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1: Subduction, slab detachment and mineralization: The Neogene in the Apuseni Mountains and Carpathians

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Abstract

The Inner Carpathians comprise several distinct Neogene late-stage orogenic Pb–Zn–Cu–Ag–Au ore districts. The mineral deposits in these districts are closely related to volcanic and subvolcanic rocks, and represent mainly porphyry and epithermal vein deposits, which formed within short periods of time in each district. Here, we discuss possible geodynamic and structural controls that suggest why some of the Neogene volcanic districts within the Carpathians comprise abundant mineralization, while others are barren. The Neogene period has been characterized by an overall geodynamic regime of subduction, where primary roll-back of the subducted slab and secondary phenomena, like slab break-off and the development of slab windows, could have contributed to the evolution, location and type of volcanic activity. Structural features developing in the overlying lithosphere and visible in the Carpathian crust, such as transtensional wrench corridors, block rotation and relay structures due to extrusion tectonics, have probably acted in focusing hydrothermal activity. As a result of particular events in the geodynamic evolution and the development of specific structural features, mineralization formed during fluid channelling within transtensional wrench settings and during periods of extension related to block rotation.

In the Slovakian ore district of the Western Carpathians, Neogene volcanism and associated mineralization were localized by sinistral, NE-trending wrench corridors, which formed part of the extruding Alcapa block. The Baia Mare ore district, in the Eastern Carpathians, reflects a transtensional wrench setting on distributed oversteps close to the termination of the Dragos Voda fault. There, mineralization was spatially controlled by the transtensional Dragos Voda master fault and associated cross-fault systems. The Golden Quadrangle Cu–Au ore district of the Southern Apuseni Mountains reflects an unusual rotated transtensional/extensional setting close to the termination of a graben system. There, fluid flow was probably localized by fault propagation at the inner tip of the graben system.

The spatial and temporal evolution of the magmatism and its changing geochemical signature from (N)W to (S)E strongly suggests a link with the contemporaneous northeastward roll-back of the subducted slab and a progressive southeastward detachment during accelerating roll-back. This geodynamic evolution is further supported by the present-day overall and

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detailed mantle lithospheric density images, the present-day heat flow patterns, the crustal architecture and its interpreted evolution, and the spatial and temporal evolution of depocentres around the Carpathian arc. In contrast to all these features, the mineral deposits in the West Carpathians, East Carpathians and Apuseni Mountains are too synchronous with respect to their individual volcanic history and contrast too much with younger volcanics of similar style, but barren, in southeastern parts of the Carpathians to simply link them directly to the slab evolution. In all three districts, the presence of magmatic fluids released from shallow plutons and their mixing with meteoric water were critical for mineralization, requiring transtensional or extensional local regimes at the time of mineralization, possibly following initial compressional regimes.

These three systems show that mineralization was probably controlled by the superposition of favourable mantle lithospheric conditions and partly independent, evolving upper crustal deformation conditions.

In the 13 to 11 Ma period the dominant mineralization formed all across the Carpathians, and was superimposed on structurally favourable crustal areas with, at that time, volcanic–hydrothermal activity. The period may reflect the moment when the (upper part of the) crust failed under lithospheric extension imposed by the slab evolution. This crustal failure would have fragmented the overriding plate, possibly breaking up the thermal lid, to provoke intensive fluid flow in specific areas, and allowed subsequent accelerated tectonic development, block rotation and extrusion of a "family of sub-blocks" that are arbitrarily regarded as the Tisia–Dacia or Alcapa blocks, even though they have lost their internal entity.

Keywords: Porphyry system; Epithermal system; Wrench corridor; Subduction; Slab break-off; Overstep; Relay structure; Hydrothermal system

1. Introduction

The formation of large-scale ore provinces is governed by many factors, amongst which the geodynamic control exerts the primary pre-condition. This is especially true for ore provinces in orogenic belts, where many different types of mineralization of various ages are found. Mineralization within collisional orogens involves a number of different processes. These mainly include: (1) Andean-type mineralization due to subduction of oceanic lithosphere pre-dating collision (e.g., Sillitoe, 1997), further enhanced by progressive changes in the angle of subduction (e.g., Mahlburg-Kay and Mpodozis, 2001), (2) mineralization associated with collisional granites (e.g., Lang and Baker, 2002; Groves et al., 2003 and references therein), (3) mineralization resulting from hydrothermal activity associated with orogen-scale faulting due to adjustment of the upper, brittle crust, sometimes caused by a lithospheric response to gravitional instability of the subducted slab after waning convergence, such as slab roll-back and slab break-off (e.g., de Boorder et al., 1998; Blundell, 2002; Drew, 2003; Lips, 2002). These processes are often associated with calc-alkaline and subsequent alkaline magmatism, including early stage porphyry and subsequent hydrothermal vein mineralization (e.g., de Boorder et al., 1998; Lehmann et al., 2000; Blundell, 2002; Richards, 2003). For all these different modes of mineralization, key questions include: (1) the general geodynamic context within which ore formation occurs; (2) the exact structural setting, which allows the formation of a plumbing system linking deep structural levels with near-surface horizons where appropriate precipitation conditions prevail; (3) the origin of metals and hydrothermal fluids, including the fluid composition; (4) the nature of the heat source driving magmatic and hydrothermal processes, and (5) the composition of magmatic melts (e.g., Hedenquist and Lowenstern, 1994; Heinrich et al., 1999; Ulrich et al., 1999; Richards, 2003). Furthermore, the possible origin of hydrothermal fluids and timing of mineralization might represent extremely important controlling factors, as the channellizing plumbing system should be available when metal-rich fluids start to rise. The origin of the fluids may lead to a distinction between highly mineralized and small, relatively barren hydrothermal systems (Heinrich et al., 1999; Oyarzun et al., 2001).

Despite these general pre-conditions, and even though the formation of hydrothermal and orogenic ore deposits includes many processes, only a few models exist which link these to specific large-scale geodynamic process such as slab break-off (e.g., de Boorder et al., 1998) or to a change of the subduction angle (e.g., Mahlburg-Kay and Mpodozis, 2001; Oyarzun et al., 2001). For example, many hydrothermal lodes and porphyry copper deposits are related to subduction of oceanic crust (Sillitoe, 1997), or further spatially controlled by the presence of slab windows within subducting slabs, as in the western USA (e.g., Haeussler et al., 1995) and the Andes (Dill, 1998). Recent observations show that many ore deposits in orogens formed rather by punctuated events and not

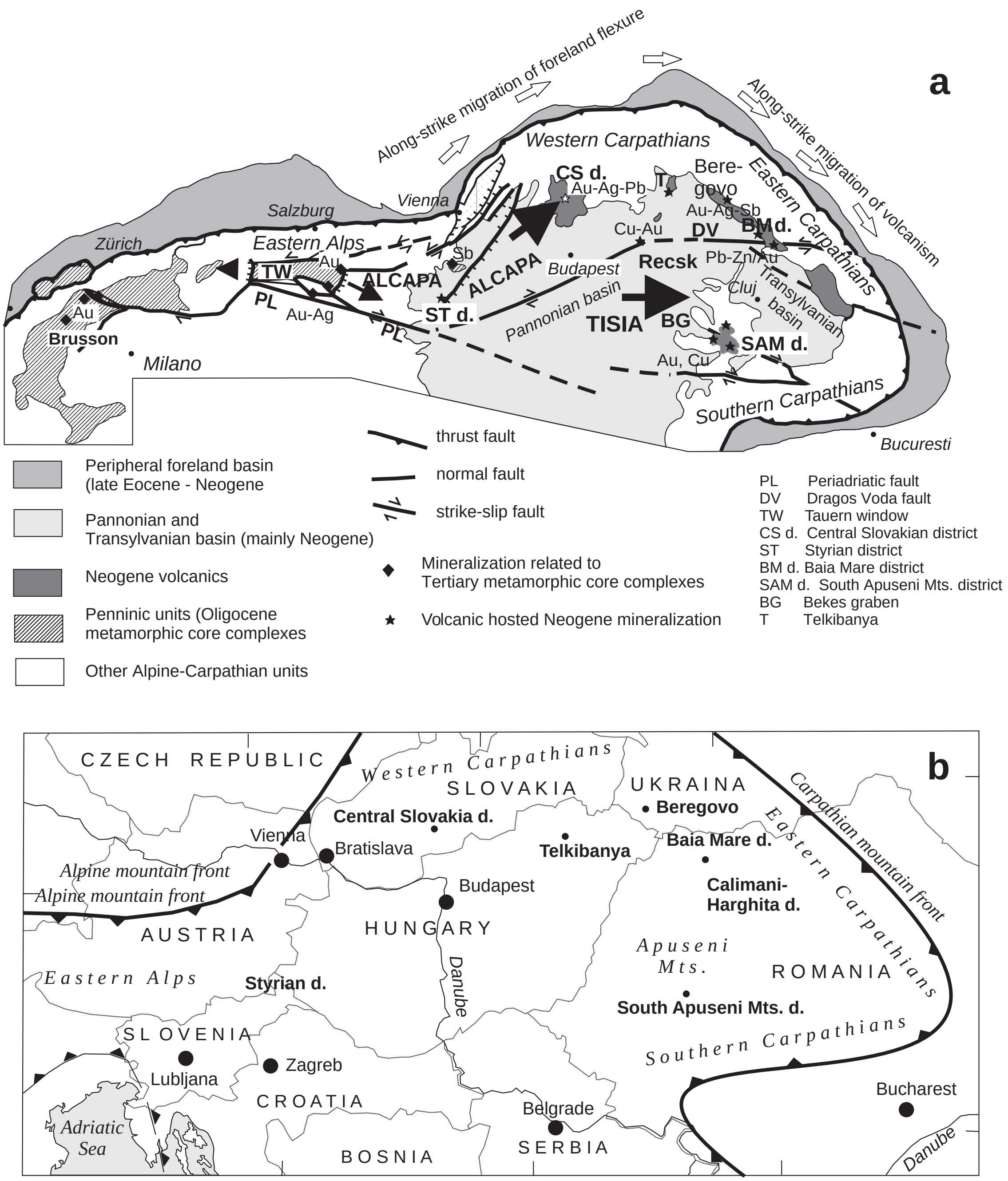

Fig. 1. a—Tectonic overview map of the Pannonian–Carpathian orogenic system displaying Neogene mineralization, magmatism and Oligocene to Neogene structures (modified after Neubauer, 2002). b—Mineralized districts (d) located within national boundaries.

by continuous processes (Mahlburg-Kay and Mpodozis, 2001; Oyarzun et al., 2001; Hall, 2002; Neubauer, 2002). The generation of distinct, adakitic melts clearly seems to play a role in the formation of giant deposits in a subduction zone environment (Oyarzun et al., 2001) and recent interpretations argue for their origin along margins of slab windows (Thorkelson and Breitsprecher, 2005). Several models have been proposed that relate ore deposits to collisional and post-collisional processes within convergent continental plates (e.g., Spencer and Welty, 1986; Beaudoin et al., 1991), including some that are possibly triggered by slab break-off (de Boorder et al., 1998). The role of deposit- and district-scale structures in the formation of hydrothermal ore deposits has not been investigated in great detail, although they act as a prime, controlling factor (e.g., Sibson, 2001, 2003; Drew, 2003; Tosdal and Richards, 2001; Richards, 2003).

Mineralization is widespread in the Alpine–Balkan–Carpathian–Dinaride (ABCD) orogen and occurs dominantly near the areas of late-orogenic extension, like the Inner Carpathians surrounding the Pannonian Basin, and the Rhodope Zone and West Anatolian Volcanic Province bordering the Aegean Sea. The Neogene mineralization of the Inner Carpathians is a particularly interesting young belt (Fig. 1a,b), as late-stage orogenic mineralization is abundant (e.g., Evans, 1975; Petrascheck, 1986; Mitchell, 1996; Borcoş et al., 1998) and varies in style along strike (e.g., Heinrich and Neubauer, 2002; Neubauer, 2002 and references therein). A deep-seated heat source, visible in mantle tomography, has been postulated recently as the cause of the magmatism and associated mineralization (de Boorder et al., 1998). Here, we discuss a number of Neogene mineralized districts in the Inner Carpathians and their geodynamic context and structural setting, and compare these with coeval, barren volcanic systems at the southeastern edge of the Alps. Data from the ABCD orogen may help to constrain models for late-stage orogenic mineralization, particularly in their geodynamic and structural control.

