Cent. Eur. J. Geosci. • 2(3) • 2010 • 207-270
DOI: 10.2478/v10085-010-0024-5

VERSITA

Central European Journal of **Geosciences**

Neogene-Quaternary Volcanic forms in the Carpathian-Pannonian Region: a review

Review Article

Jaroslav Lexa[1*], Ioan Seghedi[2], Károly Németh[3], Alexandru Szakács[24], Vlastimil Konečný[5], Zoltán Pécskay[6], Alexandrina Fülöp[7], Marinel Kovacs[7]

1 *Geological Institute of the Slovak Academy of Sciences, Bratislava, Slovakia,*

2 *Institute of Geodynamics, Romanian Academy, Bucharest, Romania*

3 *Volcanic Risk Solutions CS-INR, Massey University, Palmerston North, NewZealand,*

4 *Sapientia University, Dept. of Environmental Sciences, Cluj-Napoca, Romania*

5 *State Geological Institute of D. Štúr, Bratislava, Slovakia*

6 *Institute of Nuclear Research, Hungarian Academy of Sciences, Debrecen, Hungary*

7 *North University, Faculty of Mineral Resources, Baia Mare, Romania*

Received 27 May 2010; accepted 18 July 2010

Abstract: Neogene to Quaternary volcanic/magmatic activity in the Carpathian-Pannonian Region (CPR) occurred between 21 and 0.1 Ma with a distinct migration in time from west to east. It shows a diverse compositional variation in response to a complex interplay of subduction with roll-back, back-arc extension, collision, slab break-off, delamination, strike-slip tectonics and microplate rotations, as well as in response to further evolution of magmas in the crustal environment by processes of differentiation, crustal contamination, anatexis and magma mixing. Since most of the primary volcanic forms have been affected by erosion, especially in areas of post-volcanic uplift, based on the level of erosion we distinguish: (1) areas eroded to the basement level, where paleovolcanic reconstruction is not possible; (2) deeply eroded volcanic forms with secondary morphology and possible paleovolcanic reconstruction; (3) eroded volcanic forms with remnants of original morphology preserved; and (4) the least eroded volcanic forms with original morphology quite well preserved. The large variety of volcanic forms present in the area can be grouped in a) monogenetic volcanoes and b) polygenetic volcanoes and their subsurface/intrusive counterparts that belong to various rock series found in the CPR such as calc-alkaline magmatic rock-types (felsic, intermediate and mafic varieties) and alkalic types including K-alkalic, shoshonitic, ultrapotassic and Na-alkalic. The following volcanic/subvolcanic forms have been identified: (i) domes, shield volcanoes, effusive cones, pyroclastic cones, stratovolcanoes and calderas with associated intrusive bodies for intermediate and basic calc-alkaline volcanism; (ii) domes, calderas and ignimbrite/ash-flow fields for felsic calc-alkaline volcanism and (iii) dome flows, shield volcanoes, maars, tuff-cone/tuff-rings, scoria-cones with or without related lava flow/field and their erosional or subsurface forms (necks/ plugs, dykes, shallow intrusions, diatreme, lava lake) for various types of K- and Na-alkalic and ultrapotassic magmatism. Finally, we provide a summary of the eruptive history and distribution of volcanic forms in the CPR using several sub-region schemes.

Keywords: Carpathians • Pannonian Basin • volcanoes • volcanic forms • Neogene • Quaternary • alkali basalts • andesites • dacites • rhyolite

1. Introduction

Volcanism in the Carpathian-Pannonian Region (CPR) was active since the early Miocene through various phases in variable geotectonic and magmatic settings. As a result the CPR today is considered to be a well-defined, but relatively small region where changes of geotectonic and magmatic situations from subduction related arc volcanism to back-arc intraplate volcanism took place in a confined region. While modern arc and back arc settings are tempting areas to pursue volcanic research that commonly have a strong natural hazard element, such young volcanic regions prevent access to cores and deep zones of volcanoes due to their young age. The CPR in this respect provides an advantage due to its mature and commonly erosion-modified volcanic landforms to see the deeper architecture of formed volcanoes as well as their succession. In addition the confined nature of the CPR provides a unique opportunity to study nearly every possible manifestation of known volcanism. Such small size is comparable with a small modern volcanic arc segment such as the Caribean or Vanuatu volcanic arc, but the CPR in the same area provides an opportunity to see not only deep into its eroded volcanic landforms' deeper structures, but also in a laterally short distance small-volume monogenetic volcanoes to mighty stratovolcanoes and extensive ignimbrite sheets likely associated with large calderas. In this paper we provide a summary of known volcanic features documented in the past decades in the CPR. Such documentation has a strong scientific value to identify key elements of volcanism in a collisional zone that was active since the early Miocene until present and which, in certain parts, is considered still active. The time aspect of understanding CPR volcanic history and its volcanism is significant due to the fact that the CPR represents a nearly completed sequence of volcanism associated with subducting lithospheric slabs, back-arc extension, and post-subduction intraplate volcanism as a reflection of gradual geotectonic changes over the last 21 million years. Such complete volcanic history of a syn- and post-collisional tectonic regime related volcanism is rare, and therefore makes the CPR a unique region that could be compared with other geotectonically similar volcanic regions.

The review of volcanic forms in the Carpathian-Pannonian Region (CPR) presented here is meant to complement previous geological, geophysical and petrological investigations. Reviews have been published treating the space/time evolution of volcanic activity [1, 2], geotectonic evolution [3] and petrology and geodynamic setting of volcanic activity [4–7]. That is the reason that these subjects are covered only in the extent necessary for a proper understanding of volcanic form presentation.

