THE VARISCAN OROGEN IN SARDINIA, ITALY:
INSIGHTS INTO LATE-STAGE DEFORMATION
AS EVIDENCED IN THE LAGO MULARGIA REGION

by

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ABSTRACT

The late Paleozoic Variscan Orogen of Europe represents one member of a composite Mountain belt (including the Allegenian, Southern Appalachan, and Mauratanides among others) that formed as a response to the closure of both the Rheic and Galacia/Southern Brittany oceans between the Gondwonan and Laurentian continents, marking the final amalgamation phase of Pangaea. Field investigations as well as data from bore holes has resolved that rocks deformed during the Variscan orogeny can be delineated into six tectonic zones (Figure 4). Of particular interest to this study are the Rock of the Southern European Variscides, the most intact (minimal Pyrenean and/or Alpine deformation) section of which is located throughout the Mediterranean island of Sardinia, Italy. Variscan rocks exposed in Sardinia have been further subdivided into four distinct tectono/metamorphic zones; Mid to High Grade metamorphic rocks of the Internal zone in the north, Internal Nappes bounded by thick (~10-30m) mylonitic boundaries, External Nappes delineated by more concise nappe boundaries, and the External zone which is identified by Carmignani (1992) classic fold-and-thrust belt.

To investigate timing of deformational features developed within Variscan rocks of Sardinia this paper examines rocks contained within the External Nappes, outcropping in
the Lago Mulargia region near the villages of Siurgus/Donigala. A regional deformational sequence of events has been identified by Conti (1999) primarily consisting of a SSW shortening event followed by one 90° shift in shortening direction and again returning to a SSW directed shortening phase. Field investigations in the region along with previous mapped geology (Funedda et. al. 2010) illustrated that deformational structures in this region do not entirely fit within the context of the sequence of events identified by Conti (1999). In this study these structures are analyzed and placed within a locally defined deformational sequence, this local sequence is then placed into the context previously laid out by Conti (1999) and a new composite deformational history is created that takes into account the mapped relationships of deformational features within the Lago Mulargia region.
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The Variscan Orogen of Europe represents a segment of a 1000 K long mountain belt resulting from the collision of the Larussian continental mass with the Gondwanan continental mass (Matte, 1991). This composite mountain belt extends from the western Gulf of Mexico to Eastern Europe and the Black Sea, including the Alleghanian and Ouachita belts of the Southern Appalachians, and the Mauritanides of West Africa. Continental collision that resulted in the Variscan segment took place in the Devonian to Late Carboniferous and consisted in the closure of two large ocean basins and the amalgamation of at least two smaller continental land masses, Avalonia and Armorica, to the north and south respectively (Matte, 2001; McKerrow et al., 2000). The striking curvilinear shape of the Variscan mountain belt is thought to be due to the presence of an “African Promontory,” which acted as an indentor to the southern Larussian margin (Matte, 1991; Windley, 1995).

The Variscan Orogen of Europe consists of five distinct terrains, from North to south these are the Northern Foreland, The Rhenohercynian zone, The Saxo-Thuringian zone, The Tepla-Barrandian zone, The Moldanubian Zone, and the Southern European Variscides (or Hunic Terrains by some authors) (Stampfli et al., 2002; Windley, 1995; Matte, 1991, 2001). Of these zones, the least understood are the zones that make up the southern margin of the Variscan Orogen, whether it is interpreted as the Southern European Variscides or the Hunic Terrains. Investigations into the southern flank of the
Variscan Orogen can be carried out by examining rocks located on both the islands of Corsica (FR) and Sardinia (IT), which in contrast to other preserved Variscan age rocks in Europe show little to no later Alpine or Pyrenean deformation. The Variscan age rocks in Sardinia are of particular interest due to the fact that a complete tectonostratigraphic sequence, ranging from allochthonous high grade amphibolites in the north to autochthonous metamorphic rocks in the south, is exposed.

In Sardinia, Variscan rocks can be subdivided into three general categories with a number of prominent regional geological structures. In the northeast, the majority of rocks belong to the High-Grade Metamorphic Complex (HGMC) and are characterized by amphibolite metamorphic grade and referred to collectively as the “Internal Zone.” The southern margin of the Internal Zone is defined by the Posada Asinara Line (PAL), which has been interpreted as either the southern boundary of the Armorican plate or part of a complex series of anatomizing sutures that bound many small continental landmasses (the Hunic Terrains). To the south of the PAL, and comprising the aerial majority of Variscan rocks, the Nappe Zone consists of a series of allochthonous thrust sheets. In the southern region of Sardinia, Variscan rocks are low to non-metamorphosed. Structures found in these rocks consist of steeply dipping thrusts and fold axial planes and are collectively referred to as the “External Zone” or Variscan Fold-and-Thrust belt.

Investigations of the deformational history carried out in the Nappe zone reveal that the majority of deformation reflects NNE-SSW shortening but also includes two 90° shifts in shortening direction. Conti et al. (2001) describes these as: D1a (Gerrei Phase), SSW-directed thusting and folding; D1b (Meana Phase), main SSW-directed nappe emplacement; D1c (Sarrabus Phase) W-directed nappe emplacement; and D1d
(Flumendosa Phase) SSW directed folding. Following these contractional phases, all Variscan rocks underwent thinning by normal faulting, NW-SE directed folding, and/or crenulation development, all part of D2 (Rio Gruppa Phase) deformation.

In the Lago Mulargia area of southern central Sardinia, three of the four main tectonostratigraphic units of the nappe zone crop out. Primary structural features in the area include N-S striking Nappe emplacing faults and isoclinal folds trending 90°-140°, N-S striking normal faults with top to the west deformed porphyroclasts, as well as multiple generations of folds striking both N-S and E-W. Mapped field relationships of these structures within the Lago Mulargia area show a more complicated temporal relationship than that proposed for the region by previous authors (Conti et al., 2001). This paper will identify and analyze these Late Paleozoic structures and establish a model that not only locally describes the deformational sequence of events but also places these within the context of a more regionally identified deformational sequence.
THE VARISCAN OROGEN OF EUROPE: OVERVIEW

The Variscan Orogen of Europe is a portion of a 1000 km wide and 8000 km long
Paleozoic mountain belt that played an integral role in the formation of Pangaea (Figure
1; Matte, 1991). Remnants of this Paleozoic mountain belt extend from the western Gulf
of Mexico to Eastern Europe and the Black Sea. Other portions of this system include the
Alleghanian and Ouachita belts of the Southern Appalachians, and the Mauritanides of
West Africa (Figure 1). The Variscan segment was formed during the latter stages of the
Late Paleozoic collision of the Larussian continent with the Gondwonan continent. This
collision consisted of the complex and diachronous closure of at least three oceanic
basins, and consolidation of the Armorican (Gondwonan derived) and Avalonian
(Larussian derived) microplates, resulting in the development of several sutures that
bound at least five distinct zones or terrains located in present day Europe (Figure 2;
Matte, 2001; McKerrow et al., 2000). The assembly of these terrains (Rhenohercynian,
Saxo-Thuringian, Tepla-Barrandian, Moldanubian, and the Southern Variscides), and the
associated consolidation of the Armorican and Avalonian microplates (among others)
occurred from the Early Devonian to Late Permian time (Windley, 1995; Matte, 1991,
2001). Nearly all of the pre-Mesozoic basement rocks of Europe were deformed,
metamorphosed, and/or intruded during the Variscan orogenic event.
**Figure 1a.** Paleogeographic reconstruction showing the location of the Variscan Orogen, c.a. 340 Ma, location of Figure 1b shown in yellow: (Blakely, http://jan.ucc.nau.edu/~rcb7/340_1st.jpg)

**Figure 1b.** Post-Permian reconstruction of several members of a composite, late Paleozoic collisional mountain belt, including the Caledonian, Appalachian, Mauritanian, and Variscan Belts. Modified from Catalán (2002)
Figure 2. Hypothetical Paleozoic locations of continental masses involved in the Variscan Orogen from Middle Ordovician (465 Ma) to Early Carboniferous (340 Ma). Avalonia and Armorica highlighted in purple and orange respectively. From Matte, 2001.
Variscan rocks are now exposed in a number of “Basement Massifs” across Europe, including the Iberian, Armorican, Massif Central, Ardennes-Rhenohercynian, and Bohemian that show little to no post Variscan deformation (Figure 3). Drilling in various parts of Europe has also revealed that basement rocks of the Alpine, Pyrenean, and Baltic Cordilleran orogens are comprised of Variscan units (Matte, 1991). Previous workers (Matte, 1991, 2001; Franke, 1989; Franke et al., 2000) have made correlations among these Variscan rocks by reconstructing the position of Iberia and the Corsica Sardinia block to their pre-Mesozoic configurations, linking these stable massifs, and utilizing data from bore holes. This work shows that the European Variscan belt extended 700-800 Km in a N-S direction, 3000 Km in a E-W direction and resolves the locations of these terrains. In contrast to more linear mountain belts (i.e., Appalachians, Urals, North American Cordillera), the European Variscan belt displays a complex and sinuous pattern that is still not well understood (Matte, 2001). The aerial extent of these terrains and locations of sutures is shown in Figure 4 and they will each be discussed in the following sections.

**European Variscan Zones**

The Variscan belt of Europe consists of six main zones (Figure 4). These are, from north to south, the Northern Foreland, the Rhenohercynian Zone, the Saxo-Thuringian Zone, the Tepla-Barrandian, the Moldanubian Zone, and the Southern European Variscides. Each of these zones represent a different anatomical/temporal aspect of the Variscan Orogen. Some recent works (i.e., Stampfli et al., 2002) have found problems with the discrete boundaries and definitions of these zones (namely the Moldanubian Zone and the Southern European Variscides) and have proposed an
alternative hypothesis, referring to both of them collectively as the Hunic Terrains. Definitions from both hypotheses will be provided in the following sections.

