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Dating metamorphism and tectonic juxtaposition on Andros Island (Cyclades, Greece): results of a Rb–Sr study

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Abstract – This paper reports new geochronological data from the island of Andros, one of the less-studied islands of the Cycladic blueschist belt in the central Aegean Sea. On Andros, two tectonic units can be distinguished, the Makrotantalon unit and the Lower unit, which are separated by a low-angle normal fault, related to large-scale regional extension. Mineral assemblages indicate greenschist-facies *P–T* conditions for the last metamorphic overprint of both units. In contrast to the structurally lower unit, unambiguous indications for an earlier high-pressure stage were not recognized in rocks collected above the tectonic contact. Owing to a polyphase metamorphic evolution and incomplete resetting of the Rb–Sr isotope system during overprinting, phengite geochronology indicates a wide range in dates between *c.* 104 and 21 Ma for the Makrotantalon unit, as observed in rocks of similar structural position elsewhere in the Cyclades. The new Rb–Sr data support the interpretation, but are not conclusive evidence, that tectonic slices within the hanging wall were affected by two periods of Cretaceous metamorphism (*c.* 100–90 Ma and *c.* 80–70 Ma) and a Miocene event (*c.* 21 Ma). Tectonic juxtaposition was accomplished around *c.* 21 Ma. The Lower unit is correlative with the Cycladic high-pressure occurrences. Rb–Sr phengite dating yielded the same range in ages as determined elsewhere in the region for white mica of high-pressure rocks (*c.* 50–40 Ma) and their overprinted, greenschist-facies derivatives (*c.* 23–21 Ma). An age gradient towards the tectonic contact with the overlying Makrotantalon unit is not developed. The new results fit well into the previously established chronological framework for the larger study area. Indications for regional differences in the timing of the HP stage and/or the greenschist-facies overprint have not yet been found.

Keywords: Rb–Sr geochronology, phengite, geothermobarometry, Andros, Cyclades, Greece.

1. Introduction

Andros Island belongs to the Attic-Cycladic Crystalline Belt (Fig. 1a, b), which represents a polymetamorphic terrane within the Alpidic orogen of the Hellenides. This island occupies a central position between the well-studied high-pressure/low-temperature (HP/LT) sequences in the central part of the Cycladic archipelago and the blueschist-facies rocks that occur on Evvia and mainland Greece. The contrast between lawsonite-bearing high-pressure rocks on Evvia and the epidote-blueschist to eclogite-facies rocks in the central Cyclades (Tinos, Syros and Sifnos) suggested the existence of a metamorphic gradient in this transect (e.g. Blake *et al.* 1981). In this context, Andros is of special importance, because this island either represents a transitional zone or delineates the outer limits of the central Cycladic high-pressure realm. Despite this key location, Andros has not received the same attention as the neighbouring islands and many aspects of its tectonometamorphic evolution are poorly constrained. This is mainly due to the fact that well-preserved HP rocks are rare. Most of the island consists of pervasively overprinted, greenschist-facies sequences.

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Previous work has shown that the metamorphic succession can be subdivided into two tectonic units, the Makrotantalon unit and the Lower unit, which are separated by a subhorizontal fault (Papanikolaou, 1978). Findings of relic HP mineral assemblages indicate that the Lower unit can be correlated with the blueschist sequences of the lower main unit of the Attic-Cycladic Crystalline Belt. Judging from its structural position on top of a sequence that contains remnants of blueschist-facies rocks, it appears reasonable to suggest that the Makrotantalon unit belongs to the upper main unit of the Attic-Cycladic Crystalline Belt (e.g. Avigad & Garfunkel, 1991; Avigad *et al.* 1997), which is considered not to be affected by Tertiary HP metamorphism (Dürr, 1986; Okrusch & Bröcker, 1990). Elsewhere in the study area, greenschist- to amphibolite-facies rocks of this heterogeneous upper unit mainly yielded Cretaceous metamorphic ages. An age of *c.* 70 Ma was widely adopted as the best estimate for the time of this metamorphic episode and is deeply entrenched in the regional literature (e.g. Patzak, Okrusch & Kreuzer, 1994, and references therein). However, a relationship of the Makrotantalon unit to the upper units is as yet unconfirmed, because no age constraints for the time of metamorphism are available, and

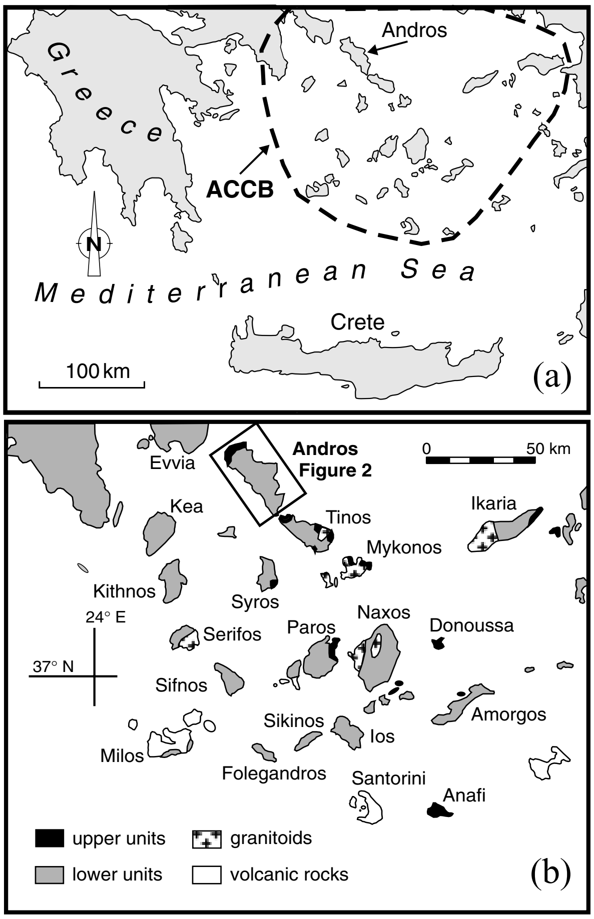

Figure 1. (a) Geographical sketch map indicating the location of Andros and the Attic-Cycladic Crystalline Belt (= ACCB). (b) Simplified geological map of the central Aegean region (modified after Matthews & Schliestedt, 1984).

an alternative interpretation suggests that this unit represents a strongly overprinted, distinct tectonic slice belonging to the lower main unit (T. Reinecke, unpub. Ph.D. thesis, Technische Univ. Braunschweig, 1982; Papanikolaou, 1987).

The focus of this paper is on the Rb–Sr geochronology of both tectonic units exposed on Andros. Samples from the Lower unit have been systematically dated in order to strengthen the petrographically based correlation with the Cycladic lower main unit and to address the question as to whether or not regional differences in the time of overprinting can be identified across the Cyclades. Furthermore, an attempt is made to identify distinct stages in the metamorphic history of the Makrotantalon unit, and to date the tectonic emplacement onto the Lower unit.

2. Geological setting

The Attic-Cycladic Crystalline Belt is built up by two major structural groups (e.g. Dürr *et al.* 1978*a*; Dürr, 1986; Schliestedt, Altherr & Matthews, 1987; Okrusch & Bröcker, 1990). The upper group is only exposed in small areas and comprises a heterogeneous sequence of unmetamorphosed Permian to Mesozoic sediments, ophiolites, greenschist-facies rocks of Cretaceous to Tertiary age, as well as Late Cretaceous medium-pressure/high-temperature rocks and granitoids (e.g. Reinecke *et al.* 1982; Altherr *et al.* 1994; Patzak, Okrusch & Kreuzer, 1994).

