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Petrology and palaeotectonic setting of Cretaceous alkaline basaltic volcanism in the Pieniny Klippen Belt (Western Carpathians, Slovakia)

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Occurrences of mafic alkaline volcanics are scattered all around Europe, being mostly related to anorogenic, extensional tectonic environments. While the widespread Cenozoic alkaline basalts have been intensively studied and are comparatively well-known, their Cretaceous precursors were often associated with the Alpine–Carpathian orogenic zones, and so their genesis and geodynamic setting are partially obscured by superimposed deformation and alteration. We describe a newly discovered body of melanephelionites inserted within the Upper Cretaceous deep-marine pelagic succession of the Pieniny Klippen Belt in Western Slovakia. The body consists of hyaloclastic lavas of nephelinitic composition. The mineralogical composition and geochemical features of the Vršatec volcanites correspond to melanephelinites. Reconstruction of the geodynamic setting of the Cretaceous mafic alkaline volcanism in the Alpine–Carpathian–Pannonian realm infers a general extensional/rifting tectonic regime that ultimately led to the opening of Penninic oceanic rift arms. However, this rifting started as basically passive and non-volcanic. Only during the later, post-breakup extension phases did the slow-spreading oceanic ridges develop, which are characterized by the MORB-type (mid-ocean-ridge basin) basaltic volcanism. Alkaline volcanic provinces have a linear character and appear to follow passive continental margins of Penninic oceanic arms opened during the Jurassic and Early Cretaceous. We infer that alkaline volcanism resulted from heating and partial melting of the subcontinental mantle lithosphere on the peripheries of asthenospheric upwellings confined to slow-spreading ridges of the Alpine Tethys. Consequently, regarding the debate about the plume *vs*. non-plume origin of the Cretaceous alkaline volcanism, the geological data from this area rather support the latter affinity.

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INTRODUCTION

Alkaline basalts and related volcanic rocks (basanites, lamprophyres *etc.*) are usually interpreted as being generated in intra-plate, anorogenic (i.e. either late post-orogenic, or pre-orogenic) tectonic environments (Wilson and Downes, 1991; Bailey and Woolley, 1999). They are commonly related to extensional tectonic regimes and occur in both continental and oceanic areas, but attained a special significance in areas characterized by continental lithosphere. There, the principal trigger of mantle-derived alkaline magmatism remains unclear – it could be either a product of small-scale, upper mantle plumes, or it formed in response to passive mantle upwelling due to lithospheric stretching and thinning (see Lustrino and Wilson, 2007 for the review).

In Western and Central Europe, mafic alkaline volcanism is common, although generally not voluminous. It occurred intermittently during the entire Phanerozoic, but is principally concentrated in two time intervals – in the Cretaceous and late Cenozoic. The late Cenozoic episode may be regarded as fully post-orogenic and extends over areas consolidated in different times during the Phanerozoic, including the young Alpine orogens. On the other hand, the mid-Cretaceous magmatism formed during the post-orogenic period in regions consolidated during the Variscan and earlier orogenies, but occurred as pre-orogenic in areas later included in the Alpine fold-thrust belts.

In the Western Carpathians, the occurrences of Cretaceous alkaline basalts can be found in various tectonic units of their central and external zones. Volcanism is represented by comparatively small portions of mostly submarine lava flows, with fewer volcaniclastic rocks, dykes and sills. Volcanism occurred in units that, at the time of its activity, were still not included in the compressional orogenic wedge but located in its foreland. In general, Cretaceous alkaline volcanism is bound to zones with an extensional tectonic regime and its effusive products always alternate with marine pelagic deposits (see Hovorka and Spišiak, 1988 and references therein).

The present paper aims at characterization of a comparatively large body of alkaline basalts occurring within an Oravic sedimentary sequence of the Vršatec Klippen area in the Púchov segment of the Pieniny Klippen Belt in Western Slovakia. This body was not known formerly and has been only recently found during geological mapping. We describe the stratigraphic position, petrology and geochemistry of these basalts, compare them to other occurrences of similar age and discuss their significance for the pre-orogenic tectonic evolution of the Western Carpathians.

GEOLOGICAL SETTING

The Pieniny Klippen Belt (PKB) is a narrow (several km), but lengthy (up to 600 km) zone dominated by Late Oligocene-Miocene wrench tectonics (Ratschbacher *et al*., 1993; Nemčok and Nemčok, 1994; Kováč and Hók, 1996; Schlögl *et al*., 2008). It separates the External Western Carpathians (EWC – "Flysch Belt", Penninic units, a Tertiary accretionary complex overriding the North European Platform) from the Central Western Carpathians (CWC – Austroalpine units, Cretaceous basement/cover nappe stack – Fig. 1), and involves Jurassic to Paleogene strata of extremely variable lithology and intricate internal structure. During more than a century of detailed research, these have been subdivided into numerous lithostratigraphic and tectonic units of distant provenances, hence witnessing excessive shortening and subsequent dispersal within this restricted zone (e.g., Birkenmajer, 1977, 1986; Froitzheim *et al*., 2008). The PKB is therefore considered as a suture, in spite of a lack of ophiolite complexes. In many places, the PKB is formed of isolated blocks of "klippen" (rigid Middle Jurassic to Lower Cretaceous limestones) embedded in the "klippen mantle" (soft Lower Jurassic and Upper Cretaceous to Paleogene shale, marl and flysch formations). Consequently, the PKB has often been characterized as a tectonic megabreccia or mélange. However, this peculiar "block-in-matrix" structure of the PKB appears to be a result of later stages of the deformation history of the PKB units, governed by Early Miocene along-strike transpressional and transtensional movements. These obliterated, in places completely, Late Cretaceous-Paleogene thrusting-related structures. Nowadays, the PKB usually centres on a positive flower structure – the outer forward-thrust limb is located within the rear parts of the frontally accreted Biele Karpaty and/or Magura units, while back-thrusts of the inner limb intensely affected the PKB/CWC boundary zones (e.g., Marko *et al*., 2005).

In accordance with its position at the boundary of two major parts of the Carpathian orogenic system, the PKB involves units derived from both. The EWC tectonic system includes the Magura and Biele Karpaty superunits consisting of Upper Cretaceous-Oligocene sedimentary formations composed mostly of flysch, and the Oravic Superunit comprising Lower Jurassic-Upper Cretaceous deposits subdivided into several nappe units, which are the most characteristic and distinctive PKB units. However, the PKB and zones adjacent to its inner side involve also nappe units derived from the CWC nappe systems, dominantly frontal partial nappes of the Fatric Superunit. In addition to these, four sets of overstepping sedimentary successions that partly seal older structures may be discerned: Senonian, Paleocene–Early Eocene, Late Eocene–Early Miocene and Mid–Late Miocene.

The Oravic Superunit includes klippen of the ridge-derived Czorsztyn Succession (including the largest klippe of this succession in the entire PKB – the Vršatec Klippen), the basinal Kysuca and Pieniny successions, and several types of "transitional" successions. These comprise sedimentary successions from the Early Jurassic onward. The Oravic Klippen occur along the northwestern margin of the PKB, forming a narrow (a few kilometres) "Klippen Belt *sensu stricto*". The broader southeastern part is mostly built of the "non-Oravic" units (Drietoma, Klape and Manín), with occasional tectonic windows of Oravic units and the Senonian-Paleogene overstepping cover (e.g., Schlögl *et al*., 2008). This zone was designated as the "Periklippen Zone" by Maheľ (1980).

The Czorsztyn Unit represents a former ridge or swell environment and is characterized by prevailing shallow-water facies during the Jurassic and Early Cretaceous (Mišík, 1979, 1994). It is interpreted as being derived from a subducted continental ribbon in a middle Penninic position (e.g., Birkenmajer, 1986; Plašienka, 2003). Due to strong dissection of the ridge caused by several rifting events, the sedimentary formations show considerable lateral and vertical variations that were used for distinction of several sedimentary successions. Thus the Czorsztyn-type successions exhibit some variability, but are generally composed of: (1) Middle Liassic to Aalenian deep-water, partly anoxic bioturbated marlstones and black shales; (2) Bajocian very shallow-water, sandy-crinoidal limestones, or in places scarp breccias; (3) after the upper Bathonian break-up unconformity, a thick succession of condensed "ammonitico rosso" nodular limestones deposited; (4) Tithonian-Berriasian coquinas, breccias, pale maiolica-type limestones and Neocomian crinoidal limestones have been preserved in places only. After an unconformity indicated by numerous neptunian dykes and partial erosion down to the Dogger deposits (Barremian-Aptian gap), the locally karstified surface (Aubrecht *et al*., 2006) was covered by an Albian hardground followed by (5) deepening Upper Cretaceous variegated "couches rouges" marlstones and (6) Maastrichtian-Eocene flysch.

The Pieniny Unit includes basinal successions with continuous stratigraphic sequences ranging from the Early Jurassic to the Late Cretaceous. The most widespread Kysuca Succession consists of: (1) lowermost Jurassic syn-rift siliciclastics; (2) Lower to Middle Jurassic hemipelagic spotted marlstones passing gradually into (3) Callovian-Oxfordian radiolarites; (4)

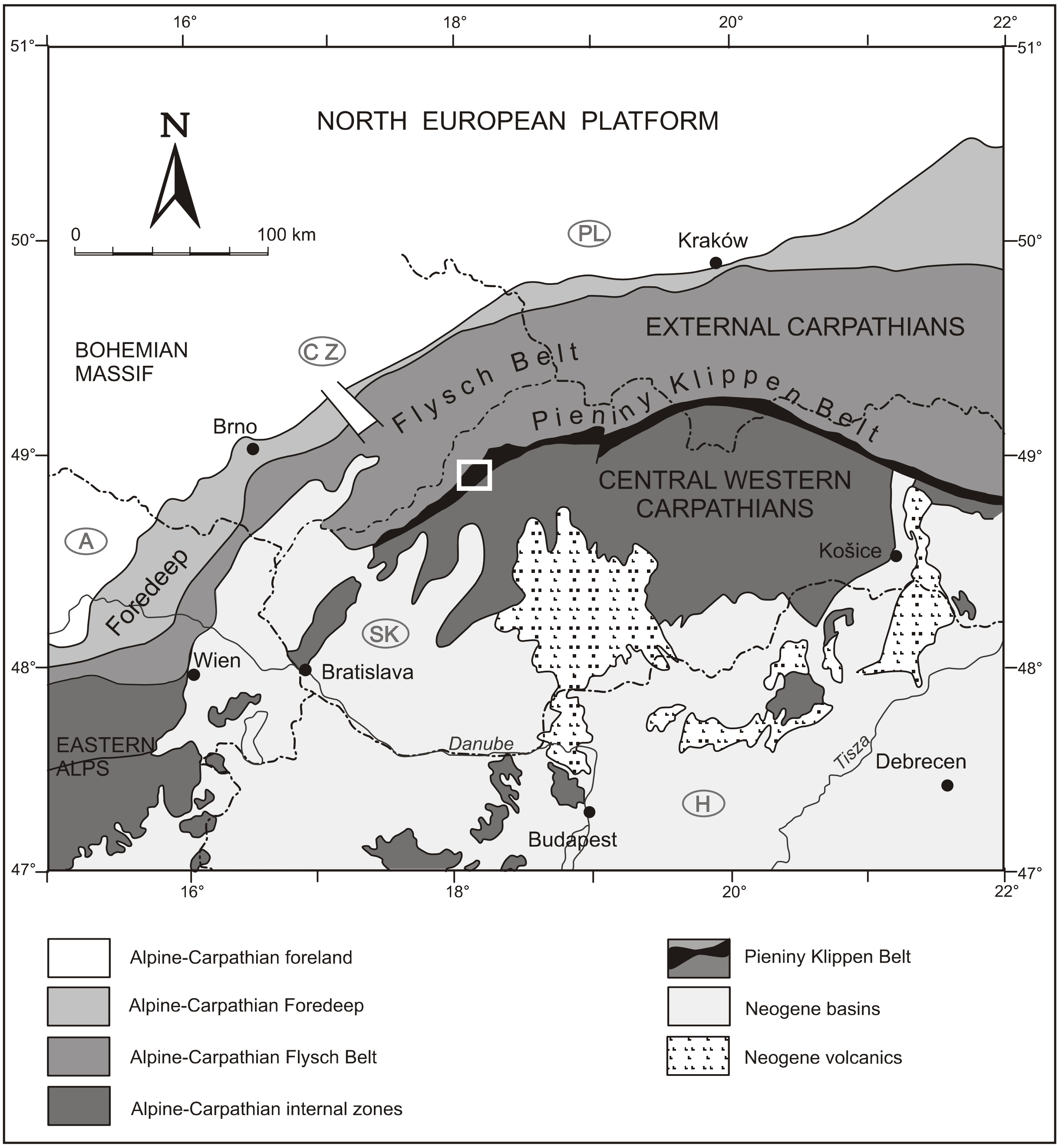

Fig. 1. General tectonic sketch of the Western Carpathians with position of the study area (white rectangle)

Kimmeridgian red nodular limestones; (5) Tithonian-Neocomian cherty limestones; (6) various mid-Cretaceous hemipelagic marlstones; (7) Turonian-Santonian coarsening-upward "exotic" flysch and (8) Campanian red pelagic "Globotruncana" marls.

