Dolomite microstructures between 390° and 700°C: Indications for deformation mechanisms and grain size evolution

Alfons Berger^{1}, Andreas Ebert², Karl Ramseyer¹, Edwin Gnos³, Danielle Decrouez³*

1: Institute of Geological Sciences, University of Bern, Baltzerstrasse 1+3, CH-3012 Bern, Switzerland

2: Geo Explorers AG, Wasserturmplatz 1, CH-4410 Liestal, Switzerland

3: Natural History Museum of Geneva, 1 route de Malagnou, CH-1208 Geneva, Switzerland

Corresponding author: Alfons Berger, Institute of Geological Sciences, University of Bern, Baltzerstrasse 1+3, CH-3012 Bern, Switzerland, alfons.berger@geo.unibe.ch, 0041 31 631 4990

Abstract

Dolomitic marble on the island of Naxos was deformed at variable temperatures ranging from 390°C to >700°C. Microstructural investigations indicate two end-member of deformation mechanisms: (1) Diffusion creep processes associated with small grain sizes and weak or no CPO (crystallographic preferred orientation), whereas (2) dislocation creep processes are related with larger grain sizes and strong CPO. The change between these mechanisms depends on grain size and temperature. Therefore, sample with dislocation and diffusion creep microstructures and CPO occur at intermediate temperatures in relative pure dolomite samples. The measured dolomite grain size ranges from 3 to 940 μm . Grain sizes at $T_{\text{max}} > 450^\circ\text{C}$ show an Arrhenius type evolution reflecting the stabilized grain size in deformed and relative pure dolomite. The stabilized grain size is five times smaller than that of calcite at the same temperature and shows the same Arrhenius-type evolution. In addition, the effect of second phase particle influences the grain size evolution, comparable with calcite. Calcite/dolomite mixtures are also characterized by the same difference in grain size, but recrystallisation mechanism including chemical recrystallization induced by deformation may contribute to apparent non-temperature equilibrated Mg-content in calcite.

Keywords: *Dolomite, marble, Naxos, mixed systems, grain sizes, CPO*

Introduction

Grain sizes and its evolution are important for the rheology of earth materials (e.g., Herwegh et al., 2005; 2011; Austin and Evans, 2007; 2009; Davis et al., 2008). The steady-state grain size includes grain growth (coarsening) and grain size reduction (recrystallization) processes and controls the rheology of rocks (Herwegh et al., 2011 and references therein). In contrast to numerous studies on calcite grain size evolution in nature and experiment (e.g., Olgaard and Evans, 1988; Covey-Crump and Rutter, 1989; Covey-Crump, 1997; Freund et al., 2001; Herwegh and Berger, 2003; Berger and Herwegh, 2004; Barnhoorn et al., 2004; 2005; Ebert et al., 2007; Austin et al., 2014), grain growth in dolomite is only experimentally investigated (Davis et al., 2011). In order to understand the rheology of dolomite, grain sizes and grain size evolution are important parameters controlling deformation and petrophysical behavior (Leiss et al., 1994; Newman, 1994; Davis et al., 2008; Delle Piane et al., 2008; Holyoke et al., 2013).

Deformation mechanisms in dolomite have been intensively investigated in nature and experiments (Wenk and Shore, 1975; Barber, 1977; White and White, 1980; Reeder and Wenk, 1983; Barber et al., 1994; Leiss et al., 1994; Leiss and Barber, 1999; Leiss and Lapp, 2001; Davis et al., 2008; Delle Piane et al., 2008; 2009; Holyoke et al., 2013; 2014). A majority of experimental data on dolomite indicate dislocation creep as the important mechanism at larger grain size, producing CPO comparable to calcite (Barber, 1977; Barber et al., 1994; Delle Piane et al., 2009). In contrast, fine-grained dolomite may deform by diffusion creep at similar conditions as calcite. The role of diffusion versus dislocation creep has been inferred from different deformation structures, and CPO in small and large grains in marbles in the Damara Orogen (Leiss et al., 1994). However, quantitative grain size data for the transition between diffusion- and dislocation creep-dominated conditions are not available.

In order to have some insights in the grain size evolution of dolomite and related deformation, we investigate grain size and microstructure of dolomite over a temperature gradient from 390°C to >700°C on Naxos. Naxos provides an ideal study area, because dolomitic lithologies outcrop over a large metamorphic spectrum and the results can be compared with studies on calcite marbles (Covey-Crump and Rutter, 1989; Covey-Crump, 1997; Krabbendam, et al., 2003; Ebert et al., 2009; 2010).