The ABCD orogen has a geological history characterized by complex interactions between various microplates squeezed in between the African and Eurasian continents, a situation which is basically similar to the present-day interactions between Australia and SE Asia in the region of the Indonesian island arc system (Hall, 1996; Blundell, 2002, Kázmér et al., 2003). For the ABCD orogen as a whole, a number of attempts have already been made to relate large-scale tectonic processes with mineralization. These include, among others, the work of Petrascheck (1986), Serafimovsky et al. (1995), Mitchell (1996), Janković (1997), de Boorder et al. (1998), Vlad and Borcoş (1998), Lexa (1999a), Lips (2002), Neubauer (2002) and Drew (2003). However, the rapid development of tectonic models for the Neogene evolution

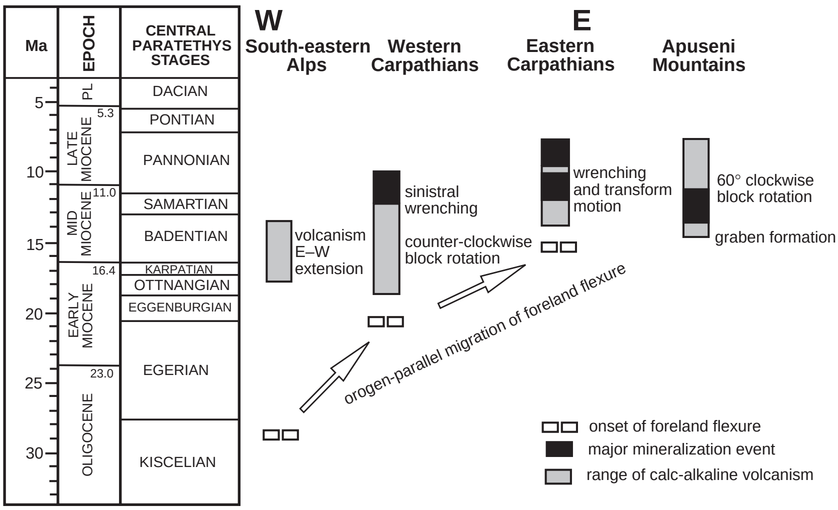

Fig. 2. Paratethys time-scale (after Rögl, 1996; Sacchi and Horváth, 2002) and major steps in the tectonic evolution of the Carpathian–Pannonian region, including mineralization in the major ore districts. PL—Pliocene.

(e.g., Fodor et al., 1999; Wortel and Spakman, 2000; Horváth et al., in press) and the availability of extensive modern data sets on mineralization make the region a natural laboratory for testing models of late-stage, orogenic mineralization.

The time-scale for the evolution of Paratethys (Fig. 2) follows calibrations proposed by Rögl (1996) and Sacchi and Horváth (2002), as the Carpathians were part of the late Eocene to Pliocene Paratethys. Because of faunistic endemism, the Paratethys time-scale is different from that of the Mediterranean area, and precise correlation is needed to relate tectonic and mineralizing processes.

In what follows we use the term strike-slip fault to characterize the master fault itself, and the term wrench corridor to typify a wide zone along both sides of a master strike-slip fault, which is affected by all kinds of secondary structures within deformable cover rock successions (e.g., Mandl, 1988).

2. Overview of major ore districts

The Inner Carpathian arc comprises a number of late-stage orogenic ore districts, which formed during Neogene collisional processes. These major districts (Lexa, 1999a; Heinrich and Neubauer, 2002; Cassard et al., 2004) include:

(1) The South Apuseni Mountains ore district ("Golden Quadrangle" or "Golden Quadrilateral") exposed in southern sectors of the Apuseni Mountains is related to the emplacement of Neogene volcanic rocks mainly between 14.9 and 9 Ma (Pécskay et al., 1995b; Roşu et al., 1997, 2001a,c, 2004). The ore deposits are of low-sulphidation and subordinate high-sulphidation epithermal type and of related porphyry Cu-type (Alderton et al., 1998; Alderton and Fallick, 2000; Milu et al., 2003; Ciobanu et al., 2004). Ore mineralogical, isotopic and fluid inclusion data constrain their formation as epithermal mineralization. Major ore deposits occur along an ESE-trending, dextral, strike-slip fault, which is considered to represent a secondary fault zone that formed within the eastward moving Tisia–Dacia block during the Neogene (Linzer et al., 1997). Drew and Berger (2001) and Drew (2003) showed that magmatism and mineralization are related to a major dextral transfer zone between sub-basins within the Neogene Pannonian–Transylvanian basin system. The ores are unusually rich in gold (and tellurium), and have been mined since pre-Roman times.

(2) In the Eastern Carpathians, the Baia Mare ore district is related to Neogene volcanics and regional faults. The Baia Mare ore district is associated with the Dragos Voda fault, which represents the northern, sinistral, confining strike-slip fault relating to a wrench corridor of the eastward extruding Tisia–Dacia and Alcapa blocks (Csontos, 1995). Bailly et al. (1998) and Grancea et al. (2002) infer a pluton at depth along the fault. The principal ore types are epithermal Pb–Zn(–Cu–Au–Ag) vein deposits. K–Ar and $^{40}Ar/^{39}Ar$ dating of igneous and alteration minerals provide a Neogene age for the magmatism (14 to 9 Ma) and a younger age (11.5 to 8 Ma) for hydrothermal activity, with dominant ore emplacement at around 11 to 8 Ma (Lang et al., 1994; Pécskay et al., 1995a,b; Kovacs et al., 2001).

(3) In the Western Carpathians, several types of Neogene ore deposits are widespread and occur intimately associated with subduction-related volcanism, particularly in the Central Slovakian volcanic field and at Telkibanya (Dill, 1998; Gatter et al., 1999; Lexa, 1999a; Lexa et al., 1999a; Molnár et al., 1999; Koděra et al., 2004, 2005). Initial andesitic volcanism is related to subduction of an ocean (outer flysch basin) during the early Miocene, and final calc-alkaline volcanism is associated with back-arc extension and possibly with slab detachment during accelerated roll-back. Nemčok et al. (2000) and Koděra et al. (2005) reported a regional control on ore veins. Major ore veins formed parallel to and contemporaneous with the movement direction during northeastward extrusion of the Alcapa block and to the WNW–ESE extension. Ore deposits are variable and include porphyry copper along with high-sulphidation epithermal gold, intrusion-related base metal and gold mineralization, as well as epithermal base metal and Au–Ag(–Sb) veins

(e.g., Gatter et al., 1999; Lexa, 1999a,b; Koděra et al., 2004, 2005).

(4) Further mineral deposits are well known in the region, specifically in the northern sectors of the Eastern Carpathians and easternmost Western Carpathians; for example, the Tokaj Mountains with the exhausted Telkibanya Au–Ag epithermal vein deposit (Molnár et al., 1999; Fig. 1). In the Eastern Carpathians, however, mineralization is limited to just a few major ore deposits, e.g., the Au(–Pb–Zn–Ag) and Pb–Zn(–Ag–Au) deposits of the Beregovo district (Vityk et al., 1994). Furthermore, it has to be noted that some large volcanic areas are nearly barren, as for example the Calimani–Gurghiu–Harghita Mts. volcanic district to the SE of the Baia Mare district in the Eastern Carpathians and the Miocene Styrian district (Fig. 1). In the Styrian district (Ebner and Sachsenhofer, 1995), extensive hydrothermal activity was associated with emplacement of volcanic rocks, forming vast areas of epithermal type alunite–illite–christobalite alteration (Weber et al., 1997). No explanation has been proposed as to why these areas are barren. In this paper, we offer a possible reason for the missing mineralization.

In the eastern sectors of the Eastern Alps, Neogene hydrothermal activity resulted in the formation of mesothermal Au–Ag(–Pb–Zn) and stibnite vein deposits (Fig. 1), both associated with low-angle normal and strike-slip fault systems around metamorphic core complexes (Weber et al., 1997; Neubauer, 2002). These deposits formed contemporaneously with volcanic-hosted ore deposits in the Inner Carpathians. However, the origin of these mineral deposits is not discussed further here.

3. Geodynamic setting

3.1. Overview

All these Neogene ore districts are part of the Inner Carpathians of the Alpine–Carpathian orocline, which was formed during Tertiary subduction/collision processes (e.g., Csontos, 1995). The Inner Carpathians and Eastern Alps expose highly deformed, pre-Alpine basement rocks and late Carboniferous to Palaeogene cover in the Western, Eastern, and Southern Carpathians and in the Apuseni Mountains, which are overlain by the Neogene infill of the Pannonian and Transylvanian basins (Horváth, 1993; Horváth et al., in press; Figs. 1 and 2). The External Carpathians comprise the infill of a Cretaceous to Palaeogene accretionary system and a Miocene–Pliocene molasse-type foreland basin (Kázmér et al., 2003; Maţenco et al., 2003). The Carpathian arc is interpreted to represent a pre-Tertiary feature within the southern continental European lithosphere, which was filled by oceanic lithosphere (Csontos, 1995; Fodor et al., 1999; Csontos and Vörös, 2004; Fig. 3a). This oceanic basin is considered to represent a lateral extension of the Penninic oceanic domain (e.g., Csontos and Vörös, 2004 and references therein). The Moesian platform is a stable promontory of the European continental plate, which has been marginally overridden by the South Carpathian/Balkan orogen and which has been approximately in its present position since Palaeogene times (Fig. 3a). To the south, remnants of the Neotethys Ocean were subducting beneath the southern sectors of the Dinarides–Hellenides–Balkan microplate, which also includes the Tisia–Dacia block (Fig. 3a).

During the Palaeogene, the Neotethys oceanic lithosphere continued to subduct northwards underneath the aforementioned continental microplate (e.g., Barr et al., 1999; Neugebauer et al., 2001). The Penninic Ocean had been consumed by slab roll-back during the latest Cretaceous to Palaeogene to form a remnant land-locked oceanic basin in the Carpathian arc (e.g., Wortel and Spakman, 2000). The previous southern European continental margin had been subducted beneath the Alps and was subsequently exhumed as Penninic continental units (Fig. 3a). These are exposed within the Tauern window (Fig. 1a) and were overridden by the Austroalpine nappe complex during the Eocene (Liu et al., 2001 and references therein). Oblique plate collision and associated stacking of lower plate continental units were followed by partitioning of convergence into northward thrusting along the northern leading edge of the orogen and emplacement of the entire Alpine nappe edifice onto European foreland units, and orogen-parallel strike-slip motions along wrench corridors due to a generally NNE–SSW shortening (e.g.,

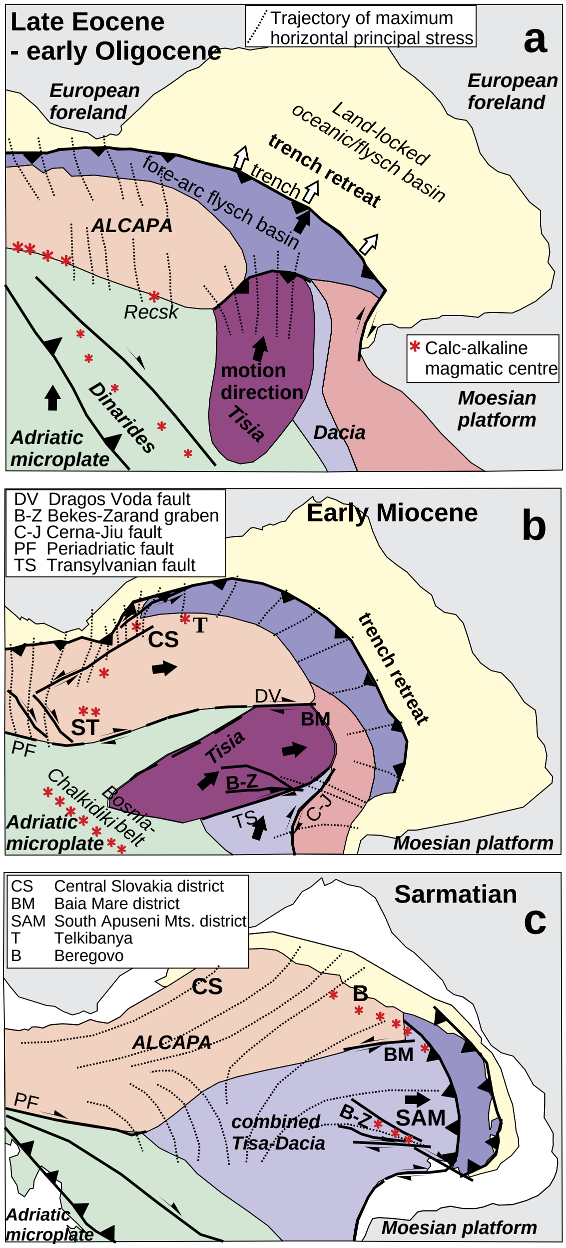

Fig. 3. Time-resolved model for Neogene evolution reflecting current understanding of orogenesis in the Carpathians (from Fodor et al., 1999, modified and updated). a—Late Eocene to early Oligocene; b—Early Miocene (~20 Ma); c—Late middle Miocene (~12 Ma). Tectonic blocks are the same as shown in Fig. 1a.

Ratschbacher et al., 1991; Sperner et al., 2002). The latter stage was also governed by indentation of the rigid Southalpine block, which formed the northern extent of the Adriatic microplate (TRANSALP Working Group, 2002). Furthermore, emplacement of post-collisional late Oligocene to early Miocene intermediate and mafic, calc-alkaline plutons and dykes along southern sectors of the Eastern Alps has been interpreted to have been the result of slab break-off of subducted continental lithosphere following continent–continent collision (von Blanckenburg and Davies, 1995; von Blanckenburg et al., 1998; Pamić et al., 2002). Except for Recsk (Fig. 1) in the easternmost extension of Periadriatic plutons (Fig. 3a) and of the Bosnia–Chalkidiki (or Serbomacedonian) belt, these magmatic rocks are largely barren of mineralization. Recsk represents a porphyry Cu prospect in Hungary (Fig. 1). The other Periadriatic plutons are eroded to deep structural levels, so that high-level plutonic apices, which in a generalized model host mineralization, are missing.

