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Provenance of the Upper Cretaceous to Eocene Gosau Group around and beneath the Vienna Basin (Austria and Slovakia)

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Abstract Chemistry of detrital garnets, chrome spinels and tourmalines of 30 selected samples in combination with the general heavy mineral distribution from 523 sandstone samples of the Upper Cretaceous to Eocene Gosau Group of the eastern part of the Eastern Alps and the western West Carpathians result in an advanced picture of sedimentary provenance and palaeogeographic evolution of that area. Garnets from Coniacian to Campanian sediments are partly derived from a metamorphic sole remnant of Neotethys ophiolites to the south. Tectonically high ophiolitic nappes, later on completely eroded, supplied mainly the paleogeographically southern Grünbach and Glinzendorf Gosau basins with ultramafic detritus, represented by chrome spinels of a mixed harzburgite/lherzolite composition, whereas no direct indications for a northern ophiolitic source, the Penninic accretionary wedge to the north of the Gosau basins, could be found. In the younger part of the Gosau basins fill, from the Maastrichtian to the Eocene, only almandine-rich garnets could be observed suggesting a southern provenance from low-grade metamorphic metapelites of exhuming Austroalpine metamorphic complexes. Ophiolite detritus is reduced in the Maastrichtian and disappears in the Paleogene.

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1 Introduction

The well documented sedimentation history of the Upper Cretaceous to Paleogene Gosau Group as part of the Northern Calcareous Alps (NCA) reflects ongoing tectonic processes at the active margin of the Austroalpine microplate during the Late Cretaceous (e.g. Wagreich 1993a; Faupl and Wagreich 2000). Several studies of heavy mineral assemblages contributed significantly to the understanding of detrital sources and hence the paleogeographic evolution of that area, including the NCA, during the Late Cretaceous to Paleogene (Woletz 1967; Faupl 1983; Stattegger 1986, 1987; Winkler 1988; Wagreich and Marschalko 1995). However, reconstructions of provenance are still puzzling and paleogeographic positions or correlations of various Gosau basins at the transition zone between the Eastern Alps and the Western Carpathians are still under debate (e.g. Wagreich and Marschalko 1995; Hofer et al. 2013). In such a case, single grain chemistry of heavy minerals provides an important tool in provenance analysis techniques (e.g. Henry and Guidotti 1985; Pober and Faupl 1988; Morton 1991; Von Eynatten and Gaupp 1999; Weltje and Von Eynatten 2004; Mange and Morton 2007; Meinhold et al. 2009; Aubrecht et al. 2009) but so far this was only applied selectively to the Gosau Group (Pober and Faupl 1988). The present study gives an updated overview of Gosau Group heavy mineral assemblages in the eastern part of the Eastern Alps and the western part of the West Carpathians using published and new data, and gives new insights from single grain chemistry with the goal to improve the provenance concept and to reconstruct the paleotectonic evolution of these areas during Late Cretaceous to Paleogene times.

2 Geological setting

The Gosau Group of the NCA comprises Upper Cretaceous to Paleogene sediments overlying folded and faulted Permian to Lower Cretaceous units (Wagreich and Faupl 1994). The NCA itself can be subdivided into several thrust units, from base (north) to top (south), including the Bajuvaric, Tirolic and Juvavic (or higher) nappe units according to the classical concepts (e.g. Mandl 2000; Schmid et al. 2008). The Gosau Group can be subdivided into a fluvial, lacustrine to shallow-marine Lower Gosau Subgroup (Turonian to Santonian/Campanian, in some areas up to Maastrichtian), and a deepwater dominated Upper Gosau Subgroup, locally starting in the Upper Santonian and reaching the Eocene as defined by Faupl et al. (1987), Wagreich and Faupl (1994) and Faupl and Wagreich (1996). Gosau Group sedimentation started in local basins (e.g. Sanders 1998; Wagreich and Decker 2001), giving way to larger and more widespread sedimentation in the Upper Gosau Subgroup (Wagreich 1993a; Krenmayr 1999; Wagreich et al. 2011).

Various Gosau Group deposits are exposed in the easternmost part of the Eastern Alps (Austria) in the area of Gießhübl near Vienna and Grünbach at the south-western margin of the Vienna Basin as well as in the westernmost part of the West Carpathians at the area of Brezová and Myjava, Slovakia (Fig. 1; see also Wagreich and Marschalko 1995). Beneath Neogene sediments of the Vienna Basin, NE–SW-striking Gosau Group deposits can be traced over several wells down to depths of more than 4,000 m (Wessely 1993; Zimmer and Wessely 1996). During the post-Eocene Alpine orogeny, sediments of the Gosau Group were folded and are today arranged in structurally complex synclines on different tectonic nappes of the NCA and the Carpathians (Fig. 1). From north to south the Gießhübl Syncline, the Prottes Gosau Group, its Slovakian equivalents at Studienka below the Neogene Vienna Basin fill, and Brezová-Myjava (here for convenience summarized as Slovakian Gosau basin), the Glinzendorf Syncline and the Grünbach Syncline have been defined (Plöchinger 1961; Wessely 1974, 1984, 1992, 1993, 2000, 2006; Zimmer and Wessely 1996). These various Gosau Group occurrences are interpreted as four, partly separated basins at least during deposition of the Lower Gosau Subgroup (Wagreich and Marschalko 1995; Wagreich and Decker 2001). We therefore refer to these various Gosau basins in the following sections.

3 Stratigraphy

3.1 Gießhübl basin

Strata of the Gießhübl basin (Coniacian to Palaeocene), which is part of the Bajuvaric Nappe system (Lunz Nappe; Faupl and Wagreich 1996) of the NCA and overthrusted by the Tirolian nappe system, starts with shallow-marine sandstones and breccias of Coniacian/Santonian age. The overlying Upper Santonian–Campanian to Lower Maastrichtian Nierental Formation (Wessely 2006) is already attributed to the Upper Gosau Subgroup and comprises mainly calcareous marls of a pelagic to hemipelagic slope facies (Krenmayr 1999; Wagreich and Krenmayr 2005; Wagreich et al. 2011). The main part of the Gießhübl basin fill is represented by the Gießhübl Formation (subdivided into a Lower, Middle and Upper Gießhübl Member), a massive, deep-water turbidite to mass-flow-complex, deposited below the calcite compensation depth (Faupl and Sauer 1978; Wagreich 2001a), with a temporal extent up to the Thanetian (Wessely 1992). Gosau Group sediments of the Gießhübl basin are reported from below the Neogene of the Vienna Basin (Wessely 2006) and crop out in the areas of Gießhübl, Alland and Altenmarkt (Fig. 1). Equivalents of the Gießhübl basin are exposed around Lilienfeld, where the Gosau Group comprises Coniacian–Santonian Lower Gosau Subgroup units followed by the hemipelagic–pelagic Nierental, Spitzenbach and Gießhübl Formations of the Campanian to Maastrichtian (Wagreich 1986; Wagreich et al. 2011).

The Prottes Gosau Group (Coniacian to Palaeocene) is a small, isolated, mainly conglomeratic subcrop complex transgressing the Mesozoic of the Tirolic nappes from underneath the Neogene of the Vienna Basin at Prottes to outcrops at Altenmarkt (Wessely 2006).

3.2 Grünbach basin

The Grünbach basin (Santonian to Palaeocene) is situated at the south-western margin of the Vienna Basin, mainly on Tirolic units in the area of Grünbach—Neue Welt (Plöchinger 1961; see Fig. 1). The connection to Upper Cretaceous rocks beneath the Neogene fill of the Vienna Basin is controversial (e.g. Hofer et al. 2011). The base of the Grünbach Gosau Group comprises Upper Santonian red conglomerates and breccias of alluvial fan environments (Kreuzgraben Formation; e.g. Erkan 1972; Summesberger et al. 2002). A transgressive trend is documented by the overlying shallow-marine Maiersdorf Formation that contains calcareous breccias, minor sandy limestones (including corals, brachiopods and gastropods) and reef-forming rudist limestones (Summesberger et al. 2002). The limnic to marginal-marine Grünbach Formation of the

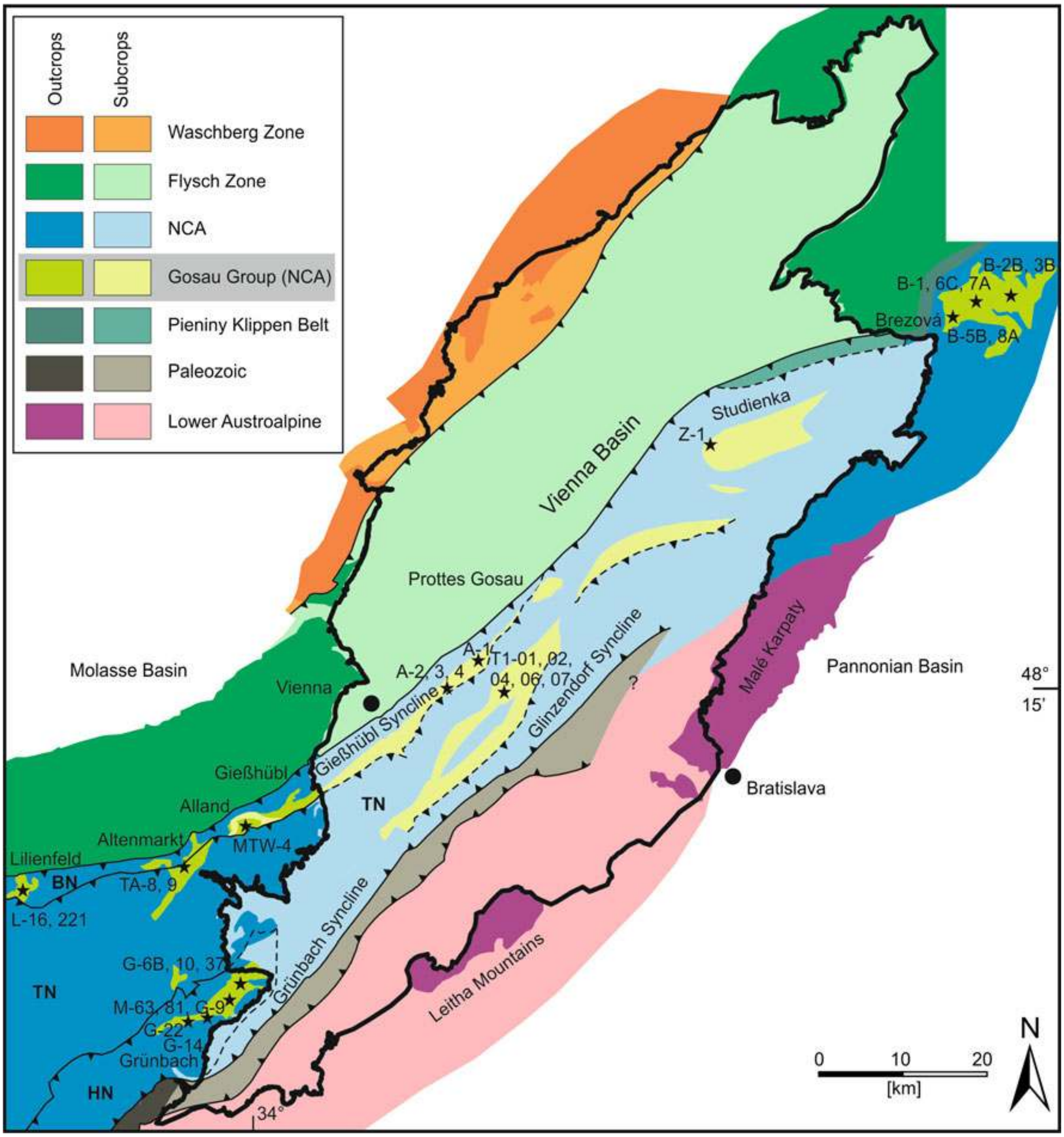


Fig. 1 Simplified geological map of the eastern part of the Eastern Alps, the pre-Neogene basement of the Vienna Basin, and the western part of the West Carparthians, with Gosau Group situated at the border of (in the Northern Calcareous Alps), and underneath, the Neogene Vienna Basin (modified after Wessely et al. 1993; Wagreich and Marschalko 1995; Zimmer and Wessely 1996). Sampled locations of wells and outcrops are marked with stars, with the sample numbers (Table 1). Suggested nappe structure of the NCA subcrops are slightly modified after Wessely et al. (1993). *BN* Bajuvaric nappes, *TN* Tirolic Nappes, *HN* higher NCA nappes

Lower Campanian contains coal layers within marls and fine-clastic sandstones as well as a massive conglomerate horizon (Summesberger et al. 2002; Wessely 2006) and is comparable to units at the base of the Glinzendorf Syncline. With the Upper Campanian to Maastrichtian Piesting Formation the facies changed to deeper-marine neritic *Inoceramus* marls and *Orbitoides*-bearing sandstones (Hradecká et al. 1999). The maximum depth of the basin was reached with the turbiditic Zweiersdorf Formation of Palaeocene age (Summesberger et al. 2002).