*E-mail: geoljalx@savba.sk

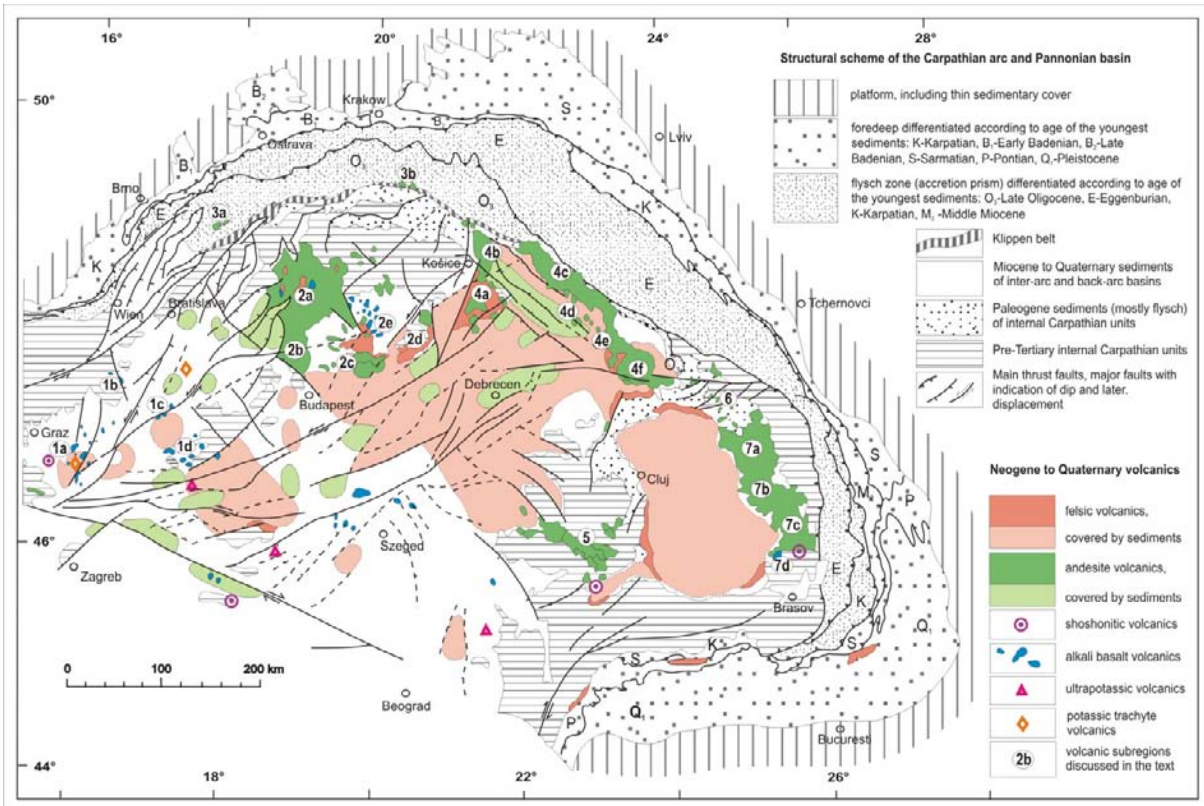

Figure 1. Distribution of volcanic formations in the Carpathian/Pannonian Region (CPR). Numbers indicate sub-regions discussed in the text. Modified after Figure 1, Pécskay et al. [2].

Knowledge of volcanic form patterns provides information about the spatial availability of magma, evolution of magma in various size chambers and ascent of magma through geologically contrasting fragments of the crust. Distribution, storage, ascent, and eruption of magmas were complex, resulting in a remarkable variety of compositions generating various volcanic forms. Naturally, volcanic forms reflect also an environment of their evolution that was mostly terrestrial and partly subaqueous (e.g. shallow marine). Further complication comes with the extent of erosion. Based on the level of erosion we distinguish: (1) areas eroded to the basement level, where paleovolcanic reconstruction is not possible; (2) deeply eroded volcanic forms with purely erosional morphology, where paleovolcanic reconstruction is still possible; (3) eroded volcanic forms with remnants of original morphology preserved and recognizable; and (4) the least eroded volcanic forms with original morphology quite well preserved.

Our presentation is based on trustworthy data for all the volcanoes in the time interval they have evolved (i.e., <21 Ma), outlining their distribution, their compositional ranges and similarities. While timing of volcanic activity is known from a map-based stratigraphy and detailed geochronology [1, 2], paleovolcanic reconstruction of volcanic forms is based on several decades of practice and collaboration. We provide characteristics of volcanic forms and a summary of the eruptive history and volcanic form distribution in the CPR using seven sub-region schemes as in Figure 1: 1 – Styermark + Burgenland + Little Hungarian Plain + Balaton Highlands; 2 – Central Slovakia Volcanic Field + Börzsöny-Visegrád-Pilis-Burda + Cserhát-Mátra + Bükk foreland + Southern Slovakia – Northern Hungary; 3 – Moravia + Pieniny; 4 – Tokaj-Zemplín + Slanské vrchy + Vihorlat-Gutin + Beregovo + Oaş + Gutâi; 5 – Apuseni; 6 – Ţibleş-Rodna; 7 – Călimani + Gurghiu + Harghita + Perşani. For each of the sub-regions we consider also succession, compositional characteristics, extent of erosion and vent distribution patterns.

