LASI III Conference
Physical Geology of Subvolcanic Systems: Laccolith, Sills and Dykes

Elba Island (Italy)
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Field Guidebook

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# Table of Contents

1. INTRODUCTION – LASI III  
   - 9

2. HISTORICAL PERSPECTIVE  
   - 10

3. GEOLOGICAL SETTING  
   - 12
   - AFRICA-EUROPE CONVERGENCE  
     - 12
   - NORTHERN TYRRHENIAN SEA REGION  
     - 14
   - ELBA ISLAND  
     - 15
       - Western Elba and the Eastern Border Fault  
         - 16
       - Central Elba and the Central Elba fault  
         - 16
       - Eastern Elba and the Zuccale Fault  
         - 17

4. RISE AND FALL OF THE ELBA INTRUSIVE COMPLEXES  
   - 19
   - WESTERN ELBA INTRUSIVE COMPLEX  
     - 19
       - Build-up of the Multilayer Laccolith Complex  
         - 19
       - Emplacement of the Monte Capanne Pluton  
         - 19
       - Intrusion of the Orano dykes  
         - 20
       - Tectonic Splitting of the Intrusive Complex  
         - 20
   - EASTERN ELBA INTRUSIVE COMPLEX  
     - 22
       - Intrusion of the Pluton  
         - 23
       - Emplacement of Felsic dykes  
         - 23
       - Ore Minerals  
         - 23
       - Tectonic Splitting of the Complex  
         - 23
   - SUMMARY  
     - 24

5. GENESIS OF ELBA MAGMAS  
   - 25
   - GEOCHEMISTRY  
     - 25
   - PETROGENESIS  
     - 30
       - Group 1 - Crustal messengers  
         - 30
       - Group 2 – Hybrids  
         - 31
       - Group 3 - Mantle messengers  
         - 32
       - Magma sources and processes  
         - 34
   - SUMMARY  
     - 38

6. THE WESTERN-CENTRAL ELBA LACCOLITH COMPLEX  
   - 40
   - INTRODUCTION  
     - 40
   - THE INTRUSIVE UNITS  
     - 40
LASI III
Field Trip Guidebook

Capo Bianco aplite 40
Nasuto Microgranite 44
Portoferraio porphyry 44
San Martino porphyry 46

MAGMA EMLACEMENT 47
Geometry of the laccolith complex 47
Construction of the multilayer Laccolith Complex 49
Filling of the laccoliths 51

SUMMARY 55

FIELD STOPS 56
Capo Bianco 56
Acquaviva-Enfola (Tuesday 16, AM) 58
Boat trip – Marina di Campo Bay (Wednesday 17, AM) 62
Cala Bardella 64
Chiessi (Wednesday 17, PM) 65

7. THE WESTERN ELBA SHEETED PLUTON 67
(MONTE CAPANNE PLUTON)

INTRODUCTION 67
THE INTRUSIVE UNITS 68
Definition of the intrusive facies 69
Map of the intrusive facies 69
Textures of the intrusive facies 73
Geochemistry of the intrusive facies 75
Biotite chemistry of the intrusive facies 76

MAGMA EMLACEMENT 77
Pre-emplacement K-feldspar–biotite growth 78
Deep origin of magma batches 80
Pluton’s internal structure 81
Emplacement sequence and mechanisms 83
Rapid pluton assembly 85
Geochronological implications 85

SUMMARY 86
FIELD STOPS 88
Cava Bontempelli 88
Pomonte 90
Sant’Andrea (Wednesday 17, PM) 90
Punta del Cotoncello (Wednesday 17, PM) 92
8. **THE LATE-PLUTONIC MAFIC DYKE SWARM (ORANO PORPHYRY DYKES)**

**INTRODUCTION**

**THE INTRUSIVE UNIT**

*Orano porphyry*

*Structural features of Orano dykes*

**MAGMA EMLACEMENT**

*Extension*

*The Western Elba Transfer Zone*

*The northern Tyrrhenian transfer systems*

*Magmatism focused by transfer systems*

**SUMMARY**

**FIELD STOPS**

*Chiessi (Wednesday 17, PM)*

*Colle d’Orano*

*Capo Sant’Andrea - Punta Cotoncello (Wednesday 17, PM)*

*Punta della Crocetta*

9. **THE EASTERN ELBA PLUTON-SILL-DYKE COMPLEX (PORTO AZZURRO-CALAMITA)**

**INTRODUCTION**

**THE INTRUSIVE UNITS**

*Porto Azzurro pluton*

*Monte Castello dyke*

*Calamita Sill-Dyke Complex*

*Magmatic-hydrothermal system*

**SUMMARY**

**FIELD STOPS**

*Fosso di Fosco (Thursday 18, AM)*

*Cala Stagnone (Thursday 18, AM)*

10. **REFERENCES**

**ACKNOWLEDGEMENTS**
1. Introduction – LASI III

LASI I was an international workshop on the physical geology of subvolcanic systems, promoted by Chris Breitkreuz and Nick Petford. LASI I was held at TU Bergakademie, Freiberg, Germany in 2002. The two-day workshop and one-day field trip led by Chris Breitkreutz and Alex Mock attracted 40 participants from 10 countries who presented papers covering a range of topics relevant to the geology and emplacement of high-level intrusions. Fourteen of the papers were subsequently published in the Geological Society, London, Special Publication 234.

LASI II helped discover how the study of high-level magmatic systems developed over the four years since the previous meeting. Ken Thomson(†) and Nick Petford brought together 53 researchers from 16 countries working in physical geology, igneous petrology, volcanology, structural geology, crustal mechanics and geophysics to discuss all aspects relating to the shallow emplacement of magma. The workshop was structured into two days with forty-three oral presentations followed by a one-day field trip led by Donny Hutton and Ken Thomson to classic Palaeocene sill complexes of the Isle of Skye. Thirteen papers presented at the workshop have been just published in the Geological Society, London, Special Publication 302.

LASI III is assembling 44 researchers from 15 countries representing These researchers will continue and update the discussion of emplacement mechanisms of shallow tabular intrusions, focussing on the discussion on geophysical imaging of tabular intrusions and their relationships with regional geology, as well as on the relationships between build up of plutons by sheeted magma batches and the working of volcanic systems. This guidebook describe the background geology, petrology and geochemistry of the late Miocene intrusive complexes of Elba Island, among the youngest plutonic-subvolcanic complexes in Europe. The Elba intrusive complexes are grouped in (i) the Western-Central Elba Laccolith Complex, (ii) The Monte Capanne Sheeted Pluton, (iii) the late-plutonic Orano Dyke Swarm, and (iv) the Eastern Elba Pluton-Sill-Dyke complex. Outcrop details are provided for the tabular intrusions to be visited during the LASI III field trip.
2. Historical Perspective

The first inhabitants of Elba Island were Musterian/Neanderthal men, about 50,000 years ago. At that time, during the Würm glaciation, the sea level was some 100 m lower than present, and it was possible to walk from mainland Tuscany to the island of Elba, Gorgona, Capraia, Montecristo and Pianosa. About 18,000 years ago, individuals of Homo Sapiens were living in Elba. Some 12,000 years ago the isthmus joining the Elba to Tuscany was submerged, but no people remained in Elba. During the Neolithic, some 5,000 year ago, people came living in Elba again, and about three thousands years ago started exploitation of copper ores from ophiolitic sequences of western, central and eastern Elba. Thereafter, Elba, with its potential of iron ores, joined the history of the main Mediterranean civilizations: the island is reported in the myth of Jason and the Argonauts, and during the VIII century B.C. the Etruscans with their king, Tyrrenhus, are reported to settle in Tuscany. Around 650 B.C., long-lasting wars in the Middle East discontinued the iron supply from the Anatolian region to the Mediterranean area. Thus the Etruscan mines of Elba Island and southern Tuscany became the major producers of iron and other metals such as copper, lead, zinc, silver. Aristotle refers to Elba when he speaks about a little island in the Etruscan region where metals, such as iron and copper, were mined, and he calls this island “Aethalia” (i.e., sooty) owing to the smoke from smelting furnaces.

At the beginning of the fourth century B.C., the Greeks from Siracusa, Sicily conquered Elba, and about a century later, the Romans took control of the island along with its valuable iron ores. Some major problems plagued mining in the Roman periods with respect to earlier Etruscan times. First, Roman exploitation-smelting techniques were less efficient and additionally massive charcoal production to supply furnace during the Roman Republican period led to almost complete deforestation of Elba Island, opposite to the Etruscans, which selected trees to cut according to their age. Then, during the II century B.C., the Romans started to take metal from Sardinia, Spain, Germany and France. During the I century B.C., the Romans finally swept out from the Mediterranean Sea the threat of the pirates and they proudly called the Mediterranean Sea as Mare Nostrum (i.e. Our Sea). At this time, Romans built beautiful villas on the island of Tuscan Archipelago, and the first Emperor, Augustus, established that no more trees could be cut
on Elba to supply furnaces. The Romans started to quarry the granite of western Elba, and columns from there can be admired today at Pantheon and Saint Paul cathedral in Rome. At this time Plinius and Virgilius referred to the island as “Ilva”, a name deriving from a people coming from Ligury and living in Elba.

The documented history of Elba begins with the Christianism and the Middle Ages, and the name “Elba” was first used by the Pope Gregorius Magnus. In the eleventh century Elba becomes a possession of Pisa, which started to exploit again the iron ores and defended the island against the Saracen pirates, with mountain- and sea-towers, fortresses (Volterraio and Marciana) and churches. From the end of XIII century to the French Revolution, Elba Island was dominated by ever changing powers, i.e. Genoa, Spain, the Medici of Florence, the Lorraines of Tuscany, the Bourbons of Naples. During the years of Napoleon’s wars, Elba Island suffered complex events, and finally, after the defeat of Lipsia, Napoleon arrived at Elba on May 1814. He built streets, gave administrative rules, supported the public health, imposed a new development of the mining industry and projected ironworks. The 26 of February 1815, Napoleon left the island with 1100 followers taking advantage of the absence of the English Commissioner. He would never come back, but Elba will never forget him. He left here his two residences, a town house, Villa dei Mulini, and a country house, Villa San Martino.

After Napoleon’s departure and his defeat, Elba was annexed to the Grand Duchy of Tuscany in 1815 and it became part of the newly born Italy by plebiscite in 1860. The Italian Elba became an important iron centre and it went through a flourishing trade development that also brought a stream of immigrants from the mainland. This industrial development was also favoured by the loss of strategic weight of the island due to the accomplishment of the Italian unification. Yet, a new threat was already hanging over Elba, and the outbreak of the World War II caused the destruction of Portoferaio. In the post-war period the island started up its tourist trade, leaving forever the industrial activities.
3. Geological setting

Africa-Europe convergence

The contractional orogens and extensional basins of the Mediterranean region (Fig. 1) have formed since Late Mesozoic time in a tectonic environment dominated by the northward relative motion of Africa with respect to Eurasia (Dewey et al., 1989). Overall, the thrust belts were largely formed by shortening the old passive continental margins created during the early Mesozoic break-up of Pangea.

![Tectonic map of the Mediterranean region, showing Mesozoic–Cenozoic contractional orogens and Neogene extensional basins. Africa–Eurasia plate boundary is shown in red (Platt, 2007).](image)

For the Apennine orogenic system, Neotethyan oceanic lithosphere is being subducted along seismically active and tomographically imaged zones in the Calabrian–Tyrrenhian systems, while in the northern Tyrrenhian system the evidence for subduction of oceanic lithosphere is fragmentary and the plate boundaries are complicated by the existence of the Adria sub-plate. During the Late Cenozoic, a distinctive pattern has evolved (Fig. 3.2) of arcuate thrust belts or accretionary wedges surrounding extensional basins floored by new oceanic crust (such as in...
the southern Tyrrhenian Sea) or thinned continental crust (such as in the northern Tyrrhenian sea) (Jolivet and Faccenna, 2000).

![Fig. 3.2 – Oligocene to Miocene evolution of the Tyrrhenian region. In (c) Elba Island is located at the time-space transition between contractional and extensional domains (Faccenna et al., 2002).]
Northern Tyrrhenian Sea region

Elba Island is located at the northern end of the Tyrrhenian Sea, a region affected by extensional processes behind the eastward progressing compressive front of the Apennine mobile belt (Fig. 1). The backbone structure of the Apennines was constructed when the Sardinia–Corsica block collided with the Adria plate (Malinverno and Ryan, 1986). This orogenic system evolved diachronously as the extensional regime migrated from west to east, trailing the retreat of the compressive regime (Brunet et al., 2000) and giving way to the opening of the extensional ensialic back-arc Tyrrhenian basin.

Fig. 3.3 - Location map for the Tuscan Magmatic Province. Also reported are the younger potassic-ultrapotassic volcanic rocks of the Roman Magmatic Province.

In this framework, magmas were generated in the mantle and interacted with crust-derived felsic magmas to generate the variety of Tuscan Magmatic Province (Fig. 3.3) intrusive and extrusive products exposed over about 30,000 km² of southern Tuscany and the northern Tyrrhenian Sea (Ferrara et al., 1989; Giraud et al., 1986; Innocenti et al.,...
This igneous activity migrated from west (14 Ma) to east (0.2 Ma) in an extensional ensialic back-arc setting as the west-dipping Adriatic plate delaminated and rolled back to the east (Serri et al., 1993).

**Elba Island**

The structure of Elba Island is made up of five tectonic complexes (Fig. 3.4) which were stacked onto each other during the eastward Apenninic compressional event before 20 Ma (Deino et al., 1992).

The three lowest complexes (I-III) have continental features, consisting of metamorphic basement and shallow-water clastic and carbonate rocks, while the upper two (IV-V) are oceanic in character (Keller and Pialli, 1990; Pertusati et al., 1993; Trevisan, 1950). In more detail, Complex IV consists of Jurassic oceanic lithosphere of the western Tethys Ocean (peridotite, gabbro, pillow basalt and ophiolite sedimentary breccia) and its late Jurassic-middle Cretaceous sedimentary cover (chert, limestone, and argillite interbedded with siliceous limestone). These rocks were deformed and metamorphosed during the Apenninic compression to form east-verging folds. Complex
V consists of argillite, calcarenite and sandy marl of Palaeocene to middle Eocene age, overthrust by an upper Cretaceous flysch sequence (Keller and Pialli, 1990).

In late Miocene time, the extensional processes affected the area of Elba Island (Bouillin et al., 1993; Jolivet et al., 1994), and several laccoliths, a major pluton, a mafic dyke swarm and a felsic pluton-sill-dyke complex were emplaced within stacked tectonic complexes. These intrusive bodies are exposed across the whole Elba Island, which is geologically and geomorphically subdivided by large-scale faults into three main zones, the western, central and eastern Elba, respectively (Fig. 3.4). These zones and their bounding faults are the key to the reconstruction of the original geometry of the intrusive complexes.

**Western Elba and the Eastern Border Fault**

Western Elba consists of the Monte Capanne pluton and its thermometamorphic carapace of Complex IV rocks containing hypabyssal porphyry intrusions. It is separated from central Elba by the Eastern Border fault that parallels the east side of the Monte Capanne pluton and put in contact the pluton’s thermally metamorphosed host rock of Complex IV with the unmetamorphosed flysch of Complex V (Fig. 3.4). The Eastern Border fault is marked, for the most part, by a distinct surface that dips moderately to steeply to the east. This fault separates a western footwall breccia of hornfelsed Complex IV rocks (ophiolitic material and deep marine cover rocks) plus fragments of the Monte Capanne pluton, locally mineralised by quartz and hematite, from an eastern hanging wall breccia made of Complex V flysch and megacrystic San Martino porphyry of late Miocene age.

Movement on the Eastern Border fault was "west side up" and juxtaposed western rocks from 4-5 km depth (Dini et al., 2002) with shallowly buried sedimentary rocks and their enclosed porphyries on the east side (Figs. 3.4 and 3.5).

**Central Elba and the Central Elba fault**

Central Elba is bounded to the west by the steep Eastern Border Fault and is separated from eastern Elba by the low-angle Central Elba fault, marked by a zone containing a tectonic mélange of rocks from Complexes IV and V (Bellincioni, 1958; Perrin, 1975; Trevisan, 1950), most notably rocks whose equivalents crop out in western Elba. These are (i) thermally metamorphosed serpentinite and basalt (Marinelli,
1955), and garnet- and wollastonite-bearing marble (Vom Rath, 1870), identical to rocks in the Monte Capanne contact aureole, and (ii) tourmaline-free aplite porphyry and K-feldspar phenocryst-bearing porphyry. The fault dips gently westward, as does the dominant fabric of the rocks resting on it, such that the highest part of the section occurs at the western edge against the steeply-dipping Eastern Border Fault (Figs. 3.4 and 3.5).

**Eastern Elba and the Zuccale Fault**

In eastern Elba, a younger low-angle detachment fault, the Zuccale fault, has been documented with eastward transport of 5-6 km (Keller and Pialli, 1990; Pertusati et al., 1993). This fault movement post-dates that on the Central Elba fault since it sliced off the leading edge of the Central Elba fault to produce a klippe of Complex V rocks in eastern Elba (Figs. 3.4 and 3.5).
Fig 3.5 – Geological map of Elba Island draped over a Digital Elevation Model.
4. Rise and Fall of the Elba Intrusive Complexes

The present setting of the Elba intrusive complexes is the result of the intrusive and tectonic-gravitational evolution that took place mainly during the late Miocene. This evolution consisted in the build-up of an intrusive complex in western Elba, soon followed by its fast tectonic-gravitational dismemberment. An analogous sequence of intrusive-tectonic events took place in eastern Elba at the very end of the Miocene.

In this chapter the build-up of the intrusive complexes (rise) and their tectonic dismemberment (fall) are outlined. The complete details of each intrusive unit are reported in the following specific chapter devoted to the description of the intrusive complexes.

Western Elba Intrusive Complex

The large-scale faults subdividing Elba Island into three main zones (western, central and eastern Elba, Fig. 3.4) are the key to the reconstruction of the original geometry of the intrusive complex. The portions of the Elba igneous complex cropping out in western and central Elba have not preserved their original emplacement geometry. All the intrusive units were emplaced within the tectonic Complexes IV and V when they were stacked above the present western Elba.

Build-up of the Multilayer Laccolith Complex

The igneous sequence of western-central Elba started with the construction of a multilayer laccolith complex, first by the emplacement of the layers of Capo Bianco aplite and Nasuto microgranite, followed in succession by the layers of Portoferraio laccolith and, finally by the intrusive layers San Martino laccolith.

Emplacement of the Monte Capanne Pluton

The deepest layers of this complex were then intruded and/or deformed by the Monte Capanne pluton and its associated late leucocratic dykes and veins.
Intrusion of the Orano dykes

Finally, about 200 mafic dykes of the Orano swarm were emplaced, cutting through the entire succession.

Tectonic Splitting of the Intrusive Complex

Shortly after the intrusion sequence was completed, the upper part of the igneous-sedimentary complex was tectonically translated eastward along the Central Elba Fault (CEF), so that the lower part is presently found in western Elba, while the upper part is in central Elba (Westerman et al., 2004) (Fig. 4.1). Following this eastward translation, a "west side up" movement occurred along the Eastern Border fault with a throw of 2 to 3 km (Fig. 4.1). This history is supported by (i) the low angle CEF at the base of the slice consisting of igneous layers and their host rocks in central Elba, (ii) shear fabrics, including E-W striations, in the fault mélange of the CEF, (iii) fragments of rocks typical of western Elba in the footwall mélange, (iv) the match of petrographic and geochemical features of the intrusive units in central and western Elba, and (v) the strongly preferred E-W orientation of Orano dykes and their restriction to the northern areas in both western and central Elba.

The minimum amount of displacement along the central Elba fault (CEF) is constrained to about 8 km by the distance from the pluton’s aureole eastward to its leading edge where fragments of hornfels from that aureole occur in the fault mélange.

The rate of displacement is constrained by the time between onset of the movement along the Central Elba fault (ca. 6.8 Ma) and the time when cobbles were deposited (close to the end of Messinian, i.e. before 5.3 Ma). Allowing for erosion and transport, a maximum estimate for the time available for the main movement on the Central Elba fault is less than 1.5 million years. Thus the eastward translation of at least 8 km occurred at an average rate in excess of 5-6 mm/yr. This movement rate is higher than movement rates reported for detachment faults (Stockli et al., 2001), while it is consistent with rate associated with gravity gliding (Fletcher and Gay, 1971), possibly triggered by magma emplacement (Merle and Vendeville, 1995).