The pre-Tertiary rocks of the Internal Carpathians are part of several fault-bounded tectonic units. Amongst these, the Alcapa block, with surface exposure in the Western and Eastern Carpathians N of the Dragos Voda fault and the Tisia–Dacia block, comprise the two most important units (Fig. 1). The Tisia block is exposed mainly in the Apuseni Mountains and the Dacia block in the Southern and Eastern Carpathians S of the Dragos Voda fault. The South Transylvanian fault is considered to separate the Tisia and Dacia blocks, although the principal fault system active during Oligocene to Miocene invasion of the Tisia–Dacia blocks was located within the southern Carpathians (Ratschbacher et al., 1993).

During late Oligocene and late Miocene to Pliocene, the Adriatic microplate rotated 30° in an anticlockwise manner (Neugebauer et al., 2001; Márton et al., 2003) and collided with the intra-Alpine microcontinent collage in the NW, whilst subduction continued in the SE (Fig. 3b). The indentation of the Adriatic microplate resulted in oroclinal bending of the Western Alps. The combined Periadriatic/Bosnia–Chalkidiki (Serbomacedonian) magmatic belt mimics the arcuate belt and comprises earlier, late Eocene to Oligocene granitoids and volcanics that are mostly exposed in the Alps and central Hungary, and mostly late Oligocene to Miocene andesitic volcanic rocks in the Dinarides and Hellenides (Figs. 1 and 3a; von Blanckenburg and Davies, 1995; Pamić et al., 2002).

The Carpathian arc is interpreted as having been filled by Oligocene to Neogene invasion of the Alcapa and Tisia–Dacia blocks into that realm. The Tisia–Dacia block moved around the Moesian promontory (e.g., Ratschbacher et al., 1993) and rotated up to ca. 90° in a clockwise manner as palaeomagnetic data indicate (Pătraşcu et al., 1994; Panaiotu, 1998). The Alcapa block escaped from the Eastern Alps and rotated ca. 60° in an anticlockwise mode (e.g., Márton, 1997). Invasion was mainly driven by northeastward slab roll-back and slab retreat of the oceanic lithosphere attached to the European continental plate. In this model, the Moesian promontory is fixed and the Alcapa block was also driven by the northward advancing Apulian block, which led the Alcapa block to escape from the Alpine collision zone. This resulted in late Eocene collision in the Western Alps and subsequent Oligocene to late Miocene migration of molasse depocentres along the Alpine–Carpathian arc (Meulenkamp et al., 1996). In the late Miocene, final collision occurred in the bend zone between the Southern and Eastern Carpathians (Maţenco et al., 2003; Bertotti et al., 2003; Tarapoanca et al., 2004). Note the delay between flexure of the foreland lithosphere as a monitor of collision and calc-alkaline magmatism, which is younger than the flexure of the continental foreland (Fig. 2). Both migrated from W to SE around the Carpathian arc, but onset of collision clearly pre-dates formation of a linear, orogen-parallel calc-alkaline volcanic belt by ca. 2 to 8 million years (Fig. 2).

A combination of slab roll-back of the remnant intra-Carpathian ocean and eastward extrusion of the Alps led to the closure of a remnant oceanic basin in the Carpathian arc (e.g., Burchfiel, 1980; Ratschbacher et al., 1991, 1993; Csontos et al., 1992; Royden, 1993; Csontos, 1995; Linzer, 1996; Linzer et al., 1997; Nemčok et al., 1998; Fig. 3b,c). Collision with the European foreland migrated from the Alps eastwards around the Carpathian arc (e.g., Linzer, 1996). The extruding blocks were affected by variable

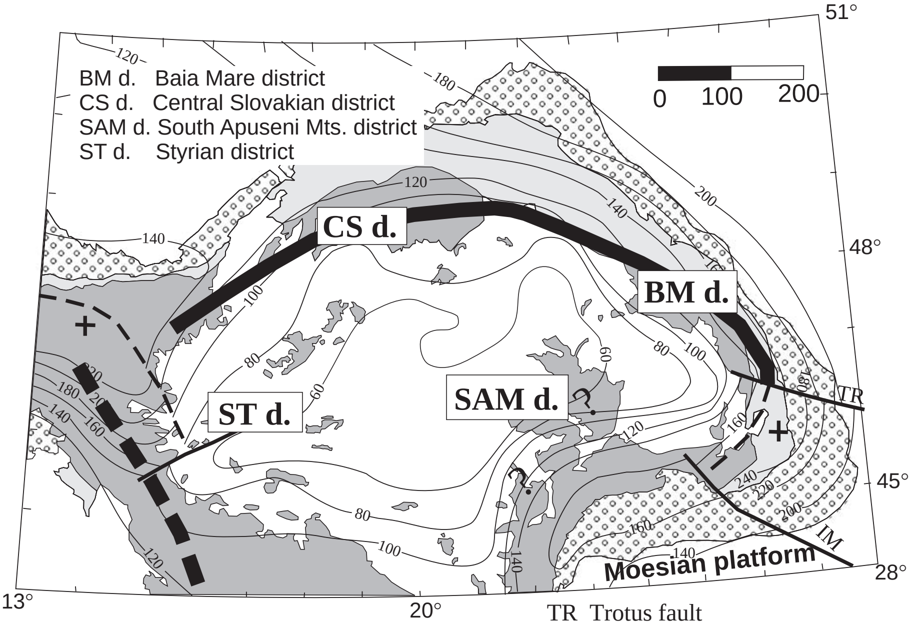

Fig. 4. Map of the base of the lithosphere (in km) of the whole Pannonian–Carpathian realm (compiled from Nemčok et al., 1998; Konečný et al., 2002; Horváth et al., 2005). Thick full and broken lines represent areas where slab break-off occurred (after Wortel and Spakman, 2000; Schmid et al., 2003; Horváth et al., 2005). Question mark represents questionable slab break-off beneath South Apuseni Mountains. Crosses represent areas of uplift.

palaeostress fields within which the maximum principal stress was mostly subhorizontal and largely subparallel to the eastward and northeastward movement direction of the extruding blocks (e.g., Peresson and Decker, 1997; Nemčok et al., 1998; Fodor et al., 1999). Note that the Alcapa block rotated ca. 50 to 80° anticlockwise and the Tisia–Dacia block ca. 90° clockwise (e.g., Pătraşcu et al., 1994; Márton, 1997; Panaiotu, 1998; Márton et al., 2003; Roşu et al., 1997, 2004).

Starting in the middle to late Miocene (~18 to 10 Ma), back-arc extension affected the whole Pannonian Basin region (Nemčok et al., 1998; Fig. 3b,c). In the Pannonian Basin, back-arc extension is associated with Pliocene to Quaternary alkaline magmatism and uplift of the asthenosphere to levels as shallow as 50 km (Horváth, 1993; Horváth et al., in press and references therein).

3.2. Mantle structure

Initial subduction and subsequent slab break-off and slab delamination have all been considered as the main mechanisms for triggering Neogene magmatism in the Western and Eastern Carpathians as well as in the Apuseni Mountains (Wortel and Spakman, 2000; Seghedi et al., 2004). Corresponding structures have been imaged by mantle tomographic methods (e.g., Fan et al., 1998; de Boorder et al., 1998; Cloetingh et al., 2003), summarized by Wortel and Spakman (2000) and Horváth et al. (in press). A simplified map of the base of the lithosphere is shown in Fig. 4 (after Konečný et al., 2002; Horváth et al., in press). These investigations revealed a wide, high-velocity body beneath the Pannonian Basin at a depth of 410 to 660 km, just above the upper to lower mantle transition (Fig. 5). This body has been interpreted to

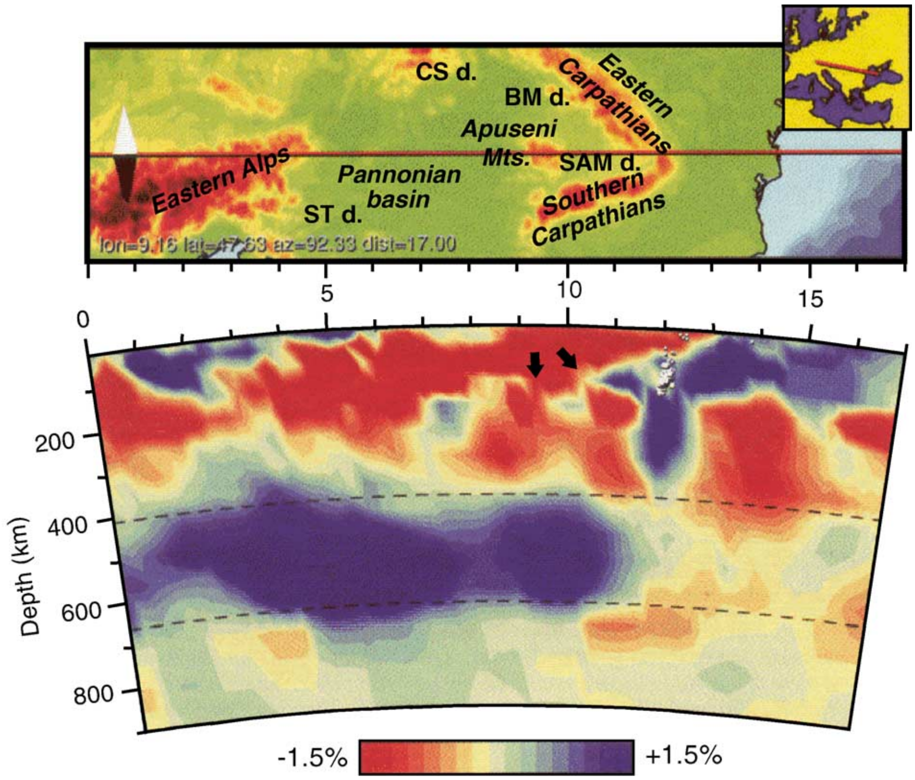

Fig. 5. Tomographic cross-section through the Eastern Alps, Apuseni Mountains and Eastern Carpathians. Blue colour denotes high-velocity, cool bodies interpreted to represent subducted lithosphere, red colour represents low-velocity, hot bodies. Note the low-velocity channels (red arrows) beneath the South Apuseni Mountains: these are interpreted to represent slab windows. Note also the still-attached subducted European lithosphere at the bend zone between the Eastern and Southern Carpathians. Figure courtesy Rinus Wortel and Wim Spakman, Utrecht.

represent subducted lithosphere that accumulated through subduction and roll-back of the remnant Carpathian Ocean. Along the margins, a low-velocity break in the lithosphere has been detected, stretching from the Western Carpathians to close to the bend zone (Bijward and Spakman, 2000). There, a heavy body is still attached to the European lithosphere (e.g., Cloetingh et al., 2003), which seems to continue beneath the Southern Carpathians (Fan et al., 1998). A similar, local break in the lithosphere has also been detected beneath the Apuseni Mountains (Figs. 4 and 5; de Boorder et al., 1998; Cloetingh et al., 2003), giving rise to the speculation that Neogene magmatism there was triggered by slab break-off.

Low-velocity bodies have been detected to depths of 70 km at the front of the Eastern Carpathians (Fan et al., 1998). These are interpreted to represent subducted marine sedimentary rocks and could represent a source of fluids rising together with magmas. No slab break has been detected beneath early Miocene volcanic rocks exposed in the eastern part of the Eastern Alps (Fig. 5), although a small N-dipping slab window has been postulated recently in the area beneath the easternmost Eastern Alps (E of the Tauern window; Lippitsch et al., 2003; Schmid et al., 2003). This slab window probably stretches to the Dinarides, where the window has already been detected (e.g., de Boorder et al., 1998). Lower Miocene volcanic rocks seemingly show no connection with that slab window (Fig. 5). In summary, these observations give rise to the notion that the prime controlling factor of Neogene to Quaternary calc-alkaline magmatism and associated late Miocene mineralization is a mantle heat source due to slab break-off, although a temporal overlap with the waning stages of subduction cannot be excluded (see Mason et al., 1996, 1998; Seghedi et al., 2004 for discussion).

3.3. Neogene crustal structure

Here we only report first-order crustal structures in the larger surroundings of the mineralized districts under discussion (Figs. 1a, 3 and 6). Internal sectors of the Western Carpathians are largely hidden by voluminous Neogene volcanic edifices (Konečný et al., 2002), so the detailed structural setting is difficult to reveal. However, the whole internal Western Carpathians represent a sinistral wrench corridor, which is segmented by extensional and pull-apart sedimentary basins. The subducted European crust is well known beneath the internal Western Carpathians (Tomek, 1993; Tomek and Hall, 1993). Collision with the European lithosphere was soft, as subordinate thick-skinned tectonics and other geophysical properties indicate.

The interior of the Carpathian arc is covered by the Neogene Pannonian Basin (Fig. 6). The Pannonian Basin formed by early to middle Miocene syn-rift extension triggered by eastward extrusion. This stage was followed by late Miocene to Pliocene post-rift subsidence. Marginal sectors of the Pannonian Basin comprise a number of synrift basins that formed due to late, early to middle Miocene rifting, largely contemporaneous with widespread calc-alkaline volcanism (Horváth, 1993; Fodor et al., 1999 and references). Some of these basins are associated with partly buried metamorphic core complexes (Csontos and Nagymaroshi, 1998). The rifting by half-graben-type extension accommodated high heat flow. The structural trend of these rifts varies from N–S in the west to NW–SE in the eastern Pannonian Basin, mainly due to later rotation. Many of these structures are cut by the NE-trending Mid-Hungarian fault zone as well as by its eastward extension in the Dragos Voda fault. This fault system represents the major confining structure between the eastward-moving Alcapa and Tisia–Dacia blocks (Fig. 1).