3.3 Glinzendorf basin

Because of the limnic to shallow-marine character with interbedded coal horizons of the Glinzendorf basin (Santonian to Campanian or maybe Maastrichtian strata on Tirolic nappes below Neogene sediments), the Glinzendorf Syncline is sometimes interpreted as an easterly continuation of the Grünbach basin (Wessely 2006). However, a continuous connection from the Grünbach Syncline underneath the Vienna Basin to the Glinzendorf Syncline is

Table 1 Sample locations and affiliation to Gosau formations of all samples with analyzed heavy minerals by electron microprobe

Gosau	Formation	Sample	Age	Sample type	Location	Mineral chemistry		
Gießhübl basin	U-Gießhübl Fm	A-1	Paleocene	Drill core	Aderklaa 84, 2,834.5 m	Grt		
Gießhübl basin	M-Gießhübl Fm	A-2	Paleocene	Drill core	Aderklaa 92, 3,011.5 m	Grt		
Gießhübl basin	L-Gießhübl Fm	A-4	Maastr-Paleoc	Drill core	Aderklaa 92, 3,112 m	Grt		Tur
Gießhübl basin	L-Gießhübl Fm	A-3	Maastr-Paleoc	Drill core	Aderklaa 92, 3,505 m	Grt	Chr	Tur
Gießhübl basin	Nierental Fm	TA-9	Campanian	Outcrop	Tasshof (near Altenmarkt)		Chr	
Gießhübl basin	Nierental Fm	TA-8	Campanian	Outcrop	Tasshof (near Altenmarkt)	Grt	Chr	Tur
Gießhübl basin	Con-San Sst	L-221	Con-San	Outcrop	Lilienfeld	Grt	Chr	Tur
Gießhübl basin	Con-San Sst	L-16	Con-San	Outcrop	Lilienfeld	Grt	Chr	Tur
Gießhübl basin	Con-San Sst	MTW-4	Con-San	Outcrop	Mitterwäldchen (near Alland)		Chr	Tur
Glinzendorf basin	Limnic-marine succession	T1-01	San-Camp	Drill core	Markgrafn. T1, 3207.8 m		Chr	Tur
Glinzendorf basin	Limnic-marine succession	T1-04	San-Camp	Drill core	Markgrafn. T1, 3272.56-60 m	Grt		
Glinzendorf basin	Limnic-marine succession	T1-02	San-Camp	Drill core	Markgrafn. T1, 3409.70-82 m	Grt	Chr	Tur
Glinzendorf basin	Limnic-marine succession	T1-07	San-Camp	Drill core	Markgrafn. T1, 3546.25-55 m	Grt		
Glinzendorf basin	Limnic-marine succession	T1-06	San-Camp	Drill core	Markgrafn. T1, 3744.10-20 m	Grt		Tur
Grünbach basin	Zweiersdorf Fm	G-14	Paleocene	Outcrop	Zweiersdorf-Oberhöflein	Grt		
Grünbach basin	Piesting Fm	G-37	Camp-Maastr	Outcrop	Malleiten-Fischau			Tur
Grünbach basin	Piesting Fm	G-10	Camp-Maastr	Outcrop	Dreistätten-Muthmannsdorf	Grt		Tur
Grünbach basin	Grünbach Fm	G-22	Campanian	Outcrop		Grt	Chr	Tur
Grünbach basin	Grünbach Fm	M-63	Campanian	Outcrop	Maiersdorf (trench)		Chr	Tur
Grünbach basin	Grünbach Fm	M-81	Campanian	Outcrop	Maiersdorf (trench)	Grt	Chr	
Grünbach basin	Maiersdorf Fm	G-9	Santonian	Outcrop		Grt	Chr	
Grünbach basin	Kreuzgraben Fm	G-6B	Santonian	Outcrop		Grt	Chr	
Slovakian basin	Jablonka Fm	B-3B	Eocene	Outcrop	Myjava	Grt		Tur
Slovakian basin	Dedkov Vrch Fm (?)	Z-1	Paleocene	Drill core	Závod 68, 3872.5-3883.6 m	Grt		Tur
Slovakian basin	Paleogene (without precision)	B-1	Paleogene	Outcrop	Solosnica (Brezová area)	Grt		Tur
Slovakian basin	Priepastné Fm	B-2B	Paleocene	Outcrop	Jablonka (Brezová area)	Grt	Chr	Tur
Slovakian basin	Podbradlo Fm	B-7A	Campanian	Outcrop	Bradlo (Brezová area)		Chr	Tur
Slovakian basin	Košariská Fm	B-6C	Campanian	Outcrop	Košariská (Brezová area)	Grt	Chr	Tur
Slovakian basin	Štvernik Fm	B-8A	Con-San	Outcrop	Štvernik (Brezová area)		Chr	
Slovakian basin	Coniacian-Santonian (without precision)	B-5B	Con-San	Outcrop	Brezová-Košariská			Tur

doubtful based on chemostratigraphy (Hofer et al. 2011). Sediments of the Glinzendorf basin appear exclusively in the subsurface, detected by several oil-industry (OMV Aktiengesellschaft and NAFTA) wells like Gänserndorf ÜT3, T3, Markgrafneusiedl 1, N1, T1, Glinzendorf 1 and T1 (Wessely 2006), Gajary 125 (Mišík 1994; Ralbovsky and Ostrolucký 1996) or Záhorska Ves (Fig. 1). Basically, limnic to marginal-marine facies is dominant in the several hundred meter thick succession starting in the Santonian with dark marls and mudstones with some minor coal layers intercalated (Sachsenhofer and Tomschey 1992; Wessely 1992, 2006). Marls, mudstones and minor conglomerates continue up to the Campanian with a marine middle part and non-marine clastics at the top (Wessely 2006). A similar succession is also reported from the easternmost part of the Glinzendorf Syncline in Slovakia (well Gajary 125) with a lower and upper conglomeratic non-marine red interval and an intercalation of a marine grey interval (Mišík 1994). Here, a Maastrichtian (Wessely 1992; Pavlishina et al. 2004) or even Palaeocene top cannot be excluded (Ralbovsky and Ostrolucký 1996).

3.4 Slovakian Gosau basin

Slovakian outcrops of Gosau Groups sediments are present in the Brezová-Myjava area, subdivided into the Upper Cretaceous Brezová Group (Coniacian to Maastrichtian) and the Palaeogene Myjava Group (Palaeocene to Eocene; e.g. Wagreich and Marschalko 1995). A direct correlation of the Gosau Group in the Studienka area, striking ENE underneath the Slovakian part of the Neogene Vienna Basin, with the exposed strata at Brezová is still being debated. However, both the Studienka and the Brezová Gosau Group can be subsumed into a Slovakian Gosau basin. Similarities of the basin fill to the Gießhübl Syncline are striking based on lithofacies and heavy minerals (Wagreich and Marschalko 1995) but also a correlation with the Prottes Gosau Group is debated due to the presence of breccias (Wessely 1992). Strata start with the Upper Coniacian coarse-clastic to marly Ostriež Formation. Above the marl-dominated Santonian Hurbanova Dolina Formation of a neritic shelf facies, calcareous marls of the Campanian Košariská Formation (similar to the Nierental Formation) document a fast deepening of the basin. Further deepening results in massive turbidite units of the Podbradlo, Bradlo and Podlipovec Formation (Upper Campanian to Maastrichtian). Carbonate-poor deep-marine turbiditic facies is recorded in the Palaeocene to Eocene Priepastné and Jablonka Formation of the Myjava Group. A deposition underneath the calcite compensation depth is supposed for the Jablonka Formation, analogue to the Gießhübl Formation (Wagreich and Marschalko 1995).

4 Palaeogeography

During their deposition the Gosau Group sediments of the NCA were located at the northern, tectonically active continental margin of the Austroalpine units, attached to the northern margin of the Adriatic plate (Faupl and Wagreich 1996, 2000). Different palaeogeographic and palaeotectonic reconstructions exist for the Cretaceous to Paleogene evolution of the Eastern Alps (e.g. Faupl and Wagreich 2000; Stampfli et al. 2002; Schmid et al. 2004, 2008; Missoni and Gawlick 2011). To the north, oblique subduction of the Penninic Ocean (Faupl and Wagreich 2000; "Alpine Tethys" of e.g. Stampfli 2000; Stampfli et al. 2002; Schmid et al. 2008; "Piemont Ocean" in Handy et al. 2010; "Alpine Atlantic Ocean" of Missoni and Gawlick 2011) underneath the Austroalpine nappe pile, i.e. the NCA, controlled the tectonic evolution of the continental margin (Wagreich 1993a). The Gosau basins represent slope basins within an early accretionary wedge (Ortner 2001) due to oblique southward subduction of the Penninic Ocean. This partly transpressional wedge accreted from the late Early Cretaceous (Von Eynatten and Gaupp 1999; Wagreich 2001b, 2003) up to the Campanian (Wagreich 1993a) and was composed of Austroalpine units including metamorphic units with the NCA at their front, and inferred obduced (South-)Penninic ophiolites (Winkler 1988; Von Eynatten and Gaupp 1999). Elevation above sea level is supposed for parts of this transpressional feature, and detritus supply (including ophiolitic detritus) is reported from the north into the Gosau basin (Dietrich and Franz 1976; Stattegger 1986, 1987; Pober and Faupl 1988: "northern provenance"; Faupl and Pober 1991). Palaeotransport direction data from sediments of the Upper Aptian to Lower Cenomanian Tannheim–Losenstein basin (Gaupp 1983; Wagreich 2003) and parts of the Lower Gosau Group (Faupl 1983; Wagreich 1988) suggest significant sediment input from the north.

A second ophiolitic belt that may have existed during this time comprises (Neo-) Tethyan ophiolites to the south (in present coordinates) of the NCA (Pober and Faupl 1988), derived by obduction from a Neotethyan ocean system (Meliata-Hallstatt and Vardar Ocean of Stüwe and Schuster 2010; Meliata-Maliac and Vardar Ocean, e.g. Handy et al. 2010) south of the Austroalpine units. The existence of a Tethys suture zone (also called Vardar/Meliata suture or Neotethys suture, e.g. Missoni and Gawlick 2011) further to the south already in the Cretaceous was inferred by Faupl and Wagreich (2000). The ophiolithic belt clearly continued into the Transdanubian Range and Dinarides (Pober and Faupl 1988; Von Eynatten and Gaupp 1999; Árgyelán and Horváth 2002; Lužar-Oberiter et al. 2009, 2012).

From the Santonian/Campanian to Eocene, subcrustal tectonic erosion of the accretionary wedge to the north of the NCA led to northwards deepening of the Gosau basins (Wagreich 1993a; 1995). The northern (Penninic) source area became deactivated at least from that time on (Faupl and Wagreich 1996; Von Eynatten and Gaupp 1999). Northwards-directed palaeocurrents and palaeoslope indicators predominate in Upper Gosau Subgroup sediments (Faupl 1983; Wagreich 1986, 1988). Today, the original palaeogeographic situation of the Gosau basins is largely obliterated due to polyphase tectonic deformation and large-scale thrusting within the Eastern Alps (e.g., Faupl and Wagreich 2000).

5 Methods

5.1 Heavy minerals

523 samples have been used to evaluate and assess the assemblages of heavy minerals of the Gosau Group in the study area. 139 of the heavy mineral analyses (all of them having the same grain size fraction of 0.063–0.4 mm) were taken from literature (Sauer 1980; Gruber 1987; Wagreich and Marschalko 1995); 384 samples were analyzed at the OMV laboratory (editor: R. Sauer) or taken from OMV in-house data (unpublished data; edited by W. Hujer, R. Sauer or unknown editors). Heavy mineral data from a different grain size spectrum (e.g. Woletz 1967: 0.063 to 0.1 mm; or Mišík 1994: unknown grain size spectrum and preparation procedure) were ignored because heavy mineral assemblages are dependent on grain size (e.g. Garzanti et al. 2009). Apatite percentages were omitted because this

mineral was at least partly dissolved by OMV preparation methods with HCl. The heavy mineral spectra were grouped in formations, members, age- or depth intervals (in case of well data) and averaged (Online Resource 1). Only samples with more than 50 counted grains (in most of the cases the number of counted grains is higher than 100) are used for statistics and for the generalized heavy mineral profiles.