The timing of displacement on CEF is constrained by (i) the occurrence of fragments of Monte Capanne hornfels in the footwall mélange of CEF some 8 km east of the nearest outcrop of thermal aureole, indicating that movement on the CEF occurred after contact metamorphism linked to the Monte Capanne pluton (ca. 6.8 Ma), and (ii)
the matching distribution of Orano dykes in western and central Elba (Fig. 1.5) suggesting that the Orano dykes were truncated and translated by the movement along the CEF, that therefore took place after their emplacement at 6.85 Ma. The timing in this scenario is further constrained by the occurrence on mainland Tuscany –50 km to the east– of cobbles and boulders of tourmaline-bearing Capo Bianco aplite and Portoferraio porphyry in conglomerates deposited close to Messinian end (Bossio et al., 1993; Marinelli et al., 1993; Testa and Lugli, 2000).

Fig 4.1 - Fall of the nested Christmas-tree laccolith complex: unroofing of the Monte Capanne pluton showing progressing stages of décollement.
Indeed, Capo Bianco aplite and Portoferraio porphyry layers were concentrated just above (and presumably below) the CEF, and the most logical mechanism to expose them without exposing the overlying San Martino porphyry units was by erosion of tilted layers (today’s tilt of laccolith layers is about 30°; Fig. 6h).

The eastward displacement of the upper part of the complex is at least partly linked to gravitational instability. In about 1 Ma, a 2,700 m thick tectonostratigraphic section was inflated by the addition of at least 2,400 m of laccolithic intrusions, leading to a total thickness for the new section of about 5,000 m. A dome with a 10 km diameter and a height of 2.5 km, was produced with a surface slope of about 25° (assuming an originally flat surface). We envision that emplacement of the Monte Capanne pluton beneath this dome caused oversteepening and triggered the main eastward displacement of the upper section. Once significant movement began, transfer of the load from above Monte Capanne towards central Elba promoted movement on the east-dipping Eastern Border fault as the unloaded pluton rose and the thickened central Elba section subsided. Final movement on the Eastern Border fault took place entirely in the brittle regime, truncating the Central Elba fault that has since been eroded in western Elba and lies almost completely buried below central Elba.

Evidence from experimental analogue models (Merle and Vendeville, 1995; Roman-Berdiel et al., 1995) and structural observations of natural examples (Gucwa and Kehle, 1978; Reeves, 1925) shows that laccolith-type magmatic intrusions can produce stresses large enough to induce thrusts and folds in the adjacent sedimentary rocks. In particular, the most efficient process seems to be gravity gliding, during which layer-parallel compression can result as rocks glide away from the topographic high created by laccolithic intrusion.

**Eastern Elba Intrusive Complex**

The eastern Elba Island tectonic stack was intruded by magmatic bodies at 5.9-5.8 Ma. A pluton, a few mafic dykes and a widespread association of felsic tabular intrusions and hydrothermal veins-masses mainly occur in the Calamita schists (Fig. 1.5), consisting of psammitic-pelitic hornfels along with minor interlayered amphibolite bodies.
**Intrusion of the Pluton**

The Porto Azzurro pluton is a megacrystic monzogranite (Conticelli et al., 2001; Innocenti et al., 1992). The pluton has only limited exposure in the south-central part of eastern Elba (Fig. 1.5), but a significant size is suggested by the extent of thermometamorphosed Calamita schists to the south. The emplacement age of the pluton is constrained at 5.9 Ma (Maineri et al., 2003). A shoshonitic dyke similar to Orano dykes was emplaced in eastern Elba at 5.8 Ma (Conticelli et al., 2001). Thus, a pattern of mafic, potassic dykes after a main pluton occurred in western Elba at 6.9-6.8 Ma, and was repeated in eastern Elba 1 Ma later.

**Emplacement of Felsic dykes**

Several generations of felsic dykes and sills crop out in the eastern part of the Calamita peninsula and in the Porto Azzurro area. They strike N40E - N160E, with sub-horizontal (sills) to sub-vertical (dykes) attitudes and para-concordant to discordant geometric relationships with host rock foliation. They invariably have a leucogranitic composition, with ubiquitous tourmaline and mutually exclusive occurrence of primary muscovite or biotite. Hydrothermal tourmaline-quartz veins cut across dykes and sills and represent a hydrothermal circulation affecting both the intrusive and hornfels rocks (Dini et al., 2008a).

**Ore Minerals**

The outcrop area of the Eastern Elba Intrusive Complex significantly overlap with the area affected by Fe-ore deposits. These deposits are found in a narrow N-S trending belt along the eastern coast of Elba Island (Dini, 2003). The two main mining areas have been for almost three millennia around the Rio Marina village (hematite) and the southern part of the Monte Calamita peninsula (magnetite).

**Tectonic Splitting of the Complex**

The tectonic evolution of the intrusive complex in eastern Elba took place following a sequence of events similar to those described for western Elba, leading to fast pluton exhumation (Keller and Pialli, 1990; Pertusati et al., 1993). Initially, dominantly sub-horizontal movement on the low-angle detachment Zuccale Fault with top-to-the east sense of shear translated the overlying rocks eastward, trimming out part of the contact aureole of the Porto Azzurro pluton as well as a 2 km wide slice.
from the front edge of the Central Elba Fault. Then, high-angle structures were activated at the eastern edge of the pluton, off-shore the eastern Elba coast.

**Summary**

The common history of western-central and eastern Elba is presented schematically in Figure 4.2 and can be summarized as follows. Initially, dominantly sub-horizontal movement on a low-angle detachment fault with top-to-the-east sense of shear (Central Elba and Zuccale faults) translated the overlying rocks eastward, trimming out part of the contact aureole of the pluton with which it is associated. Then, high-angle structures were activated mainly at the eastern edge of the pluton, i.e. the Eastern Border fault east of the Monte Capanne pluton and off-shore faults east of the Porto Azzurro pluton (Bortolotti et al., 2001). In this scenario, pluton emplacement occurred before the main faulting, and promoted its activation. This inference challenges the hypothesis that eastward shear above the pluton was synchronous with pluton emplacement, accommodating the magmatism and controlling the level of emplacement of the Monte Capanne and Porto Azzurro plutons (Daniel and Jolivet, 1995; Jolivet et al., 1998).

![Schematic summary of late Miocene fault activities at Elba.](image-url)
5. Genesis of Elba Magmas

Geochemistry

The intrusive units of the Elba igneous complexes (Table 5.1) display limited intra-unit compositional variations, but significant inter-unit geochemical variability (Table 5.2, Fig. 5.1) (Dini et al., 2002).

The intrusive units of the western Elba multilayer laccolith complex are more acidic than the Monte Capanne pluton. The Capo Bianco aplitic has extremely low TiO$_2$, FeO$_{tot}$, MgO and CaO, and a strong...
peraluminosity (average ASI = 1.42 ± 0.10, Table 5.2) with notably high Be, Cs, Rb, Nb and Ta. The Nasuto microgranite, the Portoferraio porphyry and San Martino porphyry have overlapping major element characteristics. The Nasuto microgranite is the most peraluminous of the three units (ASI = 1.29) and has slightly higher alkali content (Fig. 5.1). Major element chemistry of the monzogranitic Portoferraio porphyry (ASI = 1.18 ± 0.07) and San Martino porphyry (ASI = 1.22 ± 0.16, Table 5.2) overlap in every respect, although Portoferraio rocks are, on average, more alkaline and richer in SiO₂ and Al₂O₃ and have biotites with lower Fe no. (Fig. 5.2).

The Monte Capanne pluton shows restricted chemical variations (Fig. 5.1). The whole pluton has monzogranite composition and slightly peraluminous character (average ASI = 1.11 ± 0.05 1sd). Despite this apparent chemical homogeneity, three different facies are recognized petrographically (Farina, 2007). The trace element distribution of the Monte Capanne pluton has low variability and the distribution of REE is homogeneous throughout the pluton.

![Fig. 5.1. (a) Total alkali vs. silica (TAS) classification diagram (Le Bas et al., 1986). (b) TAS enlargement for Orano dykes; tie-lines connect core (large symbol) and rim (small symbol) of single dykes. Modified after (Dini et al., 2002).](image)
Table 5.2 - Average major element chemical compositions for late Miocene intrusive units from Elba Island.

<table>
<thead>
<tr>
<th>unit</th>
<th>Capo Biance</th>
<th>Nasuto</th>
<th>Portoferroio</th>
<th>San Martino</th>
<th>Monte Capanne</th>
<th>Monte Capanne</th>
<th>Orano normal</th>
<th>Orano hybrid</th>
<th>Porto Azzurro</th>
<th>Monte Castello</th>
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<tbody>
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<td></td>
<td>apite</td>
<td>microg.</td>
<td>porphyry</td>
<td>porphyry</td>
<td>pluton</td>
<td>leucogranites</td>
<td>dyke swarm</td>
<td>dyke swarm</td>
<td>pluton</td>
<td>dyke</td>
</tr>
<tr>
<td>n. samples</td>
<td>6</td>
<td>10</td>
<td>23</td>
<td>9</td>
<td>8</td>
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<td>2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>73.07</td>
<td>69.10</td>
<td>69.92</td>
<td>68.78</td>
<td>67.46</td>
<td>70.38</td>
<td>65.08</td>
<td>67.46</td>
<td>70.35</td>
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<td>0.29</td>
<td>0.36</td>
<td>0.56</td>
<td>0.56</td>
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<td>0.56</td>
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<td>Al₂O₃</td>
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<td>15.82</td>
<td>16.28</td>
<td>15.89</td>
<td>16.09</td>
<td>15.76</td>
<td>15.81</td>
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<td>15.15</td>
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<td>0.50</td>
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<td>0.79</td>
<td>0.53</td>
<td>1.08</td>
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<td>0.95</td>
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</tr>
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<td>FeO</td>
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<td>3.75</td>
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<tr>
<td>Na₂O</td>
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<td>3.60</td>
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<td>ASI</td>
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<td>1.18</td>
<td>1.22</td>
<td>1.10</td>
<td>1.15</td>
<td>1.09</td>
<td>1.15</td>
<td>0.85</td>
<td>1.84</td>
</tr>
</tbody>
</table>

The late-plutonic felsic products are chemically distinct from the Monte Capanne pluton and from each other (Fig. 5.1). The peraluminous Cotoncello dyke sample (ASI = 1.16) plots close to the silica-rich end of the Monte Capanne field and overlaps the compositional fields of the Portoferreaio and San Martino porphyries. In contrast, the leucogranite dykes have the highest SiO$_2$ and lowest Al$_2$O$_3$ contents of the analysed rocks from the Elba magmatic complex.

The post-plutonic Orano dyke swarm includes two compositional groups, monzogranitic samples from the core of thick zoned dykes, and granodioritic samples from borders of those dykes and from unzoned dykes (Table 5.1; Fig. 5.1). Compositions of the monzogranitic samples are intermediate between the Monte Capanne pluton and the granodioritic Orano group, and these xenocryst-rich rocks are interpreted as mixtures of Orano melt and solids from the Monte Capanne crystal mush, formed during transfer-emplacement of the dykes. The Orano porphyry has the lowest SiO$_2$ content of the Elba intrusive units and very distinctive overall trace element distribution, with extreme enrichment in Sr, Ba, and LREE, oddly coupled with high Ni and Cr contents.

The intrusive and chronological relationships of the Elba igneous complex, coupled with the overall geochemical features and the strongly variable isotopic ratios, allow to recognize three isotopically distinct groups of rocks (Dini et al., 2002).
Group 1 consists of the oldest units of the complex, i.e. the ca. 8 Ma intrusions of the Capo Bianco aplite, Nasuto microgranite and Portoferraio porphyry, along with the late-plutonic Cotoncello dyke, and is characterized by low $\varepsilon_{\text{Nd}(t)}$ values between $-9.5$ and $-10.0$ and strongly variable Sr isotopic ratios (0.7115 - 0.7228) (Figs. 5.4 and 5.5).

Group 2 is isotopically homogeneous with relatively high $\varepsilon_{\text{Nd}(t)}$ values between $-8.1$ and $-8.9$, and $^{87}\text{Sr}/^{86}\text{Sr}$ varying in the range 0.7131 - 0.7162 (Figs. 5.4 and 5.5). This group consists of the San Martino porphyry, the Monte Capanne pluton, the leucogranite dykes, and the mafic microgranular enclaves. These units were emplaced at ca. 7 Ma, and represent the bulk of the complex with a volume on the order of about 170 km$^3$.

Group 3 rocks, namely the Orano porphyry dykes, constitute the youngest units of the complex and are characterized by strongly variable isotopic compositions, including the lowest Sr and highest Nd isotope ratios of the complex: $\varepsilon_{\text{Nd}(t)}$ varies between $-7.0$ and $-10.1$, and initial Sr isotopic ratios ranges between 0.7114 and 0.7138 (Figs. 5.4 and 5.5).

Fig. 5.3 - Initial $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ plot for studied samples and Tuscan Magmatic Province samples from the literature. Modified after Dini et al. (2002).
Fig. 5.4. ASI (Alumina Saturation Index) vs. $\varepsilon_{\text{Nd}(t)}$ showing isotopically distinct groups of rocks from the Elba igneous complex. Modified after Dini et al. (2002).

**Petrogenesis**

**Group 1 - Crustal messengers**

All the older (ca. 8 Ma) pre-plutonic intrusive units, and the younger, late-plutonic Cotoncello dyke, have compositional features such as strong peraluminosity, high SiO$_2$ content, and low concentrations of ferromagnesian elements, that matches those of both natural and experimental melts derived from metasedimentary crustal sources (Patiño Douce, 1999).

For the Capo Bianco aplite, both major and trace element compositions of those rocks having the lowest Rb/Sr ratio, Al$_2$O$_3$ content, and Na$_2$O/K$_2$O are comparable with compositions obtained for experimental melts from muscovite dehydration melting of pelitic sources at 0.8 GPa and 800°C (Patiño Douce and Harris, 1998). Moreover, muscovite dehydration melting can account for the deep negative Eu anomaly in a flat REE pattern with low LREE content. In fact, monazite, apatite and zircon are commonly armoured by biotite...
and, therefore, do not supply trace elements to the liquid, thus producing a Zr-P-LREE-poor melt (Nabelek and Glascock, 1995). Such a melt would have a negative Nd anomaly linked to the high partition coefficient of Nd in the restitic monazite, relative to its neighbours Ce-Pr and Sm (Yurimoto et al., 1990). In summary, samples of the Capo Bianco aplite are interpreted as a direct product of muscovite dehydration melting of a metapelitic source.

The Nasuto microgranite and Portoferraio porphyry also belong to the 8 Ma emplacement stage, and are, like the Capo Bianco aplite, strongly peraluminous. Such a strong “crustal” imprint suggests a direct derivation by crustal melting, and the significant content of ferromagnesian elements agrees with a process that involved biotite dehydration melting. Furthermore, the CaO enrichment in these rock, relative to ferromagnesian components, is consistent with derivation from a biotite-plagioclase-rich source such as metagreywacke, rather than a simple biotite-rich metapelite (Patiño Douce, 1999). The involvement of biotite together with its accessory minerals (e.g. monazite) can explain the more typical, “crustal” REE pattern of these units.

Finally, the Cotoncello dyke is also definitely peraluminous and, as in the case of the Nasuto microgranite and Portoferraio porphyry, its distribution of major elements and REE pattern are consistent with derivation from a biotite-plagioclase-rich source. Furthermore, its high Sr isotopic composition points to a crustal origin and rules out derivation as a fractionation product of the Monte Capanne pluton.

Group 2 – Hybrids

The youngest pre-plutonic unit, the San Martino porphyry, displays geochemical and petrographic features similar to the Monte Capanne pluton, although the former is slightly more peraluminous and richer in SiO₂. Tonalitic mafic microgranular enclaves are common in this porphyry and the Sr-Nd isotopic composition of one of them is indistinguishable from that of the Monte Capanne pluton. This indicates that hybrid melts were available and mingled with peraluminous melts during emplacement of the San Martino magmas.

The Monte Capanne pluton is an acidic, voluminous body with homogeneous geochemical features. Its low silica content, along with an ubiquitous presence of abundant mafic microgranular enclaves with isotopic ratios equal those of the pluton, amphibole clots replacing
pyroxene, and plagioclase with calcic cores, all point to the involvement of more than one magma in the genesis of this pluton (Poli, 1992).

The late leucogranite dykes of the Monte Capanne pluton have overlapping Sr and Nd isotopic compositions with the pluton itself. Thus, the chemical variability within the the leucogranite group can be explained by removal of an oligoclase+biotite+K-feldspar assemblage from the least acidic, highest-Sr leucogranite samples, which in turn derive by removal of a biotite-andesine assemblage from the average composition of Monte Capanne pluton. The lack of K-feldspar in this fractionating assemblage makes the San Piero facies the most likely parent for the leucogranite melts since it was only in that facies that K-feldspar entered the liquidus assemblage late.

**Group 3 - Mantle messengers**

The Orano dyke swarm consists of two petrographic types. The monzogranitic type is rich in xenocrysts captured from an acid crystal mush that matches well with the Monte Capanne pluton: therefore, this group is not a suitable source of information about original magmas. The granodioritic to quartz monzodioritic type, with only minor xenocrysts, still displays evidence for melt hybridism. However, these melts did contain phlogopite, as well as olivine and/or clinopyroxene, as a part of their liquidus assemblage. Additionally, these rocks have high Ni and Cr content coupled, in some samples, with extreme enrichment in Sr, Ba and LREE, and strong REE fractionation.

Two samples from the Orano porphyry have been compared with mantle-derived magmas from the Tuscan Magmatic Province, i.e. the K-rich andesites from Capraia Island and lamproites from mainland Tuscany (Fig. 5.5). These products all have similar overall distributions of incompatible trace elements, with highly fractionated N-MORB normalized patterns, very high contents of the most incompatible elements, and high LILE/HFSE ratios. However, the Orano dykes most closely resemble the Capraia K-andesites with which they share overlapping Th/Ta values (Fig. 5.5). In addition, they lack the negative Sr anomaly which characterizes the lamproites.

Regarding the isotopic data, $^{87}\text{Sr}/^{86}\text{Sr}(t)$ and $^{143}\text{Nd}/^{144}\text{Nd}(t)$ values of the Orano porphyry constitute the limits found in Elba intrusive rocks, i.e. 0.71145 and 0.51227, respectively. These values are quite extreme for mantle-derived magmas; however, they are coupled with very high Sr and Nd contents (1461 and 94 ppm respectively). Because crustal
materials generally have much lower Sr and Nd contents, we conclude that these Orano isotopic ratios are very close to the original values of mantle-derived magma. Therefore, the anomalous isotopic ratios and trace element distribution are taken as evidence for an origin of Orano melts from a strongly modified mantle source, as is generally envisaged for the Tuscan Magmatic Province (Peccerillo and Panza, 1999; Serri et al., 1993).

Fig. 5.5 - N-MORB normalized spidergrams (Sun and McDonough, 1989) of the most incompatible element-enriched Orano dyke (2 samples), compared with K-rich andesites from Capraia Island (M. D’Orazio, unpublished data) and lamproites from the Tuscan Magmatic Province (Conticelli et al., 1992). Modified after Dini et al. (2002).
Magma sources and processes

The isotopic compositions of the crust-derived Group 1 units are significant in understanding the nature of the crust involved as sources for the Elba magmas. These units have low $\varepsilon_{\text{Nd}(t)}$ values, coupled with strongly variable Sr isotope ratios (Figs. 5.3 and 5.4). Two main processes can be invoked to explain these isotopic features, namely melting of independent crustal sources with different $^{87}$Sr/$^{86}$Sr ratios and uniform $\varepsilon_{\text{Nd}(t)}$ values, or melting of a Sr-Nd isotopically homogeneous crustal source under conditions able to differentiate the Sr isotopic ratios of the melt. The latter process can occur in the case of incongruent fluid-absent melting. Muscovite and/or biotite dehydration melting also involves K-feldspar and plagioclase, which have Rb/Sr ratios substantially lower than those of micas. The lack of a fluid phase can yield melts with a $^{87}$Sr/$^{86}$Sr ratio either higher or lower than that of the source (Harris and Ayres, 1998; Inger and Harris, 1993), depending on the modal proportions of the phases entering the melt (mainly K-feldspar, plagioclase and mica) and on their Sr concentrations. Isotopic modelling (Barbero et al., 1995) shows that when plagioclase makes up >30 modal% of the phases entering the melt, the melt has a $^{87}$Sr/$^{86}$Sr ratio lower than that of the protolith, with a difference up to 0.004. This type of process could be invoked to explain the differences in Sr isotopic ratios between the Capo Bianco, Nasuto and Portoferaio units (<0.003), but it is not able to explain the large difference in $^{87}$Sr/$^{86}$Sr ratio (0.011) occurring between the Capo Bianco and Cotoncello units.

