EVOLUTION OF BOUDINS UNDER PROGRESSIVELY DECREASING PORE PRESSURE – A CASE STUDY OF PEGMATITES ENCLOSED IN MARBLE DEFORMING AT HIGH GRADE METAMORPHIC CONDITIONS, NAXOS, GREECE

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ABSTRACT. During ductile deformation of marbles under high grade metamorphic conditions on Naxos, Greece, pegmatites enclosed in the marbles were deformed in a brittle fashion forming blocky boudins with quartz crystallized in the interboudin zones. We studied the three-dimensional geometry of the boudins in the field and in two large samples. The first sample was serially sectioned to observe the structural and microstructural evolution; in the second sample we mapped the surface morphology of the deformed pegmatite. In profile, the boudins can be classified as symmetric torn type boudins, evolving towards asymmetric boudins with a later domino boudin component. In three dimensions, however, the morphology of the boudined pegmatite is a simple set of normal faults with mode-I fractures in the fault tips and rotation of the fault blocks to accommodate the extension in the marble.

Deformation history was constrained by petrology and microstructures in combination with simple order of magnitude calculations of both the cooling and pore fluid pressure evolution. The dynamically recrystallized, coarse-grained calcite of the marble provides clear evidence that after the pegmatite intruded the marble and solidified, it was deformed at peak conditions of M2b metamorphism (~ 670 °C and ~ 0.6 GPa). Such pegmatitic melts contain ~ 10 percent H2O which is released during crystallization.

We infer that after crystallization of the pegmatite the pore fluid pressure in the pegmatite remained close to lithostatic due to the very low permeability of the surrounding marble, and the pegmatite was deformed at very low effective stress which led to brittle deformation of feldspar and mode-I fracturing of the pegmatite, forming torn type boudins. With time the pore fluid pressure slowly decreased, increasing the effective stress. Ongoing N-S extension thus resulted in slip along the quartz-filled interboudin zones and in block rotation, producing domino boudins.

INTRODUCTION

A marble sequence inside the high grade core on Naxos, Greece, contains an exceptionally well-exposed field example of isolated pegmatites that intruded the marble, solidified and were deformed into boudins (fig. 1). The evolution of the system is constrained by microstructural and petrological data combined with measurements from three-dimensional outcrops of the boudined pegmatites.

The aim of this study is to constrain the dynamics of the deformation of pegmatites at subsolidus conditions, to reconstruct the pore pressure evolution in the pegmatites and to give insight on deformation mechanisms, mechanical properties and stress conditions. In addition we document the three-dimensional geometry of boudins in detail and model the evolution and mechanisms of boudinage.

GEOREGICAL SETTING

Naxos is the largest Cycladic island in the Aegean Sea and belongs to the Attic-Cycladic Massif (Dürr and others, 1978), which forms an arcuate belt of metamor-
phic rocks following the trend of the Hellenic trench, the current location of NE-directed subduction of the African plate beneath the Apulian-Anatolian microplate (Jansen and Schuiling, 1976; Keay and others, 2001).

Naxos is dominated by a N-S trending elongated structural dome (fig. 2) and consists of a migmatitic gneiss core, containing rafts of marble, amphibolite, felsic schists and pegmatite intrusions, which is overlain by a series of Mesozoic metasediments (predominantly metacarbonates and metapelites). These Lower Plate rocks were affected by at least two Alpine regional tectono-metamorphic events (Urai and others, 1990). The early compressional tectonic phase during the Eocene (\(D_1\)) ended 50 to 40 Ma ago and involved subduction of continental margin material, generation of a nappe pile and regional high-pressure, low-temperature (HP-LT) metamorphism (\(M_1\); with \(T = 400\) to \(460\) °C and \(p = 0.7\) to \(0.9\) GPa) (Feenstra, ms, 1985; Avigad, 1998). This phase was followed by a period of extensional tectonics (\(D_2\)) in early Miocene (Urai and others, 1990), associated with the formation of a back-arc, thinned crust, high heat flow, rapid uplift of lower crustal rocks and intrusion of granitoid magmas (Pe-Piper and others, 1997), probably related to the southward retreat of the N-dipping subduction zone south of Crete (Urai and Feenstra, 2001). The \(D_2\)-phase was accompanied by regional greenschist facies metamorphism of Barrovian character (\(M_{2a}\); \(\sim 25\) Ma) and reached amphibolite facies in the core, associated with partial anatexis (\(M_{2b}\); \(\sim 20\) to \(16\) Ma with \(T = \) up to \(\sim 670\) °C and \(p = 0.6\) GPa) (Jansen and Schuiling, 1976; Buick and Holland, 1989; Buick, 1991a, 1991b) and the production of thermal domes and closely spaced isograds (fig. 2). Partial melting inside the leucogneiss core was associated with the intrusion of pegmatite bodies (Andriessen and others, 2002, 2003).