5.2 Heavy mineral assemblages

The stable heavy minerals zircon, tourmaline and rutile are combined to the ZTR-index (Hubert 1962), reflecting the mineralogical maturity of the sediments (e.g. Pettijohn et al. 1973; Kutterolf 2001; Garzanti and Andò 2007). During weathering and transport, these minerals accumulate and ZTR-index rises (ZTR-index increases with higher amounts of quartz and chert).

Metamorphic heavy minerals are grouped as META = garnet + chloritoid + glaucophan + epidote group + titanite + staurolite + hornblende + disthene (Von Eynatten 1996; Von Eynatten and Gaupp 1999).

5.3 Mineral chemistry

Analysis of the geochemistry of single heavy mineral grains using microprobe techniques has become widespread in sedimentology for questions of provenance in the last decades (e.g. Henry and Guidotti 1985; Pober and Faupl 1988; Morton 1991; Von Eynatten 1996; Von Eynatten and Gaupp 1999; Weltje and Von Eynatten 2004; Mange and Morton 2007; Mikes et al. 2008; Aubrecht et al. 2009; Meinhold et al. 2009). Especially garnet, chrome spinel and tourmaline are useful minerals because of their frequency and chemical variability.

5.3.1 Garnet

Garnet geochemistry is the most widely used single grain determination and discrimination tool for sedimentary provenance. This is because garnet is a frequent component in heavy mineral assemblages, it is relatively stable during weathering and diagenesis, and it shows a wide range in major element compositions, which is useful for provenance analysis (Morton 1985, 1991; Von Eynatten and Gaupp 1999; Mange and Morton 2007; Morton and Hallsworth 2007). The general formula of garnet is $(Mg, Fe^{2+}, Mn, Ca)_3(Al, Cr, Ti, Fe^{3+})_2(SiO_4)_3$ with some other possible substitutions (Deer et al. 1962). Most garnets belong to the solid solution series between the most common end members almandine ($Fe_3^{2+}Al_2[SiO_4]_3$), pyrope ($Mg_3Al_2[SiO_4]_3$), spessartine ($Mn_3Al_2[SiO_4]_3$) and grossular ($Ca_3Al_2[SiO_4]_3$). Garnets are common in various metamorphic rocks of different P–T-facies but also in plutonic igneous rocks, pegmatites or even in ultramafic varieties and some acid volcanics. Generally, garnet-rich heavy mineral assemblages are associated with metamorphic provenance (Morton 1991; Von Eynatten 1996). Dominant pyrope contents are related to higher-pressure events of the host rock (Wright 1938; Preston et al. 2002; Méres et al. 2012). Provenance discrimination is mainly based on triangle plots between various end members (Barth 1962; Von Eynatten 1996; Von Eynatten and Gaupp 1999; Preston et al. 2002; Morton et al. 2003; Méres 2008; Aubrecht et al. 2009). Almandine garnets from magmatic rocks can be classified by using binary diagram of MnO and CaO (Harangi et al. 2001).

5.3.2 Chrome spinel

Detrital chrome spinel (also chromian or brown spinel, $[Mg, Fe^{2+}][Cr, Al, Fe^{3+}]_2O_4$) is interpreted as an indicator for mafic and ultramafic provenance (such as peridotites and serpentinites) and commonly associated with ophiolites (Zimmerle 1984; Pober and Faupl 1988; Árgyelán and Horváth 2002; Mange and Morton 2007; Aubrecht et al. 2009; Meinhold et al. 2009). With regard to chrome spinel, which is relatively stable, both chemically and mechanically, a general increase in its proportions in sediments during orogeny can be observed in the geological record (Mange and Maurer 1992). Due to the different behavior of the elements Cr, Mg and Al during fractional crystallization and partial melting processes, and the temperature-dependent relation of Fe^{2+} and Mg between the mineral and silicate melt, chrome spinel geochemistry is a powerful petrogenetic tool (Dick and Bullen 1984; Arai 1992). Different chemical ratios and concentrations of spinels can be used to identify the tectonic setting of source rocks (Dick and Bullen 1984; Pober and Faupl 1988; Cookenboo et al. 1997; Lužar-Oberiter et al. 2009; Meinhold et al. 2009), although discrimination diagrams for provenance studies have also been criticized, e.g. by Power et al. (2000). Al_2O_3/TiO_2 discrimination diagram after Kamenetsky et al. (2001) separates tectonic provenance regimes mid-ocean ridge basalt (MORB), ocean-island basalt (OIB), large igneous province (LIP), island-arc magmas (ARC) and supra-subduction zone (SSZ). Several plots use Cr# (=Cr/[Cr + Al]) and Mg# (=Mg/[Mg + Fe^{2+}]) to discriminate spinel chemistry concerning questions of provenance (Dick and Bullen 1984; Pober and Faupl 1988; Arai 1992; Sciunnach and Garzanti 1997; Barnes and Roeder 2001; Hisada et al. 2004). Barnes and Roeder (2001) use the triangle plot Cr–Al–Fe^{3+} to display geotectonic settings.

5.3.3 Tourmaline

Tourmaline geochemistry is highly complex with a huge range of compositional variations (general formula: Na[Mg, Fe, Mn, Li, Al]$_3$Al$_6$[Si$_6$O$_{18}$][BO$_3$]$_3$[OH,F]) comprising of three end members: dravite (Mg-tourmaline), schorl (Fe-tourmaline) and elbaite (Li-tourmaline). Dravite and schorl as well as schorl and elbaite form a continuous solid solution series (Deer et al. 1962). Detrital tourmaline originates mainly from granites, granite pegmatites, contact and regionally metamorphosed rocks, or as recrystallized grains, or formed by metasomatic processes from schists, gneisses or phyllites (Mange and Maurer 1992). Because of its stability during weathering and diagenesis, tourmaline is a common detrital heavy mineral in siliciclastic sediments. The wide range of chemical composition provides the potential for geochemical discrimination of provenance (Henry and Guidotti 1985; Henry and Dutrow 1992; Aubrecht and Krištín 1995; Mange and Morton 2007; Tsikouras et al. 2011). Provenance discrimination is mainly based on triangle plots of Fe–Mg–Al and Fe–Mg–Ca after Henry and Guidotti (1985). Discrimination (MgO vs. FeO) between the different tourmaline varieties is given in Morton (1991).

5.4 Sample preparation for electron microprobe

30 silt to sand-sized Gosau sediment samples were taken and analyzed from outcrops and drill cores (wells Aderklaa 84, 92, Markgrafneusiedl T1 and Závod 68) of the OMV-core sample repository (Fig. 1; Table 1). The samples were crushed to small pieces and decalcified in acetic acid (at the University of Vienna) or hydrochloric acid (at OMV), which partly dissolves apatite. Heavy minerals were separated from the sieve fraction 0.063–0.4 mm by gravity settling in tetrabromethane with a specific gravity of 2.96 g/cm^3 (Mange and Maurer 1992). This grain size fraction was selected because data from different studies and OMV in-house data, which all used this range of fraction, was used and compared to our data. The grain concentrates containing the whole heavy mineral fraction were embedded in epoxy resin (araldite mount) and polished (up to 1 μm on grinding disc).

5.5 Electron microprobe

Element analyses of single heavy mineral grains were carried out using a *Cameca SX 100* electron microprobe at the University of Vienna (Department of Lithospheric Research) at operating conditions of 15 kV and 20 nA. The analyses were made and calibrated against mineral standards, followed by a PAP correction (after Pouchou and Pichoir 1984) applied to the data. 1,064 measurements on 774 single grains of garnet (263 grains), chrome spinel (269 grains) and tourmaline (242 grains) have been analyzed. B$_2$O$_3$ concentrations in tourmalines were defined between 9.660 and 9.982 as usual contents in common tourmalines. Oxygen and H$_2$O were stoichiometrically calculated. The analytical error for the elements is controlled by several factors, but generally below 5 % (e.g. Bjerg et al. 2009). Subjective influence (of a special group of grains) was tried to minimize by randomly selecting grains one after another.

Formula parameters and end members for the different minerals analyzed by electron microprobe were calculated using the software *MINSORT* (Petrakakis and Dietrich 1985) based on 12 oxygens for garnets and 32 oxygens for spinels.

6 Results

6.1 Heavy mineral assemblages

Data for the generalized heavy mineral profiles shown in Fig. 2 have been taken from the literature (Sauer 1980; Gruber 1987; Wagreich and Marschalko 1995) and OMV in-house data (unpublished data; edited by W. Hujer, R. Sauer or unknown editors). In general, garnet and chrome spinel show the most significant variations during the depositional history of the Gosau Group (e.g. Faupl 1983; Wagreich and Marschalko 1995).

The Gießhübl basin is generally characterized by a drastic rise of garnet concentrations and hand in hand with that, a drop in chrome spinel and zircon from the Coniacian/Santonian up to the Palaeocene (Figs. 2, 3). Rutile (mean: 8.2 $\pm$ standard deviation: 7.9 %) and tourmaline (5.3 $\pm$ 4.9 %) percentages are moderate/low and constant, some chloritoide is only found in the Nierental Formation (up to 8 %) and in the base of the Lower Gießhübl Formation (up to 7 %). Basal sandstones of the Coniacian/Santonian age (data from OMV wells Aderklaa 82, Schönkirchen T11 and T90, Strasshof T1, T9, T9a and T9b; $N = 27$; Online Resource 1) are dominant in chrome spinel (49.1 $\pm$ 32.1 %) and zircon (28.7 $\pm$ 21.5 %) with low garnet concentrations (7.9 $\pm$ 16.3 %). Garnet populations rise to 32.8 $\pm$ 22.2 %, while chrome spinel (21.7 $\pm$ 13.6 %) and zircon (17.8 $\pm$ 9.3 %) is depleted in samples of the Nierental Formation (data from OMV wells of Aderklaa 92; $N = 14$; Online Resource 1). The base of the Lower Gießhübl Formation ($N = 6$) is quite different to the rest of the formation and characterized by a dominant amount of chrome spinel (37.8 $\pm$ 29.3 %) with a garnet content of 22.0 $\pm$ 18.5 % (Sauer 1980). Generally, assemblages of the Gießhübl Formation are dominated by garnet (Lower G-Fm: 78.2 $\pm$ 17.9 %, $N = 139$; Middle G-Fm: 66.6 $\pm$ 29.6 %, $N = 52$; Upper G-Fm: 82.0 $\pm$ 6.9 %, $N = 52$) while chrome spinel is almost inexistent (Sauer 1980).