The Tisia and Dacia blocks are actually divided by highly curved, dextral faults, which formed during late Oligocene to Miocene motion of these blocks around the Moesian promontory (Fig. 1). The Jiu–Cerna fault formed within the interior of the Southern Carpathians. Together with other curved faults, the South Transylvanian fault is probably a second, curved fault system. Its E-trending sector is well known between the Apuseni Mountains and Southern Carpathians (Figs. 3b and 6).

3.4. Neogene magmatism

The Carpathian arc comprises many volcanic rocks, which are either areally widespread or linearly distributed. Pécskay et al. (1995a, b), Nemčok et al. (1998), Seghedi et al. (1998, 2001, 2004) and Konečný et al. (2002) consider four types of volcanic rocks (Figs. 1a and 6): (1) initial, early Miocene acidic

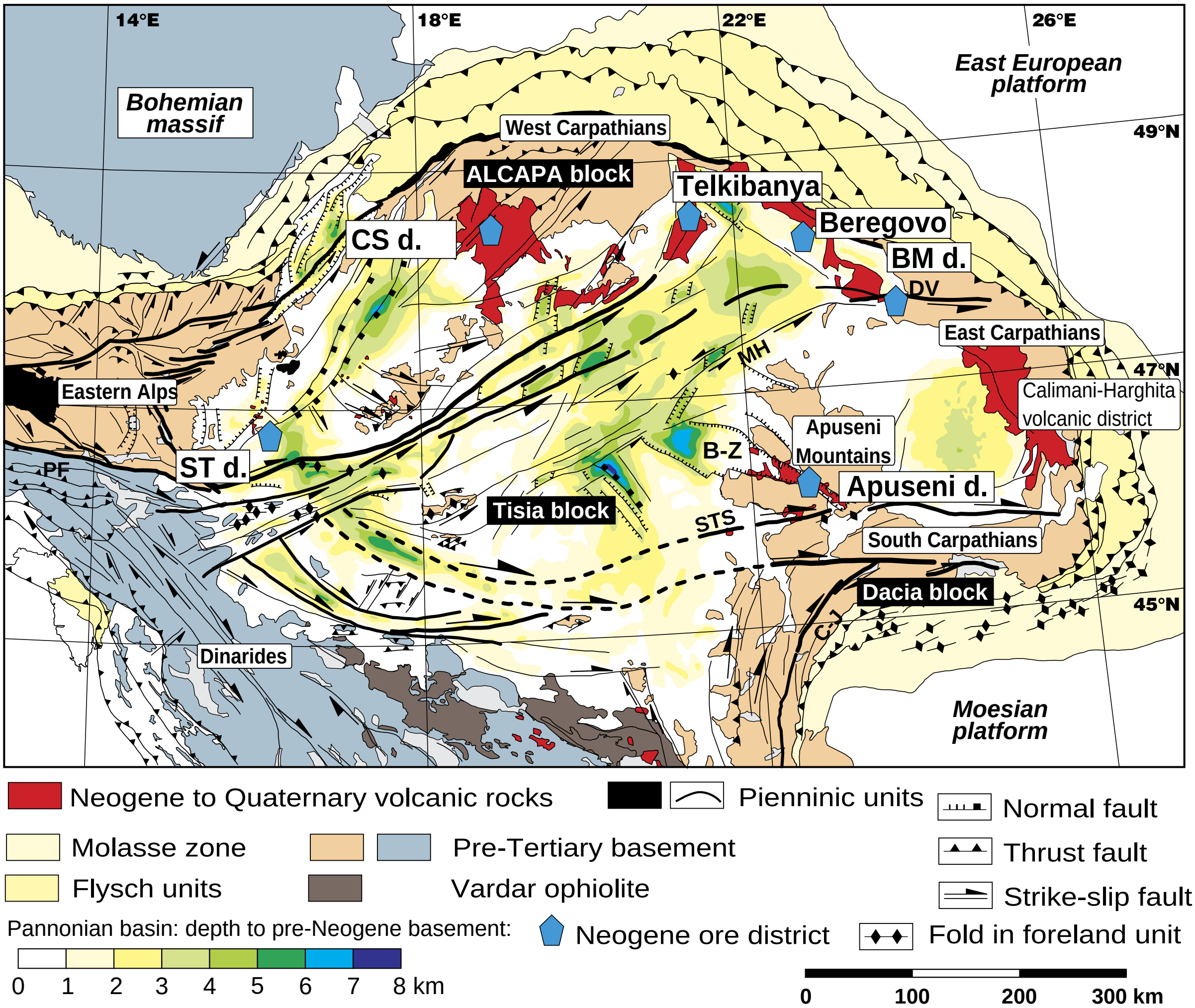

Fig. 6. Distribution of various types of Neogene sedimentary volcanic rocks and structures in the Carpathians, easternmost Alps and Pannonian basin (modified after Horváth et al., 2005). BM, ST—Baia Mare, CS—Central Slovakia (Banská Štiavnica volcano, Fig. 9), SAM—South Apuseni Mountains, ST—Styrian, d.—district. B–Z—Bekes–Zarand graben, DV—Drago Voda fault, C–J—Cerna–Jiu fault, MH—Mid-Hungarian fault, PF—Periadriatic fault, STS—South Transylvanian fault. Figure courtesy Frank Horvath, Budapest.

rocks, which are spatially widespread in the Western Carpathians and Pannonian Basin; (2) spatially widespread calc-alkaline volcanics in the interior of the Pannonian Basin (Fig. 6); (3) a linear calc-alkaline province, which extends parallel to the outer margins of the Carpathians and which linearly decreases in age from early Miocene in the easternmost Eastern Alps and Western Carpathians to Quaternary in the bend zone between the Eastern and Southern Carpathians, and (4) middle to late Miocene and Quaternary alkaline volcanic rocks widely distributed over the whole Pannonian Basin. The first group is considered to result from extension; the second and third groups from subduction and/or slab break-off, although details are highly controversial, and the fourth group from lithospheric attenuation and asthenospheric uprise due to back-arc basin formation.

As already mentioned, Miocene volcanism is widespread, and only Miocene calc-alkaline volcanism types 2 and 3 are related to mineralization, so these are the focus of the further discussion. Many studies have been carried out and a huge amount of data is now available, due to prolonged work by Mason et al. (1996, 1998), Seghedi et al. (1998, 2001, 2004, 2005), Konečný et al. (2002) and Roşu et al. (2001a,c, 2004). With regard to calc-alkaline magmatism, most recent interpretations suggest the following:

(1) Calc-alkaline melts have been produced by a fluid-contaminated, heterogeneous asthenospheric mantle source, as a result of decompression and thermal anomalies due to asthenospheric upwelling (e.g., Seghedi et al., 2004).
(2) Petrogenetic processes conducive to fluid-rich magmas and porphyry-type deposits resulted from fluids derived from partial melting of the mantle caused by decompression during lithospheric attenuation. Subducted marine sediments, as suggested by mantle tomographic studies (Fan et al., 1998), may have released high-salinity fluids, which could be incorporated into uprising magmas.
(3) The control on location is both in lithospheric breaks and upper crustal structures. In the Apuseni Mountains, volcanic rocks are related to the NW-trending graben. In the Eastern Carpathians, the volcanic rocks follow the transition between the Pannnonian basin and the inner margins of basement exposed in the Western Carpathians.
(4) Neogene calc-alkaline magmatism and associated mineralization are linked with slab tear (Linzer, 1996; Wortel and Spakman, 2000) but appear to be characteristic of normal subduction tectonics.
(5) An adakitic tendency has been observed in magmas of the southern Apuseni Mountains (Roşu, 2001; Roşu et al., 2004).

It is possible that the initiation of subduction of continental lithosphere was the underlying cause of both the calc-alkaline magmatism and associated mineralization and the initiation of slab tear. A number of geodynamic mechanisms might have been responsible for producing the magmatism, such as slab roll-back due to "normal" subduction, slab tear and slab delamination (see Seghedi et al., 2004 for discussion). However, we note that slab tear occurred in all three mineralized areas, but not necessarily at the time of mineralization. Also, slab tearing progressed further to the SE where barren or "scarcely mineralized" volcanic districts occur (Borcoş and Vlad, 1994).

The spatial and temporal evolution of the magmatism and its changing geochemical signature from W(NW) to E(SE) strongly suggests a link with the contemporaneous northeastward roll-back of the subducted slab and a progressive southeastward detachment during accelerating roll-back. This interpretation of the geodynamic evolution is further supported by the present-day overall and detailed mantle lithospheric density images, the present-day heat flow patterns, the crustal architecture and its evolution, and the spatial and temporal evolution of depocentres around the Carpathian arc.

Starting in the late–early to middle Miocene (~18 to 10 Ma), back-arc extension affected the whole Pannonian Basin region (Horváth, 1993; Nemčok et al., 1998; Fig. 3c). However, the mineral deposits in the Western Carpathians, Eastern Carpathians and Apuseni Mountains are too synchronous relative to their individual volcanic history (Fig. 2) and contrast too much with younger volcanics of similar style, but barren, in southeastern parts of the Carpathians to simply link them directly to the slab evolution. Within the Pannonian Basin, the latest stage of back-arc extension is associated with Pliocene to Quaternary alkaline magmatism and uplift of the asthenosphere to levels as shallow as 50 km (Horváth et al., in press and references therein).

4. Characteristics of Neogene ore districts

Basic features (e.g., type and age range of magmatism, type of mineralization and its isotopic characterization) of ore districts under consideration are compiled in Figs. 2 and 3, Table 1, Box 1-1 (Kouzmanov et al., 2005a), Box 1-2 (Kouzmanov et al., 2005b) and Box 1-3 (Lexa, 2005) and are discussed in the following sections.

4.1. South Apuseni Mountains district: epithermal vein Au and porphyry Cu deposits

The South Apuseni Mountains (SAM) Neogene calc-alkaline belt is famous for associated world-

Table 1
Representative data for various Neogene mineral deposits of the Inner Carpathians

District	Styrian district (Eastern Alps)	Central Slovakia volcanic district	Baia Mare district (Eastern Carpathians)	South Apuseni Mts. district ("Golden Quadrilateral")
Deposit types	Barren	Porphyry/skarn, LS epithermal gold Base–metal-rich LS epithermal veins	Epithermal low- to intermediate-sulphidation veins, cavity fillings	(1) epithermal Au–Ag veins and breccia-hosted low- and subordinate high-sulphidation ores (2) porphyry Cu–Au with stockwork, dissemination
Main metals	–	Cu, Pb, Zn, Au, Ag	Pb, Zn, Cu, Au, Ag ± Sb	Cu–Au–Ag–Pb–Te
Principal ore minerals	–	Chalcopyrite, galena, sphalerite, gold	Galena, sphalerite, chalcopyrite, gold, sulphosalts, tungstates	(1) chalcopyrite, bornite (2) gold, electrum, sphalerite, galena, tellurides, high variability of sulphosalts
Mineralization-related structures	–	N to NNE-trending veins and normal faults (releasing overstep in a wrench system)	E-trending transtensional fault, NE-trending veins	(1) E-trending Cu–(Au)-bearing veins (2) NW-trending veins (3) WNW-trending veins
Type of volcanism	Trachyandesite, latite, shoshonite	Andesite	Andesite	Andesite, dacite, adakitic-like rocks (<12 Ma)
Plutonism	No plutonic rock known	Granodiorite, diorite		Subvolcanic dacite, adakite
Age range of hosting magmatism	18 to 14 Ma	17 to 11.5 Ma	13.4 to 7 Ma	14.7 to 7.4 Ma
Age range of mineralization (alteration)	<18 Ma	12 to 10.7 Ma 12.5 to 12.2 Ma for NNE-trending veins	1.5 to 10 Ma 9.4 to 7.4 Ma	14 to 11.5 Ma
$^{208}Pb/^{204}Pb$	–	38.922 to 39.011	38.841 to 38.9931	38.476 to 38.747
$^{207}Pb/^{204}Pb$	–	15.659 to 15.682	15.652 to 15.703	15.628 to 15.659
$^{206}Pb/^{204}Pb$	–	18.810 to 18.839	18.752 to 18.876	18.497 to 18.740

Lead isotopic data are from Lexa et al. (1999a, b) and Marcoux et al. (2002). See text for sources of other data.

class epithermal vein, breccia-hosted gold and associated porphyry copper deposits (e.g., Roşia Montană; Barza, Rosia Poieni) within the historic "Golden Quadrangle" (or "Golden Quadrilateral") mining region (Ghiţulescu and Socolescu, 1941; Ianovici et al., 1969, 1976; Fig. 7; Box 1-1, Kouzmanov et al., 2005a). The high density of epithermal vein deposits, spatially and genetically related to porphyry Cu–Au-bearing deposits, gives rise to Au-rich and Cu-, Pb-, and Zn-bearing, complex magmatic–metallogenic structures (e.g., Alderton and Fallick, 2000; Udubaşa et al., 2001; Ciobanu et al., 2004). Some of the ores are unusually rich in tellurium (e.g., Săcărîmb; Ciobanu et al., 2004; Cook and Ciobanu, 2004b and references cited therein). Ca. 2000 t of gold have been produced since pre-Roman times (see discussion in Cook and Ciobanu, 2004b). Recently, the largest gold resource containing ca. 500 t Au has been identified at Roşia Montana (Leary et al., 2004).