1. Geological setting

Naxos is the largest island of the Cyclades in the Aegean Sea (Greece). It consists mainly of Eocene high-pressure-low-temperature metamorphic units, strongly overprinted by an Oligocene-Miocene medium-pressure-high-temperature (Barrovian-type) metamorphic core complex and granitoid intrusions (e.g., Andriessen et al., 1987; Urai et al., 1990; Keay et al., 2001; Avigad, 1998).

The metamorphic core complex caused by post-orogenic extension, exhumation and uplift (e.g., Lister et al., 1984) forms an elongated dome structure (e.g., Jolivet et al., 2004; Fig. 1). It has been subdivided into a crystalline basement (migmatites and gneisses), overlain by marbles alternating with schists, amphibolites, meta-bauxites and ultramafic bodies. In western Naxos a large granodiorite complex has intruded the core complex (e.g., Keay et al., 2001; Fig. 1). Relics of allochthonous ophiolites and non-metamorphic sediments of the Aegean nappe complex form the uppermost tectonic unit in NW Naxos and along the boundary between granodiorite and core complex. Neogene clastic sediments unconformably overly all other units (Fig. 1).

The lithostratigraphy of the metamorphic sedimentary sequence of the core complex comprises three units: (1) A basal unit of mica schists, gneisses, quartzites and lenses of amphibolites representing a detrital sequence with volcanic intercalations, (2) a marble unit with a basal meta-conglomerate followed by black and light-colored dolomite marbles, grey-bluish calcite marbles with meta-cherts, white calcitic marbles with dolomitic layers showing at the top karstification with metabauxites, and (3) an upper unit of mica schists and gneisses with few meter-thick bands of carbonates representing a flysch sequence with carbonate layers (Bonneau et al., 1978; Dürr and Flügel, 1979).

Only few age-diagnostic fossils, i.e., foraminifera and dasycladacea, were found in dolomitic marble (c.f., locality Ns7 in Fig. 1) indicating a late Triassic sedimentation age. At Agrilos, a small island south of Naxos, green algae of late Jurassic to early Cretaceous age were reported by Dürr et al. (1978) in a dolomitic level containing metabauxites. Thus, a Mesozoic age of the marble unit is well established (Bonneau et al., 1978; Dürr et al., 1978; Dürr and Flügel, 1979). The depositional environment of these carbonaceous sequences is less well defined but the presence of green algae (dasycladacea) indicate a shallow marine platform environment in the photic zone where dolomitization of primary calcitic sediment may occur.

The metamorphic overprint during exhumation and rapid uplift in association with the metamorphic core complex formation predominates (e.g., Jolivet et al., 2004). Lower crustal rocks were brought into contact with upper crustal units along major shallow dipping shear zones (e.g., Lister et al., 1984; Urai et al., 1990). The related deformation and associated Barrovian-type metamorphism is responsible for most of the rock fabrics nowadays exposed. The metamorphic evolution has been studied using different mineral systems, which resulted in an isograd map established by Jansen and Schuiling (1976; Fig. 1; see also Feenstra, 1985; Kreulen and van Beek, 1983). The temperature isogrades reached $>700^{\circ}\text{C}$ in the center of Naxos (e.g., Jansen and Schuiling, 1976) and are $\sim 390^{\circ}\text{C}$, at the SE border of the island. The typical deformation structures in the metamorphic core complex are km-scale isoclinal folds, coaxially refolded, resulting in N-S fold axes. The geodynamic setting has been interpreted as compressional and/or extensional (Urai et al., 1990; Gautier et al., 1993; Avigad et al., 2001). In any case, during this deformation, complex folding and thrusting affect the sequence of marbles and schists (Kruckenberg et al., 2011 and references therein). This deformation affects all sampled carbonate locations producing a dome shape foliations pattern, associated with N-S oriented stretching lineations (Urai et al., 1990).

2. Methods

Analyses of the microfabrics was performed in the X-Z-plane, with X = stretching lineation, Y = axis perpendicular to the foliation. The presented microfabric data (grain size, CPO) were derived from coupled SEM-EBSD-EDX analyses (SEM = scanning electron microscopy, EBSD = electron backscatter diffraction, EDX = energy-dispersive X-ray analysis). This coupling was necessary because etching of calcite Herwegh (2000) was not applicable for dolomite, even if higher concentrated acids were used.

