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The Arosa zone in Eastern Switzerland: oceanic, sedimentary burial, accretional and orogenic very low- to low grade patterns in a tectono-metamorphic mélange

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Abstract In the area of Arosa–Davos–Klosters (Eastern Switzerland) the different tectonic elements of the Arosa zone mélange e.g. the Austroalpine fragments, the sedimentary cover of South Penninic ophiolite fragments, as well as the matrix (oceanic sediments and flysch rocks) show distinctively different metamorphic histories and also different climaxes ("peaks") of Alpine metamorphism. This is shown by a wealth of Kübler-Index, vitrinite and bituminite reflectance measurements, and K-white mica *b* cell dimension data. At least six main metamorphic events can be recognized in the area of Arosa–Davos–Klosters: (1) A pre-orogenic event, typical for the Upper Austroalpine and for instance found in the sediments at the base of the Silvretta nappe but also in some tectonic fragments of the Arosa zone (Arosa zone mélange). (2) An epizonal oceanic metamorphism observed in the close vicinity of oceanic basement rocks units of the Arosa zone (South Penninic) is another pre-orogenic process. (3) A metamorphic overprint of the adjacent Lower Austroalpine nappes and structural fragments of the Lower Austroalpine in the Arosa zone. This metamorphic overprint is attributed to the orogenic metamorphic processes during the Late Cretaceous. (4) A thermal climax observed in the South Penninic sediments of the Arosa zone can be bracketed by the Austroalpine Late Cretaceous event (3) and the middle Tertiary event (5) in the Middle Penninic units and predates Oligocene extension of the "Turba phase". (6) North of Klosters, in the northern part of our study area, the entire tectonic pile from the North Penninic flysches to the Upper Austroalpine is strongly influenced by a late Tertiary high-grade diagenetic to low-anchizone event. In the Arosa zone mélange an individual orogenic metamorphic event is evidenced and gives a chance to resolve diagenetic–metamorphic relations versus deformation. Six heating episodes in sedimentary rocks and seven deformation cycles can be distinguished. This is well explained by the propagation of the Alpine deformation front onto the foreland units. Flysches at the hanging wall of the mélange zone in the north of the study area (Walsertal zone) show data typical for low-grade diagenetic thermal conditions and are therefore sandwiched between higher metamorphic rock units and separated from theses units by a disconformity. The Arosa zone s.s., as defined in this paper, is characterised by metamorphic inversions in the hanging wall and at the footwall thrust, thus shows differences to the Walsertal zone in the north and to the Platta nappe in the south.

Keywords Penninic–Austroalpine boundary · Diagenesis to low-grade metamorphism · Kübler-Index · Vitrinite reflectance · Bituminite reflectance · Tectono-metamorphic mélange

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1 Introduction and scope of the project

In the Arosa zone the diagenetic to metamorphic pattern is very complex and mosaic-like (Ferreiro Mählmann 1994;

Frey and Ferreiro Mählmann 1999). This tessellated pattern is the result of different, partially overlapping heating events (pluri-phase heating history, "plurifacies metamorphism" according to De Roever and Nijhuis 1963) and tectonic processes. In previous studies (Ring 1989; Dürr 1992; Frisch et al. 1994; Bevins et al. 1997) it was not possible to give a complete and absolute satisfying interpretation to explain this complex pattern in the Arosa zone. It was concluded that the ophiolithic rocks show a thermal history differing from that of the sediments and metasediments (Bevins et al. 1997). Finishing the compilations for metamorphic mapping it was striking that high diagenetic to low anchizone areas are found in the Arosa zone far to the south (Frey et al. 1999; Frey and Ferreiro Mählmann 1999; Oberhänsli et al. 2004). This very low-grade metamorphic pattern is differing from Cretaceous metamorphic epizone rocks in the hanging wall (120–70 Ma) and epizone rocks in the footwall metamorphosed in Tertiary time (45–20 Ma). It was pointed out that neither an Eoalpine nor a Neoalpine metamorphism could be the reason for this Arosa zone pattern. The discussion remained open.

The multi-method approach of this work, which is furthermore extended to almost all the lithologies of the study area, allows to draw a more refined metamorphic map and to overcome the earlier contradictions (cf. discussion in Ferreiro Mählmann 1994). In the last years the knowledge about the Alpine metamorphic history (Cretaceous to Tertiary) of the Austroalpine and Penninic units underwent important revisions, namely by the application of multi-method low temperature studies and petrologic metamorphic mapping in Eastern Switzerland (Grisons state) and Austria (Vorarlberg state) (Ferreiro Mählmann 1994, 1995, 2001; Ferreiro Mählmann and Petschik 1996; Weh et al. 1996; Weh 1998; Frey and Ferreiro Mählmann 1999; Petrova et al. 2002; Ferreiro Mählmann et al. 2002; Wiederkehr et al. 2011).

The study area is situated in Eastern Switzerland and extends from the Oberhalbstein in the South to the region of Arosa, Davos, Klosters and finally to the Swiss–Austrian boundary near the Sulzfluh in the North. Thus the complex tectonic edifice at the Penninic–Austroalpine boundary was intensely studied at a length of about 50 km (cf. Fig. 1 and inset). The aims of this work are:

1. To compile the former and new data obtained by different methods to draw a new metamorphic map for the sedimentary rocks of the Arosa zone. The new metamorphic pattern found will complete the metamorphic mappings of the adjacent units. The new data presented here will also be compared with the recent metamorphic maps of Frey et al. (1999) and Oberhänsli et al. (2004).
2. To show the relations between metamorphic grade, the structural tectonic level ("Stockwerktektonik") and the paleo-geographic origin.
3. To compare the data of diagenetic to incipient metamorphic sediments in the Arosa zone with the data of the adjacent crystalline units and their cover rocks.
4. To compile information about the timing of the main metamorphic events.
5. To fill up the gap of the missing interpretation of the very low-grade metamorphic sandwich in the Arosa zone between the low-grade metamorphic rocks of the Austroalpine and the low-grade metamorphic rocks of the Middle to North Penninic units presented in the review studies of Frey et al. (1999) and Oberhänsli et al. (2004).
6. To give a new definition of the structural limits of the Arosa zone.

2 Tectonic and metamorphic setting

An overview of the tectonic situation in the study area is shown in Fig. 1. In the northern part of this area North Penninic flysch (Bündnerschists, "Bündnerschiefer") units extend farther to the east into the half window of the Prättigau. These units are again exposed in the tectonic window of Gargellen. The flysches are overlain by Middle Penninic nappes (Falknis and Sulzfluh nappe) and by the South Penninic Arosa zone. The Arosa zone contains fragments of an oceanic crust but also slices of a continental crust with a paleo-geographic and tectonic origin from the Apulian–Austroalpine plate (Austroalpine nappes). Thus, the Arosa zone as a whole (see Fig. 1) has to be regarded as a large tectonic mélange. The Arosa zone includes ophiolites (Steinmann 1905) of the subducted South Penninic ocean, the Piemont-Ligurian ocean (Frisch 1981; Weissert and Bernoulli 1985). The Davos-Dorfberg nappe and the Tschirpen unit (Cadisch 1921; Streckeisen 1948) above the Arosa zone s.s. are attributed to be Lower Austroalpine (this two units are also referred as Tschirpen-Dorfberg nappe in simplified tectonical schemas). The highest structural units in this area are the Upper Austroalpine units. The Silvretta nappe s.l. according to Eichenberger (1986) consists of three tectonic units: the Silvretta nappe s.s., the Arosa Dolomite nappe and the Rothorn nappe. The Silvretta nappe s.s. is mainly a basement nappe. The sediments of the Silvretta nappe s.s. form two synforms at the south-western edge of this unit (Fig. 1). At Schafläger (Davos) and in the region of the Arosa Rothorn ("Rothorn nappe") Silvretta basement slices also lie beneath the Arosa Dolomites (Silvretta nappe

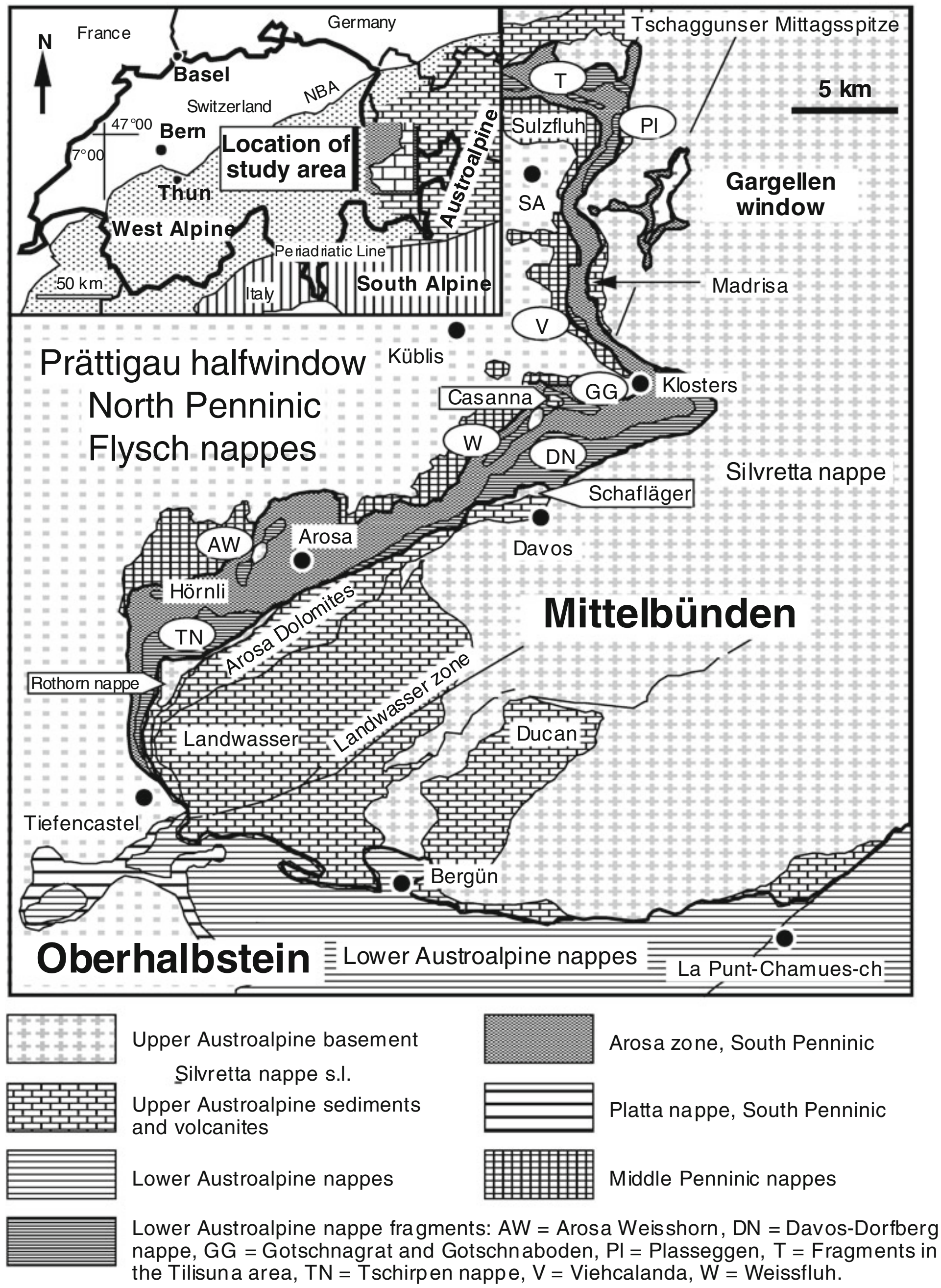

Fig. 1 Tectonic sketch map and location of the study area. *NBA* Northern boundary of the Alps, *SA* Sankt Antönien. The sedimentary cover of the Tschaggunser Mittagspitze mountain peak is part of the Rätikon mountain range and belongs to the Silvretta nappe s.s. and was thrusted to the NW on the Lechtal nappe of the Northern Calcareous Alps (Ferreiro Mählmann 1994)

s.l.). In the north of the study area other tectonic slices with basement and mesozoic rocks are located at the Penninic–Austroalpine boundary and below the Silvretta nappe s.l., namely the Madrisa zone and the Tschaggunser Mittags-spitz unit (localities see Fig. 1). A tectonic outlier (“Klippe”) of the Silvretta nappe s.l. is also located at this boundary near the Casanna (Klosters), described by Haldimann (1975) and in a very detailed review paper by Froitzheim et al. (1994).

The Arosa zone forms the basal structural unit of the Austroalpine nappes and belongs to the South Penninic domain (Trümpy 1980). In the foot-wall of the Arosa zone, metamorphism and its relation to deformation of the North and Middle Penninic nappes was studied by Thum and Nabholz (1972); Gruner (1980); Frey et al. (1980); Weh et al. (1996); Petrova et al. (2002) and Ferreiro Mählmann et al. (2002). The lowest stratigraphic units of the North Penninic flysches are overprinted by an epizonal orogenic metamorphism (sensu Miyashiro 1973) of Mid to Late Tertiary age (Thum and Nabholz 1972). The low anchi-zonal metamorphism of the Ruchberg sandstone, sedimented in Palaeocene or earliest Eocene and thus the youngest formation in this area, is related to a metamorphic overprint of the entire Prättigau area, showing an increase of metamorphic grade from the stratigraphic and structural top to the bottom (Ferreiro Mählmann 1994; Frey and Ferreiro Mählmann 1999). In the North Penninic tectonic units of the area between Sankt Anthönien and Tiefencastel (Fig. 1), no important metamorphic discontinuity of the late (Oligocene) regional metamorphic overprint was observed (Weh et al. 1996; Petrova et al. 2002; Rahn et al. 2002; Wiederkehr et al. 2011). This is however not the case in the area of Chur and to the south (Petrova et al. 2002; Ferreiro Mählmann et al. 2002).

Deformation causes a folding of the vitrinite reflectance (VR) isolines and Kübler-Index (KI)–isocrysts (Ferreiro Mählmann 1995; Petrova et al. 2002), and thus also the sub-greenschist and greenschist facies zone map pattern with the chloritoid-in isograd (sensu Winkler 1979; Bucher and Frey 1994), as shown in the “Alpine Metamorphic Map” by Frey et al. (1999). The Tertiary metamorphic patterns mapped by Frey and Wieland (1975) and Frey and Ferreiro Mählmann (1999) found in the Penninic domain is discontinuously cut at the tectonic limit to the Arosa zone (see also Figs. 6, 9, 10).

In the Arosa zone, the meta-basalts (tholeiitic basalts), phyric basalts, coarser dolerites (diabases), pillows with an intergranular to subophitic texture, basaltic pyroclastites, MORB-type rocks (Trommsdorff and Dietrich 1980; Burkhard 1987; Frisch et al. 1994) are thoroughly altered and therefore defined as spilites (Grunau 1947; Richter 1957) sensu Steinmann (1905). The traditional models postulated a steady north to south increase of metamorphic grade along the Austroalpine–Penninic boundary (Jäger et al. 1961; Frey et al. 1974; Trommsdorff and Dietrich 1980; Ring 1989; Ring et al. 1989; Frisch et al. 1994; Bevins et al. 1997). Most of these models are based on studies of ophiolites and related rocks (e.g. ophicalcites); some work was also done in the South Penninic sediment pile (stratigraphy see Fig. 2). Data for the Arosa zone was mainly published by Frey et al. (1974); Oterdoom (1978); Gruner (1980); Lüdin (1987); Weissert and Bernoulli (1984); Burkhard (1987); Ring et al. (1988) and Henrichs (1993). More recent and extensive studies on vitrinite reflectance (VR) and illite “crystallinity” (IC), in the aftermath referred as Kübler-Index (KI), were done by Ferreiro Mählman (1994, 1995, 1996, 2001) showing a more complex metamorphic pattern in the rocks of the Arosa zone and Platta nappe, units from the South Penninic domain.

The Davos-Dorfberg nappe (Figs. 1, 2), a Lower Austroalpine basement unit, contains meta-gabbros, meta-aplites, meta-pegmatites, amphibolites and meta-sediments. The petrography and also the geochronology of Davos-Dorfberg nappe are different from the adjacent Upper Austroalpine Silvretta basement near Davos, e.g. Schafläger crystalline unit (cf. Streckeisen 1948; Streckeisen et al. 1966; Giger 1985, 1986).

In the hanging wall of the Arosa zone, after the pioneer work of Dunoyer De Segonzac and Bernoulli (1976), a modern and accurate low-grade metamorphic research was carried out by Henrichs (1993) and Kürmann (1993). At the same time the area was extensively studied with the KI and VR methods by Ferreiro Mählman (1994, 1995, 1996, 2001). The tectonics, lithologies and stratigraphy of the Austroalpine sediments are shown in Fig. 2. The basic information for Figs. 1 and 2 is derived from Streckeisen et al. (1966); Tollmann (1977); Trümpy (1980); Eichenberger (1986) and Furrer et al. (1992).

Froitzheim et al. (1994) presented a tectonic evolution model for the Austroalpine units of Eastern Switzerland. These authors distinguish the following main deformation phases affecting the tectonic units at the Penninic–Austroalpine boundary:

- D1 Late Cretaceous “Trupchun phase” with a considerable westward movement of the Upper Austroalpine. D1 causes a metamorphic inversion of a M1 burial pre-orogenic diagenesis to epizone metamorphic pattern (Ferreiro Mählmann 1994, 1995);
- D2 Late Cretaceous “Ducan/Ela phase”: an extensional event in the Austroalpine nappes. D1 and D2 resulted in an orogenic metamorphism (M2) forming a post-kinematic (nappe and fold tectonic) high diagenesis to epizone pattern (Ferreiro Mählmann 2001);
- D3 Eocene “Blaisun phase” with a considerable northward movement in the Austroalpine nappes and top to north crustal shortening in the Penninic units, partly also in the Austroalpine;
- D4 Oligocene “Turba phase”: an extensional phase related to movements along the Turba mylonite zone (TMZ), the Insubric and the Engadine line (D2 according to Weh 1998; “Niemet–Beverin-phase”);
- D5 Oligocene “Domleschg phase”, a late compressional overprint affecting mainly the North Penninic flysches (“Bündnerschiefer”) in the study area (D 3a and D 3b according to Weh 1998). The Tertiary orogenic metamorphic event was referred in the chronological order as M3. In this paper the event will be renumbered as M4.