The lower group consists of a pre-Alpidic crystalline basement and a continental margin sequence of Permo-Mesozoic age (e.g. Dürr, 1986), which experienced Alpine HP metamorphism and a subsequent greenschist- to amphibolite-facies overprint (e.g. Altherr *et al.* 1979; Wijbrans & McDougall, 1986, 1988; Wijbrans, Schliestedt & York, 1990; Bröcker *et al.* 1993; Bröcker & Franz, 1998, Bröcker & Enders, 1999). This metamorphic history occurred in the context of collisional processes between the Apulian microplate and the Eurasian continent, and subsequent large-scale extension, caused either by the southward retreat of the Hellenic subduction zone and/or the westward extrusion of the Anatolian plate, due to collision between Arabia and Eurasia (e.g. Gautier *et al.* 1999, and references therein).

On Andros (Fig. 1b, 2), two tectonic units were distinguished (Papanikolaou, 1978). The structurally higher Makrotantalon unit has a thickness of up to 600 m and mainly consists of clastic metasediments and marbles. Metabasic schists are of subordinate importance. Fossil findings in the dolomitic carbonates yielded Permian ages (Papanikolaou, 1978). The tectonic boundary with the Lower unit is roughly marked by serpentinites (Fig. 2). Papanikolaou (1978) related tectonic juxtaposition of the two units to thrusting, whereas Dürr (1986), Avigad & Garfunkel (1991) and Avigad *et al.* (1997) suggested that the tectonic contact is a low-angle normal fault related to large-scale extension and exhumation processes.

The Lower unit is up to 1200 m thick and mainly consists of a volcano-sedimentary sequence that comprises marbles, carbonate-rich schists, clastic metasediments and metabasic rocks (Papanikolaou, 1978). Ferromanganoan metasediments are locally intercalated with metapelitic schists (Reinecke, Okrusch & Richter, 1985). The complete succession can be subdivided by means of four distinct marble horizons and three prominent greenschist layers (Papanikolaou, 1978). Disrupted bodies of meta-ultramafic, meta-gabbroic and meta-acidic rocks (up to several hundred metres in length) were recognized at various stratigraphic levels and were interpreted as olistoliths of meta-olistostromes (Papanikolaou, 1978; Mukhin, 1996).

The Makrotantalon and Lower units show similar tectonic features, indicating that both shared a common history, at least at some stages of the tectonometamorphic evolution (Papanikolaou, 1978, 1987). Details of the metamorphic history are unknown, but for both units mineral assemblages suggest greenschist-facies

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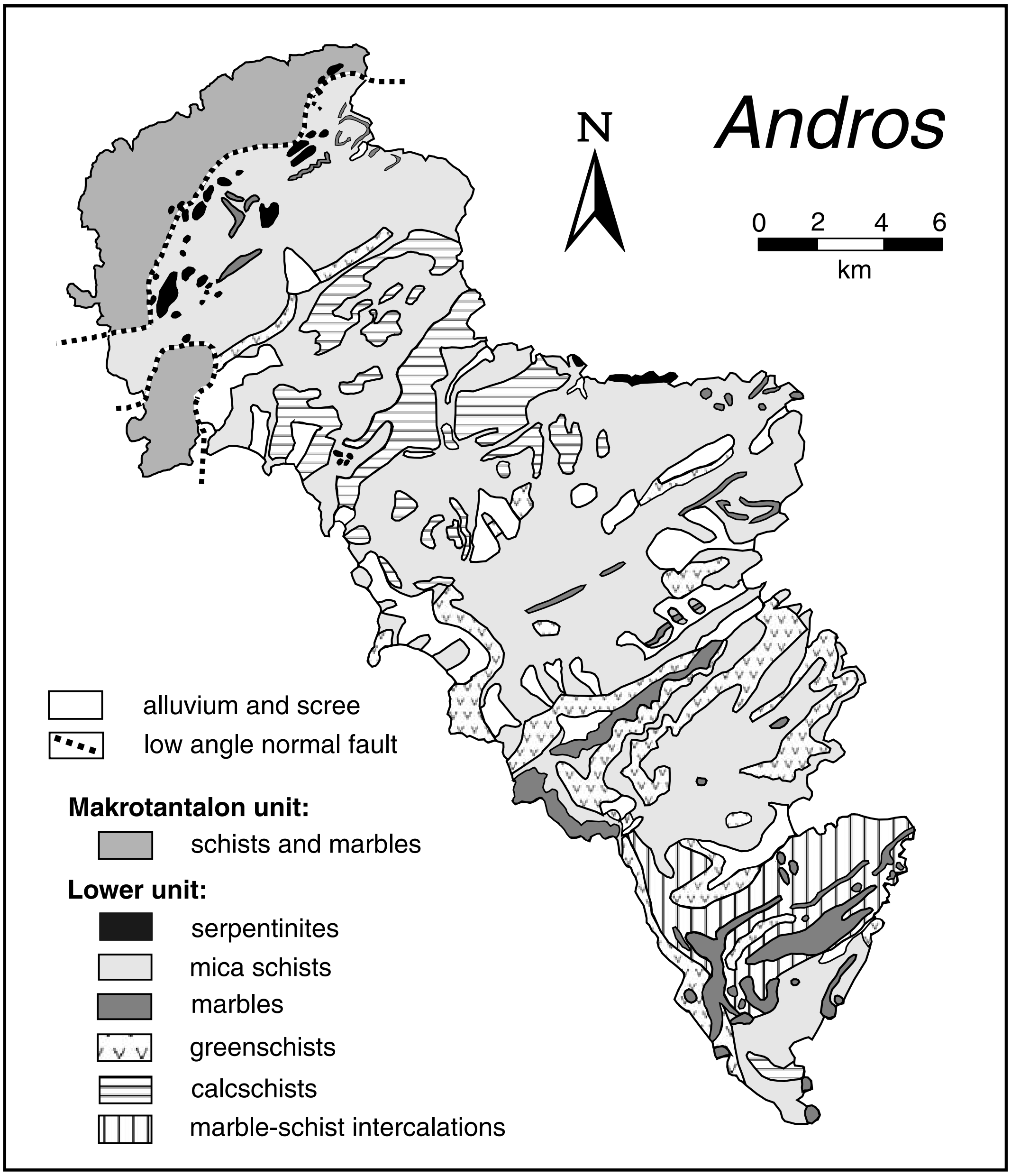

Figure 2. Geological sketch map of Andros simplified after Papanikolaou (1978). Areal distribution of the Makrotantalon unit modified after Mukhin (1996).

P–T conditions during the last metamorphic overprint (Papanikolaou, 1987). In the Lower unit, blueschist-facies relics are locally preserved, which comprise glaucophane–epidote–garnet assemblages and jadeite-rich clinopyroxene (Reinecke, Okrusch & Richter, 1985; Dekkers *et al.* unpub. data; Buzaglo-Yoresh, Matthews & Garfunkel, 1995). These findings were interpreted to suggest for the HP stage a minimum pressure > 10 kbar at temperatures similar to those attained on Tinos (*c.* 450–500 °C: Buzaglo-Yoresh, Matthews & Garfunkel, 1995). However, this interpretation is as yet unconfirmed, due to the lack of precise geothermometry. A HP stage for the Makrotantalon unit (Papanikolaou, 1987) is uncertain, because unambiguous mineralogical or textural indications for a HP episode (e.g. relic glaucophane) were not recognized.