EBSD maps were acquired at 20kV acceleration voltage, at low vacuum pressure of 20 Pa and at current of 10-13 nA. Depending on the grain size of the dolomite areas ranging from $400\text{ }\mu\text{m} \times 400\text{ }\mu\text{m}$ to $4\text{ cm} \times 2\text{ cm}$ a step size of $1\text{--}12\text{ }\mu\text{m}$ was chosen. After quantifying the second-phase content, the microstructural data were estimated from the EBSD data with OIM software. From the resulting grain areas, the equivalent circular grain diameters (D_i) were determined:

$$D_i = 2\sqrt{\frac{A_i}{\rho}} \quad (1)$$

where A_i is the area of grain i . Based on the equation given in Berger et al. (2011), the area-weighted mean grain size (D) was calculated:

$$D = \frac{D_k}{\sum_{i=1}^n A_i} \times \sum_{k=1}^K \frac{D_k}{\sum_{j=1}^J A_{kj}} \times \sum_{j=1}^J A_{kj} \quad (2)$$

where k is one of K grain-size classes with width D_k . A_{kj} is the grain area of the grain j of all grains J in the grain size class k and A_i the area of grain i of all grains n . Using the area-weighted mean grain size has the advantage that this grain size is close to the 3D grain size. In this way one does not underestimate coarse grains, like in the case of a number-weighted average (see Berger et al., 2011).

Grain sizes are presented as average area weighted values. In addition all other phases are named second phases. Their grain size (d_p) and amount (f_p) are used to calculate the modified Zener parameter (Herwegh et al., 2011 and literature therein).

$$Z = d_p / f_p \quad (3)$$

The CPO data are presented with foliation oriented E-W and the lineation on the E position.

3. Results

3.1. Outcrop and sample description

The samples cover all metamorphic zones from lower greenschist-facies up to the migmatite zone ($>700^\circ\text{C}$). The metamorphic temperature (T_{max}) reached at each locality is calculated applying a linear extrapolation between the published isogrades (Table 1). Many localities are close to known temperature estimates and extrapolation is only minor (Fig. 1). The samples can be subdivided into three temperature groups (Fig. 1; Table 1): (1) Low temperature (lower grade metamorphic zone) in the south of the island; (2) greenschist to lower amphibolite facies transition; and (3) amphibolite facies localities in the core.

There are two localities in the low temperature zone (Table 1; Fig. 1). At locality Ns7 relics of sedimentary structures are preserved, whereas locality Ns1 shows also evidence for some deformation. The latter locality shows large dolomite boudins, with calcite layers flowing around the boudins (Fig. 2). In the transition from greenschist and amphibolite facies ($450\text{--}540^\circ\text{C}$) two localities contain calcite and dolomite layers (Ns5, Ne5), whereas at the two other localities only dolomite marbles have been analysed (Ns8, Ne11). In this group, the macrostructure changes from the above described fine-grained calcite-dolomite alternations toward a coarse-grained fabric of dolomite marbles. Even in the high temperature group, some localities show boudinaged dolomite layers (Fig. 3), but most dolomite-calcite layers

are more or less parallelized (Fig. 3). The main mesostructures in the high temperature group are deformed bands of calcite and dolomite.

The samples cannot only be subdivided into temperature groups (see above), but also into groups containing different amounts of calcite and dolomite. In nearly all dolomite samples silicates are rare. This is the reason why metamorphic index minerals rarely formed (Jansen et al., 1978).

3.2. Microstructural results

In the following, the dolomite microstructures will be presented together with calcite data of the same locality/area published by Ebert et al. (2009; 2010; Table 2). The microstructures of calcite and dolomite can be subdivided into three temperature groups.

In the low temperature group, calcite and dolomite grain sizes range from ~80-140 μm and 3-33 μm , respectively. The calcite grain size variation in outcrop Ns1 (44 to 195 μm ; Fig. 2, Table 2 and 3) is positively related to the Z-value, i.e., the presence of second phases. In addition, the average second phase grain size is larger than matrix grain size (Fig. 2c). Such microstructures are difficult to explain by a single process (see discussion below; Berger et al. 2010). The associated dolomitic sample is located in a ~100 meter sized boudin (Fig. 2a), which has a grain size smaller than calcite, but much larger than those from locality Ns7 (Table 3). The dolomite sample from locality Ns7 has relics of sedimentary and diagenetic structures and no CPO. Therefore, the fine-grained dolomitic micrite represents the matrix and coarser grained parts diagenetically grown infills. The related grain size varies between 3 and 7 μm (Table 3).