During \(M_2\) metamorphism the metamorphic core complex was strongly deformed in a major crustal shear zone (Lister and others, 1984) that is now visible as a ductile
carapace separating the core from the Mesozoic metasediments (Keay and others, 2001), syn-M2 quartz mylonites (Krabbendam and others, 2003) and km-scale isoclinal folds with fold axes trending N-S. These folds are coaxially refolded by open, upright folds showing evidence for E-W shortening (Urai and others, 1990) and truncated by

Fig. 2. Simplified geological map of Naxos, Greece (after Jansen and Schuiling, 1976; Urai and others, 1990) with a detailed overview of the quarry 'LB Naxos Marble' in the high grade core, near Kinidaros (coordinate system: WGS 84).
granites emplaced at, or shortly after, the \( M_{2b} \) peak. The folding, especially in the marble horizons, was accompanied by intense boudinage, pointing to a component of sub-horizontal extension parallel to the fold axes (Buick, 1991a).

With decreasing temperatures during further extension and uplift, the deformation was strongly localized in narrow post-\( M_{2b} \) mylonites. They are commonly parallel to the local orientation of the bedding or older high-grade schistosity, characterized by extreme grain size reduction and predominantly indicate Upper Plate-to-the-north movement (Urai and others, 1990; Buick, 1991a, 1991b; Schenk and others, 2005).

The \( M_2 \) metamorphic event terminates with the emplacement of the western Naxos granodiorite body (\( \sim 14 \) to \( 10 \) Ma) (Koukouvelas and Kokkalas, 2003). Unmetamorphosed Upper Plate rocks are only exposed in E- and W-Naxos, juxtaposed above metamorphic Lower Plate rocks or the western granodiorite along detachment faults (John and Howard, 1995). The rapid uplift led to the exposure of Lower Plate rocks approximately \( 5 \) Ma before present (Wijbrans and McDougall, 1988).

**Observations**

**Field Observations**

The field area is located in a marble raft inside the Naxos high grade core, close to the village of Kinidaros at the mountain Bolibas [551 m; 37°05’12.6” 025°28’20.0”; map datum: WGS84; (fig. 2)]. Here, the marble contains thin layers of amphibolite, pegmatite intrusions and some unmelted rafts of felsic schists. Due to its high purity and commercial grade, the coarse-grained calcitic marble (grain size > 15 mm) is mined by the company “LB Naxos Marble”. Flat walls and rectangular cut-offs inside the quarry allow excellent three-dimensional observation of both boudinaged pegmatites and folded and boudinaged amphibolites (fig. 3). Additionally, some cut blocks broken along the marble-pegmatite contact expose the surface morphology of the boudinaged pegmatites (fig. 3D).

Inside the quarry, most amphibolites have a thickness of less than 5 cm. The amphibolites are folded isoclinally with (sub-) vertical axial planes (average: 088/83) and fold axes dipping towards the south, up to 45°. In the limbs of the isoclinal folds they often show pinch-and-swell structures, overprinted by brittle boudins with similar morphology and orientation as the boudins in the pegmatites discussed in detail below.

Pegmatites have thicknesses ranging from 3 to 20 cm. Most pegmatite-marble contacts are characterized by skarn formed by metasomatic reactions between the acidic melt and marble. In the central part of the quarry the orientation of pegmatites is quite variable, with an average orientation of 268/74, similar to that of the steep limbs of the folds in amphibolites. The pegmatites are boudinaged into blocks with the length (L) ranging from 10 to 30 cm. The interboudin zones are usually filled with quartz, and boudin axes typically dip towards the north, at approximately 90° to the fold axes in the amphibolites.

The outcrops show some crosscutting relationships between pegmatites and folded amphibolites (figs. 3A and 3B). The intersections are always at the quartz-filled inter-boudin zones of the pegmatites (fig. 3B). Thin amphibolite layers at a distance of a few centimeters to the fractures of the pegmatites have a wavy shape concordant with the boudins (figs. 3B, 3C and 4A).

**Macrostructural Observations**

The boudin structures were studied in detail in two blocks from the quarry. Sample 1 has a size of 25 x 30 x 90 cm and was used to study microstructures, chemistry and three-dimensional geometry of the fractures (fig. 4A). Sample 2 has a surface area of 100 x 280 cm and was investigated with a special focus on surface geometry and
Fig. 3. Photographs and schematic drawings from the quarry: \textit{in-situ} (A) – (C) and from cut blocks (D) - (F) highlighting the structural relationship between the different rock types. (A) The quarry walls offer a three-dimensional view of the pegmatite that crosscuts the folded marble-amphibolite sequence (with fold axes trending N-S); (B) detail of the floor’s surface of (A) showing pegmatites fractured preferentially at intersections with amphibolites; (C) fractured and boudinaged pegmatite (picture looking SE); note the slight folding of the thin amphibolite layer with synclines at the height of pegmatite’s fractures; (D) exposed surface of a boudinaged pegmatite with remnants of skarn; (E) marble block showing the boudinage of amphibolite with early pinch-and-swell structures overprinted by later brittle fractures which are interpreted to be coeval with the boudins in the pegmatites; (F) single block (3 x 3 x 2 m) illustrating that the pegmatite intruded parallel to the limbs of the amphibolites, and in this case into the hinge of an isoclinally folded layer.
Fig. 4. The samples studied in detail: (A) Mosaic of Sample 1 cut into 34 slices; this image sequence provides three-dimensional information on the internal evolution of the fractures inside the pegmatite; (B) in Sample 2 (100 x 280 cm) the surface of the fractured pegmatite is exposed and enables detailed measurements of displacement, length and dip of the fractures; (C) shows the analyzed fractures of Sample 2; for the analysis only the faults filled with dark gray shade were chosen, as the faults shown in bright gray represent linked fracturing.
orientation of structures of the fractured pegmatite (figs. 4B and 4C), using the nomenclature of boudins from Goscombe and others (2004). By relating the measurements to the orientation of the pegmatite occurrences in the quarry, we were able to approximate the blocks’ original orientations. The structures are summarized in figure 5.