2. Geotectonic setting

During Neogene time the Carpathian orogenic arc represented a continental margin or evolved island arc with variable parts of older continental crust. The arc migrated northward, northeastward and eastward due to subduction of oceanic and suboceanic crust of fore-arc flysch basins until it collided with the margin of the European Platform. The collision gradually evolved from the west to the east. The arc retreat into the area of former flysch basins was compensated by formation of inter-arc and back-arc extensional basins with two microplates involved – Alcapa and Tisia (Tisza) -Dacia [3]. This geotectonic framework and distribution of blocks of older continental crust influenced essentially spatial and time distribution of Neogene to Quaternary volcanic structures and their compositional evolution (Figure 1). Volcanic and/or intrusive activity took place since 21 Ma till 0.1 Ma with a distinct migration in time from west to east (Figure 2). Volcanic forms show diverse compositions that were generated in response to complex syn- and post-collisional tectonic processes [1, 2] involving subduction with roll-back, collision, slab break-off, delamination, strike-slip tectonics and block rotations of two microplates [4–7]. Major groups of calc-alkaline rock-types (felsic, intermediate and mafic varieties) have been distinguished and several minor alkalic types also pointed out, including K-alkalic, shoshonitic, ultrapotassic and Na-alkalic. Geodynamic and petrological models have clarified many of the petrogenetic processes [4–7]. In most cases the geochemical fingerprint of initial contributions from different magma sources are obscured by further evolution of magmas via differentiation, assimilation and mixing processes during ascent and storage in crustal magma chambers.

Volume and composition of magmas are essential factors in evolution of volcanic forms. Pécskay et al. [2] have distinguished the following groups of volcanic forms based on their spatial extent and essential compositional characteristics:

- Felsic calc-alkaline volcanic forms represented mostly by rhyolitic to dacitic welded and/or non-welded ash-flow tuffs, fallout tuffs and their reworked counterparts and minor domes and dome/flow complexes extended throughout most of the Pannonian and Transylvanian Basins, as well as at their margins. They associated with early extension and basin formation in the intra-Carpathian areas.

- Intermediate calc-alkaline volcanic forms in the intra-Carpathian region associated with advanced stage of back-arc extension; they are composed of mostly medium to high-K type andesites, comparable with andesites of continental margins. Rocks vary from basaltic andesites to rhyolites, often in the bi-modal type of suites. The most frequent volcanic forms are monogenetic volcanic cones, composite stratovolcanoes, effusive fields, dome/flow complexes and subvolcanic intrusive complexes.

- Intermediate calc-alkaline volcanic forms extending linearly along the internal side of the Carpathian orogenic arc associated with subduction and/or slab break-off. They are composed of medium-K andesites, similar to andesites of evolved island arcs and continental arcs. Their petrography varies from predominant basalts to andesites to subordinate dacites and rhyolites. The most frequent volcanic forms are composite stratovolcanoes, effusive fields, dome/flow complexes, subvolcanic intrusive complexes and monogenetic volcanic forms.

- Shoshonitic rocks are represented by rare occurrences in the western and eastern parts of the CPR. Their generation is subject to debate of petrologists. In fact, consensus has not been reached yet. Their source is ascribed to the lithospheric mantle.

- Rare K-trachytic and ultrapotassic volcanic forms occur in the south-western corner of the CPR and Little Hungarian Plain, but also in the south-eastern part in Banat (SW Romania). Their source is in the lithospheric mantle.

- Na-alkalic volcanic forms include nepheline basanites, alkali basalts and their differentiated counterparts such as nepheline tephrites, trachybasalts and trachyandesites. They are spread over most of the western CPR as isolated clusters of outcrops organized in more or less extended monogenetic volcanic fields of maars, diatremes, tuff cones, cinder/spatter cones and lava flows. Their source is both in the asthenospheric and lithospheric mantle.

3. Methodology

Analysis of volcanic forms characteristic for the Tertiary to Quaternary volcanic complexes and/or fields of the CPR is not as easy as it is in the case of active volcanoes. With the exception of the youngest ones, their original form has been modified, obscured and/or completely eliminated by erosion. Its extent is a function of time, climatic conditions and geotectonic setting that has controlled subsequent uplift or subsidence – erosion in areas of subsequent

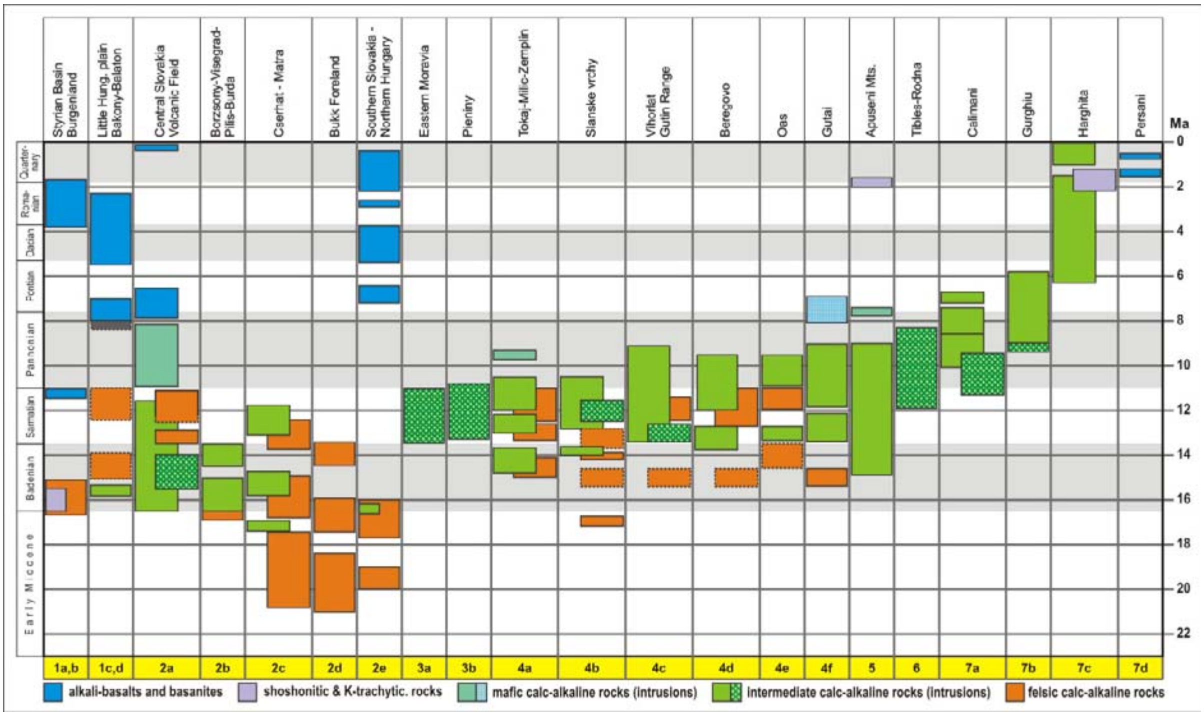

Figure 2. Timing of volcanic activity in sub-regions as shown in Figure 1. Based on Figure 2, Pécskay et al. [2].

