.R. 1113 in the southern region of the field area. Crystal tuff in the Twelve Mile volcanic unit is predominantly aphanitic and massive showing little to no volcanic layering or sorting. Sparse porphyritic crystal tuffs occur locally and contain up to 5% phenocrysts. Feldspar laths between 1 to 3 mm long are the most common phenocrysts in crystal tuff whereas biotite and quartz are less abundant. Volcanic layering is rare but where encountered, it occurs as alternating beds of tan, rhyodacitic layers and dark gray andesitic layers.

Crystal tuff in the volcanic unit has an igneous assemblage of Qtz + Mic + San + Olg (An11-12) + Bio ± Hem ± Py and a metamorphic assemblage of Wm ± Ep ± Chl ± Grt. In thin section, feldspar phenocrysts are composed of laths of plagioclase, microcline, and sanidine. Microcline occurs as both isolated phenocrysts and small inclusions within larger oligoclase crystals. Quartz phenocrysts are sub-rounded and slightly larger than feldspar laths. In one thin section, biotite concentrations define a compositional layering that is otherwise invisible in hand specimen.
Felsic lapilli tuffs and crystal lithic tuffs

Lapilli tuff makes up about 30% of the total volume of the Twelve Mile volcanic unit. The best exposures of lapilli tuff are located along Davis Branch Creek approximately 0.6 mi east of the southeastern edge of the Waxhaw pluton. The groundmass of lapilli tuff in the volcanic unit is slightly coarser crystalline than crystal tuff and contains a substantially larger percentage of phenocrysts, lapilli, and lithic fragments. Felsic crystal lapilli tuff is gray, blue, and greenish gray where fresh and weathers to a light tan, brown, and on occasion dark purple. Lapilli tuff in the Twelve Mile sequence is rhyolitic to rhyodacitic in composition, moderately to poorly sorted, and contains 10-40% subhedral feldspar laths and round blue-gray quartz phenocrysts. Feldspar phenocrysts are slightly more abundant than quartz and typically range in size from 1 to 4 mm, whereas rounded quartz crystals tend to range from 2 to 5 mm in diameter. Lapilli and phenocrysts are almost always elongate in the direction of regional cleavage. Twelve Mile lapilli and crystal lithic tuff (below) has a mineral assemblage similar to fine crystal tuffs: Qtz + Mic + Olg (An11-13) + Bio + Wm ± Chl ± Ep ± Zeo ± Hem ± Rt.

A small percentage (≤5%) of felsic rock in the volcanic unit is composed of tuffs with lapilli and lithic fragments that exceed 5 mm in size (Figure 11). The best exposure of crystal lithic tuff can be seen in an abandoned quarry in a small residential park off of Henry Harris Rd 0.3 miles east of Hwy 521. Crystal lithic tuff is typically poorly sorted and volcanic layering is rare. Groundmass is gray to blue-gray and consists of quartz, feldspar, and white mica. Lithic fragments are composed of chalky white pumice lapilli and blocks, quartz lapilli 5-10 mm in width, feldspar
Figure 11. Felsic crystal lithic tuff in the Twelve Mile volcanic unit. Lithic clasts dominantly consist of pumice lapilli and blocks that are flattened in the direction of foliation. Clasts in this outcrop are very poorly sorted and primary bedding is unrecognizable. Abandoned quarry in a small trailer park off of Henry Harris Rd. Photograph courtesy of Jim Hibbard.
phenocrysts, and minor volcanic clasts that are felsic to mafic in composition. The average aspect ratio for most lapilli and lithic fragments in these tuffs is 3:1 to 4:1 (length/width); however, some have aspect ratios as great as 9:1. Several outcrops of lithic tuff contain clasts of concentrically layered quartz rings, which are interpreted to be quartz lithophysae. Lapilli and blocks in crystal lithic tuffs are rounded and elongate in the direction of regional cleavage.

*Intermediate and Mafic tuffs (ZCiv and ZCmv)*

Intermediate and mafic volcanic rocks in the Twelve Mile volcanic unit occur as small lenses, layers, and isolated outcrops within the predominantly felsic volcanic unit. Tuffs and porphyries of intermediate composition are typically dark gray to brown and are interpreted to have the composition of andesite. Andesitic tuff occurs sporadically throughout the Twelve Mile sequence with the largest exposure located on the CSX Rail line approximately 4 km west of the town of Waxhaw. Andesitic tuff is typically fine- to medium-crystalline and contains from 10-60% phenocrysts of biotite, plagioclase, and amphibole, with little or no quartz. Mafic tuff occurs in dark green to black layers that are easily distinguishable from the light colored felsic rocks in the volcanic unit (Figure 12). The greatest concentration of mafic tuff occurs between the Waxhaw and Hancock granites approximately 6 km west-southwest of Waxhaw. Mafic tuff has the composition of basalt and usually occurs in layers that are 5 to 20 m wide that extend up to several kilometers in length. In hand specimen, basaltic tuff is fine- to medium-fine crystalline and contain amphibole needles that are 1 to 3 mm in length along with green epidote and chlorite.
Figure 12. Contact between felsic tuff and mafic tuff in the Twelve Mile volcanic sequence. Just off of Cross Creek Estates Rd approximately 1.2 km ESE of the Hwy 521/ Hwy 75 intersection. Hammer for scale.
Phyllites

The greatest abundance of phyllite occurs within the volcanic rocks of the Twelve Mile sequence. Phyllite is distributed along the GHf and also occurs within the fault zone in discrete isolated lenses approximately 1-2 m wide. Phyllites (s.s.) are slightly coarser grained and more micaceous than phyllitic-slates and as a result have a stronger phyllitic sheen when exposed to sunlight and bear mica crenulations characteristic of phyllites. They are interpreted to be the result of heterogeneous deformation within the GHfz, which resulted in the complete recrystallization of tuff and volcanic sediments. Phyllite in the volcanic unit is fine grained and silver-gray to deep blue in color. Petrographic analysis indicates that phyllite in the Waxhaw area has a mineral assemblage of Wm + Qtz + Plag + Chl ± Py ± Hem ± Ep. Aligned white mica and chlorite define foliations in phyllite, which is parallel to regional cleavages.

Rock Types: Twelve Mile Sedimentary Unit (ZCms)

Mudstone and Siltstone

The bulk of the sedimentary unit is made up of a series of monotonous, very fine-grained, and massive to finely laminated mudstone. The best outcrops of mudstone occur along Twelve Mile Creek in the northeastern portion of the field area, particularly around and just off of S.R. 1327. In outcrop, mudstones are blue-gray to greenish-gray where fresh and weather to a tan- or olive-brown. On occasion, isolated clasts of felsic and mafic tuff have been identified in several outcrops of mudstone.
All outcrops of mudstone bear a well developed slaty cleavage along which the rock breaks and weathers readily. Outcrops and hand specimens of mudstone usually lack primary features; however, about 30% of mudstone outcrop contain primary sedimentary features, the most common being fine laminations of alternating mudstone and siltstone (Figures 13a & b). Laminations occur as rhythmically alternating light and dark layers that are typically planar and continuous; however, wavy, uneven, and irregular shaped laminations also occur in Twelve Mile mudstone.

In thin section, dark layers are very fine-grained and micaceous while lighter colored layers are slightly coarser grained and quartzofeldspathic. Minor cross laminations, ripple marks, and slump folds occur in the mudstones as well. Several outcrops of mudstone contain a high concentration of fine-grained poikiloblastic pyrite.

*Sandstone*

Layers of fine- to medium-fine grained sandstone occur in the sedimentary unit and are most concentrated in the northeastern portion of the field area. Sandstone layers interfinger with mudstone and conglomerate and are observed to conformably overlie mudstone beds in several outcrops. Sandstone layers typically produce thicker and slightly more resilient sedimentary layers than mudstone and as a result bear a weaker cleavage than the mudstone layers. Sandstone layers also have more pronounced bedding and primary features for which they are most easily recognized.
Figure 13. A) Typical outcrop of laminated mudstone in the Twelve Mile sedimentary unit. B) Close-up of the same outcrop reveals that at this outcrop cleavage (subvertical) has a steeper dip than bedding (subhorizontal). Outcrop located in the Unity quadrangle on S.R. 1117 approximately 3.2 km north of Unity, South Carolina.
Figure 14. Photomicrograph of fine sandstone and siltstone laminations in an oriented sample from the Twelve Mile sedimentary unit. Top of the photograph is up. Sedimentary grading in this photograph indicates that bedding in this sample is upright. Sample location is on East Fork Twelve Mile Creek approximately 300 m west of Billy Howie Rd.
Outcrops of sandstone in the sedimentary unit are composed of graded beds that are blue-gray to dark gray in color on fresh surfaces and weather to an olive-brown. Beds range in thickness from 1 to 4 cm and contain lamellae of light colored sandstone that grades upward into darker siltstone lamellae (Figure 14). Sedimentary grading along with ripple marks and cross laminations consistently indicate stratigraphic younging direction. Grain size in Twelve Mile sandstone varies from 0.5 to 0.25 mm in fine-sand layers to $\leq 0.125$ mm in silty layers. Sandstones are predominantly composed of rounded quartz, fine mica, and poikiloblastic pyrite and hematite.

**Conglomerate**

Poorly sorted conglomerates lie in the cores of several of the large synforms in the Twelve Mile sedimentary unit (Figures 15a & b). The most spectacular outcrop of conglomerate occurs along the north bank of Twelve Mile Creek approximately 1.5 mi S20°E of the intersection of S.R. 1321 and S.R. 1315 at Union, North Carolina. Though no contact has been identified between conglomerate and other sedimentary strata, these conglomerates are interpreted to occupy the highest structural position in the field area due to their location in regional folds. Conglomeratic layers are relatively rare in the sedimentary unit only making up approximately 5% of its total volume.

Conglomerates in the sedimentary unit are greenish brown to gray when fresh and weather to a dark brown or tan. They are matrix supported and clast rich (between 40-50% clasts) with a groundmass composed of dark, fine-mudstone and
Figure 15. A) Outcrop of poorly sorted conglomerate in the Twelve Mile sedimentary sequence with clasts that range in size from coarse sand to small boulders. B) Close-up of the same outcrop indicates the predominant clast type in this conglomerate is rounded felsic volcanic clasts. Outcrop located on the north bank of East Fork Twelve Mile Creek ~2.3 km south-southwest of Union, North Carolina
siltstone. Clasts and lithic fragments in Twelve Mile conglomerates are predominantly felsic volcanic and are typically chalky white to tan. Mafic and intermediate volcanic clasts are less common and range in color from dark tan to black. Clasts are poorly sorted with no apparent bedding or grading and range in size from <1 cm to >7 cm with an average size of approximately 1 cm. Clasts have a well rounded and oblate shape where observed on faces that are perpendicular to foliation yet are angular to sub-angular and relatively undistorted when observed on rock faces parallel to foliation.

Contacts

The northwestern boundary of the Twelve Mile sequence is interpreted as the GHf, which separates both the volcanic and sedimentary units of the Twelve Mile sequence from volcanioclastic rocks in the Marvin sequence (Figure 8). While no outcrop of the contact between the Marvin and Twelve Mile sequences was observed, it is interpreted as tectonic for the following reasons: 1) cleavage intensity and finite strain in rocks in the Twelve Mile sequence increase towards the contact between the two sequences (see Structure chapter); 2) the frequency of recrystallized phyllites increases along the inferred trace of the contact; 3) the contact appears to truncate regional folds in the Twelve Mile sequence (Figure 2); 4) there is an abrupt change in rock type across the contact from feldspar-rich rhyolitic tuffs in the Twelve Mile sequence to quartz-bearing dacites in the Marvin sequence with no interfingering observed between the two sequences; 5) no clasts of the Marvin sequence were found
in the Twelve Mile sequence, which would otherwise indicate an unconformable contact.

In the southern portion of the field area, volcanic tuffs in the Twelve Mile sequence are in contact with the Waxhaw granite. Although the contact between these two rock units is not exposed, a metamorphic aureole extends up to a kilometer into the volcanic unit around the Waxhaw granite implying that these two units share an intrusive contact.

Within the Twelve Mile sequence, the volcanic and sedimentary units share an intra-sequence contact. This contact is abrupt with little interfingering of the two units and it outlines the trace of regional folds within the Twelve Mile sequence (Figure 8). It is not directly visible in any single outcrop but outcrops of felsic tuff and mudstone are observed within a distance of approximately 15 m in several locations. Near the contact, mudstones and siltstones typically become massive and in several locations lithic fragments of gray felsic tuff occur within the mudstone. Also, the orientation of volcanic layering is roughly parallel to the orientation of bedding in the sedimentary unit. Thus, it is believed the two units share a disconformable contact.

**Age and Correlation**

Volcanic and epiclastic strata in the Twelve Mile sequence have been mapped to the southeast beyond the study area where they are reportedly folded by the Troy anticlinorium (Figure 6). Previous workers have correlated felsic pyroclastic rocks and laminated mudstones in the Twelve Mile sequence with the Uwharrie and Tillery
formations of the Albemarle sequence (Schroeder, 1987, Boland, 1996) on the basis of lithologic similarity and similar stratigraphic relationships.

To date, there has been no attempt to determine a crystallization age for the volcanic rocks in the Twelve Mile sequence. An indirect age constraint comes from the Waxhaw granite that intrudes the Twelve Mile sequence. It has been dated as Late Cambrian (Fullagar, 1981), which places the Late Cambrian as a minimum age on the volcanic unit in the Twelve Mile sequence. Regional stratigraphic and structural position of the Twelve Mile sequence with volcaniclastic strata in the Albemarle, North Carolina region suggest the Twelve Mile volcanic unit may be related to the Uwharrie Formation. If this observation is correct, then it could be assumed that both units share similar crystallization ages, placing the age of the Twelve Mile volcanic unit at ca. 560 Ma.

The age of the sedimentary unit has not been determined directly by age dating or fossil data; however, it is believed to be younger than the volcanic unit based on its interpreted stratigraphic and structural position. Also, clasts of the Waxhaw granite were not observed in the sedimentary unit; and the Waxhaw granite, volcanic unit, and sedimentary unit all share a similar deformation. These observations suggest the sedimentary unit is not separated from the volcanic unit by a significant gap in age.

**Plutonic Rocks**

Plutonic rocks that occur within the study area include the Waxhaw pluton and the Hancock granite. The former is a large granitic body that has intruded and
metamorphosed part of the Twelve Mile sequence while the latter is a smaller granitic outlier to the larger pluton. A series of gabbroic lenses also intrude felsic volcanic tuffs in the southern portion of the field area. Diabase dikes intrude volcaniclastic rocks in both the Marvin and Twelve Mile sequences as well as the Waxhaw granite.

**Waxhaw Granite (Wg)**

**Distribution**

A large granitic pluton is present in the southern half of the study area and is informally named the Waxhaw granite (Boland 1996) after the nearby town of Waxhaw, North Carolina. The Waxhaw granite was mapped previously as part of a suite of quartz metaporphyry, metadiorite, and metagranite termed the Waxhaw Complex (Boland 1996). The Waxhaw pluton is a large asymmetric shaped granitic body that occupies approximately 22.5 sq. km of the southern half of the study area and extends beyond the study area to the southwest. The Waxhaw granite is more resistant to weathering and erosion than other rock types in the study area and therefore outcrops of Waxhaw granite occur as extensive boulder fields and large pavement exposures on ridges and higher topographic regions.

**Rock Type**

Granite from the Waxhaw pluton is white to light gray where fresh and light tan to dirty brown where weathered. It appears to occur in both a medium-crystalline and a fine-crystalline phase. The medium to medium-fine crystalline facies is a biotite granite with equigranular quartz and feldspar with finer crystalline biotite and
magnetite (Figure 16). Fine-crystalline Waxhaw granite resembles a feldspathic porphyry that is phenocryst rich (>50% phenocryst) in which the dominant phenocrysts are feldspar and quartz in a fine crystalline quartzo-feldspathic matrix. Biotite is present in fine-crystalline Waxhaw granite but is sparsely concentrated and difficult to see in hand specimen. Locally, the granite is deep red with a fine-crystalline, sugary texture but such occurrences in the Waxhaw granite are rare. Grain size across the entire pluton is highly variable but in general crystals never exceed 6 mm in length. The highest concentration of fine-crystalline granite occurs along the eastern flank of the pluton with a progressive increase in grain size to the west.

The Waxhaw granite has a mineral assemblage of Qtz + Mic + Ab (An4-8) + Bio + Mgn with lesser amounts of Ep ± Zoc ± Hem ± Rt ± Zrn ± Grt. Polysynthetic gridiron twinning is the distinguishing characteristic of microcline in thin section while plagioclase shows both Albite and Carlsbad twinning. Quartz percentage in the granite is between 35-45%, microcline 20-30%, plagioclase 10-25%, biotite 5-7%, and magnetite 1-3% by volume. The most distinguishing mineralogical feature of the Waxhaw granite is its high magnetite content. Almost all samples of Waxhaw granite encountered in the field are strongly magnetic and large magnetite crystals are usually recognizable in outcrop and hand specimen.

Boland (1996) noted that the Waxhaw granite is foliated where it comes in contact with the GHfz, however, quartz/biotite foliations have also been observed well within the granite body (Figure 2). Foliation in the granite is defined by aligned biotite and elongate quartz grains and foliation intensity varies from moderately
Figure 16. Foliated Waxhaw granite in a fresh boulder along Steele Hill Rd. approximately 5.5 km north of Van Wyck, South Carolina. Foliation in this boulder is defined by aligned biotite and its orientation is roughly parallel to the Swiss Army knife. Photograph courtesy of Jim Hibbard
developed to locally unfoliated. Where it comes in contact with the GHfz along the northern edge of the pluton, the granite has been deformed into lenses and pods of granitic gneiss.