Sample 1 consists of marble, pegmatite and amphibolite. It was cut into 34 slices with a thickness of about 2 cm. The pegmatite is 6 to 8 cm thick and boudinaged, while the 2 to 3 mm thick layer of amphibolite has a wavy shape following the enveloping surface of the boudins. Close (less than 2 cm) to the contacts towards both amphibolite and pegmatite, the color of the marble changes from white to yellow-brownish (see
The intrusive contact of pegmatite and marble contains skarn with a thickness of \(~2\text{ to }5\text{ mm}\).

The analysis of the single slices provides information on the boudin/fracture structure in three dimensions (fig. 6). It shows that the pegmatite is segmented into blocky boudins. Their aspect ratios \((L/W)\) range between 0.4 and 2.3, with an average value of 1.2 (fig. 4A). The evolution of a single fracture (along its length \(Q\)) is shown in figure 6C: the tips are mode-I fractures, with the relative motion (sub-) normal to the fracture walls and with quartz veins filling the fracture (fig. 7A). Towards its half-length \((Q/2)\) in Sample 1, shear displacement increases, and the fracture is gradually rotated between the single boudins. The orientation of the fracture with respect to the pegmatite is characterized by the angle \((\varphi)\) (fig. 6C) and ranges from 0 to 5° for the mode-I fractures and around 18° for reactivated and rotated mode-I fractures at half-length \((Q/2)\) (fig. 7A). Several such fractures are found in this sample; they are all subparallel to each other and characterized by the same evolution (fig. 6A).

Sample 2 exposes the surface structure of a fractured and boudinaged pegmatite (fig. 4B). It was carefully mapped at 1:10 scale, followed by measuring length \((Q)\), displacement \((D)\) and dip of the interboudin surface \((\theta)\) between interboudin surface and boudin exterior (fig. 6B). The displacements along each fracture were measured with respect to fracture length or surface morphology (for example crusts of skarn). The maximum displacement \((D_{\text{max}})\) for each fracture was recorded and plotted against the fracture length \((D_{\text{max}}/Q)\) (for example, Peacock and Sanderson, 1991) (fig. 7B). Along \(Q\) the fractures have commonly straight to arcuate traces and are bell-shaped.
In total we studied 75 fractures ranging in length (Q) from 25 to 632 mm. These fractures were subdivided into single, isolated fractures (N = 39) and faults with segment linkage (N = 36) (with branches or relay ramps transferring the displacement to another fracture) (see fig. 4C).

The maximum displacements ($D_{\text{max}}$) for single, isolated fractures range from 3 to 34 mm with an average of 12.6 mm. The displacement gradients ($D_{\text{max}}/Q$) of these fractures range from 0.039 to 0.159 (average calculated from the best fit line: 0.088).

From the tips towards the fracture’s half-length ($Q/2$) the boudin blocks are progressively rotated, which is characterized by the angle ($\alpha$) (fig. 6B) ranging between 2 and 20° (average 8.3 ± 4.4°). The asymmetric structure of the single rotated boudins...
is characterized by the angle (θ), which—for all measured interboudin surface increments in Sample 2—ranges between 39 and 78° (average: 57.1 ± 8.1°) (fig. 7A).

Microscale Observations and Geochemistry

Microscopic observations and geochemical data were collected from slice 20 of Sample 1 (fig. 8A). Composition of the rocks was analyzed by point counting (500 points), XRF on bulk samples and microprobe analysis.