The newly discovered volcanic rocks here described occur in the Chmeľová region of the so-called Vršatské bradlá group of klippen in the Púchov segment of the PKB (Fig. 1). The Oravic, Czorsztyn, Pruské and Kysuca successions crop out in a small area here, being deformed in a complex fold-fault structure (Fig. 2). To the NW, the Kysuca Unit is juxtaposed to the flysch deposits of the Biele Karpaty Unit of the EWC Flysch Belt. This contact is followed by a large wrench fault that forms the northern boundary of the PKB.

Bodies of basic volcanic rocks were discovered in two localities within the variegated marls of Late Cretaceous age. The first, smaller occurrence (a few metres in diameter) is situated on the northern slopes of Chmeľová Hill. The second, much larger occurrence is situated SSW-wards of Chmeľová Hill on the SE slopes of the unnamed crest about 1 km SW of Vršatec hotel (GPS coordinates of the approximate centre of the body: N 49°03 52.9, E 18°08 34.6). This occurrence is represented by three volcanic bodies, the largest body being up to 100 m wide and over 500 m long (Fig. 2).

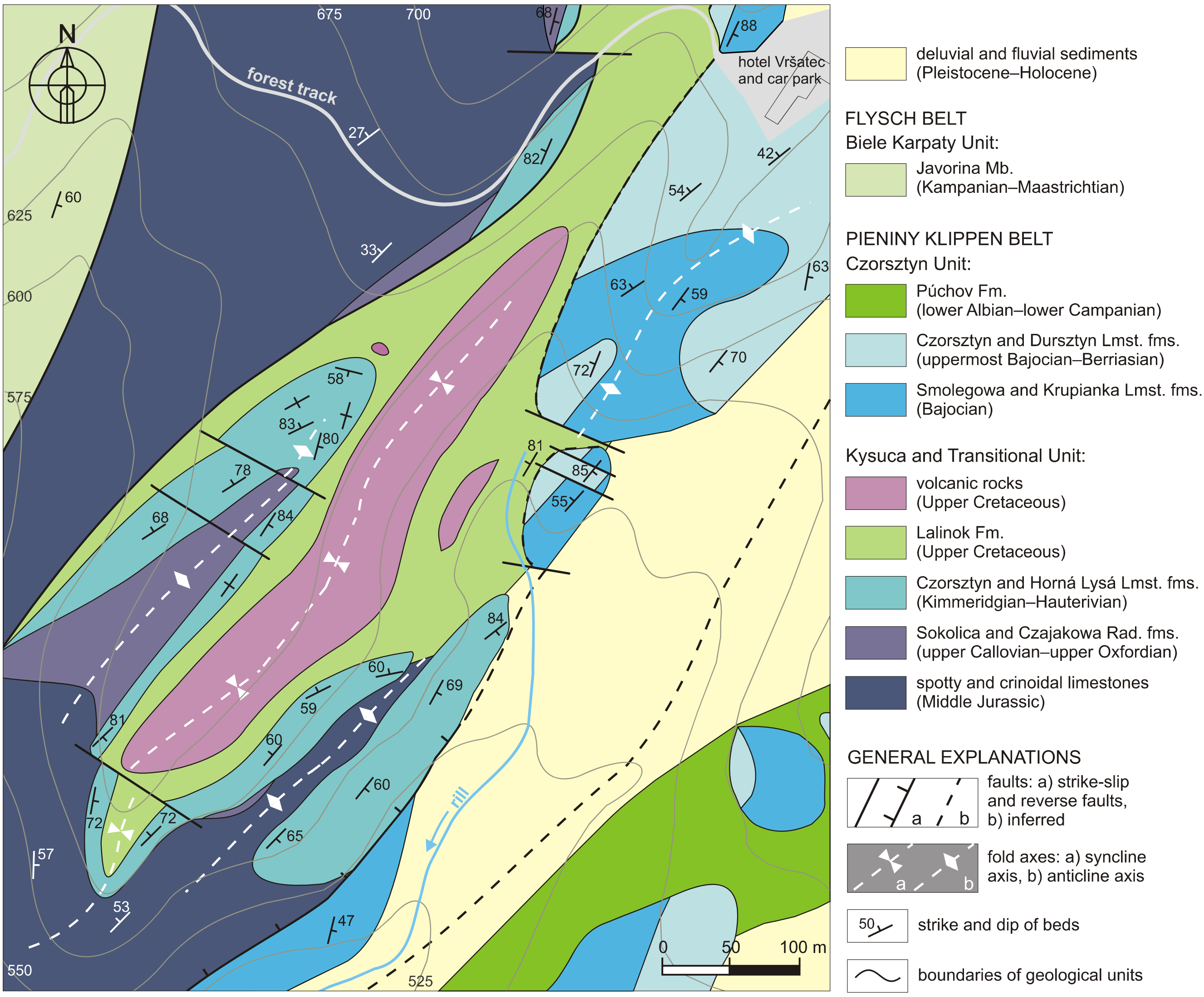

Fig. 2. Geological map of a part of the Vršatec Klippen area showing position of the volcanites studied

The area is dominated by macroscopic fold structures with SW–NE trending axes and younger, mainly strike-slip faults (Fig. 2). The macrofolds are asymmetric, non-cylindrical with a locally penetrative, steeply NW-dipping axial plane cleavage. The bulk of the volcanic rocks is situated in the core of a tight brachysyncline. The map view and analysis of fault structures reveal that folding was followed by dominantly strike-slip faulting (SW–NE and younger W–E to WNW–ESE trending faults) that finally shaped the klippen tectonic style of the area.

The sedimentary succession containing the volcanic body analysed cannot be assigned without reservations to any typical Oravic successions described in the literature. As the oldest member, it contains dark grey spotty limestones with intercalations of sandy crinoidal limestones and spongiolites, which are partly analogous to the Samášky Fm. described from the transitional Pruské Succession (Aubrecht and Ožvoldová, 1994), or to the Flaki Limestone Fm. known from the Branisko (Kysuca) Succession in Poland. This formation is overlain by greenish and red platy radiolarites (Sokolica and Czajakowa Limestone fms.), followed by red nodular limestones (Czorsztyn Limestone Fm.). The Lower Cretaceous deposits are of a special type with pinkish allodapic bioclastic limestones (Horná Lysá Fm. – Mišík *et al*., 1994). Brick-red marlstones, which can be possibly correlated with the Cenomanian Lalinok Fm., and the volcanic rocks, are the youngest members of this succession (Fig. 3). The succession described bears features of either a non-typical Kysuca Succession, or a transitional Pruské Succession. Palaeogeographically, it most probably occupied a position between these two, i.e. along a distal slope of the Czorsztyn Ridge at a transition to the Pieniny Basin. Towards the east, the succession considered is in lateral juxtaposition with the typical Czorsztyn Succession (Fig. 2) containing white and red sandy-crinoidal limestones (Smolegowa and Krupianka fms., respectively), red nodular "ammonitico rosso" limestones (Czorsztyn Limestone Fm.) and pink biodetrital "Calpionella" limestones (Dursztyn Limestone Fm.).

The volcanic rocks lie in superposition over variegated marlstones of the Lalinok Fm. In some places, a thermal contact with the marlstones was observed. The volcanic bodies consist of hyaloclastic breccias in a larger scale they represent

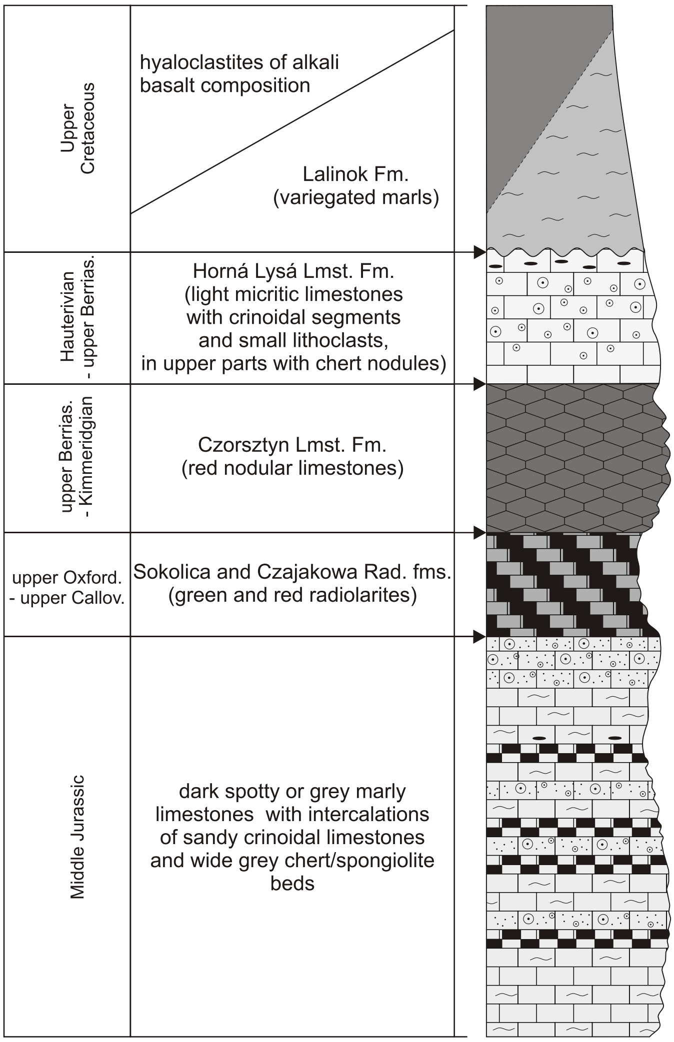

Fig. 3. Lithostratigraphic profile of the sedimentary succession with alkaline basalts studied

basaltic submarine lava flows, which discharged into a carbonate depositional environment. Determination of an approximate age of the volcanic rocks is based on the foraminiferal tests found in micritic, partly thermally recrystallized carbonate forming an interstitial substance, fissure fillings and fragments in hyaloclastites. Foraminifers, together with carbonate ooze, penetrated into the voids of the brecciated lava body either during or shortly after its solidification. The tests found represent keeled planktonic foraminifers belonging to the family Globotruncanidae (Fig. 4D; determined by Š. Józsa). These are characteristic of the Cenomanian-Maastrichtian time interval. Accordingly, the basaltic lavas were emplaced during the Late Cretaceous.

ANALYTICAL METHODS

Rock-forming minerals and Fe-Ti spinels were analysed with a wave-dispersion (WDS) electron microprobe *Cameca SX-100* and photographed in back-scattered electron (BSE) made at the Geological Survey of the Slovak Republic (Bratislava) under the following conditions: 15 kV accelerating voltage, 20 nA beam current, beam diameter 2–5 µm, ZAF corrections, standards (n – natural, sy – synthetic) – wollastonite (n) for Si and Ca, Al_2O_3 (sy) for Al, rodonite (n) for Mn, fayalite (n) for Fe, forsterite (n) for Mg, TiO_2 (sy) for Ti, NiO (sy) for Ni, Cr (sy) for Cr, willemite (n) for Zn. $Fe^{2+}$and Fe^{3+}in spinels were calculated assuming an ideal stoichiometry. The composition of the minerals is shown in Tables 1–4.

The apatite composition was measured by *Cameca SX-100* electron microprobe in the wavelength mode at the Institute of Geological Sciences, the Masaryk University, Brno (Czech Republic). The following analytical conditions were used: accelerating voltage of 15 kV, beam current of 10 nA, beam diameter of 3–5 µm, element measurement time of 20–40 s. The following natural and synthetic standards were used: apatite (for P Kα, Ca Kα, F Kα), sanidine (Si Kα), cheralite (Th Mα), metallic U (U Mβ), $CeAl_2$(Ce Lα), almandine (Fe Kα), spessartine (Mn Kα), $MgAl_2O_4$(Mg Kα), $SrSO_4$(Sr Lα), albite (Na Kα) and vanadinite (Cl Kα). The lower detection limit of the microprobe was ~0.18 wt.% for F, and 0.02–0.1 wt.% for other measured elements. Contents of As, Y, La, Pr, Nd and Ba were below the detection limit. Standard deviation was *ca*. ±1 wt.% for F, ±0.4 to 0.6 wt.% for Ca and P, and ±0.05 to 0.1 wt.% for the other measured elements. The measured data were corrected by the PAP routine.

Whole-rock powder samples were used for analyses. From a large (approx. 6 kg) generally fresh sample we separated pure dark rocks (Fig. 4A, B). By quartering we divided the sample into two and then we analysed them. The values in Table 5 are averages of two analyses. Major and trace elements were determined by ICP OES and ICP MS in Acme Analytical Laboratories Ltd. Canada (Table 5). The conditions of analyses and detection limits of elements are given in Acme Analytical Laboratories Ltd. (2008).

MINERALOGY AND PETROLOGY

The texture of the volcanics is not homogeneous, approaching that of hyaloclastites or breccias (Fig. 4A, B). In the light grey-green fine-grained vitreous matrix, there are irregular sharp-edged chips of coarse-grained volcanics (2–20 mm) and, very rarely, also sharp-edged clasts of carbonates or thin carbonate veinlets (Fig. 4B). The matrix is composed of devitrified glass, small albite grains, microlites of clinopyroxenes, amphiboles and zeolites. Volcanic clasts are traced by thin calcite veinlets. The mineral composition of the coarse-grained clasts corresponds to basalt/basanites. The most frequent mineral phases in the volcanic clasts are clinopyroxene, amphibole and illmenite, less frequent is apatite and Fe-Ti-spinel (Fig. 5). Other minerals detected are pseudomorphs after olivines, alkali feldspars and analcime.