3 Sampling and methodology

3.1 Sampling

The synthesis presented in this publication includes information of sedimentary rocks from about 380 different

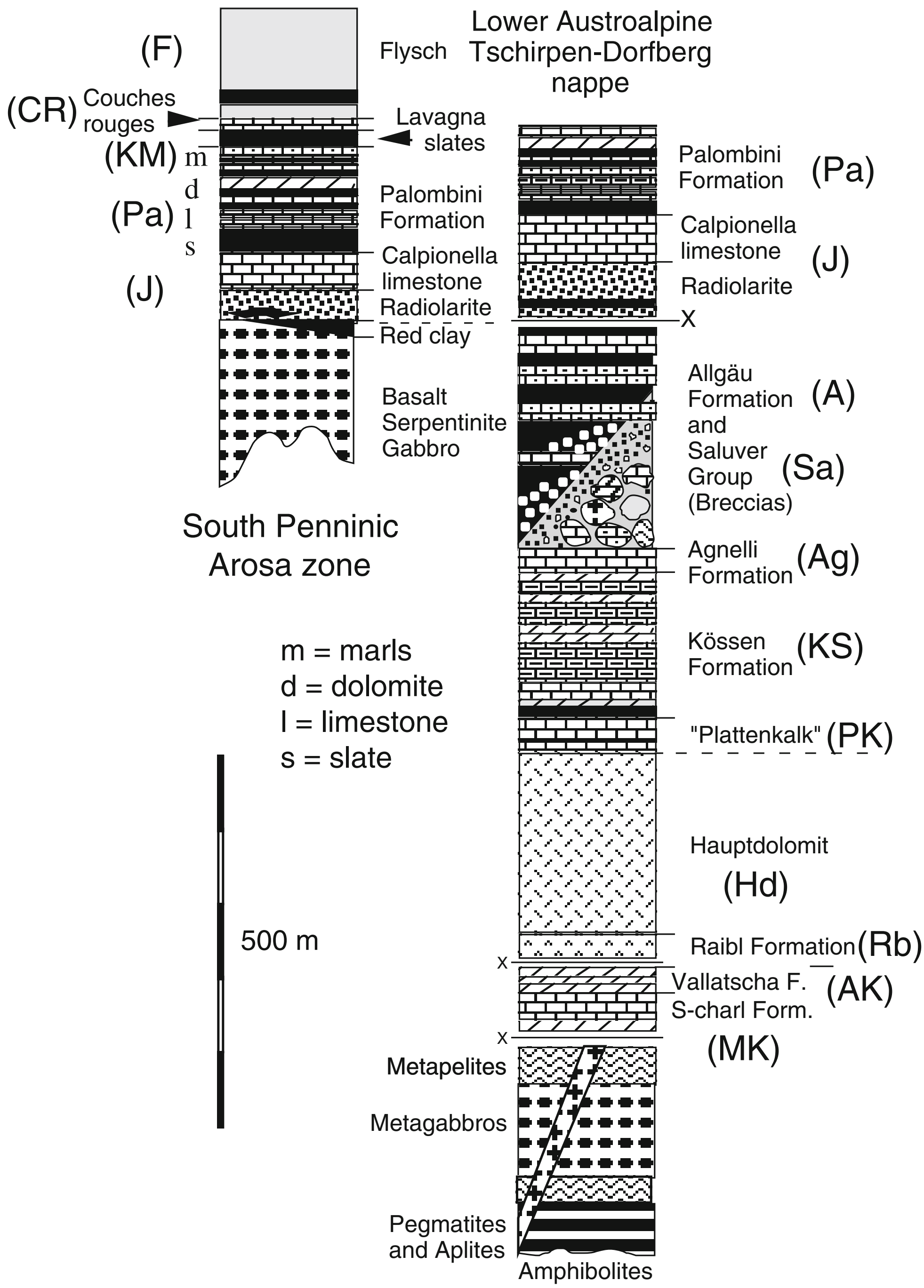

Fig. 2 Lithostratigraphic sections of the Arosa zone and the Tschirpen-Dorfberg nappe. The symbols are according to the Shell legend or labelled as shown on the figure. The abbreviations are used for the sample numbers (e.g. F for Flysch: F 27 to F 88)

localities (see Fig. 3; Ferreiro Mählmann 2001). Including the KI–VR samples from the hanging wall and footwall of the Arosa zone a total of 750 KI values and 400 VR values are used for this study (this set includes data from diploma thesis works from the Frey group at Basel University). From 350 localities a KI–VR data pair is available. 110 samples are from the Arosa zone (Table 1).

Nearly all sediment formations (cf. Fig. 2) in the Arosa zone contain materials that are suitable for XRD-CMC (clay mineral composition), VR and KI measurements (for more details see Ferreiro Mählmann 1995). The samples and the corresponding KI and VR data are listed in Table 1 (only the samples from the Arosa zone are shown).

3.2 Sample preparation

Thin sections were used for microscopic petrography. Polished resin mounted sections were prepared for VR and BR measurements as well as maceral analysis. XRD-CMC and XRD-WRC (whole rock composition) were used to get

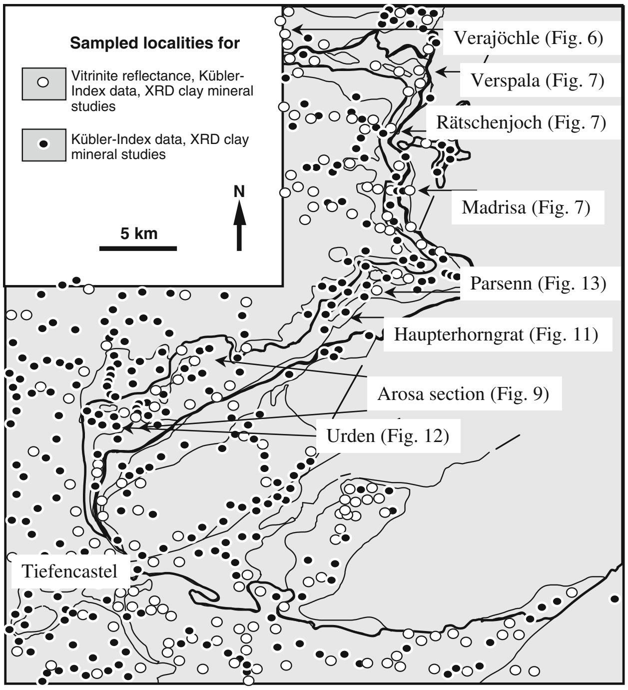

Fig. 3 Distribution of the sample localities and the methods applied, also shown are some locations of profiles and sections

preliminary information about diagenetic or incipient metamorphic grade. Sample preparation is done following the standardisation of Ferreiro Mählmann and Frey (2012). This strategy allowed the simultaneous analysis of both the inorganic and organic metamorphic parameters on the same rock specimen.

3.3 Coal petrology

Resin-mounted particulate samples where polished in sections perpendicular and in some cases, to recognize pre-graphitic structures, also parallel to bedding and slaty cleavage. In average, it was attempted to get 50 single VR values in one sample. Measurements with less than 15 measuring points (or with a high standard deviation >10 %) are not used for correlations with other methods. Maximum (%R_{max}) and minimum reflectance were analysed following the recommendations of Mackowsky (1982). The rank terms are based on the DIN-classification (cf. Stach et al. 1982: 70). VR is a parameter sensitive to temperature in the range from 0.25 up to 8.0 %R_{max} (Ferreiro Mählmann 1995, 2001; Koch 1997). The sensitivity decreases with the increasing rank. In the studied area VR ranges from 2.5 to 7.5 %R_{max}. For more detailed information on measuring procedure and evaluation of VR analysis see Petschick (1989) and Ferreiro Mählmann (2001).

Maximum and minimum reflectance was also measured for large and anisotropic bitumen fragments and vein fillings (see Ferreiro Mählmann 2001; Ferreiro Mählmann and Frey 2012). Bituminite reflectance is not shown graphically in most maps and diagrams, but mentioned in the text, because these data give valuable information about metamorphism in rocks without vitrinite phytoclasts (Jacob 1989; Ewbank et al. 1995; Koch 1997; Ciulavu et al. 2008). Bituminite reflectance (BR_{max}) correlates with VR in a 1:1 relation between a value of 2.5 and 4.5 % (Ferreiro Mählmann and Frey 2012), i.e. in the range of most Arosa zone data.

Measurements were carried out only on solid bituminite with a homogeneous bi-reflection (anisotropy not exceeding 3.0 %) without pre-graphitic or sphaerolitic-granular or undulating structures (Ferreiro Mählmann 2001; Ferreiro Mählmann and Frey 2012).

3.4 Clay mineral petrography

The samples were partly used for whole rock powder XRD studies. Only in very few cases diagnostic minerals were detected by this method. For the diffractometric determination of mineral phases the JCPDS tables were applied.

Using the X-ray identification guide by Carrol (1970); Moore and Reynolds (1989) and the table of key lines by Chen (1977) clay mineral data was interpreted. The symbols for rock-forming minerals used in this work are according to Bucher and Frey (1994: 309f).

Three specimens of every sample from 482 analysed localities were measured to determine KI (Fig. 3); therefore 1,646 KI data are available and additional 270 IC measurements from literature and unpublished diploma thesis (Ferreiro Mählmann 1994; cum biblio). The IC (Kübler 1967) is defined as the full width at half maximum intensity (FWHM) of the illite-white K mica first (10-Å) reflection of slides with a textured sedimented cover and normalized to a fraction of an angle in $\Delta°2\theta$ denominated as Kübler-Index (KI). All clay mineral analyses were carried out on the fraction <2.0 µm. The values limiting the anchizone are 0.42 $\Delta°2\theta$ and 0.25 $\Delta°2\theta$ for air-dried specimens. Sample preparation and crystallinity measurement are in accordance with the procedure described by Krumm (1984); Krumm et al. (1988) and the recommendations of Kisch (1991), also summarised by Ferreiro Mählmann and Frey (2012).

For preliminary pressure estimation the clay mineral illite/k-white mica *b* method was used with the relation $b = 6 \times d(060, 331)$. See also Fig. 14 and Table 1.

3.5 Calibration and standards used for measurements of the illite 10 Å reflection

IC data compiled from literature is treated as follows: data referring to the IC is corrected to the IC standards

Table 1 Data from the Arosa zone

Sample	Rocktype	SC a	SC l	m NN	Illite 10 Å	KI dry	KI glyc	b illite	VR R_{max} %	BR R_{max} %	n VR	s VR	n BR	s BR
F 41	Siltsone siliceous	212,845	777,820	2,330	–	–	–	–	2.47	2.47	4	0.21	26	0.30
Pa 28	Bituminitic marl	212,800	777,390	2,322	94	0.37	0.33	–	2.84	–	47	0.29	–	–
F 42a	Limestone	212,680	777,820	2,333	61	0.40	0.19	–	2.85	–	7	0.32	–	–
F 42b	Siliceous sandstone	212,680	777,820	2,333	–	–	–	–	2.80	–	5	0.25	–	–
Pa 29	Silty shale	212,440	778,530	2,175	107	0.46	0.44	–	–	–	–	–	–	–
A 195	Limestone tectonised	212,320	778,650	2,200	205	0.34	0.32	–	–	–	–	–	–	–
F 44	Siliceous sandstone	212,220	775,940	2,241	–	–	–	–	2.83	–	16	0.29	–	–
MK 55	Limestone	212,215	774,840	2,276	–	–	–	–	3.00	–	8	0.15	–	–
F 46	Turbidite sandstone	212,140	775,495	2,292	–	–	–	–	3.08	–	26	0.12	–	–
F 46bit	Turbidite sandstone	212,140	775,495	2,292	–	–	–	–	3.11	3.01	12	0.17	8	0.25
KM 33	Shale	212,075	775,650	2,220	113	0.65	0.52	–	–	–	–	–	–	–
J 32	Shale	211,985	776,740	2,240	110	0.49	0.43	–	3.20	–	21	0.39	–	–
F 29	Slitstone	211,310	784,570	2,440	–	–	–	–	2.50	–	2	0.40	–	–
F 39	Marl turbiditic	211,090	784,695	2,075	143	0.33	0.30	–	–	–	–	–	–	–
F 38	Silty limestone	210,925	784,220	2,322	–	–	–	–	2.75	–	12	0.15	–	–
F 28	Fine sandstone	210,810	785,975	2,290	–	–	–	–	3.20	3.15	77	0.26	14	0.23
MK 52	Limestone	210,665	785,505	2,196	203	0.35	0.31	–	2.83	–	8	0.23	–	–
F 40	Carbonate sandstone	210,475	786,700	1,971	240	0.30	0.27	–	–	–	–	–	–	–
F 37	Lity shale	209,545	781,700	2,332	45	0.01	0.95	–	–	–	–	–	–	–
A 162	Marl	208,910	785,915	2,290	–	–	–	–	3.17	–	3	0.40	–	–
Plas.phyl.	Phyllite	208,825	786,235	2,290	311	0.13	0.13	–	4.56	–	16	0.22	–	–
KS 130	Limestone tectonised	207,635	785,680	2,055	45	0.37	0.30	–	–	–	–	–	–	–
Pa 27	Shale tectonised	207,360	785,560	2,062	166	0.30	0.29	–	2.87	2.96	45	0.20	17	0.26
Pa 22	Shale	206,720	785,510	2,130	182	0.38	0.37	9,016	–	–	–	–	–	–
KM 44	Shale tectonised	204,715	787,975	1,580	16	0.63	0.65	–	2.57	2.54	35	0.21	10	0.23
J 49	Silty marl	204,715	788,000	1,580	11	0.43	0.43	–	–	–	–	–	–	–
J 9	Marl	200,870	784,415	2,640	–	–	–	–	2.94	–	12	0.24	–	–
KM 34	Shale	200,850	784,175	2,600	90	0.28	0.27	–	–	–	–	–	–	–
J 33	Marl	200,835	784,595	2,650	240	0.30	0.29	9,022	3.28	–	5	0.46	–	–
N 28	Shale	200,825	784,620	2,670	332	0.28	0.26	–	–	–	–	–	–	–
J 37a	Siltiger Tonstein	200,350	785,020	2,555	96	0.72	0.38	–	–	–	–	–	–	–
KM 37	Marl	198,415	784,710	2,110	200	0.37	0.24	–	–	–	–	–	–	–
KS 108	Marl	192,480	783,340	2,280	134	0.30	0.29	–	3.62	–	18	0.43	–	–
A 165	Marl	192,240	782,590	2,310	167	0.27	0.28	–	–	–	–	–	–	–
J 34	Shale	191,855	782,300	2,120	136	0.32	0.29	8,989	7.03	–	14	0.91	–	–
J 35	Shale	191,845	782,300	2,120	190	0.25	0.28	–	–	–	–	–	–	–
J 36	Limestone	191,475	782,260	2,132	284	0.29	0.27	–	3.21	–	8	0.30	–	–
Pa 55	Black shale	191,475	782,261	2,135	275	0.36	0.34	9,020	–	–	–	–	–	–
Pa 56	Shale	191,475	782,261	2,135	314	0.34	0.32	–	–	–	–	–	–	–
Pa 57	Silty shale	191,475	782,261	2,137	233	0.34	0.30	–	3.18	–	50	0.27	–	–
A 209	Marly limestone	188,755	779,410	2,515	116	0.28	0.27	–	3.2	3.25	24	0.21	21	0.19
F 80	Marly flysch	188,755	779,410	2,514	157	0.27	0.27	–	–	–	–	–	–	–
KS 138	Marl	188,755	779,411	2,513	–	–	–	–	3.72	–	44	0.20	–	–
Pa 51	Black shale	188,755	779,411	2,513	191	0.28	0.26	9,015	–	–	–	–	–	–
F 81	Marly flysch	188,755	779,412	2,512	351	0.28	0.27	–	–	–	–	–	–	–
F 82	Sandstone	188,754	779,413	2,510	68	0.27	0.25	–	–	–	–	–	–	–
F 82b	Marly siltstone	188,754	779,413	2,510	125	0.27	0.27	–	3.20	3.08	66	0.27	27	0.32
F 83	Silty shale	188,750	779,417	2,500	94	0.25	0.24	–	3.15	3.30	31	0.22	11	0.26
F 84	Silty marl	188,210	779,605	2,360	168	0.32	0.29	–	–	–	–	–	–	–
F 85	Shale claystone	188,210	779,605	2,360	237	0.35	0.31	9,014	–	–	–	–	–	–