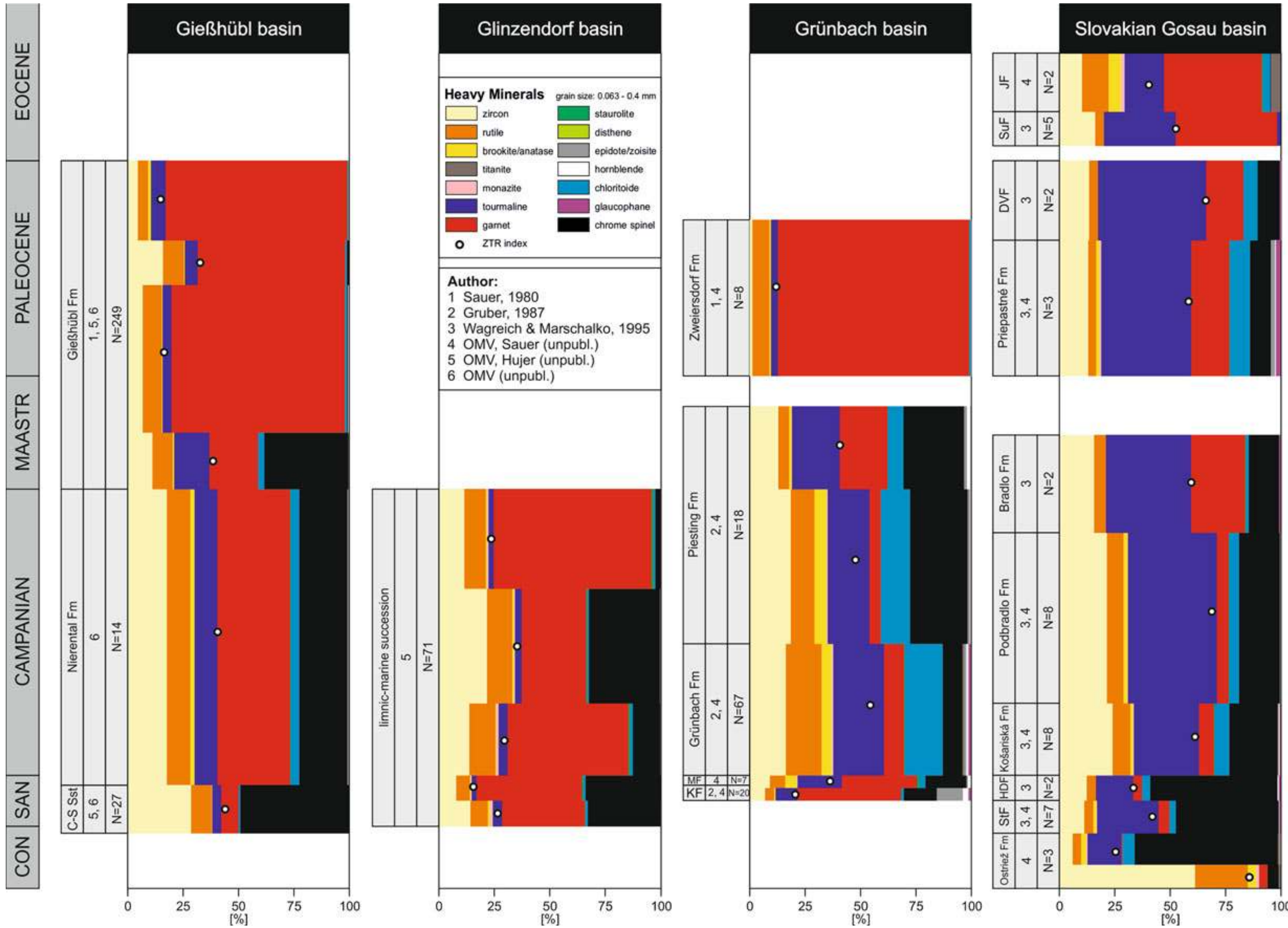


Fig. 2 Compiled heavy mineral profiles of the area of the Gießhübl, Glinzendorf, Grünbach and Slovakian Gosau basin. Number of samples (*N*) and authors (1–6) are given. Data have been taken from literature (*1* Sauer 1980; *2* Gruber 1987 and *3* Wagreich and Marschalko 1995), OMV in-house data (unpublished data; edited by *5* W. Hujer, *4* R. Sauer or *6* unknown editors) or have been edited by *4* R. Sauer at the OMV lab. Apatite percentages were omitted because it was partly dissolved. The heavy mineral spectra were grouped in formations (*C–S Sst* Coniacian–Santonian sandstones; *KF* Kreuzgraben Formation; *MF* Maiersdorf Formation; *StF* Štvernik Formation; *HDF* Hurbanova Dolina Formation; *DVF* Dedkov Vrch Formation; *SuF* Surovin; *JF* Jablonka Formation), members, age- or depth intervals (in case of well data) and averaged

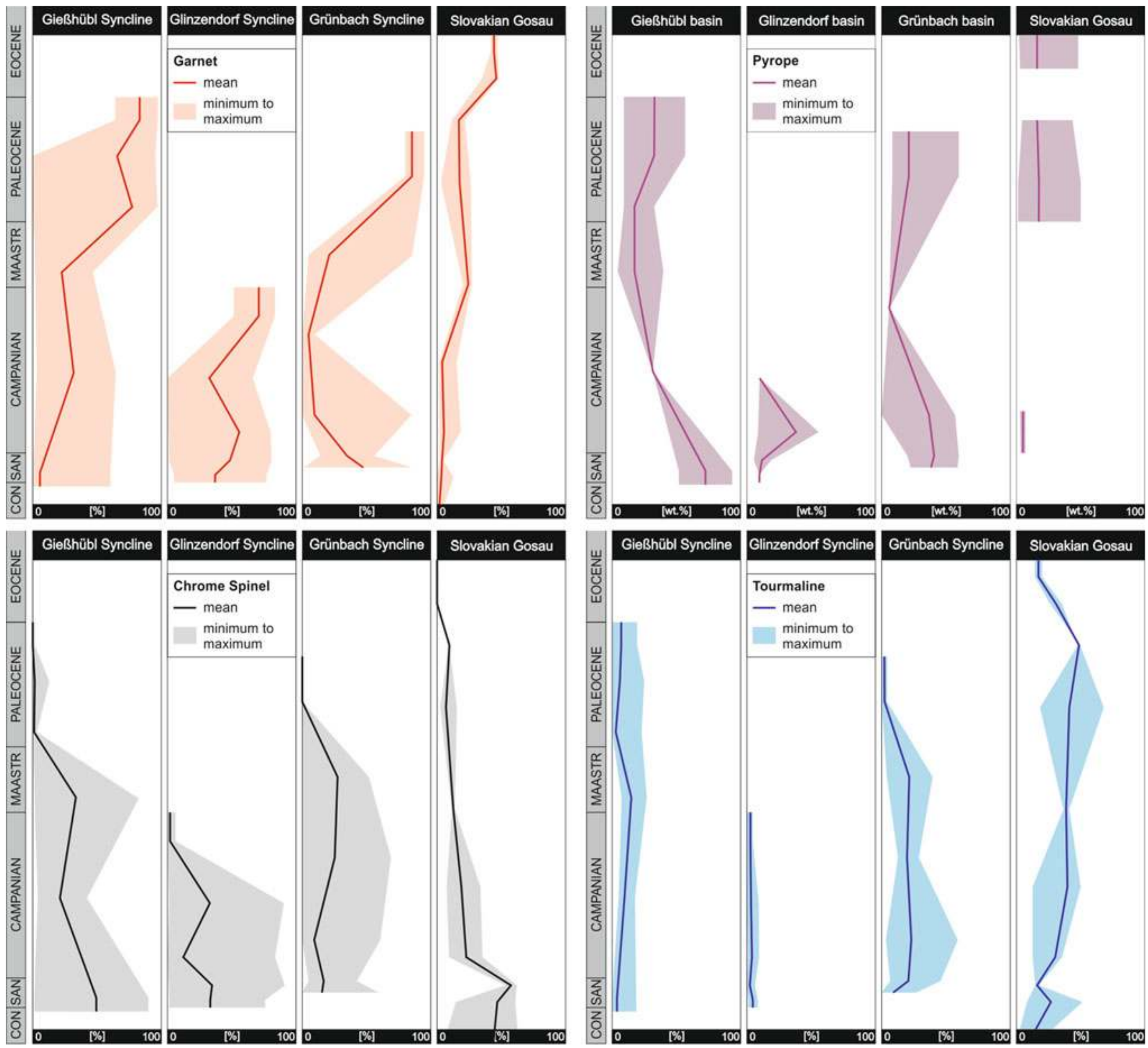


Fig. 3 Compiled profiles of garnet, pyrope, chrome spinel and tourmaline contents (mean, minimum and maximum values) of clastic Gosau sediments from the Gießhübl, Glinzendorf, Grünbach and Slovakian Gosau basin

Heavy mineral assemblages of the limnic-marine successions of the Glinzendorf basin are generally garnet-rich (45.4 ± 27.5 %) with moderate and fluctuating chrome spinel (24.4 ± 30.0 %) as well as moderate zircon (14.4 ± 12.8 %) and rutile contents (9.4 ± 6.0 %) and low tourmaline concentrations (3.1 ± 2.5 %; see Figs. 2, 3). Chrome spinel dominant parts (up to 91.7 %) correlate with non-marine strata within the wells Markgrafneusiedl T1 and Glinzendorf T1 (Hofer et al. 2013). The Santonian/Campanian (maybe Maastrichtian; Wessely 2006) section is subdivided into five depth-sections of the wells Markgrafneusiedl T1 and Glinzendorf T1 that show similar heavy mineral assemblages. It has to be kept in mind, that the correlation between the two wells by depth level is certainly not accurate, because tectonic deformation is relatively complex within the subcrop of the Glinzendorf Syncline (Wessely 1992, 2006; Mišík 1994; Ralbovsky and Ostrolucký 1996). Generally, garnet contents rise and chrome spinel concentrations drop in marine and top sections and vice versa (Figs. 2, 3). The bottom interval between 4,200 and 3,900 m is characterized by relatively high amounts of chrome spinel (32.5 ± 25.1 %; up to 77 %) and moderate amounts of garnet (37.4 ± 25.0 %). Garnet contents slowly increase between 3,900 and 3,700 m (47.7 ± 33.9 %) and 3,700 and 3,500 m (54.5 ± 22.8 %), while chrome spinel is reduced

(12.3 ± 22.7 %) between 3,700 and 3,500 m. High chrome spinel (31.7 ± 30.2 %) and zircon (21.8 ± 20.8 %) contents characterize the interval between 3,500 and 3,300 m. At the top, between 3,300 and 3,100 m, garnet accumulation dominates with 71.1 ± 9.7 %, while chrome spinel is more or less inexistent (2.3 ± 2.5 %). Similar heavy mineral assemblages (dominance of ZTR and chrome spinel) are observed for the well Gajary 125 (Mišík 1994), but, due to unknown grain size spectrum and preparation procedure, this data was ignored for further evaluation during the present study.

Late Cretaceous (Santonian to Maastrichtian) heavy mineral assemblages of the Grünbach basin are generally rich in chrome spinel, tourmaline, garnet, zircon, rutile and chloritoide, while in the Palaeocene Zweiersdorf Formation garnet is the only dominant heavy mineral phase (Sauer 1980; Gruber 1987; see Figs. 2, 3). Upper Santonian Kreuzgraben and Maiersdorf formations are dominated by garnet (46.7 ± 24.7 and 34.0 ± 17.5 %) with moderate, but fluctuating concentrations of chrome spinel (14.8 ± 14.7 and 18.4 ± 10.7 %) and tourmaline (9.9 ± 9.6 and 19.9 ± 14.2 %). Tourmaline (22.8 ± 12.8 %), chloritoide (17.1 ± 14.7 %), zircon (16.4 ± 11.4 %) and rutile (16.1 ± 12.6 %) are present in high amounts in the Campanian Grünbach Formation ($N = 67$). Chrome spinel and garnet contents are highly variable (between 0 and 61 % and 0 and 86 %, respectively), but relatively low in average (8.9 ± 12.8 % and 9.1 ± 16.0 %). With 26.5 ± 23.4 % in average, chrome spinel is the dominant heavy mineral phase in the Piesting Formation but also tourmaline (20.1 ± 9.9 %), zircon (15.8 ± 10.4 %) and chloritoide (10.2 ± 11.2 %) concentrations are moderate to high. Garnet (13.3 ± 20.1 %) already predominates in the youngest intervals (Maastrichtian) of the Piesting Formation with maximum values of 86 % (Gruber 1987). Within the overlying Palaeocene Zweiersdorf Formation this trend continues, and garnet is the dominant heavy mineral with a mean concentration of 86.3 ± 4.6 %.