Clinopyroxene (Cpx) is zonal and forms porphyric structures (Fig. 5A–E) together with amphibole, apatite and illmenite. Zoning is marked by irregular alternation of lighter and darker phases of different composition, which indicates

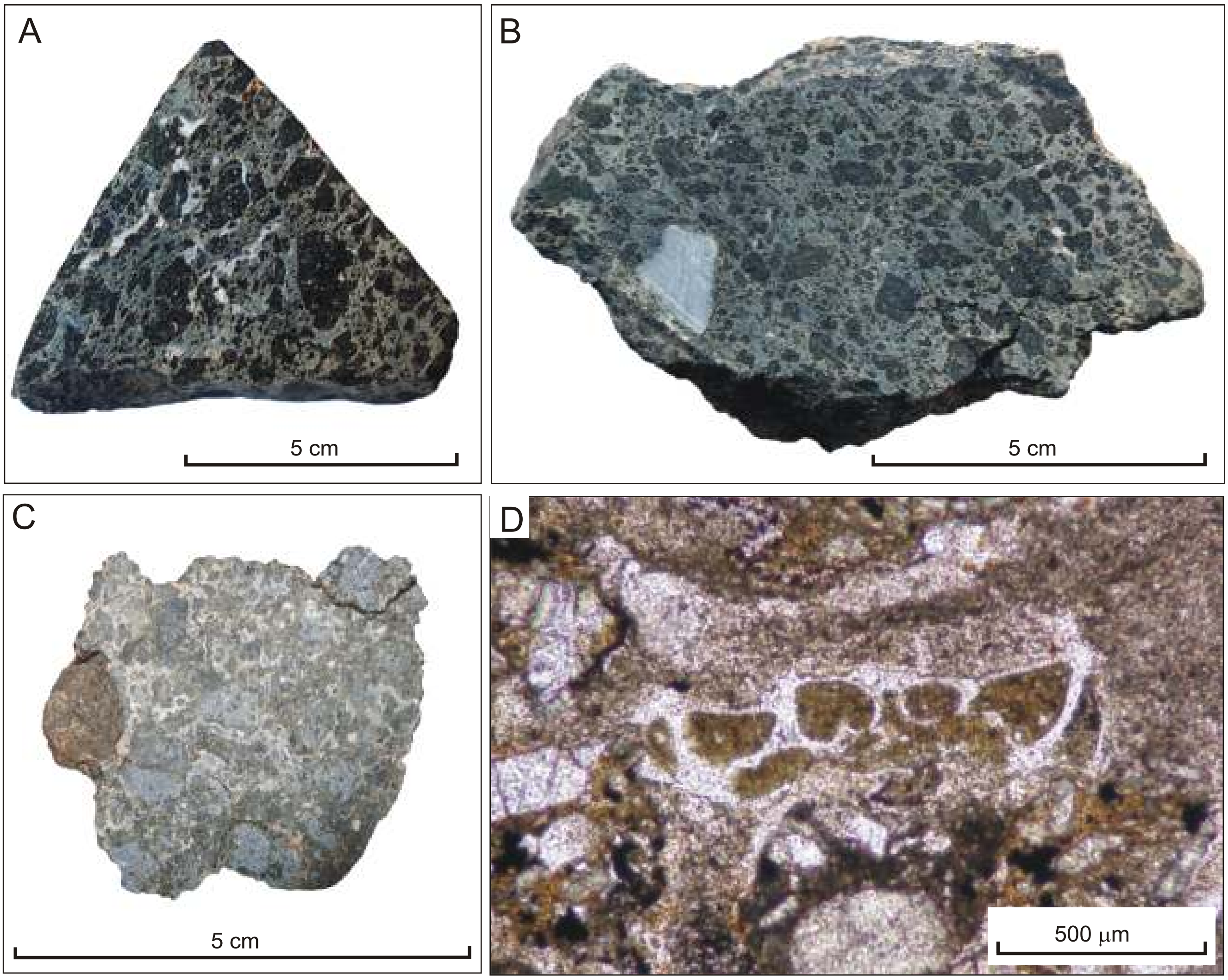

Fig. 4A–C – photomacrographs of melanephelinites with hyaloclastic brecciated structure (dark clasts of melanephelinites and pale clasts of limestones in a vitreous matrix); D – thin section of limestone with test of foraminifer from the family Globotruncanidae

rapid cooling of magma. The lighter rims are rich in Ti, Al, Fe and partly also Na and depleted in Si and Mg. Sector zoning was occasionally observed in addition to oscillatory zoning. Based on the Morimoto´s *et al*. (1988) classification, the composition of Cpx (Table 1) corresponds to diopside (Fig. 6). It correlates well with the composition of magmas and their tectonic position. We used discrimination diagrams by Nisbet and Pearce (1977) and Leterrier *et al*. (1982). The Cpx projection points are plotted in the alkali basalt fields in both diagrams. This position documents the alkali character of volcanics and their similarity to the Cretaceous alkali volcanics of the External and Central Western Carpathians (Hovorka and Spišiak, 1988). Amphiboles (Fig. 5D) are also slightly oscillatory zoned; they are less frequent than Cpx and have a specific composition (Table 2) – increased contents of TiO_2 (3.9 wt.%) as well as Na_2O and K_2O (2.5 and 1.9 wt.%, respectively). In the IMA classification (Leake *et al*., 1997) they correspond to pargasite. Some volcanic fragments contain a higher amount of apatite. Two types of apatite differing in morphology, size and composition were distinguished (Fig. 7A, B). Apatite 1 forms euhedral to subhedral porphyric crystals (phenocrysts), 150 to 800 µm in size, in association with other large phenocrysts (feldspars, Fe-Ti oxides). Apatite 2 consists of small (20 to 50 µm), long prismatic to acicular subhedral crystals in the finely crystalline to glassy matrix (groundmass) of the rock (Fig. 7B). Both apatite generations show homogeneous hydroxylapatite composition without chemical zoning across the crystals. Apatite 1 shows lower Si, Ce, Fe, Sr and F contents, and higher P, Mg and Ca contents in comparison to apatite 2; e.g., 0.3–0.5 and 0.8–1.1 wt.% SiO_2, 0.02–0.15 and ~0.3 wt.% Ce_2O_3, 0.3–0.4 and 0.6–0.8 wt.% FeO, 0.3–0.5 and ~0.8 wt.% SrO, respectively (Table 3). The F/(F+OH) ratio attains 0.25 to 0.29 in apatite 1, and 0.33 to 0.36 in apatite 2.

The alkaline character of the rock is documented also by the presence of partly resorbed leucite (Fig. 7C). Other minerals important for the magma genesis are the Fe-Ti spinels, that occur as an accessory phase. The size of the spinel grains is up to 400 µm. The spinels show strong alteration visible in BSE images (Fig. 7E, F), or in reflected polarized light. Due to alteration, spinels are inhomogeneous and their "vermicular" tex-

Table 1

Representative analyses of clinopyroxenes

No	1	4	5	6	8	9	10	11	12
SiO_2	51.50	48.70	49.33	51.69	48.97	49.61	51.75	50.28	51.81
TiO_2	1.06	1.84	1.72	1.01	1.98	1.61	1.06	1.32	1.10
Al_2O_3	2.30	3.03	4.08	2.27	3.92	3.56	2.10	2.72	2.34
Cr_2O_3	0.01	0.01	0.00	0.00	0.02	0.00	0.00	0.01	0.00
FeO_{tot}	7.62	6.02	8.21	8.03	7.75	7.61	7.58	11.15	7.35
MnO	0.23	0.20	0.20	0.21	0.17	0.19	0.25	0.40	0.24
MgO	14.52	14.49	13.11	13.97	13.64	13.28	14.11	11.10	14.35
CaO	23.39	24.01	22.25	22.48	22.89	22.75	22.73	21.54	22.80
Na_2O	0.80	0.60	0.99	0.82	0.81	0.91	0.85	1.47	0.71
K_2O	0.00	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.01
Total	101.45	98.90	99.89	100.48	100.16	99.56	100.47	99.99	100.71
Formula based on 6 O									
Si	1.90	1.84	1.85	1.92	1.83	1.86	1.92	1.91	1.92
Ti	0.03	0.05	0.05	0.03	0.06	0.05	0.03	0.04	0.03
Al	0.10	0.14	0.18	0.10	0.17	0.16	0.09	0.12	0.10
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe^{3+}	0.10	0.12	0.09	0.06	0.10	0.09	0.07	0.09	0.06
Fe^{2+}	0.13	0.07	0.17	0.19	0.14	0.15	0.17	0.26	0.17
Mn	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Mg	0.80	0.82	0.73	0.77	0.76	0.74	0.78	0.63	0.79
Ca	0.92	0.97	0.89	0.89	0.92	0.92	0.90	0.88	0.90
Na	0.06	0.04	0.07	0.06	0.06	0.07	0.06	0.11	0.05
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	4.05	4.06	4.05	4.03	4.05	4.04	4.03	4.05	4.03

FeO_{tot} – total Fe as FeO

ture is strongly porous. The major constituent of Fe-Ti spinels (titanomagnetites) is magnetite (43–46 mol%). The amount of ulvöspinel (Usp) component ranges between 32 and 34 mol%. According to the spinel classification (Deer *et al.*, 1992), the spinels studied correspond to ulvöspinels (Fig. 8). Selected microprobe analyses are shown in Table 4. The content of Al_2O_3 is up to 3.4 wt.% and MgO 4.5 wt.%. The dominant substitution trends are tetrahedral Mg $\leftrightarrow$ Fe^{2+}substitution, followed by $2Fe^{3+} \leftrightarrow Fe^{2+} + Ti^{4+}$exchange and $Fe^{3+} \leftrightarrow Ti^{4+}$substitution (Fig. 9). The charge balance is unequal due to inverse spinel structure. The substitution trends described are identical with those reported for other occurrences of Mesozoic alkali basalts in the Western Carpathians (Mikuš *et al.*, 2006), as well as in the Madeira Island alkaline lava (Mata and Munhá, 2004). Fe-Ti spinels from Vršatec and CWC alkali rocks are somewhat poorer in Ti ($Usp_{8–38}$) than those in other alkali basalt occurrences (e.g., $Usp_{52–71}$; Cornen and Maury, 1980, *etc*.). The Ti-depletion of the studied Fe-Ti spinels can be explained in terms of alteration processes (Fig. 9). Fe-Ti spinels in alkali basalts are richer in minor components (Al and Mg) than those from tholeiitic rocks and andesites. The distinctive feature of Fe-Ti spinels from alkaline rocks is their tendency towards a high Al_2O_3 content (Frost and Lindsley, 1991).

The chemical composition of the volcanics is rather specific (Table 5). Generally, these rocks are characterized by low SiO_2 contents (*ca.* 37.0 wt.%), enhanced contents of TiO_2 and P_2O_5 (3.0 wt.%) and elevated contents of incompatible elements such as Ba (1300 ppm), Sr (1100 ppm) and LREE, as well as those of Nb (217 ppm), V (161 ppm) and Zr (1050 ppm). Generally low SiO_2 contents and high CaO contents are partially influenced by carbonate content (amygdales or carbonate veins) in the samples analysed. The classification of these rocks is difficult. According to different classification diagrams (TAS, Beard *et al.*, 1998, and others) we can characterize these rocks as ultrabasic rocks and picrobasalts. For classification we used the classification of basanitic and nephelinitic volcanic rocks by Le Bas (1989). This classification is based on the CIPW norm. According to it basanites were recognized as having >5% normative *ab* and <20% normative *ne*, melanephelinites as having <5% normative *ab* and <20% normative *ne*, and nephelinites as having >20% normative *ne*. Our volcanites have 1.2% *ab* and 12.9% *ne*. This suggests that we can call them melanephelinite.