Table 1 continued

Sample	Rocktype	SC a	SC l	m NN	Illite 10 Å	KI dry	KI glyc	b illite	VR R_{max} %	BR R_{max} %	n VR	s VR	n BR	s BR
F 86	Silty shale	188,210	779,605	2,360	184	0.35	0.30	–	–	–	–	–	–	–
F 87	Sandstone	188,210	779,605	2,360	122	0.32	0.31	–	–	–	–	–	–	–
F 88	Black shale	188,210	779,605	2,360	305	0.42	0.34	–	–	–	–	–	–	–
Pa 52	Silty shale	188,210	779,605	2,359	192	0.28	0.27	–	3.20	3.18	65	0.24	24	0.33
Pa 53	Shale	188,210	779,605	2,359	85	0.30	0.29	9,011	–	–	–	–	–	–
Pa 54	Silty shale	188,210	779,605	2,359	114	0.27	0.25	–	–	–	–	–	–	–
J 101	Limestone	188,210	779,605	2,357	224	0.27	0.27	–	–	–	–	–	–	–
J 102	Radiolarite	188,210	779,605	2,357	56	0.27	0.28		3.78	–	32	0.28	–	–
J 107	Radiolarite red clay	185,000	770,195	2,122	173	0.27	0.25	8,997	–	–	–	–	–	–
KM 48a	Shale	184,905	771,040	1,860	187	0.27	0.26	9,013	4.02	–	18	0.44	–	–
KM 48b	Silty shale	184,905	771,040	1,860	91	0.38	0.30	–	–	–	–	–	–	–
MK 71a	Marly limestone	184,905	771,045	1,860	85	0.27	0.27	–	–	4.10	–	–	11	0.19
MK 71b	Marl	184,905	771,045	1,860	149	0.27	0.25	–	–	–	–	–	–	–
KM 49b	Silty shale	184,820	769,220	2,440	170	0.28	0.27	–	3.40	3.44	34	0.27	9	0.33
KM 49a	Shale	184,820	769,220	2,440	249	0.28	0.26	–	–	–	–	–	–	–
Hd 136	bituminitic limestone	184,390	768,582	2,286	55	0.24	0.24	–	–	4.81	–	–	38	0.39
KM 46	Shale	184,040	767,570	2,415	108	0.28	0.25	–	3.38	–	17	0.30	–	–
Hd 127	Shale tectonised	183,480	767,600	2,410	33	0.24	–	–	–	–	–	–	–	–
J 54	Limestone	183,195	767,680	2,367	206	0.29	0.27	–	–	–	–	–	–	–
J 63	Limestone tectonised	182,995	771,680	1,644	105	0.22	0.20	–	–	–	–	–	–	–
J 52	Limestone	182,800	766,770	2,480	97	0.20	0.19	–	–	–	–	–	–	–
J 53	Silty marl	182,555	766,825	2,430	106	0.23	0.22		3.73	–	11	0.25	–	–
Pa 33	Limestone tectonised	182,525	766,360	2,240	308	0.28	0.25	–	–	–	–	–	–	–
Pa 38b	Silty marl	182,170	771,610	1,640	138	0.22	0.20	–	3.28	–	3	0.21	–	–
Pa 38a	Limestone	182,170	771,610	1,640	–	–	–	–	3.36	–	13	0.23	–	–
F 45	Shale	182,035	766,505	2,400	32	0.29	–	–	–	–	–	–	–	–
Pa 31	Shale	182,000	766,600	2,470	27	0.26	0.25	–	–	–	–	–	–	–
KM 45b	Silty marl	181,990	766,360	2,490	514	0.16	0.14	–	5.53	5.86	19	0.43	30	0.51
J 38	Marl	181,920	766,660	2,520	118	0.23	0.23	–	–	–	–	–	–	–
Pa 30	Marl	181,920	776,625	2,500	201	0.25	0.24	–	–	–	–	–	–	–
F 30	limestone	181,860	766,600	2,490	–	–	–	–	3.32	–	14	0.32	–	–
F 31	Carbonate andstone	181,840	766,660	2,510	–	–	–	–	3.94	–	10	0.35	–	–
Pa 32	Silty marl	181,815	765,875	2,350	–	–	–	–	3.94	–	8	0.26	–	–
KM 45a	Silty limestone	181,685	766,680	2,450	236	0.24	0.23	–	3.60	–	6	0.19	–	–
Hd 126	Bituminitic marl	181,525	766,560	2,565	350	0.13	0.13	–	5.75	5.88	7	0.33	27	0.27
J 103	Radiolarite	181,205	767,510	2,365	75	0.20	0.21	–	–	–	–	–	–	–
J 104	Red clay	181,205	767,510	2,362	185	0.20	0.19	–	4.69	4.89	22	0.15	13	0.19
J 105	Sulfidic caly	181,205	767,515	2,358	234	0.22	0.21	9,004	–	–	–	–	–	–
KM 46	Marl	181,205	767,515	2,356	72	0.23	0.21	–	–	–	–	–	–	–
J 106	Radiolarite	108,205	767,516	2,355	166	0.20	0.18	–	–	–	–	–	–	–
KM 47	Slate	108,050	767,509	2,385	313	0.31	0.32	–	3.33	3.40	54	0.28	11	0.36
F 32	Turbidite sandstone	181,050	768,600	2,010	–	–	–	–	3.76	3.71	19	0.34	5	0.35
AS 39	Dolomite	177,130	762,740	1,575	162	0.28	0.25	–	–	–	–	–	–	–
Hd 131	Dolomite	171,170	766,080	1,135	284	0.18	0.18	–	5.50	–	13	0.31	–	–

SC a and *SC l* Swiss Coordinates altitude and longitude, *mNN* elevation in meter, *Illite 10 Å* intensity in counts per second of the illite 10 Å peak, *KI dry* Kübler-Index air dry, *KI glyc* Kübler-Index after glycol treatment, *b illite* K-white mica *b* cell dimension in Å, *VR* vitrinite reflectance, *BR* bituminite reflectance

(Kübler–Frey–Kisch and Frerreiro Mählmann standards) used at the University of Basel and expressed as KI values (Ferreiro Mählmann and Frey 2012). Also inter-laboratory calibrations were established among B. Kübler, M. Frey, H. Krumm, R. Petschick, A. Stahel and R. Ferreiro Mählmann (Ferreiro Mählmann and Frey 2012). Nearly all measurements from the Grisons determined in Basel, Bern, ETH Zürich, Frankfurt, and TU Darmstadt and from Thum and Nabholz (1972) can be re-calculated as KI values and fit perfectly with data from own samples for the same localities.

Data from the Universities of Tübingen (U. Ring) and Bochum (D.K. Richter, e.g. Henrichs 1993; Kürmann 1993) are not used for metamorphic mapping because a calibration of this IC data is missing or difficult to correlate (Ferreiro Mählmann and Frey 2012). Concerning the data of Ring (1989) no correlation to other institutes was found. The resulting values are irregularly high and scattering (r = 0.62; Ferreiro Mählmann 1994). For more details see also Ferreiro Mählmann and Frey (2012).

3.6 KI in function of lithologies

This chapter deals with KI variations dependent from the rock type and lithostratigraphy (petrovariations). Despite using the routine of the Frankfurt, Darmstadt and Basel laboratories (Ferreiro Mählmann and Frey 2012) some samples showed hkl lines. Evidently these samples have higher quartz and feldspar contents. In a textured sample (in a sample mount with perfect preferred orientation, the ideal case) no hkl lines should be found (Krumm and Buggisch 1991). The occurrence of hkl reflections indicates a not ordered layering of the minerals with different grain habitus. Also a 0.05 Δ°2θ broadening of KI (mean) is observed respective to a neighbouring pelite. These samples were not used for further studies (correlations).

In the Arosa zone another problem is frequent. Carbonate rocks show a smaller spread of KI values in the anchizone and high diagenetic zone. This is more prominent in limestones than in marls (Fig. 4). The graph presents 183 samples from localities where a pelite and marly limestone or marl was recovered. Between 0.25 and 0.65 Δ°2θ the correction factor of 0.055was used (Ferreiro Mählmann 1994). This strongly homogenised the data and the variance of values decreased. This correlation and KI re-calculation is probably not universally applicable and is restricted to the pressure–temperature–chemical–fluid conditions found in the Arosa zone. Due to that in the Arosa zone near Arosa a top down trend of decreasing KI was recognized.

Most lithologies of the South Penninic realm, except the rocks from the Jurassic, show a more or less important influence by detrital mica. Detrital mica-fractions strongly increase the 10 Å basal reflections, lowering the sharpness ratio (Fig. 5). Therefore, pseudo-epizonal values are frequent in the lower grade zones. These samples were also not used for further studies (correlations).

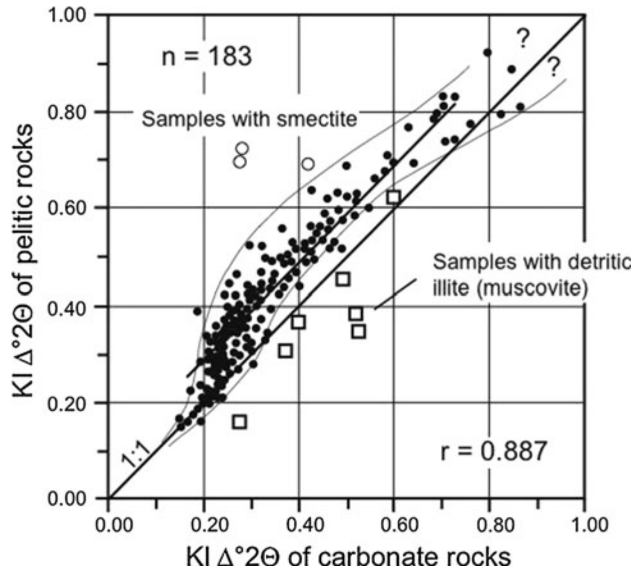

Fig. 4 Comparison of KI of claystone/shale and carbonate rock samples of the same locality to determine the petrologic variation of KI and the chemical control of the lithofacies on KI. For the correction factor used, see text. The *question marks* show results not well established or little constrained

Finally a high smectite component increases the sharpness ratio of the 10 Å basal line and broadens the KI. This is different in the effect comparing pelites with carbonate rocks (Mullis et al. 2002). Due to all this influences the KI values are plotted in the KI/OMR plot of Ferreiro Mählmann (1994) and only the values between the triangles of the high smectite containing rocks and the triangle of the detritus mica containing rocks (Fig. 5) are used for this study. This very restrictive procedure is possible because more than 100 South Penninic samples (Table 1), and including the hanging and footwalls of the Arosa zone 350 KI/OMR data-pairs are still available.

3.7 Representative data set

It is still not well established that in complex tectonic settings a high number of samples is needed to unravel the diagenetic and incipient metamorphic pattern using clay mineral or coal petrological data. Dunoyer De Segonzac and Bernoulli (1976) and Ring (1989) used a handful of data to determine zones at sub-greenschist facies conditions for a nappe structure or imbrications. Therefore these earlier studies presented misleading interpretations of the metamorphic grade in a regional

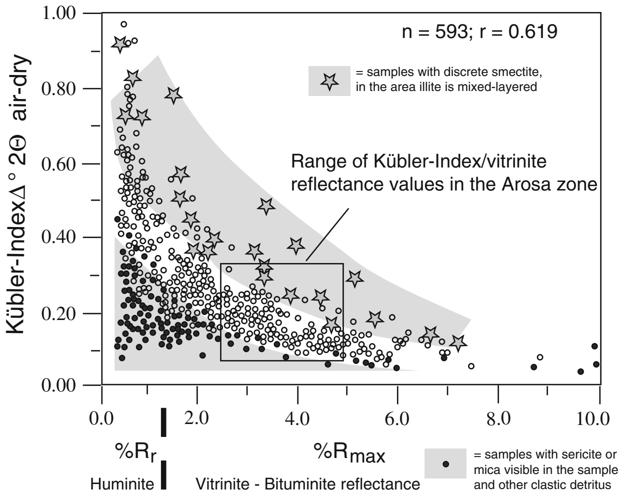

Fig. 5 KI-VR plot of all samples studies at the Austroalpine—Penninic boundary. In the *grey area* with samples marked with a *star* the high smectite content is related to weathering, retrograde diagenetic smectitisation or a very low-grade hydrothermal alteration of illite and chlorite (partly the discrete smectite peak shows interference with illite). In the *grey triangle* in the *left lower corner* of the plot (small KI at low VR) a detrital illite-muscovite component is suggested from the lithofacies or assumed due to interferences of different illite populations. In the white area between it is regarded that illite is neither affected by a detritus content nor by a high amount of smectite. For further correlation only the last sample group was used

scale. Using the same amount of samples only for one stratigraphic unit in a single tectonic structure the previous values were an element of the data variety observed by Kürmann (1993) or Henrichs (1993). To exclude effects from petrovariations, intra and inter-sample variations, uncertainties of the stratigraphic position in a section and the influences mentioned above (also the discrimination between sediment burial and tectonic load) a high number of samples is needed for a proper interpretation, including several specimens from one locality and two to four copies from one specimen (Frey et al. 1980; Frey 1987; Krumm et al. 1988; Petschick 1989; Ferreiro Mählmann and Frey 2012). Despite using a greater number of sample data and a smaller geographical grid, the complex pattern in the Austroalpine nappes and Arosa zone presented by Ferreiro Mählmann (1994) was initially not fully recognized. As done in other systematic studies (see also Ferreiro Mählmann 2001; Frings et al. 2004; Weh 2006; Ciulavu et al. 2008; Ferreiro Mählmann et al. 2012), the sample number in this work is maximized using own data and data compilations from literature, also from unpublished diploma studies and Ph.D. researches. The resulting pattern is again more precise, but also different in parts from earlier studies, thus leading also to a new interpretation.

4 Results

4.1 Diagenetic to metamorphic results derived from KI/VR data

4.1.1 Compilation of metamorphic data at the northern part of the study area

The general tectonic situation at the eastern edge of the Prättigau half window, between Klosters and the Tschagguns er Mittagsspitze is shown on Fig. 1. In the northernmost part of the study area sedimentary rocks of South Penninic and Lower Austroalpine origin occur together with fragments of ophiolite sequences in the mélange of the Arosa zone. Figure 6 shows the exemplary cross section studied at the Verajöchle (see Lüdin 1987; Ferreiro Mählmann 1994) at the west of Tilisuna (Figs. 1, 3). The tectono-metamorphic field relation in the Verajöchle-section is very representative for the Penninic–Austroalpine boundary at the northern edge of our study area (Ferreiro Mählmann 1994; Weh 1998) and this is the reason why the Arosa zone will later be re-defined as Walsertal zone in this area (Winkler 1988, see also Fig. 6 and "Sect. 5").

In the NE of this section the Upper Austroalpine Lechtal nappe is folded in a synform (Fig. 6). The anchizonal grade of the Lechtal nappe increases with stratigraphic depth (Ferreiro Mählmann 1987). The high diagenetic zone—anchizone limit is found in the Norian Hauptdolomite (Hd 117) near Lüner Lake. The anchizone—epizone boundary is established in the Permian (P) sediments. VR, KI and mineral index data indicate the same (parallel) trends of metamorphic grade (Ferreiro Mählmann 1994).

The imbricated basal Lechtal thrust of the Upper Austroalpine (forming nappe slices, "Schuppen"; Reithofer 1937; Tollmann 1977) is marked by a conspicuous "metamorphic inversion" sensu Frey (1988). Anchi-epizonal rocks of the Upper Austroalpine (Permian to Triassic) are thrusted over high diagenetic and low anchizonal (Frey et al. 1980) flysch units (F), Cretaceous marls (KM) and the Albian to Cenomanian Palombini formation (Pa). These tectonic and metamorphic relations lead to the conclusion, that the anchi-epizonal metamorphism of the Upper Austroalpine is older than thrusting and Cretaceous folding (both D1, westward movement of Upper Austroalpine according to Froitzheim et al. 1994). This Upper Austroalpine pattern is therefore related to sedimentary burial from Permian to Cretaceous (Ferreiro Mählmann 1994).

In contrast to the relatively lower grade matrix rocks of the mélange zone (Arosa zone/Walsertal zone), fragments of Triassic Muschelkalk (MK), Raibler sandstone (RS) but also the Jurassic breccias (J) and the Allgäu marls (A) show

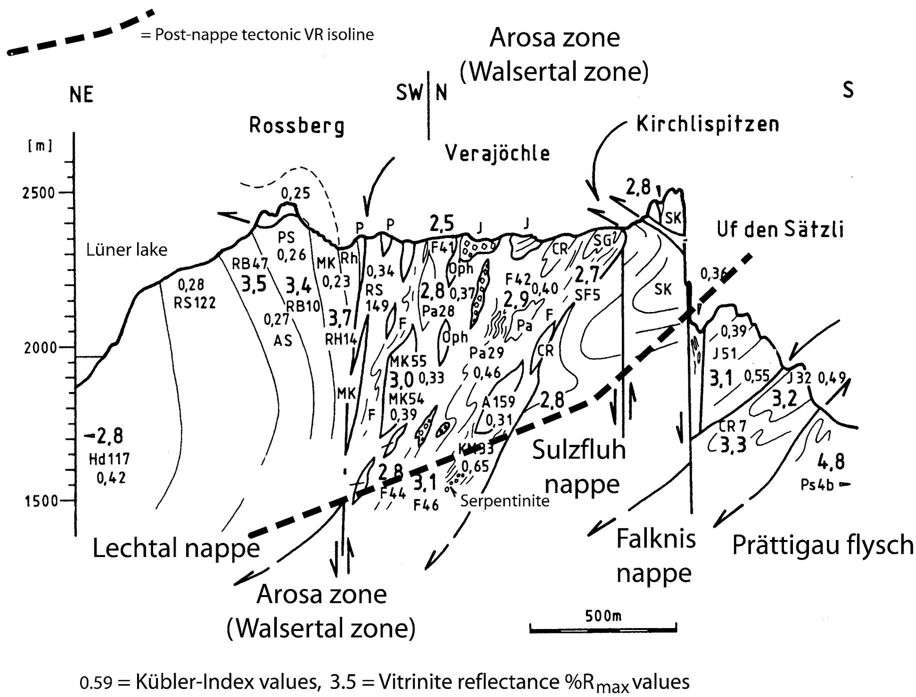

Fig. 6 The Verajöchle-section with the location of the samples used for Kübler-Index and vitrinite reflectance studies. Modified from a scanned ink-drawing from Ferreiro Mählmann 1994. Austroalpine: *P* Permian rocks, *RH* Reichenhaller formation, *MK* Muschelkalk (limestone and marls), *PS* Partnach slates, *AK/AS* Arlberg limestone/ Arlberg formation, *RS* Raibler marls, silt and sandstones, *Hd* Hauptdolomit, *KS* Kössen formation, *A* Allgäu slates and marls, *RB* samples from the Rossberg section (Ferreiro Mählmann 1987). South Penninic: *J* Jurassic rocks, *Pa* Palombini formation, *KM* Cretaceous marls, *F* Flysch, *Oph* Ophiolite magmatic rock. The tectonic fragments of MK 54, 55 and A 159 are thought to be of a lower tectonic imbrication (the Allgäu slates and marls of the Lechtal nappe show a KI of >0.45 $\Delta°2\theta$, Kralik et al. 1987; Petschick 1989)

a low to high anchizonal grade (Fig. 6). Permian to Jurassic formations and the pre-Mesozoic granite fragments in the Arosa zone (Walsertal zone) are paleo-geographically attributed to the Upper Austroalpine (Lüdin 1987; Ferreiro Mählmann 1994). This can best be explained, because the grade of metamorphism in the Upper Austroalpine fragments increases with stratigraphic depth and this is different from the Lower Austroalpine domain where structural depth controls KI, VR, clay mineralogy and the occurrence of facies critical or indicative minerals. Mineral neoformation is syn-kinematic with D1 and D2 slaty cleavage (Ferreiro Mählmann 2001).