On the basis of the above discussion, it is possible to conclude that the crustal sources activated at around 8 Ma and 7 Ma were different, at least isotopically (Dini et al., 2002). A unique crustal source (Tuscan Crust-C1; Fig. 5.6) could have been activated in the initial phases of melting to produce the Capo Bianco, Nasuto and Portoferaio magmas. In contrast, the source for the Cotoncello dyke represents a separate component (Tuscan Crust-C2; Fig. 5.6) that is isotopically comparable with the crust-derived San Vincenzo rhyolites of the Tuscan Magmatic Province (Feldstein et al., 1994; Ferrara et al., 1989).
Fig. 5.6 - (a) $^{87}\text{Sr}/^{86}\text{Sr}$ vs. 1000/Sr diagram showing possible crustal and mantle-derived components, along with possible mixing trajectories able to produce the cluster of hybrid samples with Sr≈200 ppm and $^{87}\text{Sr}/^{86}\text{Sr}≈0.7145$. Note that a mixing plus fractional crystallization process with a petrologically likely bulk DSr close to 1, would lead to the same path as pure mixing. (b) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. 1000/Nd diagrams showing possible crustal and mantle-derived components, along with possible mixing trajectories able to produce the cluster of hybrid samples. Nd content for Giglio Le Scole extrapolated from La and Ce contents.

The isotopically homogeneous, hybrid Group 2 includes most of the Elba igneous rocks. This hybrid system had to be large and homogeneous in order to sustain production of such significant amounts of compositionally constant magma. The general petrogenetic relationships of the materials involved in this process are illustrated in the $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ diagram of Figure 5.3, where the Elba data are plotted together with isotope variation fields of the whole Tuscan Magmatic Province. Overall, the available data define two curvilinear trends pointing on the right toward two distinct crustal end-members with high Sr- and low Nd-isotope ratios, and, on the left, toward a mantle component with relatively low Sr and high Nd isotope ratios. The upper trend corresponds well with the mixing trend described for the San Vincenzo volcanics that relates mafic microgranular enclaves and their cordierite-bearing rhyolite host, considered as a nearly pure crustal melt (Ferrara et al., 1989). The lower trend points to a crustal component with a lower Nd isotopic ratio, comparable with that observed in some Tuscan basement rocks (outcrops, cores and xenoliths; Fig. 5.3). Therefore, the occurrence of the two trends suggests that more than one crustal component has been involved in Tuscan Magmatic Province magmatism. The most likely candidates mantle-derived components involved in the mixing are the K-rich andesites and basalts from Capraia as well as the mafic microgranular enclaves from San Vincenzo rhyolites.

Additional insight into the evolution of the hybrid rocks of the Monte Capanne pluton and San Martino porphyries comes from examination of a $^{87}\text{Sr}/^{86}\text{Sr}$ vs. 1000/Sr diagram (Fig. 5.6a), in which most samples cluster at Sr≈200 ppm and $^{87}\text{Sr}/^{86}\text{Sr}$ r≈0.7145. These values can, therefore, be regarded as the most likely to have developed during the hybridization process. This cluster of hybrid compositions can be
obtained by a mixing process involving the Cotoncello dyke magma and Capraia K-andesites (close in space and time to the Elba igneous activity) whose source is here called Tuscan Mantle-M1 (Fig. 5.6b). Moreover, the Capraia-Cotoncello mixing trend and its end-members are very similar to those inferred to explain San Vincenzo magmatism, where hybridism is well constrained (Ferrara et al., 1989).

Examination of the plot of $^{143}\text{Nd}/^{144}\text{Nd}$ vs. 1000/Nd (Fig. 5.6b) confirms the above interpretation and suggests further insights concerning the origin of isotopic variations internal to the Orano porphyry whose samples plot well off the Capraia-Cotoncello mixing trend. The analysed samples of Orano dykes are distributed along a main linear trend with a few samples shifted horizontally into the field of hybrid magmas (Capraia-Cotoncello mixing trend). The steep trend starts from the LILE-richest sample previously identified as isotopically similar to the mantle end member for Orano magmas (from Tuscan Mantle-M2), and points to a felsic end member with a very low Nd isotopic ratio, even lower than the Group 1 end members, thus identifiable as coming from a different crustal source (Tuscan Crust-C3; Fig. 5.6b). A possible representative of this melt can be found in the late crustal leucogranite of Giglio Island (le Scole intrusion; Westerman et al., 1993). The horizontally shifted (Nd depleted) samples are interpreted as originating from a solid-liquid mixture with Monte Capanne magmas. Thus, Orano magma does not represent the mantle component of the Group 2 hybridization process, and its internal variations are instead linked to shallow level, more or less thorough mingling with a granitic crystal mush, probably during emplacement of the dykes as they passed through the Monte Capanne pluton.

In summary, the $^{87}\text{Sr}/^{86}\text{Sr}$ vs. 1000/Sr and $^{143}\text{Nd}/^{144}\text{Nd}$ vs. 1000/Nd diagrams illustrate that the most voluminous Group 2 rocks represent a hybridization trend between Capraia (from M1) and Cotoncello (from C2) end-member magmas. Group 3 Orano rocks represent a mix between a different mantle-derived magma (from M2) having geochemical-isotopic characteristics intermediate between Capraia K-andesites and Tuscan Magmatic Province lamproites, and a different crustal magma (from C3; low $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic ratio), with additional varieties derived from mingling with Group 2 magmas. In the same framework, the Group 1 Capo Bianco aplite, Nasuto microgranite and Portoferraio porphyry are crustal melts (from C1) that did not contribute to the main hybridization processes (Dini et al., 2002).
Summary

The general igneous framework of western-central Elba consists of an acidic laccolith complex intruded by a monzogranitic pluton and a slightly younger swarm of more mafic dykes. Figure 14 schematically summarizes the sequence of events that produced the Elba igneous complex (Dini et al., 2002). Capo Bianco aplite melts were produced first, apparently by muscovite dehydration melting of a metapelitic source, identified as C1 (Fig. 5.6b, inset). As the anatectic process continued, the Nasuto microgranite and Portoferraio porphyry likely originated via biotite dehydration melting of a metagreywacke source. The earliest magmas were produced without any chemical contribution of mantle melts, with melting probably linked to late Miocene lithospheric thinning and decompression, following earlier Oligocene-Miocene orogenic overthickening.

Fig. 5.7 - Schematic model for the genesis of Elba magmatic products throughout the time. Abbreviations: MME: mafic microgranular enclaves; i.r.: initial isotopic ratio. Modified after Dini et al. (2002).

After a period of quiescence, the volumetrically important phase of hybrid magmatism began, involving mantle-derived mafic magmas and peraluminous crustal melts. Mantle magma involved in this phase is never directly represented, being present most prominently as hybrid products preserved in mafic microgranular enclaves. This mantle magma is thought to be similar to the nearby coeval Capraia K-andesites and has been identified as M1. The crustal component of the hybridization (C2) was derived from a different crustal source than the peraluminous melts.
of the first phase, and is represented by the Cotoncello dyke. The first voluminous intrusion of the hybrid group, the San Martino porphyry, was produced by incipient melt hybridism. Next came the Monte Capanne pluton from a fully mature hybrid system, with internal facies representing emplacement pulses, and leucogranite dykes derived by fractional crystallisation. Finally, Orano magmas were generated from strongly modified mantle (M2), as products distinctly different than those involved in the earlier main hybridization process. During their ascent, they were first variably hybridized by mixing with a unique crustal material (C3), and then further modified by capturing material from the Monte Capanne system.

The magma formation processes recorded between ca. 8 and 6.8 Ma by the Elba magmatism, changed from crust-, to hybrid-, to mantle-dominated, as the Apennine fold belt was progressively thinned, heated and intruded by mafic magmas during late Miocene time. Very unusual melts emplaced at the beginning and at the end of the igneous activity were not volumetrically significant and did not contribute to the generation of main hybrid magmas. They do, however, emphasize the highly variable nature of crustal and mantle sources that can be involved, during a short time span, in post-collisional, extension-related magmatism.
6. The Western-Central Elba Laccolith Complex

Introduction

The Western-Central Elba Laccolith Complex was built-up by an intrusive sequence which took place between about 8 and 7.4 Ma. This sequence opened the late Miocene igneous activity at Elba Island. Detailed mapping, correlations based on petrographic and geochemical data, measured geological cross sections, along with structural data, led to the reconstruction of the original geometry and the emplacement history and mechanisms for the multilayer laccolith complex.

The intrusive units

**Capo Bianco aplite**

The Capo Bianco aplite is a white porphyritic rock with alkali feldspar granite compositions (Table 5.2). In western Elba, the outcrops occur between ultramafic rocks and argillaceous units of Complex IV, and constitute five adjacent but isolated caps (Figs. 6.1, 6.2 and 6.3) that were likely emplaced as a single sill subsequently dismembered by younger intrusions. In central Elba, a tourmaline-rich layer of Capo Bianco aplite (Fig. 6.1) was originally intruded at higher level with respect to the western Elba layer, along the tectonic contact between Eocene calcarenite and the overlying Cretaceous flysch. This layer was subsequently dismembered and encased within the later Portoferraio porphyry, as it can be seen in the northern part of central Elba (Figs. 6.1 and 6.3). Limited exposures in eastern Elba are within a block of Complex V flysch floored by the Zuccale fault.

Whole rock-muscovite Rb-Sr isochrons (Ms phenocrysts >350 µm selected to avoid secondary sericite) yielded ages of 7.95±0.1 Ma and 7.91±0.1 Ma for samples from western and central Elba, respectively (Dini et al., 2002). A 40Ar-39Ar age older than 8.5 Ma for late magmatic muscovite from central Elba is reported (Maineri et al., 2003). In minor outcrops of Capo Bianco aplite which experienced further eastward tectonic translation to eastern Elba (Pertusati et al., 1993), hydrothermal recrystallization is dated to 6.7±0.1 Ma (Maineri et al., 2003).
Fig. 6-1 - Geological map of the Western-Central Elba Laccolith Complex (Dini et al., 2006).
Fig. 6.2 – Geological map of the western-central Elba laccolith complex draped over a digital elevation model. View from the south.
Fig. 6.3 – Geological map of the western-central Elba laccolith complex draped over a digital elevation model. View from the north.
Nasuto Microgranite

The Nasuto microsyenogranite (Table 5.1 and 5.2) has a syenogranitic composition, and crops out over an area of 0.5 km² along the northern shore of western Elba (Fig. 6.1). It is entirely surrounded, as well as intruded, by the younger Portoferraio porphyry, such that its primary intrusive contacts are lost.

Portoferraio porphyry

The Portoferraio porphyry has dominantly monzogranitic, with minor syenogranitic, compositions (Table 5.1 and 5.2; Fig. 5.1), with biotite having Fe no. ranging from 0.45 to 0.50 (Fig. 5.2). These porphyries occur as major laccolith layers up to 700 m thick, commonly interconnected, and as numerous dykes (Westerman et al., 2000). The Portoferraio porphyry contains prominent phenocrysts of sanidine and has monzogranite to syenogranite compositions (Table 5.2). It occurs as four major layers up to 700 m thick, commonly interconnected and accompanied by minor dykes and sills (Figs. 6.1, 6.2 and 6.4). E-W trending dykes connected to the floor of an intrusive layer are exposed on the western side of the Golfo di Lacona (Figs. 6.1 and 6.2). Three major layers occur in western Elba. The lowest two layers, with maximum thicknesses of about 75 m, intruded Complex IV metabasalts parallel to the ENE-striking tectonic fabric. A higher layer of Portoferraio porphyry was emplaced between hornfelsed argillite above and ophiolitic rocks below. It intruded at the same level previously exploited by the Capo Bianco aplite, and now encases and crosscuts the Capo Bianco aplite and the Nasuto microgranite (Fig. 6.4). This layer terminates to the SW against the Monte Capanne pluton, but was likely connected with the Chiessi outcrops (Fig. 6.1) before the Monte Capanne pluton intruded and deformed the porphyry. Close to the pluton contact, the groundmass in this layer exhibits a strong mylonitic foliation that strikes parallel to the contact. Quartz phenocrysts here have subgrain boundaries, plagioclase cracks are cemented by micrographic quartz plus K-feldspar, and primary biotite is transformed to oriented polycrystalline aggregates. The total length of the layer was, therefore, in excess of 9 km, yet only 3 km are presently exposed along strike in the north-eastern part of western Elba.
Fig. 6.4 - Geological cross sections of the Western-Central Elba laccolith complex.
The fourth layer of Portoferraio porphyry occurs in central Elba, between Cretaceous flysch above and Eocene calcarenites below. The upper surface of this NNW-striking layer is generally subparallel to bedding in the overlying flysch that dips moderately WSW. As in western Elba, the Portoferraio porphyry here surrounds the older Capo Bianco aplite, with dykes truncating tourmaline laminations and mineralized fracture surfaces in the aplite. The maximum thickness of the layer, as calculated from cross sections, is about 400 m. Along the southern shore of central Elba, this layer splits into five tapering branches. Only small dykes and sills penetrated the overlying Cretaceous flysch. Visible contact metamorphic effects in the host rock are restricted to local biotite crystallization over a mm-thick layer at the contact with the igneous rocks.

A Rb-Sr wr-Bt isochron points to an age of 8.4±0.1 Ma, but internal isotopic disequilibrium is suggested by the fact that initial $^{87}$Sr/$^{86}$Sr ratios for plagioclase and K-feldspar deviate significantly from the isochron intercept (Dini, 1997). Thus, the interpretation of the Rb-Sr age is constrained by the field evidence showing the Portoferraio porphyry as younger than the Capo Bianco aplite and older than the San Martino porphyry. For the purpose of this work we chose to correct isotopic ratios for the Portoferraio porphyry to 8 Ma. Correction to 8.5 Ma would lead to deviations in $^{87}$Sr/$^{86}$Sr of 2 to 3 times the analytical error, and to deviations in $^{143}$Nd/$^{144}$Nd within the error.

San Martino porphyry

The San Martino porphyry has monzogranitic compositions (Tables 5.1 and 5.2; Fig. 5.1), and is characterized by K-feldspar megacrysts and biotite with Fe#. 0.53 - 0.58 (Fig. 5.2). This unit occurs in western Elba as dykes cutting the Capo Bianco aplite and Portoferraio porphyry (Figs. 3.5 and 6.1), whereas in central Elba, it occurs as three main laccolith layers up to 400 m thick and as minor crosscutting dykes (Westerman et al., 2000).

The San Martino porphyry occurs as dykes or thick layers, with prominent megacrysts of sanidine (Fig. 6d, e) set in a very fine-grained groundmass. Locally, megacrysts are aligned subparallel to intrusive contacts, with orientations steep in dykes and sub-horizontal in sills. Close to some contacts the megacryst content is very low. In western Elba, only WNW-striking dykes occur, cutting the older intrusive units and the rocks of Complex IV; the largest of these dykes extends over 2.5
km with a thickness of 25 to 50 m (Figs. 3.5 and 6.1). In central Elba, dykes of San Martino porphyry cut the fabric of the sedimentary host as well as layers of the older intrusive units, while the main bodies are four parallel, gently westward dipping layers. They were emplaced above the layers of Portoferraio porphyry, concordant to the fabric of the folded and faulted strata of the host flysch. Bifurcation along the upper surface form branches, and at Monte San Martino, a large septum of flysch maintains a simple planar geometry over a map distance of 1.5 km. The maximum thicknesses of these igneous layers are between 100 and 700 m, tapering out towards both northern and southern end. Lengths measured in the N-S direction range between 2.4 and 8.3 km, with the thickest and largest layer exposed over more than 18 km². Miarolitic cavities are found in this unit, but are notably absent in the other units, suggesting that the San Martino porphyry had the shallowest emplacement level. As with the Portoferraio porphyry, visible contact metamorphic effects are essentially absent in the host rock.

A Rb-Sr wr-Pl-Bt isochron indicates an age of 7.2±0.1 Ma (Dini, 1997). However, significant Sr isotopic disequilibrium exists within K-feldspar, and between megacrysts and the whole rock. Nevertheless, this Rb-Sr date is quite close to a sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ isochron age of 7.44±0.08 Ma (Dini and Laurenzi, 1999). Isotopic dates are in agreement with field observations and indicate that the San Martino porphyry was emplaced after a significant period of quiescence.

**Magma Emplacement**

**Geometry of the laccolith complex**

The western Elba igneous complex is composed of several sheet-like intrusions with extensively known roofs and floors, and a larger pluton with the roof exposed locally, but no known floor. All the sheet-like bodies of Orano porphyry and a few sheets of Portoferraio and San Martino porphyries cut across the structures of the host rock, and thus are dykes. The main bodies of Capo Bianco aplite, Portoferraio porphyry, and San Martino porphyry all occur subparallel to the planar structures of the host metaophiolite, metasediments, and flysch.

Based on detailed mapping (Fig. 6.1), maximum thickness values of the nine most significant intrusive layers, along with thicknesses of the intervening and overlying host rock, were determined by measurement.
in cross section (Fig. 6.4; (Rocchi et al., 2002). Since layers either thicken or thin as they project below the surface, these values are invariably equal to or less than the true maxima. Thickness values of individual layers vary over an order of magnitude, from 50 to 700 m.

Diameter values for each layer were approximated from the maximum horizontal exposed length. Layers exposed in central Elba have N-S strikes with lengths between 2.4 and 10 km, while in western Elba they have NE-SW strikes and lengths between 1.6 and 9.3 km (Rocchi et al., 2002). This latter length has been determined after reconstructing the layer before it was deformed and crosscut by the Monte Capanne pluton: this reconstruction is supported by the occurrence of mylonitic foliation at the NE and SW contacts with the pluton. Lengths perpendicular to strike are assumed to be similar since (Corry, 1988) reports differences generally less than 2 times. The maximum diameter of individual layers varies nearly an order of magnitude, from 1.6 to 10 km (Rocchi et al., 2002). All nine of the main Elba laccolith layers have large aspect ratios (diameter/thickness), varying from 12 to 33.

Emplacement of the Elba laccoliths as layers that thin toward their edges added more than 2 km thickness to the upper crust above a common magmatic centre.

Therefore, each intrusive unit is interpreted as a multilayer laccolith. The overall geometry resulting for the Elba subvolcanic complex is one of a nested multilayer, multipulse Christmas-tree laccolith sequence.

The main bodies of Capo Bianco aplite, Portoferraio porphyry, and San Martino porphyry have laccolithic shapes (Rocchi et al., 2002): (i) contacts of intrusions share the strike and dip of host flysch bedding or tectonic surfaces in the nappe sequence, (ii) the sheets clearly taper out at their visible eastern ends, (iii) detailed mapping and cross sections show that the nine main layers have convex-upward roofs and flat or convex-upward floors. The layers of each unit are connected by small dykes, and major dykes below the overlying sheets are interpreted as feeders (Fig. 6.1). The overall geometry resulting for the Elba subvolcanic complex is one of a nested multilayer, multipulse Christmas-tree laccolith sequence.

Volumes were calculated by approximating the shape of each laccolith layer to a spherical cap with height equal to T and diameter equal to L. The resulting values vary over more than two orders of magnitude, from 0.1 to 24 km³ (Rocchi et al., 2002).
Construction of the multilayer Laccolith Complex

Figure 6.5 presents a panel of cartoons illustrating construction of the pluton-laccolith complex, culminating with a “snapshot” of the resulting dome structure. The oldest intrusion is the Capo Bianco aplite laccolith (about 8.0 Ma), with a lower level at a strong discontinuity in Complex IV, and an upper level at the Complex IV-V interface. The geometry and shape of the almost coeval Nasuto microgranite is not represented owing to its dismemberment by the closely following intrusion of the Portoferraio laccolith.

Fig. 6.5 – Schematic illustration of the three main emplacement episodes of the Western-Central Elba Laccolith Complex.
The latter consists of two thin horizons and one major level occurring inside Complex IV, along with one major horizon emplaced at the Complex IV-V interface. The successive intrusion is represented by the San Martino laccolith (about 7.4 Ma); feeder dykes for the San Martino laccolith are preserved in western Elba, while the main intrusions were exclusively in central Elba within Complex V.