For various discrimination diagrams (Pearce and Cann, 1973; Mullen, 1983; Meschede, 1986, *etc*.) these volcanics correspond to OIA (ocean island alkali basalt) or WPA (within-plate alkali basalt) fields (Fig. 10). Although melanephenite typically have higher contents of some elements including REE (with a lower degree of partial melting in comparison with alkali basalts), we are comparing them with alkali basalts and/or Cretaceous alkali basalt/basanites of the Western Crapathians, as we infer their similarity in age and geotectonic setting. High contents of REE as well as those of some other elements have been influenced by a high apatite content (P_2O_5 3.0 wt.%) in the separated coarse-grained volcanites. This is a high content and it is likely to have been a result of local accu-

Table 2

Representative analyses of amphibole

Sample	B-1	B-1	B-1	B-1	B-1	B-2
No	1	2	11	12	13	6
SiO_2	40.05	40.87	40.38	40.90	40.82	41.99
TiO_2	3.80	4.04	3.86	3.85	3.82	3.64
Al_2O_3	11.45	11.90	11.63	11.78	11.88	11.15
Cr_2O_3	0.00	0.04	0.02	0.00	0.02	0.00
FeO_{tot}	13.59	12.57	13.57	12.44	13.51	12.79
MnO	0.19	0.13	0.24	0.17	0.17	0.20
MgO	12.34	12.95	12.50	13.21	12.49	13.34
CaO	11.16	11.35	11.12	11.21	11.02	11.17
Na_2O	2.49	2.50	2.57	2.55	2.53	2.68
K_2O	2.00	1.96	1.95	1.97	1.96	1.81
Total	97.37	98.47	97.95	98.50	98.46	98.95
Formula based on 23 O						
Si	6.031	6.044	6.021	6.047	6.046	6.160
Al^{IV}	1.969	1.956	1.979	1.953	1.954	1.840
Al^{VI}	0.063	0.119	0.065	0.099	0.119	0.087
Ti	0.430	0.450	0.433	0.429	0.426	0.402
Cr	0.000	0.005	0.002	0.000	0.002	0.000
Fe^{3+}	0.334	0.248	0.377	0.342	0.386	0.340
Fe^{2+}	1.377	1.306	1.315	1.195	1.287	1.229
Mg	2.771	2.855	2.777	2.913	2.758	2.918
Mn	0.025	0.017	0.030	0.021	0.021	0.024
Ca	1.801	1.799	1.777	1.776	1.749	1.755
Na^{M4}	0.199	0.201	0.223	0.224	0.251	0.245
Na^{A}	0.527	0.516	0.520	0.506	0.475	0.517
K	0.384	0.370	0.371	0.371	0.371	0.338
Total	15.912	15.885	15.891	15.877	15.846	15.855
Fe/Fe+Mg	0.332	0.314	0.321	0.291	0.318	0.296

For explanation see Table 1

mulation of apatite (and/or ilmenite) in the melt. The course of the normalized REE curve is clearly in the direction of low HREE contents without a considerable Eu-anomaly (Fig. 11). This curve is typical of nephelinitic rocks. Compared to basanites of identical geotectonic position (Pieniny Klippen Belt, Jarnuta Fm., Hanigovce – Spišiak and Sýkora, 2009) they are richer in REE. They display the same elevated REE contents as do Cretaceous alkali basalts/basanites from the Central Western Carpathians (Hovorka *et al*., 1999). Such high REE contents are typical of strongly alkali nephelinite-type rocks (e.g., Ulrych *et al*., 1998). A similar pattern is seen in the mantle-normalized trace element abundances (Fig. 12). Melanephelinites from Vršatec have higher contents of all elements compared except for Rb and Sr compared to basaltoids from the Pieniny Klippen Belt (Jarmuta Fm., Biala Woda – Birkenmajer and Lorenc, 2008; Hanigovce – Spišiak and Sýkora, 2009) and higher than in other Cretaceous alkaline basalts occurring in the Central Western Carpathians. Such a course of the mantle-normalized trace element curve for the basaltoids from other localities of the Pieniny Klippen Belt and Cretaceous alkaline basalts from the Central Western Carpathians is typical of ocean island alkali basalts (OIB).

DISCUSSION

GENERAL FRAMEWORK OF THE MESOZOIC MAFIC ALKALINE VOLCANISM IN CENTRAL EUROPE

Following the widespread and voluminous Permian continental tholeiitic to calc-alkaline, and only minor alkaline basaltic volcanism (e.g., Timmerman, 2008), the Triassic was a generally quiet volcanic period outside the Tethyan mobile belts. However, the latest Triassic-earliest Jurassic breakup of Pangaea and opening of the central Atlantic Ocean was associated with voluminous, probably plume-derived magmatism (Hill, 1991; Oyarzun *et al*., 1997; Golonka and Bocharova, 2000). Several rift arms propagated from the central Atlantic north- and westwards during the Mid Jurassic, but oceanic spreading first affected the eastern branch named the Alpine Tethys. It included the presently sutured domains known as the Piemont–Ligurian (South Penninic) oceanic realms. There the rifting was initially non-volcanic and the sea-floor was first formed by the subcontinental mantle exhumed during asymmetrical passive extension leading to continental breakup (e.g., Lemoine *et al*., 1987; Trommsdorff *et al*., 1993; Marroni *et al*., 1998). Passive mantle unroofing was then followed by MORB type volcanism along a slow-spreading centre (Koller and Höck, 1992; Dürr *et al*., 1993; Piccardo *et al*., 2004). Within-plate, mildly alkaline basalts occur rarely within the Penninic domains and were reported from the Eastern Alps (Koller, 1985; Höck and Miller, 1987; Koller and Höck, 1992; Frisch *et al*., 1994). Their age is unknown, but most probably they formed during the Late Jurassic and/or Early Cretaceous.

On the other hand, the NE-propagated Atlantic Rift arm did not result in continental breakup, but an extensive and complex rift system rejuvenated earlier Permian grabens in the future north Atlantic realm and its western European margin (e.g., Ziegler, 1988). The North Sea triple junction of the Viking, Central and Moray Firth Grabens was marked by extensive alkaline basaltic volcanism in the Forties volcanic province during the Mid Jurassic (Latin and White, 1990). Sources of this volcanism were sought in a rapid mantle upheaval and consequent partial melting of the asthenosphere and lithosphere due to lithospheric stretching and thinning (Latin and Waters, 1992).

During the Early Cretaceous, a new rift arm propagated north-eastwards from the central Atlantic, which separated Iberia and its Briançonnais promontory from Laurasia by the Galicia, Bay of Biscay and Valais (North Penninic) oceanic domains. Further to the east, the Valais Rift probably merged with the Jurassic Piemont–Ligurian Ocean (Trümpy, 1988; Liati *et al*., 2005; Schmid *et al*., 2008). However, even further east, the Early Cretaceous rifting affected the European margin as well, and formed several continental splinters, such as the Oravic (Czorsztyn) and Silesian ridges (e.g., Golonka *et al*., 2000). This rift branch was yet again essentially non-volcanic, but followed by several Late Cretaceous alkaline mafic volcanic fields at the Galician margin and Pyrenean foreland (referred to as the Iberian Province – Rock, 1982; see also Azambré *et al*., 1992; Tavares Martins, 1999). The eastern, Alpine–Carpathian sector of this rift branch (Rhenodanubian–Magura) is accompanied by volumetrically unimpor-

Table 3

Representative analyses of apatite

	1	2	3	4	5	6	7	8	9	10
P_2O_5	41.57	41.92	41.86	41.12	41.11	40.56	41.74	41.74	41.27	41.71
SiO_2	0.46	0.50	0.42	0.49	0.83	1.13	0.29	0.45	0.39	0.39
ThO_2	0.01	0.01	0.00	0.02	0.00	0.02	0.00	0.00	0.00	0.00
UO_2	0.00	0.01	0.06	0.02	0.00	0.00	0.00	0.06	0.03	0.00
Ce_2O_3	0.14	0.15	0.12	0.14	0.31	0.26	0.02	0.13	0.07	0.12
FeO	0.32	0.34	0.38	0.39	0.57	0.82	0.44	0.27	0.33	0.39
MnO	0.07	0.06	0.07	0.07	0.06	0.06	0.07	0.07	0.06	0.05
MgO	0.27	0.25	0.23	0.27	0.18	0.22	0.44	0.22	0.33	0.30
CaO	55.14	55.12	54.49	54.16	54.17	53.40	54.98	54.75	54.66	54.68
SrO	0.44	0.48	0.47	0.48	0.77	0.77	0.28	0.46	0.38	0.40
Na_2O	0.30	0.28	0.29	0.18	0.23	0.21	0.30	0.28	0.28	0.29
H_2O^*	1.30	1.32	1.29	1.26	1.12	1.15	1.27	1.26	1.23	1.29
F	0.97	0.95	0.98	1.02	1.31	1.22	1.02	1.06	1.07	0.99
Cl	0.13	0.13	0.14	0.11	0.14	0.13	0.15	0.12	0.14	0.10
O=F	–0.41	–0.40	–0.41	–0.43	–0.55	–0.51	–0.43	–0.45	–0.45	–0.42
O=Cl	–0.03	–0.03	–0.03	–0.02	–0.03	–0.03	–0.03	–0.03	–0.03	–0.02
Total	100.68	101.09	100.36	99.28	100.22	99.41	100.54	100.39	99.76	100.27
Formula based on 12 O and (OH+F+Cl) = 1 apfu										
P	2.946	2.955	2.969	2.953	2.933	2.916	2.957	2.961	2.950	2.962
Si	0.039	0.042	0.035	0.042	0.070	0.096	0.024	0.038	0.033	0.033
Sum B	2.985	2.997	3.004	2.995	3.003	3.012	2.981	2.999	2.983	2.995
Th	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
U	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.001	0.001	0.000
Ce	0.004	0.005	0.004	0.004	0.010	0.008	0.001	0.004	0.002	0.004
Fe	0.022	0.024	0.027	0.028	0.040	0.058	0.031	0.019	0.023	0.027
Mn	0.005	0.004	0.005	0.005	0.004	0.004	0.005	0.005	0.004	0.004
Mg	0.034	0.031	0.029	0.034	0.023	0.028	0.055	0.027	0.042	0.038
Ca	4.945	4.917	4.892	4.922	4.891	4.859	4.929	4.916	4.944	4.914
Sr	0.021	0.023	0.023	0.024	0.038	0.038	0.014	0.022	0.019	0.019
Na	0.049	0.045	0.047	0.030	0.038	0.035	0.049	0.045	0.046	0.047
Sum A	5.080	5.049	5.028	5.047	5.044	5.030	5.084	5.039	5.081	5.053
OH	0.725	0.731	0.720	0.711	0.631	0.654	0.709	0.702	0.694	0.723
F	0.257	0.250	0.260	0.274	0.349	0.328	0.270	0.281	0.286	0.263
Cl	0.018	0.018	0.020	0.016	0.020	0.019	0.021	0.017	0.020	0.014
Sum X	1.000	0.999	1.000	1.001	1.000	1.001	1.000	1.000	1.000	1.000

H_2O^* – total H_2O

tant, but numerous intrusions and extrusions of Lower Cretaceous mafic alkaline magmatic rocks. These embrace the southern – Austroalpine margin of the Alpine Tethys (Trommsdorff *et al.*, 1990), along with the CWC Tatric–Fatric and the Tisia terrane Mecsek-Alföld volcanic fields described below. The northern margin was affected by similar alkaline volcanism, e.g. in the St. Veit Klippen Zone of the Eastern Alps and the extensive Moravian–Silesian volcanic field of the EWC, as well as in its prolongation in the Eastern Carpathian area (see below).

During the Late Cretaceous, the alkaline volcanism moved further forelandwards into areas later influenced by a widespread Cenozoic magmatism, e.g. along the Ohře (Eger) Rift graben (Ulrych *et al.*, 2008). The final opening of the north Atlantic Ocean and the Labrador Sea in the Paleogene was then closely related to the Icelandic hot spot accompanied by voluminous plume-generated magmatism (e.g., White *et al.*, 1987; Jones, 2003). The Cenozoic alkaline volcanism in Western Europe might or might not have been related to this plume activity (Lustrino and Wilson, 2007).

Over the last two decades, the possible geodynamic sources of this volcanism have been widely debated. While some authors argued for the plume-generated opening of central Atlantic as the primary source for the subsequent passive margin alkaline volcanism (e.g., Hoernle *et al.*, 1995; Oyarzun *et al.*, 1997), others found no necessity for large-scale mantle plume activity and relate the rifting and volcanism to the interactions of local mantle upwellings and pre-existing continental lithospheric weak zones (e.g., McHone, 2000). Concerning the European Cretaceous alkaline volcanism, Oyarzun *et al.* (1997) proposed the thin-spot model whereby a strongly thinned and rifted European realm was influenced by NE-wards divected sublithospheric plume channelling.

Table 4

Representative microprobe analyses of Fe-Ti spinels from Vršatec

SiO_2	0.10	0.10	0.26	0.08	0.10	0.05	0.07	0.09	0.28	0.14	0.14	0.08
TiO_2	14.32	14.34	14.39	14.80	14.15	14.08	13.97	13.83	14.23	14.27	14.05	6.37
Al_2O_3	3.23	3.21	3.15	3.29	3.32	3.09	2.83	3.28	3.29	3.36	3.37	0.68
Cr_2O_3	0.20	0.07	0.03	0.01	0.10	0.15	0.36	0.07	0.03	0.01	0.03	0.00
*Fe_2O_3	50.69	50.67	50.66	50.81	51.07	51.93	52.28	52.17	49.71	51.97	51.36	60.70
FeO	23.41	23.12	22.74	23.90	24.33	23.95	23.86	23.92	22.94	24.25	24.00	28.83
MnO	0.62	0.71	0.85	0.67	0.56	0.61	0.71	0.72	0.96	0.64	0.60	1.49
MgO	4.21	4.23	4.47	4.07	3.81	4.12	4.13	4.10	4.05	4.02	4.02	0.15
CaO	0.04	0.11	0.04	0.02	0.04	0.02	0.05	0.02	0.03	0.02	0.02	0.00
ZnO	0.10	0.10	0.04	0.03	0.06	0.03	0.02	0.11	0.12	0.20	0.10	0.11
NiO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
Total	96.93	96.66	96.63	97.66	97.54	98.04	98.29	98.33	95.63	98.88	97.71	98.41
Formula based on 3 cations												
Si	0.004	0.004	0.010	0.003	0.004	0.002	0.003	0.004	0.011	0.005	0.005	0.003
Ti	0.407	0.408	0.409	0.418	0.401	0.396	0.393	0.388	0.410	0.398	0.396	0.187
Al	0.144	0.143	0.140	0.145	0.147	0.136	0.125	0.144	0.148	0.147	0.149	0.031
Cr	0.006	0.002	0.001	0.000	0.003	0.004	0.011	0.002	0.001	0.000	0.001	0.000
Fe^{3+}	1.440	1.443	1.440	1.434	1.445	1.461	1.469	1.463	1.431	1.449	1.449	1.779
Fe^{2+}	0.739	0.731	0.718	0.750	0.765	0.749	0.745	0.745	0.734	0.752	0.752	0.939
Mn	0.020	0.023	0.027	0.021	0.018	0.019	0.022	0.023	0.031	0.020	0.019	0.049
Mg	0.237	0.239	0.252	0.227	0.214	0.230	0.230	0.228	0.231	0.222	0.224	0.009
Ca	0.001	0.004	0.002	0.001	0.002	0.001	0.002	0.001	0.001	0.001	0.001	0.000
Zn	0.003	0.003	0.001	0.001	0.002	0.001	0.000	0.003	0.003	0.005	0.003	0.003
Ni	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000