All ore deposits of the South Apuseni Mts. district are associated with volcanic and subvolcanic rocks and cluster within three essentially NW–SE alignments (Fig. 7; Box 1-1, Kouzmanov et al., 2005a): Zarand–Brad–Săcărîmb, Zlatna–Stanija and Roşia Montana–Bucium. In addition, the Deva and Baia de Arieş deposits appear as isolated volcanic centres located away from the main alignments.

All these deposits can be characterized by three major types of mineralization (e.g., Roşu et al., 2001a,c; Ciobanu et al., 2004):

(1) Epithermal gold veins (e.g., Ruda-Barza, Brădişor and Musariu) dominate and are developed mainly as steeply dipping veins crosscutting

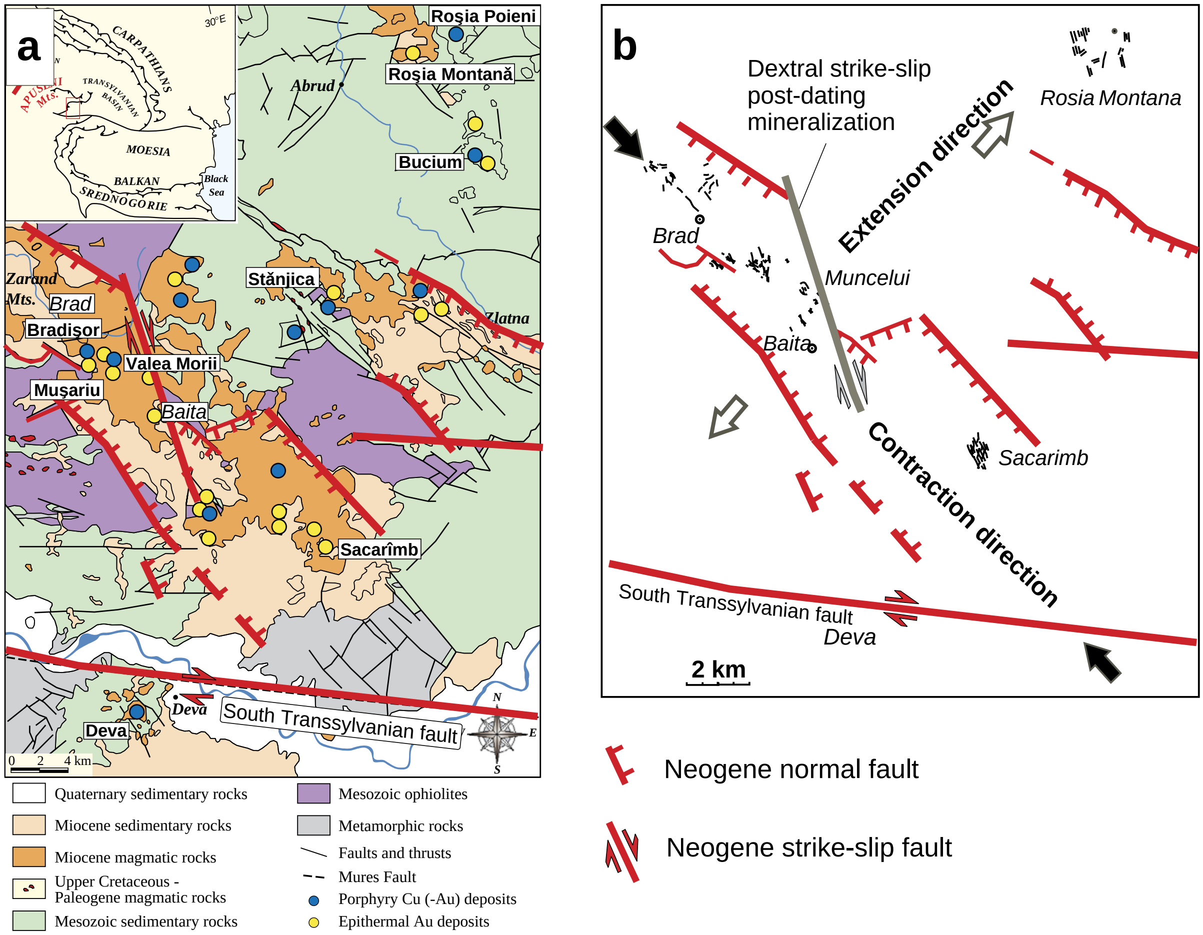

Fig. 7. a—Detailed map of Neogene structures and magmatic rocks in the South Apuseni Mountains (after Roşu et al., 2001b). b—Main structural elements and hydrothermal vein systems of the South Apuseni Mountains (adapted from Ianovici and Borcoş, 1982).

basement rocks, andesitic lavas, subvolcanic intrusions and Cretaceous or Tertiary sediments. Epithermal mineralization post-dates the porphyry-type ore formation and is commonly associated with propylitic, quartz–adularia–illite to argillic alteration (Milu et al., 2003, 2004). Minerals in epithermal veins occur in open-space filling and crustiform textures. Gold, electrum ± sphalerite, galena, and many species of tellurides (e.g., at the classical, but exhausted deposit Săcărîmb) are common in these epithermal vein deposits (Cook et al., 2004).

(2) Porphyry copper mineralization (Deva, Roşia Poieni, Valeia Morii, Musariu) is contemporaneous with associated older events of Miocene magmatism. Porphyry mineralization is hosted mainly by subvolcanic andesitic bodies and forms stockworks, disseminations and impregnations, sometimes occuring as breccia infill (Boştinescu, 1984, Berbeleac et al., 1995). Telescoping of porphyry and epithermal mineralization is common. Potassic, phyllic to argillic alteration is common in the porphyry copper deposits. The main ore minerals are chalcopyrite and bornite.

(3) Breccia-hosted epithermal deposits are common at Roşia Montană, set to become Europe's largest gold mine (Leary et al., 2004). Further sizable resources of breccia-hosted gold ore are also present at Bucium Rodu-Frasin, and

the recently closed Baia de Arieş mine with predominant breccia pipe ores.

The unusual position of SAM volcanism within the Carpathian belt, away from the main Eastern Carpathian arc, and the obvious concentration within a restricted quadrangle (1000 km^2) has been noticed and has given rise to various geotectonic models, which include subduction-related processes (e.g., Rădulescu and Săndulescu, 1973; Boccaletti et al., 1973), slab break-off and roll-back, and post-collision extension-related magmatism (Roşu et al., 1997, 2000, 2001a,b, 2004; Balintoni and Vlad, 1998; Seghedi et al., 1998).

The volcanic rocks occur within a NW-trending tectonic graben, which is filled with Miocene, mainly Sarmatian, sedimentary rocks (Fig. 7). The underlying Faţa Băii Formation constitutes a more areally extensive clastic–volcanic formation, spreading beyond the Neogene graben, which includes conglomerate, rhylolitic, rhyodacitic and andesitic tuffs and is considered of Palaeogene age (see Ciobanu et al., 2004 for discussion). Neogene volcanism following deposition of *Globigerina*-bearing marls postdates the onset of graben formation. This graben extends well into the Pannonian Basin, where it forms part of the Bekes–Zarand graben (Fig. 6), one of the deepest graben systems within the whole Pannonian Basin (Csontos and Nagymaroshi, 1998). The graben extensionally widens towards the NW by the opening of confining NW-trending normal fault systems to the NW. Structures at 10 km-scale suggest subsequent late Neogene dextral displacement along NW-trending faults (e.g., Drew, 2003). Some E-trending strike-slip faults are subordinate. The graben terminates at the Transylvanian fault in the S (Fig. 7). The two fault systems (E–W and NW–SE) are interpreted as an unusual horse-tail, pull-apart structure, which probably formed during rotation around a centre located in the SE.

Extensive palaeomagnetic evidence (Pătraşcu et al., 1994; Roşu et al., 1997, 2004; Panaiotu, 1998) shows rotation of individual crustal blocks. For example, the Apuseni Mountains block rotated clockwise by 60° between 15 and 12 Ma and demonstrates the complex history of crustal deformation. The peak of deformation was almost synchronous with magmatism, but slightly pre-dated it. As the main deformation is genetically related to involvement of continental lithosphere in subduction and subsequent collision (slab break-off, slab tear) the associated magmatism must have formed in a collisional environment by a mechanism such as partial melting of a water-rich, sublithospheric source (Seghedi et al., 2004; Roşu et al., 2004). This is possible if subduction-derived fluids were available as trap reservoirs at the base of the lithosphere and/or upper mantle and were activated by asthenospheric thermal influx and decompression.

The emplacement of magmatic structures and ore deposit genesis were controlled by strike-slip stepover extension-transfer zones and pull-apart structures, as shown by the well-expressed ring-like geometry of intrusions and the central position of productive porphyries within narrow, fault-controlled sedimentary basins (Fig. 7; Box 1-1, Kouzmanov et al., 2005a).

Correlation of emplacement ages with palaeomagnetic data demonstrates that SAM underwent extensive clockwise rotation from 15 to 12 Ma (Roşu et al., 1997, 2001a; Panaiotu, 1998). The main SAM Neogene magmatic activity, representing up to 90% of the total volume (andesites/microdiorites) and, with few exceptions, all the productive ore structures, were emplaced simultaneously with the main paroxysmal upper crust brittle transtensional deformation between 14 and 12 Ma (Roşu et al., 1997).

Steep epithermal vein networks, shallow, small-sized porphyry copper and gold structures, disseminated breccias and arrays occur within different extension and/or pull-apart basins and form complex, tectonically controlled, base-metal and gold vein and cogenetic porphyry systems. For further information, see Box 1-1 (Kouzmanov et al., 2005a).

A more complex regional pattern has resulted from spatial superposition of a pre-ore magmatic sequence (quartz–biotite–amphibole andesites and dacites) and a post-ore magmatic sequence (andesites to basaltic andesites). The pre-ore and pre-deformation sequence underwent a similar rotation to the main sequence rocks and no distinct metallogenic activity can be traced, except for the superposition of vein systems relative to the main magmatic ore-forming sequence. The post-deformation magmatic sequence, with no palaeomagnetic evidence of rotation between 12 and 7 Ma (Roşu et al., 1997) is, with few exceptions, postdating the main metallogenic activity and shows an obvious barren character.

At the scale of magmatic structures, emplacement of porphyry copper structures and epithermal vein formations was accommodated by extensional strain components and mineralized fault structures. The geometry of mineralized veins is fairly well in accord with shear-related tectonics such as steep veins parallel to extensional directions, tensional shear vein meshes (Drew and Berger, 2001; Drew, 2003), sometimes with typical flower structure, or a configuration of Riedel-oriented complex vein systems (Fig. 7, the Musariu Barza vein system) related to the South Transylvanian fault.

The ore-bearing extensional veins largely trend ca. E–W (mainly early stage quartz veins in porphyry Cu systems) and NW–SE, often crosscut by a third set of WNW–ESE veins, and bear a dextral shear character (Fig. 7; after Ianovici and Borcoş, 1982). At Săcărîmb, the dominant set trends NNE, a subsidiary set trends NNW. Except for Săcărîmb, the superposition of these three types of structures can be interpreted as representative of horse-tail structures, which are well known from analogue experiments above so-called basement faults (Mandl, 1988). Basement faults evolve when distributed strain is induced at shallow structural levels by a deep-seated, narrow strike-slip fault. In comparison with these experiments, the gold-bearing vein system of the southern Apuseni Mountains can be explained by releasing oversteps within an eastward moving block. This explanation partly contrasts with the recent explanation given by Drew (2003). At the present state of research, it is unclear whether or not the vein system rotated. At some localities, e.g., at Săcărîmb, two sets of veins are present, trending NNE–SSW and NNW–SSE. The superposition of the three (two at Săcărîmb), vein systems could be interpreted by clockwise rotation of the now NNE-trending system within a constant external stress field.

Examples of superimposed epithermal vein systems on porphyry copper structures are extensive within the SAM district (e.g., Valea Morii, Musariu, Roşia Poeni, Bolcana, etc.) and a continuity of hydrothermal activity within tensional vein meshes from porphyry to epithermal can be traced (Milu et al., 2003, 2004).

Porphyry-style networks show distinct patterns from curved trace microfractures (A-type) within a “ductile” environment, together with potassic alteration in the initial stage of porphyry formation, succeeded by straight-trace fractures (B-type) in the main porphyry stage, as a steep stockwork, around a barren core or as stress-oriented, fracture-controlled, limited stockworks (e.g., Bolcana, Musariu).

Superposition of epithermal vein meshes on porphyry intrusive structures is characteristic and controlled by the regional strain. Recirculation on dense B-type stockworks is often within the epithermal stage, together with intense sericite or clay mineral alteration. The gold-rich character of porphyry copper structures is correlated with high-grade gold within epithermal vein meshes and/or tectonically controlled, recirculated, high density B-type porphyry stockworks. Complex brecciation producing porphyry or epithermal breccias focused the fluid flow within the porphyry, allowing boiling and gold and base metal deposition. Cook and Ciobanu (2004a) suggested that sulphidation, induced by rapid opening of kin veins prior to maturity of the porphyry system, might have been an important factor in destabilizing metal complexes in the fluid during their ascent. Au scavenging by Bi–Te melt precipitates would be activated, given that the temperatures exceed 400 °C. This might have resulted in apparently missing porphyry roots of epithermal vein systems like Sacărîmb.