Emplacement depths for individual layers within the Elba laccoliths can be approximated by measuring the overburden from cross sections (plus 800 m, based on an estimate of the present mean erosion rate for Italy of 0.1 mm/yr) (Branca and Voltaggio, 1993). The emplacement depths calculated in this way range from 1.9 to 3.7 km (Rocchi et al., 2002).

As for the physical controls on emplacement, it is worth noting that most of the laccolith layers at Elba were emplaced along strong crustal heterogeneities such as thrust surfaces between Complexes, secondary thrusts inside Complexes, and bedding in the flysch. Evidence constraining emplacement mechanisms of the laccoliths includes: (i) subvertical dykes of San Martino porphyry observed and mapped at the lowest levels of the complex (Fig. 6.1), with laccoliths of that unit observed at higher levels, and (ii) laccolith layers of Portofererraio porphyry exposed along with their E-W trending feeding dykes on the western side of the Golfo di Lacona; (iii) tapering branches observed to emanate from the top surfaces of laccolith layers (at Poggio Zuffale and M. San Martino), and (iv) main sheets tapering out at their visible (eastern) ends.

The general picture is, therefore, that of vertically rising magma that stops and spreads laterally. Such a switch from vertical to horizontal magma movement could be related to reaching the neutral buoyancy level, but in this case magma should not have the energy to lift the overburden during filling of laccoliths (Hogan et al., 1998). An additional problem with the role of neutral buoyancy is represented by the progressively shallower emplacement level over time of magmas with similar densities (at liquidus conditions, calculated densities vary by less than 0.02 g/cm³), with the most dense magma (Orano) going to the shallowest levels. It has to be assumed, therefore, that the rising magma had a residual driving pressure exceeding the vertical stress when it changed its direction of movement from vertical to horizontal. Therefore, the magma spreads laterally when encountering a subhorizontal strength anisotropy that behaved as a crustal magma trap (Hogan and Gilbert, 1995; Hogan et al., 1998). Indeed, at the trap,
magma driving pressure must be sufficiently greater than the lithostatic load to allow the magma to raise the roof rather than form a stock (Hogan et al., 1998) or simply flow laterally until escaping the trap. Thus, the occurrence of intrusive layers along surfaces of strength anisotropy points out that crustal magma traps played a prominent role in controlling the emplacement level of magma at Elba (Rocchi et al., 2002).

**Filling of the laccoliths**

The growth of an intrusion can occur in various modes depending on the magma supply rate, the depth of emplacement, the availability of crustal magma traps, and the mechanisms by which country rock makes room. In the case of magmas emplaced in the upper 3-4 km of the crust, lifting the overburden is a probable mechanism for making space for the intrusion, as demonstrated in some limiting cases where overburden uplift has actually been documented (Corry, 1988). How this space is filled, however, is a matter of debate addressed by (Rocchi et al., 2002). Relevant data commonly collected to solve this problem are represented by the internal structure related to magma flow and the final shape an intrusion acquired, assuming that the shape is the result of its growth history.

Investigation of the geologically or geophysically reconstructed shapes of worldwide laccoliths, led the proposal of an empirical power-law relationship linking width (L) and thickness (T) of these intrusive bodies has been proposed (McCaffrey and Petford, 1997). The relationship has the form \( L = kT^a \) (where \( k \) is a constant and \( a \) is the slope of linear data regression in a Log-Log plot), suggesting a scale-independent mechanism of growth. A similar scale-invariant distribution of tabular shapes has been documented for plutons (Cruden and McCaffrey, 2001; McCaffrey and Petford, 1997). The mechanism generally acknowledged as most likely for laccolith growth is a two-stage process (Cruden and McCaffrey, 2001; Johnson and Pollard, 1973; McCaffrey and Petford, 1997). First, the magma spreads laterally at the emplacement level with \( a < 1 \) (the stage of horizontal elongation) until an initial sill is formed having a very high aspect ratio (\( L \gg T \)), and a width nearly equal that of the future intrusion. Then the thin intrusion thickens by dominantly upward inflation and roof lifting along a growth line with \( a > 1 \). Experiments on vertical growth due to floor subsidence suggest that tabular plutons make room by floor depression and grow.
from an initial thin sill along a line with slope \(a \approx 6\) (Cruden and McCaffrey, 2001).

The dimensional parameters collected for the Elba intrusive layers allow testing of these hypotheses. A significant power-law correlation exists between thickness (T) and diameter (L): \(\log T\) vs. \(\log L\) plot (Fig. 6.6), this correlation results in a linear fit \(r^2 = 0.93\) with the equation \(T = (0.026 \pm 0.006)L^{(1.36\pm0.14)}\) (Rocchi et al., 2002).

![Graph showing correlation between thickness and length](image)

**Fig. 6.6 - Dimensional parameters for each individual sheet of the Western-Central Elba Laccolith Complex.**

To interpret this L-T distribution, it is important to note that (i) individual Elba laccolithic sheets are thought to be only a layer of multilayer laccoliths, and (ii) the dimensional parameters of all the layers plot in the horizontal elongation field even though the slope \((a > 1)\) of the regression line is typical of vertical self-affine inflation (i.e. intrusions have the same type of shape at different scales, but L and T do not retain the same proportions as the intrusions grow). These observations led to interpret the Elba data as the first record of the occurrence of a “frozen” vertical-inflation stage in laccolith growth (Rocchi et al., 2002). In this view, the slope \((a)\) of the vertical inflation growth is represented by the value 1.4 (Fig. 6.7), rather than 6 (Cruden and McCaffrey, 2001).
Significant insights into the processes leading to “freeze” the vertical inflation stage are given by the dimensional parameters for each Christmas-tree laccolith considered as a single amalgamated unit (i.e., a simple sill with volume equal to the sum and basal diameter equal to the average diameter of its containing sheets). The resulting parameters for the Portoferraio and San Martino laccoliths plot on the regression line for pluton dimensions (Fig. 6.8). These results indicate that in the case of coalescence of the individual sheets to form three single-layer intrusions, the latter would have dimensions typical of plutons, suggesting that the growth of laccoliths and plutons goes on through the amalgamation of smaller sheet-like bodies. Additionally, it is likely that the multilayer laccoliths at Elba failed to coalesce due to an abundance of lithologic and/or tectonic discontinuities that acted as magma traps (Rocchi et al., 2002).

The filling time of a pluton like that formed by the amalgamation of San Martino laccolith layers (~21 km$^3$) can be estimated from the dimensions of its main feeding dike (~1500 m long x 10–20 m thick, Fig. 6.9, between Marciana and the northern coast). A conservative
magma-ascent rate as low as $3 \times 10^{-3} \text{ m}\cdot\text{s}^{-1}$ would result in filling on a time scale of $<10^2 \text{ a}$ (Rocchi et al., 2002).

Fig. 6.8 – Dimensional parameters for each individual layer and also of the Christmas-tree laccoliths as amalgamated layers.

Fig. 6.9 – Geological map of the Marciana Marina area draped over a digital elevation model. View from the north. Note the east-west San Martino feeder dyke (yellow).
Summary

Favourable tectonic conditions at Elba Island have allowed a detailed study of a 5-km-thick crustal section including nine intrusive layers of late Miocene age that built up three multilayer Christmas-tree laccoliths. The dimensional parameters of the intrusive layers fit a power-law distribution indicating that, after a likely first stage of horizontal expansion, the layers underwent a second stage of dominantly vertical inflation. The abundantly available crustal magma traps, in many cases, halted the supply of magma which then filled another layer. Laccoliths from Elba can be envisaged as sheet-like intrusions that did not coalesce to form single laccoliths or plutons with dimensions typically observed elsewhere. This in turn suggests that laccoliths and plutons, at least in some cases, grow by amalgamation of smaller sheet-like bodies (Rocchi et al., 2002).
Field stops

Capo Bianco

Outcrops along the coast from Capo Bianco towards the west (Fig. 6.10) are mainly constituted by the Capo Bianco intrusive unit, a low-porphyritic (less than 10 vol% of quartz, K-feldspar, plagioclase and muscovite phenocrysts), peraluminous, alkali feldspar granite, that is strongly enriched in boron. The latter feature is highlighted by the widespread presence of black-blue tourmaline orbicules and spots (Fig. 6.10). In particular, number and dimensions of orbicules increase from the bottom toward the top of this tabular intrusion (Fig. 6.11). Furthermore, increasing of tourmaline content is coupled with the development of a strong late magmatic layering evidenced by the distribution of tourmaline spots and muscovite flakes.

Fig. 6.10 – (a) Detailed geological map of the Capo Bianco area. (b) Detailed cross section of the outcrop area of Capo Bianco laccolith. After Dini et al. (2007).
Fig. 6.11 – . (a) Capo Bianco promontory. (b) Outcrop view of the Capo Bianco aplite. (c) Enlarged view of the outcrop showing the distribution of tourmaline orbicules. (d) Close-up view of tourmaline orbicules. (e) Photomicrograph of the aplite texture, with a muscovite phenocryst. (f) Photomicrograph of a small tourmaline orbicule. After Dini et al. (2007).

All these late magmatic structures developed above the solidus (no sub-solidus deformation of crystals) and overlap the early porphyritic texture of the rocks. The later emplacement of Portoferreaio porphyry and the extensional tectonics tilted the original sub-horizontal attitude of this intrusion producing the present situation where the late magmatic layering dip toward north, north-east at various degrees.
The distribution of both tourmaline and late magmatic layering, has been attributed (Dini, 1997) to the internal differentiation of the magma as B-rich fluids started to exsolve or the melt entered the immiscibility field for boron-alkali-aluminosilicate melts.

**Acquaviva-Enfola (Tuesday 16, AM)**

The sequence of emplacement events that produced the Western-Central Elba Laccolith Complex can be fruitfully investigated along the coast between Acquaviva and Capo d’Enfola (Fig. 6.12). The field trip starts from Case Acquaviva walking down the slope till the beach where the contact between the Capo Bianco aplite sill (the oldest intrusive unit) and the underlying Portoferraio porphyry laccolith (the third intrusive unit) is exposed. The contact between the two intrusive units is outlined by a thin layer (few cm to 1 m) of a black-green serpentinitic rock. The NW dipping contact surface is exposed due to the erosion of the aplite unit and the soft ophiolite material. The top of the Portoferraio porphyry in the direct contact with serpentinites was contaminated by ophiolitic material changing its colour from whitish-greyish to dark green over a thickness of few centimetres. The morphology of the contact surface is similar to the surface of *pahoehoe* lava flows (Fig. 6.13).

![Fig. 6.12 – Geological map of the Acquaviva-Enfola area draped over a digital elevation model. View from the north.](image-url)
Fig. 6.13 – Pahoehoe-like contact surface of Portoferraio porphyry laccolith layer.

Sitting on top of this contact we have a good landscape view of the cliffs where the top of the Capo Bianco sill is exposed (Fig. 6.14). The upper part of the cliffs is made by Portoferraio porphyry, while a vertical dyke of the younger San Martino porphyry crosscuts both the previous intrusive units.

Fig. 6.14 – Landscape view of Capo Bianco and Portoferraio sills crosscut by San Martino porphyry dykes.
Continuing the trip along the shoreline we follow the Capo Bianco sill along strike for about 500 metres. In this zone Capo Bianco aplite is strongly layered with small tourmaline orbicules; the porphyryc texture (small phenocrysts of quartz and feldspars) is quite visible due to the marine erosion). The attitude of layering is strongly variable due to fractures and faults but generally deep northward. A close-up observation of the layering indicate that layering was produced by multiple injection and lateral flow of magma: folded layers, laminated and truncated layers, etc. Magmatic layering is frequently crosscut by fractures and brecciated structures cemented by blue-black tourmaline (Fig. 6.15). This is a later hydrothermal event probably induced by exsolution of fluids from aplite itself. Portoferraio porphyry crosscut both magmatic layering and hydrothermal tourmaline breccias.

![Fig. 6.15 – Capo Bianco aplite hydraulically brecciated and cemented by hydrothermal tourmaline.](image)

At the end of the beach, the occurrence of several fallen blocks of Portoferraio porphyry is indicative of its presence on top of the Capo Bianco sill. The texture and mineralogy of this intrusive unit is well preserved in these particularly fresh blocks.
Climbing few metres over a big boulder we can observe a vertical dyke of San Martino porphyry crosscutting the Capo Bianco aplite. San Martino porphyry is characterised by the presence of many K-feldspar megacrysts, and quartz, plagioclase, and biotite phenocrysts embedded in a very fine-grained groundmass. The borders of the dyke are quite depleted in megacrysts and phenocrysts. K-feldspar megacrysts are perfectly euhedral and can be easily separated due to the preferential phyllic hydrothermal alteration experienced by the groundmass (Fig. 6.16). Some mafic microgranular enclaves can be observed.

Looking northwestern the Punta Sansone is visible with its big rocks protruding into the sea. Punta Sansone is made up by a vertical dyke of Portoferraio porphyry, trending NNE-SSW, that crosscuts the Capo Bianco aplite and feeds the large sill exposed on top of the Capo Bianco sill (Fig. 6.14). After the Portoferraio dyke the original magmatic contact between Capo Bianco aplite and the flysch host is exposed. Here the contact is subvertical and the portion of aplite near it contains very few tourmaline orbicules. After few metres of flysch host we enter the main body of Portoferraio porphyry. It is interesting to note the occurrence of “fluidification” structures at the contact between pelitic flysch and porphyry.
At this point we walk up on the slope to reach the main road. We follow it to the bus. Along the road several layers of flysch completely embedded into the Portoferraio laccolith are exposed, as well as the southwest prosecution of the San Martino dyke previously observed along the coast.

**Boat trip – Marina di Campo Bay (Wednesday 17, AM)**

The cliffs bounding the Marina di Campo Bay represent the best exposures of the Western-Central Elba Laccolith Complex, showing the multilayer nature of the San Martino and Portoferraio laccoliths and the geometry of laccolith layers.

The boat trip start from Marina di Campo, observing first the lowest laccolith layers at Capo Fonza (Fig 6.17). The lowest layers near Capo di Fonza belong to the Portoferraio porphyry. The tips of the two layers are easily visible on the slope of the hill. The geometry of the sills is complex and there are layers and blocks of flysch completely embedded in porphyry. In this area five layers have been mapped but moving northward they join in two main layers in correspondence of their E-W feeding vertical dykes (not visible from the Bay). The whitish colour of the porphyry is due to the pervasive phyllic hydrothermal alteration. Host rocks suffered negligible thermal effects and localised silicification and quartz veining.

![Fig. 6.17 - View from the south of Capo Fonza and the WNW-ESE oriented cliff. The sequence of tapered Portoferraio Porphyry layers dip to the west following tectonic tilting.](image)

“Climbing” up in the sequence (the overall flysch bedding is dipping to the west) toward west, four layers belonging to the large San Martino laccolith become visible. They represent the southern, thinned branches
of a large multilayer laccolith extending from the northern coast to the southern coast of the island. These layers, separated by flysch septa, merge in the central part of the island forming a unique body 700 m thick. The lower layer of San Martino porphyry continues southward into the Bay as indicated by the small outcrop named “Scoglio della Triglia” (the rock of the red mullet). The upper layer exposed along the northern coast of the Bay has a probable continuation in the porphyries exposed along the southern border of the Bay, in the Punta Bardella-Capo Poro area. Moving southward across the Bay it is possible to see the Punta Bardella outcrop where the top of the intrusive layer dips to the west. The contact zone between the San Martino porphyry and the flysch can be followed along the N-S trending cliff towards Capo Poro. The contact is characterised by undulations (Fig. 6.18), and minor intrusive layers connected to the main body by small dykes. It is interesting to note that K-feldspar megacryst are partially iso-oriented with long axes horizontal and parallel to the coast.

Fig 6.18 – Undulated upper surface of a San Martino porphyry layer.

At Capo Poro the cliff turn towards west, and along the E-W exposed vertical wall (150 m high) the geometry of the porphyrytic intrusions
change completely. Here (Fig. 6.19), the intrusive layer is not continuous showing swift thickness variations that were interpreted in the past as resulting by later faulting. However, alternatively this geometry could be interpreted as “fingers” of magmas propagating southward from the main layer exposed along the N-S trending cliff.

Fig. 6.19 – Capo Poro outcrop of San Martino porphyry.

**Cala Bardella**

Half an hour walk from the harbour of Marina di Campo to Punta Bardella. Path across Cretaceous flysch from the village to the end of Galenzana beach. Along the shore, fist-size, black, scoriaceous, heavy slags can be found. These are residua from smelting furnaces of Etruscan-Roman times. From the beach view of several laccolith layers intruding the flysch on the opposite side of the bay of Marina di Campo. At the megascale the tabular intrusions are concordant with the flysch bedding. At the very tip of Punta Bardella, the contact between flysch and San Martino megacrystic porphyry is exposed (Fig. 6.20). No thermometamorphic effects can be detected in the host rock. The porphyry is rich in K-feldspar megacrysts up to 15 cm and quartz phenocrysts up to 2 cm. K-feldspar is a high-T sanidine, testifying the magmatic origin of this mineral phase coupled with quenching of the
enclosing melt. Scattered brown-altered mafic microgranular enclaves can be observed. The groundmass is yellow-brown due to incipient weathering.

Fig. 6.20 - Contact between San Martino porphyry and Cretaceous turbidite host.

**Chiessi (Wednesday 17, PM)**

The beautiful exposures of the contact aureole along the western margin of the Monte Capanne pluton are cross-cut by the Chiessi-Colle d’Orano road. The Ligurian units and the earlier intruded (at ~ 8 Ma) Portoferraio porphyry sills were coherently deformed and re-crystallized by the emplacement of the Monte Capanne pluton (6.9 Ma). The road cuts show the Ligurian sedimentary formations (shale and limestone) transformed in pelitic hornfels and marble, and the mylonitized porphyries; both rock types were locally pierced by apophyses of the granite intrusion. The metamorphic foliation overprints the contacts between Portoferraio porphyries and hornfelses, acting also as the axial plane foliation of isoclinal folds that probably generated by the enhancement of original contact irregularities. Similar deformed sills of Portoferraio porphyry crop-out also in the Marciana-Procchio area.
suggesting that the original width of these tabular bodies was more than 9 km (Rocchi et al., 2002). The overall porphyritic texture of the rock is maintained but biotite phenocrysts have been completely re-crystallized in flattened aggregates of tiny flakes oriented along the metamorphic foliation. Quartz phenocrysts suffered sub-grain effects, flattening and formation of pressure shadows. Also K-feldspar phenocrysts were locally re-crystallised with the formation of graphic aggregates. Plagioclase and rarely K-feldspar show a brittle behaviour: phenocrysts are broken, displaced and cemented by quartz fibres. The deformed Portoferraio porphyry, the hornfels-marble aureole and the granite pluton have been later cut by microgranite dykes, pegmatite veins and mafic Orano porphyry dykes.
7. The Western Elba Sheeted Pluton  
(Monte Capanne Pluton)

Introduction

The Monte Capanne pluton in western Elba is the largest of those exposed in the Tuscan Magmatic Province (Marinelli, 1959; Poli, 1992). It is roughly circular in plan (about 10 km in diameter), and is bordered along two thirds of its perimeter by contact metamorphosed rocks mainly after Complex IV protoliths (Fig. 3.5). Contacts with surrounding country rock are mostly intrusive in nature and dip away from the pluton. Host rocks, belonging to the ophiolitic-sedimentary tectonic Complex IV and exhibiting shear fabric acquired during the Apenninic compressive phase, were overprinted by thermal metamorphism and deformed by emplacement of the pluton itself (Daniel and Jolivet, 1995); this younger deformation crosscuts the older compressive shear fabric. Lithological varieties of protoliths in the aureole preserve the reactions (i) andalusite = sillimanite, (ii) talc + forsterite = anthophyllite + H$_2$O, and (iii) the breakdown of muscovite + quartz (Spear and Cheney, 1989; Thompson, 1974; Tracy and Frost, 1991). Although these reactions depend on the compositions of fluids and solid-solution phases, taken together they suggest peak contact metamorphic conditions with temperatures in excess of 600°C at a pressure of 0.1 - 0.2 GPa. Metasedimentary xenoliths up to 10 cm are common throughout the pluton (Gagnevin et al., 2004).

The Monte Capanne pluton has a monzogranite-granodiorite composition and consists of variable proportions of plagioclase, K-feldspar (commonly as large megacrysts), quartz and biotite (Bussy, 1990; Dini et al., 2002; Gagnevin et al., 2004; Marinelli, 1959). Accessory minerals are apatite, zircon, tourmaline, monazite, allanite and ilmenite (Dini et al., 2004). Minor secondary minerals are chlorite, calcite, titanite, epidote and sericite.