*Fe_2O_3 – calculated from stoichiometry

Table 5

Chemical composition of the rocks studied (average of two analyses)

SiO_2	36.89	Ba	1357.4	Y	64.3	Zr	1052.2
TiO_2	3.30	Be	6	La	161.8	Sc	9
Al_2O_3	11.10	Co	28.5	Ce	367.2	Mo	0.9
Cr_2O_3	0.01	Cs	0.6	Pr	42.92	Cu	15.8
Fe_2O_{3tot}	15.79	Ga	31.4	Nd	178.3	Pb	10.8
MnO	0.36	Hf	22.5	Sm	29	Zn	224
MgO	6.12	Nb	217.1	Eu	8.53	As	1.4
CaO	12.61	Rb	30.3	Gd	21.7	Cd	0.5
Na_2O	2.73	Sn	5	Tb	3.16	Sb	0.1
K_2O	1.91	Sr	1107.3	Dy	13.96	Bi	0.1
P_2O_5	3.00	Ta	13.5	Ho	2.23	Au^{++}	5.4
LOI	5.70	Th	16.2	Er	5.39	Hg	0.01
Total	99.52	U	5.8	Tm	0.73	Tl	0.1
		V	161	Yb	3.98	W	0.8
		Ni	38	Lu	0.54		

Oxides in wt.%; other elements in ppm; Au^{++} in ppb; LOI – loss on ignition; Fe_2O_{3tot} – total Fe as Fe_2O_3

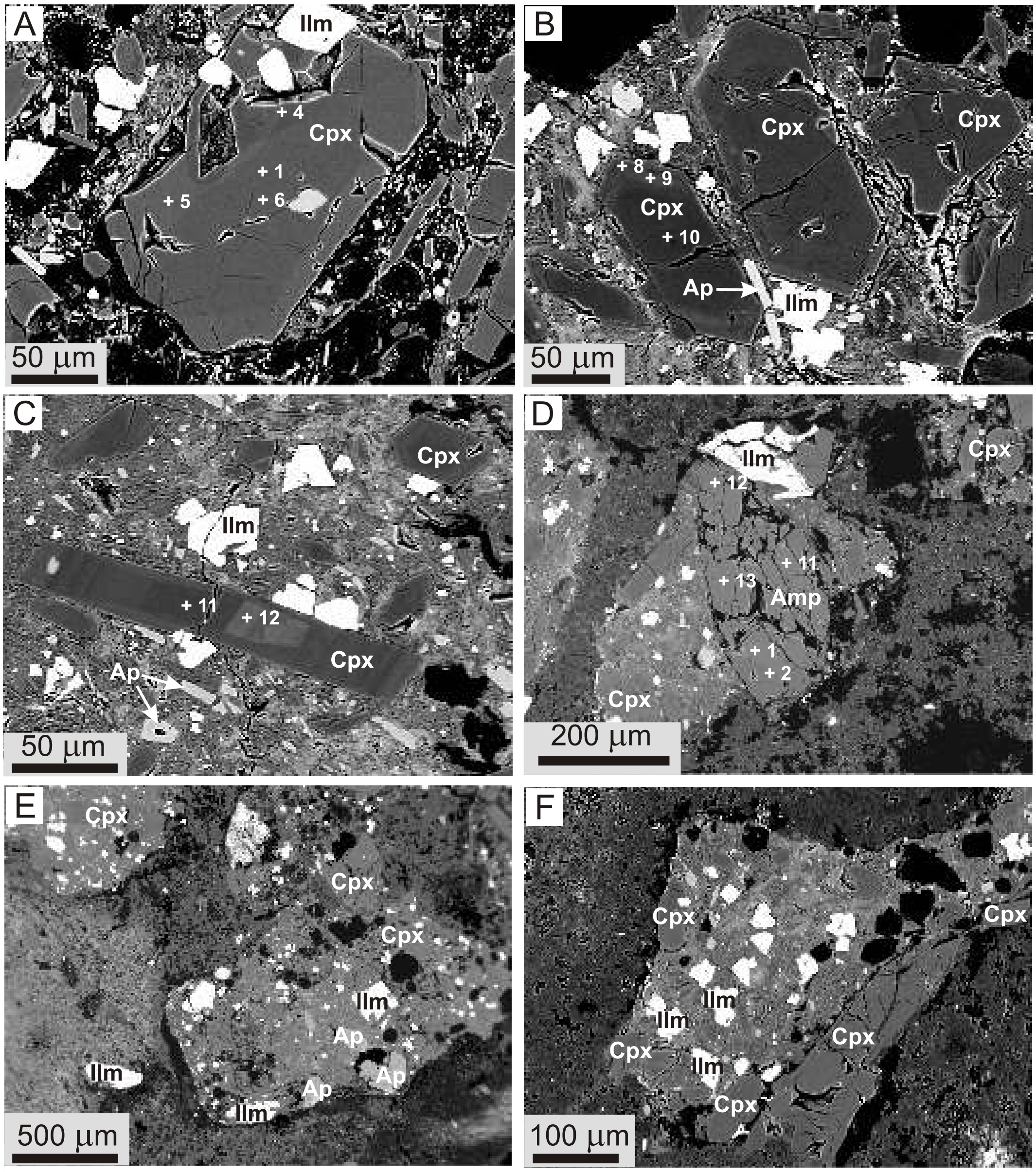

Fig. 5. Back-scattered electron images of clinopyroxenes (A–C, E, F) and amphiboles (D) with number of analyses (Tables 1 and 2)

Amp – amphibole, Ap – apatite, Cpx – clinopyroxenes, Ilm – ilmenite

OVERVIEW OF CRETACEOUS ALKALINE BASALTS IN THE CARPATHIAN–PANNONIAN REALM

The Carpathian–Pannonian area includes several Europe-derived terranes that were assembled by complex orogenic processes during the Cretaceous and Cenozoic (Schmid *et al*., 2008). Assuming a tentative mid-Cretaceous palinspastic situation (Fig. 13), the mafic alkaline volcanism occurred in the foreland of a developing orogenic wedge which propagated outwards from the internal zones adjoining the Late Jurassic Meliata Suture. These areas embraced both the northern and southern margins of the spreading Penninic oceanic basins (Alpine Tethys), and are presently dispersed in various tectonic units of the Alpine–Carpathian–Pannonian realm.

The northern Penninic passive margin is represented by the largest Carpathian Early Cretaceous alkaline volcanic region that extends in a belt some 100 km long in the Moravian–Silesian territories of NE Czech Republic and Southern Poland. The volcanics are confined to the detached sedimentary complexes of the western part of the Silesian Nappe of the EWC Flysch Belt and represent a classic area of the teschenite or teschenite-picrite association (e.g., Tschermak, 1866). These volcanic rocks are associated with deep marine clastic deposits – dark shales and calcareous turbiditic sandstones, Tithonian-Albian in age. K-Ar radiometric ages reported by Harangi and Árva-Sós (1993) range from 96 to 128 Ma, those obtained by Grabowski *et al*. (2003) are even more scattered between 63 and 148 Ma. Though affected

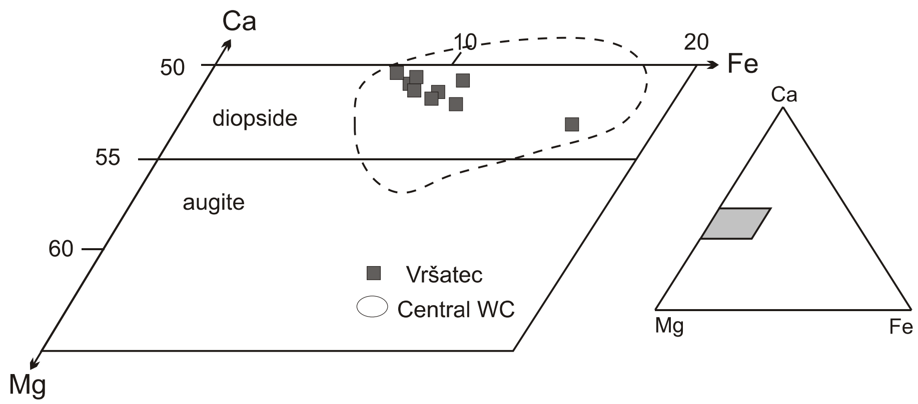

Fig. 6. Classification diagram of clinopyroxenes (Morimoto *et al*., 1988)

Central WC – composition of clinopyroxenes from Cretaceous alkali volcanites from the Central Western Carpathians (Hovorka and Spišiak, 1988; Spišiak and Hovorka, 1997; Hovorka *et al*., 1999)

by hydrothermal alteration and/or excess argon, the latter authors argue for the presence of extrusive rocks within the biostratigraphically dated Valanginian-Hauterivian deposits, therefore radiometric ages up to 140 Ma seem to be reliable. On the other hand, ^{40}Ar/^{39}Ar dating from some localities provided ages in a very narrow span from 120.4 ±1.3 to 122.4 ±1.1 Ma (Lucińska-Anczkiewicz *et al*., 2002). A characteristic feature of the whole association is a prevalence of shallow subsurface sills and veins over effusive or volcaniclastic rocks, pointing to a comparatively deep basin and a high water column. Products of the volcanic activity are fairly well differentiated (from picrites to syenites), resulting in various types of intrusive, effusive and extrusive rocks (picrites, alkali basalts, basanites, teschenites and monchiquites). On the basis of mineral composition, they have been further discriminated into less abundant types (camptonites, ankaratrites, fourchites, ouachites *etc*., for details see Mahmood, 1973; Kudělásková, 1987; Hovorka and Spišiak, 1988; Narębski, 1990; Dostal and Owen, 1998). This volcanic field is here referred to as the Moravian–Silesian Volcanic Field (MSVF).

The eastern lateral prolongation of the MSVF may be probably found in the Outer Dacide units of SW Ukraine and Northern Romania (Oszczypko *et al*., 2005). There, the Upper Jurassic to Lower Cretaceous basalts occur in several Outer Carpathian Flysch Belt units appearing in front of the Marmarosh-Bucovinian basement nappes (Vulkhovchik, Kaminnyj Potik, Trostianets, Vezhany and Rakhiv–Chivchin–Black Flysch nappes). Various geochemical signatures for volcanic rock succession in this region were reported – from ophiolitic tholeiites to intraplate alkaline basalts (Hovorka, 1996 and references therein; Varitchev, 1997; Krobicki *et al*., 2004, 2008).

Further south-east, similar volcanics are present in the Black Flysch and Ceahlău nappes (Outer Dacides – Săndulescu, 1990, 1994) in the Eastern Carpathians of Romania. Rift-related volcanism of Late Jurassic to earliest Cretaceous age is represented by calc-alkaline, tholeiitic and alkaline basalts (Lupu and Zacher, 1996; Badescu, 1997), as well as by a large intrusive syenitic body – the Ditrău alkaline massif (e.g., Dallmeyer *et al*., 1997) accompanied by numerous lamprophyre dykes, which were emplaced in the Bucovinian pre-Alpine basement (Median Dacides). This magmatism is related to the rift arm designated as the Chivchin–Ceahlău–Severin Ocean. In addition to true ophiolites (Jurassic Severin Ocean), the associated within-plate basalts indicate an intracontinental position of this narrow, probably transtensional oceanic rift (Badescu, 1997; Schmid *et al*., 2008).

In the Pieniny Klippen Belt of intra-Penninic position (Oravic continental fragment in Fig. 13), Mesozoic volcanic rocks are generally rare. Volcanics occur more abundantly in the lateral structural prolongations of the PKB, which, however, are presently considered to represent more external palaeogeographic zones with respect to the typical Oravic units of the PKB. In the Eastern Alps, Mesozoic volcanics are moderately abundant in the Ybbsitz and St. Veit Klippenzones (e.g., Schnabel, 1992). These zones separate frontal Austroalpine units of the Northern Calcareous Alps and the Northern Penninic Rhenodanubian Flysch Belt in a position analogous to the PKB. While the Ybbsitz Zone includes dismembered Jurassic MORB-type ophiolites, and is therefore ranged to the South Penninic Piemont–Ligurian oceanic realm (Schnabel, 1992; Froitzheim *et al*., 1996; Schmid *et al*., 2008), the St. Veit Klippenzone near Vienna is rich in mid-Cretaceous volcanics described as picrite dykes, lavas and tuffs intercalated within the Albian-Cenomanian red shales (Janoschek *et al*., 1956; Prey, 1975). Along with the associated Kahlenberg Nappe, the St. Veit Klippenzone is palaeogeographically placed in the southern part of the North Penninic Rhenodanubian Basin (Trautwein *et al*., 2001), i.e. in a position analogous to the Carpathian Magura (Biele Karpaty) Basin.