Compared to the metamorphic grade of the well-defined stratigraphic units in the adjacent Upper Austroalpine Lechtal nappe, the metamorphic grade of Austroalpine fragments in the Arosa zone (e.g. Verajöchle, fragments in the Tilisuna area, Fig. 1) is significantly lower when KI and mineral data are considered. But in relation to the matrix sediments of the mélange, KI values of the fragments are still indicative of a higher, inherited and transported metamorphism (Fig. 6).

Including new data from the lower structural nappes, in the Arosa zone a top–down trend from high diagenesis to low anchizone is found, but VR rock maturity increases also continuously from the Verajöchle to the south. The maturity trend lines cross cut the basal thrust of the mélange and this trend is therefore post-kinematic to thrusting. The VR increases also in the Middle Penninic Sulzfluh and Falknis nappes (3.1–3.3 $\%R_{max}$). For the North Penninic Pfävigrat formation (sample Ps 4b on Fig. 6), some 500 m to the south and below the basal Middle Penninic thrust, a value of 4.8 $\%R_{max}$ was established. The VR values probably indicate a thermal overprint post-dating the general nappe tectonics (Ferreiro Mählmann 1994; Ferreiro Mählmann and Petschik 1997). Including data of the Rupelian rocks in the north of the Lechtal nappe, the age of incipient metamorphism is post-Oligocene. But a more detailed study in the North Penninic did not yet clearly substantiate that (Petrova et al. 2002; Ferreiro Mählmann et al. 2002; Wiederkehr et al. 2011). Further work in these Penninic units is in progress.

KI, the illite-smectite (I-Sm) reaction progress and index minerals (e.g. glauconite in the Sulzfluh Gault (SG), flysch (SF) and some Jurassic rocks (J), pyrophyllite in sandy layers in the Sulzfluh limestone (SK), stilpnomelane in red Falknis Couches rouges (CR, see Fig. 6) altogether indicate a similar metamorphic trend. In the structural lower units below the Lechtal nappe some differences are observed. KI data show small or prominent metamorphic inversions at nappe boundaries. The isogrades of a post-nappe tectonic thermal event, is clearly indicated only by the VR method and the VR isoline is later offset by young vertical faults (Fig. 6). The metamorphic imbrications in the footwall of the Arosa zone documented by KI and mineral data is not verified by the VR data set.

For further information about the situation in the lower tectonic units we refer to Gruner (1980) and Frey et al. (1980). In the North Penninic Flysch and Bündnerschiefer, metamorphism increases with stratigraphic and structural depth (Weh et al. 1996; Ferreiro Mählmann and Petschik 1996; Petrova et al. 2002; Ferreiro Mählmann et al. 2002). This was also very well constrained in a earlier, systematic KI study by Thum and Nabholz (1972).

4.1.2 *Metamorphic pattern in the central part of the study area*

From the Swiss–Austrian boundary to the south (Fig. 1), KI values are more homogeneous and are indicative for a general structural depth trend very similar to that of the VR data distribution found at the Verajöchle in the north. The cross section of Fig. 7 represents the metamorphic pattern derived from KI/VR data and extends from Küblis northwards to St. Antönien and Tschaggunser Mittagsspitze (Fig. 1). Along this section, again sedimentary rocks of different tectonic units are exposed (North Penninic flysches to Upper Austroalpine sediments). Another W–E cross-section (Fig. 7) extends from St. Antönien to the Gargellen window (Figs. 1, 3). Along in this section the entire tectonic pile from Penninic flysches to Upper Austroalpine sediments is exposed.

Near Küblis, the highest metamorphic grade (low epizonal condition) is found in the N–S-section shown in Fig. 7. From Küblis northward the metamorphic grade decreases. The epizone to anchizone transition is situated near St. Antönien-Ascharina (1,320 m). From St. Antönien to the Lake Partnun (1,869 m) the drop of VR values (Fig. 7) and KI (Table 1) is indicative for a considerable decrease in the metamorphic grade. In the area Verspala—Tschaggunser Mittagsspitze, over a distance of 4.0 km, anchizone to high-grade diagenetic conditions are found, with KI values of 0.28–0.45 Δ°2θ and a rock maturity of the high semi-anthracite (‘Magerkohle”) stage (VR = 2.8–3.1 %R_{max}).

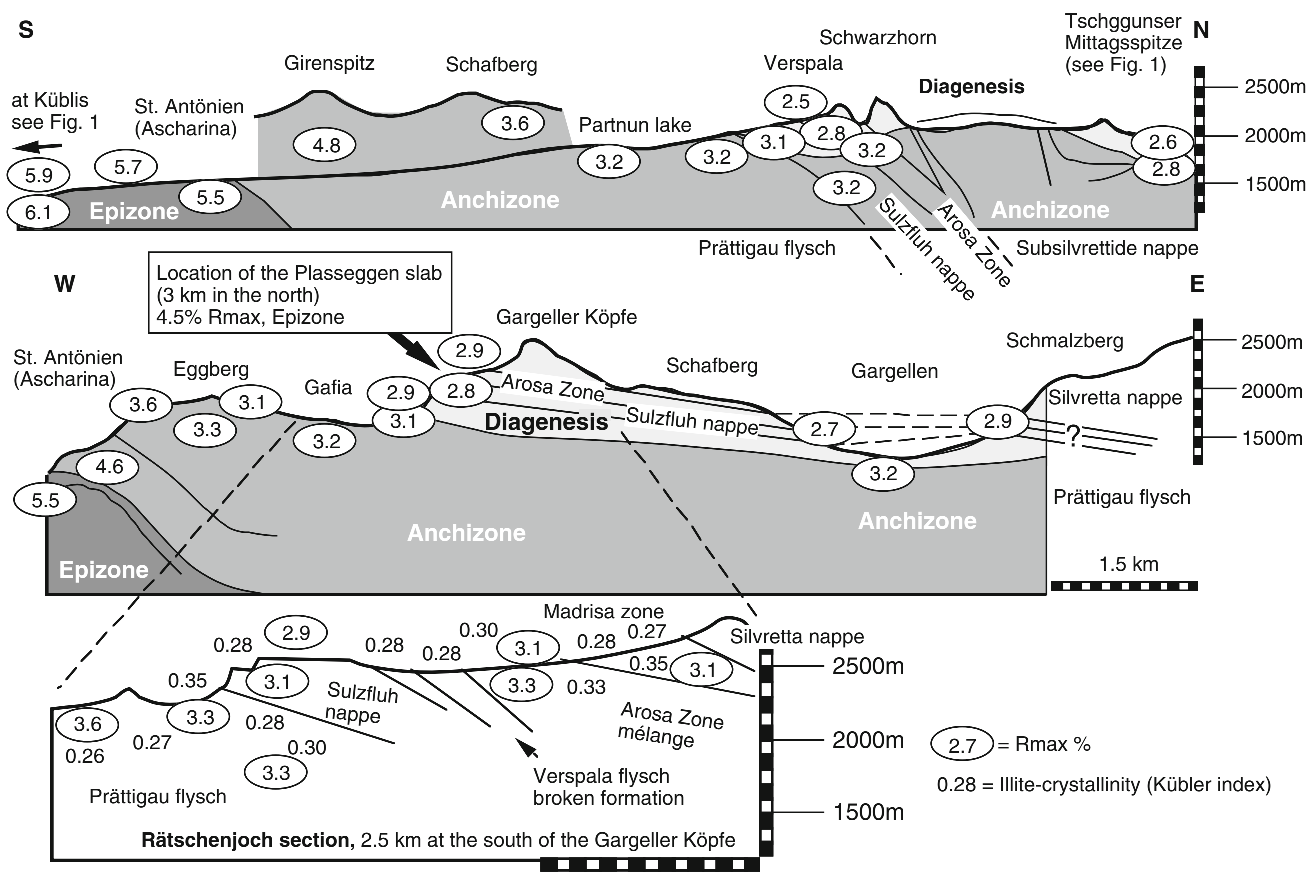

Fig. 7 Metamorphic-structural sections through the northern part of the Arosa zone. Diagenesis and metamorphic zones were defined by KI Δ°2θ. Note the inherited high metamorphic grade of the Plasseggen slab in the Arosa zone. The *question mark* shows results not well established or little constrained

Contrasting to all other rocks the broken-formation of the Verspala flysch (Cretaceous flysch formation) and some Palombini formation rocks in the Arosa zone mélange matrix show KI values of 0.40–0.50 Δ°2θ) and a rock maturity of the high semi-anthracite ("Magerkohle") stage (VR = 2.5–2.8 $\%R_{max}$). An unconformity between this units was postulated by Lüdin (1987) and might be demonstrated by our diagenesis indication. Also in other areas to the north these younger Cretaceous rocks have a significant lower diagenetic grade. In general the Arosa-zone samples show a higher Sm content than the Austroalpine samples and a lower one than the Verspala and other flysches. This relation is found in the Arosa zone (and Walsertal zone) from the north of Davos (this paper) to the deep drilling at Hindelang (Ferreiro Mählmann 1994). Comparing KI and VR values and plotting the data in the graph of Fig. 5 the Sm rich samples with a discrete Sm-interference are marked with a star. If Sm is present the KI tends to broader FWHM values (KI air-dry), thus this is an additional effect of Sm. If flysch rocks (F 42, KM 33, Pa 29; see Fig. 6; Table 1) are intercalated in older Mesozoic rocks, as found at Verajöchle, a low-grade diagenetic "island" (0.40–0.65 Δ°2θ) is dismembered between anchizone rocks (0.31–0.37 Δ°2θ). In the southern part of the Arosa zone the flysches, the lower Cretaceous and older rocks of the South Penninic domain did not show differences in metamorphic grade. This is an important difference to the Walsertal zone re-defined later in this work.

As shown in Fig. 7 the epi-anchizone transition near St. Antönien dips steeply to the north. In contrast to that the anchizone–diagenesis transition in the Gafia valley and at the Verspala has an almost planar geometry. It is only slightly curved in the Verspala region and dips gently to the north at the Tschaggunser Mittagsspitze. Restoring the diagenesis–anchizone transition to a horizontal plane and assuming a planar paleo-surface a mean coalification gradient of 0.65 $\%R_{max}$ km^{-1} can be calculated. At the thrust planes of the major tectonic units no coalification or illite aggradation hiatus is observed in this northeastern part of the study area.

VR data are also plotted in a west–east section from St. Antönien-Ascharina to Gargellen (Gargellen window; second cross-section in Fig. 7). Again the epizone–anchizone transition dips steeply to the east but the anchizone–diagenesis transition again shows an almost plane geometry. The reconstruction of a coalification gradient yields a mean value 0.70 $\%R_{max}$ km^{-1}. In both sections this gradient tends to increase from the diagenesis (0.65 $\%R_{max}$ km^{-1}) to the epizone (0.70 $\%R_{max}$ km^{-1}). KI data (not shown) substantiate the trends, but with more scattering.

The inset of Fig. 7 shows the fairly complex situation at the Rätschenjoch (6 km north of Klosters). Units of the entire tectonic pile are exposed in a section only 3.0 km long and within a topographical elevation difference of 400 m. The Madrisa zone (Upper Austroalpine), the tectonic mélange of the Arosa zone (South Penninic), the slab of the Verspala broken formation (South Penninic), the Sulzfluh nappe (Middle Penninic) and the Early Eocene Ruchberg sandstone (Prättigau flysch, North Penninic) are shown in the figure. All the above-mentioned tectonic units display almost the same grade of metamorphism in that section. At the Rätschenjoch VR is ranging between 2.9 and 3.3 $\%R_{max}$ and KI between 0.27 and 0.35 Δ°2θ. The situation at Rätschenjoch demonstrates clearly, that the main metamorphic heating affecting the sediments postdates the Eocene nappe thrusting also in this part of the study area.

It can be summarized, that in the northern and northeastern part of the study area metamorphism increases from north to south (see also Fig. 8) and from east to west. The most intense metamorphism of that area is observed between Küblis and St. Antönien in the topographic notch of Schanielatobel (Frey et al. 1999; Frey and Ferreiro Mählmann 1999). The sedimentary matrix rocks of the Arosa zone (Jurassic slates, Cretaceous Palombini formation and flysch rocks) yield data typical for an anchizonal overprint and the variation is relatively small (KI = 0.40–0.28 Δ°2θ, VR = 2.8–3.2 $\%R_{max}$). In the area from the Madrisa to Klosters metamorphism of Arosa zone matrix increases from 0.35 Δ°2θ, 3.1 $\%R_{max}$ at 2,600 m (Madrisa) to 0.30 Δ°2θ, 3.5 $\%R_{max}$ at 1,130 m (Klosters).

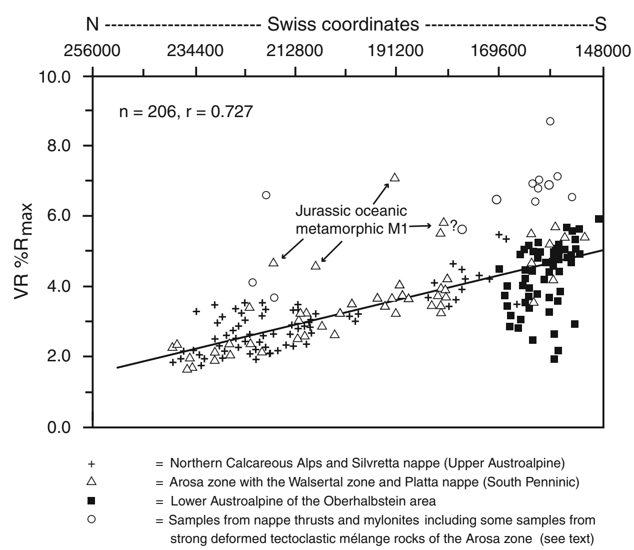

Fig. 8 Plot of vitrinite refectance versus N–S Swiss coordinates showing the general top south trend of increasing rock maturity. In the Arosa zone and more impressively, in the Lower Austroalpine of the Oberhalbstein the large VR scattering is related to different top-down trends with structural depth

This is also well demonstrated plotting the VR values according to their location in the Swiss north–south coordinate system (Fig. 8). The general post-nappe tectonic pattern described above is “disturbed” at the Plasseggenjoch. The topographic and tectonic position of the Lower Austroalpine Plasseggen fragment is shown on Fig. 7. The Plasseggen fragment consists of basement rocks (mainly biotite granite gneisses and feldspatic schists) and sediments (dolomites and phyllites of Triassic to Jurassic age). These sediments show an epizonal overprint (VR = 4.5 $\%R_{max}$, “meta-anthracite stage”, KI = 0.13 $\Delta°2\theta$). The grade of metamorphism in the surrounding rocks is reflected by the low-grade anchizone (flysch rocks and South Penninic marine sediments) giving KI = 0.30–0.39 $\Delta°2\theta$ and VR = 2.8–3.2 $\%R_{max}$. The metamorphism of this “exotic” (according to Tollmann 1977) Plasseggen fragment is again, like for other Austroalpine fragments, inherited and transported.

In the area between Tilisuna-Verspala and Madrisa (Fig. 1) the sedimentary matrix rocks of the Arosa zone (Jurassic slates, Cretaceous Palombini formation, Cretaceous marls and flysch rocks) yield data with only a small variation (Figs. 7, 8) and characteristic for a low anchizonal grade. A topographic and structural top–down trend is evident for this part of the study area.

4.1.3 *Metamorphic pattern in the southern part of the study area*

From Klosters to Tiefencastel in the South of the study area the metamorphic pattern of the tectonic pile is even more complex (see Figs. 1, 9).

In the Upper Austroalpine the metamorphism of sediments ranges from diagenesis to the epizone, as reported by Ferreiro Mählman (1994, 1995, 1996, 2001) and the grade increases with stratigraphic depth. At the base of the Upper Austroalpine the metamorphic grade increases from the North where high anchizonal conditions are observed in the Lechtal nappe to the South where epizonal conditions occur at the base of the Silvretta nappe s.s. (Figs. 9, 10). This diagenetic to epizonal pattern has been interpreted as follows:

1. Metamorphic grade of the Upper Austroalpine sediments is the sum of a complex polyphase and plurifacial metamorphic history (Ferreiro Mählmann 1994).
2. For the Permian to Raetian sediments in the Ducan and Landwasser synforms (Silvretta nappe s.s.; cf. Fig. 1) a high pre-orogenic paleo-geothermic gradient (85 °C km^{-1}) is established (Ferreiro Mählmann 1995, 1996).
3. This can be best explained by syn-sedimentary, diastathermal heating (Ferreiro Mählman 1996).