The whole intrusive mass is characterized by the widespread occurrence of mafic microgranular enclaves of strongly variable size and composition. The occurrence of these enclaves, along with amphibole clots replacing former pyroxene, and plagioclase plagioclase with oscillatory zoning (An content up to 45%) and calcic cores, as well as geochemical and isotopic data, are taken as evidence for a hybrid nature.
of the whole Monte Capanne pluton (Bussy, 1990; Dini et al., 2002; Gagnevin et al., 2004).

The pluton is cut by several leucosyenogranite dykes, occurring mainly close to the pluton's contact, within both the pluton and its thermometamorphic aureole. These dykes commonly have thicknesses up to 10’s of m.

Dates for emplacement of the Monte Capanne pluton are quite scattered (Ferrara and Tonarini, 1985; Juteau et al., 1984), with Rb-Sr and U-Pb dates between 5.8 and 7.0 Ma. None the less, dates obtained for the late, post-plutonic Orano porphyry dykes (6.83 - 6.87 Ma, e.g. Section 3.g.) suggest that the most likely emplacement age for the Monte Capanne pluton is close to the highest literature values, i.e. 6.8 - 7.0 Ma (Dini et al., 2002). Additionally, two samples from the San Piero facies display full Sr isotopic equilibrium and Rb-Sr wr-Pl-Kfs-Bt cooling ages of 6.88±0.1 and 6.75±0.07 (Innocenti et al., 1992). Therefore, initial isotopic ratios of samples from the Monte Capanne pluton have been corrected to 6.9 Ma.

The intrusive units

The pluton shows minor variability of petrographic features that led to the description of a normal and a peripheral facies on the basis of the variable content of K-feldspar megacrysts (Marinelli, 1959). The high megacryst content of the peripheral facies was interpreted as originating from a local, late-magmatic in-situ process, thus hampering any further speculations on the relationships between the pluton’s internal structure and its emplacement history. More recently, on the basis of field observations (mainly K-feldspar megacryst distribution), accessory mineral microtextures and additional minor geochemical features, the occurrence of three different intrusive facies within the pluton has been proposed (Dini et al., 2002; Dini et al., 2004; Westerman et al., 2003), stimulating the collection of further data bearing on the pluton’s emplacement history.

A quantitative field method relying on megacryst abundance has been used to map the pluton’s petrographic variability. First, minima of frequency in the abundance distribution have been identified that could be taken as indicative of the occurrence of different petrographic facies within the pluton. Then, a geopetrographic map of the pluton’s internal
variability has been constructed to show the limits between the detected intrusive facies (Figs. 7.3 and 7.4). Finally, this subdivision by facies is verified and validated with geochemical and mineralogical data in the next section.

**Definition of the intrusive facies**

Megacryst abundance parameters have been collected at 351 stations over the whole Monte Capanne pluton (about 7 stations/km²). A detailed study on the areas most enriched (close to Sant’Andrea village, Fig. 7.1) and depleted (close to San Piero village) in megacrysts (Fig. 7.2) indicate that the megacryst modal distribution (area %) shows strongly different average megacryst contents: in the San Piero zone the area % is 0.26±0.21 (1σ), corresponding to about 4 megacryst/m²; in the Sant’Andrea zone the area % is 3.37±0.56 (1σ), corresponding to about 57 megacryst/m². The measured stations are considered to belong to (i) a facies named San Piero when their megacryst area % is lower than 0.7 (i.e. the San Piero zone average +2σ), (ii) a facies named Sant’Andrea when their megacryst area % exceed 2.3% (i.e. the Sant’Andrea zone average −2σ). The modal interval between the facies defined above is attributed to a third intrusive facies here named San Francesco. Mineralogical and geochemical data validate this facies subdivision.

**Map of the intrusive facies**

All the data obtained from the 351 counting stations were used to produce a detailed contour map of megacryst distribution in the Monte Capanne pluton (Figs. 7.3 and 7.4), by the Kriging Gridding Method of the Surfer 8™ code. The megacryst content is widely variable across the pluton, ranging between 0.1 (~ 2 megacryst/m²) and 7 area% (~ 90 megacryst/m²). Using the above-defined modal thresholds between the three facies, the San Piero facies cover 35 % of the total area of the pluton, San Francesco facies 50 %, and Sant’Andrea facies 15 %.

The San Piero and Sant’Andrea facies are never in direct contact, being always separated to a variable extent by the San Francesco facies, with gradual transitions between adjacent facies. Although these transitions are required by the contouring algorithm itself, a gradual transition is effectively observed in the field. Additionally, the gradients of transitions are variable: once the effect of the station density and the intersection with morphology are taken into account, relationships to intrusive processes can be investigated.
Fig. 7.1 – Megacryst-rich Sant'Andrea facies at Sant'Andrea.

Fig. 7.2 – Megacryst-poor San Piero facies in a quarry at Seccheto.
Fig. 7.3 – Geological map of the facies of the Monte Capanne pluton draped over a digital elevation model. View from the south.
Fig. 7.4 – Geological map of the facies of the Monte Capanne pluton draped over a digital elevation model. View from the north.
The largest area of the Sant’Andrea facies occurs in the northern part of the pluton. It covers the coastal area between Capo Sant’Andrea and Colle d'Orano, and extends to the internal part of the pluton along the Monte Giove ridge. Several discontinuous outcrops are exposed on the morphologic highs from Monte Giove through the top of Monte Capanne to Pietra Murata. Additional minor outcrops occur along the western contact with the country rock. The easternmost exposure defines a N-S elongated area between the villages of Sant’Ilario and San Piero at the contact with the thermometamorphic aureole.

The San Piero facies crops out primarily in two separated zones. The widest zone forms a continuous NNE-SSW belt in the eastern part of the pluton, extending from Procchio through Cavoli to Fetovaia. The second zone extends between Pomonte and Troppolo Ridge. Additional minor outcrops are scattered in morphologic lows.

The distribution of the San Francesco facies between the San Piero and Sant’Andrea facies varies in extent, from zones a few tens of meters wide in eastern areas, to zones a couple kilometers wide in central and north-western parts of the pluton.

Overall, the Sant’Andrea facies crops out discontinuously in the external shell of the pluton, along the contact with the country rock as well as on morphologic highs. In contrast, the San Piero facies is typically found in morphological lows, where the deepest portions of the plutons have been exposed by erosion.

Understanding the internal structure and emplacement history of the Monte Capanne pluton requires determination of whether the observed internal petrographic differences (facies) were acquired at the emplacement level or resulted from the generation and emplacement of separate magma batches. To answer this question, integration of field data with microtextural, geochemical, isotopic and mineralogical data are presented in the next section.

**Textures of the intrusive facies**

The textural analysis of polished-stained slabs and thin sections for the Sant’Andrea and San Piero facies reveals important textural similarities and differences (Fig. 7.5). Both intrusive facies show features that can be described in a three-stage crystallization scheme (Bryon et al., 1994; Innocenti et al., 1997) where the first crystallizing phases form suspended in the melt, the second stage crystals build a three-
dimensional framework of touching crystals, and finally, the interstices are progressively filled with late crystallizing phases.

Sant’Andrea facies samples display a variable modal abundance (2.3-8 area %; with local extreme clustering) of large euhedral K-feldspar megacrysts (up to 20 cm in length). These megacrysts and their inclusions of biotite, plagioclase and minor quartz are representative of crystallization in suspension conditions, i.e. in the presence of a high proportion of melt. The second stage (framework) phases are represented by euhedral-subhedral quartz (up to 1.5 cm), variably zoned tabular plagioclase (2-8 mm) and subhedral biotite (> 3 mm). The interstices within the framework crystals are filled predominantly with K-feldspar, along with quartz and plagioclase that mostly forms overgrowths on existing subhedral-euhedral crystals. Tiny biotite crystals (<1 mm) are also present. Integration of field determination of K-feldspar megacryst abundance with quantitative modal analysis of rock slabs yields modal percentages of 43.5 for plagioclase, 22.3 for K-feldspar, 23.0 for quartz and 11.2 for biotite.

In the San Piero facies, the first stage of crystallization (suspension) is represented by the rounded euhedral cores (surrounded by the thick poikilitic rims grown during the final, interstitial crystallization stage) together with their biotite, plagioclase and minor quartz inclusions. The rounded cores (<0.7 area%) show the same type and pattern of mineral inclusions as the euhedral megacrysts of the Sant’Andrea facies. Euhedral-subhedral quartz, zoned plagioclase and biotite are clustered so as to locally form a three-dimensional framework and are on average smaller in size (1-6 mm) with respect to crystals of those minerals in the Sant'Andrea facies. The interstitial matrix assemblage of the San Piero facies is dominated by K-feldspar, along with minor quartz and sodic plagioclase. In particular, K-feldspar forms interstitial patches (0.1-2 mm), wholly poikilitic large plates, and poikilitic overgrowths surrounding resorbed euhedral megacrysts. San Piero modal percentages are 34.7 for plagioclase, 25.9 for K-feldspar, 24.4 for quartz and 15.0 for biotite.
Fig. 7.5 - Textural sketches of San Piero (A) and Sant'Andrea (B) facies obtained from stained slabs and thin sections line drawing. Noteworthy: 1) K-feldspar megacrysts in San Piero are partially resorbed and overgrown, in optical continuity, by poikilitic K-feldspar. 2) phenocrysts of plagioclase, quartz and biotite in the Sant'Andrea facies are significant larger than in San Piero. 3) Matrix K-feldspars in San Piero produce poikilitic, large subhedral plates while in Sant'Andrea they form smaller irregular interstitial domains.

**Geochemistry of the intrusive facies**

The Monte Capanne pluton is dominantly monzogranitic (SiO$_2$ 65-70 wt%, MgO 0.9-2.0 wt%; Table 5.2) and slightly peraluminous (ASI around 1.1). The three facies display limited but systematic compositional differences as shown on selected major and trace element variation diagrams. The Sant'Andrea facies is the most felsic, with relatively high SiO$_2$ and low CaO, MgO, Fe$_2$O$_3$ and Al$_2$O$_3$ contents. The San Francesco facies defines an intermediate compositional field partly overlapping the characteristics of both the Sant'Andrea and San Piero facies. In spite of these chemical differences and field evidence of
variable megacryst distribution, the K\textsubscript{2}O content does not display any systematic variation among the facies, ranging between 3.7 and 4.7 wt%.

Trace element distributions display limited yet significant variations among facies. Positive correlations exist between Sr and transition metals and Ba, with the San Piero facies being richest in all these elements. On the other hand, Rb, Pb, Th and Ta correlate negatively with Sr. In addition, some trace element ratios, particularly Nb/Ta, Rb/V and Sr/Y, lend further support to chemical distinction among the facies.

The Sr isotopic composition of samples from the Monte Capanne pluton is significantly variable ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7144-0.7156$; (Farina, 2007). Each facies shows a significant Sr isotope variability with a large overlap among them. Nevertheless, a systematic difference exists between the average $^{87}\text{Sr}/^{86}\text{Sr}$ ratios: 0.71506 ± 0.00032 (1σ) for Sant'Andrea facies, 0.71478 ± 0.00035 (1σ) for San Francesco facies, and 0.71474 ± 0.00023 (1σ) for San Piero facies. The Nd isotopic ratios show a restricted range ($^{143}\text{Nd}/^{144}\text{Nd} 0.512164-0.512200$), and no systematic differences between the facies have been recognized (the average value for the whole data set is 0.51218 ± 0.00001, 1σ).

**Biotite chemistry of the intrusive facies**

Overall, the biotites from the Monte Capanne pluton have annite compositions, with Fe\# = 0.48-0.57 and Al\textsuperscript{VI} = 0.10-0.55 apfu (Tischendorf et al., 2007). Throughout the pluton, two distinct textural types of biotite are recognised: 1) small, euhedral crystals included in megacrysts; 2) medium-grained, dominantly subhedral crystals occurring in the framework and as inclusions in the poikilitic portions of megacrysts. Compositional data for biotites from the matrix of the different intrusive facies define positive correlation trends between Al\textsubscript{tot}, F and Fe/(Fe+Mg) (Fe\#) coupled with a negative correlation between Si and Fe\# (Fig. 7.6). In particular, San Piero biotites have higher Si coupled with lower Al, F and Fe\# than do Sant'Andrea biotites, with San Francesco biotites showing intermediate compositions. Thus, the Fe\# is the most useful micro-chemical parameter to discriminate between the facies. San Piero biotites are characterized by the lowest Fe\# values (average 0.49±0.01) while Sant'Andrea biotites have the highest values (average 0.54±0.01). The biotites included in the megacrysts have compositions similar to those of the framework biotites of their respective host-rock.
Fig 7.6 – Biotite compositions for the facies of the Monte Capanne pluton.

**Magma Emplacement**

The Monte Capanne pluton intruded the base of the laccolith sequence. Its emplacement depth, however, is not simply the sum of laccoliths and their host rock thicknesses, since the Monte Capanne pluton punched through several layers of Portoferraio porphyry in western Elba, as demonstrated by the strong mylonitic foliation developed in the Portoferraio porphyry close to the contact with the pluton on both its NE
and SW sides. After accounting for erosion, the calculated emplacement depth for the Monte Capanne pluton is about 4.5 km, a value in agreement with peak conditions of contact metamorphism produced by the Monte Capanne intrusion (0.1-0.2 GPa; (Dini et al., 2002).

**Pre-emplacement K-feldspar–biotite growth**

The non-homogeneous distribution of megacrysts in the Monte Capanne pluton as mapped in Figures 7.3 and 7.4 allows for the definition of three intrusive facies, that also show systematic geochemical differences in major-trace element and isotopic compositions, as well as in biotite chemistry. These results raise questions about the origin, and emplacement histories, of the intrusive facies. Understanding the timing of megacryst formation and the origin of their distribution are thus pivotal in efforts to answer these questions.

The following observations support the likelihood of an early, pre-emplacement crystallization of the megacrysts: (1) the plagioclase and biotite euhedral crystals hosted in the megacrysts are distinctly smaller (0.5-2 mm) than those in the framework of the rock (euhedral-subhedral crystals > 3 mm); the relationship between megacryst faces and included plagioclases and biotites suggests that synneusis occurred, during which free movement of the large growing megacrysts in a melt-rich magma was allowed; (2) the sudden change in slope in the K-feldspar CSD at a crystal size (L) of about 20 mm indicates a shift from low to high nucleation/growth ratio, suggesting low undercooling during the megacryst crystallization stage before the shallow emplacement; (3) the higher Ba and Sr and slightly lower Rb contents of megacrysts, with respect to the matrix K-feldspar, are indicative of earlier megacryst crystallization; (4) the large Sr isotopic differences between the megacrysts (0.717-0.719) and both the bulk host granite and the matrix K-feldspar (~0.715) (Farina et al., 2005; Gagnevin, 2005; Innocenti et al., 1992) are straightforward evidence for an early megacryst origin.

It is worth noting that in the fast-cooling sub-volcanic units (porphyry dykes, sills and laccoliths) emplaced just prior to the Monte Capanne intrusive system, megacrysts are set in a very fine-grained matrix and are coherently oriented in the magmatic flow fields, further indicating pre-emplacement growth. These megacrysts, as well as those from the Monte Capanne pluton, are in Sr-isotope disequilibrium with the surrounding matrix (Westerman et al., 2003).
The integration of petrographic and chemical features of megacrysts with the textural characteristics of the mineral assemblages provides a crystallization history that provides evidence for distinct stages of formation, transfer and emplacement of magma (Fig. 7.7).

During the early pre-emplacement stage, tiny crystals of plagioclase, biotite and minor quartz nucleated and grew (source/storage stage). Afterwards, a change in physico-chemical conditions inhibited their crystallization, and K-feldspar dominated the liquidus assemblage. Low undercooling conditions then led to a K-feldspar growth-to-nucleation ratio high enough to form megacrysts. This megacrysts growth occurred in the presence of high melt percentage, allowing synneusis inclusion of the available tiny biotite-plagioclase-quartz crystals. Then, during the transfer stage, increased undercooling triggered nucleation of plagioclase, biotite and quartz phenocrysts. The crystallization sequence during these stages was similar for all three intrusive facies, except for the following significant differences: (i) the amount of megacrysts produced, and (ii) the resorption experienced by megacrysts in San Piero and in San Francesco facies, indicative of non-equilibrium P-T-X conditions. Finally, during the last stage of crystallization at the emplacement level, the magma became again saturated in K-feldspar crystallizing as part of the interstitial matrix (all three facies), as heavily poikilitic K-feldspar plates, and as overgrowths on resorbed megacrysts (only in San Piero and San Francesco facies).

Thus, the magmas generating the three different facies experienced similar physico-chemical conditions early in their history, although subsequently each of them evolved along a different path that allowed K-feldspar dissolution in the San Piero and San Francesco facies, and not in the Sant’Andrea facies. Differences in magmatic evolution were also maintained during the very late crystallization stages, when poikilitic overgrowths on K-feldspar megacrysts developed only in San Piero and San Francesco facies.
Fig. 7.7 - Crystallization sequence for Sant'Andrea and San Piero intrusive facies. Light grey - both intrusive facies; dark grey – San Piero; white – Sant'Andrea. Only rock-forming phases are considered. The crystallization sequence for San Francesco is very similar to that for San Piero but it is characterized by an higher megacrysts content. Abbreviations: Bt – biotite; Kfs – K-feldspar; Plg – plagioclase; Qtz – quartz.

Deep origin of magma batches

In order to understand the origin of these intrusive facies, the following alternative hypotheses must be considered and discussed on the basis of the whole data set: (1) the facies differentiated in-situ (i.e. at the emplacement level) from a single homogeneous batch of magma; (2) the two end-member facies represent two distinct batches of magma emplaced sequentially, while the San Francesco facies derived by in-situ mixing/mingling processes at their interface; (3) the three intrusive facies represent three sequentially emplaced, distinct batches of magma.

In the first hypothesis, a single batch of magma fractionated in-situ, giving rise to the observed differences in megacryst distribution and the other chemical parameters. This hypothesis contrasts with observations on trace element and isotopic composition of megacrysts compared with whole rocks and matrices. Even though the Ba and Sr content of Monte Capanne megacrysts is systematically higher than their host (Franzini and Leoni, 1974; Gagnevin, 2005), the Sant'Andrea megacryst-rich facies shows Ba and Sr contents lower than the megacryst-poor San Piero facies. Furthermore, the three facies show significant yet small
differences in their Sr isotopic compositions, ruling out the simple occurrence of closed-system fractionation processes. Particularly, Sant’Andrea is more radiogenic (average $^{87}\text{Sr}/^{86}\text{Sr}=0.71506$) than San Francesco (average 0.71478) and San Piero facies (average 0.71474). The higher $^{87}\text{Sr}/^{86}\text{Sr}$ of K-feldspar megacrysts with respect to their host (Farina et al., 2005; Gagnevin, 2005), coupled with the higher megacryst abundance in Sant’Andrea facies, could at a first glance be taken as responsible for the observed Sr isotope differences between the facies. However, simple mass balance modeling demonstrates that the observed isotopic differences between Sant’Andrea and San Piero facies could be accounted only for a megacrysts content for Sant'Andrea facies of about 15 vol% (i.e. about five times the observed one).

Following the second hypothesis, i.e San Francesco derived by in-situ mingling/mixing of San Piero and Sant’Andrea facies, the compositions of the earliest phases (megacrysts and their included biotites) should have a bimodal distribution reflecting the chemistry of these phases in the two parent magmas. However, the compositions of biotites included in megacrysts from the San Francesco facies have their own compositional field, separated from those of the other two facies (Fig. 7.6), pointing out that the earliest, pre-emplacement stages of crystallization occurred separately for the three facies. Additionally, the composition of framework biotites show compositional overlap with the corresponding early biotites (included in megacrysts) and are thus different in all three facies. This observation also points out that the whole crystallization process occurred separately for the three facies.

Thus only the remaining hypothesis, i.e. emplacement of three distinct pulses of magma, can actually account for all the compositional data collected. It is on this basis that the Monte Capanne pluton is inferred to be a composite intrusion fed by three distinct batches of magma that acquired their textural and geochemical features at depth. This conclusion has significant bearing on the reconstruction of the pluton’s internal structure and the emplacement mechanisms leading to that zoning.