Tourmaline is the dominant heavy mineral of the Slovakian Gosau equivalents with 31.1 ± 14.2 % in average (Figs. 2, 3). High to medium chrome spinel concentrations are characteristic for the Coniacian to Maastrichtian, medium to high amounts of garnet can be observed from the Late Campanian onwards (Wagreich and Marschalko 1995; see Figs. 2, 3). The ZTR index is higher compared to the Austrian Gosau equivalents due to the high tourmaline and constantly medium to high zircon percentages. Coniacian to Santonian strata are (except one zircon-rich sample at the base) characterized by chrome spinel (51.6 ± 14.7 %). Tourmaline contents are high as well (23.4 ± 14.0 %), while garnet is rare or even absent (4.0 ± 4.7 %) in these sediments. From the Campanian onwards chrome spinel concentrations are reduced (they typically scatter between 10 and 20 %) and do not occur in the Eocene any more. While garnet is still depleted in the Campanian (6.0 ± 5.1 %) it rises up to 24.4 ± 4.2 % in the Bradlo Formation. Tourmaline is the dominant heavy mineral in the Campanian and Maastrichtian (35.2 ± 10.4 %), zircon is relatively high and constant as well in these sediments (22.0 ± 8.0 %). In the Palaeocene, tourmaline concentrations are even slightly higher with a mean of 43.8 ± 18.1 % and a maximum content of 68 % (Priepastné Formation). Garnet percentages are higher (17.1 ± 8.7 %), chrome spinel concentrations lower (9.7 ± 5.1 %) in the Palaeocene as compared to the Late Cretaceous. Zircon (13.3 ± 3.3 %) and chloritoide contents (8.0 ± 5.7 %) are relatively high as well. In the Eocene, garnet is the dominant heavy mineral phase with an average amount of 45.2 ± 4.6 %. Tourmaline concentrations are still high (28.1 ± 7.9 %) as it is characteristic for the whole Brezová and Myjava Group. Zircon contents are similar to the Palaeocene, while chrome spinel is inexistent.

Triangle plots of metamorphic minerals (META), stable minerals (ZTR) and ophiolite detritus, respectively chrome spinel (Pober and Faupl 1988) illustrate the general decrease of ophiolitic detrital influence from the Coniacian to the Eocene with a mixed source from the Coniacian to the Maastrichtian and an ophiolitic-free provenance in the Paleogene (Fig. 4).

6.2 Correlations and logratios of heavy minerals

Pearson correlation indices of heavy minerals are shown in Online Resource 2. Ultrastable minerals zircon and rutile are (except for the Glinzendorf basin) positively correlated (r is up to 0.747). A significant positive correlation with tourmaline is not present (only for rutile and tourmaline at the Glinzendorf basin; $r = 0.476$). Negative correlations between zircon/rutile and garnet can be observed for the Gießhübl and Grünbach basin and between rutile and chrome spinel for the Glinzendorf and Grünbach basin. The highest negative correlation indices are present between garnet and chrome spinel (Gießhübl basin: $r = -0.704$; Glinzendorf basin: $r = -0.844$; Grünbach basin: $r = -0.144$; Slovakian Gosau basin: $r = -0.627$).

Logratio discrimination plots (Fig. 5) use the four dominant heavy mineral phases (garnet and chrome spinel) and indices (ZTR and META; e.g. Von Eynatten 1996; Wagreich 1998). This creates four quadrants with specific heavy mineral associations (1:1 mixtures plot on the zero-lines): Quadrant I (Q I), with samples that are dominant in garnet and chrome spinel; Quadrant II (Q II) characteristic for assemblages that are rich in garnet and other metamorphic associated heavy minerals; Quadrant III (Q III) contains samples with dominant concentrations of the stable heavy minerals zircon, tourmaline and rutile as well as

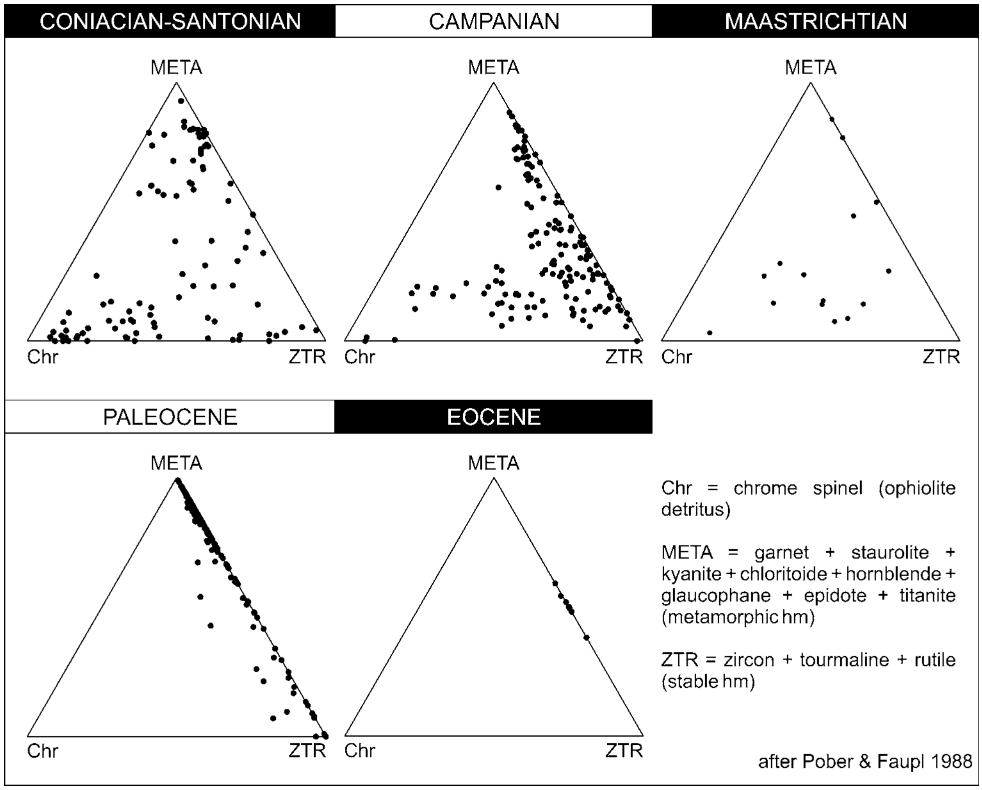


Fig. 4 Triangular plot of chrome spinel (ophiolite detritus), metamorphic (META = garnet + staurolite + kyanite + chloritoide + hornblende + glaucophane + epidote + titanite) and stable heavy minerals (ZTR = zircon + tourmaline + rutile) after Pober and Faupl (1988). Data set is grouped in age intervals

metamorphic minerals; and at quadrant IV (Q IV), samples with dominant amounts of chrome spinels and stable heavy minerals plot.

Due to strong negative correlations between garnet and chrome spinel (Online Resource 2), only two percent of the data set plot in Q I implying that a dominance of garnet hand in hand with high amounts of chrome spinel is extremely rare (Fig. 5). Samples from the Coniacian-Santonian age are either dominated by chrome spinel and stable minerals or garnet and metamorphic minerals (46 and 41 % plot in Q IV and II, respectively). Especially samples from the Gießhübl basin and Slovakian Gosau basin plot in Q IV, while samples from the Grünbach basin are dominated by garnet and metamorphic minerals. Generally, marine samples from the Glinzendorf basin plot in Q II, non-marine samples in Q IV (Fig. 5). Campanian heavy mineral assemblages of the Gießhübl basin show balanced amounts of garnet, chrome spinel, metamorphic and stable minerals and plot in the central part of Q II, III and IV. Samples from the Glinzendorf basin are mainly dominated by garnet; the Slovakian Gosau basin is rich in chrome spinel and stable minerals and plot in Q IV. Only samples from the Grünbach basin are neither dominant in garnet nor in chrome spinel and plot in Q III (Fig. 5). 69 % of Maastrichtian samples group in Q IV, symbolizing dominance of chrome spinel and stable minerals (Fig. 5). Absence of chrome spinel and dominance of garnet in the Palaeocene and Eocene result in dominant samples in Q II (80 %) and Q III (18 %; see Fig. 5).

6.3 Mineral chemistry

6.3.1 Garnet

Analyzed detrital garnets (mean values of multiple analyzes from single grains are used for further calculations) are usually almandine-rich with 63 ± 9 % in average and a maximum of 85 %. Only 12 % of 263 analyzed garnet grains have compositions lower than 50 % almandine. Grossular proportions are generally moderate to low with a mean of 18 ± 9 %. Only two grains show a dominant grossular component with 56 (B-3B) and 93 % (G-22). Pyrope contents vary between 0 and 46 % with a low average of 12 ± 7 % (half of the samples have concentrations lower than 10 %; see Fig. 3). Mean pyrope contents are summarized in Table 2. While low spessartine proportions are characteristic (6 ± 6 % in average), andratite and uvarovite is almost non-existent (Online Resource 3).

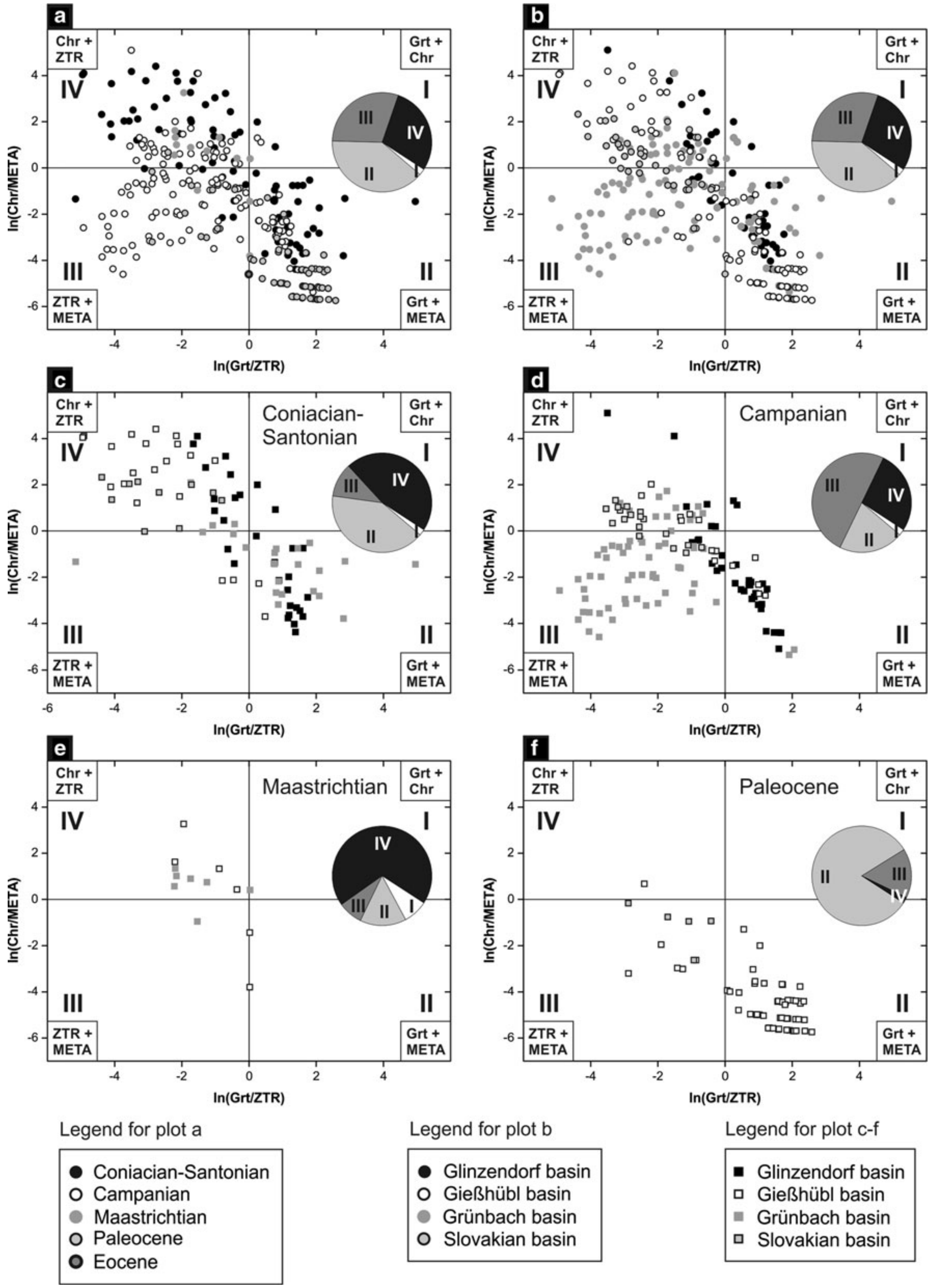


Fig. 5 Logratio discrimination plots of ln(Grt/ZTR) vs. ln(CHR/META). **a** Grouped according to age interval. **b** Grouped according to Gosau basin. **c**–**f** Grouped according to Gosau basin for individual age intervals. Grt = garnet; Chr = chrome spinel; META = garnet + staurolite + kyanite + chloritoide + hornblende + glaucophane + epidote + titanite; ZTR = zircon + tourmaline + rutile. Pie charts illustrate the percentage of samples plotting in quadrant I to IV

Table 2 Mean pyrope contents of detrital garnets from different time intervals and different Gosau basins