uplift has been much faster, reaching to subvolcanic levels while subsequent subsidence has resulted in burial by younger sedimentary deposits. As a consequence, morphological aspects could be used to define volcanic forms only in the case of Quaternary volcanoes and partially in the case of the Pliocene and Late Miocene ones. Instead, our analysis of volcanic forms is based on systematic geological mapping accompanied by facies analysis and paleovolcanic reconstruction.

Most of the areas of Tertiary to Quaternary volcanic forms in the CPR have been mapped in the scale of 1:25 000. These maps accompany open-file reports in archives of geological surveys while published maps have been reduced to the scale of 1:50 000. In the course of detailed geological mapping the approach has changed and, beside rocks, petrography recognition of genetic types of volcaniclastic rocks and principles of facies analysis were introduced. Principles of the facies analysis of volcanic forms and their application in geological mapping had been developed more than 50 years ago [9] and subsequently applied successfully to the Gutin volcanic range in Transcarpathia (Ukrainian part of the Carpathian volcanic arc) [10]. These principles were applied in lithostratigraphy, geological mapping and paleovolcanic reconstruction of the central Slovakia Volcanic Field [11, 12]. Subsequently they have been applied also in other areas of CPR, especially by authors of this paper. Naturally, principles of facies analyses developed by Maleev [9, 13] have been further elaborated along with advances in classification and understanding of volcaniclastic rocks [13–22] (and relevant references in these textbooks).

Detailed geological mapping with recognition of volcaniclastic rock facies has served as a basis of facies analysis. Volcaniclastic rocks have been assigned to genetic types (type of eruption, mobilization, transport and deposition of material) on the basis of their composition, degree of sorting, shape, texture and other characteristics of fragments. Subsequently, facies associations were used to interpret environment of deposition and eruption styles of the former volcanoes.

Under the term “paleovolcanic reconstruction” we understand reconstruction of the former volcano and its evolution, including types of volcanic eruptions and processes that modeled its form. Individual facies and types of fragments point to the type of eruption, mobilization, transport and deposition. Facies associations point to the eruption styles of the volcano and enable its division into central, proximal, medial and distal zones. Finally, facies successions imply evolution of the volcano. Paleovolcanic reconstruction is problematic or even impossible in the case of deeply eroded volcanoes.

Timing is an important aspect of paleovolcanic reconstruction. It puts limits on time needed to build up the studied volcanoes and provides information on time of their exposure to processes of erosion. Timing also elucidates migration of volcanic activity in time and space (Figure 2). Timing of individual volcanoes and volcanic complexes and/or fields in the CPR has been evaluated by Pécskay et al. [2] using more than 2 000 K/Ar datings of volcanic rocks, providing reference to original papers where results are interpreted in the context of succession and relationship to biostratigraphically dated sedimentary rocks (Figure 2). We shall refer to these data where appropriate in discussion of volcanic forms and/or evolution of volcanic fields and sub-regions.

Petrology of volcanic rocks involved in volcanic structures is beyond the scope of this paper. However, evolution of volcanoes and volcanic forms, respectively, has been influenced heavily by composition of magmas and their evolution in the course of volcanic activity. Attention is paid also to this aspect in discussion of volcanic forms and/or evolution of volcanic fields and sub-regions. General information concerning petrology of volcanic rocks in the CPR and their compositional variability has been widely reported recently [5–7, 23–27].

4. Terminology

In the case of most volcanological terms, a consensus has been reached and we follow terminology used by prominent authors of textbooks and reviews [13, 14, 16–22, 28–33]. As far as volcanic forms are concerned, such a consensus has not been always reached with those authors, and we shall define each type of volcanic form before proceeding with its characterization in the CPR. However, there are a few terms related to classification of volcanic forms that need to be clarified here, before their systematic treatment.

Volcanic form: The term has been introduced to the volcanological literature by Cotton [34] in his famous book "Volcanoes as landscape forms" where it refers to the volcano morphology as the landscape form. Schmincke [21] avoided this term and used the term "volcanic edifices" instead. However, this term is limited to constructional volcanic forms [35]. Francis [32] defines volcanic form as a landscape form resulting from both constructive volcanic processes and destructive volcanic and/or erosional processes. A scoria cone for instance is a result of constructive volcanic processes providing pyroclastic and effusive material to form a scoria cone (edifice) and tephra blanket and/or a lava field around the cone. In contrast, a maar results from dominantly "destructive" volcanic processes forming a subsidence crater surrounded by a tephra rim and distal tephra blankets. A volcanic neck results from a long lasting erosion of a formerly constructive volcanic feature. For mature stratovolcanoes, both constructive and destructive processes occur, shaping the volcanic cone which is striving to reach a steady state profile [35]. Volcanic forms are therefore not understood here as purely surficial features, and thus they are not synonyms for volcanic landforms. Instead, they involve lithologic content and internal structure as well. For instance, comagmatic intrusive bodies are usually considered as their part or, when the volcano is deeply eroded, as special types of "volcanic forms". Unlike Cotton [34] and Francis [36], we use "volcanic form" here in a broad sense to express that a volcano of any type consists of a surface feature and its associated sub-surface components, both having their own morphological, compositional and structural characteristics. The usage of this term allows characterization of old volcanoes that are today erosionally dissected, forming surface expressions (e.g. hills) that are strongly different from the original volcanic landforms. It also includes the exhumed subsurface parts (core/vent structures) of various volcanic landforms. Therefore, "volcanic form" is adequate to denominate with a common term all volcanic features and volcanism-related subsurface features exposed at the surface at the present time in the CPR as a result of a long-lasting (21 Ma) volcanic and erosional history.