**The Northern Foreland**

The Northern Foreland represents the northern-most reaches of Variscan deformation and consists of classical foreland fold-and-thrust deformation. The northern boundary is delineated by the northward decrease of Variscan deformation overprinting structures developed during the Caledonian orogenic event (540-408 Ma; Matte, 1991). The southern boundary of the Northern Foreland zone is delineated by a series of south dipping thrust ramps that flatten out and converge to form one major roof thrust striking approximately 110º stretching from the Celtic sea, off of the southern coast of Ireland, to the Hartz Mountains of eastern Germany (Figure 4; Le Gall, 1990). This structure may have acted as the main decollement between the crustal stacking wedge, consisting of Cambrian to Lower Carboniferous rocks over Middle to Upper Carboniferous flysh and mollasse deposits, over the Precambrian Caledonian basement, which incurred only mild folding during the Variscan event (Le Gall, 1990; Jones, 1991; Matte, 1991). To the east, the mountain belt is truncated by a system of dextral transcurrent faults (Elbe and Tornquist faults) and may continue to the east, connecting with the Dobrogea and Variscan Caucasua (Matte, 1991, and references therein).

**Rhenohercynian Zone**

The northernmost allochthonous zone of the Variscan system is the Rhenohercynian zone, which reflects the structural telescoping of a 300 km wide Silurian - Early Carboniferous extensional basin (Figure 4; Windley 1995; Franke, 1989). The Rhenohercynian basin remained relatively stable for much of the Paleozoic, with 5-10 km
Figure 3. Exposed Variscan Massifs in Southwestern and Central Europe used to constrain and examine the Variscan belt in Europe (Sardinia and Corsica rotated to their interpreted Carboniferous position). After Robardet, 2002.
thick sequences of Devonian elastic and bimodal magmatic rocks overlain by Late Devonian to Upper Carboniferous flysch deposits, which may reflect the prograding front of the Variscan belt (Behrmann et al., 1991). Bounded to the north by the Northern Variscan Front, the Rhenohercynian zone contains northward directed (south dipping) nappes and thrusts that accommodated ~200 km of horizontal shortening (Windley, 1995). The largest of these, the Giessen Nappe, contains MORB-type volcanics in the structurally lowest thrust sheets and Mid – Late Devonian flysch units in the structurally highest sheets (Behrmann et al., 1991). Seismic reflectors, which may represent ophiolitic material and phengite barometry of phyllites indicating pressures of 10 – 12 kbar, illustrate the presence of the Lizard – Rhenish suture (L.R.H.S. in Figure 4). Kinematic studies of the Lizard – Rhenish suture in the Rhenish massif, which marks the southern boundary of the terrain, indicate oblique convergence that produced dextral movement along this zone in the Late Carboniferous (Behrmann et al., 1991; references therein). Matte (1991, Figure 4) proposed that the Beja suture in the Iberian massif may be a continuation of the L.R.H.S.

**Saxo-Thuringian Zone**

Cropping out in the northern section of the Bohemian massif, southern portion of the Ardennes-Rhenohercynian massif, and the northern extremities of the Schwarzwald and Vosges massifs is the Saxo-Thuringian Zone (Figure 4; Robardet, 2002; von Raumer et al., 2003). This terrain is characterized by a pervasive flat-lying foliation within high grade metasedimentary and plutonic rocks of Late Precambrian to Early Paleozoic age and Lower Paleozoic sedimentary rocks affected by Variscan metamorphism (Matte, 1991). In the north, this terrain is expressed as the Mid-German Crystalline Rise
(MGCR), composed of metamorphosed Paleozoic sediments intruded by Silurian S-type granitoids (Franke, 1989; references therein). South of the MGCR, the Cambro-Ordovician development and subsequent infill of the Saxo-Thuringian basin resulted in thick sections of marine clastic rocks interbedded with bimodal volcanic rocks (Franke, 1989). The Devonian – Early Carboniferous closure of this basin formed several north-directed thrusts resulting in several inverted thrust nappes such as the Munchenberg, and lenses of 13-kbar eclogites (Windley, 1995). This unit is bounded to the north by the L.H.R.S., previously discussed, and to the south by the Tepla suture, inferred by the presence of allochthonous MORB-type rocks located at the boundary of the Saxo-Thuringian and the Tepla-Barrandian zones.

Tepla - Barrandian Zone

The Tepla - Barrandian zone, located south of the Saxo-Thuringian zone (Figure 4), contains the most complete assemblage of pre-Variscan, Cadomian basement rocks (Franke, 1989). These basement rocks include Proterozoic flysch deposits and arc-related volcanic materials that reflect Early – Mid Cambrian deformation (Franke, 1989; Franke et al., 2000). Along parts of the northern and southern margins of this terrain, MORB-type rocks in the Cadomian basement units, now altered to eclogite, suggest that a Cambro – Ordovician extensional phase reached at least the initial stages of oceanization (Franke, 1989). In the center of the Tepla-Barrandian zone, these Precambrian rocks are overlain by thin packages of Cambrian to Mid-Devonian epicontinental sedimentary and volcanic rocks that show increasing southward and northward Variscan deformation (Franke et al., 2000). Variscan thrusting has transported Cadomian basement rocks northward as least as far as the Munchenburg nappe of the Saxo-Thuringian zone,
whereas nappes in the southern portion of this terrain are south-directed (Franke, 1989). This dualism of transport direction has been interpreted to reflect the median zone of the Variscan Orogen, partitioning northward and southward directed structural features (Franke, 1989). To the south, the Tepla-Barrandian zone is bounded by the Ligerian (south Brittany)- Massif Central- Moldanubian (south Bohemia) suture (L.M.C.M.), which delineates the separation of numerous microplates from Gondwana in the Ordovician (McKerrow et al., 2000).

**The Moldanubian Zone**

Occupying the southern majority of the Bohemian Massif and delimited to the north by the L.M.C.M., the Moldanubian zone consists of two distinct tectonostratigraphic units, the Gföhl unit and the Drosendorf unit (Figure 4; Matte, 1991). These units are composed of high-grade metamorphic rocks with poorly constrained protoliths (Franke et al., 2000). The lower Drosendorf unit is made up of mainly metasedimentary and felsic meta-igneous rocks with rare lenses of eclogite structurally overlain via a ductile shear zone by the “Bunte Serie,” a package of rocks consisting of a basal sheet of orthogneiss and a wide variety of meta-sedimentary and meta-volcanic rocks (Franke et al., 2000). The Gföhl unit is a 2000 m thick thrust sheet composed mainly of ortho and paragneisses, amphibolites, metagabbros, acid and basic granulites, and peridotites (Matte, 1991). The Main Moldanubian Thrust (MMT), which placed the Gföhl unit over the Drosendorf unit, is expressed as slivers of eclogites, granulites, and HP ultramafic rocks separating the two units (Matte, 1991). NW-SE stretch lineation and shear sense indicators proximal to the MMT show southeastward transport at high-grade (amphibolite facies) conditions (Matte, 1991).
Hunic Terrains

The zonation of the southern European portion of the Variscan system has been problematic and is interpreted in various ways by different authors (Matte, 1986, 1991, 2001, 2002; Franke, 1989; Franke et al., 2000; von Raumber et al., 2002, 2003; McKerrow et al., 2000). The Hunic terrain hypothesis put forth by Stampfli et al. (2002) has been accepted by a growing number of workers in the Variscan belt as an explanation for the complexity of this portion of the mountain belt (Giacomini et al., 2006; Franceschelli et al., 2005; von Raumber et al., 2003).

The Hun terrains (or collectively, “Hun superterrain”) include all continental fragments derived from the Gondwana mainland (including the previously described Tepla-Barrandian and Moldanubian zones) and are believed to have been in lateral continuity with the Armorican block (Figures 3 and 5; Stampfli et al., 2002). This group of terrains is now expressed, along with Armorican remnants, as a ribbon-like assemblage of basement rocks throughout southern Europe (Figure 5, Stampfli et al., 2002). Sedimentological and biostratigraphic data (Paris and Robardet, 1990) and paleomagnetic data from the eastern European Hunic terrains (Schätz et al., 1997; Krs and Pruner, 1999 as referenced in Stampfli et al., 2002) suggest that these terrains were derived from the northern African margin and that no major separation between Armorica and Gondwana existed before the Early Devonian (Stampfli et al., 2002).

Syn-rift and pre-rift sequences present in components of the Hun superterrain are correlative to sequences on the northern Gondwana margin. These include Late Ordovician clastic rocks with minor interbedded volcanic packages, representing syn-rift formations, and Early Silurian graptolitic shales that show widening of the ocean basin
Figure 5. Configuration of the Hun superterrain and individual terrains therein, heavy dashed line shows the proposed southern boundary (Corsica-Sardinia block shown in red). From Stampfli, 2002.
Post-rift evolution of these terrains (primarily timing of metamorphism and flysch development) varies greatly and is mostly a function of their proximity to either the Lizard-Rhenish suture in the north or the paleo-Tethys suture to the south (Stampfli et al., 2002). Present day locations of these individual terrains within the Hun superterrain do not easily lend themselves to interpretations of their Late Devonian–Early Carboniferous paleogeography.

Tectono-thermal ages in these blocks range from Middle Devonian metamorphism in some to later Early Carboniferous (Visean) metamorphism in others, with flysch development on a few blocks beginning in the Late Carboniferous (Bashkirian) (Stampfli et al., 2002). This temporal range of deformation/sedimentation shows that not all of the terrains were involved in the initial stages of compression and that deformation was probably concentrated at the leading edge of the Hun superterrain during the formation of the Lizard-Rhenish suture (Stampfli et al., 2002).

**Southern European Variscides**

This thesis focuses on the Southern European Variscides, the most complete section of which crops out on the islands of Sardinia and Corsica (Figure 4). The island of Sardinia displays a complete tectonostratigraphic sequence ranging from the non-metamorphosed, para-autochthonous foreland in the south to a high-grade metamorphic complex in the northeast (Giacomini et al., 2006). The Posada-Asinara line, a thin mylonitic belt containing amphibolite facies micaschists, paragneisses, quartzites, and small isolated bodies of relic granulitic and eclogite facies rocks, is interpreted by some workers to be a southern Variscan suture separating the Gondwana derived Southern
European Variscides to the south from the Moldenubian (or Hun superterrain units) to the north (Carmignani et al., 1994; Stampfli et al., 2002).

South of the Posada-Asinara suture, the Southern European Variscides consist of Cambrian to Early Carboniferous sedimentary rocks, imbricated as nappes that are thrust over the Cambrian-Devonian para-autochthonous foreland in the south. Deformation patterns such as fold vergence, shear sense in mylonites, and stretching lineations within this “nappe zone” show a main tectonic transport direction of NE over SW (Carmignani et al., 1994). Stratigraphic and structural distinctions within the nappe zone allow the separation into two sub units, the “internal nappes,” with metamorphism showing a prograding Barrovian evolution from greenschist facies in the south to intermediate P amphibolite facies near the Posada-Asinara Line, and the "external nappes,” characterized by lower greenschist to non-metamorphosed rocks and the presence of a Middle Ordovician meta-volcanic sequence not present in the internal nappes (Carmignani et al., 1994).