Contacts

The Waxhaw granite is surrounded by low-grade volcanic rocks to the north and east and medium grade ductily deformed amphibolite to the southwest near the town of Van Wyck, South Carolina. The contact between the granite and surrounding rocks was not observed in outcrop but it is inferred to be an intrusive contact because a well developed zone of contact metamorphism extends into volcanic strata of the Twelve Mile sequence. Along its northern boundary the contact between the Waxhaw pluton and country rock is concordant with the strike of regional foliations, however, its eastern boundary cross-cuts the regional structural grain further supporting an intrusive relationship between the granite and the Twelve Mile sequence.

Previous mapping shows the southern boundary of the GHfz is defined by the trace of the Waxhaw pluton (Boland 1996). Sheared schists and pods of sheared out granitic gneiss along the northern contact of the pluton suggest the Waxhaw granite is overprinted by the GHfz. Previous mapping (Boland 1996) also indicates the Waxhaw pluton is in contact with a south trending branch of the GHfz (the DBsz) along its eastern flank. The results of this study indicate, however, that the Waxhaw granite is in direct contact with mildly deformed felsic volcanics and undeformed
gabbros along its eastern flank and there does not appear to be a contact aureole that has developed in volcanic rocks along its eastern flank.

**Age and Correlation**

Geochronologic data on the Waxhaw granite is limited to a Rb-Sr whole rock age of 495±21 Ma, which is interpreted as its crystallization age (Fullagar 1981). The significance of this age is questionable, however, because 1) Rb is known to be highly mobile in deformed rocks, and 2) the interpreted crystallization age of the Waxhaw granite does not fit the regional tectonic history of the Carolina zone. The Waxhaw granite is traditionally interpreted to be a component of the infrastructural Charlotte terrane. However, Cambrian and Neoproterozoic plutons in the Charlotte terrane are intensely deformed whereas the Waxhaw granite is not. Furthermore, the Waxhaw granite is bounded to the east by strata belonging to the Carolina terrane while to the west it is in contact with and contains xenoliths of amphibolite that belong to the Charlotte terrane (Boland, 1996).

**Hancock Granite (Hg)**

**Distribution**

A kilometer northwest of the Waxhaw pluton is a small, lenticular granitoid body that is informally named the Hancock granite after Hancock, South Carolina. The Hancock granite forms a long, narrow lens of granite that trends approximately N50°E, is 5.5 km long and between 0.2 and 0.5 km wide. The best exposures of the Hancock granite can be seen along the CSX rail line southwest of Hwy 521 and in an
unnamed branch of Twelve Mile Creek just east of Cross Creek Estates Rd. Because the Hancock granite is more resistant to weathering and erosion, it commonly forms ridges and pronounced topography where it occurs within the study area.

Rock Type

The Hancock granite is white to light gray and less commonly red, medium-to coarse-crystalline biotite metagranite. Unlike the Waxhaw granite, the Hancock granite is consistently medium-crystalline from outcrop to outcrop. Quartz and feldspar are equigranular, ranging in size from 4 to 7 mm. Biotite concentrations vary considerably with some outcrops containing ≥10% biotite while others contain no observable biotite. The mineralogy of the Hancock granite appears to be similar to the Waxhaw granite with the exception of magnetite, which is scarce in the Hancock granite. The Hancock granite bears a weakly developed biotite/quartz ribbon foliation that is parallel to the strike of regional foliations. This foliation can be difficult to detect in outcrops where biotite concentrations are low.

Contacts

The Hancock granite is completely surrounded by fine-grained felsic and mafic volcanic rocks belonging to the Twelve Mile volcanic unit. Though the contact between the Twelve Mile volcanic unit and the Hancock granite is not exposed in any single outcrop, it is interpreted to be intrusive based on the contact relationship between the Twelve Mile sequence and the Waxhaw granite, which is a similar in composition and morphology to the Hancock granite.
Age and Correlation

The Hancock granite is of similar composition to the Waxhaw granite, yet it is separated from the Waxhaw granite by a kilometer of deformed and metamorphosed volcanic rock. It is unclear whether the Hancock and Waxhaw granites are related, in part, because the Waxhaw granite is fine- to medium-crystalline and highly magnetic whereas the Hancock granite is coarse-crystalline and non-magnetic. To date, no contact aureole has been identified around the Hancock granite. Based on the geologic evidence, there are two possible origins for the Hancock granite: 1) it is a late phase of the Waxhaw granite in which magnetite was not able to crystallize; 2) the Hancock granite is a large stock that is unrelated to the Waxhaw granite. The Hancock granite is undated but is here considered to be related to the Waxhaw granite because of its proximity and similarity to the larger pluton.

Gabbro (Mgb)

Distribution

Several small bodies of gabbro appear to intrude felsic volcanic rocks of the Twelve Mile sequence in the southeastern half of the study area. These gabbros form elliptical pods that are oriented sub-parallel to regional foliation, however some appear to cross-cut the structural grain of the country rock. Outcrop area of gabbro ranges from 0.5 to 1.0 km in length. Gabbros in the study area are undeformed and contain no discernible foliation.
Rock Type

Gabbro in the study area is medium- to coarse-crystalline and ranges in color from dark green to black and has a salt and pepper texture. Gabbro usually occurs in small outcrops or as spheroidally weathered boulders that are non-magnetic, medium to coarse crystalline, and slightly denser than diabase. They have well preserved random, interlocking textures common to igneous rocks with a mineral assemblage of En + Dio + Lab(An57-62) ± Chl ± Opq. Pyroxene and plagioclase in gabbro are euhedral in thin section with compositional zoning common in plagioclase (Figure 17). The mineralogy of gabbro area is for the most part unaltered, however, minerals in one sample of gabbro appear to have been slightly altered by deuteric processes.

Contacts

A contact between gabbroic bodies and felsic volcanic rocks in the Waxhaw area was not recognized in outcrop. However, this study has found no evidence to suggest the contact between the gabbros and surrounding volcanic rocks is other than intrusive.

Age and Correlation

Currently, there are no age data pertaining to gabbro in the study area. Gabbro around Waxhaw is fresh, contains no discernible foliation in either hand specimen or thin section, and has a well preserved igneous mineralogy. These observations suggest that gabbro in the Waxhaw area are Mesozoic in age and related to Mesozoic rifting.
Figure 17. Photomicrograph of zoned plagioclase from a sample of gabbro collected from the study area. First order yellow minerals are enstatite while second order red and blue minerals are diopside. Slight alteration of minerals in this section is believed to be the result of deuteric alteration. Gabbro and diabase are the only rock types in the study area that are unaffected by the regional deformation. Sample located in an unnamed branch of Waxhaw Creek approximately 400 m west of S.R. 1126.
**Diabase (Jd)**

**Distribution**

Diabase occurs throughout the study area as small dikes that range anywhere from 0.5 to 3 m wide and that have an average length of 1-2 km. Diabase is typically exposed as a series of small cobbles or boulders in creeks and along hill slopes. The trend for regional dikes ranges from N-S to N40ºW with an average strike of N25ºW. The greatest concentration of diabase occurs along the northern margin of the Waxhaw pluton where diabase has been down cut by local creeks to form deep stream drainages. Diabase mapped in the Waxhaw vicinity appears to trend undeflected across the GHfz.

**Rock Type**

Diabase dikes are dark gray to black in color and most commonly occur as spheroidally weathered boulders with a characteristic rust brown weathering rind. They are composed of medium-fine crystalline plagioclase, pyroxene, and minor magnetite and they have a sub-ophitic texture. Grain size for diabase in the study area is uniform.

**Contacts**

Contacts between regional diabase and the surrounding country rock have been shown to be intrusive in the field area (Schroeder, 1987). Diabase dikes in the Waxhaw area cross-cut primary layering and foliations of the volcanic and sedimentary units described above.
Age and Correlation

Diabase dikes within the vicinity of Waxhaw are identical to other occurrences of diabase throughout the southern Appalachians and are here correlated with Mesozoic diabase dikes of Carpenter (1982) and Ragland (1991) and are interpreted to be approximately 200 Ma in age.
Structural Geology

General Statement

The original focus of this study was the structure and kinematics of the Davis Branch shear zone (DBsz) and its relationship to the GHfz. However, detailed mapping has failed to identify the DBsz. Neither strain analysis nor structural features in its reported location supports the existence of such a ductile feature. This chapter will therefore focus on the fabrics and structures that characterize the GHfz and surrounding region. Structural data presented in this chapter are the basis for new interpretations of the kinematic history of the GHfz and they add greater detail to the structural and tectonic history of the Carolina zone.

The emphasis of this chapter is the structural features that resulted from deformation within the GHfz. Three deformational events are recorded in the rocks within the GHfz and surrounding volcanic sequences (Figure 18). The most prominent is a main deformational event, \(D_M\), which is recorded in both the Marvin and Twelve Mile sequence as well as the Waxhaw and Hancock granites. Rocks in the GHfz also record a late deformational event, \(D_L\), which is only observed in rocks within the GHfz. Brittle faulting \((D_f)\) affects many of the rocks locally but it does not play a significant role in the deformational history of the GHfz.

Though structures associated with the GHfz define a single structural domain within the study area, the GHfz as well as structures associated with the GHfz appear to bend around a large flexure in the northeastern corner of the study area near the town of Weddington, North Carolina. This feature, herein termed the Weddington
Figure 18. Deformational events in the Waxhaw area. $D_M$-main deformation, $D_L$-late deformation, $D_f$-brittle deformation.
flexure, is responsible for broadly warping structures in the Carolina zone in that region, including the GHfz.

The following discussion will describe the deformational events that affect the study area starting with the main deformational event followed by the late event. Structural elements associated with $D_M$ are labeled $S_M$ (main fabric), $F_M$ (major first generation folds), and $L_M$ (mineral lineation). Structures associated with the later deformational phase are labeled $S_L$ (late fabric) and $F_L$ (major second generation folds). Sedimentary and volcanic bedding is labeled $S_0$. These sections will be followed by kinematic and strain analyses of rocks in the field area. Brittle faulting will be addressed at the end of the chapter. The Gold Hill and Silver Hill faults were not directly observed in the field, however, a few characteristic features were observed in their vicinity and will be described in the following sections.

Throughout this discussion, structural features will be described in terms of their principle axes in relation to the strain ellipsoid. Principal strain axes will be designated as follows: $X$ direction, parallel to the maximum extension direction; $Y$ direction, the intermediate stretch axis; and $Z$, the minimum extension direction. The respective principal planes of finite strain, $(XY, XZ, YZ)$ are also referred to in the following discussion.

**The Main Deformation ($D_M$)**

Structures attributed to $D_M$ are the most obvious and occur in both the Twelve Mile and Marvin sequences as well as the Waxhaw and Hancock granites. They are manifest as a regional cleavage/foliation, folds in sedimentary and volcanic bedding,
a lineation, asymmetric porphyroclasts, and boudinage of volcanic layering and mineral veins. The intensity of $D_M$ structures is greatest within the GHfz and less intense in rocks outside of the fault zone.

**Foliations ($S_M$)**

$D_M$ is responsible for the dominant foliation, $S_M$, which occurs in virtually all rock units in the study area except gabbro and diabase. In the Twelve Mile and Marvin sequences, $S_M$ occurs as a prominent spaced cleavage in most volcaniclastic rocks and as a zonal to continuous cleavage in phyllite and schist according to the foliation classification of Passchier and Trouw (1996). $S_M$ strikes roughly N60°E and dips steeply to the northwest and southeast (Figure 19). However, in the northeastern portion of the field area $S_M$ cleavages in the Twelve Mile sedimentary unit shift orientation to the north-northeast and exhibit an average strike of N25°E. $S_M$ is best developed in sedimentary rocks in the Twelve Mile and Marvin sequences where it commonly refracts through coarser- and finer-grained laminae in single layers of stratified mudstone and sandstone, thus indicating graded bedding and stratigraphic younging direction. In coarser grained crystal tuffs and porphyries, $S_M$ is less developed and is usually only recognizable in weathered outcrops.

$S_M$ in volcaniclastic rocks is defined by aligned phyllosilicates, cleavage seams composed of dark clay minerals and opaque minerals, aligned crystal and pumice lapilli, and stretched and flattened quartz phenocrysts. $S_M$ is best described as a disjunctive cleavage (e.g. Passchier and Trouw 1996) however, in finer grained sediments and volcanic tuffs, $S_M$ is a continuous cleavage. Cleavage domains in
rocks with disjunctive and zonal cleavages are typically smooth in shape, range from <0.1 to 0.4 mm in width, and vary from parallel to anastomosing depending on clasts concentration in the rock. Microlithons range in width from 0.4 to 2.0 mm and typically contain randomly oriented fabrics that grade sharply into cleavage domains.

The Waxhaw and Hancock granites contain a steep dipping foliation with similar orientations to $S_M$ cleavages in surrounding volcanic and sedimentary rocks (Figure 20). For this reason, foliations in both granites are designated $S_M$. In the Waxhaw and Hancock granites, $S_M$ is defined by aligned biotite and stretched quartz. Foliation is generally better developed in coarser grained, biotite-rich, phases of the Waxhaw and Hancock granites. Along the eastern edge of the Waxhaw pluton $S_M$ is characterized by a disjunctive foliation that becomes increasingly continuous towards the central and western portions of the pluton. Along the northwestern contact between the Waxhaw granite and volcanic tuffs of the Twelve Mile sequence, the Waxhaw granite has been deformed into a granitic gneiss with $S_M$ defined by 1 to 3 cm wide biotite and quartzofeldspathic gneissic layers.

$S_M$ is predominantly defined by cleavage seams composed of insoluble minerals that truncate primary igneous minerals in volcanic tuffs and offset bedding in stratified sedimentary rocks. These are all features of pressure solution, which is interpreted as the dominant deformation mechanism in rocks affected by $D_M$. Ductile deformation features are rare but present in rocks with $S_M$ foliations. In some plutonic and felsic volcanic rocks, quartz grains exhibit deformational features such as sweeping undulose extinction and sub-grain formation. A few quartz grains appear to have undergone dynamic crystallization by grain boundary migration and sub-grain
**Figure 19.** Southern hemisphere equal area stereonet of the contoured plot of poles to $S_M$ cleavages for the Twelve Mile and Marvin sequences.

**Figure 20.** Southern hemisphere equal area stereonet of the contoured plot of poles to $S_M$ foliations in the Waxhaw and Hancock granites.
rotation. Feldspars are deformed primarily by brittle fracturing with only a few showing minor evidence of ductile deformation, exhibiting undulose extinction and bent deformation twins. The combination of pressure solution and low-grade crystal plastic deformation during \( S_M \) formation indicates that temperatures during deformation ranged between 350° and 400°C (e.g. Passchier and Trouw, 1996).

Folds (\( F_M \))

Macroscale and mesoscale folds deform \( S_0 \) in the volcanic and sedimentary strata of the Twelve Mile and Marvin sequences. \( F_M \) folds in the Twelve Mile sequence are associated with the northwest limb of the Troy anticlinorium, which trends through the southeastern portion of the field area. No \( F_M \) folds are observed in the plutonic rocks within the field area.

\( F_M \) folds are predominantly tight to closed, symmetric, and upright to slightly inclined with interlimb angles that are estimated between 80°-60° (Figure 21). Fold hinges are sub-rounded and axial planes trend roughly N50°-60°E and dip steeply to the northwest and southeast. The majority of \( F_M \) folds verge to the southeast. In many outcrops there are also open to tightly folded quartz and epidote veins that cross-cut both \( S_0 \) and \( S_M \) at low angles but have axial planes that are roughly parallel to the axial planes of other \( F_M \) folds. In several outcrops, \( S_M \) is parallel to the axial planes of folded bedding (\( S_0 \)), which indicates that \( S_M \) cleavages are axial planar to \( F_M \) folds.

Intersection lineations between \( S_0 \) and \( S_M \) in stratified rocks as well as hinges measured from minor \( F_M \) folds in outcrop reveal that \( F_M \) folds plunge gently and
Figure 21. Northeast plunging, southeast verging folds in bedded crystal tuffs in the Twelve Mile sequence. View to the southwest. Outcrop located on Blythe Creek approximately 300 m north of Howie Mine Rd. Rock hammer for scale.
**Figure 22.** Top: Intersections between $S_0$ and $S_M$ with contoured plot for best fit great circle. Bottom: Southern hemisphere stereonet of $F_M$ fold axes.
moderately to the northeast and southwest (Figure 22). On a stereonet, both intersection lineations and fold hinges plot along a girdle that dips steeply to the northwest. Outcrop scale folds exhibit similar orientations and style as higher order parasitic folds. At one outcrop along East Fork Twelve Mile Creek near Billy Howie Road, FM folds refold soft sediment folds in the Twelve Mile sedimentary sequence, locally producing a variety of interference patterns on the same outcrop face.

**Lineation (Lm)**

A prominent lineation, Lm, is present on SM in virtually all deformed rocks in the GHfz with the exception of the Hancock and Waxhaw granites. Lm is subtle and difficult to detect in fine-grained sediments and tuffs but in coarser-grained crystal tuffs and phyllites it is well developed and easily recognized. Lm is defined by 1) preferentially aligned clusters of white mica, chlorite, and quartz; and 2) elongated mineral grains and volcanic clasts and lapilli. Because both of these lineations are mutually parallel, they have been plotted on a single stereonet (Figure 23).

Intersection lineations between S₀ and SM were also measured, however, they have been discussed in the Fm fold section (above).

Lm lineations have an average trend of N60°E/S60°W and vary in plunge from sub-horizontal to sub-vertical. Sub-horizontal (<30°) Lm lineations are the most common and occur predominantly in rocks that are located at distances greater than one kilometer from the GHf. The greatest concentration of down-dip lineations (>60°) are found in close proximity to the inferred trace of the GHf and in isolated zones of highly deformed phyllite. Lm is interpreted as a stretching lineation because
Figure 23. Southern hemisphere equal area stereonet of $L_M$ lineations.
1) the trend of $L_M$ is sub-parallel to boudinaged volcanic layers in the Twelve Mile sequence (see below); 2) epidote, mica, and quartz in thin sections are extended parallel to $L_M$ in the plane of foliation (Figure 25a).