According to their eruptive history, volcanic forms can basically be classified as monogenetic and polygenetic.

1. **Monogenetic volcanic forms** (or monogenetic volcanoes) are defined as volcanoes formed during single events (eruptions) of short-lived (days to years) volcanic activity, without subsequent eruptions [18, 32, 37–39]. Such a definition fits especially volcanic forms found in basaltic monogenetic volcanic fields and satellite edifices distributed on flanks of larger composite volcanoes. However, the same definition is applicable also to volcanic forms made of intermediate and felsic lavas – monogenetic volcanic cones of explosive, stratovolcanic or effusive type, solitary extrusive domes and solitary intrusions. Evolution of basic to intermediate composition volcanic cones may take a longer time and the best criterion to distinguish between monogenetic and polygenetic ones might be the presence or absence of a significant period of erosion implying a longer lasting break in volcanic activity. Here we slightly differ from the concepts of Francis [32], Connor and Conway [37] and Walker [40] that use a long lasting eruptive activity and larger size of these cones as a sole argument to classify them as composite or polygenetic volcanic forms, respectively. Those simple cones of Francis [32] under the heading "polygenetic volcanoes" that resulted from prolonged volcanic activity without a significant break fulfill the definition of monogenetic vol-

canoes. As an example, we can quote the volcano Izalco in El Salvador whose ca. 650 m high cone of the stratovolcanic type has been built up during almost continuous eruption of the Strombolian type lasting 196 years between 1770 and 1966 A.D [41, 42].

A number of specific monogenetic volcanic forms occur in the CPR, as presented below.

2. **Polygenetic volcanic forms** (polygenetic volcanoes) have experienced more than one, usually many, eruptive events during their history [32]. They usually associate with a long lasting and complex eruption history of a volcanic system [35]. Such volcanic systems can either consist of a single volcanic edifice, or nested volcanic edifices. Distinction from monogenetic volcanoes is rather conventional. However, many of the relatively small-volume volcanoes in a basaltic volcanic field can exhibit very complex volcanic architectures, and in many cases their eruption history shows signs of multiple and recurring activity over long periods of time [43]. The transition between small-volume but complex basaltic volcanoes and long-lived but relatively small-volume composite volcanoes is rather gradual [22, 44]. Among polygenetic volcanoes Francis [32] has distinguished simple cones, composite cones, compound volcanoes and volcano complexes. Shield volcanoes are mostly also polygenetic. Polygenetic volcanoes with upward-concave outer slopes are commonly referred to as stratovolcanoes.

In turn, polygenetic volcanoes can further be classified in composite and compound types.

2a. **Composite volcanic forms** (or composite volcanoes) have had more than one evolutionary stage in their existence, but the locus of volcanic activity has been essentially confined to a single vent or a group of closely-spaced central vents, hence the volcanoes still retain a simple conical or shield-like overall symmetry [32]. According to Davidson and DeSilva [35], they are products of multiple eruptions spanning tens to hundreds of thousand of years.

2b. **Compound volcanic forms** (or compound volcanoes) are polygenetic volcanoes that have formed from coalesced products of multiple, closely spaced vents [35]. As a result, they are characterized by a more complex landform that cannot be approximated by a simple conical morphology. Cotton [34] and MacDonald [28] have used the term "multiple volcano" for the same type of volcanic form.

A large variety of polygenetic volcanic forms has been described in the CPR. Their further classification is based on the prevailing lithology reflecting the type of dominant eruptive activity (explosive or effusive). Further classification criteria includes the type of activity at the end of volcano history, different from the dominant activity-type and the occurrence of major events during eruptive history (i.e. caldera formation, edifice collapse).

Intrusive forms (intravolcanic intrusions, subvolcanic intrusions, hypabyssal intrusions) are subsurface-solidified magma bodies that have been emplaced in an environment of older rocks [45] in both the pre-volcanic basement and the surface volcanic edifice itself. Some of them might have reached the surface (feeder dikes and necks). They occur as solitary bodies (equivalents of monogenetic volcanic forms) or in groups. Intrusions can be found associated with all types of volcanoes. However, their presence is more characteristic for the central zones of polygenetic volcanoes. Intrusions are classified as "intravolcanic" if they are emplaced in comagmatic volcanic products and as "subvolcanic" if they are emplaced in basement rocks of the volcano. They are found at surface when exhumed or in subsurface mining works or drill-holes in volcanic areas of the CPR. However, intrusions occur also in areas of Neogene magmatism in the CPR without being associated with surface volcanic activity. The term "hypabyssal intrusions" is more appropriate in those cases. Because we address here the Neogene volcanism in the CPR as a whole, we consider all of these types of intrusions as special types of "volcanic forms", whether associated with volcanic forms having a surface expression or not. In the case of intrusions resulting from a single datable intrusive event, they are considered as being equivalent with a monogenetic volcanic form and are described in the section addressing such forms.