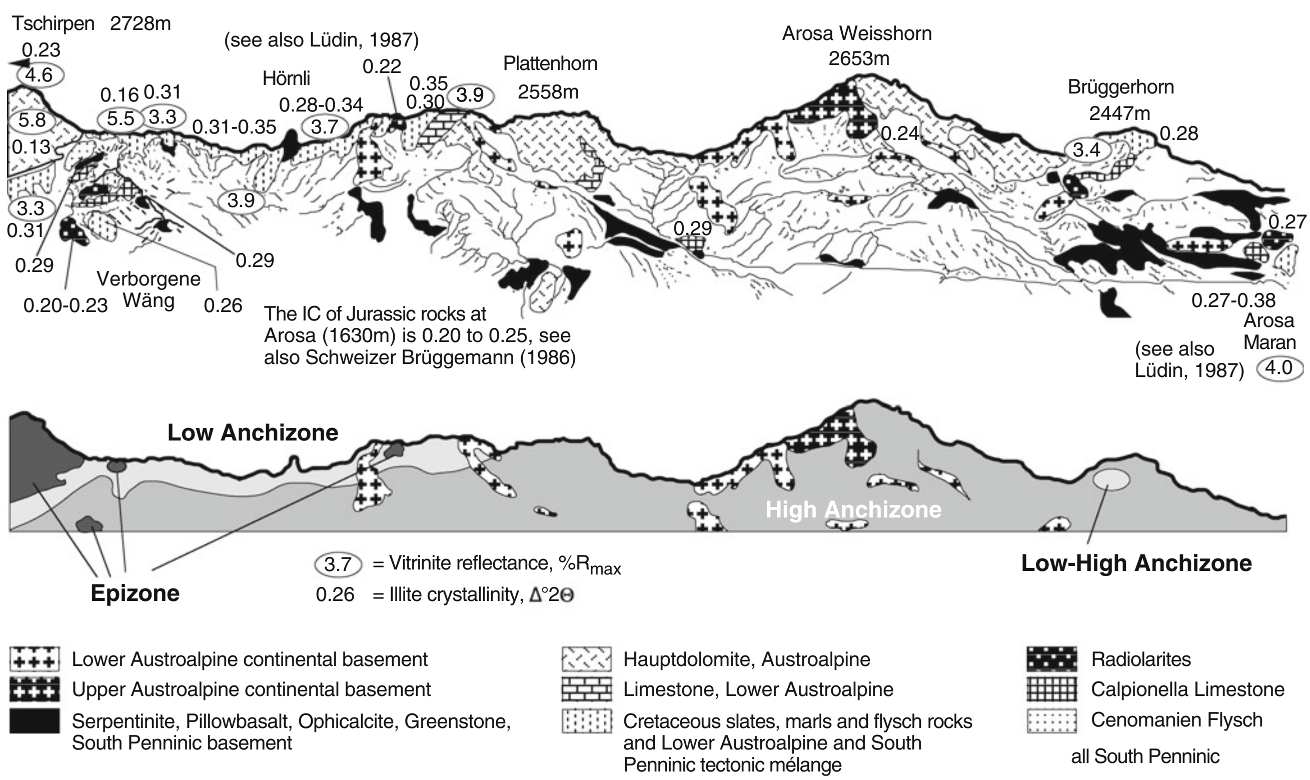

Fig. 9 The Arosa section, modified from Streckeisen 1948 with the location of the KI ($\Delta°2\theta$) and vitrinite reflectance ($R_{max}\%$) measurements. In the section below the values are combined to draw metamorphic zones. At three localities, spots of the epizone are related to occurrences with a Jurassic hydrothermal oceanic metamorphism

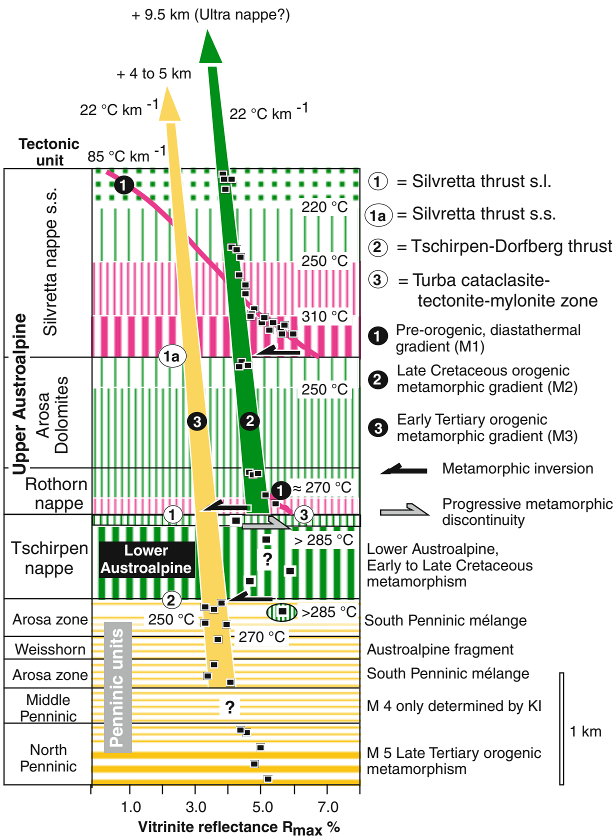

Fig. 10 Reconstruction of thermal gradients and the missing erosional overburden in the Mittelbünden section. Different metamorphic discontinuities and the different ages of the three metamorphic patterns are shown with different signatures. The first and second deformation and the first and second metamorphic phase are restricted to the Austroalpine nappe stack. The Arosa zone is overprinted exclusively by the third event of a low-temperature geotherm. The signatures of the metamorphic zones are explained in Fig. 16. The *question marks* show results not well established or little constrained

4. Normal to hyper-thermal (>60 mW/m^2) diastathermal heating took place during Permian, Anisian–Ladinian, Norian and Jurassic times (Ferreiro Mählmann 2001).

In the southwest of the study area a metamorphic inversion separates the Silvretta nappe s.s. from the Upper Austroalpine imbrications below (e.g. Arosa Dolomites; Rothorn nappe; cf. Fig. 1). The sediments of these lower imbrications of Upper Austroalpine (D1) are overprinted by a low to high anchizone metamorphism. The diagenetic to metamorphic pattern of these units is the result of the pre-orogenic burial (first metamorphic event, M1) and the Cretaceous heating during the Alpine orogenesis (second metamorphic event, "dynamothermal" orogenic M2). M2 post-dates the Upper Austroalpine thrusting and folding D1 (Ferreiro Mählman 1996).

Southwest of Arosa, Lower Austroalpine sediments of Triassic to Jurassic age are thrusted (D2, according to Froitzheim et al. 1994) over the Arosa zone (e.g. Tschirpen nappe fragment; Figs. 1, 9). The Tschirpen sediments are overprinted by a high anchizonal to low epizonal metamorphism (KI = 0.23–0.13 $\Delta°2\theta$, VR = 4.6–5.8 $\%R_{max}$; see also Fig. 9).

Small fragments (KI = 0.16 $\Delta°2\theta$, VR = 5.5 $\%R_{max}$) of the Tschirpen nappe also occur at the structural top in the Arosa zone mélange (Fig. 9). As in the northern part of the study area, the metamorphic grade of many Austroalpine nappe fragments differs markedly from the surrounding sediment matrix of the Arosa zone. Some Lower Austroalpine sediment fragments ("Schuppen" = tectonic slices) are characterized by higher anchizonal to epizonal conditions (Figs. 7, 9). Other tectonic fragments in the Arosa zone are attributed to the Austroalpine domain because of their sediment facies and lithology but they do not differ in metamorphic grade from Arosa zone South Penninic sediments. The Arosa Weisshorn fragment (Figs. 1, 9) and the Weissfluh fragment (near Davos, Fig. 1) are examples for this second type of Upper Austroalpine "Schuppen", probably better explained as outlayers (Klippen).

The following paragraph deals with the sediment matrix of the Arosa zone (stratigraphy see Figs. 2, 6, 9, 11, 13). These matrix sediments yield anomalously large KI values in the region of the Gotschnagrat and on the slope down to Klosters (cf. Figs. 1, 12). These KI values are in the range of 0.36–0.65 $\Delta°2\theta$ and therefore characteristic of the diagenesis to low anchizone. Neglecting that very low-grade anomaly near Klosters, the Arosa zone sediments generally show an anchizonal overprint (see Figs. 8, 9, 10, 11, 12). The KI values tend to decrease from the region of Davos (0.38–0.30 $\Delta°2\theta$) to the area east of Arosa (0.35–0.25 $\Delta°2\theta$). Rock maturity values range from 3.2 to 3.3 $\%R_{max}$, values typical for the 'Magerkohle-Anthrazit' stage. A general metamorphic trend is difficult to establish for the Arosa zone sediment matrix between Davos and Arosa (Figs. 8, 12). One reason is, that petrovariances of KI are frequent (Figs. 4, 12). The KI zones are generally defined for marine clays and slates (Kübler 1967; Krumm 1984; Frey 1987 for a review). The bulk rock chemistry, precursor clay mineral composition, physical rock parameters, metamorphic fluid composition and cation availabilities are influencing KI in very complex processes (Árkai et al. 2002a, b; Mullis et al. 2002; Abad et al. 2003). It is well known that metamorphic indication should be restricted to Al rich rocks (Al-micas, see Esquevin 1969). In the anchizone a correction factor of 0.055 $\Delta°2\theta$ is applied (Fig. 4) for data from the carbonate rocks. With this corrected data-set an almost homogeneous anchizonal pattern is evident showing a structural top–down trend (Fig. 9), but it is repeated that this correction can only be applied to the specific conditions in the Davos-Arosa region.

For a better understanding of the diagenetic–metamorphic conditions in the mélange matrix a Cretaceous

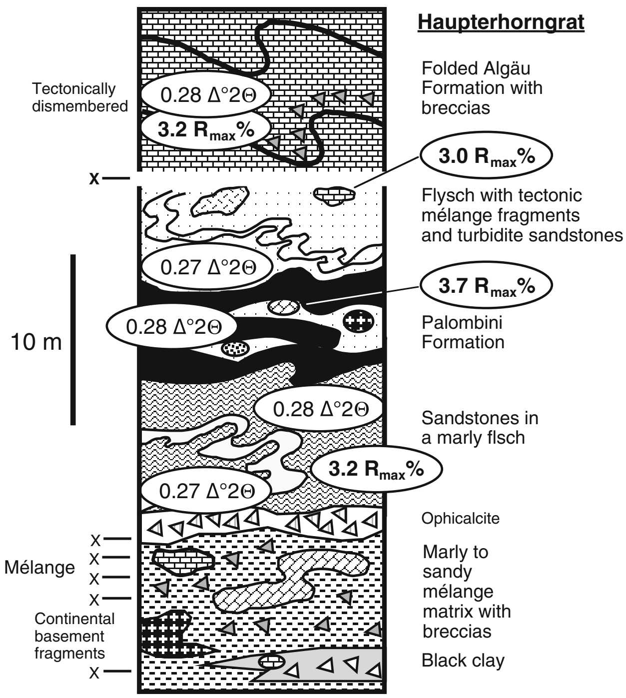

Fig. 11 The Haupterhorngrat section modified from Lüdin (1987). Conglomerate components of the Kössen formation (Rhaetian) with a high vitrinite reflectance of 3.7 %R_{max} is detritic in a Cretaceous Palombini formation sandy turbidite (Aptian to Turonian). Maturity demonstrates the Mesozoic pre-orogenic M1 metamorphism (pre-Cretaceous)

conglomerate-sandstone layer from the southwest of the Weissfluh-Schuppe (Weissfluh slice, Figs. 1, 10) was studied. The turbidite conglomerate is composed of gneisses, granites, prevailing carbonate detritus and Palombini limestones bedded in a psamitic carbonate (marl) and quartz (silty clay) matrix. In the Allgäu formation around the Weissfluh slice, the flysch rocks and pelites are overprinted by a high anchizonal metamorphism (KI = 0.27–0.28 Δ°2θ, n = 5, VR = 3.2 %R_{max}, n = 2). The VR value from the sandy, marly slate is in accordance with that (VR = 3.2 %R_{max}), but a clay-rich limestone (Austroalpine Triassic, may be from the Kössen formation) in the turbidite (section “Haupterhorngrat” number “Serie 3a” of Lüdin 1987) shows a significantly different coalification stage of 3.7 %R_{max} (controlled by a *T* test), thus similar to the values of the Kössen formation sample from the Gotschnagrat slice near Klosters (Fig. 1, and KS 107, Table 1). This means that the rock maturity of the Austroalpine fragments is older (M1 or M2) than the re-sedimentation into the oceanic basin attributed to the South Penninic domain (Arosa zone).

A similar situation was found during an excursion with Daniel Bernoulli, Gretchen Früh-Green and Helmut Weissert at Parsenn-Totalp (see also Früh-Green et al. 1990). Intercalating with radiolarites (South Penninic) and pillow basalts, a breccia with a sandy matrix and crystalline basement fragments contains breccia components of Agnelli silicate limestones with belemnites (Lower Austroalpine provenance). Also in this case the anchizonal grade is significantly higher (0.25 Δ°2θ) than the KI from the black and grey clays in the radiolarites (0.31 Δ°2θ, n = 2), the Calpionella limestone (0.30 Δ°2θ, n = 3) and the Palombini formation (0.35 Δ°2θ, n = 3).

Also during an excursion with Martin Frey and Tatiana Petrova in 1998 and later with Ulrich Szagun in 1999, Austroalpine fragments were found in the South Penninic matrix at Arosa-Maran (Fig. 9) close to the base of the Arosa zone, being in contact with the North Penninic Ruchberg sandstone in this area. South Penninic flysch and Cretaceous sediments show the same high to low anchizonal grade (0.27–0.38, mean = 0.30 Δ°2θ, n = 8, VR = 4.0, n = 2). In the Maran region it was not possible to attribute a Lower or Upper Austroalpine provenance to the different fragments, without the experiences in Austroalpine stratigraphy and help, as in other areas, by D. Bernoulli, H. Weissert and G. Früh-Green or the excellent stratigraphic study of Lüdin (1987) and his support in Basel with his rock collection.

At Parsenn-Totalp and in the southwest of the Weissfluh (Schwerzi to Haupterhorn) slightly dismembered profiles of South Penninic oceanic crust are well preserved (Lüdin 1987). In this profiles (Fig. 13) KI values increase only slightly with formation depth. However, at the bottom of the profile VR rapidly increases from 3.2 to 3.8 and 7.0 %R_{max} on a short distance. This rapid variation of VR on a short distance can be best explained by a hydrothermal effect of the adjacent ophiolithic rocks and thus represents the pre-orogenic oceanic metamorphism (ocean floor metamorphism sensu Miyashiro et al. 1971). A hydrothermal mineralisation at the contact of the pillows to the sediments is described Ferreiro Mählmann (1994). In this heated sediment rocks the greater increase of VR relative to KI demonstrates that VR adjusts more quickly to a thermal signal than KI (see also Wolf 1975). For the clay reaction smectite-to-illite the duration of heating is more important (Kisch 1987; Robert 1988; Hillier et al. 1995).

A similar sign for an oceanic metamorphism or a post-oceanic volcanic contact to the basal sediments was observed near Arosa (Verborgene Wäng; Plattenhorn; cf. Fig. 9): radiolarites and Calpionella limestones that are closely associated with ophiolithic basement rocks yield KI values of high anchizone to epizone (0.29–0.20 Δ°2θ). This Jurassic oceanic metamorphic overprint (designated as Jurassic M1; see Fig. 8) in meta-sediments is only of very local importance in the Arosa zone (Figs. 9, 12, 13).

In the area of Arosa a topographic effect is observed for the metamorphic overprint of the Arosa zone sediments (Fig. 9). Considering also the KI data of Lüdin (1987) KI is

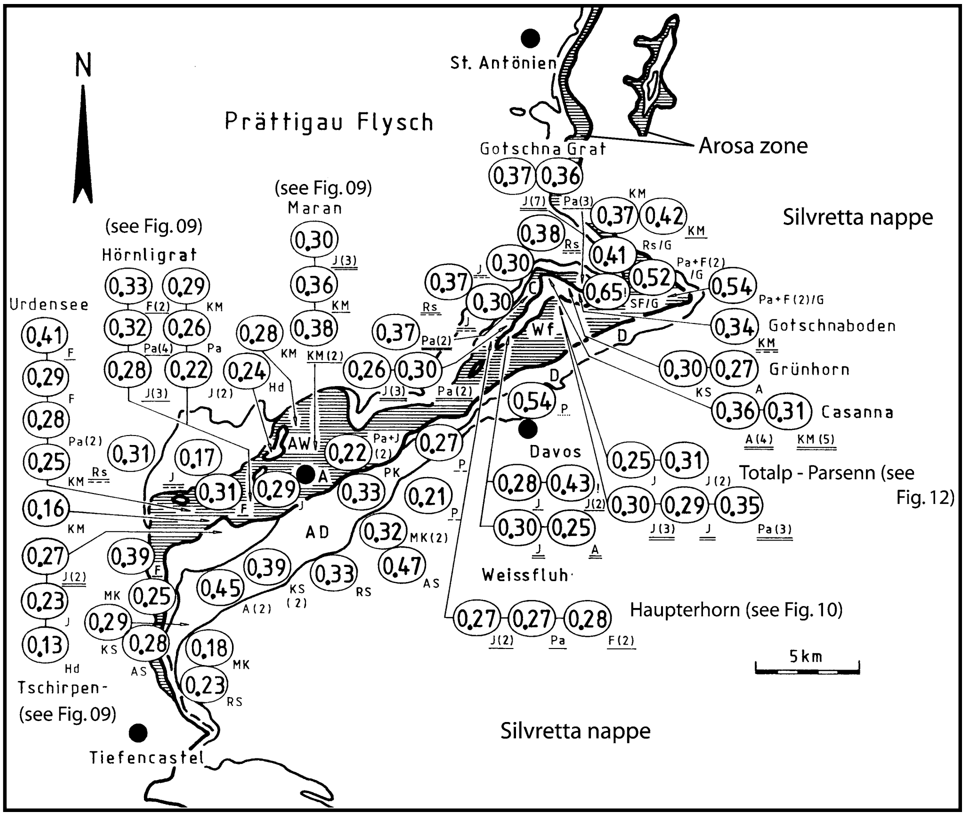

Fig. 12 Kübler Index (illite-"crystallinity") map of the Arosa zone. *P* Permian, *MK* Muschelkalk formation, *AS* Arlberg formation, *RS* Raibler formation, *Hd* Hauptdolomite, *PK* Plattenkalk, *KS* Kössen formation, *A* Allgäu formation, *J* Jurassic, *Pa* Palombini formation, *KM* Late Cretaceous formations, *F* Late Cretaceous flysch formations, *SF/G* Gault of the Sultzfluh nappe, *Rs* Ruchberg series (Eocene), (5) = mean of five samples (one sample = three individual measurements, thus five samples are represented by 15 KI values). *A* Arosa, *AD* Arosa Dolomites, *AW* Arosa Weisshorn, *D* Dorfberg nappe, *Wf* Weissfluh. Some values are marked with different types of *lines* indicating the origin from unpublished PhD. and Dilpoma thesis data (*one line* Lüdin 1987; *double line* Schweizer-Brüggemann 1986; *pointed line* Giger 1985; *solid and dashed line* Gruner 1980; *dashed line* Weber 1976). Modified from a scanned ink drawing from Ferreiro Mählmann (1994)

larger for the higher elevations (e.g. Hörnli 2,496 m and Weisshorn 2,653 m) and decreases downwards to the Lake Urden (2,249 m) and to Arosa-Maran (1,862 m). The KI values are 0.30 ± 0.05 $\Delta°2\theta$ at the top and 0.28 ± 0.02 $\Delta°2\theta$ at the bottom. On the other hand this trend is not yet well defined by VR values (Fig. 9; Ferreiro Mählmann and Frey 2012).