**Boudins**

Boudins were observed locally in outcrop throughout the field area as well as in thin section. Mesoscale boudinage is represented by stretched volcanic layering (Figure 24) and mineral veins that are extended and segmented parallel to $S_M$. Boudins have been observed on the $XZ$- as well as the $YZ$-faces of separate outcrops; however, exposure has never been sufficient to view the $XZ$- and $YZ$-faces at the same outcrop. In thin section, boudinaged quartz and epidote occur parallel to lineation ($L_M$) in the plane of $S_M$ foliation (parallel to the $XZ$-plane) as well as perpendicular to lineation (parallel to the $YZ$-plane) (Figure 25a & b). Minerals defining $S_M$ wrap around individual boudins, which, together with the observed parallelism of boudinage with $S_M$, implies that extension of the boudinaged layers and minerals took place during $D_M$. The apparent presence of two mutually perpendicular extension directions implies that flattening in the style of chocolate-tablet-boudinage (Ramsey and Huber, 1983, p. 65) occurred during $D_M$.

**Winged Objects**

Objects with symmetric and asymmetric wings are present, but uncommon, in rocks that have been affected by $D_M$ and are observable in both outcrop and thin section. Winged objects typically occur in highly deformed crystal tuffs,
Figure 24. Photograph of boudinaged volcanic layering parallel to $S_M$ in the Twelve Mile sequence. Note: cuspid boudin in-fill fold in the center of the photo below the rock hammer. Outcrop located on Davis Branch Creek near the southern edge of the study area; approximately 150 m from S.R. 1113. Rock hammer for scale.
Figure 25a. Photomicrograph of a quartz grain that has been boudinaged parallel to lineation L_M.

Figure 25b. Photomicrograph of quartz grains that have been boudinaged perpendicular to lineation L_M.
volcaniclastic rocks, and phyllites. The origin of these objects includes phenocrysts in crystal tuffs and volcanic clasts in sedimentary rocks. The cores of these objects range in diameter from 0.5 to >2.0 cm and are most commonly composed of quartz and saussuritized feldspar. Wings are typically composed of aggregates of quartz, mica, and epidote, and extend in the direction of $S_M$. Porphyroclasts with asymmetric wings occur as both sigma- and delta-shaped clasts (Figure 26) (Passchier and Simpson, 1986). Within the GHfz, wings on asymmetric porphyroclasts depict a consistent asymmetry at the scale of outcrop, hand specimen, and thin section. However, the orientation of kinematic indicators with respect to the stretching lineation ($L_M$) differs across the GHfz. This observation will be addressed in more detail below in the Kinematics section of this chapter along with the kinematic interpretation of asymmetric clasts and fabrics.

**The Late Deformation ($D_L$)**

The deformation that overprints $D_M$ is designated $D_L$, the late deformation. Structures associated with $D_L$ are not as well developed as $D_M$ and only occur within the GHfz. $D_L$ structures include a late foliation ($S_L$), intrafolial folds ($F_L$), and a discrete shear band foliation ($S_{sb}$) that overprints $S_M$.

**Late Foliation ($S_L$)**

$S_L$ is a weak to moderately developed, spaced cleavage that overprints $S_M$ and occurs locally in outcrop and thin-section. $S_L$ is defined primarily by aligned phyllosilicates and cleavage seams that locally define a crenulation cleavage. Where
Figure 26. Photomicrograph of two asymmetric winged porphyroclasts in phyllite from the Twelve Mile sequence in the GHfz. Photograph taken parallel to $L_M$ stretching lineation. Sample location on Twelve Mile Creek approximately 30 m east of State Highway S-29-93.
it occurs, $S_L$ is always weaker than $S_M$. In outcrop, $S_L$ has the appearance of a weakly
developed slaty-cleavage that is typically oriented between 15° and 40° clockwise to
$S_M$. In thin-section, $S_L$ is observed as a mica-foliation that “passively” overprints $S_M$
(Figure 27) and as a discrete crenulation cleavage that transposes $S_M$ (Figure 28). It is
unclear whether $S_L$ developed axial planar to $F_L$, however, Schroeder (1987) reported
that a late cleavage had developed axial planar to intrafolial wrench and drag folds in
the Howie Mine.

$S_{sb}$ comprises a series of locally developed shear bands that overprint $S_M$.
They are interpreted as C’-shear band cleavages (e.g. Passchier and Trouw, 1991) and
are characterized by narrow bands of sheared and aligned minerals that are oriented
slightly oblique to the bulk flow plane (i.e. shear plane) in a shear zone. Shear bands
are believed to form during general non-coaxial flow (e.g. Hanmer and Passchier,
1991). $S_{sb}$ occurs as both sporadically and homogeneously developed, non-
penetrative foliations that offset $S_M$ regional cleavages in volcanic rocks of the
Twelve Mile sequence and sedimentary rocks in the Marvin sequence. Foliations
offset by $S_{sb}$ consistently display dextral displacement on the XZ-face of outcrop and
thin-section.

Late Folds ($F_L$)

Mesoscale and microscale folds associated with $D_L$ occur locally throughout
the field area and are designated $F_L$. $F_L$ folds both $S_M$ and $S_0$ and have only been
observed in the Twelve Mile and Marvin sequences. $F_L$ folds are predominantly
asymmetric, tight to isoclinal with interlimb angles between 10° and
Figure 27. Photomicrograph of muscovite schist from the Twelve Mile sequence that bears both \( S_M \) (red-line) and \( S_L \) (blue-line). Micas that define \( S_L \) appear to overgrow earlier micas associated with \( S_M \).

Figure 28. Photomicrograph of an \( S_L \) cleavage (diagonal) crenulating \( S_M \) cleavage (horizontal).
30°, recumbent, and similar (class 2) with rounded to sub-rounded hinges (Figure 29). 

$F_L$ axes trend approximately parallel to $F_M$ folds and plunge steeply to the northeast and southwest. Where viewed down plunge, $F_L$ folds consistently demonstrate clockwise vergence with a Z-geometry that implies dextral shear. However, it is possible that $F_L$ folds are parasitic to macroscopic folds and therefore bear no kinematic significance.

**The Gold Hill fault zone and the Weddington Flexure**

The GHfz in central North Carolina constitutes all deformed volcaniclastic strata that lie between the GHf and the Silver Hill fault (SHf). Previous workers have misinterpreted the GHfz as a wide zone of ductile strain consisting of recrystallized phyllite and schist that are devoid of primary features (e.g. Gibson and Huntsman, 1988, Boland, 1996). However, the results of this investigation as well as others associated with it (Standard, 2003, Hibbard et. al., 2003) indicate that rocks within the GHfz are commonly partially recrystallized and only rarely completely recrystallized to phyllite or schist.

In the High Rock Lake area, Standard (2003) noted that ductile deformation attributed to the GHfz is only observed in isolated outcrops of phyllite and along a continuous narrow zone on the western edge of the fault zone that he interpreted as the GHf. The remainder of the GHfz around High Rock Lake was affected by low temperature/low strain deformation. Pseudotachylite was also reported along the GHfz in the vicinity of Mt. Pleasant, North Carolina (Hibbard, pers. comm., 2003). Metamorphic conditions in the study area also indicate that deformation and
Figure 29. Photograph of an $F_L$ fold that folds $S_M$ in the volcanic unit of the Twelve Mile sequence within the GHfz. Outcrop located on Six Mile Creek approximately 0.8 km south of Jim Wilson Rd. Pencil for scale.
metamorphism took place at depths no deeper than 9 km in the crust, which is generally near the brittle-plastic transition for deformation (see Metamorphic Chapter). All of these observations indicate that the GHfz is a brittle-ductile fault zone rather than a ductile shear zone.

The GHfz is a major brittle-ductile fault zone within the Carolina zone that deforms all pre-Mesozoic rock units within the Waxhaw area. The fault zone in the Waxhaw area consists of a wide damage zone that is approximately 5 to 7 km wide and extends into the Marvin and Twelve Mile sequences. Previous studies interpreted the GHf as the western boundary of the fault zone; however, this study has found that deformation from the fault zone extends up to 1.5 km into the Marvin sequence (Figure 7). From the GHf southward, deformation attributed to the GHfz extends 4 to 5 km into the Twelve Mile sequence with no certain boundary along its margin. Therefore, the GHf does not define a clear boundary for the GHfz. These observations suggest that the GHfz consists of a wide footwall damage zone to the GHf in the Twelve Mile sequence and a narrow hanging wall damage zone in the Marvin sequence.

The orientations of the structural features are uniform throughout the region with the exception of fabrics and folds in the northeast corner of the study area. There, foliations and fold axes trend approximately N15-25ºE, whereas structures in the rest of the field area trend approximately N60ºE (Figure 2). North of this structural bend, the GHfz as well as other structures in the Albemarle sequence also consistently trend N15-30ºE; however, southwest of this feature, structures in the Carolina and Charlotte terranes consistently trend N60-80ºE. Thus, it appears that a
significant first-order flexure lies in the northeast corner of the study area. This flexure is informally named the Weddington flexure because it lies near the town of Weddington, North Carolina.

**The Gold Hill Fault**

The GHf is a zone of intense deformation and strain that is estimated to be about 10 to 50 m wide and dips steeply to the northwest. Although the fault was not observed in outcrop, the approximate trace of the GHf is inferred based on the following observations: 1) the noticeable change in lithology across its trace; 2) the truncation of previously mapped folds associated with the northwestern limb of the Troy anticlinorium along the GHf (Figure 2); 3) the noticeable increase in strain intensity (see Strain Analysis section, below) in rocks near and within the inferred trace of the GHf as well as an apparent change in the intensity of deformation across the fault; 4) the consistent shear sense displayed by asymmetric fabrics and fold in rocks observed near the inferred trace of the GHf (see kinematics section, below); 5) numerous meter wide quartz veins, along with fields and quarries with quartz float ranging in diameter from centimeters to tens of meters, characterize areas proximal to the fault. These quartz veins may reflect the dissolution of silica and influx of fluids along fractures during deformation along the GHf (e.g. Schroeder et al. 1988).

**The Silver Hill Fault**

Previous workers in central North Carolina recognize the Silver Hill fault (SHf) as the southeastern boundary to the Gold Hill shear zone in that region (e.g.
Stromquist et al., 1969, Stromquist and Henderson, 1985). As with the GHf, the SHf is unexposed in the study area. Furthermore, its continuation into the study area from central North Carolina is conjectural based on geologic evidence gathered during this study. In central North Carolina, the trace of the SHf is inferred based on an abrupt change in deformational style from east to west across a relatively narrow zone (Standard, 2003, Hibbard, pers. comm., 2003). Foliation and fold intensity throughout much of the Twelve Mile sequence appears to only gradually increase northward into the GHfz with no sharp contrasts in deformation between the rocks in the southern and central portions of the study area. Strain analysis (See Strain Analysis section) indicates that rocks located outside the GHfz as inferred by Boland (1996) experienced similar deformational environments as rocks situated well within the fault zone.

Interestingly, much of the Waxhaw granite has been intensely sheared along its northwestern margin where it has the appearance of a granitic gneiss. These granitic gneisses and surrounding sheared volcanic rock possess structures indicative of a high strain zone (e.g. asymmetric fabrics, winged objects), which suggests that a shear zone is in contact with the northwestern flank of the Waxhaw granite. This shear zone may constitute the SHf. A likely explanation for the sporadic appearance of the SHf may lie in the difficulty to detect increases in strain intensity in the fine-grained volcanic and sedimentary rocks that constitute the majority of the study area northeast of the Waxhaw granite.
Timing of Deformation

The timing of motion and deformation within the GHfz has been the focus of several previous studies. Boland and Dallmeyer (1997) used $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages from amphibolite located just to the southwest of the study to constrain the timing of deformation in the GHfz to ca. 400-368 Ma. Hibbard et al. (2003) and Standard (2003) showed that Late Ordovician folding in the Albemarle sequence is tied to deformation induced by the GHfz in that region. Lavallee (2003) and Hames (pers. comm., 2005) obtained $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of ca. 360-320 Ma from white micas in the GHfz in the High Rock Lake and Mt. Pleasant regions, as well as from the GHfz in the vicinity of Catawba, South Carolina, which is located approximately 15 miles west-southwest of Waxhaw, North Carolina. Thus, a series of cooling ages spanning the early, middle, and late Paleozoic have been obtained from the GHfz.

The timing of $D_M$ in the Waxhaw study area is best constrained by the emplacement of the Waxhaw granite. Outcrop patterns of the Waxhaw granite suggest that it is both concordant with, as well as cross-cuts, $S_M$. Furthermore, most outcrops of the Waxhaw granite bear a tectonic foliation that is parallel to $S_M$ in surrounding units. Metamorphic porphyroblasts of biotite and andalusite that grew on the fringe of the Waxhaw metamorphic aureole (see Metamorphic chapter) appear to overgrow $S_M$ foliations in thin-section; yet these same porphyroblasts are slightly deformed and have weakly developed strain shadows on mineral surfaces perpendicular to $S_M$ (Figure 30). However, andalusite porphyroblasts within 100 m of the Waxhaw granite possess randomly oriented inclusions. Finally, outcrop of intermediate tuffs and basalts within the Waxhaw contact aureole contain
Figure 30. Photomicrograph of a metamorphic biotite from a semi-pelitic schist located in the contact aureole of the Waxhaw pluton. The biotite appears to overgrow $S_M$ yet is deformed by strain caps around its top and bottom edges and bears small strain shadows on surfaces perpendicular to $S_M$. These combined textures are characteristic of late syn-kinematic mineral growth.
metamorphic amphiboles that both cross-cut and lie parallel to $S_M$ cleavages. These
geologic data lead to the interpretation that the emplacement of the Waxhaw granite
was synkinematic with respect to $D_M$.

If this interpretation is correct then the timing of $D_M$ is constrained by the age
of the Waxhaw granite, which is ca. 500 Ma; however, this age is suspect for the
following reasons: 1) the unreliable nature of Rb-Sr whole rock dating on deformed
rocks; 2) the age does not fit any known model for southern Appalachian tectonism.
Depending on the error in the Rb-Sr age data, the upper limit for $D_M$ may be much
younger or older than the current age prediction. Mesozoic diabase dikes trend
undeflected across both the GHfz and the GHf. Therefore, $D_M$ and $D_L$ must be older
than ca. 200 Ma (Ragland, 1991). This constrains $D_M$ and $D_L$ to the Paleozoic and/or
late-Neoproterozoic.

Further evidence suggests that $D_M$ and $D_L$ may be closely linked or perhaps
overlap in time. The following observations are important to this interpretation: 1)
$F_L$ fold axes are roughly parallel with earlier $L_M$ lineations suggesting both features
formed during the same deformatonal regime. 2) The sense of shear determined for
both $D_M$ and $D_L$ structures consistently demonstrate southeast vergent dextral reverse-
oblique shear. 3) $S_{sb}$, a component of $D_L$ represents C’-shear band cleavages of
Passchier and Trouw (1996). Studies indicate that shear band cleavages generally
develop late during shear zone activity (e.g. Passchier and Trouw, 1996) thus $S_{sb}$ may
represent a ‘last-gasp’ of $D_M$/$D_L$ deformation. Minerals that define $S_{sb}$ formed at
similar metamorphic conditions as $S_M$ and $S_L$ which supports this interpretation. 4) In
this same vein, the structures that comprise $D_M$ and $D_L$ all formed under similar
metamorphic conditions (see Metamorphic chapter). These observations support the interpretation that $D_M$ and $D_L$ are phases of a single on-going deformation.

**Strain Analysis**

A study was carried out during this investigation with the intent to characterize the finite strain state of the GHfz. To this end, 10 samples consisting of volcanioclastic tuffs and conglomerates were collected, oriented, and analyzed using DePaor’s method of strain analysis (1988) and log $Rf/\Phi$ charts (e.g. Ramsay and Huber, 1983, Lisle, 1985) to determine the following: 1) the extent of deformation sustained by rocks in the GHfz and any strain gradients that may exist therein; 2) the southern boundary of the GHfz in terms of accumulated finite strain since the physical boundary (e.g. the Silver Hill fault) is poorly exposed; 3) if rocks within the proposed DBsz have accumulated high finite strain that is indicative of a ductile shear zone.

**Method and Assumptions**

Most methods for strain calculation are dependent upon the geometry of physical objects within the deformed host rock that act as passive markers during deformation. These ‘passive’ objects are commonly referred to as *strain markers* and can vary from metamorphic porphyroblasts to sedimentary clasts to xenoliths. The ultimate goal of finite strain analysis is to accurately gauge the relative intensity and orientation of the principal strains within a deformed rock. The most elegant means of representing strain is through the use of the finite strain ellipsoid. Because the
finite strain ellipsoid represents the total strain that has accumulated in a deformed rock, its shape can be derived arithmetically from the shapes and orientations of strain markers in a deformed rock.