Mineralogy, petrology, geochemistry and microstructure.—

Marble

Both microprobe and XRF data establish the high purity of the calcitic marble (~ 99 to 99.5 % CaCO₃). The yellow-brownish color changes close to the contacts to amphibolite and pegmatite are due to an increased amount of Fe, with the highest values measured close to the boudin necks. Inside specific calcite grains there are minor changes in chemical composition, but no systematic variation from the core towards the grain boundary. The calcite grains in the marble of Samples 1 and 2 are characterized by a bimodal grain size distribution with the small grains ranging from 0.7 to 1 mm and the large ones from 6 to 15 mm (grain size data was obtained by measuring the equivalent circular diameter in thin sections). The coarse grains have...
Fig. 8. Microstructures. (A) Slice 20 of Sample 1 with rectangles indicating the details in (B) - (E). Numbers point to three different regions: (1) unfractured region, (2) region at fracture’s half-length and (3) contact region (see table 1 for geochemical data of (1) and (3)); (B) micrograph illustrating the slip movement of the boudins parallel to the basal planes of biotite, partly altered to chlorite; (C) micrograph showing dynamic recrystallization of calcite at the contact to the pegmatite; the coarse calcite grains are characterized by lobate grain boundaries indicating grain boundary migration recrystallization, the fine grains have similar sizes to the subgrains, indicating subgrain rotation recrystallization; garnet (gt) and hedenbergite (hdb) belong to the typical assemblage of skarns (crossed polarizers); (D) fracture (thickness: 2 mm) characterized by a dominant mode-I component and filled with quartz; this vein was subjected to further fracturing indicated by a vein with a thickness of ~200 μm, indicating different pulses of brittle failure; approaching the fracture, plagioclase is progressively altered to sericite; (E) at this fracture both albite and orthoclase grains are brittle deformed and healed with orthoclase of different composition; the chemical differences are also revealed by microprobe mapping (MM) of sodium, backscatter analysis (BS) and cathodoluminescence (CL) (see also table 2).
irregular lobate grain boundaries indicating grain boundary migration recrystallization (fig. 8C). They contain subgrains with similar sizes as the small calcite grains, characteristic for slow grain boundary migration and subgrain rotation recrystallization. The co-existence of both subgrain rotation and grain boundary migration recrystallization is characteristic of dynamic recrystallization of calcite at high temperatures (Urai and others, 1986). The microstructure of the calcite is homogeneous; there are no discernible differences (for example, grain size reduction) as a function of position with respect to the displaced parts of the boudins. Some of the calcite grains show weak undulose extinction and occasional mechanical twins indicating that the marble was only weakly affected by later deformation during uplift and cooling, although marble samples mylonitized during post-M2 metamorphism at temperatures of ~ 300 °C were found in the same marble raft (Schenk and others, 2005).

Grain sizes of both fine and coarse calcite grains were used for paleopiezometry. With a grain size of the small grains of 700 to 1000 μm in the rotation recrystallization piezometer and grain size of the large grains of 6000 to 15000 μm in the migration recrystallization piezometer of Rutter (1995), differential stresses of 2.6 to 1.9 MPa and 1.2 to 0.5 MPa are calculated, respectively.

Amphibolite

The amphibolite consists of hornblende (65%; tschermakite) and feldspar (~ 28%; predominantly andesine) together with minor amounts of sphene and muscovite. All minerals are elongated parallel to the lineation. The chemical composition of the amphibolite points to a magmatic (alkalibasaltic) origin, indicated both by the amounts of Ni and Cr and the Ni/Cr ratio (Fröhlich, 1960).

Pegmatite

The mineralogical and geochemical data of the pegmatite was subdivided into (1) the central, undeformed region, (2) the fractured region and (3) the contact region towards the marble (fig. 8A and table 1).

The central, undeformed part (1) is dominated by feldspar (predominantly andesine with minor orthoclase occurrences) with grain sizes of 2 to 3 mm. Other minerals are quartz (~ 9%), and minor biotite, sericite, apatite, sphene and rare allanite. Here there is no evidence for deformation, as for example undulose extinction, deformation bands or microfractures. Myrmekites occur at contacts of plagioclase and K-feldspar. Calculated from the H2O content of amphibole and mica, the central region of the pegmatite contains ~ 0.3 weight percent H2O. These measurements indicate that the pegmatite has a granodioritic composition.

The fractured region close to the quartz veins (2) is characterized by nearly the same mineralogy as in the undeformed parts of the pegmatite: predominantly feldspar and quartz. In addition, small amounts of biotite partly replaced by chlorite, sphene, apatite and opaque iron-rich minerals are present. With decreasing distance to the quartz veins, clear mineralogical changes are observed: the feldspar crystals are successively enriched with fluid inclusions; plagioclase is altered to sericite; the An-content of plagioclase changes from 40 (andesine) towards 5 (albite). Close to the quartz veins, feldspar is microfractured (figs. 8D and 8E). The fractures are sealed with orthoclase of different composition as revealed by microprobe analysis (table 2). Also some of the euhedral grains of sphene, apatite and quartz are microfractured.

The fractures which show displacement are also filled with quartz and biotite with occasional microfolds and kinks, locally replaced by chlorite. Traces of fluid inclusions are also found in the quartz but are less frequent than inside feldspar crystals. The vein quartz is weakly deformed with large subgrains and local grain boundary migration. Locally highly altered plagioclase crystals are enclosed in the quartz vein. These microstructures all indicate that after complete crystallization of quartz these veins were weakly deformed.
Table 1
Geochemical data of two pegmatite specimens derived from Sample 1: XRF analysis of rock forming minerals, calculated CIPW norm (after Best, 1982) and modal composition