The eastern structural prolongation of the PKB is to be found in the Poiana Botizii Klippen area of Northern Romania (Birkenmajer, 1986; Oszczypko *et al*., 2005). Here a deep-water pelagic Upper Jurassic-Lower Cretaceous succession is underlain by presumably Callovian amygdaloidal basalts, basaltic tuffs (cinerites) and volcaniclastic sandstones (Săndulescu *et al*., 1982; Bombiţă and Pop, 1991). Palaeogeographically, this succession presumably represents a basinal area north of the Oravic ridge, i.e. the Magura Basin floor (Bombiţă *et al*., 1992). The geochemical character of the volcanic rocks is poorly known, however. They were described as basalts to basaltic andesites (Bombiţă and Savu, 1986).

The sporadic volcanic occurrences within the PKB *sensu stricto* were summarized by Mišík (1992). This author de-

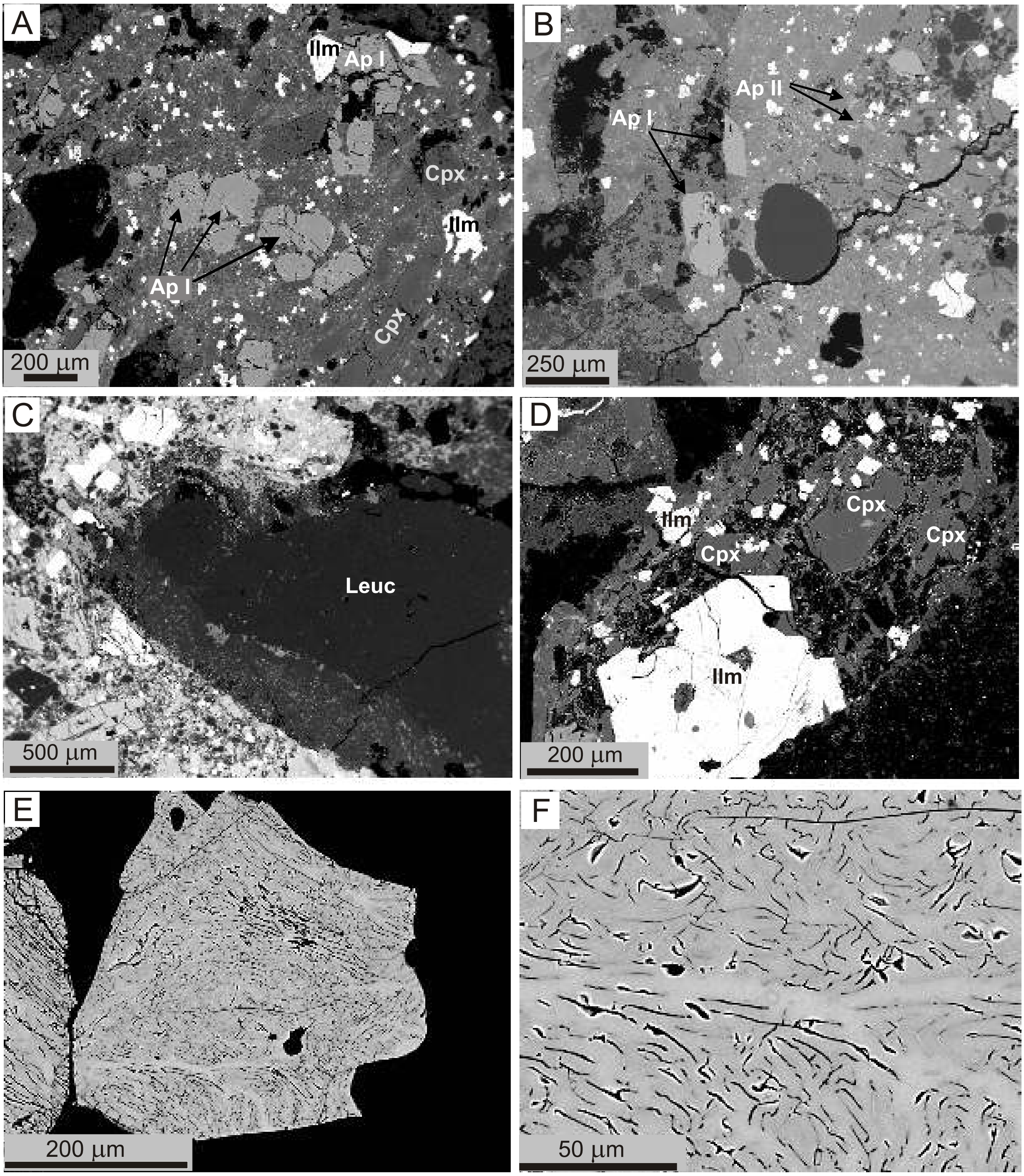

Fig. 7. Back-scattered electron images: A, B – apatite; C – leucite; D – ilmenite; E, F – strongly alterated Fe-Ti spinels

Ap I – apatite 1, Ap II – apatite 2; Leuc – leucite; for other explanations see Figure 5

scribed also pyroclastic admixtures in distal sandy turbidites in Upper Jurassic deep-water Oravic successions (Mišík *et al.*, 1991). The only larger, presently known volcanic occurrence in the PKB was described from the Velyky Kamenets quarry near Novoselica in Western Ukraine (Krobicki *et al.*, 2004, 2008 and references therein). The Berriasian or younger amygdaloidal basalts and basaltic tuffs show a within-plate, alkaline character (Krobicki *et al.*, 2008).

In the Orava region of NW Slovakia, a small exposure of strongly altered basaltic tuffs (palagonites) inserted within the Upper Cretaceous variegated marls of the Czorsztyn Unit was described by Aubrecht (1997). Some other patchy occurrences of Upper Cretaceous tuffite layers are cited by Mišík (1992).

In Western Slovakia, small basalt exposures have been reported from a few places. In vicinity of Bošáca village near Trenčín, hyalobasanitic lavas crop up within lower Albian marly limestones of the Drietoma or Manín Unit (Kullmanová and Vozár, 1980). However, these units most probably represent the frontal elements of the CWC Fatric nappe system located in the Periklippen Zone neighbouring the PKB *s.s.* Con-

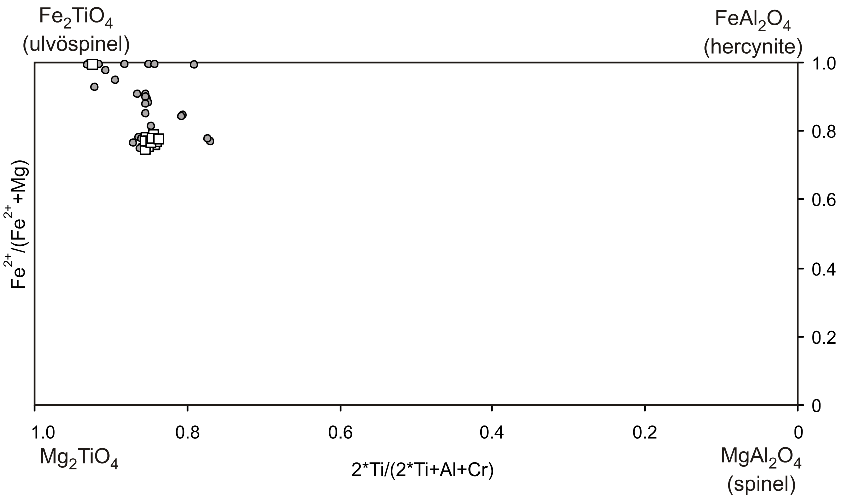

Fig. 8. Nomenclature and composition of spinels based on classification of Deer *et al*. (1992)

Studied spinels (white squares, Table 4) are compared with Fe-Ti spinels of the Central Carpathian Mesozoic alkali basalts – grey circles (Mikuš *et al*., 2006)

ing of two samples yielded ages of 110.6 ±4.2 Ma and 120.3 ±4.5 Ma (Birkenmajer and Pécskay, 2000). Two similar basaltic olistoliths residing within the Proč (Jarmuta) Fm. near Hanigovce village in eastern Slovakia have been recently studied by Spišiak and Sýkora (2009). From the point of view of their position, petrology and geochemistry, they are almost identical to the Jaworki olistolith.

The drifted European continental fragments south of the Penninic Ocean include the Austroalpine, Central Western Carpathians (CWC – eastern prolongation of the Austroalpine domain) and the Tisza–Dacia terrane (e.g., Csontos *et al*., 1992). In the CWC, the primitive mafic alkaline volcanic rocks occur in the Tatric, Fatric and probably also in the Hronic superunits (*cf*., Zorkovský, 1949; Hovorka and Spišiak, 1988; Hovorka *et al*., 1999 for a review). Volcanites of the Tatric–Fatric Volcanic Field (TFVF) are mostly of basanite type (fewer picrites), or their volcaniclastics. In the Tatric sedimentary cover, hyalobasanitic lava flows form stratiform bodies within the Lower Cretaceous deep-marine sequences in the Malé Karpaty, Low and High Tatra Mts. Besides these, basanite dykes in the Tatric basement granitoid rocks have also been reported (Hovorka *et al*., 1982*a*, *b*; Spišiak *et al*., 1991). Six K-Ar datings from these dykes are dispersed between 93 and 115 Ma (Spišiak and Balogh, 2002).

More frequent, though small TFVF lava bodies are known from the Barremian-Albian hemipelagic sedimentary successions of the Fatric Krížna and Manín nappes (Zorkovský, 1949; Hovorka and Sýkora, 1979; Hovorka and Spišiak, 1988). The products of this volcanic activity are weakly-differentiated rocks of basalt/basanite type and rarely picrites that occur as veins in the Triassic carbonates (Spišiak and Hovorka, 2000). Volcaniclastic rocks, mostly hyaloclastites, are quite common. The age of this volcanism, based on stratigraphic and geochronological data, is mid-Cretaceous – generally Aptian to early Albian (K-Ar isotopic ages 106 and 116 Ma; Bujnovský *et al*., 1981). Picrite veins near Poniky village in the Banská Bystrica district penetrate the Triassic carbonates of either the

sequently, they belong to the Tatric–Fatric Volcanic Field described below. A few other, very small occurrences of hyaloclastic basalts (augitites – Zorkovský, 1949) were reported from the Púchov sector of the PKB in the vicinity of Mikušovce and Streženice, only several km NE of our locality. Their position amidst Upper Cretaceous variegated marls also closely resembles our finding.

In addition to these volcanic occurrences in the PKB, Mesozoic igneous rocks of various characters can be found as pebbles in conglomerate bodies of the Cretaceous to Paleogene flysch formations. Their provenance is questionable; commonly they are thought to be derived from an "exotic" source located south of the present PKB (see e.g., Mišík and Sýkora, 1981; Birkenmajer, 1988; Mišík and Marschalko, 1988; Demko *et al*., 2008).

At Biała Woda near the village of Jaworki in the Polish PKB, a remarkable solitary, few metres in diameter basaltic olistolith resides within the uppermost Cretaceous conglomerates of the Jarmuta Formation (Birkenmajer and Wieser, 1990). This basalt has an alkaline, within-plate geochemical signature and was interpreted as being derived from the Czorsztyn Ridge (Birkenmajer and Lorenc, 2008). K-Ar dat-

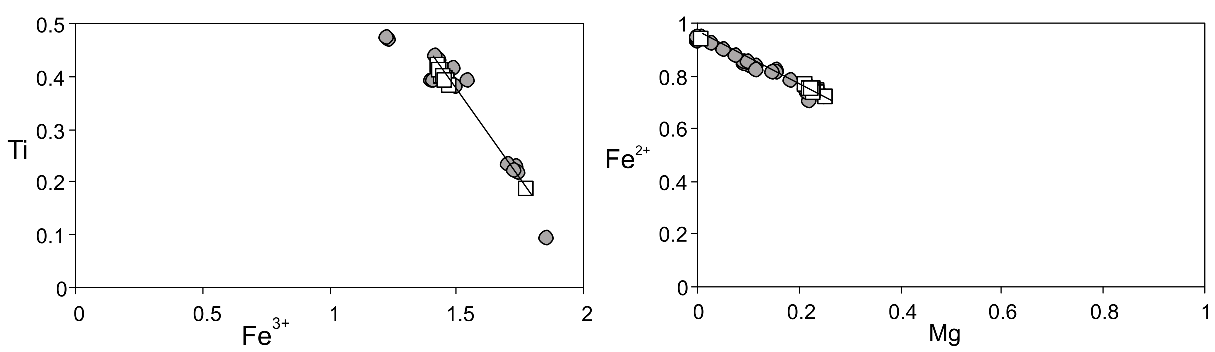

Fig. 9. Correlation of major elements in Fe-Ti spinels from Vršatec

Studied spinels (white squares) are compared with Fe-Ti spinels of Mesozoic alkali basalts (grey circles; Mikuš *et al*., 2006)