The shape of a strain marker after a given strain is dependent upon four factors: 1) the initial shape of the marker ($R_i$), 2) the shape of the strain ellipsoid ($R_s$), 3) the orientation of the strain ellipsoid ($\Phi$), 4) the marker’s orientation with respect to the strain ellipsoid ($\Phi'$). Heterogeneous volume loss during deformation can also affect the final shape of the strain ellipsoid. To effectively characterize the finite strain state of a deformed rock, the orientation ($\Phi$) and principal elongations ($e_1$, $e_2$, $e_3$) of the finite strain ellipsoid are required. Unfortunately in most cases it is difficult to determine either the principal elongations or the orientation of the strain ellipsoid from a given sample directly. The only data that can be obtained from a deformed rock is the orientation ($\Phi'$) of the long axis of each strain marker with respect to a given direction and the principal stretches that correspond to the major and minor axes for each strain marker ($S_1$, $S_2$, $S_3$) where $S_n = (1 + e_n)$. The final ellipticity ($R_f$) of a strain marker is derived from the ratio of its principal stretches:

$$R_f = \frac{S_a}{S_b}$$

; where $S_a$ corresponds to the major axis and $S_b$ is the minor axis

From these data, $R_s$ can be calculated through a series of arithmetic and statistical calculations that comprise the $R_f/\Phi$ technique of Ramsay (1967) and Robin (1977).

The $R_f/\Phi$ technique relies on a set of core assumptions: 1) the deformation is homogeneous, 2) the initial orientations of the long axes of the strain markers were randomly distributed, 3) there is little to no competency contrast between the strain
markers and the surrounding groundmass during deformation. For a sample in this study to be considered suitable for strain analysis, further assumptions had to be made: 4) the XY-plane of the strain ellipsoid coincides with the orientation of the tectonic cleavage ($S_M$) in each sample, 5) strain measured in each rock sample is interpreted to be the finite strain accumulation during $D_M$ and $D_L$ events. Furthermore, even though strain accumulation in a shear zone is innately heterogeneous, at the scale of each sample finite strain accumulation is considered homogeneous.

One of the main issues that must be taken into consideration when performing strain analysis in low-grade metamorphic terranes is the role of pressure-solution during deformation. Deformation induced by pressure-solution involves removal and redeposition of material along discrete surfaces (e.g. Rutter, 1983, Tada et al., 1987, Renard et al. 1999) and is therefore inherently heterogeneous. However, it has been shown that pressure-solution does not significantly interfere with $Rf/\Phi$ methods for strain calculation (Onasch, 1984). Pressure solution can only significantly affect $Rs$ when either dissolution seams are spaced in intervals greater than the average width of the strain markers or if the displacement across a dissolution surface exceeds the minimum width of the strain marker (Onasch, 1984). Displacements across pressure-solution seams in the samples chosen for this study are relatively small and most strain markers are in contact with at least one dissolution surface. Therefore, values for $Rs$ and $\Phi$ are considered accurate even though there appears to be some manipulation of the strain markers by pressure-solution.
For the purposes of this study, ten samples of deformed volcaniclastic tuffs and conglomerates were oriented and cut parallel to their XY- and XZ-faces. The hyperbolic stereonet method of DePaor (1988), a derivative of the $R_f/\Phi$ technique, was used on all samples with exception of two, W-9 and C-209. Those two samples were analyzed using the log $R_f/\Phi$ charts of Lisle (1985) in fulfillment of a class assignment during the Spring 2004 semester. The hyperbolic stereonet is preferred over the more classical log $R_f/\Phi$ charts because the stereonet is not as data intensive requiring only a minimum of 16 data points for accurate analysis as opposed to 40 for the log $R_f/\Phi$ charts. The hyperbolic stereonet is therefore a faster and more efficient method than log $R_f/\Phi$ charts when analyzing a large number of samples. The $R_f$ and $\Phi$ of 20 to 40 deformed clasts and phenocrysts were measured on the XY- and XZ-face of each sample in order to determine the $R_s$ and $\Phi$ of the strain ellipsoid. $R_{xy}$ and $R_{xz}$ were determined by the techniques above while $R_{yz}$ was determined by calculation. K-values were calculated arithmetically using $R_{xy}$ and $R_{yz}$ of each strain ellipse.

Results

The results for each analysis are listed in Table 1. Data points corresponding to the $R_{xy}$ and $R_{yz}$ faces of the strain ellipsoid are plotted on a Flinn graph (Figure 31) and the location and shape of the strain ellipse for each sample is plotted on a general geologic map of the study area (Figure 32). The results of the strain analyses indicate that the greatest accumulation of finite strain occurred in rocks within the immediate vicinity of the GHf. Elsewhere in the study area, rocks sustained lower
values of finite strain. The orientation of the X-axes of the strain ellipsoids (Figure 33) do not vary significantly from the orientations of fold axes and stretching lineations ($L_M$) associated with $D_M$. The plunge of the principal axes for each strain ellipsoid varies with respect to its relative position to the GHf. Ellipsoids that lie in close proximity to the GHf have steep plunging X-axes while strain ellipsoids located more than one kilometer from the GHf have shallow plunges except in isolated regions of high strain within the GHfz. This variation in plunge is also consistent with the variation of $L_M$ plunges with respect to the GHf (see $L_M$ section).

One of the main reasons for conducting this analysis was to establish a southern boundary for the GHfz by determining a strain gradient between rocks deformed by the GHfz along its southern edge and comparing the strain in those rocks to rocks outside the southern boundary of the fault zone. However, the results from this analysis were unable to reveal a southern boundary for the GHfz because rocks within the fault zone appear to have experienced similar strain conditions as rocks outside the zone (Figure 32). This situation leads to the interpretation that strain in the footwall damage zone of the GHf likely grades southward into the surrounding volcanic strata with no certain southern boundary.

There appears to be a contrast in deformation between rocks situated near the GHf and rocks located at significant distances (>1 km) from the GHf. This is best observed in the Flinn graph where the majority of samples have apparent K-values that
Table 1. Geometry of Strain Ellipsoids for rocks in the study area

<table>
<thead>
<tr>
<th>Sample</th>
<th>distance from fault</th>
<th>$R_{xy}$</th>
<th>$R_{yz}$</th>
<th>$R_{xz}$</th>
<th>App. $K$</th>
<th>X-axis</th>
<th>XY-Plane</th>
</tr>
</thead>
<tbody>
<tr>
<td>C-276</td>
<td>10 m</td>
<td>3.20</td>
<td>2.20</td>
<td>7.00</td>
<td>1.83</td>
<td>75/235</td>
<td>252/82</td>
</tr>
<tr>
<td>C-329</td>
<td>650 m</td>
<td>2.21</td>
<td>1.31</td>
<td>2.90</td>
<td>3.90</td>
<td>77/279</td>
<td>258/85</td>
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<tr>
<td>C-40</td>
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<td>1.73</td>
<td>3.55</td>
<td>1.44</td>
<td>80/079</td>
<td>050/80</td>
</tr>
<tr>
<td>C-197</td>
<td>1.10 km</td>
<td>1.60</td>
<td>2.26</td>
<td>3.61</td>
<td>0.48</td>
<td>55/239</td>
<td>068/80</td>
</tr>
<tr>
<td>W-216</td>
<td>1.50 km</td>
<td>1.15</td>
<td>2.79</td>
<td>3.21</td>
<td>0.08</td>
<td>10/065</td>
<td>247/80</td>
</tr>
<tr>
<td>C-209</td>
<td>2.10 km</td>
<td>1.34</td>
<td>4.10</td>
<td>5.50</td>
<td>0.11</td>
<td>54/060</td>
<td>062/85</td>
</tr>
<tr>
<td>W-138</td>
<td>3.40 km</td>
<td>2.00</td>
<td>1.90</td>
<td>3.80</td>
<td>1.11</td>
<td>19/066</td>
<td>240/71</td>
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<tr>
<td>W-9</td>
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<td>1.60</td>
<td>2.06</td>
<td>3.30</td>
<td>0.57</td>
<td>07/066</td>
<td>060/70</td>
</tr>
<tr>
<td>W-96</td>
<td>9.20 km</td>
<td>1.90</td>
<td>1.87</td>
<td>3.56</td>
<td>1.03</td>
<td>30/231</td>
<td>036/80</td>
</tr>
</tbody>
</table>

*distances expressed in both meters and kilometers

Figure 31. Flinn graph (after Flinn, 1962) of $R_f/\Phi$ analyses from the GHfz. Samples near the GHf plot within the field of apparent constriction while samples collected at distances greater than 1 km of the GHf plot within the field of apparent flattening.
Figure 32. Map showing the shapes of XZ principal plane strain ellipses superimposed on a general geologic map of the Waxhaw study area.
Figure 33. Southern hemisphere equal area projection of the orientations of major principal axes to strain ellipsoids in the study area.
plot within, or very near, the field of apparent flattening. These data points represent samples that are located in the GHfz but more than one kilometer from the GHf as well as samples located outside the fault zone. The data cluster indicates that bulk strain in the GHfz is predominantly apparent flattening. This assessment is consistent with the earlier observation that flattening was a component of $D_M$ (see Boudins section).

A smaller cluster of three samples plots within the field of apparent constriction. These represent samples that are located within several tens to hundreds of meters of the GHf. The data here indicate that deformation along the GHf was predominantly constrictional provided that volume loss/gain is negligible. The consistency of the data however suggests that deformation on the GHf was constrictional. The occurrence of constrictional deformation in these rocks could be the result of a greater component of stretching involved with deformation along the GHf. In summary, the GHf experienced increased values of finite strain relative to the GHfz and was dominated by apparent constriction, whereas the majority of the GHfz experienced low values of finite strain dominated by apparent flattening.

An investigation was also carried out on a sample of lapilli tuff (W-9) collected from Davis Branch Creek with the intent to determine if it had undergone bulk strain indicative of a high strain zone (such as the DBsz). Prior to this analysis, there was some question concerning the existence of a ductile shear zone in the region of Davis Branch Creek. Geologic mapping around the reported location of the DBsz found that orientations of regional foliations do not trend parallel to a north-south striking shear zone, nor were any mylonitic fabrics or shear sense indicators found in
rocks within the reported trace of the DBsz. Strain analysis on lapilli tuff W-9 revealed that bulk strain in the sample did not significantly vary from strain experienced by rocks outside the GHfz. This observation, together with the other data presented, indicates the DBsz is not a ductile shear zone.

**Kinematic Analysis**

Kinematic indicators associated with the GHfz and the GHf occur in both the Marvin and Twelve Mile sequences. Kinematic indicators within high strain zones are typically described in relation to their orientation with respect to a stretching lineation and regional foliation. Early models of high strain zones depicted a deformational regime dominated by progressive simple shear (e.g. Ramsay and Graham, 1970) and predicted that kinematic indicators (e.g. asymmetric winged objects, strain sensitive fabrics) form parallel to a linear shape fabric (stretching lineation) and perpendicular to the planar shape fabric (foliation) in a shear zone. In these early models, the stretching lineation was taken to represent the direction of bulk transport (e.g. Simpson and Schmid, 1983, Hanmer and Passchier, 1991).

However, more recent studies show that within transpressional fault systems where the deformation has both a simple and pure shear component, sub-vertical and sub-horizontal stretching lineations can form as the result of a single deformation (Sanderson and Marchini, 1984, Robin and Cruden, 1994, Tikoff and Greene, 1997, Lin et al., 1998). In a transpressional shear zone, the orientation of the stretching lineation reflects the orientation of the principal axis of the finite strain ellipsoid.
rather than the bulk transport direction (Robin and Cruden, 1994, Tikoff and Greene, 1997).

Modern structural geologists use the *vorticity vector* to describe the kinematics of high strain zones. The vorticity vector can be thought of as an axis of rotation about which particles and material lines in a shear zone rotate during deformation. The direction particles rotate around the vorticity vector, either clockwise or counter-clockwise, defines the sense of shear of a high strain zone, that being dextral or sinistral respectively. Kinematic indicators will therefore be best developed on the plane that is perpendicular to the vorticity vector.

During progressive simple shear, the vorticity vector is perpendicular to the stretching lineation and parallel to the foliation in a shear zone. Hence, kinematic indicators form on the XZ-face (parallel to lineation and perpendicular to foliation) of a high strain zone in which progressive simple shear is the dominant deformation. However, in transpressional shear zones the vorticity vector can be parallel or oblique to both lineation and foliation. The former is characteristic of wrench-dominated transpressional shear zones dominated by pure-shear while the later is a characteristic of triclinic shear zones. One of the features of triclinic shear zones is that kinematic indicators form on planes both parallel and perpendicular to the stretching lineation.

Kinematic indicators in the GHfz are uncommon, predominantly ductile though a few brittle indicators have been observed (Figure 34), and define a sense of movement that is consistent throughout the study area. The geometry of kinematic indicators define two separate domains within the GHfz (Figure 35) in which the orientation of shear sense indicators differs relative to the stretching lineation. The
**Figure 34.** En-echelon pair of quartz filled extensional veins. The geometry of extensional vein arrays such as these consistently indicate dextral shear on planes parallel to layering (sub-horizontal). Outcrop location on Twelve Mile Creek approximately 0.5 south of Jim Wilson Rd. Foot for scale.
Figure 35. Map illustrating the two kinematic domains within the GHfz. The *light gray* region represents domain-Kw while the *dark gray* region represents domain-Ks. White regions are areas that do not contain kinematic indicators or rocks that have not been deformed by the GHfz.
larger of the two domains is labeled Kt (triclinic kinematics) and comprises the bulk of the GHfz while the smaller domain follows the trace of the GHf and is labeled Kw (wrench kinematics). Kinematic indicators have not been observed outside of the GHfz.

In domain Kt, kinematic indicators occur predominantly on the XZ- and YZ-faces of outcrops and thin-sections. The asymmetry of mica fish, winged porphyroclasts, and C’-shear bands on the XZ-face indicate that the dominant motion within the GHfz involved dextral shear (Figure 36a) with a component of northwest over southeast thrusting of the Marvin sequence over the Twelve Mile sequence. The presence of kinematically consistent shear sense indicators on the YZ-face of rocks in the domain Kt is indicative of general shear (pure shear + simple shear) and of a vorticity vector that is oblique to both foliation and lineation (Figure 36b). These kinematic data as well as geometry of kinematic indicators in Kt are characteristic of triclinic deformation.

Domain Kw consists of rocks that are deformed by the GHf. Rocks in this domain are typically highly strained and bear a noticeable steep plunging lineation. In Kw, kinematic indicators are found predominantly on a sub-horizontal surface, perpendicular to both foliation and lineation (YZ-face). Like domain Kt, the asymmetry of S-C fabrics, mica fish, winged clasts, and C’-shear bands indicate that dextral shear (Figure 37a) coupled with northwest over southeast thrusting was the dominant motion along the GHf. Kinematic indicators are not present on either the XY- and XZ-faces of rocks in Kw and all clasts observed on these faces are stretched and elongate in the vertical direction (Figure 37b).
Figure 36. Deformation features in domain-Ks. A) Shows delta-clasts indicating dextral shear parallel to lineation. B) Shows two sigma clasts indicating dextral shear perpendicular to lineation. C) Illustration of positions of A and B.
Figure 37. Deformation features from domain-Kw. A) Sigma-clast indicating dextral shear oriented perpendicular to lineation. B) Foliation surface from same outcrop as A showing a strong down-dip lineation. No kinematic indicators are present on the surface orthogonal to foliation. C) Illustration of positions A and B.
Though the GHfz can be divided into separate domains based on varying kinematics and style of deformation, models for transpression along with the geologic evidence from the study area suggest the bulk of the deformation that occurred in the GHfz is the result of a single deforming event. Within domain Kt, the orientation of the vorticity vector appears to range from near-perpendicular to oblique to the X-axis of the finite strain ellipsoid (stretching lineation), which for most of Kt is subhorizontal. However, the orientation of the vorticity vector may shift into parallelism with the X-axis of the strain ellipsoid within domain Kw while the plunge of both the vorticity vector and X-axis of the strain ellipsoid becomes increasingly subvertical. Thus, vertical stretching lineations with shear sense indicators subparallel to a horizontal surface occur in rocks along the GHf. A model for transpressional shear zones developed by Robin and Cruden (1994) predicts that within oblique transpressional shear zones the vorticity vector will be vertical near the center of a shear zone and shallows towards the margins of the shear zone. This model fairly accurately describes the kinematic conditions within the GHfz except that domain Kw does not occupy the geographic center of the fault zone; rather it encompasses an area slightly broader than the GHf. The model also predicts that the zone consisting of subvertical vorticity vectors will become narrower with a greater horizontal shortening component of deformation (Robin and Cruden, 1994).

The change in orientation of the strain ellipsoid across the fault zone is also predicted by models of transpressional deformation. Tikoff and Greene (1997) model a similar phenomenon in the Rosy Finch-Gem Lake shear zone in central California. They found in the case of wrench-dominated transpression where the angle of
convergence is relatively low (α<20°), the long axis of the finite strain ellipsoid will initially form at a sub-horizontal position and become increasingly vertical with increasing deformation (Tikoff and Greene 1997). This pattern is similar to the GHfz where accumulated finite strain values for domain Kt are relatively low resulting in sub-horizontal finite strain ellipsoids while domain Kw experienced greater finite strain and therefore contains sub-vertical strain ellipsoids.

In summary, kinematic indicators suggest the GHfz is a zone of wrench-dominated transpression where the dominant motion is dextral transpression with a component of northwest over southeast thrusting. Models for transpression suggest that deformation along the GHf had a significant horizontal shortening component manifest by a narrow zone with vertical vorticity vectors and strain ellipsoids (e.g. Robin and Cruden, 1994, Tikoff and Greene, 1997). This conceptualization suggests an angle of convergence (α) of approximately 19° along the GHfz according to the model developed by Tikoff and Greene (1997). This model also accounts for sub-horizontal lineations and strain ellipsoids in domain Kt where finite strain values are less than domain Kw.