The transition from low temperature to amphibolite facies samples (Group 2) shows dolomite grain size increase from 32 and 182 μm (Table 3). Equivalent calcite marble grains are significantly larger (from 110 and 347 μm ; Table 2). The samples show variable Z-values (Fig. 4), but have microstructures as lobate grain boundaries, twinning and elongated grains, related to intracrystalline deformation of calcite and dolomite. The CPO shows variations between strong and weak CPO's. The strong CPO is similar as for the high temperature examples (Fig. 5).

The high temperature samples (Group 3) show a variety of average grain sizes of calcite (1266-2806 μm) and dolomite (113-942 μm) depending on temperature (Table 2 and 3). The variation in grain sizes is counterbalanced by more homogeneous microstructures of the calcite and the dolomite layers. Inside the high temperature group, high degree of CPO is

common, but some rare samples with a weak CPO exist. The observed microstructures are typical for intracrystalline deformation, as indicated by the occurrence of lobate grain boundaries, elongated grains with shape preferred orientations, twinning, and crystallographic preferred orientations (Fig. 6). The observed strong CPO of this group shows as for calcite a maximum of the $\langle c \rangle$ axes perpendicular to foliation and preferred orientation of the $\langle a \rangle$ axis in the lineation (Fig. 6).

4. Discussion

The calcite grain size evolution on Naxos shows a typical trend in an Arrhenius type diagram (Fig. 7; Ebert et al., 2010; Herwegh et al., 2011). This fits well with the observation of calcite grain sizes in tectonic settings characterized by high strain deformation (Ebert et al., 2007; 2008; Herwegh et al., 2011). This grain size-temperature relationship is further supported by experimental data of calcite with different second phase particles (Austin et al., 2014). In the following the coexisting dolomites will be discussed.

4.1. *The low temperature samples*

In the low temperature localities different grain sizes occur in dolomite (3-30 μm). Neither dolomite nor calcite grain sizes fit with the general trend (Fig. 7). In locality Ns1, second phase silicates are larger than matrix calcites (Fig. 2c). In contrast, the calcite samples from locality Ns1 show mainly increasing grain sizes of the second phases with decreasing calcite grain-sizes (Fig. 2). This is the opposite of a coupled grain growth trend, which shows a systematic grain size increase of both matrix grains and second phase particles (Berger et al., 2010). Possible causes for this relationship are either a fast metamorphic growth of the silicates without coarsening of the calcite or a grain size reduction of the calcite without grain-size reduction of the mica. The latter is supported by indications of calcite recrystallization. However, this interpretation requires a two-stage process: A first grain growth stage, followed by a second stage of recrystallization. This hypothesis fits with the apparent too large grain size of dolomite in comparison with all other samples (grey symbols in Fig. 7). The apparent too large dolomite grain size in Ns1 locality would then reflect a preservation of an older stage inside the boudin (Figs. 2a and 7). The apparent too large grain sizes of calcite and dolomite are probably relics of older HP/LT metamorphism, which is overprinted in all other areas with the exception of locality Ns1. This grain sizes are

consistent with the estimated >12 kbar/ >470 °C conditions of the HP/LT metamorphism in SE Naxos (Avigad, 1998). The sampled rock body was apparently less deformed and metamorphosed during the major M2 metamorphic stage. Anyhow, the final grain sizes of samples Ns7 and Ns1 do not fit in the Arrhenius type evolution linking grain size and temperature using the M2 metamorphism.

4.2. Deformation at high temperatures

Deformation at high temperatures (600-700°C) is dominated by dislocation creep and dynamic recrystallization of calcite and dolomite (Fig. 3). Microstructures and CPO of the high temperature calcite and dolomite are similar and related to the subparallel mesostructural layers (see section 4.1). However, also at 620°C boudinage has been observed, which indicates that dolomite is stronger than calcite. The overall textures indicate slip systems similar as in calcites (see also Leiss et al., 1994; Delle Piane et al., 2009). There are mainly basal (001) slip in direction of $\langle a \rangle$ ($= [100]$) and rhombohedral slip in direction of $\langle r:f \rangle$ ($= [102]:[104]$; Barber et al., 1994; Fig. 6). Available deformation mechanism maps support this observation, that at low differential stresses and the measured grain size dislocation creep is the dominant process (Fig. 8). The dominance of dislocation creep in dolomite above 500-550°C has already been observed in experiments (Holyoke et al., 2013; Fig. 8). Experimental data also indicate that calcite is weaker than dolomite, consistent with our observed boudinage structures (Fig. 3).