Additional petrological work on Andros clearly is needed to substantiate models suggesting a metamorphic gradient across the Cycladic blueschist belt. Powerful arguments for this interpretation have not been presented up to now. This concept was originally supported by the absence of eclogitic assemblages in sporadic findings of relic HP rocks on Andros. The possibility that this apparent difference results from incomplete field knowledge or from a higher degree of overprinting was not taken into consideration. In the meantime, additional HP relics were recognized on Andros (e.g. Buzaglo-Yoresh, Matthews & Garfunkel, 1995), which are compatible with the interpretation

Table 1. Mineral assemblages, mineral chemical characteristics and *PT*-estimates for greenschists of the Lower unit (LU) and the Makrotantalon unit (MU)

Sample	Unit	Mineral assemblage	Si Hbl	X_{Mg} Hbl	X_{Mg} Chl	Al Ep	Si Ph	$PT_{max.}$ Triboulet (1992)
1466	LU	Act–Ep–Chl–Ph–Ab–Cc	7.538–7.851	0.60–0.65	0.46–0.49	2.443–2.532	3.347–3.475	407 °C; 6.0 kbar
1801	LU	Act–Ep–Chl–Ph–Ab	7.552–7.881	0.80–0.83	0.68–0.70	2.278–2.702	3.301–3.449	439 °C; 8.1 kbar
1807	LU	Act–Ep–Chl–Ph –Ab	7.669–7.859	0.76–0.81	0.66–0.69	2.141–2.370	3.042–3.355	370 °C; 6.6 kbar
1811	LU	Mg–Hbl/Act/(Gln)–Ep–Chl–Ph–Bt–Ab	7.364–7.734	0.71–0.77	0.57–0.59	2.145–2.429	3.084–3.507	445 °C; 7.7 kbar
1813	LU	Act–Ep–Chl–Ab	7.668–7.903	0.76–0.79	0.64–0.67	2.493–2.636	–	400 °C; 7.2 kbar
1820	LU	Mg–Hbl/Act–Ep–Chl–Ab	7.360–7.874	0.70–0.83	0.64–0.66	2.246–2.360	–	523 °C; 8.7 kbar
1426	MU	Act–Ep–Chl–Ph–Ab–Cc	7.507–7.906	0.60–0.77	0.54–0.61	2.163–2.273	3.313–3.55	455 °C; 5.4 kbar
1439	MU	Act–Ep–Chl–Ab	7.781–7.922	0.72–0.78	0.59–0.6	2.144–2.352	–	350 °C; 4.6 kbar
1843	MU	Act–Ep–Chl–Ph–Ab	7.693–7.877	0.71–0.76	0.56–0.59	2.156–2.634	3.070–3.509	430 °C; 4.1 kbar

Mineral abbreviations after Kretz (1983) and Ph – phengite. Note that Gln in sample 1811 is a relic of the blueschist metamorphism.

that the *P–T* conditions during the HP stage were not significantly different from those attained in the central Cyclades (Katzir *et al.* 2000).

3. *P–T* estimates for the last metamorphic overprint

Exact metamorphic conditions during the last overprint are difficult to constrain, due to the lack of suitable petrological tools for greenschist-facies mineral assemblages. In both tectonic units, plagioclase is almost pure albite (An_{0-3}) in all rock types and metabasic rocks consist mainly of actinolite, albite, chlorite and epidote. Magnesiohornblende only occurs in the Lower unit, in subordinate amounts. With use of the amphibole–albite–chlorite–epidote–quartz geothermobarometer of Triboulet (1992), six metabasic rocks from the Lower unit indicate temperatures of 370–520 °C at 6.0–8.7 kbar (Table 1), which broadly correspond to previous *P–T* estimates, suggesting for the overprint temperatures between 350 and 450 °C at a pressure of 5–6 kbar (T. Reinecke, unpub. Ph.D. thesis, Technische Univ. Braunschweig, 1982). This data highlights a segment of the retrograde *PT*-path through the epidote–amphibolite- and greenschist-facies, which developed subsequent to the HP-metamorphic overprint (Fig. 3). Three metabasic samples from the Makrotantalon unit yielded temperatures of 350–455 °C at 4.1–5.4 kbar (Table 1). Due to the analytical uncertainty of the microprobe results and uncertainties in the calculated Fe^{3+}-concentrations, a minimum error of ± 30 °C and ± 1 kbar should be assumed. These results suggest that the *PT*-conditions recorded in the Makrotantalon unit are identical with the lowest *PT*-conditions recorded in the Lower unit (Fig. 3), but need to be further substantiated by a detailed petrological evaluation of metamorphic conditions, which is beyond the scope of this study.

4. Sample characteristics and phengite compositions

From the Makrotantalon unit, seven samples of metasediments were selected for geochronological studies. These samples comprise five mica schists, which repre-

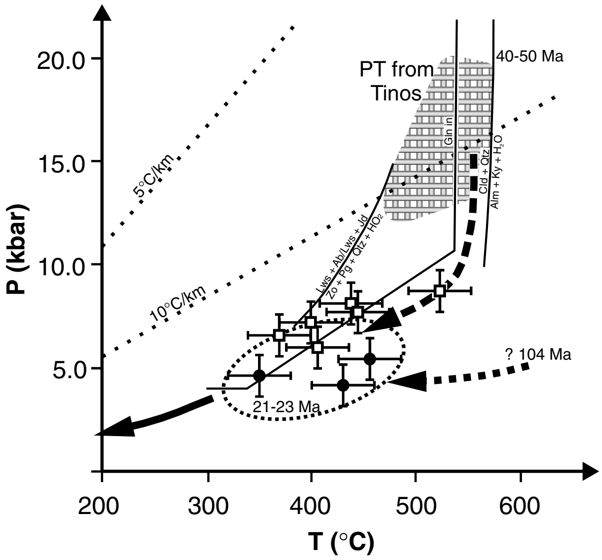

Figure 3. Results of thermobarometry and presumed *P–T* trajectories for the Lower unit (open squares and broken line) and the Makrotantalon unit (filled circles and dotted line). See text for details. Fine dotted ellipse indicates the common metamorphic overprint of the Lower unit and the Makrotantalon unit in the Miocene. *P–T* field for HP-metamorphism (squared field) based on data reported for the neighbouring island of Tinos (Bröcker, 1990).

sent the dominant rock type in the Makrotantalon unit, one calcschist and one chlorite schist. Sample locations are shown in Figure 4. Mica schists mainly consist of quartz, albite, phengite, chlorite, biotite and graphite (± opaque minerals, ± titanite/altered Ti-phase, ± zircon, ± apatite, ± tourmaline). The mineral assemblage of the studied calcschist (sample 1435) comprises calcite, quartz, phengite, chlorite, albite and graphite. The chlorite schist (sample 1845) is mainly composed of chlorite, albite, quartz, phengite, Mn-rich epidote and opaques.