Only limited KI and VR data are available for the tectonic base of the Arosa zone, thus no coherent rock maturity and metamorphic grade can yet be established for the "footwall" of the Arosa zone. In some areas, e.g. west of Weissfluh (between Klosters and Arosa; Fig. 1) preliminary data seems to indicate another metamorphic hiatus between Arosa zone (low anchizone) and Middle Penninic units (high anchizone, unpublished data from U. Szagun, see Frey et al. 1999). It seems, that west of Weissfluh high anchizonal Middle Penninic sediments are squeezed between the low anchizonal Arosa zone sediments (tectonically above) and the low anchizonal Penninic Bündnerschiefer (tectonically below).

The tectono-metamorphic relations in the Middle Penninic nappes seem to be much more complicated than formerly recorded by Gruner (1980); Frey et al. (1980); Ferreiro Mählmann (1994) and Weh (1998), and these units are still not well documented in the metamorphic maps (Ferreiro Mählmann and Petschik 1997; Frey et al. 1999; Oberhänsli et al. 2004).

4.2 The correlation of Kübler-Index and vitrinite reflection values

Soon after the introduction of the IC method by Kübler (1967, 1968) as well as Frey and Niggli (1971) to study the very low grade metamorphism, a first calibration/correlation was attempted with the well established coalification determination (based on VR), a method already a century old. Kübler and Frey noted very early that a general correlation is difficult to establish and recommended to refer to the tectono-metamorphic history (Kübler et al. 1979; Frey et al. 1980). For a better understanding of the KI/VR

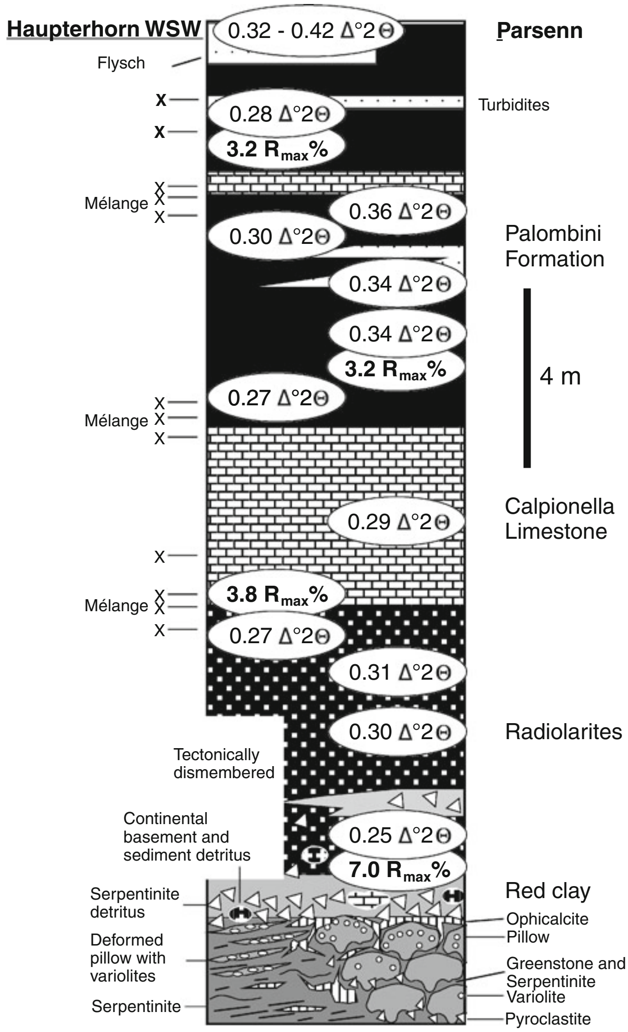

Fig. 13 The Haupterhorn WSW and Parsenn section (Totalp) modified after Lüdin (1987). The high Kübler-Index (Δ°2θ) and vitrinite reflectance (R_{max}%) gradient is related with a hydrothermal epidote-amphibole hornfels facies in the red clay, the ophicalcite and the greenstone. The younger Radiolarite is not included. Thus the hornfels can be related with an oceanic metamorphism. The gradient in the Haupterhorn WSW section is probably a remnant of the thermal history during rifting found in the sediments of the South Penninic ocean floor cover

correlation trend every tectono-paleo-geographic unit has to be plotted in a separate graph. In an extended study on 1,124 samples Ferreiro Mählmann (1994) has recognized that every Austroalpine nappe (7 nappes) and Penninic nappe (8 nappes) has it own typical correlation. The same is known from the Helvetic nappes (Frey et al. 1980; Rahn et al. 1995; Schmidt et al. 1997; Ferreiro Mählmann et al. 2012). It was implied that the correlation in each case is dependent on kinetics in various ways.

Correlating all samples of the Arosa zone where both methods (KI/VR) were applied (but excluding those with high Sm content or detrital mica, Fig. 5), 48 samples document a general trend with a broad variation with a Δ VR of 1.0–2.5 %R_{max} at a specific KI value. The correlation using only samples from the Arosa mélange zone related to the South Penninic sedimentation realm is clearly of much higher significance (Fig. 14). South Penninic samples contrast to the trend found in the samples from Lower and Upper Austroalpine fragments. The difference in KI/VR correlation is more obvious with increasing metamorphism, showing a slow progression for South Penninic samples (Fig. 14).

The KI/VR slope found for the South Penninic samples contrasts also with the slopes published from the Upper Austroalpine nappes (Ferreiro Mählmann 1994; 1995) and from the Lower Austroalpine nappes (Ferreiro Mählmann 1995, 2001), but also from the relation found in the Penninic nappes in the footwall of the Arosa zone (Petrova et al. 2002; Ferreiro Mählmann et al. 2012). Comparing KI and VR, the rock maturity is generally higher in the Austroalpine fragments (Fig. 14). Because metamorphism (M1 and M2) in the Austroalpine fragments is inherited prior to the re-sedimentation in South Penninic rocks, in dismembered sediment strata (broken formations) and tectonic “Schuppen” displaced into the Arosa zone mélange, the KI/VR (OMR) correlation cannot give additional information about the grade and metamorphic conditions of the mélange matrix. Therefore for further correlation and metamorphic grade determination the study concentrates on the samples of South Penninic provenance (“matrix sediments” of the mélange).

Flysch, pelites and the marly matrix of the Arosa zone is in many places highly deformed and a slaty cleavage developed. Also from Arosa to the south the quartz brittle–ductile transition is surpassed and schistosity formed with a mineral stretching lineation. Strain combined with high anchizonal to epizonal grades enhances VR and the “bireflectance of vitrinite and bituminite” (e.g. Teichmüller 1987) as compared to KI (Ferreiro Mählmann 2001; Árkai et al. 2002b; Ciulavu et al. 2008). But strain as an influencing factor on the KI/VR correlation (Ferreiro Mählmann et al. 2012) is recognised only from a few tectonites in the Arosa zone and mostly related with the basal Austroalpine thrust (Fig. 8).

Figure 14 show plots of VR versus KI data from metasediments of different tectonic provenance and displaced together in the tectonic mélange of the Arosa zone. The important differences in the correlation of these data are obvious. It is evident that in the Austroalpine KI values correlate with higher VR values and in the Penninic units correlate with lower VR values. All KI/VR values plot in the integrated data area typical for orogenic diagenesis and metamorphism as postulated by Ferreiro Mählmann et al. (2012). The data set of the South Penninic Arosa zone samples plots slightly displaced to the left in the VR/KI plot (Fig. 14). Such correlations are reported from

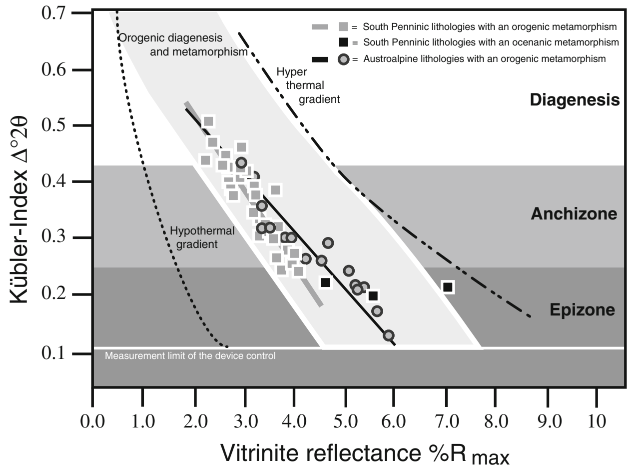

Fig. 14 Graph showing the correlation between vitrinite reflectance (VR) R_{max}% and Kübler-Index (KI) Δ°2θ, using the samples exclusively from the Arosa zone (Table 1). The KI-VR trend from samples of Lower Austroalpine fragments shows a regression of high significance and a moderate slope to high maturity. The second correlation in the graph, neglecting the low epizonal samples with a very high maturity, and using only samples from South Penninic rocks, shows a steep trend definitely different from the Lower Austroalpine. Comparing the KI versus VR data from the South Penninic rocks of the Arosa zone with data from literature (Ferreiro Mählmann et al. 2012), the Arosa zone data fit better with typical trend found for normal to slightly lower heat flow conditions and orogenic metamorphic settings

geodynamic settings with a slightly lowered orogeneic geothermal gradient. On the other hand the data set of the Austroalpine samples plots nearly in the middle of the VR/KI plot related to normal orogeneic geothermal gradients.

The data variation of KI and VR values is low for the entire Arosa zone and the bulk can be attributed to the anchizone. In the north the South Penninic rocks have a distinctly lower grade (including diagenesis at Verajöchle, Fig. 6) than the adjacent Austroalpine and to the south the Austroalpine fragments have mostly a much higher grade including the epizone. So, in both areas the Austroalpine fragments can be mostly distinguished from the South Penninic matrix by grade of metamorphism. If Austroalpine fragments with a inherited burial anchizonal grade occur, they cannot be distinguished by metamorphic grade, but the KI/VR relation is still different!

Three South Penninic sedimentary rocks of Jurassic age are of high epizonal grade (Figs. 13, 14). These epizonal red clay rocks (also radiolarites) are stratigraphically concordant to ocean floor basement rocks of the Arosa zone (cf. Figs. 9, 13) and this overprint is interpreted as a fingerprint of pre-orogenic, oceanic metamorphism (Jurassic M1). A strong shift in the KI/VR graph to the right side is frequently related with hyper-thermal conditions. A significantly higher VR relative to illite aggradation is well documented in the KI/VR plot for these samples and can be explained with a faster reactivity of VR during temperature increase (Wolf 1975; Ferreiro Mählmann et al. 2012).

From the KI/VR (OMR) correlation graph, hyper-thermal conditions are not indicated for the rest of the Arosa zone data. The correlation gives a relatively steeper slope of the KI/VR-values (Fig. 14) if compared with the Austroalpine. It was assumed, that the relatively low VR values were caused by retardation of VR through a pressure effect (Ferreiro Mählmann 1994). In experimental petrology there is a controversial discussion about such an effect (see Dalla Torre et al. 1997; Ernst and Ferreiro Mählmann 2004; Le Bayon et al. 2011, 2012a, b). Experimental data verifies pressure retardation at low-grade diagenetic conditions (Le Bayon et al. 2012b) but an enhancement with increasing metamorphic grade. Therefore, including these new findings, the correlation does not indicate a relative increase in pressure in comparison with the KI/VR trends found in the Austroalpine units.

The plot of Fig. 14 shows that the anchizone is defined by different VR values in different tectonic settings. Thus, as mentioned above, the KI/VR relation is strongly controlled by the thermal history, therefore a KI/VR correlation can only be valid for a specific tectonic unit. From Fig. 14 it can be predicted that a single linear to cubic regression formalism, e.g. the most frequently used VR geo-thermometry of Barker and Pawlewicz (1986), or the KI geo-thermometry by Mullis et al. (2002), but also the VR/BR equation used by Schönherr et al. (2007) cannot be applied to determine temperatures (Le Bayon et al. 2011; Ferreiro Mählmann et al. 2012). Thus in this study the KI/VR (OMR) correlation is only used to characterise the paleo-geothermal conditions for a further geodynamic discussion.

From Fig. 14 it is obvious that the zone of diagenesis to anchizone transition in the Arosa zone is not very well-constrained with a calculated value of 2.6 %R_{max}. The high-grade boundary of the anchizone is not determined because orogenic epizone grade is not found in the matrix of the Arosa zone s.s. The diagenesis–anchizone transition in the Lower Austroalpine fragments from the Arosa zone gives a calculated correlation value of 2.9 %R_{max}. According to Ferreiro Mählmann (2001) the diagenesis–anchizone transition in the Lower Austroalpine of the Oberhalbstein region yields a value of approximately 3.1 %R_{max}. For the Lower Austroalpine fragments of the Arosa zone the same tectono-metamorphic, normal thermal orogenic evolution is proposed as for the Lower Austroalpine nappes at the south of Tiefencastel (Fig. 1). Ninety percent of the Arosa zone is metamorphosed at anchizone grade (Fig. 14), thus a further subdivision of the anchizone was done for a better visualization of the trends (Fig. 9). The median KI value of the anchizone (0.25–0.42) with KI = 0.335 Δ°2θ was chosen for the limit between the low and high-grade anchizone.

4.3 K-white mica b dimension determination

As the Lower Austroalpine represents the tectonic units with deeper orogenic burial in the Austroalpine edifice (Dunoyer De Segonzac and Bernoulli 1976), higher pressures conditions can be assumed than in the Upper Austroalpine. This is proven with the k-white mica b cell determination by Henrichs (1993). This should be more accentuated in the Arosa zone representing the suture of the Piemont Ligurian Ocean. Elevated pressures derived from facies indicative minerals are thought to be related to the Cretaceous subduction event during the accretionary collision of the Lower Austroalpine and the South Penninic Piemont Ligurian Ocean (Weissert and Bernoulli 1985; Lüdin 1987; Winkler 1988). In the sedimentary South Penninic rocks no facies indicative minerals were found in this petrological and XRD study. Not having found any indication for high-pressure/low heat flow conditions during metamorphism based on KI/VR correlation another test with the k-white mica b dimension determination was tried.

For preliminary pressure estimation the clay mineral illite/k-white mica *b* method was used, with $b = 6 \times d$ (060, 331). Illites with *b* values lower than 9.020 Å (using the method of Sassi and Scolari 1974) are typical for the South Penninic meta-sediments of the Arosa zone (Fig. 15; Table 1). The values of K-white mica *b* dimensions in the cumulative frequency plot trends to the field of the low to intermediate pressure facies sensu Guidotti and Sassi (1986). Note that only ten samples were used for *b* determination and further investigations are necessary. Quartz was used as an internal standard—all pelite—meta-pelite samples had the restricted assemblage quartz, albite, K-white mica, pyrite ± coalified organic matter, chlorite, and small amounts of calcite. Thus, the *b* determination can be used for a qualitative pressure-indication. The main problem is the low number of data leading to poor statistics because frequently the (060, 331) line was too low in intensity. Nevertheless the few data are important for the discussion of a potential high-pressure event in the South Penninic.