4.3. Dolomite/calcite mixed systems

The effects of second phases on the stable grain sizes have been intensively investigated in natural samples and experimentally (e.g., Herwegh and Berger, 2004; Ebert et al., 2007; Brodhag and Herwegh, 2010; Herwegh et al., 2011; Austin et al., 2014). In all these studies, the second phases are fundamentally different from their matrix (e.g., mica in calcite), which is different for dolomite/calcite pairs. However, the Zener relations show similar relationships as for other systems (Fig. 4). In addition, the related experimental data indicate a differentiation of strain between calcite and dolomite (Delle Piane et al., 2009; Kushnir et al., 2015; see also Davis et al., 2011).

The high temperature examples displaying a mixture of calcite and dolomite occur at localities Nk2, Nz2 and Nz1 (Fig. 1). However, a well-developed CPO is preserved in several

other samples (Fig. 9; Ebert et al., 2009) indicating a dynamic recrystallization process. Therefore, the role of deformation and chemical recrystallization has to be clarified (e.g., Hay and Evans, 1987; Stünitz, 1998). Chemical recrystallization includes change of composition and/or mineral reaction (Stünitz, 1998). A special situation is given in dolomite-calcite mixtures, where the crystallographic lattice is very similar and where dolomite and calcite show an incomplete solid-solution series (Mg-calcites). Recrystallization requires maintaining calcite and dolomite compositions (specified in terms of crystallography and chemistry) to temperatures of around 900°C. However, during dynamic recrystallization of dolomite chemical components of calcite can be accommodated in dolomite. The same is true for recrystallization of calcite. This requires a change in the Ca/Mg ratio of the minerals, slowing down the mechanism of recrystallization. The observed microstructures result in local islands of dolomite in calcite and vice versa (Fig. 9). In these microstructures, dolomite relics in calcite crystals have the same crystallographic orientation as the host calcite (Fig. 9). The carbonates underwent most likely a combined mechanical and chemical recrystallization mechanism (Herwegh et al., 2003; Delle Piane et al., 2009). Such a combined recrystallization-mechanism of physical and chemical driving forces may be responsible for the apparent non-equilibrated Mg-content in calcites as reported for example in Delle Piane et al. (2009). The relics of dolomite in calcite may be a sign of such a recrystallization process, which occur in submicron scale in the investigated Naxos samples (Fig. 9).

4.4. Grain size evolution

The available grain size data of relative pure dolomite fit well in an Arrhenius type plot ($1/T$ versus $\log$ grain size; Fig. 7). The stabilized grain size depends on material parameters, temperature and Zener values (Figs. 7 and 4; Herwegh et al., 2005; Ebert et al., 2007). In a situation of steady state, the stabilized grain size (D) can be expressed as (Herwegh et al., 2005):

$$D = c' \exp(-Q/RT) * (d_p/f_p)^{m*} \quad (4a)$$

Where c' is a constant, Q an activation energy, R the gas constant and T the deformation temperature. In a pure system, the equation reduces to:

$$D = c' \exp(-Q/RT) \quad (4b)$$

This is similar to a grain growth equation (e.g., Davis et al., 2011). The grain-coarsening rate is an important parameter for the paleowattmeters and flaw laws (Austin and Evans, 2007; Herwegh et al., 2014). Experimental data of pure grain growth (Davis et al., 2011) or grain

size of deformed dolomite (Kushnir et al., 2015) are systematically smaller than their calcite equivalent. The absolute difference depends on the stabilizing processes (see above; Herwegh et al., 2005). Absolute values of a grain growth constant are only available for the grain growth constant K (different from value c' in equation (4)). However, for a given temperature and time interval a linear relation between K and c' exists. Therefore, we compare our data with the K -value of Davis et al. (2011) using a growth time of 1 Ma and using equation (5):