**Collisional History**

**Ocean Formation**

The pre-collisional history of rocks now part of the Variscan orogen is poorly constrained due, in part, to the low abundance of exposed ophiolitic material. The magmatic and sedimentary records suggest that two, possibly three, oceanic basins developed during a Cambro-Ordovician rifting of the northern passive margin of Gondwana (Matte, 1991; Franke, 1989; references therein, Figures 3 and 6). MORB-type rocks exposed both in the Massif Central and the external Western Alps have been dated
Figure 6. Tectonic cartoon showing possible plate movement of continental blocks between 465 and 340 Ma, shown from Iberia to Newfoundland. AV, Avalonia; GSB, Galicia-Southern Brittany ocean; AR(HT), Armorica (Hunic Terrains). After Matte, 2001
at 490–480 Ma. Of the basins separating these rifted blocks, Avalonia to the north and Armorica located between Avalonia and the Gondwana mainland, only the basin between Avalonia and Armorica developed into a wide oceanic basin (Figures 2 and 6; Franke et al., 2000). Similarities in faunal assemblages, sedimentary facies, and paleomagnetic data, along with the absence of arc-related magmatic rocks, suggest that the Galicia-Brittany ocean between Armorica and Gondwana was comparatively small (Windley, 1995). An alternative explanation for these similarities that allows for a large separation between Avalonia and the Gondwana mainland is the distribution of individual members of the Hun superterrain.

In this model, no one large oceanic basin separated Armorica and Gondwana but several smaller oceanic basins collectively formed the separation (Figure 2). Late Ordovician carbonates in southern Scandinavia require a much warmer climate than faunal assemblages present on these terrains and illustrate a large latitudinal separation between the Avalonian microplate and the Laurussia mainland (Franke, 1989).

Collision and Accumulation

Tectonic development of the Variscan orogenic belt was complex and probably reflected the diachronous closure of multiple ocean basins and the amalgamation of at least two microplates between the southern margin of Laurussia and the northern margin of Gondwana, occurring from Devonian to late Carboniferous time (Windley, 1995, Figures 2, 6). The occurrence of early Barrovian metamorphism and crustal melting in Southern Brittany, Massif Central, and western Spain, along the southern suture (CCS and MCS) of the Variscan system suggest that the initiation of collision occurred around 380 Ma (Matte, 1991). The presence of extensional Early Carboniferous intracontinental
basins and the intrusion of post-kinematic S-type granites in the Massif Central provide a minimum age for this deformation at 350-360 Ma (Matte, 1991).

Opposite subduction directions are inferred for the Rheic and Galicia-Southern Brittany oceanic basins based on structural geometry and HP/UHP metamorphism (430 and 360 Ma; Matte, 2001). This subduction resulted in the formation of intracontinental basins and the separation of the Armorica microplate into two sub-units, the Saxo-Thuringian and an un-named southern unit between the Tepla and Central Moldenubian sutures (Matte, 2001). Van der Voo (1988), based on palaeomagnetic data, concluded that large scale geographic separation was developed between Gondwana and southwestern Europe during the Late Devonian. Calc-alkaline volcanic rocks of Devonian age located in the northern portion of the Massif Central and southern Spain strengthen the argument that southward oceanic subduction operated contemporaneously with continental convergence in the Variscan system. Conversely, arguments by other workers (Matte, 1986; Windley, 1995) utilizing tectonic, palaeomagnetic, palaeontological, and palaeobotanical data suggest that oceanic subduction was limited and may have only occurred on small scales. The Hun superterrain hypothesis presented by Stampfli et al. (2002) could again be called upon to produce a larger geographic separation with smaller microplates interspersed, linking these floral and faunal populations. Due to this controversy, paleogeographic reconstructions here are presented as tentative (Figures 2 and 6).

Convergence continued in all three of the oceanic basins during the Late Devonian–Early Carboniferous time. Initiation of collision in these basins is constrained to Early Carboniferous in the Rheno-Hercynian Basin, Late Devonian (pre-Famennian) in
the Saxo-Thuringian Basin, and poorly constrained at pre-340 Ma in the Moldenubian Basin (Franke et al., 2000). This contraction occurred in an overall regime of dextral transpression and accommodated shortening of c. 50% in the Rhenohercynian and Saxo-Thuringian zones, with minimum transport magnitudes of 200 km and 180 km, respectively (Franke et al., 2000; Windley, 1995; Behrmann et al., 1991).

Following the cessation of collision, a series of large-scale low-angle normal faults attenuated the thickened orogenic crust, resulting in the formation of Late Carboniferous and Early Permian intracontinental basins that may have resembled the current day Basin and Range province of the USA (Matte, 1991; Malavieille et al., 1990). Intrusion of granites dated at 290 and 280 Ma represents the last magmatic event associated with the Variscan System, and may be related to thinning and heating of the lower crust (Matte, 1991; references therein).

**Geochronological Constraints**

The timing of tectonic events in the formation of the Variscan Orogen has been interpreted from a variety of geochronological methods, including biostratigraphy and isotope geochronology.

Rifting of the Avalonian microplate from the northern margin of Gondwana and formation of the Iapetus oceanic basin is marked by broad areas of Gondwana-derived detrital sediments and carbonates located within the Gondwana platform (von Raumer et al., 2003). The youngest detrital zircons found in these packages have been dated at 600 Ma and establish a maximum age for rifting of this segment of the Iapetus oceanic basin (Schaltegger and Gebauer, 1999 as referenced in McKerrow et al., 2000). The separation of Avalonia (then accreted to Laurussia) was geographically far enough and temporally
persistent enough to prevent the migration of ostracodes across the basin until Devonian times (McKerrow et al., 2000). Development of the Tournquist Sea is thought to have been contemporaneous with the development of the Iapetus Sea at 600 Ma. Biostratigraphic data suggest that it was closed during the Shelveian Orogeny of western England in the Late Ordovician (McKerrow et al., 2000; references therein).

The rifting of Avalonia from the Gondwana mainland and genesis of the Rheic oceanic basin occurred in the Early Ordovician (Arenigian) as suggested from diverging similarities in faunal assemblages of each continental block (McKerrow et al., 2000, Cocks & Fortey, 1982). U-Pb isotope analysis of Zircons from rocks associated with the closure of the Rheic ocean, outcropping in the Northern Phyllite Zone and the Mid German Crystalline High, all are younger than c. 400 Ma (Franke et al., 2000; references therein). Following the closure of the Iapetus and Tourquist ocean basins (Late Ordovician), the Rheic basin remained between the Armorica-Hunic Terrains and the Avalonian-Laurussia continent until its closure in Early Devonian (Emisian) time (Figure 6, McKerrow et al., 2000). Post-Emisian felsic volcanism in the northern portion of the Rhenish massif, along with the presence of the Giessen, Harz, and Lizard ophiolites, suggests a renewed pulse of extension in the late Emisian concentrated proximal to the Rheic suture, but this potential phase is not well understood (Franke et al., 2000, McKerrow et al., 2000). This late Emisian extension may have played a role in the development of the Hun superterrain proposed by some workers (von Raumer et al., 2003).
Figure 7. Comparison of fault patterns in the European Variscan system (A) and the Himalayan system (B), showing a general curvilinear pattern. From Matte, 1991.
Final closure of these oceanic basins is thought to have occurred in the Late Devonian to Early Carboniferous. Closure of the Rheic ocean is recorded in the lowest sections of the Hörre-Gommern nappe of the Rheno-Hercynian zone; sandstone turbidites containing detrital micas of possible “Caledonian” origin are interbedded with greywacke turbidites derived from the Mid German Crystalline High (Franke et al., 2000). Complimentary to the interbedded turbidites, MGCH derived turbidites of Early Carboniferous (Tournaisian) age are found on the autochthonous foreland in the Harz massif (Franke et al., 2000). Closure of the Saxo-Thuringian basin is bracketed by U-Pb zircon, Ar hornblende and mica ages of 380 and 365 Ma and again Tepla-Barrandian (Hunic Terrain) derived greywacke turbidites of Late Devonian (Famennian) age are found in the Saxo-Thuringian “foreland” (Franke et al., 2000, references therein). Internal to the Hunic terrains, collision in the Moldanubian zone is shown to pre-date the intrusion of the Central Bohemian batholith (c. 340), and Early Carboniferous (late Visean) volcanic and sedimentary rocks in the Vosges Mountains (Franke et al., 2000).

**Geometry**

The cross-sectional “Fan-like” nature of the Variscan belt along with metamorphic and deformational age data show that deformation within the mountain belt migrated from the median zone to both the Northern and Southern forelands between 380-330 Ma and 380-340 Ma respectively (Figure 6; Windley, 1995; Franke et al., 2000). Subduction, collision, and contraction resulting in this bi-vergent orogen with contemporaneous thickening in both forelands is consistent with progressive evolution in zones of continent-continent collision (Matte, 1991).
By far the most prominent large-scale aspect of the Variscan system is the unusual curvilinear shape. Matte (1991) attributes this shape to a promontory of the African (Gondwanan) continent acting as an indentor. Symmetrical distribution of strike-slip faults on either side of the Ibero-Armorican arc has been compared to strike-slip fault distribution in the western syntaxes of the Himalayan system (Figure 7; Matte, 1991; Windley, 1995). This model may help explain the proximity of opposing shortening directions in some portions of the system (Matte, 1991).
THE VARISCAN OROGEN IN SARDINIA, ITALY

The most complete exposure of southern flank of the Variscan system (Southern European Variscides) is located throughout the Corsica-Sardinia continental block. Many rocks observed in mainland Europe known to have been deformed during the Variscan orogenic event have undergone further deformation due either to the Pyrenean Orogeny or the Alpine Orogeny. In contrast, outcroppings of rocks deformed during the Variscan Orogeny found throughout the islands of Corsica and Sardinia show little to no Alpine or Pyrenean deformation, thus providing a unique opportunity to study the development of the southern flank of the Variscan Orogeny without significant overprinting. Although both Corsica and Sardinia display Variscan rocks, the island of Sardinia, in particular, contains a complete tectonostratigraphic sequence ranging from allochthonous high grade amphibolites in the north to autochthonous low grade metamorphic rocks in the south. Because of this remarkably complete section, analysis of the rocks on this island will be used to help gain a better understanding of the development of the Southern Variscan as a whole.