From the Lower unit, eight samples were selected for phengite dating (Fig. 4), representing different lithologies (mica schist, calcschist, greenschist, meta-acidite). Most samples represent pervasively overprinted greenschist-facies rocks with mineral assemblages dominated by albite, chlorite, phengite, quartz and

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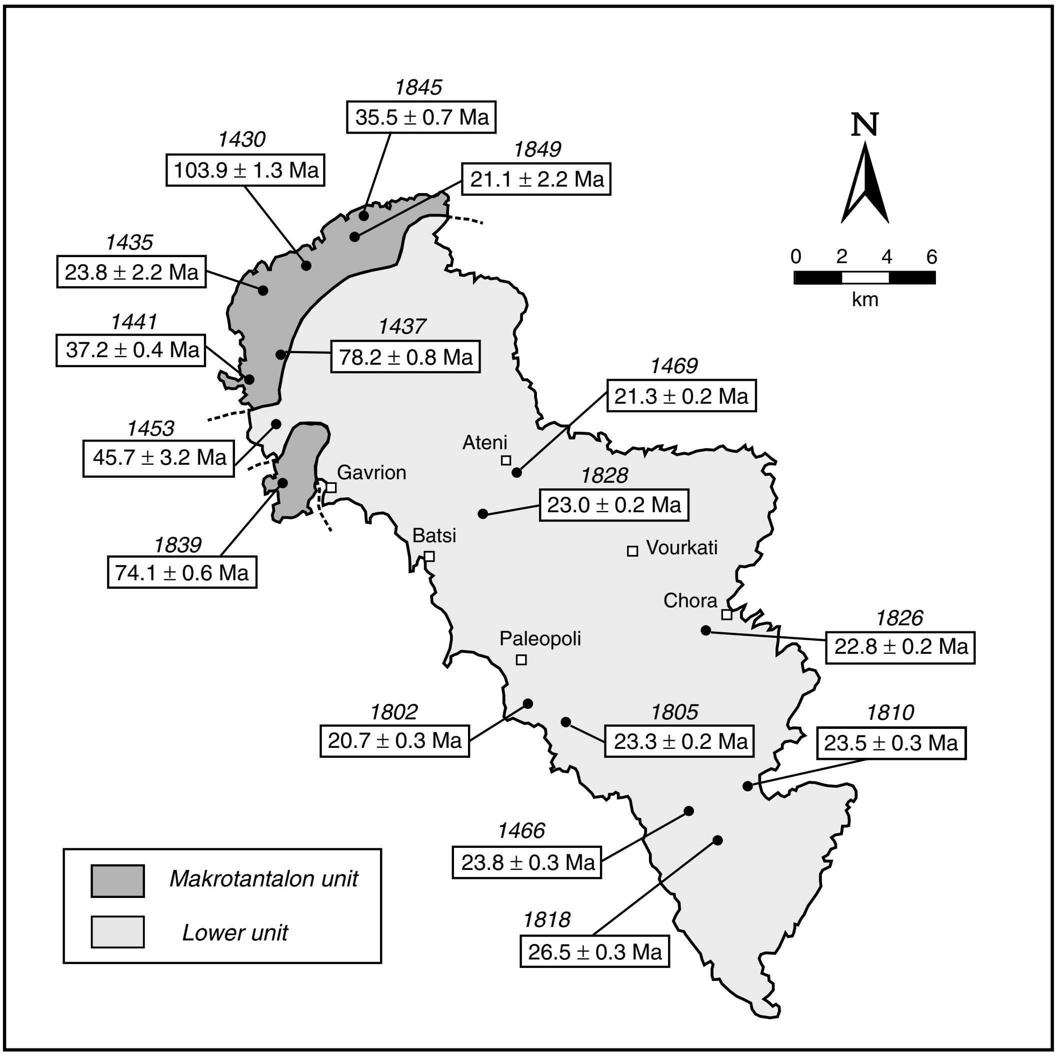

Figure 4. Sample locations and results of Rb–Sr dating. Areal distribution of the Makrotantalon unit after Papanikolaou (1978) and Mukhin (1996).

epidote/clinozoisite (± calcite, ± opaque minerals, ± titanite/altered Ti-phase, ± zircon, ± apatite, ± tourmaline). Sample 1453 is a well-preserved HP mica schist and mainly consists of glaucophane, garnet, phengite, epidote and quartz.

To characterize the white mica populations used for geochronological studies, phengite compositions were determined in thin-sections of the dated samples with the electron microprobe. Most phengite plots along the Al-celadonite–muscovite join (Fig. 5), illustrating the importance of the Tschermak's substitution [Si, (Mg, Fe^{2+}) = Al^{VI}, Al^{IV}] for the studied white mica. The data for sample 1845 from the Makrotantalon unit (Fig. 5a) and sample 1828 from the Lower unit (Fig. 5d) suggest extensive trioctahedral $FeAl_{-1}$ substitution. Phengite from metasediments of the Makrotantalon unit mostly has Si-values < 3.4 p.f.u., whereas phengite from different lithologies of the Lower unit more often yields maximum Si-values > 3.4 p.f.u. Samples from both tectonic units are characterized by a considerable range in Si-contents (MU: 3.14–3.42 p.f.u.; Lower unit: 3.12–3.55 p.f.u.), indicating compositional heterogeneity. This variability suggests that interpretation of apparent Rb–Sr ages will be fraught with problems related to mixing of different mica generations and/or inheritance.

5. Rb–Sr geochronology

Makrotantalon unit. Rb–Sr ages of the mica schists were calculated using the isotope characteristics of phengite–plagioclase pairs (Table 2). The absence/low modal abundance of Sr-rich phases (epidote, calcite, apatite) provides a reasonable explanation for the relatively high Sr-concentrations (69–174 ppm) in phengite from chlorite and mica schists (Table 2), because in such mineral assemblages white mica is a major sink for Sr. The Makrotantalon unit samples yielded phengite Rb–Sr dates between *c.* 104 and *c.* 21 Ma (Table 2; Fig. 6), which are not distributed in a systematic regional or lithostratigraphic pattern (Fig. 4). Replicate phengite analyses based on different aliquots of