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<thead>
<tr>
<th>XRF analysis</th>
<th>(1)</th>
<th>(3)</th>
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<tr>
<td>SiO₂</td>
<td>64.37</td>
<td>63.83</td>
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<tr>
<td>TiO₂</td>
<td>0.13</td>
<td>0.55</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>18.84</td>
<td>17.88</td>
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<tr>
<td>Fe₂O₃</td>
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<tr>
<td>FeO</td>
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<td>MnO</td>
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<td>0.04</td>
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<tr>
<td>MgO</td>
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<td>0.47</td>
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<tr>
<td>CaO</td>
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<tr>
<td>Na₂O</td>
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<td>K₂O</td>
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</tr>
<tr>
<td>P₂O₅</td>
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<td>0.18</td>
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<tr>
<td>Cr₂O₃</td>
<td>&lt; 0.005</td>
<td>&lt; 0.005</td>
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<tr>
<td>SO₃</td>
<td>&lt; 0.20</td>
<td>&lt; 0.20</td>
</tr>
<tr>
<td>V₂O₅</td>
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<td>&lt; 0.005</td>
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<tr>
<td>LOI</td>
<td>0.38</td>
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<td>total</td>
<td>99.83</td>
<td>98.21</td>
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<table>
<thead>
<tr>
<th>CIPW norm³</th>
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<tbody>
<tr>
<td>Q</td>
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<tr>
<td>Ne</td>
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<tr>
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<td>Il</td>
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<tr>
<td>Ap</td>
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<tr>
<td>total</td>
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<tr>
<th>Modal composition⁴</th>
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<tbody>
<tr>
<td>quartz</td>
</tr>
<tr>
<td>plagioclase</td>
</tr>
<tr>
<td>kali-feldspar</td>
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<tr>
<td>biotite</td>
</tr>
<tr>
<td>amphibole</td>
</tr>
<tr>
<td>apatite</td>
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<tr>
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</tr>
<tr>
<td>garnet</td>
</tr>
<tr>
<td>Ca-Al-silicates</td>
</tr>
<tr>
<td>scapolite</td>
</tr>
<tr>
<td>total</td>
</tr>
</tbody>
</table>

¹(1): specimen from undeformed region (Point 1 in figure 8A)
²(3): specimen from contact region towards the marble (Point 3 in figure 8A)
³abbreviations: Q: quartz; Ne: nepheline; Or: orthoclase; Ab: albite; An: anorthite; Di: diopside; Hy: hypersthene; Wo: wollastonite; Mt: magnetite; Hm: hematite; Il: ilmenite; Ap: apatite
⁴based on counting 500 minerals
Some calcite veins crosscut the quartz veins and may have formed by replacement reactions of plagioclase (Tröger, 1969), or during a later deformation phase, which however had to be weak, as both marble and pegmatite are not subjected to grain size reduction.

At the contact of pegmatite and marble (3) — macroscopically visible as dark rim—minerals such as scapolite, andradite (garnet) and hedenbergite are observed (fig. 8C and table 1). Such an assemblage is characteristic for metasomatic reactions between granitic magma and carbonates, also known as skarn (Meinert, 1992). The chemical composition of the garnet and pyroxene point to a tungsten skarn, a type that is often associated with pegmatitic or aplitic dikes in a high grade environment, conditions which existed inside the high grade core of Naxos during M2b metamorphism.

**DISCUSSION**

The pegmatites exhibit many interesting structural aspects: (1) mode-I fractures with quartz veins subnormal to the pegmatite and with the fragments progressively rotated towards domino-type boudins; (2) shear fractures with extraordinarily high displacement gradients; (3) feldspar deformed in a brittle fashion under medium-pressure, high-temperature conditions. On the one hand, mode-I fractures and quartz veins suggest high pore fluid pressures, which reduced the effective stress in the pegmatites, so that the low stress in the slowly deforming, dynamically recrystallizing marble was sufficient to cause brittle fracturing. On the other hand, the rotated boudins would not have formed under lithostatic pore pressures, because this would have allowed the fractures to widen progressively without rotation of the fragments.

In what follows we present a model that explains the formation of these boudins during progressive deformation of the marble, illustrated by simple order of magnitude calculations of the rates of the process (figs. 9A and 9B).

**Cooling and Crystallization of the Pegmatite, Origin of High Pore Fluid Pressure and Mode-I Fracturing**

During peak $M_{2b}$ metamorphism (20 to 16 Ma), the high grade core was at medium-pressure, high-temperature conditions (~ 670 °C and ~ 0.6 GPa) (Jansen and Schuiling, 1976; Buick and Holland, 1989), causing dynamic grain growth of calcite inside the marble towards the migmatite complex (Covey-Crump and Rutter, 1992).

<table>
<thead>
<tr>
<th>Table 2</th>
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<tr>
<td>Microprobe data of feldspars shown in fig. 8E.</td>
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<td></td>
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<tr>
<td>SiO₂</td>
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<tr>
<td>Al₂O₃</td>
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<tr>
<td>TiO₂</td>
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<tr>
<td>FeO</td>
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<tr>
<td>CaO</td>
</tr>
<tr>
<td>K₂O</td>
</tr>
<tr>
<td>Na₂O</td>
</tr>
<tr>
<td>Total</td>
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</table>

1 Ab: albite with anorthite component of ~ 20%.
2 Or1: albite with orthoclase component of ~ 95%.
3 Or2: note higher Na₂O content of ‘Or2’ as the result of the brittle fracturing and subsequent sealing of ‘Ab’ and ‘Or1’
1989; Urai and Feenstra, 2001). Swarms of pegmatites intruded the marbles synkine-

matically at these peak M2b conditions, crosscutting B1 and B2 folds in the amphib-

olites, which already had a complex history of deformation.