The Posada-Asinara Line (PAL)

The role of the Variscan rocks of Sardinia in the evolution of the Variscan orogenic system is well understood in some respects but still controversial in others, most notably the nature of the Posada-Asinara Line (Figure 8; PAL). Matte (1986, 2001) proposed that
**Figure 8.** Generalized tectonic map of Variscan rocks in Sardinia. Here a complete tectonostratigraphic sequence from NNE-SSW is visible. From Carosi & Palmeri., 2002
Figure 9. Location of the PAL (bottom right) in relation to the Central European Variscan Belt, as interpreted by Matte, 2001. (Corsica and Sardinia are rotated to their interpreted Carboniferous position.) See Figure 4 for legend.

Figure 10. Corsica-Sardinia continental block (red) within the Hunic terrain model, as proposed by Stampfl et al., 2002. (Heavy dashed line represents southern boundary of Hunic Terrains.)
the PAL is part of a larger, composite suture system (Galicia-Southern Brittany-Southern Bohemia, Alpine-Maures-Corsica-Sardina suture) that forms the southern boundary of the Armorican microplate (Figure 9). This hypothesis has been used by several authors (e.g., Carmignani et al., 1994; Carosi & Palmeri, 2002) to explain the structural juxtaposition of the high-grade metamorphic complex (Armorican provenance) above the low to medium-grade nappes of the Internal Nappe Zone (Gondwanan provenance). In this model, the occurrence of relic eclogites and highly deformed metabasites within the PAL are interpreted as tectonically dismembered high-grade metamorphic pieces of oceanic crust overriding the nappes (Giacomini et al., 2006).

In contrast, Stampfli et al. (2002) and von Raumer et al. (2003) propose that the PAL represents part of a complex system of structurally bounded small terrains (Gondwanan provenance) that collided diachronously with the Gondwanan margin and now form the Hun terrains (collectively, “Hun superterrain”; Figure 10). Recent structural, geochemical, and geochronological studies (e.g., Helbing & Tiepolo, 2005; Franceschelli et al., 2005; Giacomini et al., 2006) show conspicuous similarities between rocks both north and south of the PAL. The most recent published papers concerning the Variscan system and the role of Sardinia therein show no general consensus on the origin of the PAL and its relationship to the Hun superterrain (Franke et al., 2000; Matte, 2001; Stampfli et al., 2002; von Raumer et al., 2003).

**Tectono-Metamorphic Zones**

Despite the uncertain nature of the PAL, the Variscan rocks of Sardinia have been divided into three main tectono-metamorphic zones: the “Axial Zone” or Internal Zone in the northeast, the Nappe Zone (subdivided into both the Internal Nappe Zone (INZ) and
the External Nappe Zone (ENZ)), and the External Zone or Foreland (Carmignani et al., 1994; Figure 8).

**Internal Zone**

The Internal Zone of Sardinia has been further subdivided into two metamorphic complexes that together represent the innermost part of the Variscan chain in Sardinia: the High Grade Metamorphic Complex (HGMC) and the Low-Medium Grade Metamorphic Complex (L-MGMC) (Franceschelli et al., 2005). The HGMC makes up the aerial majority of the Internal Zone and is comprised of migmatites and gneisses that are characterized by amphibolite metamorphic grade, reaching the sillimanite + K-feldspar zone (Franceschelli et al., 2005). Sporadically embedded within the HGMC are isolated bodies of mafic and ultramafic metamorphic rock, whose protoliths are probably of Precambrian age, and which are intruded by Ordovician orthogneisses (Carmignani et al., 1994). Some of these ultramafic bodies contain relic granulites that are interpreted to be portions of layered basic bodies that intruded in the lower crust (Carmignani et al., 1994). The L-MGMC occupies the area closest to the PAL, cropping out on the southern portion of the island of Asinara, northern Nurra, Anglona, and the northern Baronie regions, and is structurally juxtaposed with the Nappe Zone (Figure 8; Franceschelli et al., 2005).

**Nappe Zone**

South of the PAL, a sequence of Cambrian to Early Carboniferous rocks have been stacked into a series of SW-directed nappes, presumably during the collision of either the Armorican microplate or the Hun superterrain with the Gondwanan mainland (Carmignani et al., 1994; Giacomini et al., 2006). Kinematic indicators within the Nappe
Zone (e.g., fold vergence, tectonic lineations, and shear sense in mylonites) show a main SSW-directed tectonic transport (Carmignani et al., 1994). Based on stratigraphic, metamorphic, and structural features, the Nappe Zone is sub-divided into the External and Internal Nappe Zones, occupying the northern and southern portions of the Nappe Zone respectively (Carmignani et al., 1994). The most complete exposure of the entire Nappe Zone is located on the flanks of the Flumendosa Antiform, a regional-scale (10’s of km) E-W trending fold, located in the Gerrei region of Sardinia (Figure 8; Carmignani et al., 1994).

**Internal Nappe Zone**

The Internal Nappe Zone is exposed from the PAL southward to the Flumendosa Antiform where it overthrusts the northern edge of the External Nappe Zone (Figure 8). This zone comprises low to medium-grade sequences of metasandstones, quartzites, phyllites, marbles, and a limited amount of metavolcanic rocks (Franceschelli et al., 2005; Carmignani et al., 1995). Sequences of the Internal Nappe Zone are similar to those of the External Nappe Zone except for the low abundance of Ordovician metavolcanic packages and Devonian age marbles, and the greater abundance of characteristic Siluro-Devonian graphitic phyllites (Figure 11; Carmignani et al., 1994). A Barrovian metamorphic facies is pervasive in the Internal Nappe Zone, ranging from greenschist facies in the Barbagia, Goceano, and southern Nurra regions to intermediate P amphibolite facies proximal to the PAL (Carmignani et al., 1994; references therein).

Stretching lineations in the low-grade metamorphic units are oriented approximately parallel to the orogen (within 10°); whereas minor folds commonly trend...
Figure 11. Generalized stratigraphy of both the Internal and External Nappe zones of central Sardinia. From Funedda et al. (2010)
Figure 12. Relative tectonostratigraphy of the nappe zone, highlighting the relationships between the Internal nappes, external nappes, and the foreland zone. Simplified from Conti et al., 2001.
normal to the orogen. Axes of these orogen-normal folds are commonly cut obliquely by their own schistosity (Carmignani et al., 1992). This fold/schistosity pattern has been observed in the Appalachian belt (termed “Transected Folds”), where it has been called upon as evidence for regional strike-slip movement (Carmignani et al., 1992; references therein). Along with the transected folds, the stretching lineations have been drawn on to provide evidence for the interpretation that large scale strike-slip movement accompanied the tectonic shortening (Carmignani et al., 1992). This dextral transpressional component of Variscan deformation in Sardinia fits well with the known transcurrent tectonics of the Ibero-Armorican arc described by Matte (2002; Figure 8).

**External Nappe Zone**

The External Nappe Zone occupies the area south of the Flumendosa Antiform, southward to the Iglesiente region of SW Sardinia (Figure 8). It is composed of lower greenschist facies to non- or very weakly metamorphosed rocks. Within each nappe, the lowest part of the sequence is made up of Cambrian metasandstones, metasiltites, quartzites, and phyllites grouped into the “Arenarie di San Vito” and “Arenarie di Solanas” Formations (Figure 11; Carmignani et al., 1994). Above the Cambrian package, an unconformity, the “Sarrabese” Phase, marks the abrupt transition into a Middle Ordovician volcanic complex, which is composed of metavolcanic rocks ranging from basalts to intermediate-felsic dominated rhyolites and, in some cases, metavolcanoclastic rocks (Carmignani et al., 1994). The Middle Ordovician volcanic units are overlain by an Upper Ordovician meta-sedimentary package consisting of quartzites, metarkoses, metaconglomerates, fossiliferous shales, and metalimestones (Carmignani et al., 1994). Silurian rocks in the sequence consist of graptolitic black shales and smaller interbedded
packages of fosiliferous limestones that grade upward into the more competent Devonian
limestones (Carmignani et al., 1994). In the outermost areas (southwestern-most) of the
External Nappe Zone, syntectonic flysch deposits of Lower Carboniferous age are found
locally (Carmignani et al., 1994).

**Deformational history of the Nappe Zone**

The tectonic evolution of the Nappe Zone of Sardinia is best understood by
looking at the rocks in the External Nappe Zone (Figure 8). Outcroppings of Paleozoic
rocks in this zone show evidence for a complex series of Early Carboniferous
contractional deformation events (collectively $D_1$) and post-thickening, extensional

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<th>Table 1: Correlation of deformational events in the Variscan belt of Sardinia (from Conti et al., 2001)</th>
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<tr>
<td><strong>Tectonic Setting</strong></td>
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<td><strong>Crustal Thickening</strong></td>
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<td><strong>Tectonic Exhumation</strong></td>
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events ($D_2$; Table 1). Numerous workers have recognized evidence for SW-directed
tectonic transport as the most important transport direction in the whole of the Nappe
Zone. However, the overall structural evolution is more complex and consists of at least
three recognized deformational episodes and two 90° shifts in transport direction (Conti et al., 1999). Following D1 contraction, all tectonostratigraphic units were affected by either postcollisional extensional features, such as km-scale normal faulting and vertical thinning, or by granitic intrusion (Conti et al., 1999).

**Contractional Deformation and the Development of the External Nappes**

Initiating in the Early Carboniferous, the main (D1) deformation produced three primary decollements: 1) at the base of the Ordovician volcanic package, 2) the top of the Ordovician volcanic package, and 3) the bottom of the Cambrian metasedimentary units (Carmignani et al., 1992). D1 deformation also produced W-to-SW-facing tight-to-isoclinal folds with an S1 foliation developed in the Chlorite zone of the lower greenschist facies (Carmignani et al., 1994). This SSW-directed contraction resulted in a structural stack consisting of four main nappes (from bottom to top), the: 1) Castello Medusa-Riu Gruppa unit (here, simply “Riu Gruppa unit”), 2) Gerrei unit, 3) Meana Sardo unit, and 4) Barbagia unit (Figure 12; Conti et al., 1999). A minimum SSW-directed transport magnitude, obtained by digitally backing each nappe off of its underlying counterpart and repositioning them along a NE-SW shortening axis, is shown to be ~ 85 km (Figure 13). Later W-directed thrusting placed the Sarrabus unit over both the Gerrei and Meana Sardo units exposed only in SE Sardinia (Conti et al., 1999).