Pluton’s internal structure

The internal architecture of the Monte Capanne pluton can be reconstructed on the basis of the geometric distribution of outcrops of the three facies, the surface morphology, and the projected “threshold”
surfaces separating them as derived from the facies contour map (Figs. 7.3 and 7.4).

The Sant’Andrea megacryst-rich facies crops out on morphologic highs and near contacts with the enclosing country rock, while exposures of the San Piero facies typically occur in morphologic lows. The San Francesco facies commonly crops out between the two other facies, with variable thickness. In the south-eastern area, the extreme gradient of megacryst content indicates the San Francesco facies has negligible thickness there, with the San Piero facies appearing in direct contact with the Sant’Andrea facies.

The threshold surfaces between the facies have subhorizontal attitudes at the highest elevations, close to the inferred pluton summit, while in the peripheral zones these surfaces show an outward dip, more or less parallel to the pluton-country rock contacts. These observations led to the interpretive cross section reported in Figure 7.8 (bottom panel), which presents an interpretation of the pluton’s internal structure. Attitude variations of the threshold surfaces depict the first-order geometry of the facies, yet potential short wavelength variations (undulations, apophyses) typical of intrusive contacts, are not resolvable.

Overall, the intrusion shows a shell structure with the facies constituting three sheets extending over the whole pluton. The Sant’Andrea facies represents the uppermost sheet, partly removed by erosion; the pluton’s current morphological surface mimics the original contact between the Sant’Andrea facies and the country rock, as supported by the occurrence of scattered caps of thermometamorphic host rock preserved over the granite, e.g. in the Monte Perone area. The San Francesco facies represents the underlying intermediate sheet. For the San Piero sheet, the upper boundary against the overlying San Francesco sheet is reconstructed by the contour map (Figs. 7.3 and 7.4), while the lower boundary is not constrained.

Assuming a roughly circular shape of the intrusion, the steep dip of the pluton/country rock contacts allows to infer that the present total diameter of the pluton is similar to its mapped diameter (about 10 km). The total thickness of the pluton has been estimated using magnetic data at about 2.5 km (Dini et al., 2008b). As for individual facies, the Sant’Andrea sheet shows a mean diameter of about 9.5 km and a thickness of 250 m, the San Francesco sheet is about 9 km in diameter and 650 m thick, and the San Piero facies has an exposed diameter of about 8 km. If no unexposed facies occur at depth, the thickness of the San Piero facies can be estimated at about 1.5 km. Noteworthy, the
dimensional parameter of San'Andrea and San Francesco facies are comparable with the largest sheets of the multi-layer laccolith complex emplaced in central and western Elba (Rocchi et al., 2002).

**Emplacement sequence and mechanisms**

The bulk of geologic-petrographic-geochemical data collected for the Monte Capanne pluton indicates that its internal sheet structure resulted from the emplacement of three distinct batches of magma that arrived sequentially as the Sant'Andrea, the San Francesco and the San Piero batches.

This emplacement scenario is supported by a variety of evidence and indications, including the fact that despite extensive outcrop exposure, no feeding structures to the uppermost Sant'Andrea sheet have been observed crosscutting lower sheets. The uppermost Sant’Andrea and San Francesco sheets display swift thickness variations up to local elision, that are interpreted to indicate lateral squeezing resulting from the latest intrusion of San Piero facies. Additionally, the sequential emplacement of increasingly more mafic facies (from Sant’Andrea to San Piero facies) follows the overall temporal trend of progressive increase of mantle contribution to the magmas as reported for the whole western-central Elba magmatism (Fig. 5.7; (Dini et al., 2002). In this scenario, the large amount of mafic microgranular enclaves in the Sant’Andrea facies suggests early extraction of Sant’Andrea magma from the source region before mixing processes effectively hybridized subsequent magmas as represented by the San Francesco and San Piero facies. Finally, it is worth noting that this downward stacking of magma batches is analogous to the growth of endogenous domes in volcanic environments (Barker, 2007; Nakada et al., 1995).

Thus, an emplacement model is depicted schematically in Figure 7.8. First, the Sant’Andrea magma batch reached the emplacement level and spread laterally to form a sill-like intrusion most likely thicker and shorter than presently observed. Then the San Francesco batch underplated the Sant’Andrea sheet, partially displacing it. Finally, emplacement of the most voluminous San Piero batch elevated and deformed the two older sheets to generate the observed geometry.

Overall, space for magma was created by uplift of the overburden in a process of downward building of a thick laccolithic body. The resulting large surface uplift led to gravitational instability, in turn triggering eastward translation along a detachment fault of the whole
overburden soon after pluton emplacement (Westerman et al., 2004). Moreover, minor contribution to space generation was from subsolidus deformation with development of mylonitic structures in the thermometamorphic country rock (Dini et al., 2002; Rossetti et al., 2007).

Fig. 7.8 – Schematic time-sequence model for progressive evolution of the Monte Capanne pluton. Emplacement of different sheets and progressive inflation. Abbreviations: h: contact metamorphic cap with horizontal attitude; v: contact metamorphic cap with vertical attitude.
Rapid pluton assembly

The lack of sharp contacts between the facies implies that the time lag between subsequent injections had to be short enough to have melt preserved in the contact zone between the magma batches. To constrain the order of magnitude of the maximum time lag required to preserve magmatic contacts between sequential magma batches, defined as “critical time” by (Saint-Blanquat et al., 2006), three-dimensional thermal modeling was performed using the HEAT code (Wohletz, 2003). The following conservative conditions have been assumed: magma temperature at emplacement = 800 °C, intrusion top 6 km deep, country rock T=150°C (corresponding to a geothermal gradient of 25°C/km), thermal conductivity = 3.0 W m⁻¹ K⁻¹ for the magma and 2.5 for the ophiolitic country rock. After the emplacement of the first magma batch (thickness = 250 m, width = 9.5 km), the critical time before the arrival of the second batch is about 0.5 ka (the first magma batch would cross its solidus at 650 °C after 550 a). Then, the critical time between the emplacement of San Francesco and San Piero batches, taking into account the thermal perturbation produced by the Sant’Andrea body (local temperature of 300 °C), is about 7 ka. These results imply fast magma supply rates, via a system of dikes connecting the source region to the emplacement level. In this frame, the time scale over which the Monte Capanne pluton formed is essentially determined by the cumulative critical times between successive magma pulses (Menand, 2008), i.e. in the order of tens of ka.

Geochronological implications

Such a rapid pluton assembly contrasts with magma production rates (thousand of ka; (Petford et al., 2000), prompting some considerations relative to isotope geochronological resolution of incrementally assembled plutons. On one hand, in systems built up in the relatively hot middle-lower crust, diffuse contacts can be generated even during pluton assembly that spans several millions of years, and thus isotopic dating can successfully contribute to unravel the succession of magma pulses (Coleman et al., 2004; Glazner et al., 2004). On the other hand, for systems assembled in the relatively cold shallow crust, such as the Monte Capanne system, simple thermal modeling indicates that diffuse contacts resulted from the incremental addition of magma on time scales smaller than current analytical errors in isotope chronology. In the case of Monte Capanne pluton, the history of rapid transfer-emplacement of
magma hampers the possibility of geochronologically resolving the emplacement of magma pulses. Nevertheless, for systems characterized by prolonged magma production, coupled with rapid transfer, dating of different minerals can provide information on different stages of the evolution of the magmatic system. For the Monte Capanne pluton, these considerations can explain the apparent discrepancy existing between the isotopic ages obtained using Rb-Sr and ⁴⁰Ar-³⁹Ar methods (micas and sanidine, 6.8-6.9 Ma: (Dini and Laurenzi, 1999; Innocenti et al., 1992; Maineri et al., 2003) and U-Pb systematics (zircon, 7.1-8.0 Ma: (Daly et al., 2007). Indeed, taking into account the high zircon saturation temperature for acidic (peraluminous) melts (Watson and Harrison, 1983), U-Pb zircon ages should be indicative of the very first stages of crystallization, likely close to the magma generation event. These ages could not represent the emplacement/solidification age, just as U-Pb zircon ages of silicic volcanic rocks do not represent eruption ages (Simon et al., 2008). On the other hand, isotopic dating of late-crystallizing minerals can give a better estimate of emplacement age. Thus we propose that magmas feeding the Monte Capanne pluton were generated in the deep crust starting at 8 Ma, and emplaced in a rapid three-batch sequence at ca. 6.9 Ma.

Summary

The Monte Capanne pluton shows variability of petrographic and chemical parameters throughout, but integrated textural-geochemical analysis has revealed otherwise cryptic internal diffuse contacts between intrusive facies.

The Monte Capanne pluton was assembled incrementally by downward stacking of three slightly different magma batches, building a sheeted pluton in the intermediate-shallow crust. This result fills the gap of knowledge about the depth of occurrence of sheeted plutons, previously documented only for intermediate-deep crustal plutonic settings and very shallow subvolcanic levels.

The magma batches formed at depth, acquiring unique geochemical features that were preserved after ascent and emplacement.

The three batches were emplaced in a short time sequence, thus inhibiting the development of sharp contacts between sequential magma
batches and hampering geochronological efforts to unravel age differences between internal facies.

The Monte Capanne case study indicates that integration of textural and geochemical data can shed light on the internal structure of plutons, and that similar analyses may reveal cryptic contacts in many other plutons.
Field stops

*Cava Bontempelli*

In this abandoned quarry, the San Piero Facies (Fig. 7.2) of the Monte Capanne pluton is characterized by a very low content of clearly defined K-feldspar megacrysts, and by the unique occurrence of “ghost” K-feldspar megacrysts, i.e. strongly poikilitic cm-sized crystals only visible when the sun shines on properly oriented cleavages (Fig. 7.9).

![Fig. 7.9 - K-feldspar poikilitic plate, in San Piero facies: “ghost” type (left), visible as reflecting cleavages (right).](image)

The following observations support the likelihood of an early, pre-emplacement crystallization of the megacrysts: (1) the plagioclase and biotite euhedral crystals hosted in the megacrysts are distinctly smaller (0.5-2 mm) than those in the framework of the rock (euhedral-subhedral crystals > 3 mm); the relationship between megacryst faces and included plagioclases and biotites suggests that synneusis occurred, during which free movement of the large growing megacrysts in a melt-rich magma was allowed; (2) the sudden change in slope in the K-feldspar CSD at a crystal size (L) of about 20 mm indicates a shift from low to high nucleation/growth ratio, suggesting low undercooling during the megacryst crystallization stage before the shallow emplacement; (3) the higher Ba and Sr and slightly lower Rb contents of megacrysts, with respect to the matrix K-feldspar (e.g. Franzini et al., 1973), are indicative of earlier megacryst crystallization; (4) the large Sr isotopic differences between the megacrysts (0.717-0.719) and both the bulk host granite and the matrix K-feldspar (~0.715) (Farina et al., 2005; Gagnevin, 2005; Innocenti et al., 1992) are straightforward evidence for an early megacryst origin.
It is worth noting that in the fast-cooling sub-volcanic units (porphyry dykes, sills and laccoliths) emplaced just prior to the Monte Capanne intrusive system, megacrysts are set in a very fine-grained matrix and are coherently oriented in the magmatic flow fields, further indicating pre-emplacement growth. These megacrysts, as well as those from the Monte Capanne pluton, are in Sr-isotope disequilibrium with the surrounding matrix (Westerman et al., 2003).

The integration of petrographic and chemical features of megacrysts with the textural characteristics of the mineral assemblages provides a crystallization history that provides evidence for distinct stages of formation, transfer and emplacement of magma (Fig. 7.7). The following discussion focuses on the end-member facies of Sant’Andrea and San Piero, with San Francesco facies displaying textural features similar to those of San Piero facies.

During the early pre-emplacement stage, tiny crystals of plagioclase, biotite and minor quartz nucleated and grew (source/storage stage). Afterwards, a change in physicochemical conditions inhibited their crystallization, and let K-feldspar dominate the liquidus assemblage. Low undercooling conditions then led to a K-feldspar growth-to-nucleation ratio high enough to form large megacrysts. This growth of large megacrysts occurred in the presence of a high melt percentage, allowing synneusis inclusion of the available tiny biotite-plagioclase-quartz crystals. Then during the transfer stage, increased undercooling triggered nucleation of plagioclase, biotite and quartz phenocrysts. The crystallization sequence during these stages was similar for all three intrusive facies, except for the following significant differences: i) the amount of megacrysts produced, and ii) the significant resorption experienced by megacrysts in San Piero and in San Francesco facies, indicative of non-equilibrium P-T-X conditions. Finally, during the last stage of crystallization at the emplacement level, the magma again became saturated in K-feldspar with crystallization as part of the interstitial matrix (all three facies), as wholly poikilitic K-feldspar plates, and as overgrowths on resorbed megacrysts (only in San Piero and San Francesco facies).

Thus, the magmas generating the three different facies experienced similar physicochemical conditions early in their history, but each subsequently evolved along a different path that allowed K-feldspar dissolution in the San Piero and San Francesco facies, but not in the Sant’Andrea facies. Differences in magmatic evolution were also maintained during the very late crystallization stages with the formation
of poikilitic overgrowths on K-feldspar megacrysts only in San Piero and San Francesco facies.

**Pomonte**

From the church of Pomonte towards Monte Capanne, fifteen minutes walk through the village and along the stream to the lowermost of three small abandoned quarries. The granite represents the San Francesco facies, with moderate content of K-feldspar megacryst. The main smooth surface of the granite correspond to the main jointing of the granite, and was exploited for quarrying blocks. On this surface, mafic microgranular enclaves can be observed. Some are dark with included K-feldspar megacrysts. Some enclaves have size in excess of 1 m and looks almost like ghost enclaves, with petrographic and textural features almost indistinguishable from the main granite. Here and, more prominently, in some quarries in Chiessi, enclaves and granite can be distinguished only if the contacts is exposed. As a whole, a wide range of hybridism degree can be observed in this site.

On the NW wall, fibrous clusters of cummingtonite with biotite corona can be observed. These are likely pseudomorphs after euhedral mafic crystals, most probably orthopyroxene, that has been rarely found as relic in mafic enclaves from a different site in the Monte Capanne pluton.

**Sant’Andrea (Wednesday 17, PM)**

From Zanca, twenty minutes walk along the road down to the shore of the small Sant’Andrea inlet. From the beach, walk to the west towards Capo Sant’Andrea. The granite facies cropping out here is called Sant’Andrea facies, and is characterized by a high content of K-feldspar megacrysts (Fig. 7.1) and quartz phenocrysts, set in a coarse grained matrix. All along the path along the shore cliff, abundant mafic microgranular enclaves can be observed, with size up to several metres (Fig. 7.10). The shape is sometimes ellipsoidal, sometimes irregular, although always rounded, with sharp, lobate borders. The enclaves shows embayments and trails of “granitic” material (mainly medium to coarse grained quartz and plagioclase crystals) connecting K-feldspar megacrysts and the main granite (Fig. 7.11). K-feldspar megacrysts can be observed straddling the enclave-granite contact. “Double” enclaves can be observed, with a fine grained enclave hosted within a medium grained enclave.
Fig. 7.10 – Mafic microgranular enclaves in Sant’Andrea facies of the Monte Capanne granite. Note some “double” enclaves.

Fig. 7.11 – Sequential stages of entrapment of a K-Keldspar megacryst from the granite magma into the mafic enclave magma.

At the end of the path, on the rather flat surface of Capo Sant’Andrea, a variety of K-feldspar megacryst shape can be observed. The granite is crosscut by a 15 cm-thick dyke of the dark Orano...
porphyry, that hosts megacrysts grabbed from the Monte Capanne crystal mush. The contact is always sharp, but show dramatic variations of the geometry in a few metres apart, from straight to sinuous.

**Punta del Cotoncello (Wednesday 17, PM)**

At Punta del Cotoncello a dyke of leucocratic facies crosscut the main Sant’Andrea facies (Fig. 7.12, left). The Cotoncello dyke has fine-grained groundmass and lower pheno-megacryst content with respect to Sant’Andrea granite facies. The Cotoncello dyke is only occurrence in Elba of a high $^{87}\text{Sr}/^{86}\text{Sr}$ crust-derived magma (see main text). The contact shows a complex geometry, owing to the interplay of brittle and plastic relative behaviour of the intruding dyke and the host Sant’Andrea crystal mush. Clusters and trails of K-feldspar megacrysts, along with megacryst-laden schlieren are common in the area (Fig. 7.12, right). Size and abundance of mafic microgranular enclaves are lower than observed in Sant’Andrea outcrop.

![Fig. 7.12 – Left: contact between megacryst-rich Sant’Andrea facies of the Monte Capanne granite and fine-grained Cotoncello dyke. Right: magmatic biotite schlieren outlining magmatic flow fields, with iso-oriented K-feldspar megacrysts.](image)
8. The Late-Plutonic Mafic Dyke Swarm
(Orano porphyry Dykes)

Introduction

The Orano dyke swarm (ODS) intruded all of the other intrusive units of the Western-Central Elba Laccolith Complex and the Monte Capanne Pluton. The ODS was emplaced before complete consolidation of the Monte Capanne plutonic system at 6.85 Ma, representing the closing event of igneous activity in western Elba (Dini et al., 2002).

In western Elba, the ODS occurs only northwest of a geomorphic lineament named the Pomonte-Procchio line (PP line, Fig. 8.1) (Dini et al., 2008b). This is a prominent topographic feature seen in satellite imagery as an alignment of valleys and ridges extending from Pomonte through Monte Perone to the Procchio area (Fig. 8.2). The PP line splits the Monte Capanne pluton (including its contact aureole) into two regions characterized by different morphological textures: (i) the northwestern part displays a deeply incised drainage pattern, with valley orientations clustered in WNW and NE directions, (ii) the south-eastern part shows a smoother morphology with an overall NW-trending, poorly incised drainage system.

In central Elba, Orano dykes are found only in the northern half of the region (Fig. 8.3). Indeed, Orano dykes here belong to the eastward-displaced uppermost part of the swarm that was rotated approximately 15-20° clockwise during the eastward translation from western to central Elba (Fig. 4.1).

The intrusive unit

Orano porphyry

The Orano dyke swarm (ODS) includes more than 200 darkly-coloured dykes (Fig. 8.4) with dominant quartz monzodioritic to granodioritic compositions (Fig. 8.5). Orano dykes are porphyritic with plagioclase, biotite, clinopyroxene and amphibole phenocrysts and quartz and K-feldspar xenocrysts set in a very fine-grained groundmass of plagioclase, K-feldspar and phlogopite.
Figure 8.1. (a) Interpretive geological map of Orano dykes (black) in western Elba. The south-eastern limit of the area affected by dyke intrusion is indicated by the bold dashed line. Shades of grey within the Monte Capanne pluton represent the
pluton’s internal facies: white: K-feldspar megacryst-rich S. Andrea facies; dark grey: K-feldspar megacryst-poor S. Piero facies; grey: S. Francesco facies, with intermediate petrographic features. Dotted lines bound areas dominated by dykes with ENE trends (major system) and NNE or NW trends (minor system). Oblique ruling represents the pluton’s contact aureole; (b) strike frequency distribution and rose diagram of western Elba Orano dykes; (c) rose diagrams of strike for zones dominated by the major and minor systems. The diagram on the right has a radius scaled 4x with respect to the one on the left to emphasize the strike distribution within the minor system.

Fig. 8.2 – Geological map of the the Orano dyke swarm and the Pomonte–Procchio lineament draped over a digital elevation model. View from ESE.
Figure 8-3 - (a) Interpretive geological map of central Elba showing the distribution of Orano dykes (black, modified after Dini et al. 2004). Oblique ruling: Monte Capanne pluton country rock; grey: central Elba multilayer laccolith complex; horizontal ruling: laccolith country rock (Complex V: Cretaceous and Paleocene flysch); oblique dashed ruling: Complex IV (Jurassic ophiolite sequence and its ocean floor sedimentary cover). (b) Strike frequency distribution and rose diagram of central Elba Orano dykes.
Figure 8.4 - a) Typical Orano dyke within Monte Capanne monzogranite; hammer for scale. Chiessi village and Colle San Bartolomeo in the background. b) Orano dyke with horn-and-bridge (side step) geometry at Colle d’Orano type locality. Small sledge for scale. The dyke also contains exotic K-feldspar megacrysts and granite blocks; c) example of brittle emplacement style of Orano dyke in the Monte Capanne monzogranite, near Sant’Andrea. Lens cap to the right of the circle for scale. Note within the circle a K-feldspar megacryst of the granite cut across by the dyke; d) example of plastic emplacement style of Orano dyke in the Monte Capanne monzogranite, near Pomonte; hammer for scale; e) example of Orano dyke crosscutting an aplite dyke within the Monte Capanne monzogranite, near Pomonte; hammer for scale; f) examples of sudden orientation changes of Orano dykes, along with closely occurring plastic and brittle emplacement styles. Locality near Chiessi (see also Fig. 8.12 for location); hammer for scale.
Some dykes, generally among the thickest, are zoned. Outer border zones, typically a few 10’s of centimetres thick, are distinguished from inner zones by (i) finer-grained groundmass, (ii) lower content of K-feldspar and quartz xenocrysts, and (iii) higher ferromagnesian mineral concentrations. The inner portions of zoned dykes have monzogranitic compositions, and their xenocrysts are petrographically very similar to minerals of the Monte Capanne pluton. The borders of zoned dykes are petrographically comparable to the unzoned dykes and have granodiorite to quartz-monzodiorite compositions (Fig. 4). These unzoned Orano dykes show evidence of early magma mingling including (i) three populations of plagioclase phenocrysts having strong differences in composition and texture, (ii) groundmass biotite attaining lower Fe no. with respect to phenocrysts (Fig. 8.5), (iii) rounded and embayed quartz and Kfs xenocrysts, and (iv) coexisting quartz and olivine plus clinopyroxene phenocrysts (generally replaced by tremolite-actinolite, Cr-clinochlore and Mg-phyllosilicates).