Considering the composition of the pegmatite (table 1), it is reasonable to assume

the emplacement temperature of the pegmatitic melt to be around 750 °C (Carm-

michael and others, 1974; Whitney, 1988). Rough approximations based on simple
calculations using a Stefan-model (for example, Turcotte and Schubert, 1982) indicate

that 750 °C hot, pegmatitic melt (with a thickness of 7 cm) can be completely solidified

in the range of hours to days after its emplacement into a host rock with a temperature

of 670 °C. Further cooling would reduce the temperature difference between pegma-

tite and marble until the regional temperature would have been achieved after

months. Such a rapid crystallization of the pegmatitic melt under water-saturated

conditions is also proposed by Matthews and others (2003), who observed graphic

intergrowth of quartz and feldspar in occurrences at the contact of high grade core

and metasediments of the Lower Series. These authors estimated that also the

buoyancy-driven propagation of the dikes was very rapid inside the high grade core:

pegmatites with a thickness of 20 cm are estimated to reach a distance of 500 m in ~ 70
days (Matthews and others, 2003). Another, albeit less likely scenario is that the

pegmatite melt intruded at the temperature of the marble and crystallized slowly

during regional cooling; we return to this later in the discussion.

Most pegmatites crystallize from H2O saturated magmas with the fluids commonly

being in supercritical state (Jahns, 1982). As a consequence of the crystallization of

predominantly anhydrous crystals, the amount of fluids in the residual melt would

increase, allowing an aqueous vapor phase to exsolve and replacement reactions to

take place. This is enhanced by the viscosity contrast between melt and co-existing

aqueous fluids, the latter being about 8 orders of magnitude higher at 800 °C under

high pressures (Jahns, 1982).

The amount of dissolved H2O in peraluminous granitic melts at conditions of

the pegmatite’s emplacement is about 10 percent (Behrens and Jantos, 2001). After

the pegmatite melt intruded the marble, cooling was accompanied by crystallization of

mainly anhydrous minerals and evolution of a hydrous fluid with metasomatic reac-
tions producing skarn at the contact zone between pegmatite and marble.

The mode-I fractures in feldspar at peak M2b conditions and at differential stresses

of about 2 MPa in the marble require near lithostatic pore fluid pressures in the

pegmatite to reduce the effective stress. This produced the quartz veins, the abundant

fluid inclusions towards the fractures, and the fractured and healed feldspar crystals.

We interpret these observations to indicate that the hydrous fluids released during the

crystallization of the pegmatite were retained for some time in the pegmatite and pore

pressures were lithostatic, allowing mode-I fracturing of the pegmatite during slow

deformation of the surrounding marble.

The pegmatite at present contains less than 1 percent H2O. Therefore fluids must

have migrated from the pegmatite into the marble (see also Matthews and others,

2002). Marble however, is characterized by very low permeabilities at elevated tempera-
tures (Baker and others, 1989; Fisher and Paterson, 1992; Buick and others, 1997;
Matthews and others, 2002).

Starting from the simple assumption that the far field stress and kinematics of
ductile flow in the marble did not change during pegmatite intrusion, the geometry of
this system is puzzling. Pegmatite intrusion into an isotropic marble would imply σ3

normal to the pegmatite, while the fractures in the pegmatite constrain σ3 to be

parallel to the pegmatite. More work is needed to clarify this point, but one explana-
tion may be that under these conditions of close-to isotropic stress in the marble
Fig. 9. Timetables to illustrate the relationship of metamorphism and deformation. Note that the time axes are not scaled. See text for detailed description. Annotations point to references: (a) Jansen and Schuiling, 1976; Feenstra, ms, 1985; Wijbrans and Dougall, 1988; Buick and Holland, 1989; (b) Covey-Crump and Rutter, 1989; (c) Urai and others, 1990; Wijbrans and Dougall, 1988.
A case study of pegmatites enclosed in marble deforming at high grade...
[\sigma_1 \approx 600 \text{ MPa}, (\sigma_1 - \sigma_3) \approx 2 \text{ MPa}] the orientation of the pegmatites was controlled by the anisotropy in the marble caused by the amphibolite layers.

Before the pegmatite was completely solidified, stresses in the melt were \((\sigma_1 = \sigma_2 = \sigma_3 = \rho_f)\), represented by a point in the origin of the Mohr diagram (state 1 in fig. 10). Fluids formed during crystallization remained in the cooling pegmatite due to the very low permeability of the marble during peak M2b conditions, although the very high pore pressures may have increased permeability of the marble. With a differential stress of \(\sim 2 \text{ MPa}\) in the marble, the magma pressure was very close to the vertical stress \((\sigma_3 = 0.6 \text{ GPa})\), so that the pressure of the hydrous fluid \((\rho_f)\) was initially also close to \(\sigma_3\). Therefore the effective stress \((\sigma')\) in the pegmatite

\[
\sigma' = \sigma - \delta \rho_f
\]  