Con-San	Mean	SD	*N*
Mean pyrope contents of detrital garnets (wt%)			
GHB	0.36	0.14	2
GDB			0
GBB	0.20	0.05	30
SGB			0
Campanian	**Mean**	**SD**	***N***
GHB	0.16		1
GDB	0.14	0.09	18
GBB	0.17	0.07	19
SGB	0.03	0.01	3
Maastrichtian	**Mean**	**SD**	***N***
GHB	0.10	0.06	31
GDB			0
GBB			0
SGB			0
Paleocene	**Mean**	**SD**	***N***
GHB	0.09	0.04	62
GDB			0
GBB	0.11	0.06	39
SGB	0.09	0.06	52
Eocene	**Mean**	**SD**	***N***
GHB			0
GDB			0
GBB			0
SGB	0.09	0.09	6

GHB Gießhübl basin, *GDB* Glinzendorf basin, *GBB* Grünbach basin, *SGB* Slovakian Gosau basin

Various triangle plots show similar systematic trends of garnet populations in time and space (Fig. 6; different Gosau basins). Generally, garnets of such compositions cannot be explicitly assigned to a specific lithology, because garnet chemistry is dependent on paragenesis and coexisting mineral phases. Higher amounts (around 20 and 30 %) of pyrope and grossular are commonly associated with amphibolites, blueschist-associated eclogite or granulite source rocks. Pyrope contents of <50 % exclude garnet peridotites and associated eclogites as possible source (Von Eynatten and Gaupp 1999). Components of high-grade metamorphic garnets supposed of being derived from amphibolites and granulites (mainly expressed by high contents of pyrope and grossular) are prominent from the Coniacian to the Campanian and disappear in the Maastrichtian to the Eocene, where dominant almandine-rich garnets suggest biotite schist and granite source. Mean pyrope values are high in the Coniacian–Santonian (20.6 ± 7.0 %) and Campanian (14.5 ± 8.7 %) and significantly lower in the Maastrichtian (9.7 ± 5.8 %), Palaeocene (9.8 ± 5.4 %) and Eocene (8.8 ± 9.1 %). Discrimination diagram (almandine + spessartine – pyrope – grossular; Fig. 6) of Preston et al. (2002) indicates 49 % of amphibolites and 29 % of eclogite provenance in Coniacian–Santonian age. This "high-metamorphic" source is dominant compared to lower metamorphic-grade derived garnets from biotite schists (10 %) and granites/pegmatites (10 %). This influence of a high metamorphic character of garnets is still present (but reduced) in the Campanian (35 % indicate amphibolites, 15 % eclogites, 23 % biotite schists and 17 % granities as possible provenance). From the Maastrichtian to the Eocene there are no indications for high-grade metamorphic garnets any more. Biotite schists (around 70 % plot in this field) are the supposed dominant source. The same trend can be demonstrated using the boundary conditions after Morton et al. (2003) with dominant garnets derived from amphibolites/granulite facies from the Coniacian to the Campanian in contrast to prevailing low-grade metamorphosed almandine-rich garnets from the Maastrichtian onwards (Fig. 6). The general retrograde metamorphic character in time can also be observed in the almandine – pyrope – grossular plot (Fig. 6) after Méres (2008): The bigger part of Coniacian to Campanian garnets suggest compositions from granulite, eclogite and high amphibolites to granulite facies, and high-pressure eclogites, high-pressure mafic granulites and amphibolites as interpreted source rocks. Maastrichtian to Eocene detrital garnets plot more or less in the field of amphibolite facies and suggest a lower degree of metamorphism (e.g. Aubrecht et al. 2009).

There was also a difference in the spatial distribution of detrital garnets observed. The Grünbach and Glinzendorf basins, which were palaeogeographically located in the south of the Gosau depositional area, are generally characterized by a higher amount of garnets derived from a higher-metamorphic source compared to the northern Gießhübl and Slovakian basins. A statistical evaluation of this observation is not possible, because the different time periods are not always represented by samples from all Gosau basins. In the Slovakian basin, garnets do not suggest input of high-grade metamorphic (e.g. upper-amphibolitic, eclogitic) source rocks (Fig. 6).

6.3.2 *Chrome spinel*

MgO contents of the analyzed detrital chrome spinels (mean values of multiple analyses from single grains are used for further calculations) range from 6.8 to 19.7 % with 13.3 ± 2.4 % in average. Mean Cr_2O_3 concentrations of 40.5 ± 9.9 % fluctuate between 10.7 and 58.5 %. Mg#

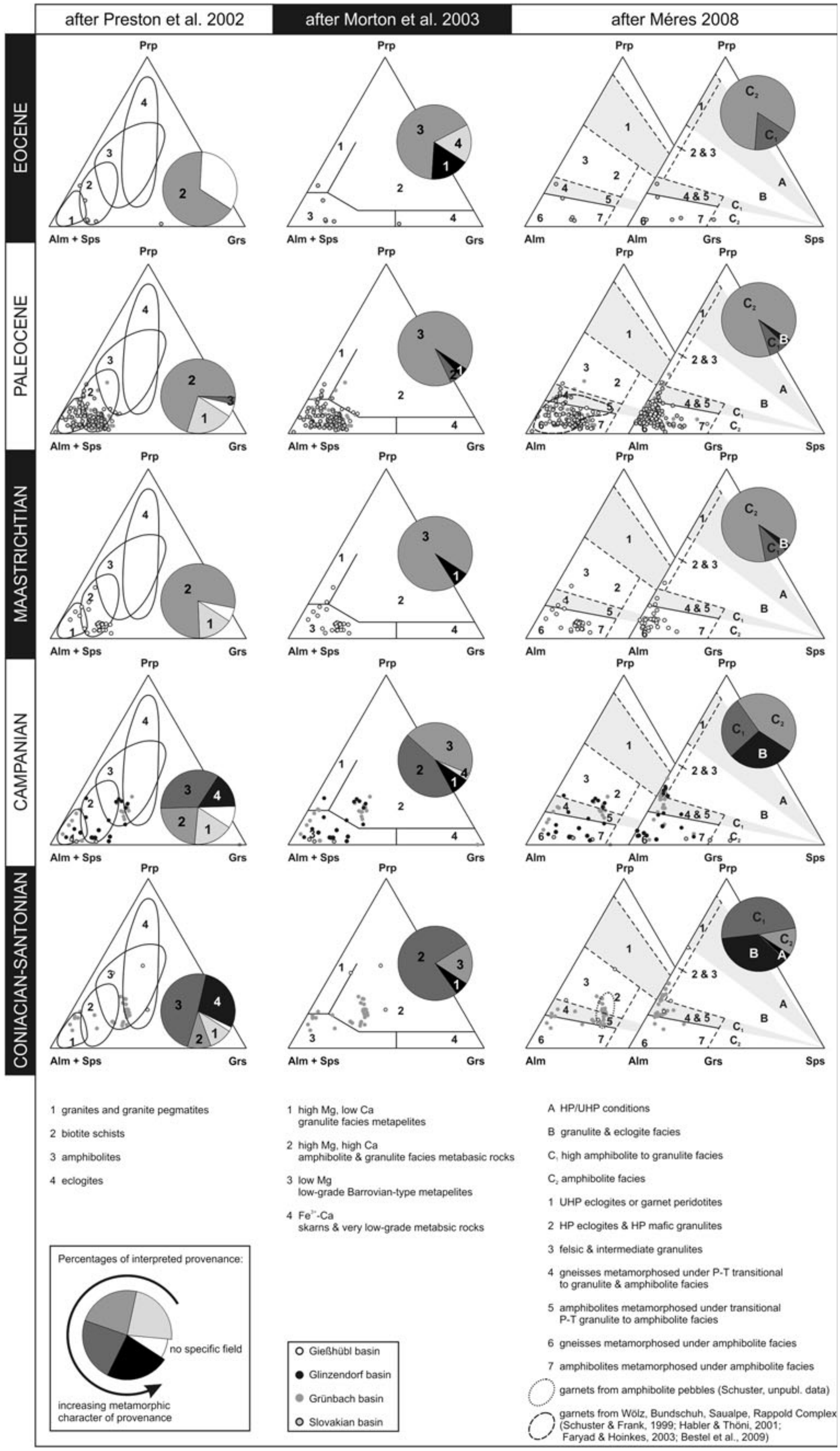


Fig. 6 Garnet chemistry variations displayed as triangle plots for detrital garnets proposed by different authors (data from Online Resource 3). *Left column* almandine (Alm) + spessartine (Sps) − pyrope (Prp) − grossular (Grs) triangle plot after Preston et al. (2002). *Middle column* the same after Moore et al. (2003). *Right column* almandine − pyrope − spessartine triangle plot after Méres (2008). Plots are shown for individual age intervals with samples grouped in different Gosau basins (Gießhübl, Glinzendorf, Grünbach and Slovakian Gosau basin). Fields for interpreted provenance (lithologies or metamorphic facies) from where the garnets are supposed to be derived are given below the plots according to the authors. Pie charts illustrate the percentage of samples plotting in the different fields of provenance (*gray scale* indicates the metamorphic character of the provenance)

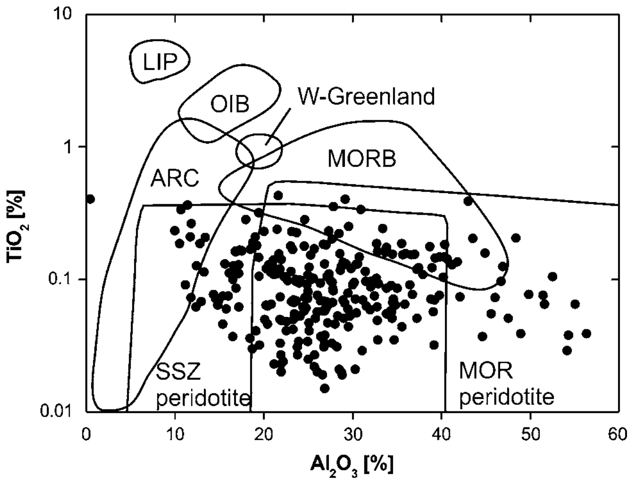


Fig. 7 Chrome spinel discrimination diagram (Al_2O_3 vs. TiO_2), showing the suggested fields of data derived from various types of mafic and ultramafic rocks after Kamenetsky et al. (2001). *LIP* large igneous province, *OIB* ocean-island basalt, *MORB* mid-ocean ridge basalt, *ARC* island-arc basalt, *SSZ peridotite* supra-subduction zone, *MOR peridotite* plotted data from Online Resource 4

and Cr# show a broad range between 0.29 and 1.0 as well as 0.2 and 0.88, respectively, with means of 0.62 ± 0.14 and 0.65 ± 0.14. Trace element concentrations are constantly low with TiO_2 <0.43, NiO <0.38, MnO <0.33 and ZnO <0.31 (Online Resource 4). Generally, zoned spinels could not be observed thus representing unaltered chemical composition (Pober and Faupl 1988).

According to the Al_2O_3/TiO_2 diagram of Kamenetsky et al. (2001), most of the detrital chrome spinels were derived from supra-subduction zone (SSZ) and MOR-type peridotites. Only a few samples trend to volcanic rock sources from island-arc basalts or MORB-type rocks (Fig. 7).

Generally, chrome spinels fit well to the field of ophiolites using the Mg# vs. Cr#-diagram of Pober and Faupl (1988), as shown in Fig. 8. The influence of metamorphic spinel is negligible. Most of the samples plot in the field of type II ophiolite (peridotite) provenance, which is a transitional classification comprising both type I and III (Pober and Faupl 1988). Samples derived from type I (Cr# <0.6; lherzolite composition) and III (Cr# >0.6; harzburgite composition) ophiolites are similarly distributed for the whole dataset. About two-thirds of the analyzed grains are of harzburgitic provenance. Looking in detail, there is a significant difference in time and space. In the Coniacian–Santonian, harzburgite composition is with around 75 % dominant compared to 14 % of lherzolite related source. This dominance of harzburgite peridotites is reduced in the Campanian (53 % harzburgite compared to 41 % lherzolite composition). The mean of Cr# of the Coniacian–Santonian is with 0.68 ± 0.11 significantly different from the Campanian with 0.61 ± 0.16. A significance of 0.000 (T = 3.773) is proved by Welch's *t* test, which was used because equality of variance is not given. 63 and 30 % of Maastrichtian to Palaeocene spinels are related to harzburgite, respectively, lherzolite source. Palaeogeographically northern basins have a significant higher influence of harzburgite character (74 % in the Gießhübl basin; 65 % in the Slovakian Goasu basin) compared to southern basins (44 % in the Glinzendorf basin; 47 % in the Grünbach basin).