Observations by previous workers at both the outcrop and thin section scale have shown that the contractional history was complex and consisted of at least four distinct deformational phases (Figure 14, Conti et al., 1998; Conti et al., 1999). These phases consist of the following:
Figure 13. Minimum transport magnitudes obtained by digitally backing each nappe off of each other along an axis concordant with main nappe emplacement direction.
Figure 14. Time series showing development of the external nappes, ranging from early folding and thrusting to post-thickening normal faulting. From Conti et al., 1999
Gerrei phase (D1a)

Early SSW-directed folding and thrusting, creating folds and mylonites in the Riu Gruppa, Meana Sardo, and Barbagia units, also developing isoclinal folds in the Gerrei unit (Conti et al., 1999; Figure 14a).

Meana phase (D1b)

SSW-directed main Nappe emplacement, developing mylonites along the main thrusts and within the Meana Sardo and Barbagia units (Conti et al., 1999; Figure 14b).

Sarrabus phase (D1c)

W-directed emplacement of the Sarrabus unit over both the Meana Sardo and Gerrei units along the Villasalto thrust (Conti et al., 1999; Figure 14c).

Flumendosa phase (D1d)

A late SSW-directed shortening stage that developed upright E-W striking folds (i.e., Flumendosa antiform, Barbagia synform, and Gennargentu antiform) that refold earlier thrusts and folds (Conti et al., 1999; Figure 14d).

Post-Thickening Deformational Events

Structures that developed as a response to post-collisional (D2) thinning are located throughout Variscan rocks of Sardinia and were produced at different crustal levels (Carmignani et al., 1994). These extensional structures include km-scale high and low-angle normal faults, asymmetric eastward overturned and recumbent folds, and granitic intrusion (Conti et al., 1999). In the lower crustal levels, cropping out in northern Sardinia, this extensional deformation is expressed as a transecting foliation in basement pendants comprised of micaschists, paragneissuses, and migmatites (Carmignani et al.,
Further south and at a higher crustal level, this D2 phase is best expressed in the flanks of major collisional structures such as the Flumendosa Antiform.

The latest contractional deformation stage (Flumendosa Phase) oriented pre-existing structures and weak stratigraphic levels into geometries that served to localize D2 deformation (Conti et al., 1999). Low-angle normal faults preferentially developed on both the northern and southern limbs of E-W striking antiforms, unroofing their hinge zones (Figure 14e; Conti et al., 1999). Normal sense (D2) displacement along these boundaries is differentiated from D1 thrusts rotated into a S-dipping orientation by displacement of marker horizons, minor drag folding, shear sense of attenuated porphyroclasts, and shear bands in mylonites (Conti et al., 1999).

Figure 15: Drawing illustrating the deformation of structures developed during shortening phases (labeled S1) by structures accommodating extension (normal fault, S2 foliation)
From Conti et al., 1999, drawing by Ausonio Ronchi
At shallower crustal levels, some of the main nappe emplacing thrusts have been re-activated as normal faults. The most notable of these is the Villasalto thrust, located on the southern flank of the Flumendosa antiform. This fault is characterized by a thick cataclastic zone that internally displays normal sense kinematic indicators that in some locations cut D1 foliations in the footwall (Conti et al., 1999; Carmignani et al., 1994). This D2 deformation of earlier D1 structures is illustrated in Figure 15, showing refolding of S1 foliation and recumbent fold development produced by D2 normal faulting. Other E-W striking, higher angle normal faults (i.e., the Corr ‘e Cerbo fault) can be located along strike for as long as 15 km and are observed to truncate earlier low-angle normal faults (Conti et al., 1999). Interactions between D1 and D2 structures in the External Nappe Zone, specifically in the Lago Mulargia region of Sardinia, can provide some insight to the late evolution of the Variscan system in Sardinia and will be further examined in following chapter.

External Zone

The External Zone, cropping out in the Iglesiente-Sulcis region of SW Sardinia (Figure 8), is composed of a Cambrian to Devonian sedimentary sequence and has features of a classic foreland fold-and-thrust belt with steeply dipping thrusts and fold axial planes (Carmignani et al., 1992). One of the most striking structural features in the External Zone is the fold interference pattern produced by E-W trending folds (accentuated by both pre- and early Variscan deformation) and N-S striking, purely Variscan folds (Carosi et al., 1992). The resultant dome-and-basin structures expose the
Figure 16. Stepwise diagram showing the development of backthrusts in a Cambrian basin, due to inversion of tectonic regime.
oldest rocks in the center of large anticlines bounded by narrow synclines such as the “Arenarie Dome” north of Iglesias (Carosi et al., 1992; Carmignani et al., 1994). Another important structural feature of the External Zone is the presence of northeastward, hinterland-directed back-thrusts that affect the entire Cambro-Ordovician sequence. This Cambrian-Devonian succession is interpreted to have been deposited in a basin bounded on the west by a slope and deepening eastward along a ramp (Carosi et al., 1992).

The Cambrian normal faults that created this basin may have been reactivated at the onset of Variscan shortening as thrust faults, with top to the west displacement. As the basin closed at the peak of shortening, back-thrusts developed, possibly due to a buttressing effect of the listric normal fault geometry, resulting in the inversion of the basin (Carosi et al., 1992; Carmignani et al., 1994). Figure 16 illustrates the kinematics of this possible buttressing effect of listric fault geometry. Carmignani et al. (1992) suggests that a decollement likely exists at the base of the Cambrian sequence and that it may serve as a common linkage to all of the major thrusts in the External Zone.
THE VARISCAN OF THE LAGO MULARGIA REGION, SARDINIA

The Lago Mulargia region of south-central Sardinia (Figures 8 and 17) contains Variscan structures that have been attributed to both D1 contractional deformation and D2 extensional deformation. Due to the unique exposure and interactions of these structures in this area, the Lago Mulargia region serves as an excellent “field laboratory” to further investigate and further define the timing and deformational characteristics of these (D1a-D1d) events. Here, three of the four main tectonostratigraphic units, namely the Riu Gruppa unit, the Gerrei unit, and the Meana Sardo unit, crop out to varying degrees (Plate 1). Major D1 structures in the area include nappe-emplacing thrust faults, isoclinal folds trending 90-140°, and stretching lineations with similar trends. D2 structures in the Lago Mulargia region consist of eastward overturned folds trending N-S, plunging an average of 20° S, and generally N-S striking normal faults.

Tectonostratigraphic Units

Rock types contained within each of these Nappes are, for the most part, correlative (Figure 18). Physical differences (thickness, grain size, mineral content) within each specific unit as displayed in different thrust sheets are attributed to the non-homogeneous nature of terrigenous sedimentary and volcano-clastic rocks. Nappe descriptions here will include only rocks located within the mapped region. All units have undergone lower
Figure 17: Geology of South Western Sardinia, Lago Mulargia area shown by dashed box. From Carosi et al., 1992
Syn-tectonic deposits

Terrigenous succession and Carbonates of the Late Ordovician to Early Carboniferous

Mid Ordovician Volcano-clastic succession

Cambrian to Early Ordovician Terrigenous succession

Figure 18: Correlation or rock units in the Lago Mulargia area within each nappe, see Figure 12 for rock description. From Funedda et al., 2010.
Greenschist facies metamorphism, and for the purpose of clarity descriptions in Table 2 will concern their protoliths.

**Data**

Table 2: Rock unit descriptions subdivided according to tectonostratigraphic units, translated from Funedda et al., 2010

<table>
<thead>
<tr>
<th>NAPPE</th>
<th>UNIT SYMBOL</th>
<th>UNIT NAME</th>
<th>DESCRIPTION</th>
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<tbody>
<tr>
<td>Riu Gruppa</td>
<td>PMN</td>
<td>FORMAZIONE DI PALA MANNA</td>
<td>Alternating beds of siltstones, micaceous arenites, and quartz arenites. Contains some limited zones of chert conglomerates and breccias. PMN&lt;sub&gt;a&lt;/sub&gt;; massive, dark quartz arenites (Lower Carboniferous)</td>
</tr>
<tr>
<td>ASU</td>
<td></td>
<td>MARMi DI ASUNi</td>
<td>Dolomitic marbles, limited occurrence of calc-shists, rare zones of crinoids (Devonian-Lower Carboniferous)</td>
</tr>
<tr>
<td>SGA</td>
<td></td>
<td>SCISTi A GRAPTOLiTi</td>
<td>Calcareous pelites and siltstones with intercalated nodules of black chert (Silurian-Devonian)</td>
</tr>
<tr>
<td>SGA&lt;sub&gt;a&lt;/sub&gt;</td>
<td></td>
<td>SCISTi A GRAPTOLiTi (a)</td>
<td>Dark grey carbonates and lighter fossil bearing carbonates (Silurian-Devonian)</td>
</tr>
<tr>
<td>Gerrei</td>
<td>SGA</td>
<td>SCISTi A GRAPTOLiTi</td>
<td>Carbonic pelites and siltstones with intercalated nodules of black chert (Silurian-Devonian)</td>
</tr>
<tr>
<td>SGA&lt;sub&gt;a&lt;/sub&gt;</td>
<td></td>
<td>SCISTi A GRAPTOLiTi (a)</td>
<td>Dark grey carbonates and lighter fossil bearing carbonates (Silurian-Devonian)</td>
</tr>
<tr>
<td>ACN</td>
<td></td>
<td>ARGILLOSCISTi DI RIO CANONi</td>
<td>Grey-green pelitic rocks, calcareous siltstones with intercalated carbonates, locally silicified (Upper Ordovician)</td>
</tr>
<tr>
<td>MGM</td>
<td></td>
<td>METARCOSi DI GENNA MESA</td>
<td>Arkose and Greywacke, quartz arenite and quartz conglomerates, thickly bedded or massive (Upper Ordovician)</td>
</tr>
<tr>
<td>PRF</td>
<td></td>
<td>PORFIROIDi</td>
<td>Rhyolite and Dacite, zones of epiclastic material, isolated packages throughout the area contain large (2-10cm) phenocrysts of K-feldspar (Middle Ordovician)</td>
</tr>
<tr>
<td>Meana Sardo</td>
<td>ORR</td>
<td>FORMAZIONE DI ORROLEDEDU</td>
<td>Grey to green pelitic rocks and fossiliferous siltstones (Upper Ordovician)</td>
</tr>
<tr>
<td>MSV</td>
<td></td>
<td>FORMAZIONE DI MONTE SANTA VITTORIA</td>
<td>Volcanic Greywacke and Andicite, MSV&lt;sub&gt;a&lt;/sub&gt;; aphanitic Rhyolite, MSV&lt;sub&gt;a&lt;/sub&gt;; Volcanic Greywacke with rare packages of Andicite, epiclastic (Middle Ordovician)</td>
</tr>
<tr>
<td>SVI</td>
<td></td>
<td>ARENARlE DI SAN VITO</td>
<td>Alternating packages of quartz arenite, micatious arenite, and pelitic rocks (Middle Cambrian-Lower Ordovician)</td>
</tr>
</tbody>
</table>
Field work consisted of 1:10,000 scale field mapping focused on documentation of fault geometry, distribution of fold orientations and wavelengths, and collection of oriented samples near these faults. Initial regional mapping of the area was carried out by Funedda et al. (2010). Further reconnaissance mapping proved, for the most part, to corroborate with and, in some instances, further refine the work done by Funedda et al., (2010). Analytical work comprised of creating contour fault plane resolution diagrams (Figure 19), examination of microstructures displayed in thin sections from oriented samples, and stereographic analysis of data gathered from these thin sections.