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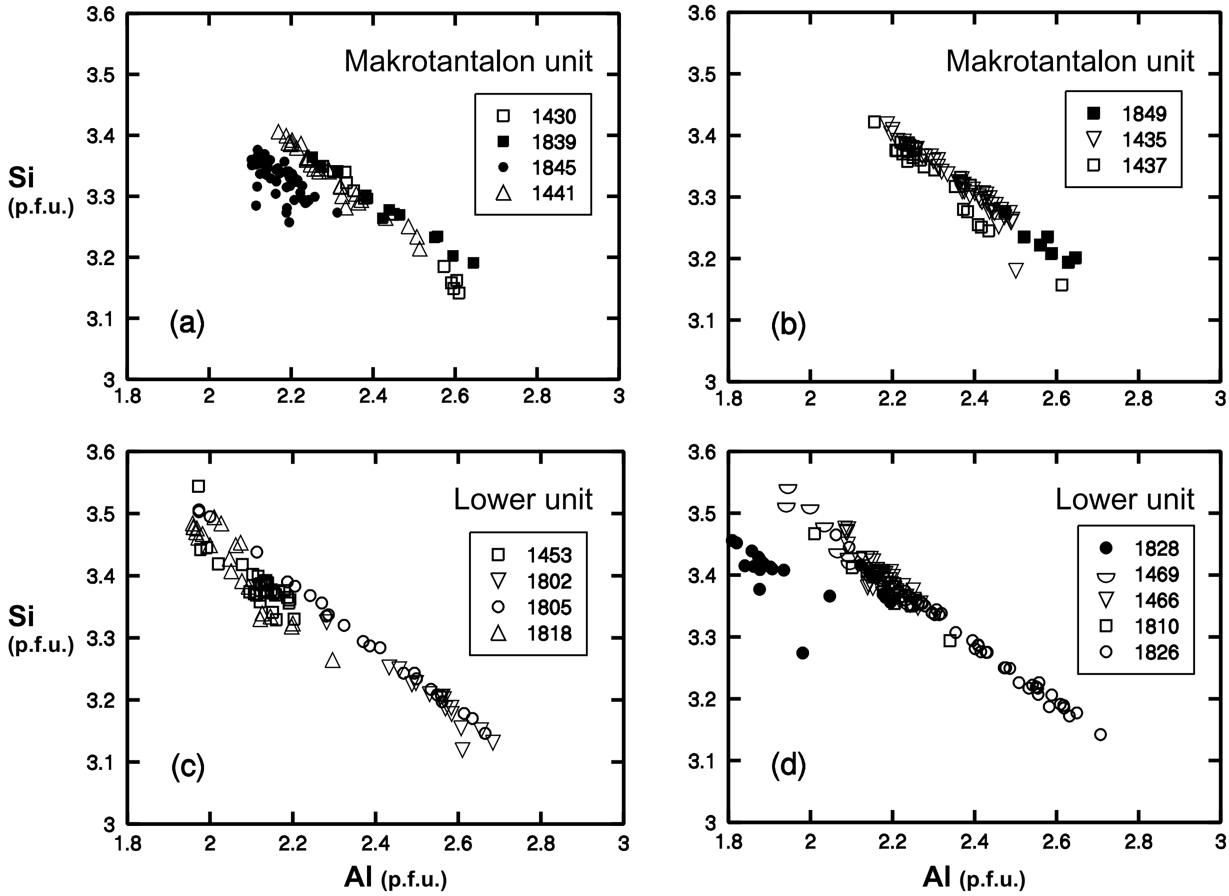

Figure 5. Si–Al^{total} variation in phengite from the Makrotantalon unit (a, b) and the Lower unit (c, d); p.f.u. – per formula unit.

Table 2. Rb–Sr isotope data for samples from the Makrotantalon unit (Andros, Greece)

Sample	Rock type	Mineral	Grain size (μm)	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$[a]	$^{87}Sr/^{86}Sr$	$\pm 2\sigma$	Age (Ma) $\pm 2\sigma$
1430	Mica schist	Phengite	250–180	406	98.8	11.91	0.732949	0.000017	103.9 ± 1.3[b]
		Phengite[c]	250–180	405	101	11.57	0.732656	0.000041	
		Albite	250–180	7.06	44.2	0.4625	0.716151	0.000012	
1435	Calcschist	Phengite	250–180	170	58.9	8.334	0.711548	0.000015	23.8 ± 2.2[b]
		Phengite[c]	250–180	239	19.1	36.26	0.721592	0.000021	
		Calcite	250–180	0.95	533	0.00514	0.708886	0.00001	
1437	Mica schist	Phengite	250–180	414	74.9	16.04	0.741179	0.000020	78.2 ± 0.8
		Albite	250–180	6.26	28.4	0.6386	0.724051	0.000010	
1441	Mica schist	Phengite	250–180	345	68.9	14.51	0.729855	0.000014	37.2 ± 0.4
		Albite	250–180	6.63	29.7	0.6462	0.722522	0.000011	
1839	Mica schist	Phengite	355–250	340	132	7.507	0.731433	0.000012	74.1 ± 0.6
		Phengite[c]	355–250	686	78.4	25.42	0.750316	0.000106	
		Albite	250–180	6.42	19.9	0.9367	0.724514	0.000019	
1845	Chlorite schist	Phengite	250–180	155	93.7	4.795	0.709711	0.000016	35.5 ± 0.7
		Whole rock		73.9	162	1.324	0.707963	0.000016	
1849	Mica schist	Phengite	355–250	242	174	4.032	0.716449	0.000013	21.1 ± 2.2
		Phengite[c]	355–250	246	153	4.658	0.716738	0.000017	
		Albite	250–180	3.41	51.6	0.1916	0.715347	0.000011	

[a]The $^{87}Rb/^{86}Sr$ ratios were assigned an uncertainty of 1% (2σ); [b]uncertainties on ages based on three data points only consider scatter, no Student-t factor; [c]replicate analyses based on newly hand-picked mineral separate with different sample weight.

samples 1435, 1839 and 1849 largely confirm the dates obtained in the first run, but indicate considerable differences in Rb and Sr concentrations, and corresponding $^{87}Rb/^{86}Sr$ ratios (Table 2; Fig. 6). Highly variable Rb and Sr concentrations (samples 1435 and 1839) suggest that the analysed aliquots represent different mixtures of a compositionally heterogeneous mineral separate. However, in both cases, the influence on the age calculations is of minor importance, at least for the questions raised in this study, indicating that compositional heterogeneity does not correspond to a significant age diversity. In case of sample 1849, it seems