6.3.3 Tourmaline

Generally, tourmaline chemistry of the analyzed formations (mean values of multiple analyzes from single grains are used for further calculations) range to a great degree in major element concentrations: Al_2O_3: 22.4–39.3 (mean: 32.1 ± 2.2), FeO: 0.3–23.4 (mean: 8.1 ± 2.8), MgO: 0.0–14.8 (mean: 6.6 ± 1.6), CaO: 0.0–17.4 (mean: 0.5 ± 1.1; see Online Resource 5). Mixtures of dravite and schorl are most common with no specific differences in time or space. Minor and trace element concentrations fluctuate between 0.0 and 3.1 (Na_2O), 1.9 (TiO_2), 1.4 (F), 1.0 (K_2O) and 0.4 (MnO). Chemistry of cores and rims vary differently strong between almost zero and a few weight percentages (Online Resource 5).

Provenance-discrimination plots (Fe–Mg–Al and Fe–Mg–Ca after Henry and Guidotti 1985) indicate a generally mixed source of various metapelites, Fe^{3+}-rich quartz-tourmaline rocks and granitoids (Fig. 9). Neither in time, nor in space, i.e. palaeogeographic separated Gosau basins, distinct changes in the source rocks is identifiable. According to these plots, there is also no clear evidence for the presence of tourmalines derived from meta-ultramafic rocks and Cr, V-rich metasediments, which would be indicative for an ophiolitic provenance. All in all, the tourmaline chemistry data point to strongly mixed provenance and do not show a significant fundamental change in source areas during Late Cretaceous to Palaeocene times.

7 Discussion

Heavy mineral assemblages of Coniacian–Campanian sediments are characterized by prevailing chrome spinel, followed by zircon, tourmaline and garnet in all of the investigated Gosau basins (Fig. 2). This indicates erosion of an ophiolitic source during sedimentation of the Lower Gosau Subgroup. Originally, this detritus was interpreted to have been only derived from a northern Penninic ("Alpine Tethys") source (Dietrich and Franz 1976; Stattegger 1986). Besides other arguments (Decker et al. 1987), the chemical analysis of detrital chrome spinels

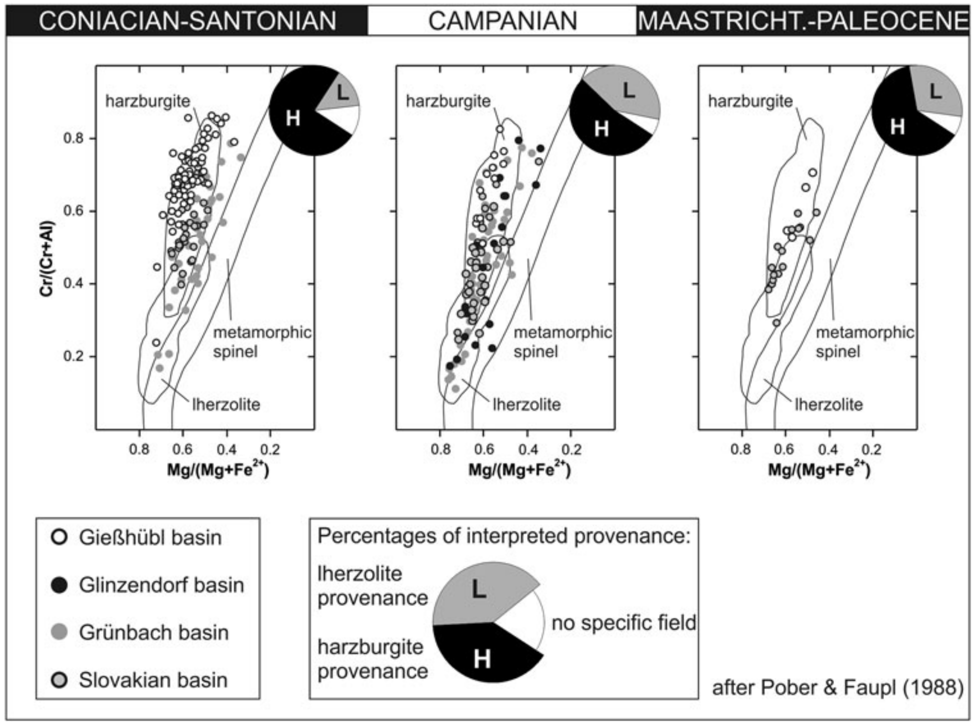


Fig. 8 Chrome spinel discrimination diagrams Mg/(Mg + Fe^2) vs. Cr/(Cr + Al), showing the suggested field of provenance after Pober and Faupl (1988). Plots are shown for individual age intervals with samples grouped in different Gosau basins (Gießhübl, Glinzendorf, Grünbach and Slovakian Gosau basin). Pie charts illustrate the percentage of samples plotting in harzburgite (H) or lherzolite (H) provenance field. Plotted data from Online Resource 4

(Pober and Faupl 1988) and sedimentological features of serpentinitic sandstones (Wagreich 1993b) introduced a possible southern provenance for ophiolitic detritus in the NCA (Fig. 10). This southern ophiolite complex supplied clastic material at least since the Early Cretaceous (Decker et al. 1987; Von Eynatten and Gaupp 1999) up to the Maastrichtian and is now completely missing, probably fully eroded. Ophiolitic thrust sheet(s) on top of the NCA due to Neotethys obduction can be inferred as the main source (e.g. Wagreich 1993b; Missoni and Gawlick 2011). Today, the only evidence for this ophiolitic thrust unit(s) is provided by basic volcanic components in conglomerates (Gruber et al. 1992), coarse serpentine fragments in sandstones (Wagreich 1993b), extremely chrome spinel-rich heavy-mineral assemblages (this paper), and the presence of amphibolitic metamorphic sole pebbles (unpublished data by Ralf Schuster, Geological Survey, Austria) in Cretaceous formations of the NCA. These obducted Vardar ophiolites (West-Vardar ophiolites overlying Meliata ophiolites; Schmid et al. 2008) and mélanges shed chrome spinel-rich detritus also at a larger regional scale (Pober and Faupl 1988; Árgyelán 1996; Von Eynatten and Gaupp 1999; Lenaz et al. 2009) and can be connected into the Dinarides (Lužar-Oberiter et al. 2009, 2012). In addition, high amounts of Cr and Ni in non-marine fine-grained clastic sediments of the Glinzendorf basin suggest a dominant ophiolitic detrital supply in proximal, palaeogeographically southern basins (Hofer et al. 2013).

Chrome spinels from northern Gosau basins have characteristically higher Cr# (mean: 0.67 ± 0.10) compared to spinels from southern basins (0.60 ± 0.18) and indicate harzburgite source (Fig. 8). Spinels from Penninic ophiolites from the Tauern, Unterengadin (Pober and Faupl 1988) and Rechnitz window (Mikuš and Spišiak 2007), which are today's equivalents to the inferred northern ophiolitic provenance, commonly have lower Cr# indicating lherzolitic composition. However, the chemistry of chrome spinels is hardly distinguishable between the Vardar and Penninic (Alpine Tethys) ultramafics (Pober and Faupl 1988; Mikuš and Spišiak 2007). Hence, on the one hand, chrome spinel data do not confirm dominant detrital input from northerly located ophiolites belonging to a Penninic Ocean source. On the other hand, inferred south Penninic oceanic successions, now situated at the northern margin of the NCA, do contain harzburgitic chrome spinels in Cretaceous formations (e.g. Arosa and Ybbsitz zone, Pober and Faupl 1988). This means that a Penninic derivation and thus a northern provenance cannot be ruled out based on chrome spinel chemistry alone, although this is also suggested by palaeotransport data, e.g. for the Tannheim-Losenstein basin in the Early Cretaceous (Wagreich 2001b).

Chrome spinel chemistry of southern Gosau basins points to a provenance from a mixed harzburgite and lherzolite source of obducted Neotethys ophiolites or a Tethys suture sensu Pober and Faupl (1988) and Faupl and Wagreich (2000). Analogue detrital chrome spinels from Dinaric Cretaceous basins (Lužar-Oberiter et al. 2009) and

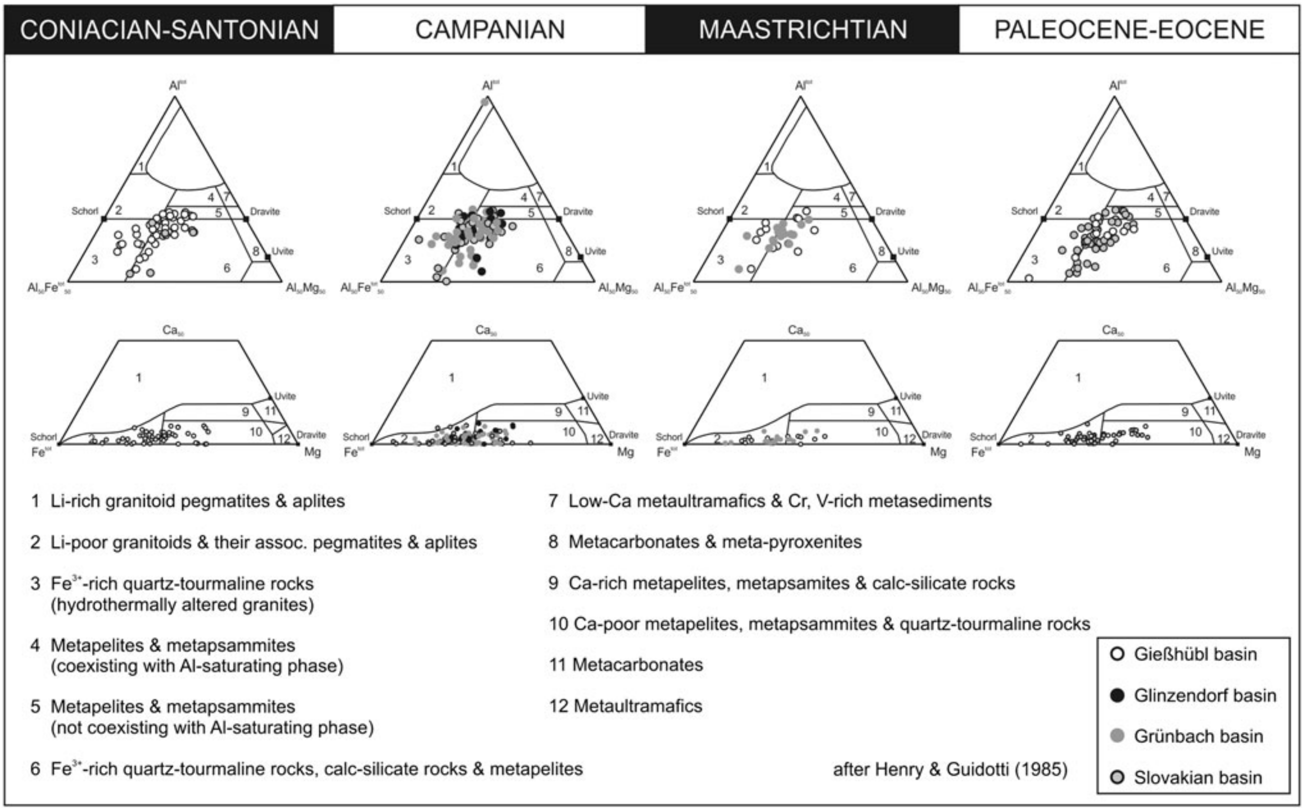
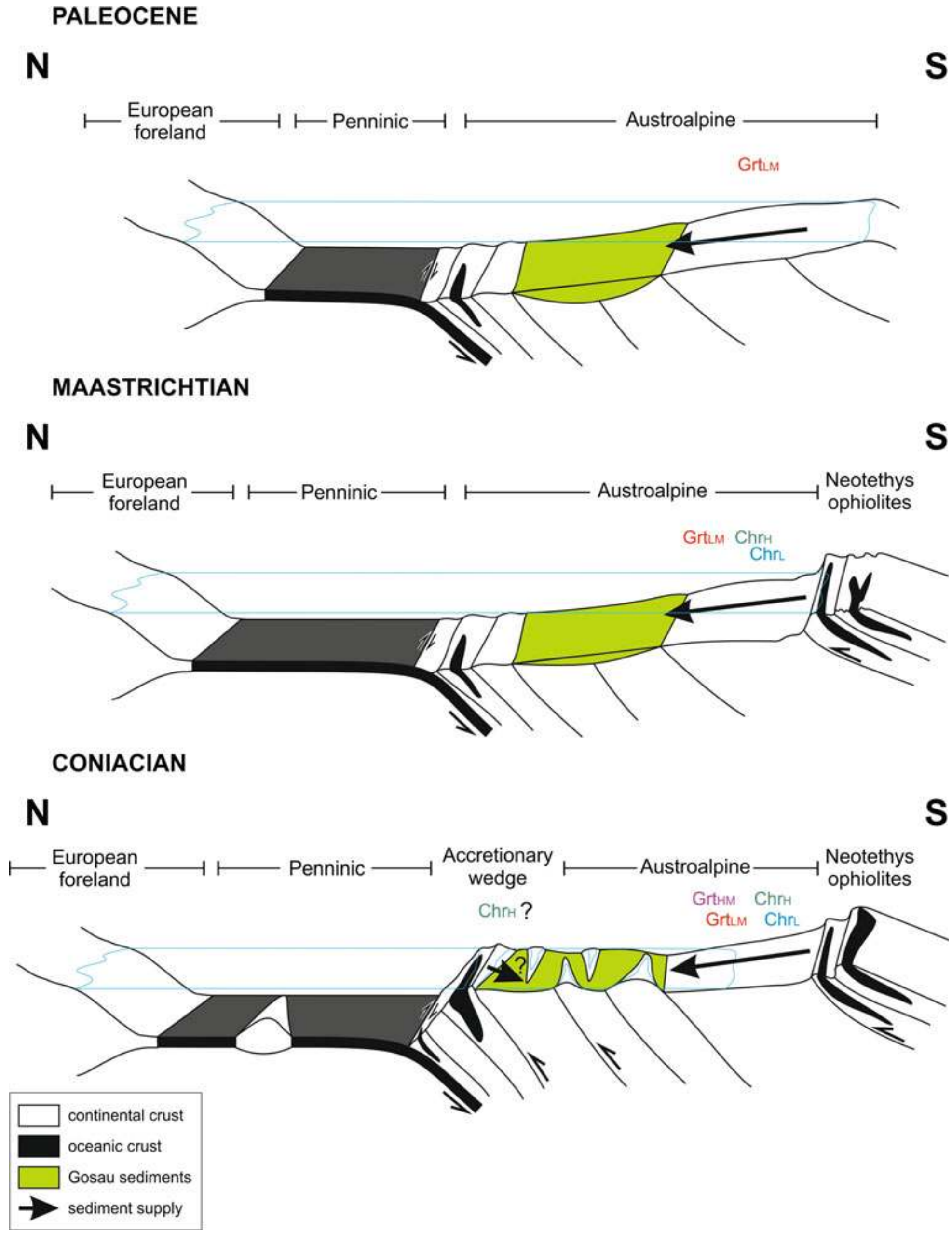