Fields of volcanic forms (or volcanic fields). Individual volcanic forms generally occur in groups forming volcanic fields. Volcanic fields can be characterized and named according to the prevailing volcanic forms they include. Monogenetic volcanic fields are composed of mostly monogenetic volcanic forms and they can be described using shape, size, and volcano density as parameters. Commonly they show an areal extent. Polygenetic volcanic fields (volcanic complexes according to Francis [32]) consist of spatially, temporally, and genetically related major and minor volcanic edifices with their associated lava flows and volcaniclastic rocks. Frequently they consist of aligned volcanic forms organized in volcanic ranges (or "chains", as traditionally used in the East Carpathians) forming a part of the Carpathian volcanic arc. Longer volcanic ranges (e.g. the Călimani-Gurghiu-Harghita volcanic chain – no. 7 in Figure 1) can be subdivided in segments (e.g. Călimani, Gurghiu and Harghita segments – no. 7a, 7b and 7c in Figure 1). However, polygenetic fields of areal extent also occur in the back-arc region (e.g. Central Slovakia Volcanic Field, Börzsöny-Visegrád-Pilis-Burda volcanic field, Apuseni volcanic field – no. 2a, 2b and 5 respectively in Figure 1).

One special type of volcanic field is that in which the dispersion and accumulation of volcanic products is controlled by a structural-morphological feature, such as a graben, instead of being controlled by individual central volcanoes so that individual volcanic sources are either located (and possibly deeply eroded) outside the field considered or cannot be readily identified. Such fields are commonly named in the relevant literature as volcano-tectonic depressions (or grabens). They can be subdivided into subtypes according to the prevailing lithology of the volcanic fill of the depression (lavas, volcaniclastic rocks or mixed).

Assemblages of individual intrusions that cannot be associated with volcanic fields are termed "intrusive complexes" and they can be viewed in analogy with volcanic fields as "intrusive fields".

5. Volcanic forms of the Carpathian-Pannonian Region (CPR)

This part of our paper is devoted to a systematic description of volcanic forms occurring in the CPR. Volcanic forms are highly variable in size, age, extent of denudation, rock types, composition of lavas and other properties.

In the first step we divide volcanoes (volcanic forms) into monogenetic and polygenetic ones using criteria already mentioned in the terminology part. Assignment of volcanic forms to one of these categories is not always straightforward.

In the second step we use a division based on essential compositional characteristics (e.g. Na-alkalic or basic, intermediate, felsic calc-alkaline). Such a division has two reasons: (1) it corresponds to the first order lithostratigraphic division of volcanic products; and (2) composition of magma stands behind contrasting eruption styles and different volcanic form assemblages.

5.1. Monogenetic volcanoes

In the CPR, monogenetic volcanoes are characteristic especially for alkali basalt monogenetic volcanic fields and sporadic occurrences of ultrapotassic rocks where maars, tuff-rings, tuff-cones, scoria-cones with related lava flows and flow fields, necks and/or plugs, dikes and sills, diatremes and rare small shield and dome-flows have been recognized.

Monogenetic volcanoes associated with calc-alkaline basaltic andesite and andesite magmas are represented by solitary domes, dome-flows, tuff cones, small simple cones, dikes, sills, laccoliths, irregular protrusions and stocks. Surficial forms occur usually at parasitic vents of mature stratovolcanoes and in areas of bimodal andesite/rhyolite volcanic activity with multiple vents. Solitary intrusions may associate with parasitic vents of andesite stratovolcanoes; however, they are characteristic especially for strongly uplifted and deeply denudated areas, as the zone where Carpathian volcanic arc overlaps with the internal zone of the accretion prism.

Calc-alkaline felsic extrusive lava domes and dome-flows with associated pyroclastic rocks as well as minor intrusions could form fields of dispersed monogenetic volcanic forms similar as the alkali basalt volcanic fields.

It is important to note that many of the described monogenetic volcanic forms, if incorporated into the structure of polygenetic volcanoes, represent their integral part.

5.1.1. Maars

Maars are small, so called "negative" volcanoes, and form a "hole-in-the-ground" morphology (Figure 3) [39, 46–48]. The formation of a maar volcano is directly related with some sort of subsidence generated by mass-deficit beneath the surface [49–51] as a consequence of explosive interaction between hot magma and water and/or water saturated sediments [52, 53].

Maar volcanoes, gradually subside long after their formation as a result of resettling of the water-soaked debris accumulated in the volcanic conduit beneath the crater that is commonly referred to as a diatreme [54, 55].

Maar volcanoes are also unique sedimentary traps [39, 56]. Their crater can act as a deep and sheltered sedimentary micro-basin that collects sediments from a continental catchment area [57].

5.1.2. Maars related to Na-alkalic rocks

Maar volcanoes in the Pannonian Basin are common [58] due to a long lasting production of Na-alkalic mafic magmas through the Late Miocene to Pleistocene time [1, 5, 24, 25, 27, 59]. Such intraplate, intracontinental magmatism released mafic magma along weakness zones (e.g. commonly along deep and old structural weakness zones of the basement) that encountered a shallow subsurface zone of water-saturated filled sedimentary basins in the region [5]. The various thickness of the shallow marine and fluvio-lacustrine siliciclastic Neogene sedimentary infill [60–65] over the fractured, commonly karstic Mesozoic and Paleozoic carbonatic and metamorphic rock units [66–68], provided a combined aquifer that the rising mafic magma interacted with. As a result, the Pannonian Basin was a playground of phreatomagmatic volcanism through the Late Miocene to Pleistocene [58, 69, 70]. Maar volcanoes are formed by both soft-and-hard-substrate environment, and as a result generated various maars. In areas where the water-saturated soft-substrate infill was thick (e.g. Little Hungarian Plain Volcanic Field–Figures 1–