$$(D^n - D_o^n) = k_0 \exp(-Q/RT) (t - t_0) \quad (5)$$

The K -value of Davis et al. (2011) has to be adapted to different temperatures and activation-energy. Using the activation energy from calcite (Covey Crump, 1986: 174 kJ/mol) then the K -value of Davis et al. (2011; $\sim 5 * 10^{-5} \mu\text{m}^3/\text{s}$) translates into a value $\sim 5 * 10^{-3} \mu\text{m}^3/\text{s}$. The resulting grain size versus temperature curve fits with our measured data for calcite (Fig. 10). This indicates that the relative difference in grain size and the natural and experimental data fit well between these two carbonates. In the presented data set of steady state deformed calcite and dolomite, the relative difference of the Arrhenius type relation for dolomite is five times smaller than for calcite (Fig. 7).

5. Conclusions

The dislocation creep occurs in our example at a temperatures of 530°C, which fits very well the experimental data (Holyoke et al., 2014). Comparing experimental data (Holyoke et al., 2014), data from Naxos (this study) and other reported examples (see below), the data corroborate the published deformation mechanism maps (Fig. 8 and reference therein). The observed other plastic deformation at lower T may be a combination of diffusion and dislocation creep (compare Leiss et al., 1994; Newman and Mitra, 1994; Delle Piane et al., 2009). In the examples of the Tennessee area, dolomite underwent diffusion creep at conditions of $\sim 300^\circ\text{C}$ (Newman, 1994). A dolomitic shear zone in the Eastern Alps, which developed at 400°C, shows strong CPO accompanied with small grain sizes (Conti, 1997). The example from the Damara-orogen shows the transition from diffusion creep to dislocation dominated samples depending on grain size. The metamorphic condition reported in the latter

example is upper greenschist facies (Leiss et al., 1994; Leiss and Barber, 1999). This may be related to temperatures in the range of 450°C-500°C. The different observations are comparable with the presented Naxos data.

The comparison of grain sizes and CPO's of calcite and dolomite over a temperature range of 390-700°C indicates for dolomite the following:

- dolomite shows a grain size evolution depending on the Z-value and temperature as already shown for other minerals (Herwegh et al., 2011)
- strong CPO of dolomite occurs at temperature >500°C
- above ~500°C dislocation creep is the dominant deformation mechanism in pure dolomite at grain sizes above ~100 µm
- the grain size of steady state deformed dolomite is five times smaller than the equivalent calcite grain size (using relative pure carbonate samples).

Acknowledgements:

The study was financed through Swiss National Science Foundation grant No. 200021-109231/1. D. Bitzios is thanked for logistic support in the field and for organizing export permission. We thank G. Molli for the constructive review and M. Herwegh for fruitful discussions.

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Figure captions:

- Fig. 1 Geological map of Naxos modified after Jansen (1973) with investigated sample locations (see Table 1).
- Fig. 2: Summarizing the situation at low temperature locality Ns1. The grey calcite sample contains as second phase mainly mica with a high aspect ratio. (a) Outcrop photo of dolomite boudin (see text). (b) Zener relations for calcite and dolomite samples from this outcrop. (c) Second phase grain size versus calcite grain size. (d) Microphotograph of calcitic samples containing large white mica.
- Fig. 3: Boudinage of dolomite bands in calcite marble. (a) Locality Ns5. Note small boudins in the lower part of image. (b) and (c) Locality Nz1. (b) Boudinaged and stretched dolomite bands. Dolomite is well visible, because it is weathering back. (c) Strongly stretched dolomite boudins resulting in alternating dolomitic and calcitic bands.
- Fig. 4: Zener relations for dolomite and calcite (see Table 2 and 3). Note the general smaller grain size of dolomite at similar temperatures and the similar behaviour of both carbonates in presence of second phase particles (Zener relation).
- Fig. 5: Dolomite CPO in samples with different peak metamorphic temperatures. Samples from top to bottom: Ns1e, Ns5h, Ne5f, Nk3e (see Tables 1 and 3).
- Fig. 6: Typical dolomite CPO in high temperature sample Nk7.
- Fig. 7: Arrhenius type diagram for pure calcite and dolomite marbles.
- Fig. 8: Deformation mechanism map redrawn from Holyoke et al. (2013). Data for 600°C of this study are indicated by boxes. Both minerals should deform by dislocation-creep, which fits with the observation in Naxos. (ltp = low temperature plasticity; see Holyoke et al. (2013) for further explanations).