Fig. 9 Detrital tourmaline diagrams using the Fe^{tot}–Al^{tot}–Mg and Fe^{tot}–Ca–Mg triangle plots and provenance field suggested by Henry and Guidotti (1985). Plots are shown for individual age intervals with samples grouped in different Gosau basins (Gießhübl, Glinzendorf, Grünbach and Slovakian Gosau basin). Fields for interpreted provenance from where the tourmalines are supposed to be derived are given below the plots according to the authors. Plotted data from Online Resource 5

chrome spinels from Dinaric ophiolites (Maksimovic and Majer 1981), probably similar to the fully eroded NCA southern provenance, show identical chemistry.

Supplementary to dominant spinel portions, metamorphic derived minerals are important in Coniacian to Campanian heavy mineral assemblages (Fig. 2). Garnet chemistry indicates a mixed source of high- to low-grade metamorphosed provenance (Fig. 6). High amounts of garnets derived from HP/UHP conditions of granulite, eclogite and amphibolite facies can be observed in the Gießhübl, Glinzendorf and Grünbach basin. Garnets from the Slovakian Gosau basin do not suggest a higher metamorphic character, which can also be an effect of low number of samples. Garnet chemistry and spatial distribution of these high-grade metamorphosed grains point to high metamorphic units to the south as the most likely source area. The now more southerly located high-grade metamorphic basement of the Eoalpine (Cretaceous-age) eclogite belt of the Austroalpine basement nappes (e.g. Schmid et al. 2004) cannot be the source for these garnets, because cooling ages of these units are at ca. 90-80 Ma (e.g. Thöni 2006), i.e. definitely not eroded during the given Coniacian–Santonian interval. Furthermore, investigations on amphibolite pebbles (upper amphibolites facies) from Gosau Group conglomerates of the Grünbach basin suggest remnants of the metamorphic sole of Neotethyan (Vardar) ophiolite sheets as ultimate provenance (unpublished data by Ralf Schuster, Geological Survey, Austria). Similarities (e.g. symplectitic textures and cooling ages) to amphibolites from the metamorphic sole at the base of the Dinaric ophiolites are conspicuous, and are not known from Austroalpine metamorphic units. These (upper) amphibolite-facies metamorphic soles generally indicate basaltic and sedimentary protoliths (Carosi et al. 1996; Dimo-Lahitte et al. 2001; Elitok and Drüppel 2008; Mikes et al. 2008). Garnet chemistry from Ralf Schuster (unpublished data from amphibolite pebbles of the Lower Gosau Group) and Balen et al. (2003; data from the Dinaric Krivaja-Konjuh ultramafic massif, Bosnia) are identical to our results from the detrital garnets of the Lower Gosau Subgroup (Fig. 6). Similar compositions of detrital garnets are also observed in the Dinaride Ophiolite Zone mélange in Bosnia, which are attributed to the sub-ophiolitic metamorphic sole (Mikes et al. 2008).

Fig. 10 Plate tectonic model for the Eastern Alps and Western Carpathians during the deposition of the Gosau Group with associated souce areas of detritus. Types of detrital garnets and chrome spinels are marked from various provenances: Grt_{LM} low-grade metamorphosed garnets, Grt_{HM} high-grade metamorphosed garnets, Chr_{H} chrome spinels of mainly harzburgite-provenance, Chr_{L} chrome spinels of mainly lherzolite-provenance



The influence of ophiolitic detritus is present up to the Maastrichtian, with chrome spinels from an harzburgite-dominant source (Figs. 2, 4, 8, 10). At this time, detrital input only comes from the south (Wagreich and Faupl 1994). Nappes of high metamorphic facies in the south are already eroded at this time and garnet chemistry generally indicates low metamorphic metapelites as provenance (Fig. 6).

Paleogene heavy mineral assemblages are only dominated by garnet (around 80 %) and tourmaline in the Slovakian basin (Fig. 2). Chrome spinel is non-existent, except minor amounts in the Slovakian Gosau, probably due to reworking of older Gosau sediments. Garnet chemistry is analogue to the Maastrichtian from low metamorphic, probably metapelitic source (Fig. 6). Garnet chemistry is, analogous to the Maastrichtian, from the today's exposed Austroalpine Crystalline Complexes of the Eastern Alps, such as the Wölz, Rappold, Radenthein, Plankogel, Koralpe and Saualpe Complexes (Schuster and Frank 1999; Habler and Thöni 2001; Faryad and Hoinkes 2003; Bestel et al. 2009; see Fig. 11). Abundant amounts of garnet mica schist lithoclasts e.g. in the Gießhübl Formation corroborate these units as the most likely sources (Sauer 1980).

During this time interval, ophiolitic nappes including metamorphic soles, from a southern derivation were thus already eroded (Fig. 10). Local ophiolitic remnants only provided clastic material for the easternmost Slovakian

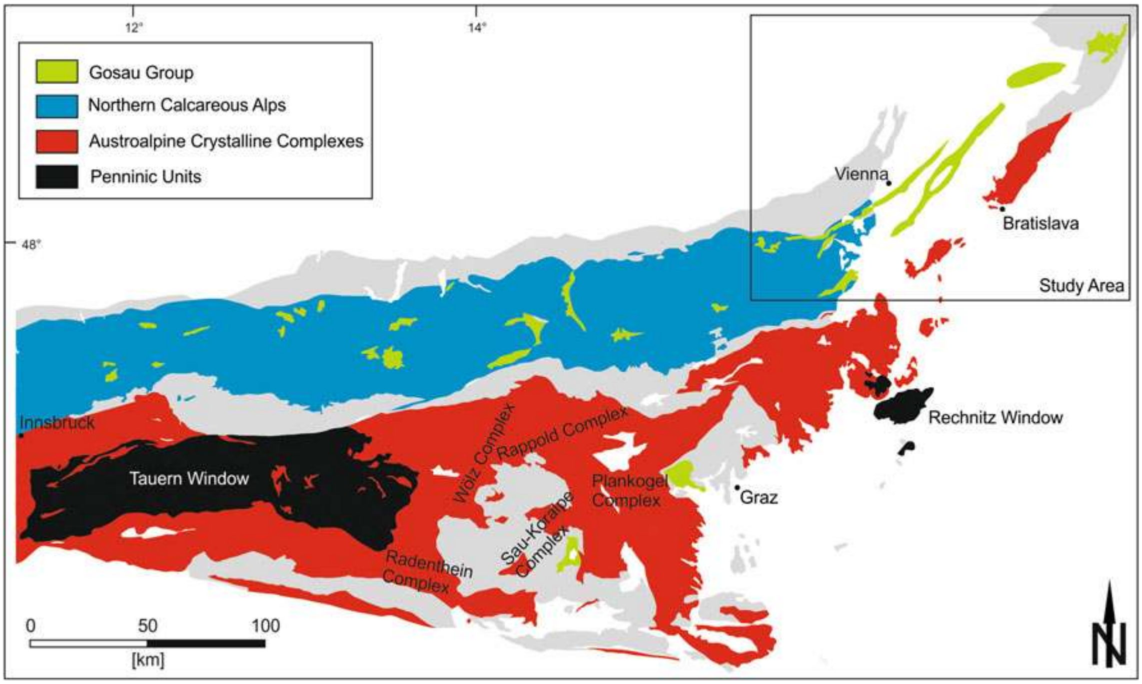


Fig. 11 Simplified geological map of the Eastern Alps (east of Innsbruck) after Egger et al. (1999). Highlighted Austroalpine crystalline complexes are supposed to be the source area for Paleogene almandine-rich garnets of the Gosau Group in the study area (grey colored units include the Rhenodanubian Flysch Zone, the Greywacke Zone, the Graz Palaeozoic, the Drau Range and the Male Karpaty Mountains in Slovakia; see also Schmid et al. 2004)

Gosau basin. Equivalent crystalline complexes, such as those in the Eastern Alps, do not exist in the Carpathians. Significantly higher ZTR-indices and slightly higher degrees of roundness of garnets from Paleogene heavy mineral assemblages in the Slovakian part compared to the Gießhübl and Grünbach basins imply a more distal palaeogeographical position of that sedimentation area.

8 Conclusions

The occurrence of detrital garnets from high-metamorphic sources, and of chrome spinels of a mixed harzburgite and lherzolite provenance in the Coniacian to Campanian Gosau Group suggests dominant sediment input from obducted ophiolites situated on top and/or south of the NCA. At this time, a mélange of high metamorphic soles and ophiolitic nappes from the hanging wall of the northwards thrust onto the Austroalpine realm acts as source area for the Gosau basins. Detrital garnets with high proportions of pyrope (and grossular) may indicate erosion from relics of a metamorphic sole. Harburgite-dominant chrome spinel, especially in the palaeogeographically northern basins, may have been derived also from a southern source and do not directly indicate influence of the Penninic accretionary wedge as a second provenance, as is also suggested by strongly mixed metapelites and granitoids tourmaline data which do not change in time. From the Maastrichtian onwards, only low- to medium-grade metamorphic garnets, mainly from metapelites of southern provenance, can be observed. Moderate chrome spinel contents represent ultimate erosion of ophiolitic structures in the south. In the Paleogene nearly no ultramafic detritus is being provided by the hinterland and garnets from low grade metamorphic source dominate the heavy mineral assemblages. These almandine-rich garnets show the same chemical composition as garnets from the Austroalpine Crystalline Complexes of the Eastern Alps.

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