Fig. 8.5 – (a) Total Alkali vs. Silica classification diagram (Le Maitre 2002) showing the composition of ODS compared to the other Miocene igneous units of Elba. Worthy of note is the compositional overlap of Orano dykes from western and central Elba; (b) enlarged view with tie-lines connecting inner (small symbols) and outer parts (large symbols) of zoned dykes.
**Structural features of Orano dykes**

In western Elba, over 80,000 m total cumulate length for Orano dykes are seen NW of the PP line on the interpretive map (Fig. 8.1a), occurring primarily within the Monte Capanne pluton and to a lesser extent in the surrounding contact aureole. Orientations of Orano dykes in western Elba are shown in Figure 8.1b as a histogram and as a rose diagram, both constructed from the weighted data set described above. A primary system of dykes made of a set accounting for 67 of the 80 km cumulative length, shows a normal distribution with average orientation N75E, and is referred to hereafter as the ENE major system. Additionally, a minor system of dykes occurs, consisting of two sets with average orientations N38W (hereafter referred to as NW) and N22E (hereafter NNE). The areal distribution of the dykes belonging to the different systems in not uniform: Figure 8.1a shows how dykes with different orientations are concentrated in distinct ENE oriented “belts”. Figure 8.1c illustrates separately the strike distribution of dykes in the zones dominated by the major and minor systems, respectively, in order to better visualize the patterns of the minor systems.

Orano dykes in central Elba belong to the eastward-displaced uppermost part of the swarm, and they crop out with 9,300 m of cumulate length (Fig. 8.3). Patterns of orientation and relative frequency are similar to those in western Elba (Fig. 8.1) even though the orientation pattern in central Elba is rotated approximately 15-20° clockwise. This rotation likely took place during the eastward translation of the uppermost part of the ODS from western to central Elba. As a consequence, only the western Elba dykes are found in their pristine emplacement location and have their original attitudes preserved.

The entire ODS is steeply dipping with local dip variations between 90° and 40°, and rare sub-horizontal small dykes. Orano dykes in western Elba are numerous (157 dykes on the interpreted map), with thicknesses ranging from a few cm to 50 m, and lengths ranging from metres to about 7 km. Most dykes have sharp and planar contacts that commonly change orientation abruptly (Figs 8.4). Thicknesses along strike are generally maintained, independent of abrupt variations in attitude, indicating that the direction of opening was perpendicular to dyke walls. This direction of opening is further supported by observations where Orano dykes cut a variety of older lithologic contacts, such as between the pluton and country rock, older aplite dykes, or microgranular magmatic enclaves (Fig. 8.4e). Given the
constraints that the major system is overwhelmingly dominant, that dyke
dips are steep, and that opening is perpendicular to dyke walls, the
overall direction of opening was horizontal in the NNW direction.

Dykes in both the major and minor systems show the same
rheological behaviour with respect to their host: although they generally
have sharp and continuous contacts, some dykes locally display flow
"re-adjustment" of the dyke walls. Embayments and crenulated contacts
are common and testify that the granite was not completely crystallized
during dyke intrusion (Fig. 8.4d). Additionally, individual Orano dykes
have been traced continuously through significant changes in orientation
without petrographic variation (Fig. 8.4). These dykes also show
additional evidence of complex geometry, such as sudden side-steps
(Fig. 8.4c,f), horn structures and bridges (Fig. 8.4b), and broken bridges.
Finally, no systematic geochemical differences exist for Orano dykes of
different orientations (Dini et al., 2008b). Taken together, these lines of
evidence indicate that the major and minor systems constitute a single
dyking episode.

Fig. 8.6 – Close-up of the interpretive geological map of the Orano dyke
swarm draped over a digital elevation model. View from the east.
Magma Emplacement

Extension

These attitude and geometry data for the ODS have then been used to derive information about the distribution and degree of extension needed to accommodate dyke emplacement. With this aim, starting from the interpretive geological mapa of the Orano dykes in western Elba (Figs. 8.1 and 8.6), a map of “percent extension” was created (Fig. 8.7), by contouring the extension percentages calculated along the N15W-S15E direction. This choice effectively documents the extension produced by the ENE major dyke system, while the extension due to the minor systems is not accounted for. Nevertheless, extension determined along directions orthogonal to the strikes of minor systems is always below 2%, thus not affecting the results described in the following paragraph.

Fig. 8.7 – Contour map of percent extension (indicated by the numbers on the contour lines) as determined in the N15W-S15E direction for Orano dykes in western Elba (Dini et al., 2008b). Oblique ruling: contact aureole. A-A’ indicates the location of magnetic cross-section reported in Fig. 8.8.
The pattern of extension reveals three maxima (up to 15%) located in a restricted area in the central part of the dyke swarm. This area is characterized by a positive magnetic anomaly, modelled as a sill-like mass of moderate magnetic susceptibility (Fig. 8.8) and thought to represent the immediate source region for the Orano dyke magma (Dini et al., 2008b). These maxima have shapes elongated parallel to the strike of the major system. It is worth noting that extension values abruptly decrease starting from the maxima to the peripheral zone of dyking. Thus, the amount of extension (with respect to the ENE strike of the major dyke system) decreases both transversally and longitudinally, indicating that most of the magma input was focused in a limited area.

Fig. 8.8 - Magnetic model on a NW-SE section (see Fig. 8.7 for location) (Dini et al., 2008b). Magnetic susceptibilities used in the model were measured in the field and are reported in parentheses for each rock type in $10^{-3}$ SI units.

The south-eastern margin of the ODS is marked by the NE-trending PP line, a geomorphic lineament that had structural significance prior to ODS emplacement. Indeed, the slightly older leucocratic dykes have different orientations NW and SE of the PP line (Fig. 8.9a). Orientations of leucocratic dykes in the north-western part of the pluton correlate
strongly with pluton radii (Fig. 8.9b), suggesting dykes exploited radial cooling fractures. Leucocratic dykes SE of the PP line trend NNE-SSW, perpendicular to magnetic lineations in the pluton and to mineral lineations in the surrounding country rocks (Bouillin et al., 1993).

Fig. 8.9- (a) Orientation of aplite dykes (Boccaletti and Papini, 1989; Dini et al., 2008b; Pezzotta, 2000) and magnetic lineations (Bouillin et al., 1993); (b) frequency distribution in NW Monte Capanne zone of aplite dyke orientations (angle between dyke strike and pluton radius through dyke centre; (c) stereogram of poles of representative Orano and aplite dykes from NW and SE zones.
The Western Elba Transfer Zone

The ODS emplacement was not controlled either by a local, cooling-related stress field or by the overall regional late Miocene E-W extension (Pascucci et al., 1999). Indeed, the contrast in orientations of the radial leucocratic dyke system and the slightly younger ODS in the NW portion of the pluton and its surroundings, indicates that the local stress fields ruling aplite dykes emplacement shifted to a different stress field. This latter stress field is not compatible with the overall E-W extension in the northern Tyrrhenian Sea since the dykes are oriented nearly parallel to that direction.

The stress field controlling ODS emplacement can be constrained by the structural patterns of the dykes (Fig. 8.1). The major system of Orano dykes shows a pattern of linked offset fractures, best seen in the largest dykes weaving through the heart of the pluton (Fig. 8.1) that look like infill of connected side-stepping en échelon fractures. This pattern is characteristic of primary fractures (R1) of a dextral Riedel shear system trending NE-SW. A key zone for the understanding of such a geometry is the south-western region of the swarm where the segmented, ENE-trending dyke of San Bartolomeo is almost continuous over a length of about 3 km. The two main segments are characterised by eastern terminations striking WNW, parallel to the small dykes intruding the host rock to the south of the major dyke. Additional evidence for dextral shear in that region is represented by the prominent ENE-trending Chiessi-Capanne-Perone dykes in combination with the nearby NW-oriented dyke set (Fig. 7).

Mechanisms controlling emplacement of the ODS could be related to differential rates of horizontal movement and an irregular distribution of stretching in the brittle portion of the crust on a regional scale. This requires strain accommodation along transfer faults that typically develop at relatively high differential extension rates (Acocella et al., 2005), perhaps as a result of differential slab retreat (Scrocca, 2006). The ODS possibly formed in a narrow NE-trending transfer fault zone, the western Elba transfer zone (WETZ), with the PP line marking the SE border of the zone. In this scenario, the major dyke system is aligned in correspondence with R1 shear fractures of predominantly dextral strain. Such transfer fault systems in back-arc regions are characteristically complex (Dewey et al., 1998) such that variations in the rates of translation of adjacent blocks within the transfer fault zone produce sectors with opposing senses of strain, even if the overall shear couple
remains constant (Corti et al., 2003). Within the ODS, coexisting opposite senses of shear are pointed out by the occurrence of zones dominated by one or the other set of minor dykes. Where dextral shear dominated, the NW-trending set of minor dykes formed along R2 fractures (e.g. the minor system to the north of Monte Capanne; Fig. 8.1); in zones dominated by sinistral shear, NNE-trending R2 fractures were exploited to make the second dyke set (e.g. the minor system in the north-western area of the ODS).

Overall, thickness and spacing of Orano dykes in the WETZ are not homogeneously distributed: the extension values related to dyke emplacement depict three ENE-elongated ellipses for highest values (3-15%) enclosed in a larger elliptical area with the same orientation, bounded by 0.5% extension value (Fig. 8.7). Thus, the distribution of extension values indicates that significant magma volume was emplaced in a narrow, restricted zone within the overall outcrop area of the ODS. This suggests that, even though the ODS structural pattern was controlled by the WETZ, the restricted zone of rising melts appears to be linked to a local discrete feeding core. This inference is supported by the positive magnetic anomaly observed for this area and interpreted as evidence for a susceptible, mafic igneous body just below the north-western part of the Monte Capanne pluton.

Restricted geographical distribution is characteristic not only of the ODS, but also of the entire episode of western Elba magmatism that included intrusion of seven lithologically distinct units comprising a total estimated volume of about 200 km$^3$ (Rocchi et al., 2002). The igneous activity evolved over 1.5 Ma from crustal products to hybrid products involving multiple crustal- and mantle-derived sources (Dini et al., 2002; Westerman et al., 2003). In all likelihood, the ODS stress field was active during the previous igneous episodes, as indicated by the 7.4 Ma San Martino laccolith for which E-W trending feeder dykes are preserved in western Elba (Westerman et al., 2004). It is impressive that such numerous and petrogenetically distinct magma batches were emplaced at the same narrowly constrained centre, suggesting the persistence of a discrete system for magma feeding-ascent that was intermittently activated between c. 8 Ma and 6.85 Ma. Noteworthy, the north-western zone of the Monte Capanne pluton has also been indicated as the feeding zone for that entire pluton on the basis of AMS data (Bouillin et al., 1993).
The northern Tyrrhenian transfer systems

The WETZ is parallel to several subvertical transverse structures oriented at high angles to the Oligocene-Aquitanian compressional (thrusts, fronts) and post-Burdigalian extensional (basins, normal faults) features of the Apennine orogenic belt (Fig. 8.10). These ENE- to NE-trending structures have long been recognized on the basis of geological studies (Fazzini and Gelmini, 1982; Ghelardoni, 1965; Signorini, 1935) and geophysical data (Bartole, 1995; Cassano et al., 2001), and have been interpreted as transfer fault zones (Bartole, 1995). Furthermore, large-scale transfer zone activity has been recently inferred also for the central sector of the Apennine chain on the basis of detailed seismological reconstruction of the thrust front (Scrocca, 2006).

In the Tyrrhenian-Tuscan area, transverse structures bound or offset N- to NNW-oriented geological-geophysical features such as sedimentary basins (Acocella and Rossetti, 2002; Carmignani et al., 1994; Liotta, 1991; Pascucci et al., 1999) and magnetic anomalies (Cassano et al., 2001; Chiappini et al., 2000). The NE-trending lineaments are thought to represent inherited Mesozoic structures that initially (1) developed as transform faults during Mesozoic growth of the Tethys Ocean lithosphere, then (2) served as lateral ramps and transpressive structures during the closure of the Tethys oceanic realm and the consequent Oligocene-Aquitanian continental collision, and finally (3) focused transfer fault zone development during post-Burdigalian to recent extension (Bartole, 1995; Carmignani et al., 1995; Liotta, 1991). It is worth noting that the pre-Burdigalian activity along these structures was characterised by left-lateral transpression, as testified to by several occurrences described in the basement rocks of Corsica and Sardinia (Carmignani et al., 1995).

One of the most important NE-trending lineaments corresponds with Elba Island and shows an apparent left-lateral offset of the very prominent N-S Elba Ridge. That ridge coincides with a positive magnetic anomaly more than 300 km long (Fig. 8.10a; (Cassano et al., 2001; Chiappini et al., 2000), interpreted as a suture zone dominated by ophiolitic rocks stacked during construction of the Apenninic nappe and thrust complex (Bartole et al., 1992). The left-lateral offset of this N-S structure can be related to the pre-extensional tectonic activity. We propose that, with the onset of extension, the NE-trending structure was reactivated as a transfer zone (the WETZ) that controlled the late Miocene emplacement of the ODS. The SE margin of this transfer zone
across western Elba corresponds with the PP line, that not coincidentally, is located along the SW continuation of the most prominent NE-trending lineament on land, i.e. the Piombino-Faenza line (Bartole, 1995).
Fig. 8.10 – (a) Distribution of magmatic centres and regional transversal lineaments. Red stars: magmatic centres; yellow stars: intrusive bodies inferred on the basis of geophysical data; green contoured areas: positive magnetic anomalies (values in nT); dashed thick blue line: alignment of magnetic maxima; black thin lines: isobaths in metres below the sea level; diagonal hatch pattern: main post-Burdigalian sedimentary basins; diagonal ruling: Corsica Alpine tectonic units; crossed hatch: Variscan Sardinia-Corsica basement; solid black lines with ticks on the down-dropped side: normal faults; grey dashed lines: transversal geological-geophysical lineaments; yellow stripes: linear array of magmatic centres of the Tuscan Magmatic Province. Main rock association of magmatic centres: (1) Sisco, (2) Capraia Island, (3) Orciatico, (4) Montecatini Val di Cecina, (5) western Elba Island, (6) San Vincenzo, (7) Campiglia, (8) Larderello-Travale subsurface complex, (9) eastern Elba Island, (10) Gavorrano, (11) Castel di Pietra buried complex, (12) Roccastrada, (13) Montecristo Island, (14) Monte Amiata, (15) Radicofani, (16) Giglio Island, (17) Torre Alfina, (18) Tofla, (19) Monte Cimino, (20) Manziana, (21) Cerite; (b) tectonic sketch of a theoretical west-dipping, back-rolling subduction system (simplified after Doglioni, 1991) developing dextral transfer zones in its northern sector; (c) close-up view of the Western Elba Transfer Zone linking offset zones of differential extension as indicated by the lengths of the white arrows. Modified after Dini et al. (2002).

This local situation of ODS emplacement controlled by dominantly dextral transtensional activity (Fig. 8.10) is well framed in the general scenario of the northern Tyrrhenian-Apennine system that is characterized by a strongly arcuate subduction zone, slab rollback, and associated back-arc extension. In such a setting, theoretical models (Doglioni, 1991), require dextral transtension along ENE-trending structures (Fig. 8.10b) to accommodate strain between domains affected by differential extension rates. In the area of western Elba, the interplay between the inherited pre-Burdigalian sinistral offset of the N-S Elba Ridge and the later differential extension helped to focus prolonged igneous activity in a small segment of a dextral transfer zone (Fig. 8.10c).

**Magmatism focused by transfer systems**

The western Elba intrusive complex is part of the Tuscan Magmatic Province, made up of some twenty exposed or cored igneous centres and several additional intrusions inferred from geophysical data (Fig. 8.10a) (Dini et al., 2005; Dini et al., 2003; Serri et al., 2001). The distribution of igneous centres in Tyrrhenian-Tuscan area appears to be geographically linked to segments of seven main NE-trending
lineaments (Fig. 8.10a). In particular, we identify alignments of igneous centres along NE-trending structures (Fig. 8.10a).

The well known, eastward-younging overall distribution of the ages of these igneous centres (Innocenti et al., 1992) also occurs along each of these seven alignments. This suggests an eastward sequential activation of segments of the NE-transversal structures triggering extraction, rising and emplacement of magmas.

Igneous activity at most of these igneous centres was short-lived, resulting in the emplacement of a single magmatic mass. In contrast, the formation of the western Elba intrusive complexes and the Larderello-Travale buried intrusions, were long-lived events (1.5 to 2.5 Ma). The fact that these two igneous centres lie on the “Western Elba-Piombino-Faenza” transversal line suggests that it represents the most significant alignment of transfer zones in the northern Tyrrhenian-Apennine system.

Summary

The ODS in western and central Elba where are constituted by ca. 90 km of dykes that have been mapped over an original vertical extent exceeding 4 km. In western Elba, where most dykes were emplaced in the Monte Capanne monzogranitic pluton, two distinct systems of dykes are recognized, a major system made of one set trending ENE, and a minor system consisting of two sets trending NNE and NW. The ODS were emplaced in a complex NE-SW dextral shear zone within which differential strain created minor local zones of sinistral shear.

Emplacement of the ODS, marking the end of magmatism in western Elba, is the first documented evidence for the activity of a transfer fault zone in the northern Tyrrhenian region. This long-lived magmatism in western Elba suggests the transfer zone was active for up to 1.5 Ma before activity migrated north-eastward along the Pomonte-Prochico-Piombino-Faenza transversal lineament. Similar distribution of all the magmatic centres of the Tuscan Magmatic Province suggests they can be explained as related to activity in transfer fault zone segments that developed sequentially along reactivated ancient lineaments, as back-arc extension migrated north-eastward across the Tyrrhenian-Tuscan region.
Field stops

Chiessi (Wednesday 17, PM)

Along the coast NW of Chiessi village, several Orano dykes are well exposed, either fresh (dark grey) or hydrothermally altered (green-grey). These dykes show cross-cutting relationships with the metamorphic aureole-pluton contact.

![3D rendering of the geological map of Chiessi area.](image)

The dykes have variable rheological behaviour with respect to their host: although they generally have sharp and continuous contacts, some dykes locally display flow "re-adjustment" of the dyke walls. Embayments and crenulated contacts are common and testify that the granite was not completely crystallized during dyke intrusion (Fig. 8.4d). Additionally, individual Orano dykes have been traced continuously through significant changes in orientation without petrographic variation (Fig. 8.4). These dykes also show additional evidence of complex geometry, such as sudden side-steps (Figs. 8.4c,f and Fig. 8.12), horn structures and bridges (Fig. 8.4b), and broken bridges. Resorbed K-feldspar xenocrysts commonly occur.
Figure 8.12 – a) Detailed geological map of an Orano dyke near Punta del Timone, Chiessi (W side of Monte Capanne pluton). Grey: Orano dyke; white: Monte Capanne pluton; oblique ruling: contact aureole. This single, complex dyke shows (i) variable intrusive styles from plastic to brittle and (ii) the whole range of strikes shown by the overall ODS (see also Fig. 8.4a,f). Topographic contour lines are in metres above sea level.