4.2. Baia Mare district of the Eastern Carpathians

The Baia Mare ore district of the Eastern Carpathians is controlled by the intersection of the E-trending Dragos Voda fault system and the NNW-trending late Miocene volcanic belt (Figs. 1, 6 and 8); Box 1-2, Kouzmanov et al., 2005b). Nearly all ore deposits are of vein type and comprise Pb–Zn(–Cu–Au–Ag) (Borcoş et al., 1974, 1975, 1976; Borcoş and Gheorghita, 1976; Lang, 1979; Borcoş and Vlad, 1994; Kovacs, 2001; Kovacs et al., 2001; Marias et al., 2001; Marcoux et al., 2002; Grancea et al., 2002, 2003). The Baia Sprie deposit is located within an E-trending graben up to 800 m wide (Figs. 1 and 8a; Box 1-2, Kouzmanov et al., 2005b) that is filled with Upper Badenian to Pannonian sedimentary rocks (Ciulavu, 1998; Bailly et al., 2002). This structure signifies the transtensional nature of the otherwise

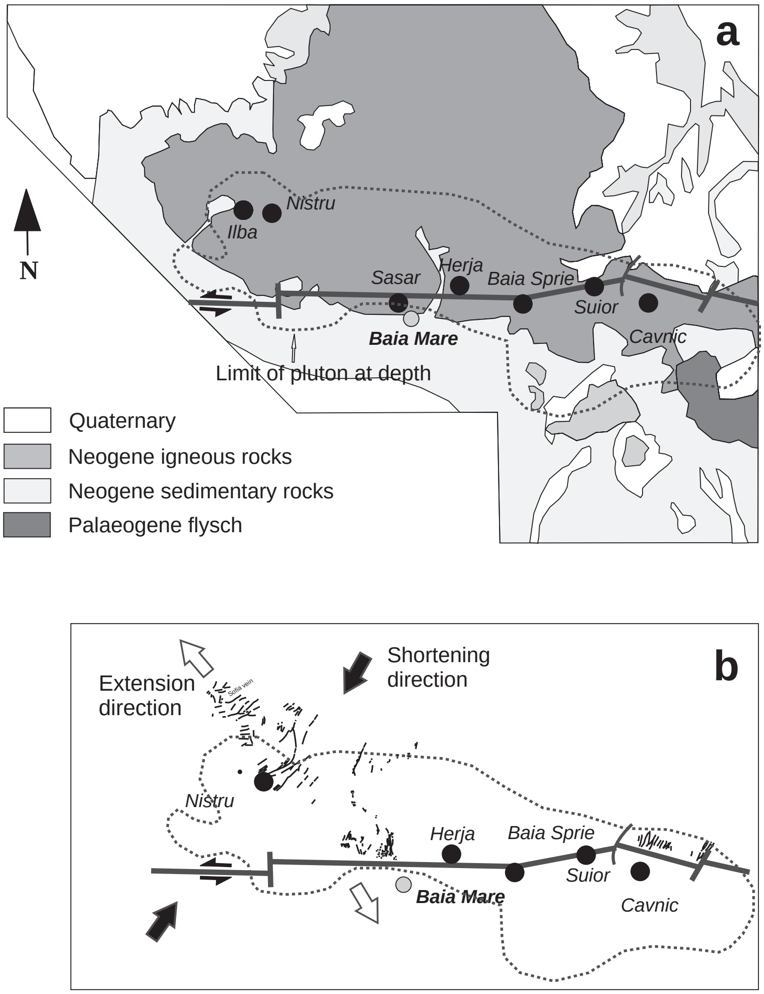

Fig. 8. a—Structural map of central sectors of the Baia Mare district, including distribution of mineralization (modified after Ianovici and Borcoş, 1982; Bailly et al., 2002). b—Interpretation of vein systems of the Baia Mare district (from Ianovici and Borcoş, 1982) based on basement faulting experiments.

almost pure strike-slip Dragos Voda fault. Other ore-bodies are along faults well known to the N of the Dragos Voda master fault. These faults mostly trend NNW–SSE, but some further faults are WNW-trending (Ianovici and Borcoş, 1982).

The Baia Mare district is associated with middle to late Miocene to Pliocene calc-alkaline volcanics and plutonic rocks. Volcanism ranges from 13.4 to 6.9 Ma and is contemporaneous with deposition of Upper Badenian to Pannonian sedimentary rocks. The Baia Mare district comprises, from NW to SE, the following major deposits and camps: Gutai Mountains with Ilba and Nistru, Baia Mare, Baia Sprie and Cavnic. Previous studies established the presence of a narrow, major E-trending pluton along the Dragos Voda fault, at a depth of 1 to 4 km beneath the surface (Borcoş and Vlad, 1994; Bailly et al., 1998, 2002). Consequently, the overall structural control is the transtensional nature of the Dragos Voda transform fault, which allowed the emplacement of a major linear pluton at depth.

Mineralization is in vein systems that are several hundred m to 5 km long, 0.3 to 10 m wide and can be followed to depths of 300 to 1,000 m. Veins strike E–W in Baia Sprie, NNE–SSW in Cavnic, N–S to NNE–SSW in Baia Mare, and NE–SW-trending at Nistru

(Fig. 8b). Both vertical and horizontal zoning is common in these veins (Box 1-2, Kouzmanov et al., 2005b). Vertical zoning includes, from margin to centre, Au–Ag, Pb–Zn and Cu, horizontal zoning Pb + Zn, Cu ± Pb + Zn, Cu + Au. Details of mineralogy are described by, amongst others, Cook (1997), Cook and Damian (1997), Kovacs et al. (2001) and Marias et al. (2001). Hydrothermal systems were controlled by tectonic and magmatic events, injection of magmatic fluids and their mixing with meteoric water, and boiling processes. Complex mineral/metal systems include earlier Fe–Sn–W–Bi systems, which evolved into late-stage Au–Ag–As–Sb systems. Lead isotopic data show some crustal contamination of the magma-derived metals from the local basement (Marcoux et al., 2002).

Based on earlier investigations (Borcoş et al., 1974, 1975, 1976; Borcoş and Gheorghita, 1976; Lang, 1979; Borcoş and Vlad, 1994), Bailly et al. (1998, 2002) and Grancea et al. (2002) proposed a five-stage evolution of mineralization: (1) an initial Fe precipitation stage; (2) Cu–(Bi)–W stage; (3) Pb–Zn stage; (4) Sb stage, and (5) Au–Ag stage. Alteration assemblages include quartz–illite/sericite–adularia for Au–Ag systems and quartz–calcite–rhodochrosite–rhodonite for Pb–Zn mineralization. Two periods of hydrothermal activity were dated: 11.5 to 10.0 Ma and 9.4 to 7.4 Ma, the latter for a base–metal and epithermal vein system, e.g., of Baia Sprie and Cavnic (Lang et al., 1994). Sericite of the first alteration type was dated at 7.8 to 8.0 Ma (Fig. 2).

Compared with analogue experiments summarized by Mandl (1988, 2000), this configuration is well established in wrench corridor regions between master faults arranged en echelon or extensional oversteps, again in basement fault systems. The hydrothermal veins give the general NE–SW to NNE–SSW orientation of the maximum principal stress axis of the external stress field. These veins can become curved in a sigmoidal manner during wrench movements (e.g., to the N of Baia Mare).

4.3. Western Carpathians

Inner sectors of the Western Carpathians expose middle Miocene calc-alkaline, mainly andesitic, stratovolcanoes, which host a number of ore deposits (Figs. 1 and 9; Box 1-3, Lexa, 2005). These stratovolcanoes mainly include the Central Slovakia volcanic field with the prominent Štiavnica and Kremnica, Javorie and Pol'ana stratovolvanoes (Fig. 8; Lexa et al., 1999a,b), and the Tokaj Mts. (Fig. 1; Molnár et al., 1999). Among these, we summarize data from the Central Slovakia volcanic field, based on the work of Lexa (1999b) and Lexa et al. (1999a).

Mineral deposits of the Central Slovakia Neogene volcanic field are hosted by the central zones of large andesite stratovolcanoes (Fig. 9) involving volcanotectonic depressions, resurgent horsts, extensive subvolcanic intrusive complexes and complexes of differentiated rocks. Thirteen types of mineralization have been recognized, but only three of them have formed major ore deposits (Box 1-3, Lexa, 2005): (1) resurgent horst and rhyolite related low sulphidation Au–Ag deposits: Kremnica, Pukanec, Nová Baòa, Banská Belá; 2) caldera related low sulphidation Au deposit: Hodruša-Rozália; 3) intermediate sulphidation Pb–Zn–Cu–Ag–Au deposits: Banská Štiavnica, Hodruša.

The Kremnica deposit is situated on marginal faults at the eastern side of the Kremnica resurgent horst (Fig. 9) in the central part of the N–S-trending Kremnica volcanotectonic graben. Epithermal veins are hosted by propylitized andesites in the central zone of a large andesite stratovolcano (around 16.2 to 15.0 Ma). Epithermal veins were much younger, contemporaneous with the uplift of the horst and the late stage rhyolite volcanic activity. Radiometric dating of sericite and adularia has given the age as 11.1 to 10.1 Ma (Kraus et al., 1998). The 6-km-long N–S trending system of epithermal veins is dominated by a listric fault dipping 60° at the surface to 50° at a depth of 1300 m (Fig. 10). Whilst the vein is relatively narrow at depth, it gradually opens towards the surface up to the maximum of 80 m, and branches into a funnel-shaped system of veins and veinlets, including complementary antithetic veins. Tops of the veins are exposed about 200 m below the former surface. The gangue of the veins is represented by banded and cavernous quartz with a variable proportion of carbonate. Darker quartz/chalcedony bands with fine dispersed pyrite and/or marcasite are also enriched in gold and silver. The rare presence of hydrothermal breccias implies that precipitation of sulphides and gold was probably related to boiling. This is supported by the occurrence of

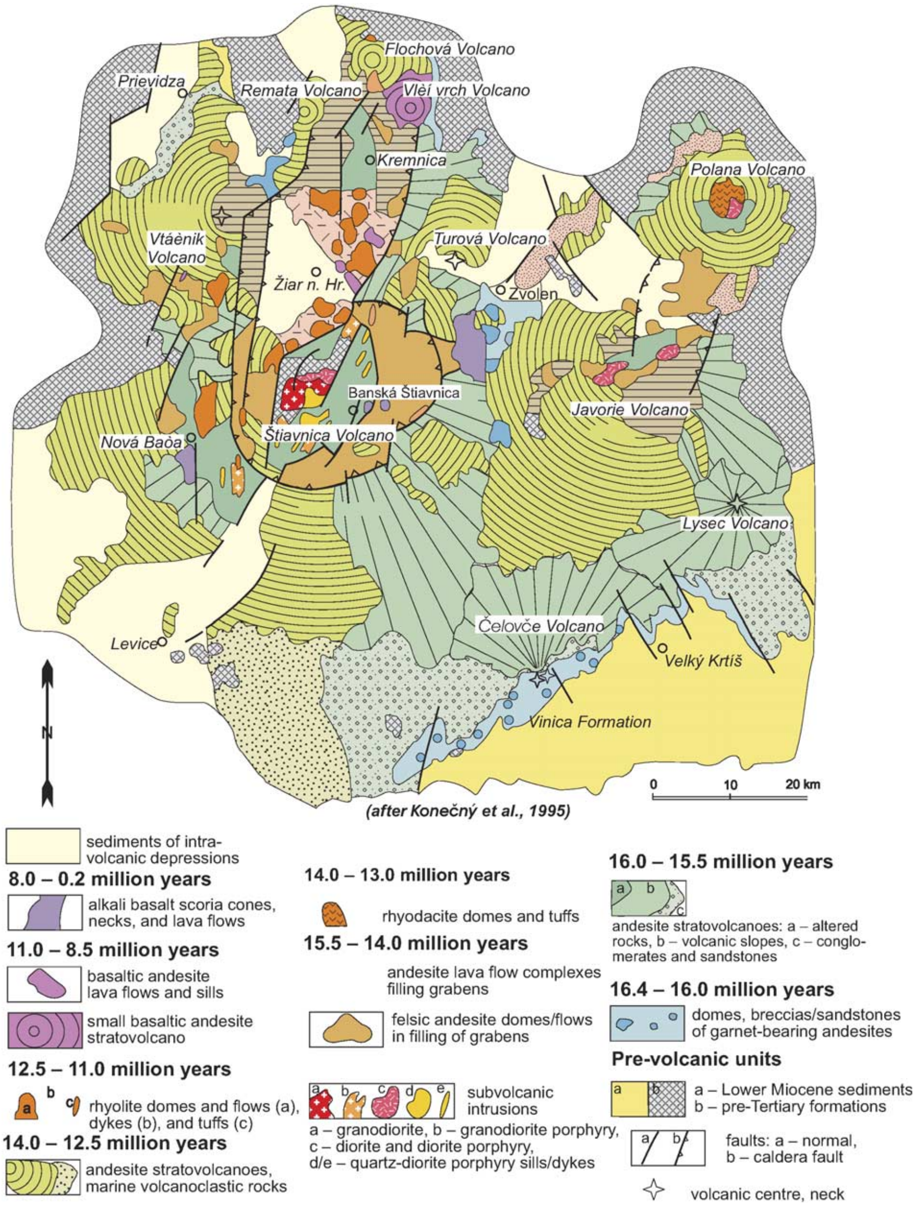

Fig. 9. Geological map of the Central Slovakian volcanic field including the Banská Štiavnica stratovolcano (modified after Konečný et al., 1995; Lexa et al., 1999a).

bladed carbonate crystals as well as by the occurrence of adularia in veins and surrounding rocks. Microscopic gold is present as electrum in pyrite and quartz. Silver sulphosalts and base–metal sulphides are present in minor amounts only. Their content increases with depth. The epithermal system is accompanied by extensive hydrothermal alterations (Kraus et al., 1994). Strong silicification, adularization and/or sericitiation close to the veins passes outwards into a zone with mixed-layer I/S and CH/S argillic minerals and finally into a zone with montmorilonite. Structural aspects, mineral associations, zoning and alteration at the Kremnica ore deposit define the deposit as a typical low-sulphidation epithermal system. Despite the fact that the system is located in the central zone of the older andesite stratovolcano with intrusion-related alteration and minor mineralization, its evolution was related rather to the much younger resurgent horst uplift and related late-stage rhyolite magmatism.