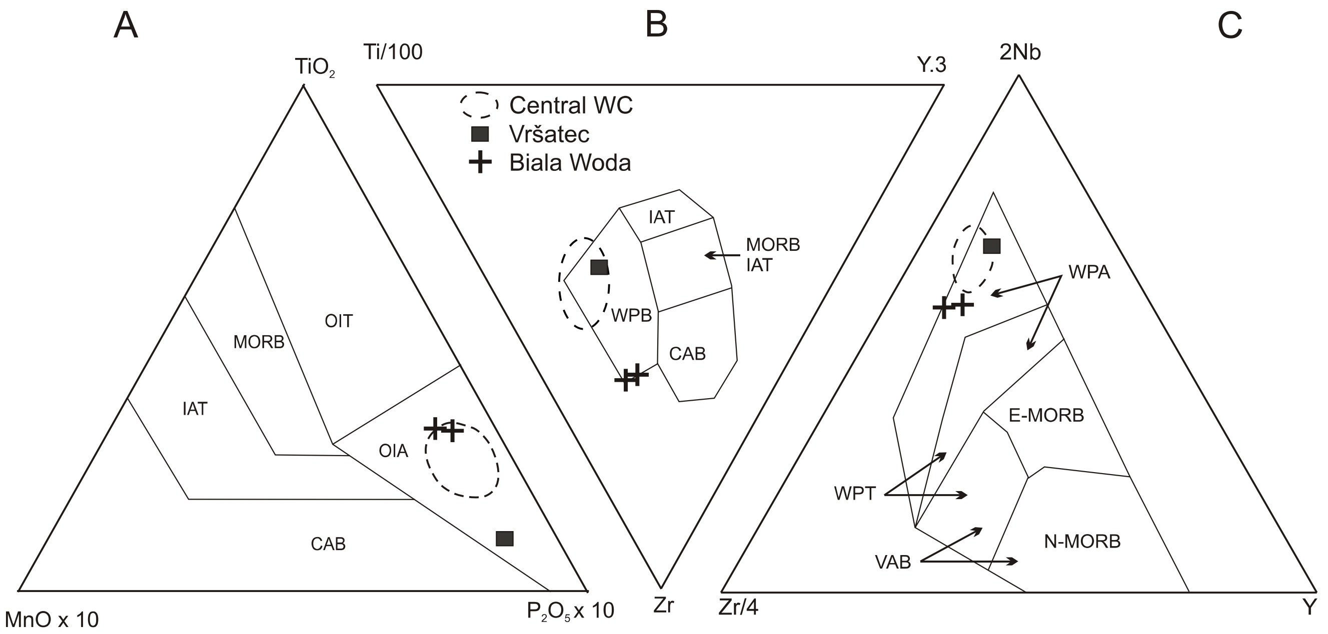

Fig. 10. Discrimination diagrams for basalts: A – MnO × 10–TiO_2–P_2O_5 × 10 (Mullen, 1983); B – Zr–Ti/100–Y.3 (Pearce and Cann, 1973); C – Zr/4–2Nb–Y (according to Meschede, 1986)

Dashed circle – composition of Cretaceous alkali volcanites from the Central Western Carpathians (Hovorka and Spišiak, 1988; Hovorka *et al*., 1999), CAB – calc-alkaline basalt of volcanic arcs, E-MORB – E-type mid-ocean ridge basalt, IAT – island arc tholeiites, MORB – mid-ocean ridge basalt, N-MORB – N-type mid-ocean ridge basalt, OIA – oceanic island alkali basalt, OIT – oceanic island tholeiites, VAB – volcanic arc basalt, WPA – within-plate alkali basalt, WPB – within-plate basalt, WPT – within-plate tholeiites

Fatric Krížna, or the Hronic Choč Nappe (Hovorka and Slavkay, 1966; Spišiak and Hovorka, 2000).

The intra-Carpathian Tisza (Tisia) terrane, which is another Europe-derived fragment, includes inselbergs and substrata of the southern part of the intra-Carpathian Pannonian Basin and the Apuseni Mts. Early Cretaceous volcanism is particularly widespread in the western part of the Tisza Unit, in the present Mescek Mts. and subcrop of the Alföld Basin of Southern Hungary (e.g., Harangi, 1994; Haas and Péró, 2004). This terrain was designated as the Mecsek–Alföld Igneous Field (MAIF) by Harangi *et al*. (2003). Volcanic rocks include alternating pillow lavas and lava breccias, hyaloclastites and volcaniclastics, as well as numerous subvolcanic dykes and sills (Harangi *et al*., 2003). Effusive and extrusive products once possibly formed huge seamounts and were closely related to coeval atoll-like carbonate build-ups dated as Valanginian to Albian (Császár and Turnšek, 1996). K-Ar radiometric data range between 135 and 110 Ma (Harangi and Árva-Sós, 1993).

Upper Cretaceous (Campanian to Maastrichtian) alkaline basalts are rarely reported from the Carpathian area. Mafic to ultramafic dyke swarms (predominantly monchiquite) were described from the Transdanubian Range (e.g., Kubovics *et al*.,

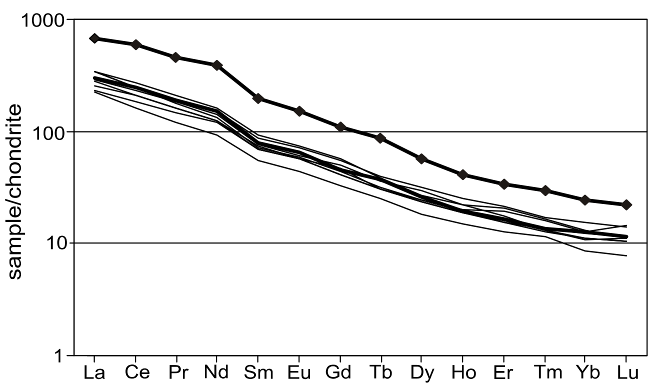

Fig. 11. Chondrite-normalized rare-earth element abundances of the Vršatec melanephelinites (squares), Hanigovce basanites (Pieniny Klippen Belt, Spišiak and Sýkora, 2009 – heavy line) and Cretaceous alkali volcanites of the Central Western Carpathians (Hovorka *et al*., 1999 – thin lines)

Normalizing values after McDonough and Sun (1995)

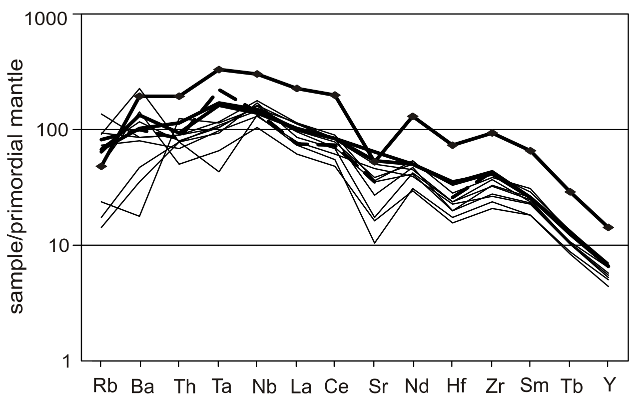

Fig. 12. Mantle-normalized trace element abundances of: squares – Vršatec volcanites, heavy lines – basaoids from Hanigovce (Jarmuta Fm., Pieniny Klippen Belt; Spišiak and Sýkora, 2009), dashed line – basaoids from Biala Woda (Jarmuta Fm., Pieniny Klippen Belt, Birkenmajer and Lorenc, 2008) and thin lines – Cretaceous alkali volcanites of the Central Western Carpathians (Hovorka *et al*., 1999)

Normalizing values after McDonough *et al*. (1992)

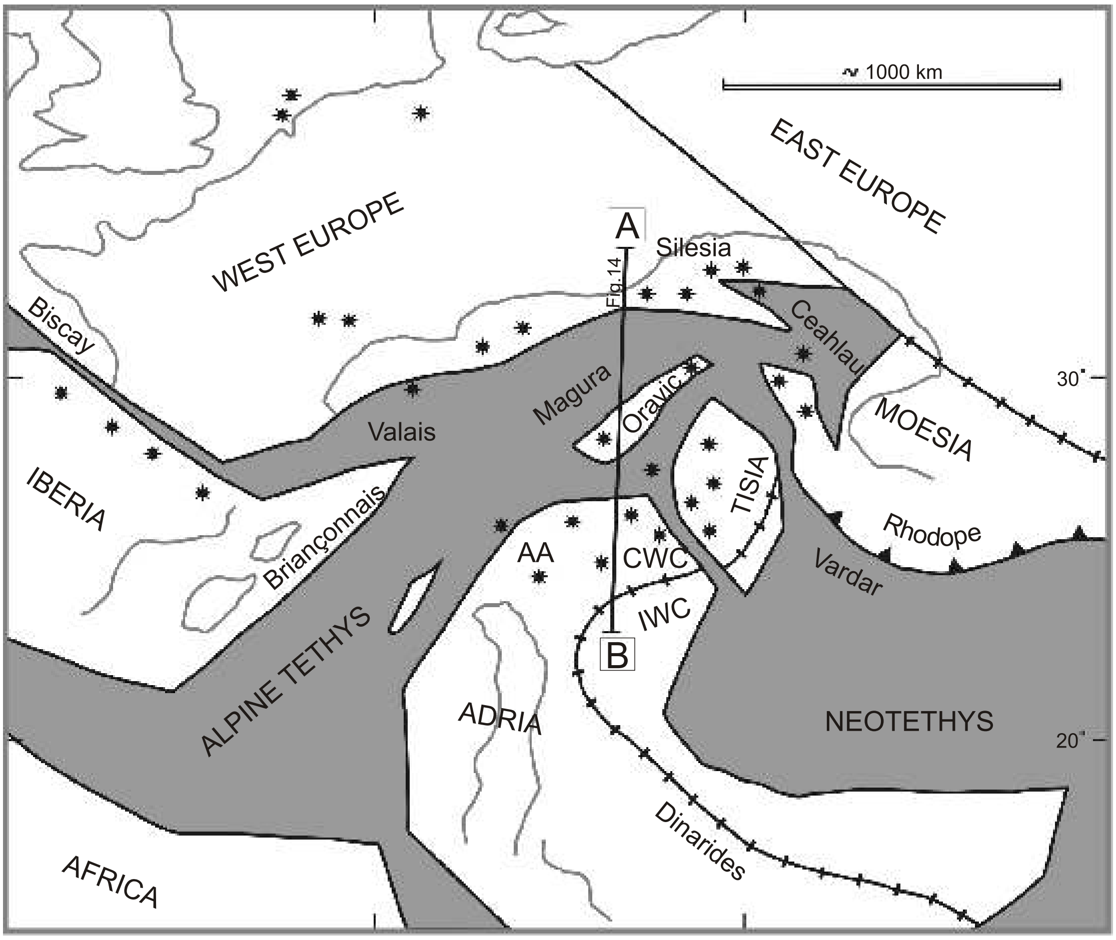

Fig. 13. Palaeogeographical scheme for the western Tethyan region in mid-Cretaceous times

Blank areas – continental crust (AA – Austroalpine, CWC – Central Western Carpathians, IWC – Internal Western Carpathians); grey areas – oceanic crust; crossed line – Meliata and related Jurassic sutures; barbed line – active subduction; asterisks – inferred position of alkaline basalt volcanic centres; compiled based on reconstructions of various authors mainly, Ziegler (1988), Stampfli and Kozur (2006) and Schmid *et al*. (2008)

1990; see also review by Hovorka, 1996 and references therein). Along with the lamprophyre dykes of similar age from the Villány Mts. in Southern Hungary (Tisza megaunit), they bear signs of subduction-related enrichment of the parental asthenospheric magma (Nédli and Tóth, 2007). Contamination of the lithospheric mantle by the subducted material is then recorded also in the Plio-Pleistocene alkali basalts in the intra-Carpathian area (Kovács *et al*., 2004). The increasing amount of crustal material that polluted the mantle sources of alkali magmatism from the Late Cretaceous onwards can be ascribed to the global increase of subducted crust in the Alpine–Carpathia–Mediterranean areas due to ongoing Europe-Africa convergence and subduction of various Tethyan branches (e.g., Piromallo *et al*., 2008).

GEODYNAMIC BACKGROUND OF THE EARLY CRETACEOUS ALKALINE VOLCANISM

The products of Cretaceous alkaline volcanism are known from both the EWC and CWC, i.e. north and south of the PKB. These rocks, in spite of restricted volume and area coverage, are of key importance for the definition of geotectonic conditions at the time of their formation (Hovorka and Spišiak, 1988, 1993; Spišiak and Hovorka, 1997).

Available geochronological and biostratigraphic data indicate that the Cretaceous alkaline volcanism in the Western Carpathian area started during the earliest Cretaceous (at *ca*. 140 Ma) and culminated during the Aptian and early Albian, i.e. roughly from 125 to 100 Ma. This can be postulated despite the isotopic age data usually showing a considerable scatter while even the K-Ar ages from a single volcanic body may differ by 10 Ma (*cf*. Birkenmajer and Pécskay, 2000). Biostatigraphical dating of sediments, in which the submarine lava flows are inserted, is often of low resolution as well. However, the late Early Cretaceous age of their origin fits well to the majority of occurrences of all volcanic fields described above. The alkaline basalts described here from the PKB appear to be a little younger, although probably not younger than 90 Ma (late Turonian).