Fig. 9: Calcite-dolomite mixtures. (a) A dolomite rich part in the mixed system. Note the calcite islands in dolomite and vice versa (sample Nz1c). (b) Dolomite-calcite grain boundary with islands of dolomite in calcite and their crystallographic relations. Note the identical crystallographic orientation of the dolomite relics inside calcite (c-e) $\langle 001 \rangle$ CPO of sample Nz1, Nz2e. These samples show high content of calcite in dolomitic layers.

Fig. 10: Grain growth modelling result using k_0 : $5 \cdot 10^{-3} \mu\text{m}^3/\text{s}$, Q : 174 kJ/mol and $t-t_0$: 1 Ma. In addition, the dolomite data from Naxos are plotted (compare Fig. 7).

Figure1

Figure2

Figure3

Figure4

Figure5

Figure6

Figure7

Figure8

Figure9

Figure10

Table1

Table 1: Sample locations and identifications (see Fig. 1) and maximum temperature reached at the Oligocene-Miocene metamorphic overprint.

Locality	Latitude	Longitude	Calcite samples	Dolomite samples	$\sim T_{\max}$ [°C]
Ns7	36°57.57' N	25°32.46' E	Ns7c	Ns7a	380?
	36°56.00' N	25°28.55' E	Ns1a – Ns1d	Ns1e	390
	37°02.54' N	25°34.34' E	Ne6a, Ne6b		435
Ns5	37°02.43' N	25°29.06' E	Ns5a, Ns5e	Ns5h	470
Ns8				Ns8a	485
Ne5	37°05.47' N	25°33.86' E	Ne5a, Ne5b	Ne5f	500
Ne11	37°05.01' N	25°32.72' E		Ne11b, c	530
Ns4	37°02.72' N	25°29.18' E	Ns4a, Ns4f	Ns4k	560
Nz1	37°04.35' N	25°29.30' E		Nz1c, Nz1d	620
Nn7	37°09.40' N	25°33.10' E		Nn7a, Nn7b	630
Nn5	37°10.73' N	25°32.92' E	Nn5a		630
Nz2	37°05.10' N	25°29.60' E	Nz2a – Nz2c	Nz2'a,	650
				Nz2'b	
				Nz2'e,	
				Nz2'g	
				Nz2c	
Nk3	37°07.84' N	25°28.23' E		Nk3e	680
Nk2	37°04.39' N	25°27.10' E	Nk2a – Nk2l	Nk2h	700
Nk7	37°05.75' N	25°28.06' E		Nk7a	700

Table 2: Area-weighted mean calcite grain-size and modified Zener parameter (Herwegh et al., 2011) of the sample. Data from Ebert et al. (2009; 2010)

Sample	D [μm]	Z
Ns7c-k	79	1476
Ns1a_g	155	67574
Ns1b_ff	133	3852
Ns1c_i2-3	45	77
Ns1c_k2	104	1273
Ne6a_a	81	20000
Ne6b_a	108	6496
Ns5a	188	20000
Ns5e_f	204	1872
Ne5a	317	20000
Ns4d	138	252
Ns4a_w	1148	90000
Ns4f_b	160	116
Nn5a	1266	20000
Nz2a	2287	90000
Nz2b	1746	90000
Nz2b_i	864	4000
Nz2c-i	252	335
Nk2b	1494	44798
Nk2c	1466	12474
Nk2f	1816	95861
Nk2h	581	706
Nk2k	2501	36204

Table 3: Area-weighted mean dolomite grain-size and modified Zener parameter of the sample

Sample	D [μm]	Z
Nk3e	768	
Nk7a	943	
Nk2h	581	
Ns4k	132	
Ne5f	59	
Ns8a	27	
Ns5h	32	
Ns7a f	3	
Ns7a c	7	
Nn7a	424	7000
Nn7b	113	500
Ns1e_a	33	10000
Nz1c	294	1000
Nz1d	192	133
Nz2'a	612	2873
Nz2'b	413	3000
Nz2c_1	312	412
Nz2c_3	392	1796
Nz2c_4	350	2233
Nz2'e	213	344
Nz2'g	316	736
Ne11b_oben_b	156	10938
Ne11b_b	115	1554
Ne11b_1	182	13427
Ne11b_a	88	203
Ne11c_a	108	500