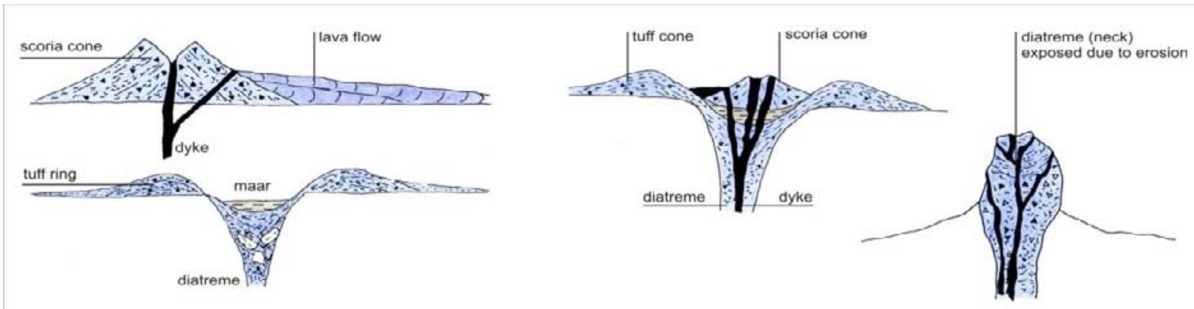

Figure 3. Volcanic forms of alkali basalts.

1c, 9), broad maars formed, that are inferred to be transitional to tuff rings [58]. In areas where hard-substrate aquifers were closer to the surface, deep maars formed [58].

Recognition of maar volcanoes in the CPR is however not always easy. Due to the fast erosion of the crater rim deposits, maars that preserved their original morphology are relatively rare in the CPR. Maars with recognizable surface morphology are common in southern Slovakia (Figure 1-2e), such as the Hajnáčka, Jelšovec and Pinciná maars (Figure 4) that also act as a major accumulation sites of maar lake sediments [71–76]. Evidence for the presence of former maar volcanism and existence of maar volcanoes is abundant in the CPR especially due to the great exposures of diatremes [58] representing the subsurface part of a volcano (Figure 3), correctly referred to as a maar-diatreme volcano [77, 78]. It seems that much evidence points to the fact that the phreatomagmatic volcanic fields, including their maars, were probably dominated by soft-substrate aquifer controlled, shallow and broad maars with occasional mixed aquifer controlled maars that formed over fracture-controlled aquifers. Another paper in this volume presents some detailed account on the potential analogy for the Pannonian Basin phreatomagmatic volcanoes [79].

5.2. Tuff-rings / tuff-cones

Tuff rings and tuff cones (Figure 3) are also phreatomagmatic monogenetic volcanic forms [80]. They represent end members of a spectrum where magma interacts with an increasing abundance of external water [80, 81]. In the case where the magma and water interaction is controlled by shallow surface water, the resulting volcanic landform will be a tuff ring [51]. It seems that there is a gradual transition among maars and tuff rings, and in many cases, the distinction between these two landforms cannot be drawn easily. Sensu stricto, maars should have a crater floor that has subsided below the syn-eruptive surface, and therefore the pre-volcanic rock units must crop out in the crater wall [51]. In the case of a tuff ring formation, such subsidence is not expected, due to the shallow surface magma and water interaction that cannot create significant mass deficit beneath the surface and cause collapse of the crater floor. Tuff cones are clearly constructional volcanic landforms and commonly associated with hydromagmatic Surtseyan-style emergent volcanism in shallow submarine or sub-lacustrine settings. The abundance of the water in the surface results in an eruptive condition which is wet, and potentially an eruptive environment which is at least partially subaqueous [82, 83]. The dump conditions over the vent and the abundant free condensed water in the deposition surfaces provide a deposition environment of the pyroclasts that is very unique. Textural features of soft-sediment movements such as oversteepened and plastered beds, mud cracks, and plastic deformation features beneath larger pyroclasts are among key features tuff cones can preserve in their sedimentary record [84–86]. Before a subaqueous volcano emerges above the water table, the accumulating pyroclastic deposits will accumulate on a subhorizontal depositional surface from a radially moving pyroclastic density currents. A pyroclastic mound with subhorizontal bedding and characteristic wet depositional features will be later buried by steeply dipping beds to complete the evolution of the tuff cone [82, 87].

5.2.1. *Tuff-rings / tuff-cones related to Na-alkalic rocks*

In older phreatomagmatic volcanic fields such as many in the CPR area, the distinction of such forms is not straightforward. The above mentioned process was likely operating in many cases of the CPR monogenetic fields, especially in those areas where the thick Neogene sedimentary infill provided substantial water saturated sediments to fuel phreatomagmatism. Due to the advanced stages of erosion in many CPR monogenetic volcanic fields, the preserved volcanic landforms nearly in each case represent proximal facies, the core of a phreatomagmatic vol-