Faults

Faults in the Lago Mulargia region can be divided into three general groups according to their orientation, dip direction, and offset magnitude (Table 3), Nappe emplacing thrust faults (Fault group 1, FG1), N-S trending, East dipping faults (Fault group 2, FG2), and generally E-W trending normal faults with small offset magnitudes (Fault group 3, FG3). Mapped field relationships (Plate 1) of these faults establish a relative temporal sequence starting with the FG1 followed by the FG2 and then the FG3. To illustrate these fault relationships, a simplified map showing only nappe boundaries and faults has been created (Figure 19).

The two faults in FG1 together provide the local boundaries of the Gerrei unit. The lower boundary of the Gerrei unit is marked by the western-most thrust fault (FG1-A), which delineates the boundary between the Gerrei unit and the Riu Gruppa unit. The upper boundary of the Gerrei unit, and its contact with the Meana Sardo unit, is marked by the eastern-most thrust fault (FG1-B). Both FG1-A and FG1-B dip to the east, contour
a) location used to determine dip, FG1-A, 25° E

b) location used to determine dip, FG2-A, 36° E

c) location used to determine dip, FG1-B, 33° E

Figure 19: Fault plane resolution diagram illustrating the orientation of all major faults in the Lago Mulargia area. Fault plane geometry resolved by creating a network of known point elevations along the trace of the faults and interpolating contours at 25 m increments.
Figure 20: Simplified geological map of the Lago Mulargia region, showing only tectonostratigraphic units, Post-Variscan cover, and relevant faults.
fault plane resolution shows the FG1-A to be dipping ~21° E and the FG1-B to be
dipping ~33° E (Figure 19).

Two N-S trending faults that comprise the FG2 are mostly continuous through the
mapped region and off-set both the FG1-A and the FG1-B (Plate 1, Figure 20). The

<table>
<thead>
<tr>
<th>Fault Group</th>
<th>Members</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fault Group 1 (FG1)</td>
<td>Two nappe emplacing thrust faults, one providing the lower boundary of the Gerrei unit (FG1-A) and the other forming the upper boundary of the Gerrei unit (FG1-B)</td>
</tr>
<tr>
<td>Fault Group 2 (FG2)</td>
<td>N-S striking faults not associated with nappe emplacement, one fault entirely contained within the Gerrei unit (FG2-A) and another that places upper units within the Meano Sardo unit in contact with the Gerrei unit (FG2-B)</td>
</tr>
<tr>
<td>Fault Group 3 (FG3)</td>
<td>Numerous small scale normal faults ubiquitous to the map area, these faults generally trend E-W and truncate faults in faults groups FG1 and FG2</td>
</tr>
</tbody>
</table>

eastern-most of these (FG2-A) is almost completely contained within the Gerrei, indicating that vertical separation of this fault cannot be greater than the total thickness of the Gerrei unit. To the west, a fault juxtaposes the Meana Sardo unit with the Gerrei unit (FG2-B), this provides a much larger bracket for off-set magnitudes. Discontinuous East-West trending faults of FG3 are common in to the area and display much smaller off-set magnitudes than those of the other two fault groups.
Field relationships of these faults can provide some insight into the late Variscan evolution of southern central Sardinia. Previous workers (Carmignani et al., 1994; Conti et al., 1999) have interpreted the eastern-most N-S striking fault (FG2-A) as a thrust. Further investigation of the mapped area shows that the interpretation of shear sense along this eastern-most N-S striking fault carries implications that lead to three possible models that could result in the mapped relationship.

Hypothesis one: this fault is an out-of-sequence thrust propagating below FG1-A and ramping stratigraphically upward into the overlying Gerrei and Meana Sardo units (Figure 21a). This model would suggest that kinematic indicators along this fault would be consistent with the SSW-directed transport direction of both the FG1-A fault and the FG2-A fault.

Hypothesis two: following nappe emplacement, a hinterland directed (NW dipping) normal fault truncated the nappe stack (Figure 21b). Implications of this model would be SSW-directed shear indicators along both faults in the FG1 and top-to-the-NW indicators along the FG2-A.

Hypothesis three: after initial emplacement, extension was facilitated by reactivation of both or either of the faults in the FG1 as hinterland-directed normal faults and accompanied by the development of new hinterland-directed normal faults (Figure 21c). Kinematic indicators in this model could include both top-to-the-SW and top-to-the-NW on both the FG1 faults and limit the FG2-A fault to top-to-the-NW indicators.

To test the viability of these hypotheses, work was focused on establishing a sense of offset along both the FG2-A fault and the FG1-A fault and further analyzing the field relationships that exist between these two faults. New insights into the nature of the FG2-
a) HYPOTHESIS ONE
After initial nappe emplacement, an out-of-sequence thrust propagating below the Gerrei unit truncates the nappe stack thrusting a portion of the Gerrei unit and the overlying Meana Sardo unit over a portion of the Gerrei unit and the Riu Gruppa unit.

b) HYPOTHESIS TWO
Following nappe emplacement, the nappe stack is truncated by a hinterland-directed normal fault.

c) HYPOTHESIS THREE
Once contractional deformation has seceded, extensional deformation is facilitated by reactivation of either both or one of the initial thrust faults, followed by the development of a hinterland-directed normal fault.

Figure 21: Hypotheses proposed to accommodate mapped geology in the Lago Mulargia area
A fault will help to test hypotheses pertaining to deformational events occurring late in the Variscan Orogen in southern central Sardinia.

Contour fault plane analysis of both the FG1-A fault and the FG2-A fault shows that both faults dip to the east with the former dipping ~25° E and the latter dipping ~36° E and also highlights a junction of the two faults (Figure 19). Here it can be noted that the fault segment to the north of the junction possess a dip consistent with that displayed by the FG2-A. From this observation, it is concluded that the FG1-A fault is truncated by the FG2-A fault.

Approximately 1 km north of the junction of these two faults, there is a poorly preserved package of PRF~130m long and 45m wide, bounded on the east by rocks in the Riu Gruppa nappe and bounded on the west by the SGAa unit of the Gerrei nappe (Figure 22a). This package is thought to be a “sliver” of the lower Gerrei nappe with its western boundary formed by the FG1-A fault. This relationship would suggest that the FG2-A fault exploited the existing plane of weakness formed by the FG1-A fault.

Folds

Folding in the mapped area is most evident in the Gerrei and Riu Gruppa units, with each unit displaying unique fold patterns. Fold patterns displayed within the Gerrei unit consist of ~1 km scale E-W trending folds accompanied by 2-15 cm scale asymmetric folds (Figures 22 and 23). These folds are truncated in the Gerrei unit by the FG2-A fault and although outcrop scale folds in an E-W orientation are visible to the east of this boundary, the larger (~1 km scale) are not discernable. In the Riu Gruppa unit, folding is comprised of ~1 km scale folds trending N-S. In the mapped area, a south
Figure 22: Map insets highlighting a) a small (~130m X 45m) tectonic “sliver” derived from the lower section of the Gerrei unit and b) the truncation of the FG1-A fault by the FG2-A fault.
Figure 23: Folds in the Lago Mulargia area, consisting of ~1Km scale E-W trending folds accompanied by ~2-3 Km scale N-S trending folds. Insets highlight E-W trending folds present in the lower portion of the Gerrei unit (a) and fold interference patterns (b).
dipping synform-antiform sequence affecting the SGA, ASU (laterally discontinuous here) and PMN units may be traceable across the FG1-A fault and expressed in the southwestern portion of the Gerrei unit as a zone of “Dome and Basin” fold interference patterns (Figure 23). The most notable mapped relationship of the N-S trending folds is the near parallel trend of these folds and the FG2-A fault. This fault-fold limb relationship has been noted in numerous locations within the external nappe zone, most notably the Villasalto thrust and the Corr ‘e Cerbo fault, although in both of these instances the faults/folds trend E-W.

**Thin Sections of Oriented Samples**

To aid in determining which proposed hypothesis is more viable, oriented samples were collected in proximity to both the FG1-A and the FG2-A faults (Plate 1). All samples were cut perpendicular to strike and parallel to lineation. Each sample represents a vertical plane perpendicular to foliation with the inclination determined by the stretching direction of attenuated clasts. Stereographic analysis of these samples was carried out by representing each sample at a line, where the trend is determined by the azimuth of the slide plane of the sample (being a vertical plane) and the plunge determined by the main stretching direction delineated by attenuated clasts within each sample.

Of all samples examined, each shows one stretching direction with the exception of sample 23, located in a package of PRF bounded on the west by the FG1-A fault and bounded on the east by the FG2-A fault (Plate 1). This particular sample contains euhedral pyrite crystals that have developed quartz strain fringes that appear to have been
Plate 2: Thin section of sample 23 (Figure 24 for location) showing quartz strain fringes on euhedral pyrite crystals. It is apparent that a first set of fringes existed (aligned with primary foliation) but has been re-crystallized and overprinted by a second set developed with a different strain axis.