Sources of the Cretaceous and Cenozoic alkaline basaltic volcanism in Western and Central Europe were looked for in an extensive, sheet-like subcontinental mantle reservoir designated as the Common Mantle Reservoir (CMR; Lustrino and Wilson, 2007). The CMR should have originated during the Cretaceous and resided in the asthenosphere since then, occasionally forming upwellings and feeding volcanic edifices during periods of suitable extensional tectonic regime in the overlying lithosphere (Harangi *et al*., 2003; Piromallo *et al*., 2008). Several conceptual models attempting to explain the origin of Cenozoic alkaline basalts in the circum-Mediterranean area were summarized in a comprehensive review by Lustrino and Wilson (2007). The published opinions range from a deep mantle plume hypothesis, through "thin spot", "mantle fingers", plume channelling, passive rifting and mantle upwelling models, up to mantle lithospheric decompression melting due to lithospheric stretching. These variable views reflect the geochemical signals in alkaline basalts from various parts of Europe that, in spite of sharing many common features, may differ significantly. Hence, they do not provide unequivocal evidence for the general application of any model considered. Nevertheless, the presence of small mantle plumes, diapirs or "fingers" beneath certain volcanic provinces has been independently supported by seismic tomography (Granet *et al*., 1995), but some other models lack such evidence. An extensional tectonic regime is not an inevitable prerequisite either for instance Dézes *et al*. (2004) documented a general compressional tectonic regime in the Alpine and Pyrenean foreland during the Paleocene, when the Cenozoic alkaline magmatic provinces of Central Europe (provinces of Western and Central Europe, Rhenish Massif, Black Forest and Bohemian Massif) were initiated.

The same applies for the Cretaceous alkali basalts all around Europe. As pointed out by Harangi *et al*. (2003), there is no simple model that can explain all aspects of Cretaceous primitive mafic volcanism in Western and Central Europe. Genetic relations to Atlantic plumes are uncertain, although envisaged by e.g. Oyarzun *et al*. (1997) and Wilson (1997). According to these authors, the alkaline magmatism is confined to peripheries of the central and/or north Atlantic plumes and affected the European passive margin.

The period between 120 and 85 Ma was a time of exceptional planetary-scale geodynamic process, recorded by the normal magnetic polarity superchron, a global sea level highstand, increase in the spreading rate in the Pacific Ocean, plate movement reorganization, and so on. These phenomena have been interpreted as the result of an important mantle event – the rise of huge mantle upwellings named superplumes or superswells (e.g., Larson 1991*a*, *b*). Extensive mantle melting produced large igneous provinces as the Ontong-Java or Kerguelen plateaus and the mantle superplume ultimately led to Gondwana splitting and the opening of the south Atlantic and Indian Oceans. However, the superplume model has been criticized (e.g., Loper, 1992; Anderson, 1994), especially concerning the apparent coincidence of the magnetic superchron and the plume-related superficial flood basalts. The superplume obviously would have needed some tens of Ma to rise from the core-mantle boundary to the base of the lithosphere. Therefore Larson and Kincaid (1996) modified the superplume hypothesis and proposed elevation of the 670 km thermal boundary layer as a primary trigger of the onset of mid-Cretaceous volcanism and Pangaea splitting.

However, all models relying on some sort of rising mantle plume as a source of the Cretaceous alkaline basaltic volcanism in Europe do not take into consideration the tectonic style of rifting that governed the opening of several oceanic arms of the Alpine Tethys (Penninics). The geological record suggests a passive, essentially non-volcanic rifting mode during the pre-drift stage, resulting from accumulation of tensional stresses in the European lithosphere located in the foreland of the Alpine–Carpathian orogen prograding from the hinterland Tethyan (Meliata) Suture (e.g., Froitzheim *et al*., 2008; Fig. 13). The first phases of rifting encompassed wide areas, the ensuing breakup occurring along lithospheric detachment faults that caused sublithospheric mantle exhumation. Oceanic crust production commenced only later due to passive mantle upwelling. This simple shear, "Wernicke-type" (Wernicke, 1985; Lister *et al*., 1986) rifting mode has been reconstructed for both the Mid Jurassic opening of the Ligurian–Piemont (South Penninic) Ocean (Lemoine *et al*., 1987; Bernoulli *et al*., 1993; Froitzheim and Manatschal, 1996; Schaltegger *et al*., 2002; Piccardo *et al*., 2002), as well as for the Early Cretaceous opening of the North Penninic Valais Ocean (Florineth and Froitzheim, 1994; Froitzheim and Rubatto, 1998; Manatschal *et al*., 2006) and Magura Oceans (Plašienka, 2003). The Cretaceous continental breakup events affected the southern passive margin of the North European Platform and formed intra-Penninic continental splinters as the Briançonnais in the west and Oravic (Czorsztyn) Ridge in the east (Fig. 13). The Cretaceous North Penninic Rift System (Valais Ocean) partly re-rifted the Jurassic South Penninic Ocean (Liati *et al*., 2005). Even further to the west, the present-day uninverted Galicia passive margin of the Atlantic Ocean shows a similar evolution (e.g., Manatschal and Bernoulli, 1999; Manatschal, 2004; Manatschal *et al*., 2006).

Recently, Manatschal and Münterer (2009) have proposed a new tectonic/magmatic scenario for the evolution of the Alpine Tethys ophiolites. They characterize the magma-poor ocean/continent transition (MP-OCT) as a complex association of subcontinental mantle exhumed along a low-angle detachment, fragments of continental basement, tectono-sedimentary breccias, post-rift sediments and various basic volcanites ranging from MORB-type to alkaline. Evolution of such margins progressed from a non-volcanic asymmetric rifting stage to breakup accompanied and followed by MOR mafic magmatism. Alkaline magmas were emplaced later during a post-breakup stage.

Furthermore, White *et al*. (1987) showed that plume-influenced asthenosphere 100–150°C hotter than normal would generate enormous rift-related volcanism, a feature which is not observed along the Alpine–Carpathian Mesozoic rift zones. On the contrary, geochemical modelling of Cretaceous alkali basalts suggests a low-degree (3–6%), low pressure (60–80 km) partial melting of a garnet peridotite source at the spinel/garnet transitional zone (Harangi, 1994; Harangi *et al*., 2003). An enriched, HIMU/OIB affinity parental magma indicated by trace element distribution and Sr and Nd isotopic ratios probably resulted from mixing of asthenospheric and metasomatized subcontinental mantle material (Trommsdorff *et al*., 1990; Latin and Waters, 1992), or by introduction of recycled crustal components into the depleted upper mantle (Lustrino and Wilson, 2007).

During the Early Cretaceous, the Western Carpathian orogenic foreland was subjected to marine sedimentation organized into to wide basinal and narrow ridge areas, as indicated by the sedimentary record (Michalík *et al*., 1996; Plašienka, 1998). This rugged topography resulted from preceding Jurassic rifting events, which lack any signs of rift-related volcanism. There were several additional rifting events starting from the Berriasian onwards, which ultimately led to the opening of the North Penninic–Magura Ocean (Plašienka, 2003). The main phase of Cretaceous mafic alkaline volcanism partly coincides with development of the Urgonian carbonate platforms (Barremian–Aptian), which grew in places on submarine ridges of the Oravic and Tatric domains (Michalík and Soták, 1990; Michalík, 1994; Józsa and Aubrecht, 2008). Since Urgonian platforms are locally inserted within comparatively deep-water pelagic sequences, a temporary lowering of bathymetry of some hundreds metres must have occurred. Furthermore, contemporaneous emersion and terrigenous clastic input was documented in certain parts of the Tatric and Oravic areas (Jablonský *et al*., 1993; Aubrecht *et al*., 2006). In addition to the eustatic sea level drop, the differential uplift was most probably related to horst formation and extensional block tilting. Accordingly, an extensional tectonic regime governed wide areas that surrounded the Penninic oceanic domains. As an alternative, the Barremian-Aptian surface uplift might be attributed to mafic magma underplating. However, the small amount of effusive volcanic material, low degree of its differentiation and crustal contamination, as well as the primitive geochemical characteristics of the Lower Cretaceous volcanism rather suggest a rapid ascent of tiny portions of mantle melts that pierced an overlying lithosphere attenuated by preceding and coeval rifting events (Fig. 14).

Summing up, the geological record from the Western Carpathians indicates a passive, volcanic-poor rifting mode during the Jurassic and Early Cretaceous. Rifting and continental breakup were consequences of build-up of tensional stresses in the lithosphere located in the foreland of the developing orogenic wedge. Short-term spreading and oceanic crust production resulted from divergent plate movements (Europe *vs*. Adria/Africa) with restricted production of MORB-type basalts in axial zones of comparatively narrow, slow-spreading and/or MP-OCT Penninic rifts of the Alpine Tethys. Contemporaneously, the alkaline basaltic volcanism occurred on passive continental margins of the Penninic rifts, being generated by low-degree melting of lithospheric mantle or mixed sources on the peripheries of mantle upwellings that fed the axial oceanic rift zones. Global mantle processes, such as the Cretaceous superplume and/or the thermal boundary layer elevation, might have contributed to the ambient upper mantle temperature rise and enhanced the upper mantle upwellings ("diapiric instabilities" – Lustrino and Wilson, 2007) below the extending, thermally softened European continental lithosphere. Nevertheless, the volcanic centres do not form clusters indicating hotspots above local mantle plumes, fingers, or diapirs, but spread for thousands of kilometres along margins of the passive Penninic rifts. Alkaline volcanism generally postdated the main rifting phases, hence it appears to be the product, not the cause of rifting. In view of that, the geological data rather support a non-plume origin of the Lower Cretaceous mafic alkaline volcanism in the Alpine–Carpathian–Pannonian realm.

CONCLUSIONS

The newly discovered body of basaltoid within the Pieniny Klippen Belt forms a comparatively large lava body composed mostly of hyaloclastites. Volcanics are intimately related to coeval marly deposits containing Globotruncanid foraminifers, thus their age is Cenomanian or younger. The sedimentary succession in which the volcanics occur belongs to the Oravic Superunit that can be placed palaeogeographically at the transition between the Czorsztyn Ridge and the Kysuca–Pieniny Basin, i.e. within the Penninic oceanic realm (Alpine Tethys).

The mineralogical composition and geochemical features of the Vršatec volcanites studied correspond to melanephelinites. The rare element and trace element patterns are higher than those typical of magmas generated in an intraplate setting and are similar to those of OIB or WPA provinces. High contents of REE as well as those of some other elements have been influenced by high apatite (P_2O_5 3.0 wt.%) contents. These volcanics were probably generated by decompression melting during passive asthenospheric mantle upwelling associated with lithospheric stretching and thinning.

Though the chemical composition of the melanephelinites from Vršatec is different (higher contents of P_2O_5, Rb, Sr, REE) from the other vulcanite occurrences in the Pieniny Klippen Belt (Jarmuta Fm.; Biala Woda – Birkenmajer and Pécskay, 2000; Birkenmajer and Lorenc, 2008; Hanigovce –

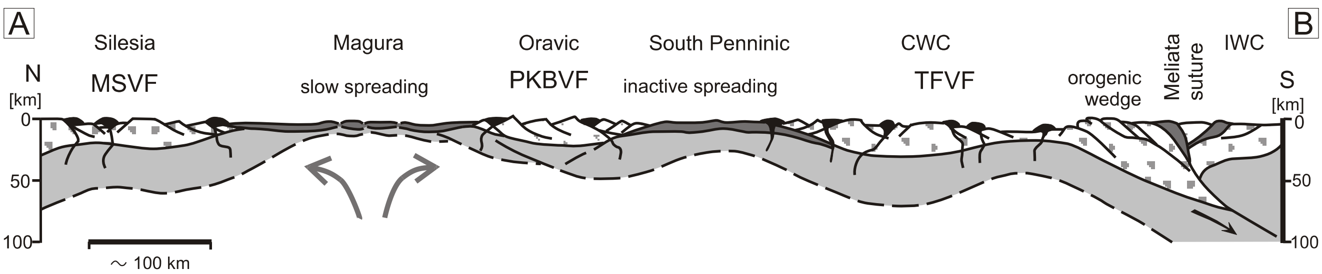

Fig. 14. Lithosphere-scale profile of the future Western Carpathian area showing presumed position and sources of Cretaceous alkaline basalts

MSVF – Moravian–Silesian Volcanic Field, PKBVF – Pieniny Klippen Belt Volcanic Field, TFVF – Tatric–Fatric Volcanic Field, dotted – continental crust, dark grey – oceanic crust, light grey – lithospheric mantle, tailed black buttons – alkaline volcanics (exaggerated); for the position of the section see Figure 13

Spišiak and Sýkora, 2009) the age and geological position is very similar to them and to other Cretaceous alkaline basalts occurring in the External (Silesian Unit, MSVF) and Central Western Carpathians (Tatric nad Fatric units, TFVF), as well as to some volcanic rocks from the Eastern Alps (AAVF), Eastern Carpathians and Pannonian Basin (Tisza terrane – Mecsek Mts., MAIF) and also outside the Alpine–Carpathian–Pannonian realm (e.g., Pyrenean foreland, Iberian/Atlantic passive margin). Considerations of the geodynamic setting and tectonic evolution of all these zones suggest that they developed along passive rift arms of the Alpine Tethys governed by an extensional tectonic regime due to build-up of tensional stresses in the foreland of compressive Alpine orogenic wedges prograding from the hinterland. Rifting started as non-volcanic, and in places was followed by restricted oceanic spreading and MORB type volcanism. Accordingly, the alkaline volcanism was related to passive mantle upwelling resulting from lithospheric stretching. Small portions of mantle-derived melts originated by low-degree partial melting of a HIMU-like protolith on the peripheries of mantle upwellings rising below the rift axes. Their ascent and piercing of the overlying lithosphere was enhanced by the general extensional regime and considerable lithospheric thinning as a result of preceding and coeval rifting events. Therefore, the basalts were emplaced off rift axes, along the passive margins rimming the Penninic basins. Possible contribution of mantle plume material related to a mid-Cretaceous mantle “storm” cannot be ruled out, but appears to be of minor relevance in explaining of the Cretaceous alkaline mafic volcanism in the Alpine–Carpathian area.

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