4.4 Paleogeothermic gradients

Another approach to identify metamorphic and tectonic relationship is the reconstruction of metamorphic field gradients or paleo-geothermic gradients in tectonic cross sections (Fig. 10). Assuming, as proposed for the Arosa zone (excluding the Austroalpine fragments) that the main heating event post-dates the main episode of deformation and faulting in a single tectonic complex the resulting rock maturity—and illite aggradation—pattern (VR/KI data set) will reveal an increasing metamorphic grade from the top to the base of that structural complex. In the nappe pile in

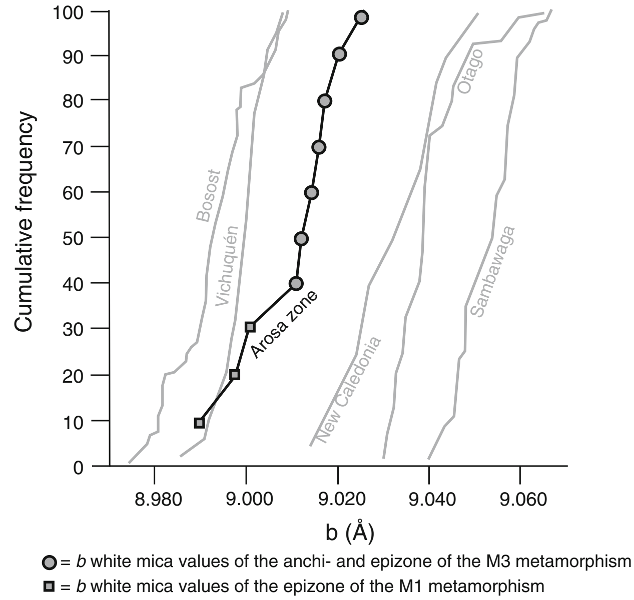

Fig. 15 Cumulative curves of K-white mica b cell dimensions for 10 samples of the Arosa zone (10 samples from the South Penninic rocks). Reference *lines* from the areas Bossot (HT-LP), Otago (LT-HP) and Sanbagawa (LT-HP) are from Guidotti and Sassi (1986), from Vichuquén (HT-LP) is from Belmar et al. (2002) and from New Caledonia (HP-MT overprinted by LT-LP) is from Potel et al. (2006). The samples from the epizone of the Arosa zone, trend to low-pressure facies series, all the other values are indicative for an intermediate pressure facies series

the hanging wall successive episodes of tectonic transport and subsequent reheating did lead to a disturbance of the formerly continuous and relatively simple paleo-geothermic patterns.

Based on KI/VR-data from a profile some characteristic values of the paleo-geothermic regime can be reconstructed, e.g. cal cm^{-1} s^{-1} $°C^{-1}$, mW/m^2 or °C km^{-1}. Generally the coalification gradient (%R_{max} versus depth, mostly %R_{max} per km) is proportional to the paleogeothermic gradient (temperature growth vs. depth) whereas the heat flow (cal cm^{-1} s^{-1} $°C^{-1}$, mW/m^2) is inversely proportional to it (Robert 1988; Sweeney and Burnham 1990). The coalification gradients shown in Fig. 10 (but also the KI gradients) can therefore be regarded as a representation of the paleo-geothermal gradients (Johnsson et al. 1993). The rate of coal rank increase is mainly dependent on the geothermal gradient but also partly on the heat conductivity of the rocks (Teichmüller and Teichmüller; in Stach et al. 1982, p. 56).

In the studied area sediment facies differentiations within formations of more than 300 m thickness (excluding the Hauptdolomite, see Fig. 2) are not pronounced and the overall lithological effects are therefore negligible. Thus the VR-data in our study area is useful for the calculation of paleo-geothermal gradients. The paleo-geothermic gradients attributed to the profile (Fig. 10) were obtained by linear or cubic regression analysis of VR versus depth. Depth was chosen as the independent variable.

4.4.1 KI/VR gradient from the Austroalpine nappes

The present multi-phase pattern of the entire tectonic pile is fairly complex (Figs. 10, 16). The temperatures plotted in Fig. 16 in the Austroalpine are attributed to two different metamorphic events. M1 is attributed to a hyper-thermal syn-sedimentary, pre-orogenic metamorphism of Permian to Jurassic age and M2 is the consequence of nappe tectonics and a hypo-thermal orogenic metamorphism of Cretaceous age (Ferreiro Mählmann 1994, 1995, 2001). Note, that the Jurassic M1 pattern of the Arosa zone attributed to the oceanic metamorphism (see Figs. 9, 13) is too local to bee presented in Fig. 10.

4.4.2 KI/VR gradient from the Arosa zone

Ferreiro Mählmann (1994) attributed the metamorphic pattern of the Arosa zone and the first deformation (Froitzheim et al. 1994) to the subduction of Piemont-Ligurian ocean under the Austroalpine–Apulian plate (see also Trümpy 1980; Ring et al. 1989; Henrichs 1993; Froitzheim et al. 1994). However data published by Ferreiro Mählman (1995, 1996, 2001), Nievergelt et al. (1996) excludes a continuous, Cretaceous, post-mélange tectonic maximum heating of the entire South Penninic together with the Austroalpine nappe system. This can be concluded because the diagenetic to anchizonal maturity level of the Arosa zone cannot be in accordance to the KI/VR trends of the M1 or M2 metamorphic events in the Austroalpine. Therefore the Arosa zone in Mittelbünden was not overprinted by the M1 and M2 heating event (Frey and Ferreiro Mählmann 1999).

The mélange tectonics in the South Penninic Arosa zone sediment rocks do not appear to deform VR isograds in the area of Davos–Arosa–Klosters (Figs. 9, 10, 11, 12). Therefore, the top–down trend found in the Arosa zone is now considered as a consequence of the M3 event and the formation of the mélange ceased prior to maximum M3 metamorphism.

For the South Penninic rocks a very conservative estimation of KI/VR gradient was applied. A mean gradient is calculated (cf. Fig. 15) that is about 25 % higher than the difference between the gradient of the Vichuquén Basin, Chile (Belmar et al. 2002) and the Austroalpine (Ferreiro Mählman 1996) as shown on Fig. 14. Also a simple and conservative thermal gradient adjustment was applied to the matrix sediments of the Arosa zone, yielding a thermal gradient of 22 ± 2 °C and a barometric gradient of 300 bar per vertical km. These gradients are comparable to those of the hypothermal M2 event of the Austroalpine (Ferreiro Mählmann 2001).

Main deformation in the Arosa zone and the Austroalpine fragments show the same micro-structural elements (Froitzheim et al. 1994). M3 can be cogeneticaly related to that main post-mélange deformation, including the Austroalpine fragments. That deformation will be considered as D3 in the following.

4.4.3 KI/VR-gradient from the footwall of the Arosa zone

On the lower margin of the Arosa zone a complex metamorphic discontinuity is found. Adjacent Eocene sedimentary rocks (Trümpy 1980) of the Falknis nappe are of high diagenetic grade. The metamorphic grade and history of the Middle Penninic units differ from the evolution in the Arosa zone, e.g. in the area west of the Weissfluh (Gruner 1980; Frey et al. 1980) and in the north at the southern border of the Rätikon mountains (Ferreiro Mählmann 1994). But this metamorphic hiatus between the Arosa zone and the Middle Penninic domain cannot continuously observed around the Prättigau half-window (Frey et al. 1999; Frey and Ferreiro Mählmann 1999; Oberhänsli et al. 2004). Altogether, these observations lead to the conclusion that the top–down trend of anchizonal grade in the Arosa zone must be explained with another heating and is therefore also not related to the thermal event observed in the Middle Penninic nappes (M4, Fig. 10). In the discussion we will refer later to D4 as the Turba Mylonite phase.

4.5 Metamorphic map with new data from the Arosa zone

This chapter considers the wealth of sediment data obtained by the methods mentioned in this paper. Sediments in the Upper Austroalpine, the Lower Austroalpine and the Arosa zone yield clearly different KI-VR-T °C-mineral association relations (see Figs. 6, 7, 8, 9, 10, 11, 12, 13, 14). These patterns are partly controlled by the duration of the heating episode and by the state of final equilibrium reached by the indicative reaction systems used to estimate metamorphic grade.

The metamorphic map (Fig. 16) shows the maximum grade reached since the sedimentation. For temperature estimations given for the Upper and Lower Austroalpine units see Ferreiro Mählmann (1996, 2001). It is also attempted to attribute time to the metamorphic events (pre-orogenic, Cretaceous, Tertiary) based on the data presented.

KI/VR correlations shown on Figs. 10 and 14 are hetero-chronic (M1 to M3). All the younger events (M4 and M5) are simplified together as Tertiary (footwall of the Arosa zone). Based on the correlations, tectonic and stratigraphic observations the following, above-mentioned heating episodes can be distinguished:

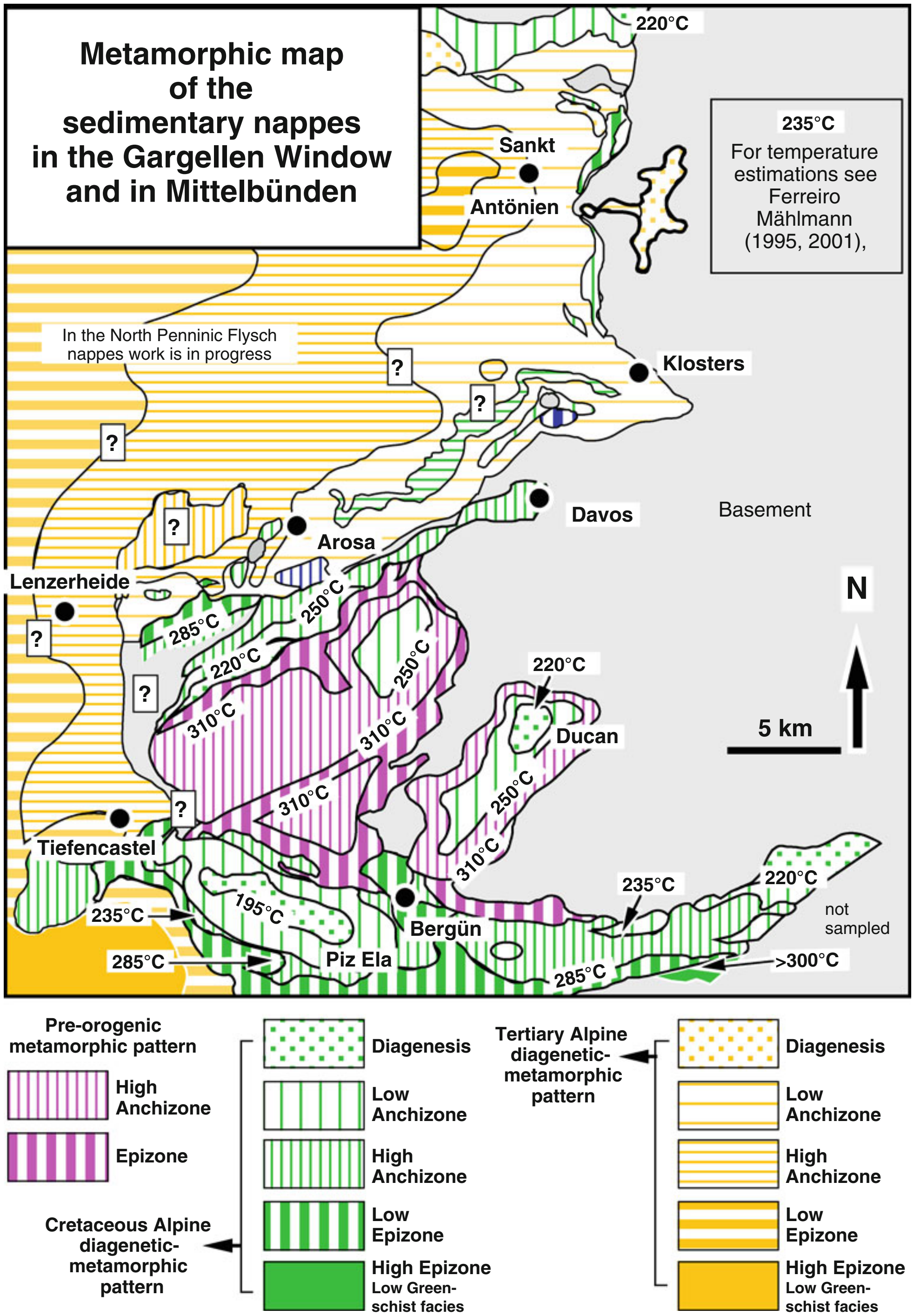

Fig. 16 The metamorphic map of Mittelbünden and the Prättigau half window. Three different metamorphic patterns of different age were mapped due to the tectono-metamorphic relationships established. Temperatures are derived from multi-methodical correlations and maturity modelling techniques (see text). The *blue-signed areas* show the locations were rocks of the South Penninic are strongly affected by Jurassic oceanic metamorphism. The *question marks* show results not well established or little constrained

1. M1: a Permian to Jurassic pre-orogenic, diastathermal event in the Upper Austroalpine (very pronounced in the Silvretta nappe s.s.). This is marked in the metamorphic map with a lavender colour (M1a). Spots with an oceanic metamorphic pattern (also pre-orogenic) in the Arosa zone are marked in blue (M1b).
2. M2: a Cretaceous orogenic pattern (tectonic burial) in the lower imbrications of the Silvretta nappe s.l. and in

the Lower Austroalpine (e.g. Tschirpen nappe) is shown in green.

3. M3: a Tertiary thermal episode post-dating the deformation of Cenomanian and Turonian flysches (Richter 1957; Winkler 1988) and pre-dating Oligocene extension (Turba Mylonite phase, D4, Nievergelt et al. 1996) in the Arosa zone. The Coniacian flysch age postulated for the Arosa zone is not well constrained (Lüdin 1987). M3 is related with the Eocene “Blaisun phase” D3, but post-dates the mélange deformation.
4. M4: a possible heating event affecting Middle Penninic units in Tertiary, shown in yellow on the map.
5. M5: below the Austroalpine edifice and below the South Penninic nappes a Tertiary heating episode did overprint the North Penninic nappes. Paleocene to Eocene sediments in the Prättigau (Nänny 1948; Thum and Nabholz 1972), in the Oberhalbstein (Ziegler 1956; Eiermann 1988) and in the Engadine window (Rudolph 1982) are affected by this latest metamorphic episode (cf. Fig. 16).

5 Discussion

5.1 Temperature estimations from the KI/VR correlation

The KI/VR relation is strongly controlled by thermal histories therefore a KI/VR correlation is only valid for a specific tectonic unit (Kübler et al. 1979; Frey et al. 1980; Ferreiro Mählmann 1994). The anchizone defined by KI correlates with different VR values in different tectonic settings. A calibration through mineral data and including maturity modelling techniques (TTI, EASY%R_0) gives obviously different temperatures for KI-zone boundaries (Fig. 10) in the Upper Austroalpine (Ferreiro Mählmann 1994) and in the Lower Austroalpine nappes (Ferreiro Mählmann 2001). A KI–VR-T °C generalisation, widely published in the literature, should clearly be avoided (Kisch 1980, 1987, 1990; Kisch and Frey 1987). Such generalizations, formerly also applied to the study area (Ring et al. 1988, 1989; Henrichs 1993; Kürmann 1993) caused a temperature overestimation of KI values and other IC measurements. The overestimation of metamorphic grade by Ring et al. (1988), referring the Arosa zone to be of low grade (greenschist facies) is due to an error propagation caused by the generalization and the use of non-calibrated IC values. Because also a compilation of mineral data and new maturity modelling is planned for the Arosa zone, no temperature values will be proposed in this study. Therefore, the forward discussion will be restricted to a metamorphic grade determination related to KI-zones.

5.2 The northern tectonic limit of the Arosa zone based on structural-metamorphic arguments

From the Tilisuna area (Fig. 1) to the north fragments of Lower Austroalpine provenience with high anchizonal to epizonal grade are missing (except a phyllitic slice in Lichtenstein, Ferreiro Mählmann 1994). The Northern Calcareous Alps did not have a Lower Austroalpine base when thrusted on the South Penninic tectonite of the mélange zone. Due to the differences in fragment composition (Lechtal nappe and subsilvrettide fragments) and in accordance with the different grade of metamorphism (high grade diagenesis) and also based on the differences of the non-metamorphosed to low grade sub-greenschist facies ophiolitic oceanic rocks (Lüdin 1987), it is reasonable to differentiate between the Arosa zone in the south of the Rätikon mountains and the Walsertal zone to the north according to Winkler (1988). The metamorphic (high grade diagenetic) pattern in the mélange matrix in the Walsertal zone shows a geographically homogeneous topographic top–down trend of increasing diagenesis. The Walsertal zone contains low grade diagenetic flysches bounded by brittle tectonites without clay mineral neoformations (Ferreiro Mählmann 1994) and this is again different to the Arosa zone s.s. defined in this paper. This deformation–diagenesis relationship in the Walsertal zone (see Fig. 6) is attributed to late movements in the mélange zone during retrogressive temperature and pressure conditions.