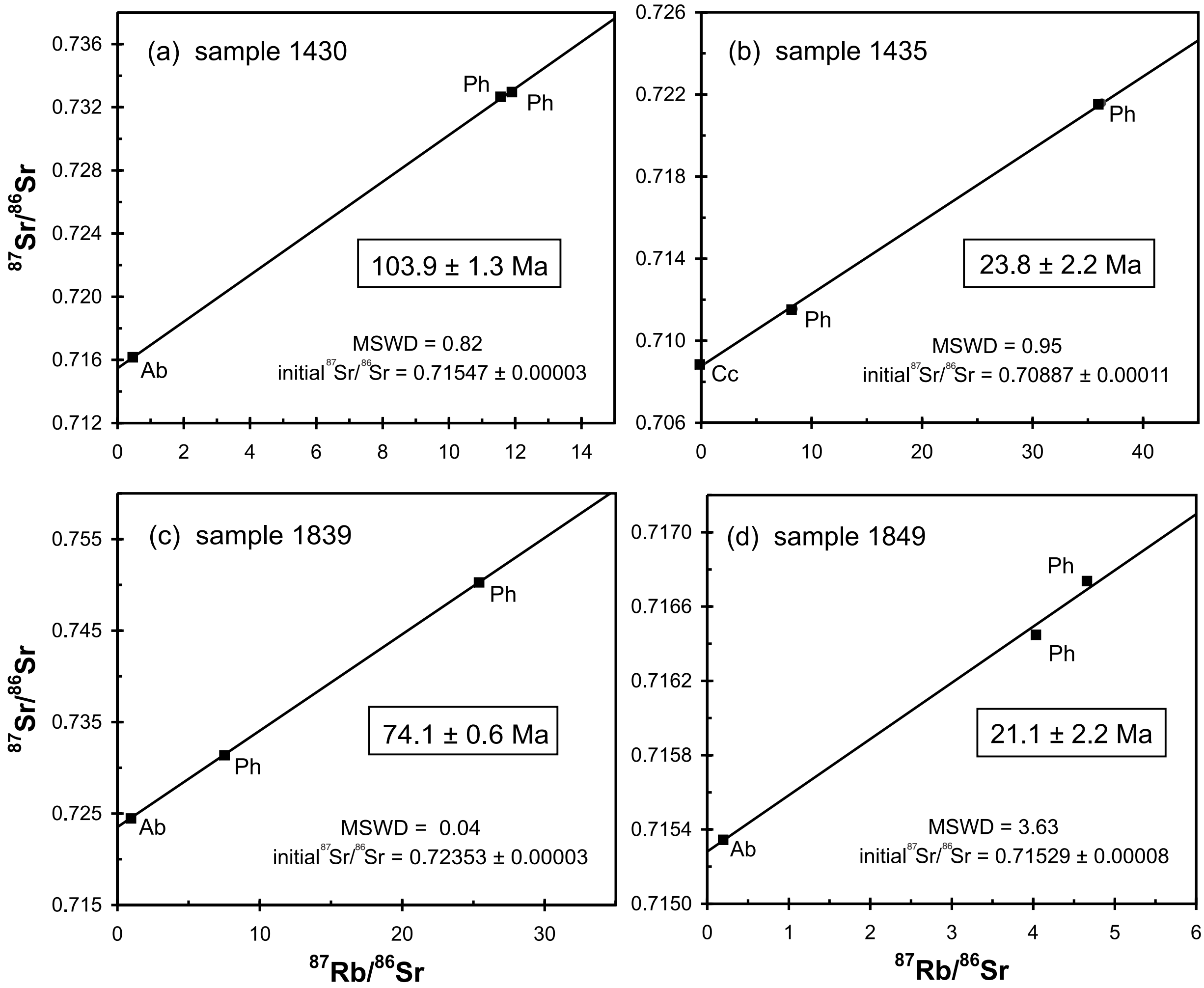

Figure 6. Selected Rb–Sr isochron diagrams for samples from the Makrotantalon unit. See Table 2 for results of additional samples. Mineral abbreviations following Kretz (1983).

more likely that one aliquot of a homogeneous mica population was slightly contaminated by a Sr-rich mineral phase, such as an inclusion of apatite.

Lower unit. The blueschist-facies sample 1453 yielded phengite–epidote and phengite–glaucophane ages of 45.2 ± 0.5 Ma and 45.0 ± 0.5 Ma, respectively. An internal isochron, which includes all three phases, indicates an age of 45.7 ± 3.2 Ma (Fig. 7a; Table 3). The greenschist-facies rocks yielded dates between 20.7 and 26.5 Ma (Fig. 6b, d; Table 3). A regional or lithostratigraphic gradient in the age distribution is not developed.

6. Discussion

Interpretation of the new Rb–Sr dates from Andros is severely hampered by general limitations of multigrain dating in polymetamorphic terranes. Due to problems related to incomplete resetting and/or mixing of different mica generations, correct identification and dating of distinct metamorphic stages may be obscured. Multigrain dating of such phengite populations cannot provide precise time constraints for specific events and, in most cases, an unambiguous interpretation is impossible. The geological significance of such dates can only be tested by application of different isotope systems to the same samples and/or by application of microsampling and/or microbeam techniques. For samples from Andros, such data is currently not available, and thus our conclusions are mainly based on inference with widely accepted facts about the regional geology.

Rb–Sr dating of samples from the Makrotantalon unit yielded highly variable apparent ages ranging between *c.* 104 and *c.* 21 Ma. Although the database for such a heterogeneous age population is still rather small and insufficient to tightly constrain the geological relevance of distinct subsets, it makes sense for the following discussion to distinguish four groups of Rb–Sr dates (*c.* 104 Ma, 78–74 Ma, *c.* 37–35 Ma and *c.* 21 Ma). What is the geological significance of these numbers? The following alternatives need to be considered: (a) the inconsistent age pattern indicates variable degrees of isotopic disturbance (e.g. incomplete

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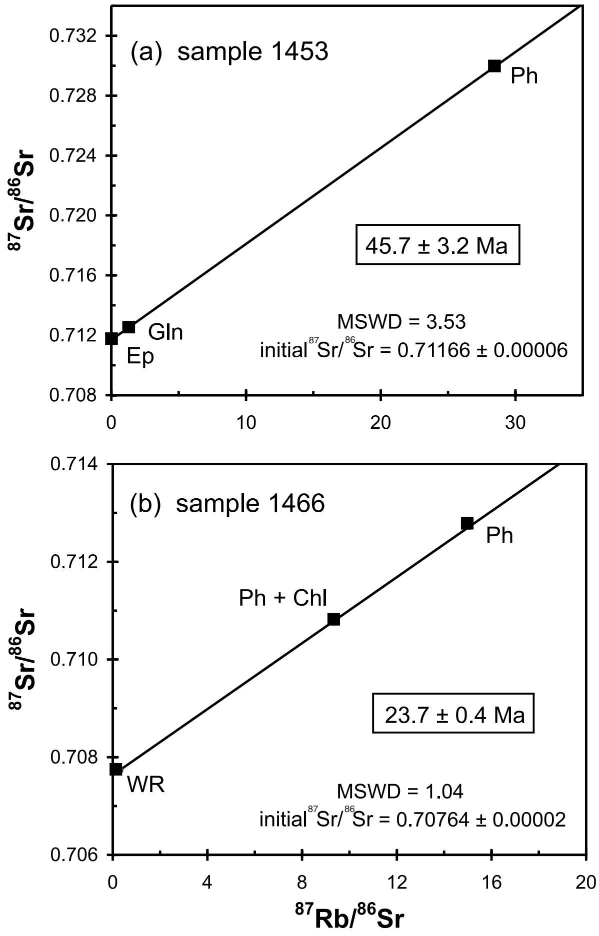

Figure 7. Selected Rb–Sr isochron diagrams for samples from the Lower unit. See Table 3 for results of additional samples (two-point isochrons). Mineral abbreviations following Kretz (1983) and Ph – phengite.

rejuvenation during metamorphic and/or deformational overprints; late-stage fluid/rock interaction); (b) the range in dates is related to mixing of different age populations, due to relic preservation of older phengite generations or within-grain age zonation in rocks that record a complex tectonometamorphic history and (c) despite superimposed modifications, the Rb–Sr results are geologically meaningful and approximate the time of distinct *P*–*T*-deformation stages. It is worth pointing out that most dates for the Makrotantalon unit fit into the broader picture and correspond to well-established age groups within the study area, as constrained by other multigrain or microbeam dating techniques. For example, different aliquots of sample 1430, representing sample weights between *c.* 1 and 3 mg, yielded a reproducible date of *c.* 104 Ma (Table 2). Although not impossible, it is unlikely that random grain selection during hand-picking has produced the same mixing properties in both cases, but it can be argued that this date is related to incomplete resetting of pre-104 Ma phengites during subsequent overprinting. On the other hand, several studies related to the upper units of the Attic-Cycladic Crystalline Belt have provided indications for a tectonothermal event at *c.* 100–90 Ma. Sanchez-Gómez, Avigad & Heimann (2002) reported K–Ar and $^{40}Ar–^{39}Ar$ phengite ages of 100–80 Ma for clasts from allochthonous sedimentary units on Mykonos and Paros and suggested a derivation from a Pelagonian nappe lying over the Cycladic blueschists in the Tertiary. Metamorphic ages around 100–93 Ma were also recognized in the greenschist-facies upper unit on Tinos (Rb–Sr phengite: Bröcker & Franz, 1998) and the amphibolite-facies Vari sequence on Syros ($^{40}Ar–^{39}Ar$ and Rb–Sr phengite, SHRIMP U–Pb zircon: Tomaschek *et al.* 2000). K–Ar ages of