**Kink Folds (F_k)**

Deforming all planar fabrics in all units within the study area except gabbro and diabase is a set of north-northwest trending kink and box folds designated F_k. F_k kink folds occur in most outcrops throughout the field area and have both sinistral and dextral asymmetry. Kink folds are spaced at irregular intervals in outcrop and have a width of 0.5 to 1.5 cm. Box folds are rare and are small, only 2 to 3 cm in width with
low amplitudes and an average wavelength of 3 cm. The axial surfaces for \( F_k \) folds have an average trend of N25\(^\circ\)W (Figure 38) and a near vertical dip. The trend of \( F_k \) axial surfaces appears to strongly correlate with the strike of regional diabase dikes in the Waxhaw area. Also, there appears to be an increase in concentration of kink and box folds in outcrop near Mesozoic diabase dikes. Thus, there may be a genetic relationship between \( F_k \) and the intrusion of regional Mesozoic diabase. This would also explain the dual asymmetry of kink folds in the study area. \( F_k \) is considered to post-date the major deformational phases that affect the study area (e.g. \( D_M, D_L \)) because \( F_k \) overprints structures attributed to the three main deformational events.

**Brittle Faulting (\( D_f \))**

The latest deformational event to affect the Waxhaw field area produced a set of northwest trending brittle faults (\( D_f \)). These faults are characterized by small scale brittle fractures that offset planar fabrics and mineral veins, isolated zones of silicified breccia, and inferred faulting based on abrupt and uncharacteristic changes in lithology. Brittle faults observed in outcrop are typically narrow, steep dipping and have an average trend of N25\(^\circ\)W. Separations along brittle faults indicate both sinistral and dextral slip with apparent displacements that range from several centimeters to over 50 meters. \( D_f \) is interpreted to post-date all deformational events in the field area because \( D_f \) faults cross-cut \( D_M \) and \( D_L \). Because faults associated with \( D_f \) are roughly parallel to the strike of regional diabase dikes that intrude the field area, \( D_f \) is believed to be the result of Mesozoic extension that is pervasive along the Atlantic margin of North America.
Figure 38. Symmetric Rose diagrams of kink fold axes (top) and regional trend of diabase dikes (bottom).
Summary and Conclusions

The rocks in the Waxhaw area show evidence of three ($D_M$, $D_L$, $D_f$) distinct deformational episodes. $D_M$ is the most pervasive deformation in the field area and is responsible for regional axial planar cleavages in the Marvin and Twelve Mile sequences as well as foliations in the Waxhaw and Hancock granites. $D_L$ is responsible for fabrics that overprint $D_M$ structures as well as a set of isoclinal folds that fold both $S_0$ and $S_M$. $D_L$ is interpreted to be roughly coeval with $D_M$, in which $D_M$ and $D_L$ represent distinct phases of a single progressive deformation. Structures attributed to $D_M$ lie both inside and outside the GHfz; however, they are most intensely developed and concentrated in within the GHfz. $D_L$ structures are only observed in the fault zone, plus $D_M$ and $D_L$ appear to be end members of a single deformation. Therefore, $D_M$ and $D_L$ are attributed to deformation along the GHfz.

The effect of features attributed to $D_f$, while noted, had little effect on preexisting features and constitute the last known deformational activity in the study area.

The GHfz is the major structure within the study area and constitutes a 4 to 5 km wide damage zone in the footwall Twelve Mile sequence and a 0.5 to 1 km wide hanging wall damage zone in the Marvin sequence. The GHf is a steep dipping dextral reverse fault with southeast vergence. While one of the original objectives of this investigation was the characterization of the DBsz, detailed mapping around the purported location of the shear zone revealed that it does not exhibit features common to a ductile shear zone. This conclusion is supported by finite strain analysis in the vicinity of Davis Branch Creek.
Deformation within the GHfz is interpreted to be wrench-dominated transpression where the sense of motion was dextral shear coupled with a component of northwest over southeast thrusting. Rocks in close proximity to the GHf experienced the greatest accumulation of finite strain while rocks situated more than one kilometer from GHf accumulated less finite strain. The occurrence of isolated pockets of intensely strained phyllites within the GHfz is likely the result of heterogeneous transpression or slip partitioning in a manner suggested by Tikoff and Greene (1997) and Jiang et al. (2001). Models for transpression indicate that deformation may have initiated along the current trace of the GHf (domain Kw) and over time nucleated into the surrounding country rock. This process accounts for the geometry of kinematic indicators and strain ellipsoids in domains Kw and Kt. Over time the GHf would have experienced greater strains than the surrounding GHfz, thus strain ellipsoids in domain Kw that were initially sub-horizontal would switch to a sub-vertical orientation while in domain Kt, strain ellipsoids would remain sub-horizontal. A gradual progression of strain from the GHf into the footwall damage zone would also account for the lack of a definite boundary along the southern trace of the GHfz.
Metamorphism

Introduction

This chapter examines the distribution of metamorphic mineral assemblages in the Waxhaw study area to determine the number, conditions, and timing of metamorphic events that affect the region. Results from this study indicate that rocks in the field area have undergone two metamorphic events designated $M_C$ (contact event) and $M_M$ (regional event). $M_C$ minerals are the result of local contact metamorphism during the emplacement of the Waxhaw granite of semi-pelitic and volcanic rocks in the Twelve Mile sequence. $M_M$ assemblages belong to a regional lower greenschist facies event that affects both the Twelve Mile and Marvin sequences, define fabrics produced during $D_M/D_L$, and produced a set of retrograde minerals that replace $M_C$ mineral assemblages. Understanding the nature of these metamorphic events and the conditions at which they took place is important to the understanding of key interpretations set forth in the previous chapter.

A total of 51 thin sections, including 4 polished sections for microprobe analysis, were examined in order to determine the mineral assemblages present, their textures, and to interpret the metamorphic history of the Marvin and Twelve Mile sequences. Anorthite content (An) in plagioclase was determined using a combination of the Michel-Levy method (Tobi and Kroll, 1975) and electron microprobe analysis. This chapter discusses the metamorphic history of the Waxhaw study area in the following sections: 1) contact metamorphic mineral assemblages and textures; 2) regional metamorphic mineral assemblages and textures; 3) metamorphic conditions of both the regional and contact episodes; 4) timing of metamorphic events
including their relation to regional deformational events discussed in the structure chapter.

**Contact Metamorphic Assemblages and Textures (MC)**

MC mineral assemblages occur within a metamorphic aureole that has developed along the northwest side of the Waxhaw pluton and extends from 0.5 to 1.0 km into the Twelve Mile sequence (Figure 39). Minerals associated with the metamorphic aureole define an andalusite-in isograd at its periphery, a gedrite-in isograd near the center of the aureole, and a sillimanite-in isograd within 100 m of the Waxhaw pluton. The progression of sillimanite + andalusite near the Waxhaw pluton to andalusite + biotite near the Hancock granite indicates that contact metamorphism is the result of heating related to the intrusion of the Waxhaw granite and not the Hancock granite. Therefore, the following mineral assemblages are attributed to contact metamorphism by the Waxhaw granite rather than the Hancock granite. Characteristic prograde contact metamorphic mineral assemblages from three selected samples within the contact aureole are given below.

**Twelve Mile Sequence:**

- **C-164** (semi-pelite): And + Chl(1)* + Bio + Hem + Ms + Qtz
- **C-159** (semi-pelite): And + Str + Chl(1)* + Bio + Ms + Qtz + Sil
- **C-169** (basite): Ged + Ath + Crd + Plag(An99) + Qtz + Trm + Opq

*(1) indicates prograde growth while (2) indicates secondary growth

Locations and mineral assemblages of all other samples studied within the Waxhaw contact aureole are displayed in Figure 39 and Table 2, respectively. In all
Figure 39. Geologic map illustrating the extent of the Andalusite-, Gedrite-, and Sillimanite-in isograds in the Waxhaw contact aureole. Also plotted are selected sample locations within the contact aureole including samples C-159, C-164, and C-169. Mineral assemblages for all samples are displayed in Table 2.
Table 2. Representative mineral assemblages for rocks within the Waxhaw contact aureole.

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<tr>
<th>Sample #</th>
<th>s-pelitic**</th>
<th>felsic v.</th>
<th>mafic v.</th>
<th>v. fine</th>
<th>fine</th>
<th>medium</th>
<th>Qtz</th>
<th>Plag</th>
<th>Bio</th>
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<td>C-83</td>
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The symbol X indicates the mineral is present in hand specimen and/or thin-section. Tourmaline is implied for samples C-83, C-99, and C-180 based on chemical and optical analysis on the gedrite bearing sample C-169.

* This symbol indicates the mineral is present as a porphyroblast as well as in the groundmass for that sample.
** s-pelitic = semi-pelitic; felsic v. = felsic volcanic; mafic v. = mafic volcanic
samples studied, $M_C$ minerals have been partially to completely replaced by mineral assemblages that correspond to a later regional greenschist facies event ($M_M$); however, those later retrograde minerals and related textures will be discussed in the *Regional Metamorphic* section of this chapter. The following section will review the characteristic mineral textures within the Waxhaw metamorphic aureole starting with the andalusite-isograd and moving up-grade to the mineral assemblages that typify the sillimanite-isograd.

Sedimentary and felsic volcanic rocks in the contact aureole are coarser grained than similar rocks situated outside the aureole. Felsic volcanic rocks within the gedrite and sillimanite isograds are typically composed of medium-grained muscovite, biotite, and quartz. Many felsic volcanic rocks affected by $M_C$ contain large, round, quartz idioblasts that range from 1.5 to 2 cm in width and porphyroblasts of biotite and chlorite that are 5 to 7 mm in length. Outside the gedrite isograd, felsic volcanic rocks appear similar in hand specimen to felsic volcanic rocks that have not undergone $M_C$ metamorphism.

Semi-pelitic schist in the andalusite isograd is fine-grained and contains large porphyroblasts of chlorite, biotite, and andalusite. In hand specimen, andalusite porphyroblasts are euhedral and range in length from 0.5 to 2.0 cm. Euhedral chlorite and biotite porphyroblasts are dark brown and finer-grained than andalusite porphyroblasts. In thin section, biotite and chlorite porphyroblasts are slightly to moderately deformed by a schistosity defined by white mica that corresponds to $S_M$ yet they also contain inclusion trails that are parallel to $S_M$ schistosity (Figure 30). It
Figure 40. Photomicrograph of a sericitized andalusite pseudomorph under cross polarized light from sample C-164. The core of the andalusite porphyroblast still remains. All andalusite porphyroblasts outside the sillimanite isograd have been partially to completely replaced by sericite such as this one.
should be noted that biotite porphyroblasts in these schists have been completely replaced by retrograde chlorite; however, they are interpreted to have originally been biotite porphyroblasts that grew during contact metamorphism because biotite is very common in low-T/medium-P prograde metamorphic assemblages. Also, felsic tuffs in the contact aureole contain similar biotite porphyroblasts that have been only partially replaced by chlorite. Andalusite porphyroblasts have been largely pseudomorphed by sericite; however, many andalusite porphyroblasts possess small cores of poikilitic andalusite (Figure 40). Andalusite pseudomorphs in C-164 envelope chlorite and biotite porphyroblasts and contain inclusion trails that are also oriented parallel to the surrounding $S_M$ foliation.

Semi-pelitic rocks become increasingly coarser-grained and micaceous in the gedrite and sillimanite isograds. In outcrop and hand specimen, semi-pelitic phyllites and schists in the sillimanite isograd are compositionally zoned into micaceous and quartzofeldspathic layers that are parallel to $S_M$ and range in thickness from 2 to 5 mm. Micaceous layers are composed of medium-grained muscovite and biotite, along with varying amounts of fine-grained garnet, medium to coarse andalusite porphyroblasts, and fine needles of white sillimanite. In thin-section, andalusite is poikiloblastic with inclusions of fine-grained quartz, muscovite, and staurolite (Figure 41). Andalusite porphyroblasts inside the sillimanite isograd also lack foliation-parallel inclusion trails unlike metamorphic porphyroblasts outside the sillimanite isograd. Large porphyroblasts of metamorphic biotite are rare and chlorite porphyroblasts are absent. Staurolite occurs as very-fine inclusions in garnet and
Figure 41. Photomicrograph of an andalusite porphyroblast in plane polarized light from sample C-159. Note the two cleavages present in this sample and that it does not contain preferentially oriented inclusion trails. Many porphyroblasts in C-159 contain randomly oriented inclusions indicating growth prior to $D_M$. 
Figure 42. Photomicrograph of garnet, chloritoid, and staurolite in sample C-159. The high relief, very fine-grained yellow inclusions within the garnet are interpreted to be staurolite. Staurolite inclusions were also observed in andalusite.
Figure 43. Photomicrograph in plane light of very fine, fibrous sillimanite in sample C-159. Sillimanite is rare in thin-section and usually occurs in bundles such as seen here or as individual needles in quartz.
andalusite (Figure 42). Fibrous sillimanite occurs as individual grains or in small bundles within quartz (Figure 43).

A series of contact metamorphosed basaltic layers between 4 and 15 m wide and up to 3 km long occur inside the gedrite isograd between the Hancock and Waxhaw granites. In outcrop, these layers are fine-grained, greenish-black to salt and pepper textured, foliated, and possess dense clusters of fine to medium grained amphibole needles that are randomly as well as preferentially oriented parallel to cleavage. Microprobe analysis reveals these amphiboles to be the orthoamphibole gedrite, which is a common mineral in low-P/medium-T metabasites. Cordierite porphyroblasts are visible in polished slabs of hornfelsed metabasalt.

In thin section, groundmass in hornfelsed basalt is composed of very fine-grained, granoblastic plagioclase, quartz, and opaque minerals. Gedrite occurs as long, prismatic needles that overgrow the surrounding groundmass and contain small inclusions of quartz and plagioclase (Figure 44). Exsolution lamellae are discernible within larger gedrite crystals. These lamellae are interpreted to be anthophyllite, which commonly exsolves into fine lamellae within gedrite crystals that have cooled through the gedrite-anthophyllite solvus (Spear, 1993) (Figure 45). Subhedral cordierite porphyroblasts have a poikiloblastic to skeletal texture with numerous small inclusions (Figure 46) of fine-grained quartz, feldspar, tourmaline, and opaque minerals.
Figure 44. Photomicrograph under plane polarized light of optically homogeneous gedrite blade in sample C-169. Retrograde chlorite sheets have grown in the groundmass surrounding the gedrite crystal.
Figure 45. Photomicrograph under plane polarized light of fine-scale exsolution lamellae (parallel to violet line) in gedrite crystal from sample C-169.
Figure 48. Photomicrograph under cross polarized light of poikiloblastic cordierite porphyroblasts.
Regional Metamorphic Assemblages and Textures (M_M)

M_M regional metamorphism affects all pre-Mesozoic rocks in the study area. M_M mineral assemblages define a prograde metamorphic event in the Twelve Mile and Marvin sequences and a retrograde event in rocks affected by M_C. Mineral assemblages attributed to M_M are listed below:

Marvin sequence:
- Felsic volcanic rocks: Qtz + Olg(An28-30) + Bio + Wm ± Chl ± Ep ± Grt
- Chlorite schist: Chl + Wm + Ep + Qtz + Plag

Twelve Mile sequence:
- Felsic volcanic unit: Qtz + Olg(An11-12) + Bio + Chl + Ep ± Wm ± Grt
- Sedimentary unit: Clay minerals + Qtz + Chl + Wm ± Ep ± Py ± Opq
- Basaltic rocks: Act + Ep + Chl + Plag(Ab?) + Sph + Opq
- Phyllite: Wm + Chl + Plag + Qtz + Opq

Plutonic Rocks:
- Waxhaw Granite: Qtz + Mic + Ab(An4-8) + Bio + Mgn ± Ep

Rocks with M_C mineral assemblages:
- C-164 (semi-pelite): Chl(2)* + Wm + Ms + Qtz
- C-159 (semi-pelite): Ctd + Grt + Ms + Hem + Mgn + Wm + Fe-Chl
- C-169 (basite): Bio + Chl
The Marvin and Twelve Mile sequences are underlain predominantly by semi-pelitic and felsic volcanic rocks with minor mafic tuffs and chloritic schists. The similarity in rock-type between the two sequences makes it relatively easy to compare their metamorphic condition. Mineral assemblages in the Waxhaw granite were not helpful in the determination of regional metamorphic grade although deformation mechanisms from the granite did prove useful to this end. The following section describes the minerals and textures in the Twelve Mile and Marvin sequences that resulted from \( M_M \) metamorphism. Rocks in the Twelve Mile sequence that experienced \( M_C \) contact metamorphism prior to \( M_M \) regional metamorphism possess unique mineral assemblages and textures that are indicative of retrograde metamorphic conditions. Those mineral assemblages and textures will be discussed at the end of this section.

Regional metamorphic conditions are fairly uniform throughout the field area with no significant difference in mineral assemblages between the Marvin and Twelve Mile sequences. Most rocks affected by \( M_M \) retain their primary volcanic and sedimentary textures. Metamorphic minerals are scattered throughout each sample and occur in small isolated patches along cleavage seams. Phyllites and schists possess the greatest abundance of \( M_M \) minerals.

The most common \( M_M \) metamorphic assemblages in semi-pelitic rocks from both the Twelve Mile and Marvin sequences are clay minerals, chlorite, and sericite along with poikiloblastic pyrite. Texturally, some sedimentary rocks in the GHfz appear unaffected by \( M_M \) metamorphism. Fine-grained biotite, chlorite, sericite, and epidote are the most common metamorphic minerals in felsic volcanic rocks from
both sequences. Metamorphic biotite is distinguished from biotite phenocrysts in felsic tuffs and porphyries by its smaller grain-size and preferential growth along cleavage seams. Garnet, while rare, is interpreted to be metamorphic in origin and was identified in hand specimens and thin sections in felsic tuff from both sequences.

Mafic tuffs and mafic-derived sedimentary rocks in both sequences yield the most diagnostic $M_M$ metamorphic mineral assemblages. In the Marvin sequence, mafic schists are composed of fine- to medium-grained chlorite and muscovite while plagioclase is almost completely saussuritized. Basalt from the Twelve Mile sequence has a metamorphic assemblage of actinolite, chlorite, and epidote. Biotite is absent in mafic rocks from both sequences.