Fig. 10. Simplified Mohr diagrams illustrating the failure conditions and fracture types. At regional M2b conditions inside the high grade core (0.6 GPa and \(\sim 670 ^\circ \text{C}\)); the dynamically recrystallized calcitic marble indicates a differential stress \((\sigma_1 - \sigma_3) = 1 \text{ to } 3 \text{ MPa}; 1)\) the pegmatitic melt intruded and crystals and melt co-existed at these conditions; 2) simple order of magnitude calculations show that high pore fluid pressures were able to be generated, shifting the Mohr circle towards the tensile field, such that the reduced effective stress allowed mode-I fractures and brittle deformation of feldspar; 3) at a later stage fluids were able to escape from the pegmatite system; subsequently the pore fluid pressure is reduced, so that the Mohr circle does not touch the failure envelope anymore; due to the extensional regime during the M2b event the previously formed mode-I fractures evolved towards reactivated and rotated mode-I fractures [to observe at the fractures’ half-length \((Q/2)]\).
was initially close to zero. Deformation of the surrounding marble caused extension of the pegmatites, so that stress became anisotropic with $\sigma_3$ increasingly tensile (state 2 in fig. 10) until it reached the tensile strength $T$ of the pegmatite (we estimate this to have been a few MPa but less than 5 MPa). Evidence for this is given by i) the brittle deformation of feldspar and fracture healing by feldspar of different composition, ii) by mode-I fractures filled with quartz and iii) by the presence of euhedral quartz grains that grew in the veins (see figs. 4B, 4C and 6A).

Measurements of the dynamically recrystallized marble point to low differential stress $(\sigma_1 - \sigma_3)$ of $2 \text{ MPa}$. Assuming a flow law of regime 3 (Schmid and others, 1980),

$$\dot{\varepsilon} = A \cdot \exp\left(\frac{-Q}{RT}\right) \cdot (\sigma_1 - \sigma_3)^n, \quad (2)$$

and using the parameters ($A = 18.4 \text{ [lnA[MPa^{-1}s^{-1}]]}$, $Q = 427 \text{ [kJmol^{-1}]}$, $R = 8.314 \text{ [JK^{-1}mol^{-1}]}$, $T = 943 \text{ [K]}$, $(\sigma_1 - \sigma_3) = 2 \text{ [MPa]}$, and $n = 4.2$) (Schmid and others, 1980; de Bresser and others, 2002), the resulting strain rate in the marble during this deformation event is $4 \times 10^{15} \text{ s}^{-1}$.

An extension ($\epsilon$) of $\sim 1 \text{ percent}$ was estimated for the mode-I fractures (see for example fig. 8D). With the calculated strain rate in the surrounding marble, this event lasted for approximately 80,000 years, during which pore pressure must have remained close to lithostatic.

If we assume a cooling scenario in which the pegmatite melt and surrounding marble were at almost identical temperatures, crystallization would have been very slow during regional cooling and would have taken place in the range of millions of years. To allow mode-I fracturing after solidification, the aqueous fluids would have to remain in the pegmatite during this much longer period, maintaining the high pore fluid pressure. Thus, regardless of the exact duration of the crystallization, the brittle deformation of the feldspar and the mode-I fractures inside the pegmatite were the result of high pore fluid pressures maintained for $\sim 80,000 \text{ years}$ or longer, at or close to peak M$_2$b conditions.

An important question is why the mode-I fracturing is commonly not perpendicular to the pegmatites’ layering and why the initial boudins already have a rhombic shape (see fig. 4A). One explanation for this could be that at the start of the deformation the pegmatites were systematically oriented at an angle to the minimum principle stress in the marble (fig. 11A), so that the fractures developed at less than $90^\circ$ to the pegmatite. Ongoing deformation resulted in progressive rotation of the pegmatites’ enveloping surface into parallelism with the principle extension direction (fig. 11B).

**Evolution of Boudinage**

If the pore fluid pressure had remained lithostatic for longer, the boudins would have been separated without rotation, with the gaps progressively filled with quartz veins. This is not the case, and also the surrounding marble was not able to flow into the gaps between the fragments, so that extension of the marble was accommodated by rotation of the fragments and sliding on the preexisting fractures and shearing of the quartz veins, as shown by undulose extinction, large subgrains and grain boundary migration in quartz. This requires an increasing effective stress to prevent progressive widening of the mode-I fractures (fig. 10; state 3).

Palinspastic restoration of the rotated boudins at the fracture’s half-length ($Q/2$) into its original state, points to a bulk extension ($\epsilon$) of $\sim 14 \text{ percent}$, which in turn gives a duration of the extensional event of $\sim 1.1 \text{ million years}$ using the calculated strain rate of $4 \times 10^{15} \text{ s}^{-1}$ in the surrounding marble. We infer this to have taken place at peak metamorphic conditions, based on the microstructures in the quartz vein, and on the homogeneous microstructures in the marble. If this deformation happened later, for example at greenschist facies metamorphic conditions, the marble would show
grain size reduction in the regions close to the sliding fractures (Schenk and others, 2005). Along the reactivated mode-I fractures, there is no evidence for skarn (fig. 8B). This again indicates that during this movement the pegmatite was already crystallized.