Table 3. Rb–Sr isotope data for samples from the Lower unit (Andros, Greece)

Sample	Rock type	Mineral	Grain size (μm)	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$[a]	$^{87}Sr/^{86}Sr$	± 2σ	Age (Ma) ± 2σ
1453	Mica schist	Phengite	200–160	248	25.2	28.52	0.729958	0.000013	45.7 ± 3.2[b]
		Epidote	160–63	16.1	1116	0.0418	0.711676	0.000010	
		Glaucophane	160–125	3.30	7.49	1.275	0.712557	0.000020	
1466	Greenschist	Phengite	180–125	228	43.8	15.05	0.712737	0.000014	23.7 ± 0.4[b]
		Ph+chl[c]	180–125	17.4	5.35	9.418	0.710780	0.000014	
		Whole rock		47.6	497	0.2768	0.707735	0.000011	
1469	Calcschist	Phengite	250–180	231	3.90	172.3	0.760962	0.000026	21.3 ± 0.2
		Whole rock		29.5	269	0.3169	0.708943	0.000016	
1802	Meta-acidite	Phengite	250–180	141	42.7	9.516	0.709824	0.000016	20.7 ± 0.3
		Whole rock		40.2	142	0.8213	0.707268	0.000014	
1805	Mica schist	Phengite	355–250	223	39.8	16.22	0.714501	0.000011	23.3 ± 0.2
		Whole rock		46.6	599	0.2251	0.709206	0.000012	
1810	Mica schist	Phengite	355–250	130	65.8	5.697	0.709433	0.000012	23.5 ± 0.3
		Whole rock		28.7	356	0.2332	0.707612	0.000012	
1818	Meta-acidite	Phengite	180–125	170	19.9	24.72	0.716161	0.000014	26.5 ± 0.3
		Whole rock		54.1	167	0.9384	0.707202	0.000029	
1826	Mica schist	Phengite	355–250	323	46.2	20.27	0.715850	0.000011	22.8 ± 0.2
		Whole rock		72.7	371	0.5676	0.709481	0.000013	
1828	Meta-acidite	Phengite	180–125	94.0	10.2	26.70	0.714825	0.000013	23.0 ± 0.2
		Whole rock		19.1	110	0.5008	0.706269	0.000012	

[a]The $^{87}Rb/^{86}Sr$ ratios were assigned an uncertainty of 1% (2σ); [b]uncertainties on ages based on three data points only consider scatter, no Student-t factor; [c]Ph+chl–phengite–chlorite mixture.

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c. 100 Ma were also described for garnet–muscovite–albite gneisses, which occur as isolated klippen in the topmost part of the Cretan nappe pile (Seidel *et al.* 1981). In all these cases, the studied samples were collected from tectonic units that occur on top of blueschist units. Although the database for the *c.* 100–90 Ma event in the Attic-Cycladic Crystalline Belt is still very small (both on Andros and Tinos only one of the dated Rb–Sr samples falls into this group), the remarkable concordance of such dates on a regional scale, as well as the concordance of different phengite dating methods (multigrain K–Ar, ^{40}Ar–^{39}Ar, Rb–Sr) with SHRIMP U–Pb zircon results, suggests that these numbers are more than fortuitous. It is a plausible interpretation, yet not conclusively proven, that these dates indicate an episode of regional metamorphism, whose geochronological record has largely been erased by later overprints.

Previous work on several Aegean islands and Crete has provided K–Ar dates (hornblende, muscovite, biotite) between 84 and 59 Ma for low-pressure/high-temperature rocks of the upper units (e.g. Seidel *et al.* 1976, 1981; Dürr *et al.* 1978*b*; Reinecke *et al*. 1982; Altherr *et al.* 1994; Maluski, Bonneau & Kienast, 1987; Patzak, Okrusch & Kreuzer, 1994). Based on a large body of literature data and their own studies, Patzak, Okrusch & Kreuzer (1994) concluded that the upper units record the existence of a regionally important crystalline terrane of Upper Cretaceous age, which comprises rocks of different metamorphic grade and various types of granitoid intrusions. Two metamorphic rocks from the Makrotantalon unit yielded Rb–Sr dates of 74.1 ± 0.9 and 78.2 ± 0.8 Ma (Table 2) and, in combination with its structural position, strongly support a correlation with this crystalline terrane.

A characteristic feature of many geochronological datasets related to the upper units is a large spread of apparent metamorphic ages which cannot convincingly be explained by regional variations in temperature conditions or by differences in the cooling history. For example, all amphibolite-facies samples studied by Patzak, Okrusch & Kreuzer (1994) on Tinos were collected within a relatively small outcrop area, representing a lithostratigraphic thickness of *c.* 300–400 m. K–Ar hornblende ages of amphibolites range from 77 to 66 Ma and K–Ar ages of muscovites from interlayered paragneisses vary from 59 to 52 Ma, without systematic distribution on the outcrop scale. Bröcker & Franz (1998) reported apparent Rb–Sr ages (phengite-whole-rock) between *c.* 92 and 21 Ma for greenschist-facies upper unit rocks from Tinos. This scatter was recently further substantiated by K–Ar and ^{40}Ar–^{39}Ar dating of samples representing the same structural position (Zeffren *et al.* 2005). These studies have provided strong indications for variable resetting of both the K–Ar and Rb–Sr systems, due to deformational and metamorphic processes, which most likely occurred under greenschist-facies conditions in the Miocene (Bröcker & Franz, 1998; Zeffren *et al.* 2005). Such a non-pervasive, partial rejuvenation was related to tectonic juxtaposition, and the youngest date obtained from a sample collected close to the tectonic contact was considered to approximate the timing of this event (Bröcker & Franz, 1998; Zeffren *et al.* 2005). Other samples from the studied sequences mostly yielded geologically meaningless dates. This scenario also provides a plausible explanation for the new dataset from Andros, and our favoured interpretation is to suggest that dates of *c.* 37–35 Ma have no geological relevance. It is important to note that the youngest ages obtained for samples from the Makrotantalon unit (*c.* 24–21 Ma) cannot be distinguished from similar values for the timing of greenschist-facies metamorphism in the Lower unit. This apparently contemporaneous metamorphism with similar *P–T* conditions on both sides of the tectonic contact lends support to the interpretation that tectonic juxtaposition was already completed at that time, and is also in accord with the assumption that tectonic juxtaposition occurred during the metamorphic event which caused greenschist-facies overprints in both units (cf. Bröcker & Franz, 1998). A systematic younging towards the tectonic contact with the Lower unit, as described from Tinos (Zeffren *et al.* 2005), cannot unambiguously be shown for the Makrotantalon unit. Available field observations suggest that rocks with higher ages persisted at various lithostratigraphic levels of this tectonic unit, further suggesting that resetting during tectonic emplacement correlates with localized shearing and/or fluid infiltration. This interpretation also holds true for the dataset reported by Patzak, Okrusch & Kreuzer (1994).