Both $S_M$ and $S_L$ are defined by $M_M$ minerals, which includes chlorite, sericite, biotite, and clay minerals that have grown parallel to cleavage seams in phyllitic-slates in both sequences. In mafic rocks, $S_M$ is defined by aligned chlorite and actinolite with long dimensions oriented parallel to the foliation. $L_M$ mineral lineations are defined by aligned clusters of biotite and chlorite as well as aligned amphiboles. Chlorite, biotite, and sericite define $S_L$ in felsic tuffs from the GHfz.

Rocks affected by $M_M$ exhibit variable degrees of recrystallization. Volcaniclastic rocks proximal to the GHf as well as in isolated lenses of phyllite within the GHfz display the greatest degree of recrystallization. Quartz in these rocks has experienced intense grain-size reduction by dynamic recrystallization and feldspar has been almost completely saussuritized. In the majority of phyllitic-slates in the Twelve Mile and Marvin sequences, quartz has undergone partial
recrystallization while feldspar has deformed primarily by cataclasis. Similar
deformation mechanisms are observed in quartz and feldspar in the Waxhaw granite.

Rocks that were affected by $M_C$ prior to $M_M$ metamorphism bear retrograde
$M_M$ mineral assemblages that replace $M_C$ metamorphic minerals. In semi-pelitic
schists that lie between the andalusite and sillimanite isograds, andalusite
porphyroblasts are partially to completely replaced by fine-grained sericite while
large biotite porphyroblasts have been replaced by chlorite. Biotite porphyroblasts in
felsic volcanic rocks have also been partially replaced by retrograde chlorite.
Retrograde chlorite and biotite overgrow the edges of cordierite and gedrite crystals
in metabasites inside the gedrite isograd.

Schists in the sillimanite isograd possess a retrograde mineral assemblage that
includes garnet and chloritoid. Garnets are fine-grained, poikiloblastic, and overgrow
$M_C$ minerals as well as fine sericitic micas that define $S_M$ foliation in thin-section.
Chloritoid is only identifiable in thin section and commonly overgrows $M_C$ minerals.
Chloritoid porphyroblasts are euhedral, fine-grained, and only occur around
andalusite porphyroblasts (Figure 47) where they typically grow within biotite.
Andalusite porphyroblasts in the sillimanite isograd are relatively unaffected by
sericitization. The edges of some chloritoid and andalusite porphyroblasts have been
replaced by a late phase of greenish-brown to olive brown chlorite. Retrograde
chlorite also replaces porphyroblastic biotite. The growth of this retrograde chlorite
is interpreted to be late during $M_M$ retrogression because of its minor abundance in
thin section and because it replaces chloritoid, which is interpreted to have grown
early during $M_M$ retrograde metamorphism.
Figure 47. Photomicrograph under plane polarized light of chloritoid grains that have crystallized around an andalusite porphyroblast. Photo taken from sample C-159.
Conditions of Metamorphism

The following section explores the conditions of metamorphism during both $M_C$ and $M_M$ by examining textures, mineral assemblages, and deformation mechanisms in metamorphosed rocks in order to establish P-T paths for each event. This section begins with the regional metamorphic event ($M_M$) because all rocks, including those affected by $M_C$, experienced $M_M$ metamorphism. Also, the metamorphic conditions during $M_M$ play a significant role in the interpretation of the conditions for retrograde metamorphic assemblages in rocks affected by $M_C$.

The contact metamorphic section will follow the regional metamorphic section and discuss the conditions that led to peak metamorphism during $M_C$ followed by the P-T conditions that resulted in $M_M$ retrograde metamorphism of $M_C$ mineral assemblages. The final section discusses the results of microprobe analysis on a sample of metabasite from the contact aureole that contains an orthoamphibole-cordierite mineral assemblage and the possible metamorphic conditions implied by that assemblage.

Regional Metamorphic Conditions ($M_M$)

The interpreted regional metamorphic grade in the study area is based upon the characteristic mineral assemblages in the Marvin and Twelve Mile sequences and the observed products of deformation mechanisms in all units affected by $M_M$. The An-composition of plagioclase in felsic volcanic and plutonic rocks was not considered in the interpretation of $M_M$ conditions because their chemistry is
interpreted to reflect conditions during primary igneous crystallization rather than metamorphic recrystallization.

Basaltic rocks from the Twelve Mile sequence exhibit mineral assemblages indicative of the chlorite zone of the greenschist facies. Chlorite-rich schists in the Marvin sequence reflect a regional greenschist facies metamorphic event in that sequence. Replacement of plagioclase by epidote and sericite and replacement of biotite by chlorite in rocks within the GHfz is interpreted to be the result of hydrothermal alteration during progressive deformation within the fault zone. Such alteration is common in many greenschist facies shear zones (e.g. Barker, 1990, Gates and Speer, 1991). Felsic and mafic volcanic tuffs retain many of their original igneous features, which is also common in volcanic rocks that have experienced chlorite grade metamorphism (Yardley, 1989). Sedimentary rocks in both sequences contain minerals that are too fine-grained to be identified in thin-section except along cleavage seams where common metamorphic minerals include white mica and chlorite. Metamorphic garnets in felsic tuff from the Marvin and Twelve Mile sequences are interpreted to be spessartine, which can grow at low to middle greenschist facies (Barker, 1990).

The dominant deformation mechanism in volcaniclastic strata in the Marvin and Twelve Mile sequences is pressure solution, which is a process most commonly associated with sub- to lower-greenschist facies (T<300°C)(e.g. Gratier, 1987, Passchier and Trouw, 1996). However, recent studies indicate that pressure solution can be operative at much higher temperatures (Niemeijer et al., 2002). Undulatory extinction and recrystallization by grain boundary migration and sub-grain rotation
are observed in quartz in all deformed rock units. Feldspar is commonly deformed by cataclasis with a few grains displaying patchy undulose extinction and internal kinking. All of these features are consistent with low to moderate temperature (350-400ºC) deformation (e.g. Passchier and Trouw, 1996).

The similarity of mineral assemblages and deformation mechanisms in the Marvin and Twelve Mile sequences indicate that metamorphic conditions are similar in both sequences. Based upon these observations, $M_M$ mineral assemblages and deformation mechanisms observed in the Waxhaw area are interpreted to have formed during lower to middle greenschist facies conditions. Temperatures and pressures under these conditions range from 350 to 450ºC and 2 to 8 kbars (Figure 49) (Yardley, 1990). Deformation mechanisms further constrain maximum temperatures to 400ºC. If a classic Barrovian P-T path is assumed for $M_M$ regional metamorphism, then peak pressures are estimated between 3 to 4 kbars, which corresponds to roughly 9-12 km depth within the crust (assuming a pressure gradient of approximately 1 kbar/3 km).

Contact Metamorphic Conditions ($M_C$)

The mineral assemblages from samples C-159 and C-164 will be referred to throughout the following section in order to characterize the prograde and retrograde metamorphic conditions that affected rocks in the Waxhaw contact aureole. Sample C-164 bears a mineral assemblage representative of semi-pelitic rocks between the andalusite and gedrite isograds while sample C-159 is representative of semi-pelitic schists inside the sillimanite isograd. Throughout the discussion, prograde is used to
indicate progressive heating that resulted from contact metamorphism while

*retrograde* refers to cooling and re-equilibration of M$_C$ assemblages to the M$_M$ regional event.

Sample C-164 is a fine-grained schist from inside the andalusite isograd that contains coarse grained andalusite, biotite, and chlorite porphyroblasts. Many of the andalusite porphyroblasts have been replaced by fine sericite and quartz with a few remnant cores of unaltered andalusite in some of the larger grains. Biotite has been completely replaced by chlorite. Andalusite pseudomorphs envelope the smaller biotite porphyroblasts implying that biotite growth initiated before andalusite. Both andalusite and biotite porphyroblasts in C-164 have been deformed and contain inclusion trails that are parallel to S$_M$ foliation.

Schist in the sillimanite isograd bears a prograde assemblage of andalusite, biotite, staurolite, and minor fibrous sillimanite. The absence of recognizable staurolite pseudomorphs and its relatively minor presence as inclusions in garnet and andalusite indicates that staurolite growth was not significant during M$_C$. The minor abundance of fibrous sillimanite and the relatively unaltered state of andalusite also suggests that rocks inside the sillimanite isograd were heated to temperatures that were barely sufficient for the growth of sillimanite.

The growth of garnet and chloritoid in C-159 and the replacement of andalusite and biotite porphyroblasts in C-164 by sericite and chlorite, respectively, is interpreted to mark the beginning of M$_M$ retrogression. This retrogressive event appears to overlap with the end of M$_C$ metamorphism (see below). The initial retrograde M$_M$ assemblage in C-159 includes garnet, chloritoid, and Fe-oxide
minerals and is followed by a later stage of $M_M$ retrograde metamorphism marked by the replacement of andalusite and biotite by Fe-chlorite. Chloritoid occurs exclusively around andalusite porphyroblasts where it grows within as well as between $M_C$ biotite and andalusite porphyroblasts together with magnetite, hematite, and muscovite. Both garnet and chloritoid are interpreted to belong to the $M_M$ retrograde assemblage because both minerals overprint $M_C$ assemblages as well as $S_M$ schistosity.

The absence of inclusion trails parallel to $S_M$ in most sillimanite isograd porphyroblasts suggests that $M_C$ assemblages inside the sillimanite isograd are pre- to syn-kinematic with respect to $D_M$. However, $M_C$ prograde porphyroblasts outside the sillimanite and gedrite isograds are deformed by $S_M$ and clearly possess $S_M$-parallel inclusion trails, thus they are interpreted as late-synkinematic. Based on these observations, the intrusion of the Waxhaw pluton and subsequent heating of the country rock is considered synkinematic with respect to $D_M/D_L$ progressive deformation.

The following progression of prograde and retrograde contact metamorphic mineral reactions is interpreted from the geologic evidence from semi-pelitic schists within the Waxhaw contact aureole. Mineral reactions and P-T paths were estimated based on textural evidence as well as upon the following assumptions: 1) all semi-pelitic rocks are interpreted to have a high $X_{Fe}$ content due to the abundance of Fe-oxides and the absence of cordierite in all semi-pelitic hornfelsed rocks; 2) Mn-content in C-159 is considered to be high enough ($MnO \sim 0.25-0.3$ wt%) to lower the
garnet stability field (e.g. Mahar et al., 1997, Martinez et al., 2002) and allow garnet growth at pressures estimated for $M_M$ retrograde metamorphism.

The P-T grid that contains mineral reactions for low-P pelites is displayed in Figure 48. P-T paths for C-159 and C-164 are plotted onto the same P-T grid in Figure 49. AFM diagrams for each P-T point are displayed in Figures 50 and 51. The majority of AFM diagrams correspond to the mineral assemblages observed in C-159 because that sample contains the most diverse mineral assemblage and therefore the most data relating to $M_C$ contact and retrograde metamorphism. It should be noted that an assumed Mn-component present in C-159 would affect the samples bulk composition such that it would not plot perfectly onto an AFM triangle. However, for the purpose of this investigation, topology and mineral reactions are considered correct.

$M_C$ metamorphism initiated with the emplacement of the Waxhaw granite into the Twelve Mile sequence. Lithostatic pressures during emplacement could not have exceeded 4.5 kbars because of the presence of andalusite in semi-pelitic rocks in the contact aureole. Pressures can be further constrained to a range of 2.0-2.5 kbars during $M_C$ prograde metamorphism because staurolite, while able to grow, was not a significant $M_C$ metamorphic phase (Martinez et al., 2002). Based on these pressure estimates, the first prograde mineral assemblage is chlorite and biotite through the reaction:

\[ \text{clay minerals + MnO} \rightarrow \text{chlorite + biotite + quartz + H}_2\text{O} \quad [1] \]
Figure 48. P-T grid for pelites in the KFMASH system. Grid is after SPaC 2000 grid of Spear et al. (2000, personal website). Selected KFMASH and KFASH reactions are in black and red lines, respectively. Aluminosilicate stability fields are represented by green lines according to Holdaway (1971).
Figure 49. P-T grid from Figure 48 showing two P-T paths for contact metamorphism. Light-blue arrows represent P-T path for C-159 while dark blue arrow represents P-T path for sample C-164. Numbered points correspond to mineral reactions and corresponding AFM diagrams. Box labelled $M_M$ represents the P-T conditions of $M_M$ regional lower greenschist facies metamorphism.
Figure 50. The series of AFM diagrams for prograde mineral reactions 1 through 4. Numbers 1 - 4 refer to P-T points labeled in Figure 49. The plus symbol represents the interpreted bulk composition for both samples C-159 and C-164.
Figure 51. The series of AFM diagrams for prograde mineral reaction 5 and retrograde reactions 6 and 7a,b. Numbers 5 - 7 refer to P-T points labeled in Figure 49. The plus symbol represents the interpreted bulk composition for both samples C-159 and C-164.
Reaction 1 initiated at temperatures between 300 and 450ºC and pressures of 2.0 to 2.3 kbars. As temperatures exceeded 450ºC, andalusite began to crystallize at the expense of chlorite according to the reaction:

\[ \text{chlorite} + \text{muscovite} \rightarrow \text{andalusite} + \text{biotite} + \text{quartz} + \text{H}_2\text{O} \]  \[2\]

Staurolite first appeared when temperatures reached approximately 550ºC. Pressure is still estimated to have been approximately 2.0 to 2.3 kbars which would allow for limited staurolite growth by pushing the P-T path of sample C-159 into the lower limit of the staurolite stability field and facilitating the reaction:

\[ \text{andalusite} + \text{biotite} \rightarrow \text{staurolite} + \text{muscovite} + \text{Fe-Ti oxides} + \text{quartz} \]  \[3\]

During reaction 3, staurolite is interpreted to have grown at the expense of andalusite and biotite. Most of the Mn in C-159 has until this point been assumed to have fractionated into chlorite and biotite. However, at this point, the breakdown of biotite allows for the transfer of Mn into Fe-Ti oxides. As staurolite broke down to form garnet, Mn from Fe-Ti oxides fractionated into garnet as part of the reaction:

\[ \text{staurolite} + \text{muscovite} + \text{Fe-Ti oxides} + \text{quartz} \rightarrow \text{garnet} + \text{biotite} + \text{andalusite} \]  \[4\]

Garnet here is interpreted to be primarily spessartine and almandine. The interpreted growth of garnet and andalusite at the expense of staurolite is supported by the presence of staurolite inclusions in garnet and andalusite.

The growth of garnet also marks a shift in metamorphic conditions from \( M_C \) to \( M_M \) metamorphism. \( M_M \) is accompanied by an increase in pressure that pushes the P-T path of C-159 into the garnet stability field as well as all for sillimanite growth at lower temperatures. Temperatures during garnet growth are estimated to be approximately 580ºC and are interpreted to have continued to increase because of the
presence of sillimanite in C-159. Thus, an apparent overlap between $M_M$ and $M_C$ mineral assemblages is implied with the growth of sillimanite according to the reaction:

$$\text{andalusite} \rightarrow \text{sillimanite}$$  \[5\]

Reaction 5 is estimated to have occurred above 600°C with pressures estimated at 2.5 to 3.0 kbars. Reaction 5 also marks the peak metamorphic assemblage for $M_C$ metamorphism inside the sillimanite isograd.

Retrograde metamorphism inside the sillimanite isograd is marked by the growth of garnet and chloritoid. As temperatures began to cool, the mineral reactions in C-159 are interpreted to have shifted back to reaction 4 before proceeding to chloritoid growth through the reaction:

$$\text{andalusite} + \text{biotite} + \text{garnet} \rightarrow \text{chloritoid} + \text{muscovite} + \text{quartz} + \text{Fe-oxides}$$  \[6\]

This reaction is interpreted to have begun when temperatures dropped below 580°C while pressures continued to increase to approximately 3.0 to 3.5 kbars. The breakdown of andalusite was a significant factor during this reaction because chloritoid is only observed around andalusite porphyroblasts. Furthermore, symplectite is observed in andalusite grains that are in contact with chloritoid (Figure 52). Fe$^{2+}$ and Fe$^{3+}$ from biotite and garnet that did not fractionate into chloritoid are believed to have been incorporated into magnetite and hematite. The final reaction to occur during $M_M$ retrogression was replacement of chloritoid and biotite by Fe-chlorite according to the reaction:

$$\text{biotite} + \text{chloritoid} \rightarrow \text{Fe-chlorite} + \text{white mica} + \text{quartz}$$  \[7a\]
P-T diagrams indicate that reaction occurs between 520-540°C and 3-4 kbars (Figure 48). These pressures correspond to pressures estimated for $M_M$ metamorphic conditions.

$M_C$ prograde metamorphic conditions between the andalusite and gedrite isograds were similar to those interpreted for rocks inside the sillimanite isograd except that temperatures did not exceed 550°C as indicated by the lack of garnet, chloritoid, and sillimanite in those rocks. The first metamorphic minerals to grow in sample C-164 were chlorite and biotite according to reaction 1. Peak conditions were experienced in C-164 during the growth of andalusite through reaction 2. Temperatures and pressures for that reaction are estimated to range from 500-550°C and 2.0-2.5 kbars. $M_M$ retrogression in C-164 is then interpreted to have commenced with the replacement of andalusite by sericite and biotite by chlorite according to reaction 7b at temperatures below 500°C and pressures between 3-4 kbars.

$$\text{biotite + andalusite } \rightarrow \text{ Fe-chlorite + white mica + quartz}$$  \[7b\]

In summary, metamorphic conditions during $M_C$ are interpreted to have proceeded along an isobaric path up to approximately 580°C within the sillimanite isograd. Between the gedrite and andalusite isograds, peak metamorphic temperatures did not likely exceed 550°C. $M_M$ retrogression is interpreted to have begun with an increase in pressure associated with $M_M$ regional greenschist facies metamorphism, which induced garnet growth in C-159. Temperatures inside the sillimanite isograd are interpreted to have continued to increase until peak conditions were achieved at approximately 600-625°C and 2.5 to 3.0 kbars. Temperatures within the contact aureole are then interpreted to have dropped facilitating retrograde
Figure 52. Photomicrograph under plane polarized light of symplectite in an andalusite porphyroblast that is in contact with chloritoid, leading to the interpretation that chloritoid is crystallizing at the expense of andalusite. Note also that greenish-brown chlorite is replacing the edges of both chloritoid and andalusite.
chloritoid growth within the sillimanite isograd and the replacement of metamorphic porphyroblasts inside the andalusite isograd by retrograde sericite and chlorite.