Colle d’Orano

Type locality for Orano porphyry dykes, soon obliterated by fast-growing lichens. The Sant’Andrea facies of the Monte Capanne pluton is here crosscut by several Orano dykes showing sudden side-steps, horn structures, bridges and broken bridges (Fig. 8.4b). Granite blocks and resorbed K-feldspar xenocrysts are common occurrence.

Capo Sant’Andrea - Punta Cotoncello (Wednesday 17, PM)

On the shore cliff at Capo Sant’Andrea some Orano porphyry dykes cut across the Monte Capanne granite (coarse-grained Sant’Andrea facies) with sharp rectilinear contacts (Fig. 8.13) swiftly changing to side-step geometry (Fig. 8.4c).

Along the coastal path between Capo Sant’Andrea and Punta del Cotoncello, several strongly altered brown Orano dykes can be observed. The higher alteration degree of the dykes with respect to the granite host generate decametre-sized embayment of the coastline.
The dykes crosscut the structures of the Monte Capanne granite, with the contact breaking across K-feldspar megacrysts of the granite (Fig. 8.14). Commonly, the Orano porphyry contains large xenocrysts of rounded K-feldspar and ocellar quartz with amphibole rim (Fig. 8.15).
Fig. 8.15 – Left: Contact between Orano porphyry dyke (below) and megacryst-rich Sant’Andrea facies of the Monte Capanne granite. Right: Close-up of the Orano porphyry with xenocrysts of rounded K-feldspar and ocellar quartz.

**Punta della Crocetta**

From the road downslope through a small gate towards the cliff shore. Jurassic pillow metabasalts of the Ligurian-Piemontese ocean basin are crosscut by an Orano mafic dyke parallel to the path (Fig. 8.16). Chilled margin of Orano dyke can be observed against host rock. Also visible is a small apophysis from the main Orano dyke within the host pillows.

Fig. 8.16 – Orano dyke crosscutting Jurassic lava pillows at Punta della Crocetta.
9. The Eastern Elba Pluton-Sill-Dyke Complex  
(Porto Azzurro-Calamita)

Introduction

The eastern Elba Island tectonic stack was intruded by magmatic bodies at 5.9-5.8 Ma. A pluton, several felsic tabular intrusions and hydrothermal veins-masses mainly occur in the Calamita schists (Fig. 3.5), consisting of psammitic-pelitic hornfels along with minor interlayered amphibolite bodies.

The intrusive units

Porto Azzurro pluton

The Porto Azzurro pluton is a megacrystic monzogranite (Tables 5.1 and 5.2) with compositions similar to the most acidic portions of the Monte Capanne monzogranite (Conticelli et al., 2001; Innocenti et al., 1992). The pluton has only limited exposure in the south-central part of eastern Elba (Figs. 3.5 and 9.1), but a significant size is suggested by the extent of thermometamorphosed Calamita schists to the south. The emplacement age of the pluton is constrained at 5.9 Ma (Maineri et al., 2003). The reconstruction of paleoisograds and their offset constrains the extent of movement on the younger Zuccale fault noted above (Pertusati et al., 1993).

Monte Castello dyke

A brownish-grey porphyritic dyke occurs in eastern Elba (Conticelli et al., 2001). The rock is quite altered, and the original phenocryst assemblage consisted of plagioclase, clinopyroxene, olivine and scattered K-feldspar megacrysts of likely exotic origin (Table 5.1). The dyke had an original shoshonite composition, and an emplacement age of 5.8 Ma (Conticelli et al., 2001). Its petrographic and geochemical features resemble those of the Orano dykes. A pattern of emplacement of mafic dykes after a main pluton occurred in western Elba at 6.9-6.8 Ma, and was apparently repeated in eastern Elba 1 Ma later.
Calamita Sill-Dyke Complex

Several generations of felsic dykes and sills crop out mostly in the eastern part of the Calamita peninsula and in the Porto Azzurro area. They strike N40E - N160E, with sub-horizontal (sills) to sub-vertical (dykes) attitudes and para-concordant to discordant geometric relationships with host rock foliation.
The felsic tabular bodies are coarse-grained pegmatitic or microgranitic and invariably have a leucogranitic composition. They are composed of quartz, K-feldspar, plagioclase and and mutually exclusive primary muscovite or biotite. Tourmaline is ubiquitous as an early crystallizing phase, as testified by its euhedral shape (Fig. 9.3a) and indicate an original high boron content of the magma. These felsic tabular bodies are locally associated to massive boron metasomatic effects (e.g. Cala Stagnone). Hydrothermal tourmaline-quartz veins cut across dykes and sills and represent a hydrothermal circulation affecting both the intrusive and hornfels rocks (Dini et al., 2008a).

Coarse-grained pegmatitic and microgranitic felsic dykes have been observed trending NNW-SSE with sub-vertical attitudes and thicknesses less than 1.7 m (Fig. 3a and 4).

**Magmatic-hydrothermal system**

In the Cala Stagnone area, magmatic dykes, hydrothermal veins and metasomatic masses are hosted within biotite–andalusite-feldspar-bearing hornfels. The main hornfels fabric strikes NW with SW dip, and represents the axial plane foliation of tight folds.

Tourmaline at Cala Stagnone is common in the magmatic rocks and dominant in the related hydrothermal and metasomatic products, mainly occurring as tourmaline-quartz hydrothermal veins and metasomatic bodies. Metasomatic bodies have developed at the contact between leucogranites and the host hornfels over a distance of few cm up to one metre, with preferential replacement of biotite-rich layers by black microgranular tourmaline and quartz (Figs. 9.2b and 9.3b). The metasomatic bodies occur as black massive rocks with relics of folded quartz veins maintaining the fabric of the hornfels. Metasomatic bodies are associated with small tourmaline veins (A-veins, see below) that propagate from the felsic dykes into the hornfels.

Tourmaline pleochroism varies from pale brown through dark brown to dark green, reflecting changes in mineral chemistry. The analysed tourmalines plot in the alkali group and can be classified as schorl (low MgO content in the range 0.4-1.9 wt%, Fig. 9.3 and 9.5).

Two sets of tourmaline-quartz veins (A-veins and B-veins) characterise the hydrothermal system; both vein sets cut across felsic dykes, hornfels rock and metasomatic bodies. They show sharp contacts with host rocks and off-shoots. The observed mutual cross-cutting relationships indicate a coeval development of the two vein sets.
Fig. 9.2 – (a) Felsic dykes at Cala Stagnone. (b) Metasomatic tourmaline replacing biotite-rich layers in hornfels close to the felsic dyke contact. (c) Tourmaline rich A-veins cutting across Calamita Schist. Maximum exposed thickness of vein (black arrow) is about 2 cm. (d) Tourmaline rich B-veins cutting across Calamita Schist. The distance between the two veins (black arrows) is about 1.2 m. e) B-veins (black arrows) cutting across hornfels, felsic dyke and metasomatic tourmaline. (f) Polished slab of a late stage B-veins showing the results of multiple infill process. The cataclastic fabric consists of brownish fine–grained uvitic tourmaline cementing clasts of early black schorl-dravite tourmaline and host rock. After Dini et al. (2008a).

The A-veins are centimetre to decimetre-spaced N- to NNW-striking veins moderately to steeply dipping to the W (Figs. 9.2c and 9.4). Veins
widths range from 0.2 to 8 cm and exposed lengths between 0.9 to 3 m. These veins, with nearly constant width and single tapering terminations, are completely filled with fine-grained black tourmaline and minor quartz with isotropic fabric (Fig. 9.3c).

Fig. 9.3 – (a) Doubly terminated euhedral crystals of schorl in felsic dyke. All crystals are distinctly zoned, both in sections cut perpendicular and parallel to the c-axis of the tourmaline grains. In (001) sections, the crystals have irregular cores, pale-brown in colour, with darker euhedral rims. Sections parallel to the c-axis show a different texture, with a wedge-shaped zone on either side of the crystal, recalling the hourglass zoning. In complete tourmaline crystals (see inset), this hourglass consists of a short green wedge on the analogous c- pole enriched in Ti (TiO2 = 0.9-1.0 wt%), a more developed pale-brown wedge on the antiloggous c+ pole depleted in Ti (TiO2 = 0.3-0.4 wt%), embedded in a darker brown tourmaline (TiO2 = 0.6-0.7 wt%). (b) Polygonal aggregate of metasomatic tourmaline with interstitial quartz. The crystals show a concentric zoning with a progressive Mg enrichment towards the rims. Quartz, and rarely tourmaline, contains numerous fluid inclusions (~10-30 µm) with halite crystals (see inset). (c) Cataclastic texture in a B-vein: fragments of the first tourmaline generation (coarse grained; schorl-dravite) cemented by the very fine-grained tourmaline of second generation (dravite-uvite); (d) Detail of the second tourmaline generation in B-veins: note the difference in grain size and crystal shape of this tourmalines with respect to the first hydrothermal generation and the metasomatic tourmaline.
The B-veins consist of decimetre to metre spaced E-W to WNW-ESE veins dipping gently to S-SW (Figs. 9.2d and 9.4). Thicknesses range from 0.1 to 25 cm and exposed lengths between 3 and 15 m. The set B shows marked variations of width along-strike, with single tapering, Lower hemisphere equal area stereonet showing poles to felsic dykes A-veins and B-veins.forked terminations, branching structures, off-shoots and intense cataclasis of host rock in the intervening space (Fig. 3d, e, f). The B-veins are characterised by variable fabrics and infilling materials. Some veins are filled only by fine-grained black tourmaline and quartz and their width is commonly about 2 cm. Most of B-veins show a composite fabric: the margins consist of fine-grained isotropic black tourmaline and quartz (as for A-veins), while the inner parts exhibit a coarse-grained cataclastic fabric with angular fragments of host rock and early tourmaline vein infill set in a very fine-grained matrix of brownish tourmaline and quartz (Figs. 9.2f and 9.3c, d).

Hydrothermal and metasomatic tourmalines are petrographically and chemically different from magmatic ones. Metasomatic bodies are made
by a fine-grained (100-500 µm) polygonal aggregate of subhedral to
euhedral tourmaline crystals with short prismatic habit (Fig. 9.3b).
Tourmaline crystals show a concentric zoning with a progressive Mg
enrichment towards the rim with schorl-drauvite brown cores and dravitic
dark-brown to green rims. The calcium content is higher (0.4-0.9 wt%) than in magmatic tourmaline from felsic dykes (0.1-0.4 wt%) (Figs. 9.5 and 9.6).

The black tourmalines from A-veins and the first generation in B-veins are prismatic and their aggregates display variable textures from fine-grained equigranular (100-500 µm) to very fine-grained (50-200 µm), with intergrown elongated crystals (Fig. 9.2c and d). Tourmaline from the coarser type shows both reverse concentric zoning (from schorl to dravite) and patchy-zoned grains; pleochroic colors are similar to those of metasomatic tourmalines. Finally, very fine-grained elongated tourmaline crystals display an intermediate schorl-dravite composition with low calcium content (0.3-0.7 wt%) and negligible zoning (Figs. 9.5 and 9.6).

The very fine-grained brownish tourmalines (second generation in B-veins) display an isotropic intergrowth of elongated wedge-shaped crystals (50-150 µm) with triangular section and skeletal terminations, set in a cryptocrystalline groundmass (5-30 µm; Fig. 5c and d). Pleochroism ranges from pale-brown to brown without any significant zoning. The high calcium and magnesium content (1.7-2.9 and 8.3-10.8 wt% respectively) and the titanium abundance (1.1-1.7 wt%) are indicative of the significant uvite component (Figs. 9.5 and 9.6).

Metasomatic and hydrothermal quartz and tourmaline host a wealth of fluid inclusions containing halite crystals and several daughter/entrapped minerals (e.g. inset in Fig. 5b).

Cala Stagnone is an example of hydrothermal system evolving from
confined, centimetre- to metre-scale (metasomatic bodies), to large-

scale, tens to hundreds of metres (A- and B-veins) fluid circulation,

triggered by the intrusion of boron-rich magmas. The boron-rich nature

of the leucogranitic magma feeding Stagnone dykes is emphasized by

the widespread occurrence of tourmaline-rich metasomatic bodies and

hydrothermal veins as well as by the ubiquitous occurrence of
tourmaline in the dykes. The widespread tourmalinization of country
rock depends on the low Fe and Mg content of acidic magmas as a

limiting factor in the precipitation of magmatic tourmaline, leading to

massive exsolution of boron-rich fluid that can be buffered by biotite-

rich host rocks.
Fig. 9.5 - Classification diagrams for the tourmaline group minerals. The grey shaded area represents the composition of the tourmaline from the Porto Azzurro monzogranite pluton. a) Most of the Stagnone tourmalines plot in the Alkali Group
field, however, the late tourmalines from B-veins define a trend of progressive Ca enrichment (grey arrow) that enter the Calcic Group field. **b)** Triangular diagram for Alkali Group tourmalines showing the chemical evolution of tourmalines from magmatic, through metasomatic, to hydrothermal systems. Note that the Stagnone tourmalines do not follow the classic evolutive trend of granite-pegmatite systems (dashed arrow), being characterised by a reverse evolution (black arrow) controlled by the metasomatic reactions with biotite-rich hornfels and amphibolites. **c)** Triangular diagram for Calcic Group tourmalines showing that the B-vein tourmalines mostly enriched in Ca are true uvites.

The early development of metasomatic bodies is followed by tourmaline veins as the result of transient conditions of fluid-rock interaction. During the early stage, the boron-rich fluid exsolved from felsic dykes and concentrated along the dyke-host rock boundary, leading to the formation of decimetre- to metre-thick metasomatic masses through pervasive, local fluid circulation at the grain scale. Accumulation of boron-rich fluids locally built-up fluid overpressure conditions favouring hydro-fracturing and development of tourmaline vein sets. According to Oliver (1996), both vein sets are channelized fractured open systems at the mesoscopic scale within closed wall rock system, as supported by the lack of any significant recrystallization-metasomatism in the wall rock. The close spatial relationship of the two vein sets (Figs. 3 and 4) document a local switch in the minimum principal stress ($\sigma_3$ nearly sub-horizontal for A-veins and nearly vertical for B-veins) related to fluctuation in fluid overpressure (Sibson, 2000). Tourmaline textures and grain size reflect the different conditions of crystallization, ranging from fine-grained polygonal equigranular (metasomatic bodies) to very fine-grained intergrowth with elongated wedge-shaped skeletal crystal set in cryptocrystalline groundmass (veins). The former suggests a progressive replacement of biotite-rich layer in the hornfels, the latter point to fast crystallization due to cyclic pressure release in a fault valve behaviour (e.g. Sibson 2000) with cataclasis of early infilling. The multiple opening-infill process in B-veins is emphasised by at least two main stages of pressure build-up and release as testified by the cataclasis of the early black tourmaline infill sealed by the very fine-grained brownish tourmaline.

The multiple hydro-fracturing processes at Stagnone is further characterised by variable tourmaline compositions that could provide information on the fluids involved. The tourmaline hosted in felsic dykes has typical schorl composition reflecting the very low Mg content of the acidic magma (Fig. 6 and 7). In contrast, the schorl-dravite
composition of the metasomatic tourmaline indicates chemical interaction with biotite-rich layers in hornfels rocks. Indeed, the progressive biotite replacement is recorded by the reverse zoning of tourmaline. The same tourmaline composition is found in the A-veins and in the early stage of B-veins. This suggests a direct link between the metasomatic fluids, equilibrated with hornfels, and fluids involved in the early stage of hydro-fracturing.

Fig. 9.6 - Variations of Mg, Fe, Ti and F vs. Ca (atoms per formula unit, apfu) of the Stagnone tourmalines (symbols as in Figure 9.5). In these diagrams the different chemical evolution of tourmaline from metasomatic/early hydrothermal (low Ca and Ti) and late hydrothermal stages (high Ca and Ti) is clearly visualised. Also fluorine is positively correlated with calcium content.

The drastic change in texture from first- to second-generation of tourmaline in the B-veins is coupled with a significant shift towards uvite composition, that indicates Ca-rich fluid circulation. In particular, Ca enrichment is positively correlated with Ti content (Fig. 7), suggesting a different fluid source distinct from the magmatic one. Interaction with psammitic hornfels rocks (CaO ~0.3 wt%; TiO2 ~0.8 wt%; Puxeddu et al., 1984) could not account for Ca and Ti enrichment of fluids. Alternatively, the interaction of boron-rich fluids fluid with mafic rocks, as the amphibolite bodies interbedded within the hornfels (CaO ~9.2 wt%; TiO2 ~2.3 wt%; Puxeddu et al., 1984), can be regarded as an effective mechanism.
Fig. 9.7 - Conceptual model of the Stagnone magmatic-hydrothermal system. Letters indicate A- and B-veins, while different arrows show the sequential stages of fluid circulation as discussed in the text. Metasomatism occurred early in this history, as boron-rich fluids were made available by felsic dyke crystallization. Pervasive infiltration of fluids (modified through the reaction with biotite-rich rocks) at the dyke/hornfels contacts led to local fluid overpressure giving rise to the first hydro-fracturing event and deposition of black schorl-draviritic tourmalines into A-veins and early B-veins. Later, new fluid pressure build-up occurred, presumably by the emplacement of new felsic dykes into an amphibolite-bearing portion of the Calamita unit, and Ca-Ti rich fluids exploited again B-vein structures producing cataclastic textures cemented by brownish tourmaline (dravite-uvite).

The examined system is inferred to result from multiple magma pulses interacting with different host rocks, thus producing distinct hydrothermal fluids recorded by multiple hydro-fracturing events and
different tourmaline composition. The evolution of Stagnone magmatic-hydrothermal system can be summarised as follows (Fig. 9.7): (1) felsic dyke emplacement in the hornfels rocks with boron-rich fluid release; (2) development of a confined metasomatic-hydrothermal system with fluid evolution towards Mg-rich composition; (3) build-up of fluid pressure and hydro-fracturing with A-veins and early B-veins development; (4) build-up of fluid pressure and hydro-fracturing triggered by Ca-Ti-rich fluids with growth of uvitic tourmaline in the late B-veins.

This model emphasizes the transition from confined-pervasive to channelized fluid circulation, reflected by the tourmaline fabric and composition. Moreover, the model highlights the role of fluid circulation and overpressure associated with, and generated by the emplacement of multiple sheet-intrusions in the formation of fracture systems.

**Summary**

The hydrothermal system associated with the eastern Elba igneous complex affected an area exceeding 20 km$^2$, thus representing a major hydrothermal event in the upper crust. The boron-rich fluids were highly saline, as indicated by the presence of halite in fluid inclusions at room temperature ($\geq$ 26% NaCl equivalent). Similar high-salinity fluids characterised the hydrothermal circulation in contact aureoles of Larderello-Travale geothermal field. In eastern Elba the boron-rich saline fluids circulated close to their magmatic source, as indicated by the spatial distribution of tourmaline veins in the Stagnone area. Such a confined circulation of saline fluids could have temporarily decreased the resistivity of the crustal layer hosting this hydrothermal system. Thus the Stagnone metasomatic-hydrothermal system could represent a proxy of the anomalously conductive crust characterizing the Larderello-Travale geothermal field (Dini et al., 2008a) (Gianelli et al., 1997).
Field stops

*Fosso di Fosco (Thursday 18, AM)*

Along the unpaved road ringing Monte Calamita from Capoliveri to Cala Stagnone, the Calamita schists are exposed. These are Paleozoic siliciclastic sequences regionally metamorphosed during the Oligo-Miocene Apennine belt construction, then overprinted by contact metamorphic effects linked to the intrusion of late Miocene Porto Azzurro pluton.

At the U-turn of Fosso di Fosco, the road deeply penetrates the Calamita schist formation, reaching the Eastern Elba Sill-Dyke Complex. Right on the road cut a thick sill is exposed, while in the upper part of the section, thin leucogranite layers connected by small feeder dykes represent a “pocket” example of a Christmas tree laccolith complex (Fig. 9.8).

Fig. 9.8 – “Pocket” Christmas laccolith at Fosso di Fosco.

*Cala Stagnone (Thursday 18, AM)*

From over Capo Calvo, a small unpaved road reaches the shore at Cala Stagnone. Here, subvertical, tourmaline-rich leucogranite dykes are exposed along the shore cliffs. Striking observations can be made of the metasomatic tourmaline masses and hydrothermal tourmaline veins/breccias described in the previous sections.
Fig. 9. 9 – Leucogranite dyke with tourmaline layering.

Fig. 9. 10 – Tiny pegmatite vein surrounded by metasomatic tourmaline showing preferential replacement of former biotite-rich layers in the Calamita schist host.
10. References


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