Chlorite replacing biotite often forms under greenschist facies conditions in the presence of fluids. This is consistent with a minor deformation of the marble at lower temperatures (strain of up to 1%), as indicated by the mechanical twins and slight undulose extinction in the marble. Another mechanism for the formation of chlorite is at high temperatures during the final phase of crystallization and autometasomatism (Tröger, 1969). In addition the decreasing An-content of plagioclase with the formation of sericite and calcite is characteristic for autometasomatism (Tröger, 1969). More work is needed to determine the conditions of chlorite formation, but both scenarios discussed above are in agreement with the main deformation and rotation of boudins to have taken place at or close to peak metamorphic conditions.

During continuing M2b extension, deformation in the surrounding ductile marble produced shear stresses acting on the predetermined, inclined mode-I fractures inside
the pegmatites (Ramberg, 1955), resulting in the progressive rotation of single boudins about their centers (fig. 11B), similar to experiments and theoretical considerations of Mandal and Khan (1991) and Kusznir and others (1991).

According to Mandal and Khan (1991) during layer-normal compression of a viscous fluid, rhombic segments undergo rotation with either their separation or interfacial slip along the interboudin zones as a function of aspect ratio (W/L) and orientation of the fracture (\(\phi\); orientation of the fracture with respect to the pegmatite layering) (fig. 12). Plotting our data derived from Sample 1 into this diagram, they predominantly fall into the field of \(k_r < 1\), that is into the field that predicts separation of the single boudins, structures that we did not observe either in the samples or in the quarry. This may be explained by two processes: i) the fractures form a three-dimensional network, so that, even if in a two-dimensional section the pegmatite is fractured, connection in three dimensions provides some coherence; in addition, the finite length of fractures (Q) and high displacement gradients would shift the field-separating graph (\(k_r = 1\)) towards the left for our case; or ii) there was some cohesion between the boudins due to crack-seal processes and precipitation of quartz veins, which inhibited the ductile marble from flowing into the interboudin regions.

The deformational behavior of brittle pegmatite and ductile marble is comparable with the flexural cantilever model for lithospheric extension during the formation of sedimentary basins (Kusznir and others, 1991, and fig. 10 therein). This model of coupled simple and pure shear during continental extension suggests that the

![Diagram showing orientation of fractures with respect to pegmatite layering and aspect ratio](image-url)
brittle upper crust deforms by faulting along planar faults, while the ductile lower crust and mantle deform by distributed pure shear. It also explains the familiar domino-style block-rotation in the upper crust with the hanging- and footwalls being always kept in contact, similar in style to the pegmatite fragments that rotated in the matrix of the ductile deforming marble.

In profile plane, the boudinaged pegmatites suggest the development from initially dilational asymmetric torn boudins (characterized by high-angle quartz-filled interboudin zones) towards a later domino boudin component documented by block rotation (fig. 4A). As discussed above, we interpret these structures to be the result of one single event, during which only the pore fluid pressure decreased progressively.

In recent papers, such asymmetric boudins are proposed as important shear sense indicators (Goscombe and Passchier, 2003; Goscombe and others, 2004). The three-dimensional geometry of the pegmatites of our study reveals that the fractures (Q) are very similar to those of normal fault series observed in extensional regimes (for example, Kusznir and others, 1991; Schlische and others, 1996; Ackermann and others, 2003; Cowie and others, 2005). Inside a general regime of N-S extension and coaxial flow of the marble, normal faulting is able to account for the boudin block rotation along predetermined fractures with an initially developed dominant mode-I component. We infer that there is no need for non-coaxial flow in the marble to produce such asymmetric boudin structures, and that the use of asymmetric boudins in a foliation-parallel boudin train as shear sense indicators needs to be regarded with caution.

The fracturing of the pegmatite inside the ductile marble may be used to constrain the mechanical properties of pegmatites during medium-pressure faulting at subsolidus conditions. We plotted our data of maximum displacement (D_max/2) against the length of the fracture (Q/2) with displacement gradients being in the range of ~ 0.04 to ~ 0.16

Fig. 13. Graphs of displacement (D_max/2) versus distance (Q/2) for the fault profiles of Sample 2 compared with faults in unconsolidated sand from Idni, Morocco (Wibberley and others, 1999). The very high displacement gradients of the pegmatite point to its weakness at medium-pressure, high-temperature conditions, comparable to unconsolidated sand (note that in this diagram the pegmatite’s data of both displacement and length are divided by the factor 2 to facilitate the comparison with the published data).
CONCLUSIONS

The exceptional outcrop conditions of pegmatites in a marble in the high grade core of Naxos at M2b conditions represent a natural laboratory to study i) the dynamics and deformation mechanisms at subsolidus conditions and ii) the three-dimensional structure of asymmetric boudinage.

Mode-I fractures in the pegmatites developed after solidification, during slow deformation of the surrounding marble, under lithostatic pore fluid pressures, with pore fluids produced from the crystallizing pegmatite melt, and the high pressure maintained by the very low permeability of the surrounding marble.

As pore fluid pressure dropped, continuing extension resulted in rotation of the boudins with slip along the interboudin zones. The fractures are characterized by extremely high displacement gradients, suggesting that pegmatites are weak under these conditions.

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REFERENCES

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