*Cordierite-Orthoamphibole assemblage*

A sample of hornfelsed basalt (C-169) with coarse porphyroblastic amphiboles was collected from within the Waxhaw contact aureole between the Hancock and Waxhaw granites. The amphibole porphyroblasts and coexisting minerals were analyzed using wavelength dispersive electron microprobe analysis for the purpose of determining their compositions. The results indicate that two orthoamphibole phases are present in C-169: gedrite and anthophyllite. The data obtained here are important to the interpretation of the metamorphic history of rocks within the Waxhaw contact aureole.

The location of sample C-169 can be viewed in Figure 39. Rocks that contain orthoamphibole are primarily metabasalts that occur in narrow layers and lenses within the Twelve Mile sequence between the Hancock and Waxhaw granites. The first occurrence of orthoamphibole in these layers marks the gedrite-in isograd within the Waxhaw contact aureole. These basalts are dense, greenish-black to salt and pepper textured, fine-grained, and typically bear a reddish-brown weathering rind common to mafic rocks in the field area. Orthoamphiboles occur as large bladed crystals 1-5 mm wide and 0.5-2 cm long or in densely clustered masses of fine-grained needles (Figure 53). They lie in the foliation plane but also cross-cut it, indicating that orthoamphibole growth occurred late during the main deformational ($D_M$) phase.
Figure 53. Photograph taken of an outcrop of metabasalt C-169. The amphiboles in this basalt are the mineral gedrite. Notice that gedrite here has crystallized as both individual crystals and in dense clusters and dikelets. Photograph taken on an unnamed dirt road approximately 0.25 miles north of Steele Hill Rd. Mechanical pencil for scale.
A polished section of C-169 was made and analyzed optically before microprobe analysis. In thin-section, gedrite occurs as large, bladed, prismatic crystals that are pleochroic pale-blue and gray in plane polarized light and exhibit low to middle 2nd order inference colors in cross polarized light. Individual anthophyllite crystals are not apparent in thin-section and primarily occur as fine lamellae within coarse gedrite blades. Other minerals in C-169 include medium-grained, poikiloblastic cordierite, and fine-grained quartz, plagioclase, tourmaline (Figure 54), and opaque minerals. Retrograde biotite and chlorite have grown around the edges of gedrite and cordierite grains.

Gedrite is a Si-poor, Al- and Na-rich orthoamphibole typically found in regions of low to medium pressure metamorphism (Deer et al., 1992). The formula for an ideal gedrite contains a half-filled A-site composed principally of Na and two Al atoms in tetrahedral coordination with Si (e.g. Robinson et al., 1971, Cameron and Papike, 1979, Spear, 1980). The structural formula for ‘ideal’ gedrite was proposed by Robinson et al. (1971) as:

$$\text{Na}_{0.5} \ R^{2+}_{2} \ (R^{2+}_{4}R^{3+}_{1}) \ \text{Si}_{6} \ \text{Al}_{2} \ \text{O}_{22} \ (\text{OH})_{2}$$

However, later studies revealed the A site for gedrite can contain significantly larger concentrations of Na (Berg, 1985). Anthophyllite is a Si-rich and Al-poor orthoamphibole that commonly occurs with gedrite. Deer et al. (1992) list the ‘ideal’ chemical formula for anthophyllite as:

$$\text{(Mg, Fe}^{2+}_{7} \ \text{Si}_{8} \ \text{O}_{22} \ (\text{OH, F})_{2}$$

However, natural anthophyllite typically contains an appreciable amount of Al in its chemical structure and therefore the boundary between gedrite and anthophyllite is
Figure 54. Photomicrograph under plane polarized light of fine-grained tourmaline in a cordierite porphyroblast in metabasalt C-169. The presence of tourmaline points to the importance of hydrothermal fluids during the metasomatism and subsequent metamorphism of C-169.
usually taken at the composition with Si;Al in tetrahedral sites (e.g. Deer et al., 1992). Therefore, spot analyses that possess Si in excess of 7.0 are interpreted to be anthophyllite rather than gedrite.

Approximately 30 spot analyses were obtained from four orthoamphibole crystals in sample C-169. Nearly the same number of spot analyses (n=29) were conducted on individual grains of plagioclase. Tables 3 and 4 list selected spot analyses from gedrite and plagioclase grains, respectively. Analyses span a moderate range of anthophyllite and gedrite with the greatest Si cations per unit formula in anthophyllite at 7.24 to the lowest in gedrite with 6.47 Si per formula unit. No K$_2$O was detected in any analyses of orthoamphibole from C-169. An-content in plagioclase is listed along the bottom row in Table 4. Plagioclase grains in C-169 are extremely Ca-rich with An-values that range from 95.8 to 97.5.

Structural formulas for orthoamphibole are based on 23 oxygens (anhydrous basis for calculation) while plagioclase formulas were calculated on an 8 oxygen basis. Al was added to tetrahedral sites until the sum of tetrahedral Si and Al equaled eight in the structural formulas of orthoamphibole. All remaining Al was allocated to octahedral sites. Cations of Fe, Mg, Mn, Ti, Ca, and Na were added to M1, 2, 3, and 4 sites unit a sum of seven was reached. Any remaining Ca and Na was placed into A-sites. All Fe is assumed to be Fe$^{2+}$ and no attempt was made to calculate ferric-iron content in orthoamphibole; however, Fe$^{3+}$ fractionation into gedrite is thought be low (e.g. Robertson et al., 1982, Berg, 1985). Analyses with cation totals in excess of seven, however, are interpreted to reflect the presence of at least some Fe$^{3+}$ in the chemical structure of gedrite and anthophyllite (Berg, 1985).
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</table>

**Total An** 97.5 97.1 96.4 96.4 95.8 96.4 96.7 97.2
It has been noted in other studies (e.g. Spear, 1980, Robinson et al., 1982, Berg, 1985, Stoddard and Miller, 1990) that the most common substitution mechanisms within the orthoamphibole solid-solution series are the edenitic substitution:

\[ \square^A + \text{Si}^{iv} = \text{Na}^A + \text{Al}^{iv} \]

and the (Mg-) tschermakitic substitution:

\[ \text{Mg}^{vi} + \text{Si}^{iv} = \text{Al}^{vi} + \text{Al}^{iv} \]

Comparison of the data gathered in this analysis confirms that similar substitution mechanisms occur in orthoamphiboles from C-169. The chart in Figure 55a displays a strong positive correlation between tetrahedral Al and A-site occupancy that is indicative of edenitic substitution. A plot of Fe- and Mg- cations in M1, M2, M3, and M4 sites versus octahedral Al (Figure 55b) demonstrates that tschermakitic substitution involved Mg and not Fe-substitution. A visible negative correlation exists on a plot between Fe-cations vs. Si-cations (Figure 55d), which indicates that Fe-exchange did occur during crystallization and cooling of orthoamphibole and that Fe-fractionation was greater in Si-poor orthoamphiboles. However, a possible Si-Fe exchange mechanism is unknown at this time.

Spear (1980) attempted to define a miscibility gap in the anthophyllite-gedrite series and suggested that a solvus exists between the two end-member orthoamphiboles, the crest of which lies at approximately 600°C at orthoamphibole compositions with \( \text{Al}^{iv} = 0.75, \text{Al}^{vi} = 0.3 \) and A-site occupancy of 0.15. Spear (1993) also suggested the crest of the solvus increases in temperature with increasing pressure. A plot of total Al vs. A-site occupancy shows that most of the gedrites
Figure 55. a) Calculated A-site occupancy plotted against tetrahedral Al for all gedrite analyses. Positive linear trend is indicative of edenitic substitution. b) Octahedral Al versus total Fe and Mg cations. Negative trend of Mg indicates that it and not Fe is active during tschermakitic substitution. c) Compositions of orthoamphiboles plotted according to A-site occupancy and total Al. Compositions for the most part lie above the orthoamphibole solvus. d) A plot of Fe versus Si illustrating the inverse relationship between those cations in orthoamphibole.
analyzed plot above the orthoamphibole solvus (Figure 55c) and display continuous variation with no pronounced gap between Al-poor anthophyllite and Al-rich gedrite. The absence of two visibly distinct orthoamphibole phases, the presence of exsolution lamellae, and continuous variation in composition indicates that orthoamphiboles in C-169 experienced temperatures that exceeded 600ºC. The presence of a single orthoamphibole phase in thin-section with abundant exsolution lamellae suggests super-solvus crystallization of gedrite that was followed by a period of slow cooling through at least a portion of the orthoamphibole solvus.

These data are consistent with hornblende hornfels facies conditions for C-169 which coincides with peak metamorphic temperatures during $M_C$. The growth of orthoamphibole did not occur until late during $M_C$ metamorphism when temperatures exceeded 600ºC which was adequate for sillimanite growth in schists within the contact aureole. Textural evidence indicates that orthoamphibole growth must have continued after peak $M_C$ conditions with little $M_M$ retrogression of the assemblage except for minor replacement of orthoamphiboles by biotite and chlorite.

Many studies indicate that metamorphic growth of cordierite-orthoamphibole assemblages must be preceded by metasomatic alteration of the parent rock (e.g. Humphris and Thompson, 1977, Bowers et al., 1985, Schumacher and Robinson, 1987, Stoddard and Miller, 1990) usually by sea-water. Hydrothermal alteration of basalt by sea-water removes CaO and enriches the parent rock in Na$_2$O. The presence of tourmaline in C-169 does indicate a fluid phase was present in C-169; however, CaO was not removed from the rock. Instead, it appears that CaO was driven into plagioclase during metamorphism which accounts for the high An-contents C-169
plagioclase. All of the gedrites analyzed also have Na-contents less than the ‘ideal’ gedrite (Na ~0.1-0.3). It therefore appears unlikely that sea-water altered the parent rock chemistry prior to metamorphism. More likely, water from the intruding pluton and/or from metamorphic dehydration reactions in semi-pelitic schists around the basaltic layers contributed to metasomatism of C-169, which was then followed by orthoamphibole growth during peak M\(_C\) conditions.

**Timing of Metamorphism**

The timing of metamorphism in the Waxhaw area is constrained by the following factors: 1) the crystallization age of the Waxhaw granite; 2) the correlation of the age of deformation with metamorphic mineral growth, because M\(_M\) minerals define D\(_M\) and D\(_L\) fabrics. Age data obtained from rocks deformed by the GHfz are inconclusive regarding the timing of deformation and metamorphism. \(^{40}\text{Ar}/^{39}\text{Ar}\) ages from white mica within the GHfz to the southwest and northeast of the study area constrain the timing of cooling of white mica within the fault zone to approximately 360-320 Ma (Lavallee, 2003). However, \(^{40}\text{Ar}/^{39}\text{Ar}\) analysis on amphibolite deformed by the GHfz south of Van Wyck, South Carolina constrains cooling within the GHfz to ca. 400-368 Ma (Boland and Dallmeyer, 1997). If these ages reflect cooling after successive periods of metamorphism and deformation within the GHfz, then they suggest that metamorphism and deformation attributed to activity in the GHfz occurred throughout the middle and late Paleozoic. However, the results of this study indicate that D\(_M\) and D\(_L\) deformation, and subsequently M\(_M\) was roughly coeval with
the intrusion of the Waxhaw pluton, which has been dated to 495±21 Ma (Fullagar, 1981).

Textural relationships of metamorphic mineral assemblages along with the interpreted conditions of metamorphism indicate the timing of $M_C$ and $M_M$ overlap. This is supported by microtextural relationships between $M_C$ porphyroblasts and $S_M/S_L$ foliations, which suggest contact metamorphism related to the Waxhaw granite initiated before $D_M$ and continued late into $D_M/D_L$ progressive deformation. Based upon the relationships of the synkinematic Waxhaw pluton and implied sequence of metamorphic events in the Waxhaw area, the timing of metamorphism along the GHfz in the Waxhaw area appears to have occurred ca. 500 Ma. However, this age is suspect as it is derived from the Rb-Sr whole rock age of the Waxhaw pluton.

A loss or gain of Rb during deformation would alter the Rb-Sr whole rock age of the granite, accordingly. The best interpretation is an implied loss of Rb during deformation, which would result in an apparent age for the Waxhaw granite that is older than its true crystallization age. This interpretation allows for a middle to late Paleozoic crystallization age for the Waxhaw granite, which fits with the middle to late Paleozoic $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages obtained from the GHfz, implying that metamorphism occurred during the middle and/or late Paleozoic.

**Summary**

Rocks in the Waxhaw study area have been affected by two metamorphic events which appear to overlap in time. The first metamorphic episode was a contact event ($M_C$) associated with the intrusion of the Waxhaw pluton, which
metamorphosed semi-pelitic and volcanic rocks in the Twelve Mile sequence to hornblende hornfels facies. $M_C$ metamorphism was also responsible for the growth of gedrite-cordierite assemblages in basaltic rocks within the metamorphic contact aureole. The later metamorphic episode ($M_M$) was a regional lower greenschist facies event that affected both the Marvin and Twelve Mile sequences and was likely to be coeval with deformation along the GHfz. Both metamorphic events are broadly contemporaneous with the intrusion of the Waxhaw pluton.
Discussion and Conclusions

Introduction

The Waxhaw area was chosen for this investigation because it was reported to contain a sinistral shear zone (DBsz) that was potentially coeval with the GHfz (Boland, 1996). As discussed in previous chapters, detailed mapping has failed to identify the DBsz. This study has therefore focused on the nature and timing of deformation within the GHfz in order to revise the current deformational history of the study area and shed new light on the structural and metamorphic history of the GHfz. This chapter reconstructs the geologic history of the Waxhaw, North Carolina area based on interpretations derived from the lithostratigraphic, structural, and metamorphic features outlined in the previous chapters. The following geologic history will describe the major features associated with the study area, as well as the timing of deformational and metamorphic events in the Waxhaw area from oldest to youngest. In addition, some of the major observations and conclusions of this study will be discussed in the context of their regional significance with respect to the tectonic history of the southern Appalachians. The chapter will conclude by outlining the major conclusions of this study and briefly discuss remaining unanswered questions concerning the geology of the Waxhaw area.

Geologic History of the Waxhaw, North Carolina Area

The oldest rocks in the study area belong to the Marvin sequence, which is predominantly composed of mildly deformed quartz dacites and associated shallow intrusions along with mafic-derived sedimentary rocks that have locally been
recrystallized into chloritic schist. The Marvin sequence is similar in appearance, composition, and structural position to volcanic sequences around Mt. Pleasant, North Carolina, which have been dated to ca. 615 Ma (Miller, pers. comm., 2005), suggesting the Marvin sequence is of a similar age.

The Twelve Mile sequence comprises the bulk of the GHfz and can be divided into two separate units: a felsic volcanic unit and sedimentary unit. The majority of the felsic volcanic unit is composed of unseparated rhyolitic to rhyodacitic crystal tuff, lapilli tuff, and crystal lithic tuff that are interlayered with lesser amounts of andesitic and basaltic tuff. The sedimentary unit includes all unseparated mudstones, sandstones, and conglomerates in the Twelve Mile sequence that lie disconformably above the felsic volcanic unit. Deposition of the Twelve Mile sequence is interpreted to have occurred ca. 560 based on stratigraphic and structural similarities between it and the Uhwarrie and Tillery Formations in the Albemarle Group in central North Carolina.

Plutonic rocks that intrude the southeastern portion of the Twelve Mile sequence include the Waxhaw and Hancock granites. The relationship between the Hancock granite and Waxhaw granite is unclear; however, the proximity of the Hancock granite to the Waxhaw granite suggest it could be genetically related to the Waxhaw granite despite variances in mineralogy and morphology between the two. The Waxhaw granite has yielded a Rb-Sr whole rock age of 495±21 Ma (Fullagar, 1982); however, this age is considered tenuous at best because of the known discrepancies regarding Rb-Sr methods on deformed rocks and because the
crystallization age does not fit any known tectonic model for peri-Gondwanan terranes in the southern Appalachians.

The emplacement of the Waxhaw granite resulted in the local metamorphism (M\textsubscript{C}) of semi-pelitic and volcanic rock in the Twelve Mile sequence. The resultant metamorphic aureole around the northwest flank of the pluton extends up to one kilometer into the Twelve Mile sequence and progresses from a sillimanite-in isograd in close proximity to the granite to an andalusite-in isograd around the perimeter of the aureole. M\textsubscript{C} conditions were also responsible for the formation of gedrite-cordierite assemblages in mafic tuff within the volcanic unit. Temperatures during peak M\textsubscript{C} conditions within the sillimanite and gedrite isograds exceeded 600\degree C while temperatures in between the gedrite and andalusite isograds did not exceed 550\degree C. Pressures during M\textsubscript{C} ranged from 2.0 to 2.5 kbars.