The Banská Štiavnica–Hodruša ore district is localized in the central zone of a large and complex andesite stratovolcano (Fig. 9) involving seven types of mineralization, but only Pb–Zn–Cu–Ag–Au epithermal veins were the subject of extensive exploitation, which dates back to the 12th Century and continued until 1947 (Koděra et al., 2004, 2005 and references therein). The system of Pb–Zn–Cu–Ag–Au epithermal veins extends across an area of 100 km^2 (Box 1-3, Lexa, 2005) and involves 120 veins and vein branches with lengths up to 8 km and thickness up to 10 m in ore shoots (Lexa et al., 1999a,b; Koděra et al., 2004, 2005). Three types of epithermal veins occur in a zonal arrangement: (a) base–metal veins $\pm$ Ag, Au with transition to Cu $\pm$ Bi mineralization at depth in the east/central part of the system, (b) Ag–Au veins with minor base–metal mineralization, mostly at the western part of the system, and (c) Au–Ag veins located around the outskirts of the system. Epithermal veins localized on faults of the post-caldera resurgent horst (Fig. 9) are hosted by massive andesites, andesite porphyry, quartz–diorite porphyry, granodiorite, diorite, and, in the western part, by basement rocks. Host rocks are generally affected by propylitic alteration. Veins are accompanied by zones of silicification, adularization and sericitization, especially those poor in sulphide and base metals. Most of the veins have banded textures and hydrothermal brecciation is frequent. Fluid inclusion studies point to a system that was frequently boiling (Kova-

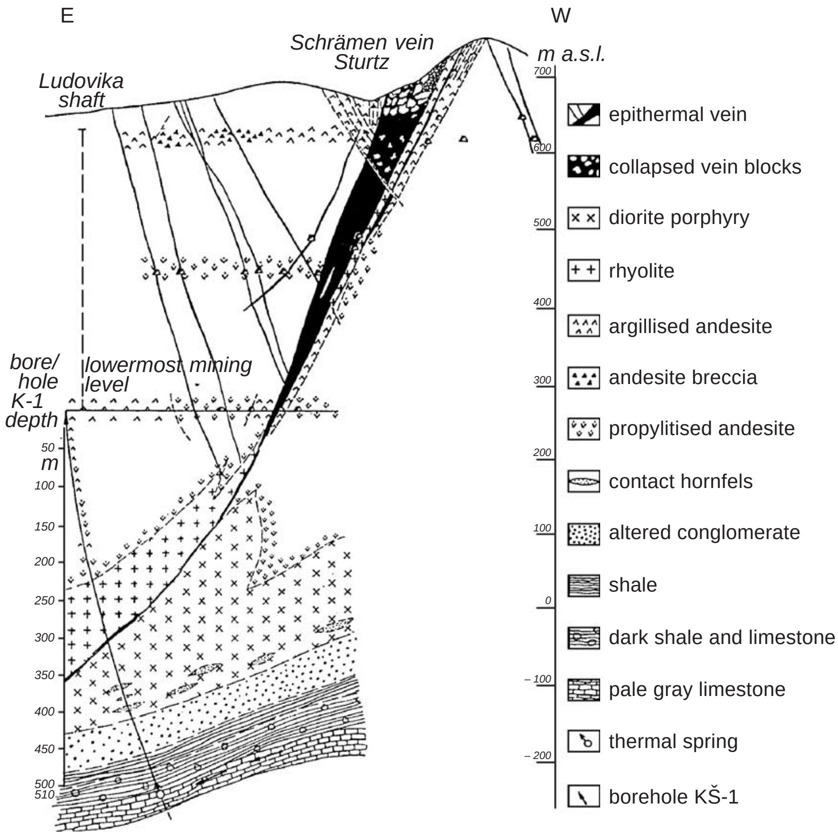

Fig. 10. Cross-section across Kremnica deposit (modified from Lexa and Bartalský, 1999).

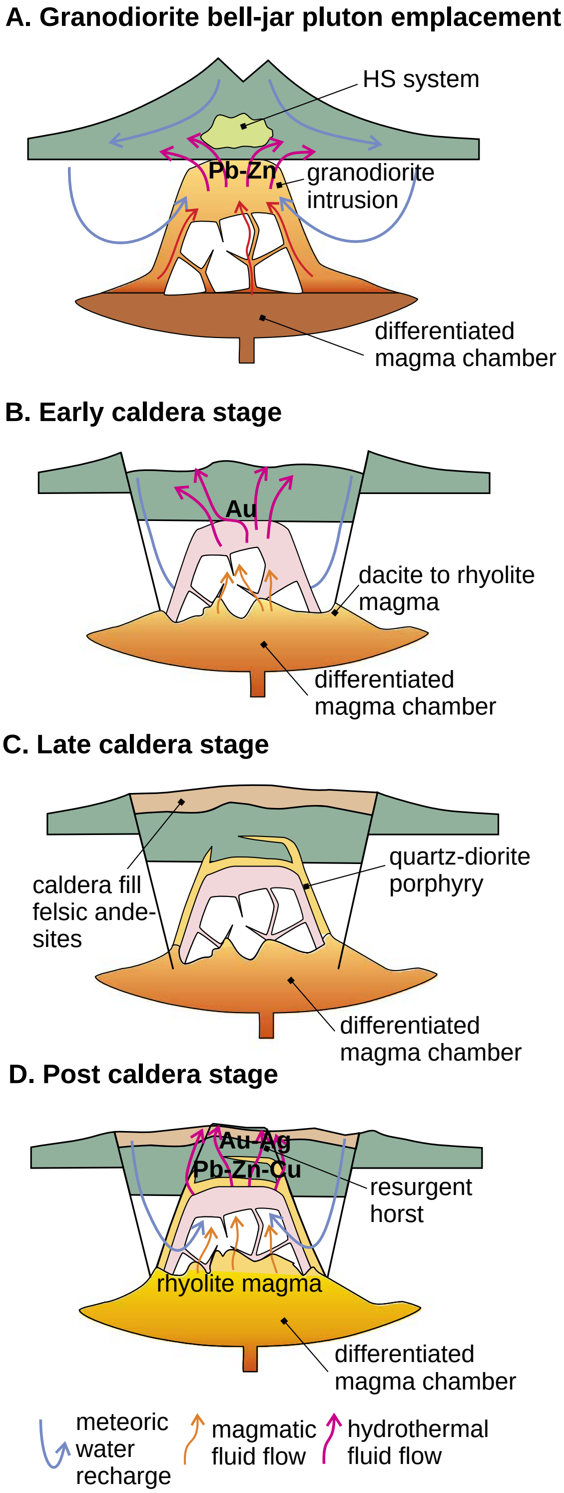

Fig. 11. Evolutionary model of the Banská Štiavnica stratovolcano including various types of mineralization (after Lexa, 1999a; Lexa et al., 1999a; Koděra et al., 2005).

lenker et al., 1991). Isotopic compositions of oxygen and hydrogen in hydrothermal fluids indicate mixing of magmatic and meteoric components (with a generally increasing proportion of meteoric component towards younger periods of mineralization). Veins are accompanied by zones of silicification, adularization and sericitization, indicating a low- to intermediate-sulphidation environment (Fig. 11).

The Hodruša–Rozália deposit represents another type of epithermal gold mineralization. It occurs in the form of subhorizontal veins and veinlets in extensively altered pre-caldera stage andesites, not far from the roof of a granodiorite pluton (Fig. 9). The mineralization is dismembered by a younger set of quartz–diorite porphyry sills and segmented by faults of the post-caldera resurgent horst. Banded veins and veinlets of quartz, rhodonite, rhodochrosite and adularia with minor pyrite, sphalerite, chalcopyrite and Au of moderate fineness are 0.1 to 2 m thick. Fluid inclusion studies (Koděra et al., 2005) proved that fluids of low salinity and moderate temperatures were boiling at pressures between 45 and 114 bars. Oxygen and hydrogen isotope data point to homogeneous mixed magmatic and meteoric fluids.

The resurgent horst in the central zone of the Štiavnica stratovolcano also hosts other mineralization types (Lexa et al., 1999a; Lexa, 2001). Koděra and Lexa (2003) have summarized the following metallogenetic model (Fig. 11):

- There is no mineralization related to the *pre-caldera andesite stage* (16.0 to 15.5 Ma).
- The diorite intrusion of the *subvolcanic intrusion stage, phase 1* is a parental body to the *barren high sulphidation (HS) system at Šobov*.
- The granodiorite bell-jar pluton of the *subvolcanic intrusion stage, phase 2* is a parental body to the *magnetite skarn mineralization* at contacts with Mesozoic carbonate rocks and *stockwork disseminated base–metal mineralization* in apical parts of the pluton accompanied by HS alterations in overlying andesites.
- The granodiorite to quatz–diorite porphyry stocks of the *subvolcanic intrusion stage, phase 3*, host *porphyry/skarn type Cu ± Mo, Au mineralization*.
- Volcanic rocks of the *caldera stage* (14.5 to 14.0 Ma) filling the caldera host several *hot spring type mineral deposits* representing outflows of epithermal systems, including the low sulphidation (LS) epithermal system of the Hodruša–Rozália Au deposit.
- Marginal faults of the *resurgent horst* host an extensive and long lasting (12.5 to 10.5 Ma) system of intermediate sulphidation *(IS) to LS epithermal base–metal and precious-metal veins* (Banská Štiavnica and Hodruša deposits).

Pre-caldera mineralization types are related closely in space and time to the emplacement of individual subvolcanic intrusions, which in turn are related to evolution of magmas in a high level magma chamber following maturity of the pre-caldera stage stratovolcano. Owing to the emplacement of magma into the low-pressure environment, mineralization shows an evolution in two stages: the first, including HS systems, is related to fluid exsolution and partition of magmatic fluid into brine and vapour phases during the final stage of crystallization; the second, ore-bearing stage is related to fluids of mixed meteoric and magmatic origin (exsolution at supercritical regime at depth). A change of morphology due to long-lasting denudation and caldera collapse resulted in a completely new fluid flow pattern in the stratovolcano. The infiltration zone of meteoric water moved to the caldera margins and slopes of the stratovolcano, allowing an outflow of hydrothermal fluids from the heated central part of the caldera. A differentiated magma in the shallow magma chamber itself was the source of heat and magmatic fluids. Due to the governing stress field at the time of the initial caldera collapse, related epithermal mineralization is hosted by subhorizontal E–W extensional structures. The evolution of the LS epithermal systems was suddenly interrupted by a further subsidence of the caldera and by the emplacement of quartz–diorite porphyry sills and dykes at subvolcanic level. Post-caldera epithermal mineralization is related to the contemporaneous resurgent horst uplift and rhyolite magmatism. The uplift was most likely caused by accumulation of low-density rhyolite magma and by transformation of crustal rocks above the shallow magma chamber into anatectic magmas. Heat and fluids of the evolving magma, and opening of extension faults of the resurgent horst, renewed conditions for a large-scale circulation of fluids in the caldera. The system of epithermal veins and their zoning reflect a complex interaction of the structural evolution of the resurgent horst with the evolving (cooling) hydrothermal system. The evolution of the magma in the high-level chamber towards a rhyolitic composition by extensive anatexis created a spatial relationship between the early intrusion-related mineralization and the much younger, extensive low sulphidation epithermal mineralization.

Hydrothermal ore veins consistently trend NNE–SSW all over the Banská Štiavnica stratovolcano (Figs. 9 and 10; Lexa et al., 1999a; Nemčok et al., 2000). This can be interpreted to represent WNW–ESE extension linked with widespread normal faults. The magmatism and associated vein-type hydrothermal mineralization of the Tokaj Mountains are basically similar. Volcanism occurred within a NNE-trending graben structure. Veins trend ca. N–S (Fig. 12) and hydrothermal minerals are dated at 12.2 to 12.5 Ma and cut subvolcanic dacite domes (Molnár et al., 1999). This is consistent with E–W extension in the Pannonian Basin.