Primary foliation: Trend 105°, Plunge 23° E
Plate 3: Quartz strain tail development of euhedral and semi euhedral pyrite grain boundaries. Strain tail overprinting relationship highlights a strain axis inclined c.30° from the primary foliation.
Primary foliation: Trend 105°, Plunge 23° E
Figure 24: Stereographic plots of oriented thin sections

24a, stereographic plot of oriented samples proximal to the FG1-A fault. Both representations of sample 23 (A and B) shown in orange. Best fit great circle determined by single orientation samples (shown in red), 026°, 32°E.

24b, stereographic plot of oriented samples proximal to the FG2-A fault. Both representations of sample 23 (A and B) shown in orange. No particular orientation if found to be common within pints proximal to the FG2-A fault (shown in yellow).
re-crystallized and overprinted by a second set of strain fringes that developed with a different stretching axis (Plates 2 & 3). Recrystallized quartz grains in this first set of strain fringes are equigranular and display interlobate grain boundaries. Intact strain fringes that overprint the re-crystallized quartz throughout the sample are parallel in all occurrences and tend to extend about 12% of the first re-crystallized quartz tails. For the purpose of stereographic analysis, this sample is assigned one trend (slide plane, 105°) and two plunge values; one of the primary stretching direction (23° E, 23a) and one inclined approximately 30° (53°, 23b) to the primary foliation.

When plotted, the samples collected along the FG1-A fault show a distinct alignment represented by a best fit great circle with an orientation of 026° strike, 32° E dip (Figure 25a). Fault plane analysis (Figure 19) of the FG1-A fault resolved a strike ranging from 050° to 030° with a consistent dip of ~25°-30° E. This is somewhat comparable to the best fit great circle when 1km scale gentile folds of the fault plane are taken into account. When compared with sample 23 (resolved as 23a and 23b as previously described), orientation of the initial stretching direction, 23a, is congruent with the best fit great circle determined by the samples collected along the FG1-A fault.

Stereographic analysis of the FG2-A fault is not as straightforward, samples here show two distinct populations (Figure 24b). Samples 20, 17, and 5 all trend approximately 090°, with corresponding plunge magnitudes ranging from flat lying to 30°. Other data collected along the FG2-A fault are not distributed in any systematic way. When compared with the data collected from FG2-A, sample 23a and 23b have similar trends to samples 20, 17, and 5 (Figure 24). Sample 23b shows a much steeper plunge more similar to the general dip magnitude shown for the FG2-A fault in Figure 19.
**Interpretation of Deformational History**

To illustrate the deformational history inferred for the Lago Mulargia region, a time series of block diagram cartoons will be examined, ranging from Early Carboniferous to Permian time. Although not every detail resolved on the geologic map (Plate 1) is accounted for, these cartoons serve to show a general deformational history and culminate in a geometry that closely approximates the mapped geology. For the purpose of clarity in this discussion and to allow for comparison to deformational histories resolved by previous workers, identified deformational events here will be presented as D1, D2, etc. continuing in chorological order and summarized in Table 4.

The earliest deformation event evidenced in the Lago Mulargia region (D1) is characterized small (500 m wavelength) E-W trending folds most visible in the SW portion of the map (Figure 23b). It is clear from mapped relationships that these D1 structures are truncated by later developed structural features. Following D1 deformation, D2 deformation consisted of main SSW nappe emplacement along newly developed large scale NNE dipping thrust faults (Figure 25) stacking thrust sheets of mildly deformed (D1) material from NNE to SSW. Analysis of the map at a larger scale shows no evidence of the Gerrei Unit west of the Lago Mulargia area and is thus interpreted to represent the lateral termination of the nappe, resulting in fault boundaries that trend N-S despite an emplacement direction of NNE to SSW. This “lateral ramp” geometry is represented by a general dip to the east of nappe emplacing faults (Figure 25). At this stage, small E-W trending folds (D1) that developed prior to nappe emplacement are
visible throughout the nappes but not pervasive through nappe boundaries. This phase also shows the initial configurations of both faults in the FG1.

Following the cessation of SSW-directed D2 deformation, a ~90° shift in shortening direction occurred. This D3 phase of deformation resulted in the west-directed thickening of the nappe stack and is characterized by km scale N-S tending folds that affect both D1 (E-W 500m scale folds) and D2 (Large scale NNE dipping thrust faults) structures (Figure 26). During W-directed (D3) deformation, some shortening is interpreted to have been concentrated at Nappe boundaries, resulting in the formation of drag folds proximal to these preexisting fault planes and causing a general thickening of the nappe stack (Figure 26, inset). Here it is also worth mentioning that there may be a larger scale set of weakly defined N-S trending folds, which provides a mechanism for both faults in the FG1 to rotated from NNE dipping to be east dipping.

**Table 4:** Outline of recognized deformational phases in Lago Mulargia region

<table>
<thead>
<tr>
<th>Phase</th>
<th>Deformation direction</th>
<th>Characteristic structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>NNE-SSW</td>
<td>~500 meter scale E-W trending folds</td>
</tr>
<tr>
<td>D2</td>
<td>NNE-SSW</td>
<td>large scale nappe emplacing thrust faults, initially NNE dipping, development of primary strain fringes</td>
</tr>
<tr>
<td>D3</td>
<td>E-W</td>
<td>~1km scale N-S striking folds, ~200 m scale N-S overturned drag folds concentrated along D2 structures</td>
</tr>
<tr>
<td>D4</td>
<td>E-W</td>
<td>N-S trending, E dipping normal fault (FG2-A) total offset less than thickness of Gerrei Nappe, overprinting of strain tails</td>
</tr>
<tr>
<td>D5</td>
<td>NNE-SSW</td>
<td>Gentle 500m scale E-W trending folds affecting D1, D2, and D4 structures</td>
</tr>
<tr>
<td>D6</td>
<td>E-W</td>
<td>N-S striking normal fault, juxtaposing undeformed rocks of Permian age with those in the Meana Sardo and Gerrei Nappes</td>
</tr>
</tbody>
</table>
Figure 25: Block diagram illustrating fault configuration following D2 deformation. Nappe emplacement faults truncate minor E-W trending folds developed in the previous D1 phase.
Figure 26: W-directed D3 deformation characterized by ~1 km scale N-S trending folds. Inset highlighting drag folding along D2 developed thrust faults.
Figure 27: Showing location (a) and sense of displacement (b) and post erosional configuration (c) along D4 N-S trending, E dipping normal fault (FG2-A) total offset less than thickness of Gerrei Nappe.
Figure 28: Diagram showing orientation and distribution of units post D5 deformation (a), E-W directed extension juxtaposing Permian age rocks (b) and final configuration of nappes (c, color convention from Figure 20).
At this point, the entire nappe stack underwent a “relaxation” period characterized by D4 extensional deformation. In the Lago Mulargia region, D4 deformation is evidenced by a E dipping fault, FG2-A as previously described. Along the majority of its trace, the FG2-A fault juxtaposes rock from the upper and lower sections of the Gerrei Nappe, thus offset magnitude must be less than the total thickness of the Gerrei Nappe in this area. Stereographic analysis of N-S striking faults shows a consistent eastward attenuation of clasts. Here, the majority of deformation attributed to the D4 phase is interpreted to be exploiting previously existing planes of weakness (initially developed as D2 structures, later deformed by D3 deformation) that strike approximately N-S (Figure 27).

When analyzed in thin section, rocks proximal to the FG1-1 fault show that the last deformation event recorded by strain tail development shows an average trend of 026° dipping 32° to the East (Figure 24). From this data, we can assume that at least some of the extensional deformation was localized along the initial nappe emplacing thrust faults, resulting in normal sense reactivation. The remaining extensional deformation was accomplished by the creation of new E-dipping normal faults that partially utilize the pre-existing fault planes. The development of FG2-1 can be seen in Figure 27, exploiting the planes of weakness formed by both of the FG1 faults. Of particular importance is that the propagation of the FG2-1 fault did not follow these planes of weakness precisely, rather it followed them more generally, resulting in packages of rock belonging to different nappes left in their post-emplacement relationship. This is the process called upon to explain the small “sliver” of PRF derived from the Gerrei Unit that remains in its initial position with respect to the Riu Gruppa Nappe (Figure 22a).
Thin section analysis of sample 23 in particular, taken from within this “sliver,” shows two distinct deformation orientations with an easily determinable sequence (Plates 2 and 3). The first of these extensional events (shown by quartz fringes that have undergone dynamic recrystallization) occurs along a plane roughly parallel with the FG1-A fault, and the second event (shown by quartz fringes that have not been recrystalized, thus younger) occurs along a plane with a similar orientation to the FG2-A fault.

Following D4 E-directed normal faulting, we again interpret a ~90° shift in shortening direction similar to that of both D1 and D2 events. Visible in the fault plane resolution diagram (Figure 19) faults both of the FG1 and FG2 groups show gentle ENE-WSW ~500m scale folding. Hinge lines of individual folds can be noted affecting faults in both fold groups (FG1-A and FG2-A in particular) thus having to occur post D4 E-directed extensional deformation. This event is interpreted to have been characterized by gentle folding rather than reactivation of FG1 faults due to D3 and D4 deformation, essentially "locking up" and rotating previous structures developed with a NNE-SSW shortening direction.

The final deformation event to affect the Lago Mulargia area again represents a ~90° shift in deformation axis. Here it is characterized by a N-S striking normal fault, FG2-B, expressed mainly in the lower portion of the Meana Sardo Nappe although along some segments it forms the contact between rocks of the Gerrei Nappe and relatively undeformed rocks of Permian age (Figures 20 and 28). Although the orientation, timing, and other characteristics of the FG2-B fault are significant to the overall geology of the Lago Mulargia area; in this case, these areas of investigation fall outside the scope of this specific project and will not be discussed in further detail.
Figure 29: Comparative diagram of mapped geology and culmination of time step diagrams.
As previously mentioned, time step block diagrams approximately documenting the progression of these events culminates in a diagram that shows a striking resemblance to the mapped geology. To highlight and further facilitate these comparisons, a diagram showing Nappes and post-Variscan rocks as color blocks is provided and can be compared to a map that does the same (Figure 29).