Rock units in the study area record a significant brittle-ductile tectonothermal event that was responsible for D\textsubscript{M}/D\textsubscript{L} deformational phases as well as an accompanying lower greenschist facies metamorphic event, M\textsubscript{M}. The timing of deformation is constrained by the emplacement of the Waxhaw pluton. Metamorphic porphyroblasts between the andalusite and gedrite isograds are deformed by S\textsubscript{M} yet also contain inclusion trails parallel to S\textsubscript{M}, whereas porphyroblasts in the sillimanite isograd possess both randomly oriented as well as foliation parallel inclusion trails. The microtextural evidence seems to indicate that emplacement of the Waxhaw granite and subsequent M\textsubscript{C} metamorphism is synkinematic to late-synkinematic with respect to the regional D\textsubscript{M}/D\textsubscript{L} deformational episodes.
$D_M$ is interpreted as the older phase of the two progressive events and is responsible for the most noticeable penetrative foliation ($S_M$) in the study area, which is observed in all pre-Mesozoic rock units. In the Twelve Mile and Marvin sequences, $S_M$ is a prominent steep dipping pressure solution cleavage that is axial planar to folded sedimentary and volcanic bedding ($F_M$). $S_M$ also forms a steep dipping foliation in the Waxhaw and Hancock granites. Other structures attributed to $D_M$ include boudinage of volcanic layering and mineral veins and a set of mineral stretching lineations ($L_M$) within the GHfz that vary in plunge from subhorizontal to subvertical. The orientation of asymmetric fabrics and winged objects associated with $D_M$ consistently displays dextral shear coupled with northwest over southeast reverse motion.

$D_L$ is the younger phase of deformation, and it affects all pre-Mesozoic rocks in the GHfz only. It is responsible for isoclinal folds ($F_L$) that fold $S_M$, an associated late foliation ($S_L$) that overprints $S_M$ and is likely axial planar to $F_L$, and a set of locally penetrative shear band cleavages ($S_{sb}$). The vergence of $F_L$ folds as well as the geometry of the late-forming $S_{sb}$ consistently display dextral shear.

The timing of $D_M$ with respect to $D_L$ is vague but there is structural evidence to support the interpretation that both were part of a single progressive deformation. First, minerals that define both $D_M$ and $D_L$ fabrics formed under similar metamorphic conditions. Second, kinematic indicators from both events consistently display the same sense of motion within the GHfz. Third, $D_M$ and $D_L$ structural features have the same orientation indicating they formed during the same event. Fourth, microtextural relationships observed in the Waxhaw contact aureole indicate the field area only
experienced a single deformational event. Fifth, structures attributed to D_L overprint D_M structures. These observations indicate that although D_M and D_L were phases of the same progressive deformation, D_L outlasted D_M.

The progressive D_M/D_L deformational episode is interpreted to be the result of dextral wrench-dominated tranpressional shear in the vicinity of the GHf. The result of progressive transpression was the spread of deformation from the GHf into the footwall Twelve Mile sequence and hanging wall Marvin sequence, thus forming the GHfz. Transpressional deformation within the GHfz is indicated by the following lines of evidence: 1) the coexistence of subvertical and subhorizontal mineral stretching lineations in the GHfz; 2) the change in attitude of strain ellipsoids across the GHfz is consistent with models for transpression; 3) the dominant strain recorded in the GHfz is apparent flattening; 4) the asymmetry of fabrics and winged objects suggest the dominant sense of motion along the GHf was dextral-reverse-oblique shear that involved thrusting of the Marvin sequence over the Twelve Mile sequence; 5) all fabrics in rocks associated with the GHfz are defined by minerals that formed at similar metamorphic conditions. Furthermore, retrograde metamorphic mineral assemblages within the Waxhaw contact aureole record an increase in lithostatic pressure in the Twelve Mile sequence during D_M/D_L progressive deformation, which is best explained by thrusting the Marvin sequence onto the Twelve Mile sequence.

A regional metamorphic event (M_M) is responsible for prograde metamorphic assemblages in the Twelve Mile and Marvin sequences and retrograde assemblages in rocks within the Waxhaw contact aureole. M_M mineral assemblages define D_M and D_L fabrics, which indicate that regional metamorphism was coeval with regional
deformation ($D_M/D_L$). Peak conditions during $M_M$ are interpreted to be 350-400°C and 3-4 kbars, which corresponds to lower greenschist facies. The increase in pressure and subsequent cooling of the Waxhaw contact aureole during the regional metamorphic episode allowed for the growth and replacement of the $M_C$ prograde assemblages in the contact aureole by $M_M$ retrograde assemblages.

Outcrop and map scale brittle faults ($D_f$) affect all of the units within the study area with the exception of diabase dikes and gabbro. These faults do not appear to record any significant displacements. The trend of brittle faults is roughly parallel to the trend of diabase dikes, suggesting that $D_f$ is related to Mesozoic extension.

The youngest rocks in the Waxhaw area are Mesozoic diabase dikes and a suite of undeformed, unaltered gabbro that is interpreted to be Mesozoic in age. Diabase occurs in long, narrow dikes that cross-cut all other units and trend undeflected across the GHfz at an orientation of approximately N25ºE. Although diabase and gabbro do not contribute significantly to the regional geology, their unaltered and undeformed nature indicates that no significant displacements have occurred along the GHfz after their emplacement during the Mesozoic.

**Regional Implications**

The GHfz is a first order structure located within the peri-Gondwanan Carolina zone and is interpreted to juxtapose Neoproterozoic metavolcanic rocks within the Carolina terrane. Accretion of the Carolina zone to Laurentia is poorly understood, mainly because Alleghanian shortening along the central Piedmont shear zone has obscured the original suture between the Piedmont and Carolina zones. The
GHfz, however, appears to predate Alleghanian deformation (Hibbard et al., 2002, Hibbard et al., 2003, Miller et al., 2003, Standard, 2003) and may therefore be related to the docking of the Carolina zone (Hibbard et al., 2000).

There are currently two competing models for the accretion of the Carolina zone to Laurentia. One model involves Late Devonian to Early Mississippian dextral-oblique convergence of the Carolina zone to Laurentia with subduction of the Piedmont zone beneath the Carolina zone (e.g. Hatcher et al., 1999, Bream et al., 2000, Hatcher et al., 2005). The other model calls for Late Ordovician sinistral-oblique docking to the orogen with subduction of the Carolina zone beneath the Piedmont zone (Hibbard, 2000). If the deformation within GHfz records the collision between the Carolina zone and the Laurentian margin, then its structural history can provide some insight to the accretionary history of the Carolina zone.

Previous work in central North Carolina indicates the GHfz is a sinistral reverse fault system with southeast vergence (Standard et al., 2003, Hibbard et al., 2003, Standard, 2003). Age dating in this region has revealed the zone was active throughout the Paleozoic; however, earliest motion along the GHfz appears to be coeval with a set of Late Ordovician counter-clockwise en echelon regional folds in the Albemarle Group. These reports support the model for sinistral-oblique accretion of the Carolina zone to the Laurentian margin. However, kinematic reports from central North Carolina conflict with the kinematics of GHfz in the Waxhaw area, as well as other areas in northern South Carolina (Lawrence, 2001). As previously described, the GHfz in the Waxhaw study area comprises a dextral wrench-dominated transpressive damage zone in both the hanging wall and footwall of the GHf.
There are some similarities between the GHf in the Waxhaw area and in central North Carolina. Around High Rock Lake and Mt. Pleasant, North Carolina, the GHf is reported as a steep dipping, southeast vergent reverse fault that thrusts mildly and moderately deformed volcanioclastic strata over the Albemarle sequence. This reported geometry is similar to that of the GHf in the Waxhaw area except that lateral motion along the fault in the Waxhaw study area is dextral rather than sinistral. Also, rocks in the hanging wall of the GHf in both central and south-central North Carolina are characteristic of the Virgilina sequence rather than the Charlotte terrane on the basis of age, composition, and metamorphic grade. If the Virgilina sequence occurs in the hanging wall of the GHf, then the GHf can not constitute a terrane boundary as previously interpreted by other workers (e.g. Goldsmith et al., 1988, Butler and Secor, 1991).

The timing of deformation between the GHfz in central North Carolina and the Waxhaw area presents a significant issue. In central North Carolina, deformation in the fault zone is fairly well constrained to the Late Ordovician based upon $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages from micas sampled from axial planar cleavages to synkinematic folds in the Albemarle Group. In the Waxhaw study area, however, the timing of deformation is constrained by the emplacement of the Waxhaw pluton, the age of which is tenuously constrained at best. Because the available Rb-Sr whole rock age of $495\pm21$ Ma for the Waxhaw pluton is likely not geologically significant, several alternate ages are here suggested along with supporting evidence for each age and there implications for deformation in the GHfz.
(1) *The Waxhaw pluton is Late Neoproterozoic in age.* The Waxhaw pluton is bounded to the southwest by ductily deformed amphibolites attributed to the Charlotte terrane and contains xenoliths of similar amphibolites along its western margin (Boland, 1996). However, low-grade volcanic slates of the Carolina terrane bound the pluton to the northwest, east, and north, indicating the Waxhaw granite is a stitching pluton between the two terranes. Furthermore, the Waxhaw granite is synkinematic to a regional deformation that has affected both the Charlotte and Carolina terranes (Boland, 1996). These relationships are similar to those of the Longtown metagranite in north-central, South Carolina, which is a late Neoproterozoic synkinematic stitching pluton between the Charlotte and Carolina terranes in that region (Barker et al., 1998). If the Waxhaw granite is late-Neoproterozoic, then deformation in the GHfz may record the collision of the Carolina terrane with the Charlotte terrane in a manner similar to deformation recorded by the Longtown metagranite (Barker et al., 1998).

(2) *The Waxhaw pluton is Late Ordovician in age.* The Waxhaw pluton is synkinematic to deformation within the GHfz, which in central North Carolina has been constrained to the Late Ordovician. If the GHfz in the Waxhaw area is analogous to the GHfz in central North Carolina, then it implies that the Waxhaw pluton is Late Ordovician in age, as well. This is difficult to reconcile, however, because there are currently no known Late Ordovician plutons in the Carolina zone.

(3) *The Waxhaw pluton is Carboniferous in age.* The Waxhaw pluton is only mildly to moderately deformed, which is common for many Carboniferous plutons in the Carolina zone. There are several granitic plutons located near the study area in
northern South Carolina that are Early to Middle Pennsylvanian in age, namely the Catawba pluton and the Liberty Hill plutons. Furthermore, the Liberty Hill pluton in northern South Carolina has locally metamorphosed volcanic rocks in the Carolina terrane (Speer, 1981). Interestingly, Middle to Late Carboniferous and Early Permian $^{40}$Ar/$^{39}$Ar cooling ages have been obtained from phyllite and gneiss deformed by the GHfz in northern, South Carolina (Lawerence 2001, Lavallee 2003), suggesting the GHfz is an Alleghanian fault zone.

A geochronologist at Texas A&M is currently working to obtain more precise U-Pb zircon ages from the Waxhaw pluton to resolve this issue. However, until a more precise age is forthcoming, the deformational history of the GHfz in south-central North Carolina must remain a matter of speculation. To this end, an Alleghanian age for deformation is the simplest interpretation when all of the geologic evidence is taken into consideration. The basis for this interpretation is Late Mississippian and Early Pennsylvanian cooling ages gathered from the fault zone in several locations across North and South Carolina and because the kinematics of the GHfz are similar to other known Alleghanian shear zones in the southern Appalachians.

There are some discrepancies with the $^{40}$Ar/$^{39}$Ar cooling ages obtained from within the GHfz. Immediately southwest of the study area, Boland and Dallmeyer (1997) obtained Middle to Late Devonian cooling ages from amphibolites deformed by the GHfz, while two independent studies in northern South Carolina obtained Carboniferous to Early Permian cooling ages from the fault zone (Lawerence 2001, Lavallee 2003). Lavallee (2003) also reported Early Carboniferous ages in the GHfz.
in central North Carolina. Thus, it appears the GHfz may have experienced several separate deformational episodes; however, the tectonic fabric recorded in the Waxhaw study area suggests that a single progressive deformational event ($D_M/D_L$) occurred in that region.

One possible explanation for the distribution of $^{40}$Ar/$^{39}$Ar cooling ages obtained from an apparently single tectonic fabric is that rocks in the GHfz experienced a series of coaxial deformational episodes. This interpretation accounts for multiple cooling ages obtained from rocks that possess only one ‘apparent’ fabric. Another possibility is that $D_M$ and $D_L$ are two unrelated episodes of deformation within the GHfz with $D_M$ representing a Middle to Late Devonian event and $D_L$ representing Alleghanian reactivation of the fault zone. However, structural and metamorphic evidence suggests that $D_M$ and $D_L$ are end-member phases of a single deformational event.

Interestingly, the Middle and Late Devonian ages obtained from the GHfz seem to imply deformation occurred within the fault zone in a manner compatible with the model for Late Devonian accretion of the Carolina zone proposed by Hatcher et al. (1999). However, this model is difficult to reconcile because of the lack of 1) Middle to Late Devonian magmatism in the Carolina zone, which is the purported upper plate; 2) a Devonian fore-land basin in the Carolina zone, which would suggest the encroachment and subduction of a large crustal block beneath the Carolina zone; 3) other structures within the Carolina zone that can be tied to the collision and subduction of the Piedmont zone beneath the Carolina zone. If the Late Devonian model is not correct, then Middle and Late Devonian ages obtained within the GHfz
may reflect hybrid ages resulting from partial Alleghanian resetting of Late Ordovician ages.

In summary, the GHfz in south-central North Carolina exhibits the characteristics of an Alleghanian transpressional fault zone based on its structure, kinematics, and \(^{40}\text{Ar}/^{39}\text{Ar}\) ages collected within the fault zone. If the GHfz in central North Carolina is analogous to the GHfz in the Waxhaw area, then this interpretation implies reactivation of the GHfz during the Alleghanian orogeny. An interpretation for Alleghanian reactivation is further supported by Early to Middle Carboniferous \(^{40}\text{Ar}/^{39}\text{Ar}\) cooling ages obtained from the GHfz around High Rock Lake, North Carolina (Lavallee, 2003). This new interpretation of the GHfz does not explain the absence of sinistral shear sense indicators in the Waxhaw area; however, whatever the reason for the distribution of dextral and sinistral shear sense indicators across the GHfz, resolution of this problem is likely related to its curved surface trace around the Weddington Flexure.

**Conclusions**

The major conclusions of this study are:

1) This study documents the southern continuation of the Gold Hill fault zone from central North Carolina into the Waxhaw area and helps confirm that the Gold Hill fault zone is not a wide zone of recrystallization and ductile shear as described by previous workers. Rather, it comprises a wide brittle-ductile footwall damage zone and narrow hanging wall damage zone to the Gold Hill fault. This study also found
that the Gold Hill fault does not constitute the western boundary to the Gold Hill fault zone in the Waxhaw study area but instead the fault lies within the Gold Hill fault zone.

2) Mildly deformed and metamorphosed volcanioclastic sediments that lie on both sides of the Gold Hill fault zone indicate that it is likely not the boundary between the Charlotte and Carolina terranes as previously inferred.

3) Structural mapping and strain analysis in the Waxhaw area was unable to identify the existence of the Davis Branch shear zone around its previously reported location southeast of the Waxhaw granite.

4) The Gold Hill fault zone in Waxhaw, North Carolina is a dextral transpressive fault system. No evidence of sinistral asymmetric fabrics as described in central North Carolina was found. Kinematic and metamorphic data gathered from the study area suggest that movement along the Gold Hill fault zone occurred during a progressive deformation that involved southeast thrusting of the Marvin sequence over the Twelve Mile sequence coupled with a component of dextral strike-slip motion.
Future Research

This study has provided new observations concerning the timing and kinematics of structures attributed to the Gold Hill fault zone. Although many questions have been answered during this study, some questions remain unanswered and many new questions have been generated. Listed below are topics that need further investigation in order to confirm interpretations of this study and to help better understand the geology of the Waxhaw area.

1) The age of the Waxhaw granite is important to constraining the exact timing of \( D_M / D_L \) progressive deformation in the Waxhaw area. Though a crystallization age exists for the granite, it is likely invalid. What is the age of the Waxhaw granite and how does that age compare with other deformational ages from the GHfz?

2) Several studies have produced Late Devonian and Early Mississippian \(^{40}\text{Ar}/^{39}\text{Ar}\) cooling ages from white mica within the GHfz. Are they geologically significant and if so what do they represent?

3) The GHfz in the Waxhaw study area is a dextral reverse fault zone, whereas in High Rock Lake and Mt. Pleasant, North Carolina, the GHfz is a sinistral reverse fault zone. Is the GHfz in these two regions kinematically related or do they represent two different faults with different deformational histories? To this end, does the Weddington Flexure represent a continuous feature about which structures in
the Carolina zone are folded or is the GHfz in Waxhaw, North Carolina a younger fault that truncates the GHfz mapped in central North Carolina?

4) Mildly deformed volcaniclastic rocks occur on both sides of the Gold Hill fault indicating that it is not the terrane boundary between the Charlotte and Carolina terranes. However, several workers have indicated the boundary between the two terranes is tectonic. Where is the Charlotte-Carolina terrane boundary and what is its relationship to the Gold Hill fault zone, if any?
References Cited.


Boland, I., 1990. The geology of the southern half of Rock Hill East and the northern half of Catawba quadrangles, South Carolina. MS Thesis, The University of South Carolina, 118 p.


Bream, B., Hatcher, R., Miller, C., Fullagar, P., 2000. Paragenesis, geochemistry, and preliminary ion microprobe geochronology of detrital zircons from the southern


Geologic Map of the Waxhaw Area

within portions of the Catawba NE and Waxhaw USGS 7.5-minute quadrangles
Mapped by John Stefan Allen 2003-2005

Geologic Cross-section of the Waxhaw Area, N.C.-S.C. from A to A’
Dashed lines indicate bedding; Wavy lines indicate structural trends

Figure 2.