5.3 The southern tectonic limit of the Arosa zone based on structural-metamorphic arguments

The lithofacies and the internal structure of the Arosa zone are characteristic of mélanges in subduction complexes (Ring et al. 1990). The Arosa zone is a mélange unit composed of South Penninic and Austroalpine rocks (at the basis also Middle Penninic rocks). The mélange is postulated to be Cretaceous to Tertiary in age (Trümpy 1980). As indicated in earlier chapters it is to assume that the main deformation and mélange formation in the Arosa zone could correspond to D3 of the higher tectonic units (Fig. 17). Because D2/M2 is of Late Cretaceous age (70 ± 5 Ma, Ferreiro Mählmann 2001) and the tectono-metamorphic pattern in the Arosa zone is fairly different, a mélange formation of Cretaceous age seems problematic. This needs therefore some more specification referring to the Arosa zone in particular:

First of all, an old concept must be excluded. Many authors, have postulated a maximum metamorphism following nappe emplacement in Tertiary times (45–25 Ma; the so called ‘Lepontine metamorphism’) and affecting the entire nappe pile in Eastern Switzerland (e.g. Bearth 1962; Niggli and Zwart 1973; Frey et al. 1974; Trommsdorff and

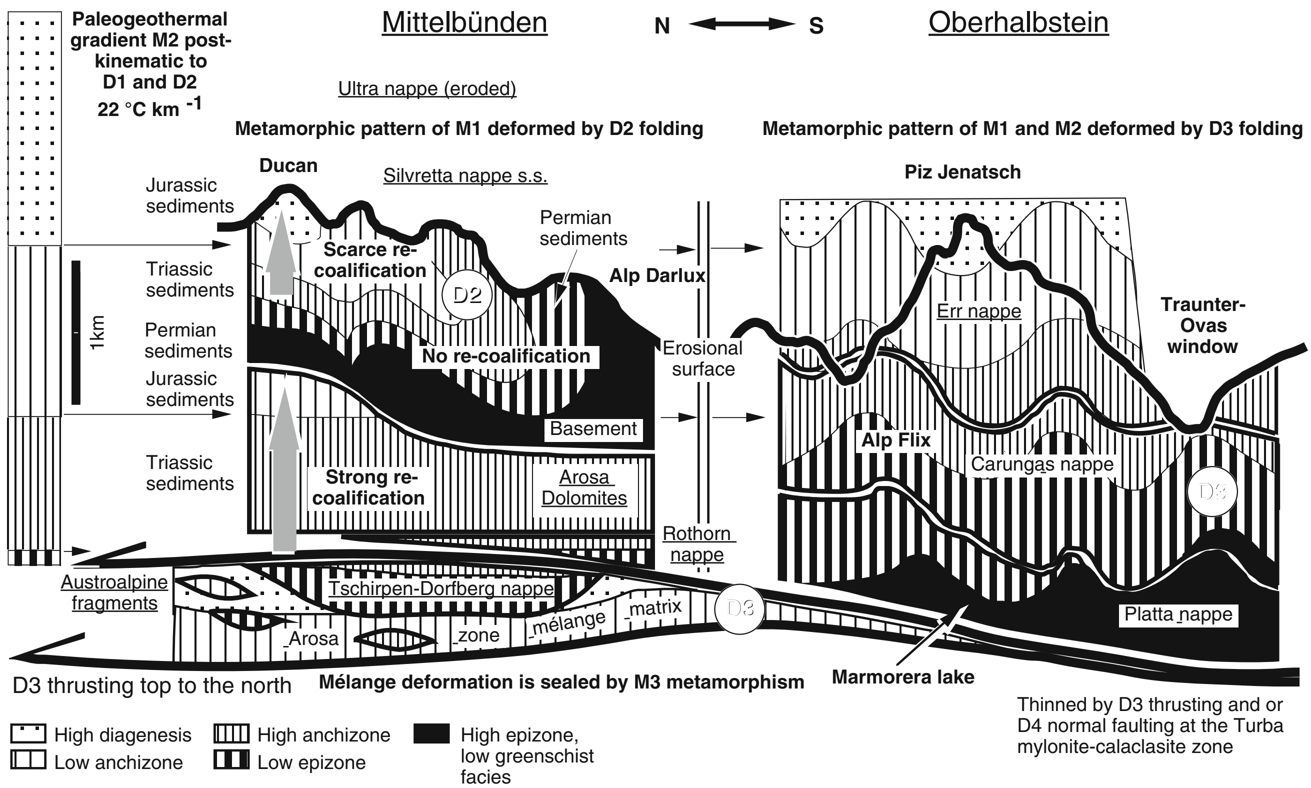

Fig. 17 Tectono-metamorphic model during D3 deformation. In the Oberhalbstein the nappe stack and the metamorphic zones are folded. In the Mittelbünden area the Austroalpine nappe stack is partly re-coalified by the second metamorphism, but thrusted without major deformation of the second metamorphic pattern on the Arosa zone. The second metamorphism is characterized by a metamorphic inversion to the Arosa zone. In the Arosa zone, Austroalpine fragments dismembered during D3 show an inherited transported metamorphism. After ceasing of the mélange formation the matrix of the mélange was overprinted by a third thermal event. The last event was the result of thermal re-equilibration after D3 thrusting and crustal thickening

Dietrich 1980; Ring et al. 1989; Kürmann 1993). More recent low temperature studies of sedimentary units by Ferreiro Mählman (1994, 1995, 1996, 2001) and vitrinite thermal modelling (Ferreiro Mählmann 2001) revealed a much more complex thermal history in the Austroalpine of Eastern Switzerland and the study area in particular (see Sect. 5 above and Fig. 16). This very complex diagenetic–metamorphic pattern contrasts strongly to the former model of a continuous north–south increase of metamorphism from diagenesis to medium grade metamorphism (Niggli and Zwart 1973; Ring et al. 1989). Frey (1986) included only the Oberhalbstein area to be affected by the Tertiary Lepontine metamorphism, but this is only valid for some parts of the lower structural Penninic units in the footwall of the Turba mylonite zone (Nievergelt et al. 1996).

Furthermore the metamorphic history of the Upper Austroalpine differs from the thermal evolution of the Lower Austroalpine in the hanging wall of the Turba mylonite zone. Here and in the lower imbrications of the Silvretta nappe, peak conditions of a Cretaceous low anchizonal climax is syn-kinematic to D1 and D2 (Fig. 17), both of Cretaceous age (Ferreiro Mählmann 1996, 2001). Peak temperature metamorphism (M2) in the Lower Austroalpine units in the south (Oberhalbstein, Fig. 17) occurred during Cretaceous Alpine orogeny (Guntli and Liniger 1989; Handy et al. 1996) and this is also reflected in the new metamorphic maps (Frey et al. 1999; Oberhänsli et al. 2004). Dietrich et al. (1974) recognized that petrographic data in the Arosa–Platta mafics suggest a progressive increase of grade in the field, but they pointed that in some places tectonic mixing within the ophiolite thrust sheets may have destroyed that picture.

The tectono-metamorphic patterns found, are in conflict with a simple model of a coherent South Penninic tectonic complex with a trivial, unique and synchroneous deformation history. Cornelius (1950) described slices of Austroalpine Permo-Mesozoic rocks between the Platta nappe and the basement of the Austroalpine Err-Bernina nappe and Carungas nappe. Ring et al. (1990) interpreted large part of the Carungas nappe as part of the Arosa zone due to ophiolitic intercalations, the structural position, and the mélange character, but also in deeper structural levels complex mélange units are known (Cornelius 1950; Dietrich 1969, 1970). Frisch et al. (1994) considered the Platta nappe as a separate tectonic unit located immediately beneath the Arosa zone (Ring et al. 1989, 1990; Dürr 1992). Also a coherent "Eo-Alpine tectonic nappe structure" was postulated (also Staub 1934, 1971; Cornelius 1950; Tollmann 1959, 1977). It is preferred to use the term Cretaceous Alpine orogeny in this context.

Nevertheless it is evident that early D1 deformations top west (Late Cretaceous) are frequent in the Carungas nappe and also in the fragments of the Arosa zone (Ring et al. 1989; Biehler 1990; Nagel 1996). Together with other authors we conclude that older thrust sheets (imbricated fragments) rode passively on top of younger ones. Thus thrusting began in the uppermost unit. But the S–N (D3) deformation overprints an E–W to SE–NW (D1) deformation and some extensional structures (D2). Also in the Austroalpine fragments from the Arosa zone occasionally early stage D1-fabrics are preserved and characterised by an angular unconformity with the main mélange cleavage (D3).

The main problem results in the ambiguous correlation of the D1 to D3 deformation series as defined by Froitzheim et al. (1994) with the deformation series according to Ring (1989) and Frisch et al. (1994). Ring et al. (1990) differentiated three deformations in the Carungas–Platta unit (their southern extension of "Arosa zone"). Thus in the Platta nappe the third deformation was described as an extensional one. It is not clear, if the extensional one can be related to the Ducan phase (D2) or Turba phase (D4) sensu Froitzheim et al. (1994) and Nievergelt et al. (1996) because the sub-parallel SE–NW D1 and D2 phases were not recognized (see Nievergelt et al. 1996). Ring et al. (1990) mentioned brittle fractures related to D3, but Turba mylonite structures from Tinizong to the south into the Bivio–Margna region (Nievergelt et al. 1996) are ductile.

The metamorphic studies are strongly supporting the model of Froitzheim et al. (1994) and Nievergelt et al. (1996). The Platta nappe was metamorphosed to peak conditions (D1/M1 and D2/M2) in post nappe tectonic time together with the Lower Austroalpine (Ferreiro Mählmann 1994, 2001). The Platta nappe (also part of the dismembered lower Carungas nappe) is located in the same structural level as the Arosa zone in the north, but in the Arosa zone the thermo-tectonic history and mélange formation is post D2/M2. Thus in both South Penninic units metamorphism peaked at different times. The D2 mélange formation in the Carungas–Platta unit and D3 mélange formation in the Arosa zone s.s. (as defined in this paper) probably documents two propagating deformation stages to the north. The deformation in the Platta nappe is of Cretaceous age and deformation moved to a lower structural level after D2, into the Arosa zone (Cretaceous to Eocene, D3; see below). The Lower Austroalpine and Platta nappe was at the same time deformed by D3 folding (Blaisun phase). The Platta nappe is a part of the "orogenic lid" (Laubscher 1983) since D3 and the limit of the Arosa zone s.s. to the south has to be located at the village Tiefencastel. At that locality a tectonic boundary was suggested by Ferreiro Mählmann (1995) coinciding with the inverted metamorphism at the boundary of the Platta-Austroalpine nappe pile to the Arosa zone s.s. (see Fig. 17).

5.4 South Penninic oceanic metamorphism in the Arosa zone

In the Arosa zone remnants of the South Penninic (Piemont-Ligurian) oceanic plate (according to Trümpy 1975) are also preserved. Jurassic oceanic basement rocks and their pelagic sedimentary cover of Jurassic to early Cretaceous age are highly dismembered and imbricated as slices with distal continental margin rocks (Lower Austroalpine) do occur. Imbrications with lithologies of continental margin origin occur preferentially along the upper and lower boundary of the Arosa zone. Although the major lithologies of an ophiolite suite are present in the South Penninic mélange in the structural middle part of the Arosa zone, no complete section through oceanic crust has been preserved. These oceanic sequences therefore record the tectono-sedimentary evolution of a transform domain along the northern margin of the Austroalpine–Apulian promontory (Weissert and Bernoulli 1985).

At three locations of the Arosa zone (Hörnli, Verborgene Wäng, Parsenn) oceanic metamorphism was postulated to explain the epizonal grade in the oldest red clays of the radiolarite formation and also the rapid decrease of VR and the larger spread of KI (Figs. 9, 12, 13) in stratigraphic younger formations. Also a hydrothermal hornfels facies metamorphism is well preserved by the typical mineral paragenesis (Ferreiro Mählmann 1994 and references therein). Peters (1963) postulated an Alpine genesis of the hornblende, diopside and garnet bearing rocks. This was also another reason for the concept of a post-kinematic upper Eocene–Oligocene high T-metamorphism of the Arosa zone (Niggli and Zwart 1973; Dietrich et al. 1974). Because these occurrences of hornfelses are very local with some chemically appropriate educt rocks in the vicinity (Fig. 13), which did not evolve to hornfels facies rocks, former postulates of high epizonal or even mesozonal regional overprint of the Arosa zone can be ruled out. The high-grade scenario is only of historical interest, but it is the reason of prevailing discussions and overestimations of metamorphic grade (Ring et al. 1989; Frisch et al. 1994; Bevins et al., 1997). The local occurrence of epizonal red clays is just a remnant of oceanic metamorphism strongly blurred by younger orogenic phases in most places. There is also a geo-chronological clue to the Jurassic onset of that oceanic metamorphism (M1 in the oceanic plate): Peters and Stettler (1981) reported a

$^{39}Ar/^{40}Ar$ plateau age of 165 Ma yielded by a phlogopite from the South Penninic serpentinite complex of Totalp (Davos).

Taking into account and comparing the evolution of the Platta nappe and the Arosa zone the most cited assumption, that the mélange formation and displacement of the Austroalpine is of Cretaceous (Laubscher 1970; Weissert and Bernoulli 1985) or Cretaceous to Tertiary age (Ring et al. 1988) has to be questioned and also revised by a more differentiated history. Even older models with a static post-dynamic overprint of the entire tectonic nappe pile in Tertiary ("Lepontine") are in the same manner only of historic interest in this context.

6 Conclusion

The new model with a complex, multiphase evolution is based on the following arguments:

1. Folding, thrusting and normal faulting in the Tschirpen-Dorfberg nappe units is of Cretaceous age (D1 and D2 are inherited and characterized by transported metamorphism) and also thrusting of the Platta nappe pre-dates the D3 mélange formation of Arosa zone s.s.
2. Deformations like imbrications, folding and mélange tectonics are sealed by the high diagenetic, anchizonal to partly epizonal metamorphic pattern on the South Penninic sedimentary rocks. Deformation may have been initiated during the latest Cretaceous (post D2; post 65–75 Ma, Ferreiro Mählmann 2001) at D3. Therefore, main dislocation to the north is of Tertiary age (D3 for Arosa zone s.s.) and result of a propagating deformation front.
3. Syntectonic sediments were partly unlithified and subjected to soft sediment deformation, this coinciding with early stages of subduction and accretion in the Lower Cretaceous (Winkler 1988).
4. Middle and North Penninic units with Eocene–Oligocene rocks (Nänny 1948; Trümpy 1980; Steinmann 1994) are constituents of the footwall thrust of the Arosa zone mélange. Thus the D3 accretional deformation was active until Eocene–Oligocene time, but shifted to the lower structural part of the Arosa zone s.s. and into the imbrications of the Lechtal nappe.
5. The metamorphic top-down and north–south pattern of the South Penninic Arosa zone shows transported metamorphism on the Oligocene rocks of the lower Penninic units (Figs. 10, 16, 17). Therefore metamorphic overprint of Arosa zone s.s. (M3) must be earlier than Oligocene.
6. The pattern M4 is restricted to the Middle Penninic. The metamorphic pattern of the Middle Penninic is limited by a metamorphic gap, in places an inversion, in places a progressive hiatus at the tectonic limits of that structural unit. The progressive hiatus ("colder" rocks on top of relatively "hotter" rocks) is caused by an extensional fault in the Gurgaletsch area (SW of Arosa). Weh (1998) found micro-structural evidences that the normal faults are related with the Oligocene Turba mylonite zone (D4).
7. Anchizonal values in the Arosa zone, showing a maturity of 3.3 bis 3.9 % R_{max} and assuming a similar gradient for M3 as for M2 (22 °C km^{-1}) can only be explained by an important crustal thinning. M2 was the result of a crustal thickening with a load caused by the Ultra nappe (Ferreiro Mählmann 1994). Extension in the Upper Austroalpine is not prominent. If the low anchizonal overprint in the Arosa zone was caused under the present nappe pile an erosional difference has to be postulated (erosion of the Ultra nappe). But >9.5 km (Ferreiro Mählmann 1995) of eroded rocks in a short time span of 20 Ma results in a high erosion rate of about 500 m Ma^{-1}. Considering also the diagenetic flysches near Klosters and Madrisa, the erosion gap needed to explain the data would further increase!
8. At the basal thrust of the Silvretta nappe to the Tschirpen-Dorfberg nappe a progressive hiatus is again evident. This can be the result of a normal fault between the Upper and Lower Austroalpine (Ferreiro Mählmann 1994) but structural clues are missing. This metamorphic hiatus may be the northernmost continuation of the brittle part of Turba mylonite zone as normal fault and is therefore probably related to D4 (D2 according to Weh 1998). Tectonic elements related to Turba mylonite zone and its extension, the Gürgaletsch shear-zone (Weh 1998), show a transition from ductile shear to brittle faults, but could not yet be convincingly recognized in the northern part of the study area. We assume that the Arosa zone basis is the northern prolongation of the Turba mylonite zone in a very brittle structural position: (1) due to the tectono-metamorphic pattern, (2) due to the fact that in brittle parts normal faults propagate to higher structural units, and (3) Turba elements of normal faulting are known in the south of our study area in an equivalent nappe pile level (Weh 1998).

Using all the structural and stratigraphic information, the D3–M3-event observed in the Arosa zone (geotherm 3 in Fig. 10) can be bracketed into the time span between the Austroalpine D2–M2 event and the middle Tertiary M4 event in the Middle Penninic as well as the Oligocene "Turba phase" (D4). Probably, some 0–6 km of crust was displaced by normal faults and some 0–6 km was eroded on top of the nappe pile at the same time. Independent of the assumptions for a missing crust profile, the Turba phase structures deform the D3/M4 patterns. All together it can be concluded that metamorphic pattern in the Arosa zone is of Paleogene age.

Generally the metamorphic evolution model of Dietrich et al. (1974) was much closer to the findings of our present study than all the more recent studies. These authors proposed a multi-metamorphic overprint. After a Jurassic to lower Cretaceous oceanic metamorphism, followed by a Cretaceous to Eocene subduction metamorphism (low T–high P) a post-kinematic upper Eocene–Oligocene metamorphism was assumed. As a conclusion, it is possible to differentiate between a tectono-metamorphic history in the Cretaceous including the Austroalpine and the Platta nappe from a Tertiary history in the footwall of Austroalpine–Platta complex including the rocks of Arosa zone s.s. (South Penninic). Clear high-pressure indications could not be found in the Arosa zone.

A low-pressure facies or partly an intermediate pressure facies according to Guidotti and Sassi (1986) is also indicated by the illite *b* cell values of 8.989–9.022 Å sampled from the South Penninic lithologies. The values are much lower than the pressure of 3.0–6.5 kbar proposed according to Ring et al. (1989) and lower than K-white mica *b* cell values from the Lower Austroalpine and Upper Austroalpine (Henrichs 1993). Similar to the TTI maturity model used by Ferreiro Mählmann (1994) a maximum burial of about 9 km is determined and this is also in agreement with the K-white mica *b* cell values and geothermal reconstructions. If any early Alpine high-pressure event would have controlled the illite *b*-value, the *b* cell dimension is robust to later greenschist facies re-equilibration, also if the KI would adjust to the new P–T conditions (Potel et al. 2006).

Based on studies of ultramafics in the Arosa zone, Peters (1969) concluded, that the geological setting, the paleogeography and tectonics but also the whole complexity of metamorphism should be considered. This is still not really ascertained for all the South Penninic nappes and additional work is needed. A comparison of mineral data from literature, own data from ultramafics and the new metamorphic pattern presented is planed, including unpublished data of D. Robinson, R. Bevins, M. Frey, S. Th. Schmidt and P. Árkai.

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