**Comparison to Regional Deformation History**

Previous workers investigating the deformational history of the Variscan Orogen in Sardinia have established a regional deformational history that shows a significant resemblance to the sequence resolved by the rocks in the Lago Mulargia area (Conti et al., 1999). Upon further investigation, it is clear that although these two deformational histories are similar, there is some significant incongruence. Here we will discuss these similarities and differences, thus placing the deformation of the Lago Mulargia region in context of a regional sequence of events. To facilitate this discussion, a table outlining the regional history will again be presented (Table 1) and comparative discussions will also refer to locally established deformation as outlined in Table 4.

The primary regional event (D1) is recognized locally as small (~500m) scale folds primarily visible in the Gerrei unit (Table 4). This is congruent with the folding and regional schistocity resulting from early S-directed folding and thrusting of the Gerrei Phase (Table 1, Conti et al., 1999, 2001). Following what we now recognize as Gerrei Phase deformation, locally the development of D2, main nappe emplacing S-directed thrust faults stacks mildly deformed rocks into a series of three nappes (Table 4). Regionally, the Meana Phase of deformation consisted of SSW-directed shortening and nappe stacking and is thus congruent with locally recognized D2 deformation (Table 1).
Table 1: Correlation of deformational events in the Variscan belt of Sardinia (from Conti et al., 2001)

<table>
<thead>
<tr>
<th>Tectonic Setting</th>
<th>Characteristic Features</th>
<th>Deformation Phase</th>
<th>Phase Names</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crustal Thickening</td>
<td>Early mylonitic deformation in the Barbagia, Meana Sardo and Riu Gruppa units, folding and main regional schistocity in the Gerrei Unit</td>
<td>Early S-directed Thrusting and folding</td>
<td>Gerrei Phase</td>
</tr>
<tr>
<td></td>
<td>Final Emplacement, late mylonitic deformation, main foliation development in the Meana Sardo and Barbagia units</td>
<td>Main S-directed Nappe emplacement</td>
<td>Meana Phase</td>
</tr>
<tr>
<td></td>
<td>Emplacement and main folding in the Sarrabus and Arburese units</td>
<td>W-directed nappe emplacement</td>
<td>Sarrabus Phase</td>
</tr>
<tr>
<td></td>
<td>Large-scale upright antiforms and synforms, crenulation cleavage</td>
<td>Late stage N-S shortening</td>
<td>Flumendosa Phase</td>
</tr>
<tr>
<td>Tectonic Exhumation</td>
<td>Normal faulting, NW-SE folds, crenulation cleavage</td>
<td>D2</td>
<td>Rio Gruppa Phase</td>
</tr>
<tr>
<td>?</td>
<td>NE-SW folds, crenulation cleavage</td>
<td>D3</td>
<td></td>
</tr>
</tbody>
</table>

As resolved in the Lago Mulargia region, D2 (now recognized as Meana Phase) deformation was followed by D3 deformation. This phase is characterized by a 90° shift in shortening direction resulting in the development of ~1 km scale N-S striking upright folds and ~200m overturned folds proximal to D2 (Meana Phase) emplacing thrust faults (Table 4). These small scale overturned faults are interpreted to be a result of W-directed re-activation of previously existing (D2) planes of weakness. Conti et al. (1999) also recognizes a 90° shift in shortening axis following the Meana Phase of deformation (Table 1). Regionally, the Sarrabus phase is characterized by the W-directed emplacement of the Sarrabus and Arburese Nappes. Although neither of these nappes are found within the Lago Mulargia area, we interpret that D3 (Sarrabus Phase) deformation most likely represents the same W-directed shortening and the Lago Mulargia region is simply further afield than the leading edge of the Sarrabus nappe.
At this point Conti et al. (1999) recognizes another 90° shift in shortening direction, returning to a similar axis called upon for both Gerrei and Meana phase deformation. The Flumendosa phase is marked by the regional development of 10-15 Km scale upright folds, namely the Barbagia Synform, Flumendosa Antiform (Figure 17), and the Gennargentu Antiform that refolded previously developed folds and thrusts. Deformation during the Flumendosa Phase was not able to utilize previously developed planes of weakness in the same manner observed in the previous phases. I believe that this is due to the character of deformation attributed to the Sarrubus Phase, essentially “locking up” the FG1 faults by folding them into orientations no longer conducive to the shortening direction that developed them. Although these structural features are two of the most apparent features in the Nappe Zone at regional scale, the Lago Mulargia area is located on the North-dipping flank of the Flumendosa Antiform (Figure 17), thus the only deformation visible in the study area is a general dip to the N-NE, possibly accompanied by minor sympathetic folds similar in geometry and orientation to those developed in during the Gerrei and Meana Phases. Although no map scale evidence for the Flumendosa phase is visible in the Lago Mulargia area, this phase still affected rocks in the area and thus must be included in a deformatonal history of the area.

Following D3 (Sarrabus Phase) deformation a N-S trending, E-dipping normal fault (FG2-A, Figure 28) truncates the nappe stack and is primarily contained within the Gerrei Nappe. This W-directed extensional phase is locally defined as D4 (Table 4). No direct field evidence was found that relates Flumendosa phase deformation with locally defined D4 deformation: in this case, we default to the sequence established by Conti et al. (1999). The regionally defined Rio Gruppa phase follows the Flumendosa phase
(Table 1) and is characterized by NW-SW trending normal faults. Examination of structural features indicate that the Rio Gruppa phase is most likely the same event recognized in Table 1 as D4.

At this point, another 90° shift in deformation direction is indicated by the structures located in the Lago Mulargia region. The FG2-A fault (developed during the W-directed Rio Gruppa extensional phase) shows distinct E-W striking folds that are pervasive through other previously developed structures (Figure 19). This NNE-SSW directed phase of deformation that post-dates Rio Gruppa phase deformation may be isolated to the Lago Mulargia region and is not recognized in the regional deformation history. I interpret that this D5 phase indicates that the Variscan evolution at least in the Lago Mulargia region was more complex than previously interpreted and consisted of multiple ~90° shifts in deformation axis.

The D5 deformation event that post-dates the Riu Gruppa Phase marks the final phase of Variscan Orogenesis. Thinning and vertical attenuation of the nappes throughout the island continued into the Permian and may be linked to the development of Permian age basins. In the Lago Mulargia area, this Permian thinning is marked by the development of a normal fault of similar orientation and geometry to FG2-1, identified as D6 in Table 4. The basin developed by this fault was subsequently filled with Permian age sediment that now makes up the majority of the rock visible in the eastern portion Lago Mulargia area (Figure 29). Due to the nature of some deformation phases, not all phases that affected rocks in the Lago Mulargia region are visible at map scale and were not included in Table 4. To illustrate the relationships of these defined phases, a composite table was constructed, including regional events not observed and locally
resolved events not included in the regional history (Table 5). When visible in the mapped area, characteristic features in Table 5 include only features present in the Lago Mulargia area. For the purpose of continuity deformation phases identified will default to those identified by Conti et al. (1999) when applicable.

**Table 5: Correlation of regional scale and locally resolved deformation phases**

<table>
<thead>
<tr>
<th>Characteristic Features</th>
<th>Deformation Phase (Conti et al., 1999)</th>
<th>This Paper</th>
<th>Phase Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>~500 meter scale E-W trending folds</td>
<td>Early S-directed Thrusting and folding</td>
<td>D1</td>
<td>Gerrei Phase</td>
</tr>
<tr>
<td>large scale nappe emplacing thrust faults, initially NNE dipping, development of primary strain fringes</td>
<td>Main S-directed Nappe emplacement</td>
<td>D2</td>
<td>Meana Phase</td>
</tr>
<tr>
<td>~1km scale N-S striking folds</td>
<td>W-directed nappe emplacement</td>
<td>D3</td>
<td>Sarrabus Phase</td>
</tr>
<tr>
<td>~200 m scale N-S overturned drag folds concentrated along D2 structures</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Large-scale upright antiforms and synforms, crenulation cleavage¹</td>
<td>Late stage N-S shortening</td>
<td>NA</td>
<td>Flumendosa Phase</td>
</tr>
<tr>
<td>N-S trending, E dipping normal fault (FG2-A) total offset less than thickness of Gerrei Nappe, overprinting of strain tails</td>
<td></td>
<td>D2</td>
<td>Rio Gruppa Phase</td>
</tr>
<tr>
<td>Gentle 500m scale E-W trending folds affecting D1, D2, and D4 structures²</td>
<td></td>
<td>D5</td>
<td></td>
</tr>
</tbody>
</table>

**Relationship to the Variscan Orogen as a Whole**

Variscan geologic features observed in the Lago Mulargia area provide further insight into the later stages of orogenic development. The area underwent a period of NNE-SSW directed thickening (Gerrei Phase, Meana Phase), followed by a period of E-W directed shortening (Sarrabus Phase), followed in turn by a return to NNE-SSW directed shortening (Flumendosa Phase). This shortening was then followed at least locally by E-W directed extension (Rio Gruppa phase) and finally by another ~90° shift in direction and return to shortening (D5). Based on this study, the area underwent a higher degree of deformation as a result of E-W directed shortening than previously
Figure 30: Primary Gondwona/Laurentia collision direction shown by red arrow, marginal shortening axis show by yellow arrows.
thought. As a result of this E-W deformation, extensional structures in the area (i.e., normal faults) developed utilizing previously developed planes of weakness that trend in a more N-S direction than those found throughout the island, which typically trend E-W.

When the Variscan chain is observed at continental scale, it displays a pronounced curvilinear geometry, thought to have been a result of the African promontory on the leading edge of the Gondwanan plate. This sinuous geometry has resulted in strike-slip fault development of opposing sheer sense on either side of the promontory. Dextral sense deformation markers are observed in the Variscan age rocks of Sardinia most notably in rock packages proximal to the PAL as observed by Carosi & Palmeri, (2002) and others.

We believe that this dextral transpressive regime can be called upon to accommodate the particular series of deformational events suggested by the geology in the Lago Mulargia area. This dextral transpressive regime provides a mechanism for both NNE-SSW and E-W directed thickening as observed in the region and also predicts that extensional features would trend in both E-W and N-S (Figure 31). Although regional E-W extensional faults, most notably the Corr ‘e Cerbo fault on the southern flank of the Flumendosa Antiform, accommodate the majority of the offset due to crustal thinning, some aspect of tectonic thinning was accommodated by N-S striking features such as those observed in the Lago Mulargia region. This pattern of extensional tectonic features suggests that during the latest stages of orogenic development, the Variscan rocks in Sardinia, and the Southern European Variscides by association, underwent a post orogenic thinning stage that was directed in a more heterogeneous pattern than resolved by other workers.
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