Textural and petrological observations indicate that the structurally Lower unit on Andros was affected by blueschist- to eclogite facies metamorphism and a subsequent greenschist-facies overprint, suggesting a correlation with the lower main unit of the Attic-Cycladic Crystalline Belt. This conclusion is further corroborated by the geochronological results. A blueschist-facies metasediment yielded a Rb–Sr age of *c.* 45 Ma, and pervasively overprinted greenschist-facies rocks provided apparent ages of *c.* 24–21 Ma. Both age groups fit well into the general geochronological framework established elsewhere in the study area for the lower main unit using a variety of different dating techniques (K–Ar, ^{40}Ar–^{39}Ar, Rb–Sr white mica multigrain: e.g. Altherr *et al.* 1979, 1982; Wijbrans & McDougall 1986, 1988; Bröcker *et al.* 1993; Bröcker & Franz, 1998; Bröcker *et al.* 2004; ^{40}Ar–^{39}Ar phengite single grain: Wijbrans, Schliestedt & York, 1990; SHRIMP U–Pb zircon: Tomaschek *et al*. 2003; Lu–Hf garnet: Lagos *et al.* 2005; *in situ* UV-laser ablation ^{40}Ar–^{39}Ar phengite: Putlitz, Cosca & Schumacher, 2005). However, as indicated by variable Si-contents in phengite (Fig. 5), the samples from Andros do not contain a compositionally homogeneous mica population. Thus, their multigrain phengite dates cannot be considered to provide precise ages for distinct *P–T* stages. Such white mica populations can only indicate an upper time limit for the last overprint (*c.* 24–21 Ma) and a lower time limit for an earlier event (*c.* 45 Ma).

7. Summary and conclusions

The Makrotantalon unit occupies the highest structural level of the metamorphic succession on Andros. Mineral assemblages and geothermobarometry testify to greenschist-facies metamorphism. The time constraints presented in this study, together with the general field relationships, indicate that the Makrotantalon unit belongs to the group of upper units, which occurs in scattered occurrences throughout the Attic-Cycladic Crystalline Belt. Owing to a polyphase metamorphic evolution, Rb–Sr dates for the Makrotantalon unit range from 104 to 21 Ma, as observed in rocks of similar structural position elsewhere in the Cyclades. Despite superimposed modifications during subsequent overprints, it is here suggested that most of these dates are geologically relevant, approximating the real age of distinct tectonometamorphic events. Published and new geochronological data are compatible with the interpretation that the upper units of the Attic-Cycladic Crystalline Belt record indications for two distinct episodes of Cretaceous metamorphism (*c.* 100–90 Ma and *c.* 80–70 Ma) and a Miocene event (*c.* 24–21 Ma). During the last metamorphic overprint greenschist-facies *P–T* conditions were attained. On Andros, the metamorphic conditions during the Cretaceous events remain enigmatic, because petrological information for this part of the complex *P–T–t* path is not available. This is mainly due to the penetrative last overprint under greenschist-facies conditions, which erased all relic features of foregoing events. Elsewhere in the larger study area, metamorphism of this age is associated with low- to medium-pressure/high-temperature conditions. The scatter in apparent ages most likely records the influence of deformational and metamorphic processes under greenschist-facies conditions during the Miocene, which caused variable resetting of the Rb–Sr and K–Ar systems (this study; Bröcker & Franz, 1998; Zeffren *et al.* 2005). As previously also suggested for Tinos (Bröcker & Franz, 1998), tectonic juxtaposition of the Makrotantalon unit onto the Lower unit is interpreted to be associated with the processes causing contemporaneous greenschist-facies metamorphism at *c.* 23–21 Ma in the hanging wall and footwall of the fault zone. After this overprint, both units experienced a common exhumation (Fig. 3). Refinement of regional geodynamic considerations will require a better understanding of the local tectonometamorphic history, in combination with precise and unequivocal age information for both tectonic units exposed on Andros.

The upper units of the Attic-Cycladic Crystalline Belt bear many similarities to the uppermost tectonic unit of Crete (e.g. Patzak, Okrusch & Kreuzer, 1994; Langosch *et al.* 2000). In both cases, Cretaceous metamorphism (*c.* 70 Ma) and granitoid intrusions are well documented. In the Cyclades, there is increasing evidence that these crystalline slices experienced an earlier tectonometamorphic event at *c.* 100–90 Ma. We speculate that additional geochronology in correlative rocks on Crete may provide unambiguous evidence for similar complexities in their metamorphic evolution.

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Appendix 1. Analytical methods

Mineral compositions were determined with a JEOL JXA-8900R electron microprobe with five spectrometers at the TU Bergakademie Freiberg. Operating conditions were 15 kV acceleration voltage, 20 nA beam current and counting time of 20 s for Si, Al, Mg, Ca, Sr, and K, and 30 s for Ti, Cr, Fe, Mn and Na. The beam diameter was set at 1 μm for all phases except for mica and plagioclase which were analysed with a 3–5 μm beam diameter. For standardization, reference materials from MAC™ (Micro-analysis Consultants Ltd, UK) were used.

Isotope analyses were carried out at the Institut für Mineralogie, Universität Münster. For sample preparation, whole rocks were crushed in a steel mortar or using a jaw-breaker. Whole rock powders were prepared in a tungsten carbide mill. For mineral separation, crushed material was reduced in size either by grinding for only a few seconds in a tungsten carbide mill or by use of a disc mill. Following sieving, fines were removed and minerals were enriched by use of a Frantz magnetic separator and, in case of phengite, by adherence to a sheet of paper. After hand-picking, optically pure mineral concentrates were washed in ethanol (p.a.) in an ultrasonic bath, repeatedly rinsed in H_2O (three times distilled) and dried at *c.* 40–50 °C in an oven overnight.

Whole-rock powders (*c.* 100 mg) and mineral separates (phengite: *c.* < 1–36 mg; plagioclase: *c.* 6–23 mg; epidote: *c.* 0.3 mg; glaucophane: *c.* 15–28 mg) were mixed with a ^{87}Rb–^{84}Sr spike in teflon screw-top vials and dissolved in a HF–HNO_3 (5:1) mixture on a hot plate overnight. After evaporation and drying, 6N HCl was added to the residue. This mixture was again homogenized on a hot plate overnight. After a second evaporation to dryness, Rb and Sr were separated by standard ion-exchange procedures (AG 50W-X8 resin) on quartz glass columns using 2.5 N and 6 N HCl as eluents. Rb was loaded with H_2O on Ta filaments; Sr was loaded with TaF_5 on W filaments. Calcite (*c.* 5 mg) was dissolved in 2.5 N HCl. Mass-spectrometric analysis was carried out using a VG Sector 54 multicollector mass spectrometer (Sr) and a NBS-type Teledyne mass spectrometer (Rb). Correction for mass fractionation is based on a $^{86}Sr/^{88}Sr$ ratio of 0.1194. Rb ratios were corrected for mass fractionation using a factor deduced from multiple measurements of Rb standard NBS 607. Total procedural blanks were less than 0.1 ng (mostly < 0.05 ng) for Rb and less than 0.24 ng (mostly < 0.1 ng) for Sr. Based on repeated measurements, the $^{87}Rb/^{86}Sr$ ratios were assigned an uncertainty of 1% (2σ). The uncertainties of the $^{87}Sr/^{86}Sr$ ratios are reported at the $2\sigma_m$ level. In the course of this study, repeated runs of NBS standard 987 gave an average $^{87}Sr/^{86}Sr$ ratio of 0.710316 ± 0.000028 (2σ, n = 16). All ages and elemental concentrations were calculated using the IUGS recommended decay constants (Steiger & Jäger, 1977) using the least squares regression technique of York (1969), considering only scatter but no Student-t multiplier. Rb–Sr isochrons were plotted using Isoplot/Ex (Ludwig, 2001).

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