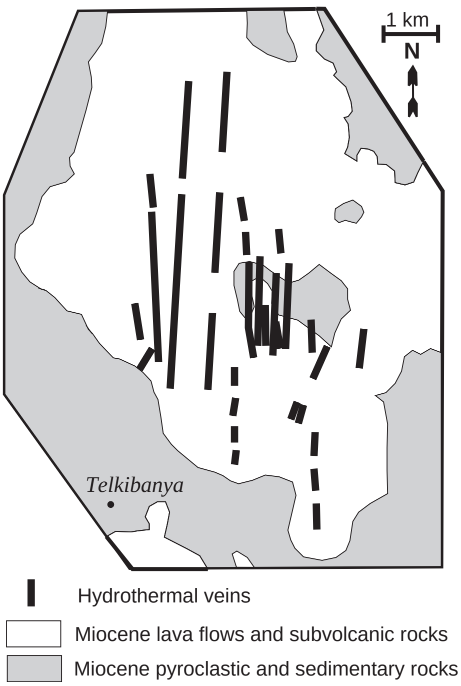

Fig. 12. Vein systems of the Telkibanya stratovolcano (after Molnár et al., 1999).

4.4. The barren Styrian volcanic field

The Styrian volcanic field of mainly Karpatian–Badenian age (17 to 13 Ma) is located on the South Burgenland high in the southeasternmost Eastern Alps (Ebner and Sachsenhofer, 1991, 1995; Fig. 1). Rocks comprise mainly high-K trachyandesites, which are exposed in a major stratovolcano (Gleichenberg), and in the subsurface (Lannach). The latter is penetrated by numerous boreholes for hydrocarbon exploration. These rocks are highly altered to an illite–alunite–christobalite assemblage (Weber et al., 1997), and exploration for Au was unsuccessful (Walter Prochaska, pers. comm.). Illite is mined for industrial purposes, so exposure for study of possible structures is good. In the Neogene Styrian volcanic field, there is a lack of any kind of fault and hydrothermal vein systems that are contemporaneous with or post-date emplacement of volcanic rocks. There are nearly no subvolcanic, shallow level plutons, except for a single trachytic dome. Consequently, shallow-level plutonic systems that could have released hydrothermal fluids were not in existence.

4.5. The "scarcely mineralized" Calimani–Gurghiu–Harghita volcanic district

SE from the Baia Mare district of the Eastern Carpathians lies the large volcanic district of the Calimani, Gurghiu and Harghita Mountains. The volcanic rocks are characterized by dominantly calc-alkaline andesites and additional basalts, basaltic andesites (especially in the Calimani Mountains) and dacites to shoshonites (in the South Harghita Mountains), and have ages ranging progressively, from NNW to SSE, from 9 Ma to nearly the Present (Pécskay et al., 1995a,b; Mason et al., 1998; Seghedi et al., 2004, 2005). Although a geodynamic setting comparable to the main mineralized districts occurred here over the last 10 Ma, with subduction, roll-back and subsequent slab tear (Wortel and Spakman, 2000), this very large volcanic district hosts very little significant base metal and Hg mineralization including aborted porphyry systems (Borcoş and Vlad, 1994).

5. Discussion

The data from Neogene ore districts of the Inner Carpathians show a number of similar features and remarkable facts:

(1) A slab window within the subducted lithosphere has been imaged by seismic tomography beneath all three ore districts under consideration, although these are present-day structures and may not necessarily reflect the structure of the middle to late Miocene.
(2) All the major Neogene mineralization is intimately associated with explosive calc-alkaline magmatism, including explosive volcanism and shallow-level plutonism. Mineralization in each district was short-lived as was the underlying volcanism. Hydrothermal activity was associated with magma-derived fluids that were variably mixed with meteoric water due to the near-surface sites of ore precipitation.
(3) Both volcanism and mineralization are unevenly distributed. Major volcanism across a wide area is mainly located in tectonic graben structures, indicating their intimate relationship with extension. These graben structures have a distinct trend and were formed or activated during rotation and final emplacement of crustal blocks. Transtensional deformation during rotation may have facilitated the upward flow of magma and release of hydrothermal fluids.
(4) There is an overall geodynamic control on the siting of the three ore districts that contrasts remarkably with those regions where calc-alkaline volcanism is related to pervasive alteration but no mineralization, particularly as some areas remained barren even though hydrothermal systems had been established. Mineralization is predominantly in the southern Apuseni Mountains, in the Baia Mare region and the Western Carpathians.
(5) In the Western Carpathians the main mineralization occurred at the end of the dominant period of volcanism, while in the Eastern Carpathains the main mineralization is temporally constrained in the middle of the period of volcanism, and in the Apuseni Mountains the mineralization occurred in the early phases of volcanism. Although the main phase of volcanism is diachronous from W to E and lasted longer before and/or after the mineralization, ore mineralization was limited to a relatively narrow time span between ca. 13 and 10 Ma in all three districts (Fig. 2). This suggests a prime control by some factor other than magmatism alone. This could have been a distinct structural event of extension or transtension, combined with large-scale block rotation.
(6) Hydrothermal lodes display a pronounced and distinct trend in each deposit, different from the overall extensional structures. The preferred vein orientation mitigates against opening mechanisms by pure fluid overpressure in all three districts, but relates these veins to regional tectonic stress orientations. The obliquity between large-scale graben orientations and hydrothermal vein orientations suggests transtensional motion during hydrothermal activity.
(7) The particular geodynamic setting also resulted in particular structures that were part of the plumbing systems. Rotation and wrenching implies subvertical orientation of the intermediate principal stress axis (σ_2) and, therefore, formation of subvertical to steep open extensional gashes and subordinate normal faults, which would have allowed preferred subvertical fluid flow.

From the data presented above, a relationship between large-scale geodynamic processes, surface structures and mineralization is clear. The Cenozoic geodynamic evolution of the Carpathian arc and of the Pannonian Basin within that arc is constrained by strong evidence of lithospheric subduction roll-back, followed by slab tear and detachment along the margins of the orogen during Neogene times. Tectonic events along the Carpathian arc itself were dominated by invasion of the Intracarpathian Alcapa and Tisia–Dacia blocks and subsequent "soft" collision with the European and Moesian plates.

Palaeomagnetic evidence shows rotation of individual crustal blocks: e.g., the Apuseni Mountains block rotated clockwise by 60° between 14 and 12 Ma whilst the Alcapa block rotated 30° anticlockwise during the same period, coinciding with the dominant mineralization. It is possible that the initiation of subduction of continental lithosphere was the underlying cause of both the magmatism and mineralization and the initiation of slab tear. Neogene calc-alkaline magmatism and associated mineralization are linked with slab tear but appear to be characteristic of normal subduction tectonics.

The data demonstrate that in all three regions, an additional heat source must have existed, possibly due to slab break-off that probably contributed much to weakening of overlying continental crust. However, the specific control was due to other factors within the crust, such as the complicated vein systems, which were responsible for upward fluid transfer. Vertical fluid transfer is particularly effective in transtensional strike-slip tectonic settings because subvertical extensional gashes open. In all three cases, a wide distribution of mineralized fault structures can be observed, which operated as fluid channels. We suggest a relatively narrow fault zone at depth and a wide zone of wrenching, which affected the regions of our ore districts. Comparison with analogue modelling results simulating wrench corridors (e.g., Mandl, 1988, 2000) can be used to explain the structural assemblage found in all three ore districts. The ore districts were controlled by upper crustal extensional structures, and the relationship of major ore districts as well as volcanism with graben structures is well established. Hydrothermal vein systems are partly oblique to the boundaries of graben structures, so that transtensional motion can be implied. The overall geometry fits well with wrench corridors, following analogue models of basement faults (Mandl, 1988). This implies subhorizontal orientations of maximum and minimum principal normal stresses σ_1 and σ_3 and vertical orientations of extensional veins. This may be an important pre-condition as this allows opening of vertical tensional fissures conducive to pronounced and effective vertical fluid transfer.

Lead isotopic data show a mantle origin of metals in mineralization associated with calc-alkaline magmas within all three districts (Table 1), with some added, variable contamination from the lower crust. This variable lower crustal contamination is particularly important in the Baia Mare and Central Slovakia districts and fits well with variable crustal contamination of hosting magmatic rocks (e.g., Mason et al., 1996; Seghedi et al., 2004).

The data compiled from different regions of the Inner Carpathians provide some additional constraints on generalized models of the geodynamic control of late-stage orogenic systems. Note that the causes of volcanic-hosted mineralization are not the same over the entire Inner Carpathians, but vary from district to district. Generalized model diagrams are compiled in Fig. 13. Tomographic and structural data show that slab roll-back and subsequent tearing during rotation may have played the prime role in the Apuseni Mountains (Fig. 13). Late-stage adakitic melts derived from a subducted lithosphere were an additional control. Magma and fluid channelling by intersection of a transtensional transform fault and a volcanic chain was the localizing factor in the Baia Mare district (Fig. 13). There, the origin of fluids in subducted mantle lithosphere was a possible source. Wrench corridor type magma and fluid channelling played an important role in the Western Carpathians. There, calc-alkaline magmatism and hydrothermal activity resulted from extension caused by slab retreat during subduction and initial soft collision of the invading Alcapa block with the European foreland (Fig. 13).

In all three well mineralized districts (South Apuseni Mts., Baia Mare and Central Slovakia districts), the presence of magmatic fluids released from shallow plutons and its mixing with meteoric water were critical for mineralization, requiring transtensional or extensional local regimes at the time of mineralization, possibly following initial compressional regimes. These three systems show that mineralization was probably controlled by the superposition of favourable mantle lithospheric conditions and partly independent, evolving upper crustal deformation conditions.

In the 13 to 11 Ma period the dominant mineralization formed all across the Carpathians, and was superimposed on those structurally favourable crustal areas where there was volcanic–hydrothermal activity at that time. The period may reflect the moment when the (upper part of the) crust failed under lithospheric extension imposed by the slab evolution. This crustal failure would have fragmented the overriding plate, may have broken up the thermal lid to provoke intensive fluid flow in selected areas, and allowed subsequent accelerated tectonic development and block rotation and extrusion of a "family of sub-blocks". These are arbitrarily regarded as the Tisia-Dacia or Alcapa blocks, even though they have lost their internal entity.

Finally, the differences between the three mineralized Inner Carpathian districts and the barren Styrian

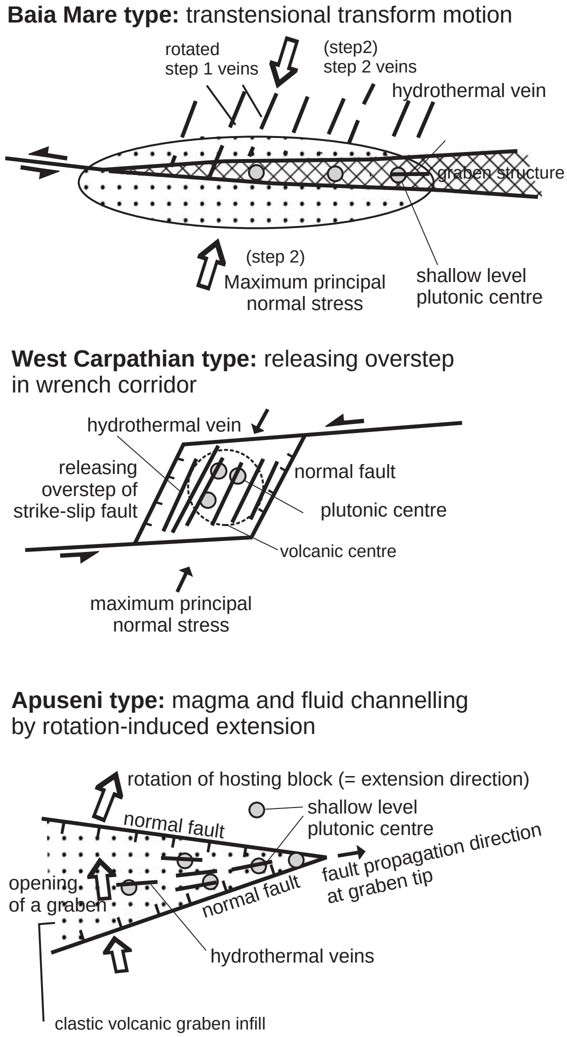

Fig. 13. Generalized models shown in plan view of the geodynamic control of late-stage orogenic systems, based on observations in the Inner Carpathians. a—Magma and fluid channelling by intersection of a transform fault and volcanic chain (type Baia Mare). b—Wrench corridor type of magma and fluid channelling (type West Carpathian). c—Slab window and extension (type Apuseni).

volcanic system are: (1) there is a lack of any kind of fault and hydrothermal vein systems in the Styrian volcanic system that are contemporaneous with or post-date emplacement of volcanic rocks, and (2) there are nearly no subvolcanic, shallow level plutons, except for a single trachytic dome. Consequently, shallow-level magma systems that could have released hydrothermal fluids were not in existence. Widespread alteration of the illite–alunite–christobalite type was probably derived mainly from meteoric water. In conclusion, the missing shallow-level plutons as well as the missing pathways could represent an explanation of why the Styrian volcanic field is barren. Similar reasoning could be applied to the scarcely mineralized Calimani–Harghita volcanic district of the Eastern Carpathians.

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