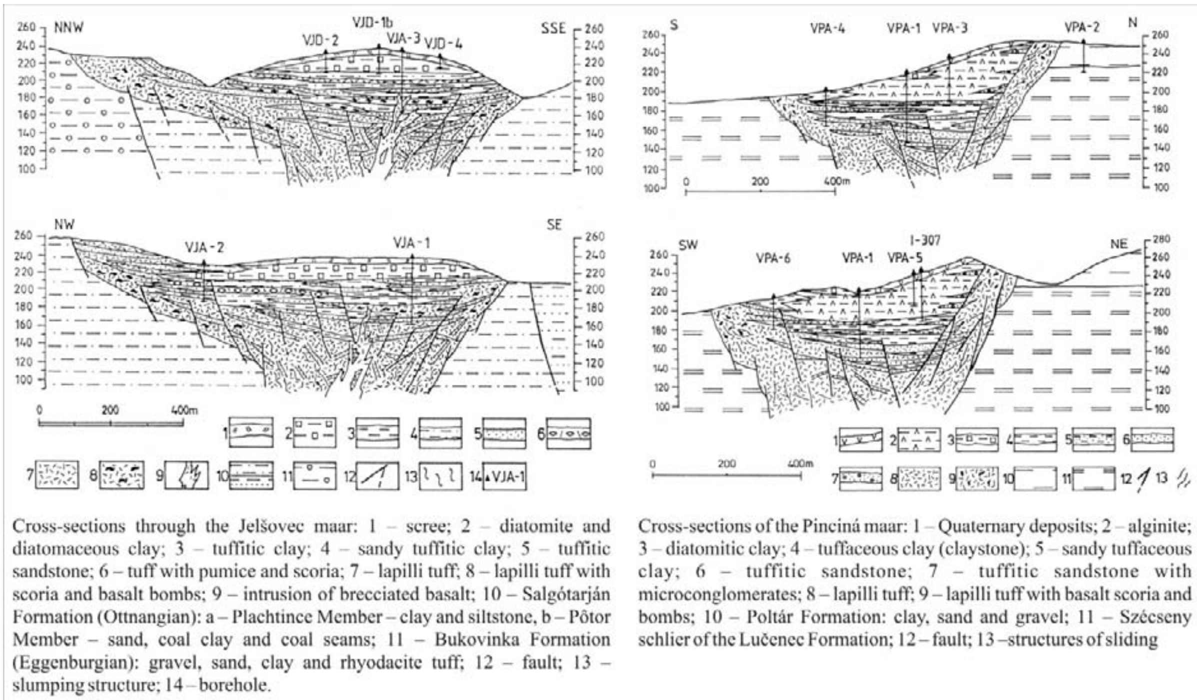

Figure 4. Sections of the maars Jelšovec and Pinciná in southern Slovakia – examples of maars with diatomite resp. alginite deposits of maar lakes with no subsequent volcanic activity [74, 358].

cano without definitive reference points outside of the preserved core of the volcano [58]. This makes it highly speculative to distinguish between shallow soft-substrate controlled maars from tuff rings in the CPR area. While this problem still persists, the textural study of the preserved pyroclastic rocks can help to define the preserved volcanic landforms. Pyroclastic rocks of maars in any substrate environment are expected to be rich in accidental lithic fragments. Their origin clearly represents the depth of explosion locus. In case of the explosion locus in the shallow soft-substrate, the pyroclastic rocks then should contain a high proportion of fragments derived from these rock units. In the case of such textural characteristics and other field evidences of potentially shallow craters and broad, lensoidal shape volcanic landforms, we can interpret it as an erosional remnant of a shallow maar. While the same 3D architecture of a preserved volcanic landform with pyroclastic rock units is rich in glassy juvenile pyroclasts, we can establish it as an erosional remnant of tuff rings [88].

It seems that tuff rings are more common in the Little Hungarian Plain Volcanic Field (Figure 1-1c), Styermark (Figure 1-1a), and Slovenia [58, 88] than in the Bakony-Balaton Highland Volcanic Field (BBHVF, Figure 1-1d), where shallow maars prevail [58]. Either way, the fact that the majority of the pyroclasts are glassy juvenile particles, and that the accidental lithic fragments are derived from the immediate underlying pre-volcanic soft-substrate suggests that the magma fragmentation level in these volcanoes of the CPR was shallow.

Recognition of tuff cones in old eroded terrains such as the CPR is not easy. Key features for reconstructing a tuff cone origin are the evidence to support the volcanic erosional remnant as a constructional landform, the pyroclastic rocks, rich in chilled glassy pyroclasts, a very small proportion of accidental lithic fragments and abundant evidence of soft sediment movement and water saturated depositional environments. A general link between subaqueous (lacustrine, shallow marine) non-volcanic sedimentary environment and coeval phreatomagmatic volcanic successions can also help to establish the eruptive environment.

In the CPR area, tuff cones have been identified from southern Slovakia (Figure 5) and some indications of tuff cones known from the Steyermark and Slovenian monogenetic volcanic fields [58, 75, 89]. It is unlikely that tuff cones were formed in the BBHVF. Phreatomagmatic volcanoes of the LHPVF are also more consistent with their tuff ring origin.

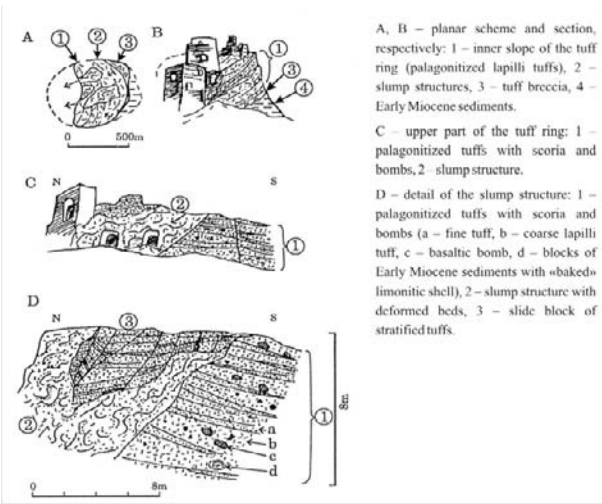

Figure 5. Castle hill in Fiľakovo, southern Slovakia – remnants of a tuff-cone with exposed internally dipping succession [359].

5.2.2. Tuff-cones related to intermediate calc-alkaline rocks

As in the case of monogenetic alkali basalt volcanic fields (see above), tuff-cones associated with andesitic volcanic activity are products of phreatomagmatic eruptions. Remnants of two eroded tuff-cones have been identified during geological mapping of volcanic fields in Slovakia.

One of the tuff cones is close to Košický Klečenov in Slanské vrchy (Figure 1-4b) and may represent a parasitic vent of the Bogota or Strechov andesite stratovolcano of Sarmatian age [90, 91]. A flat lava topped hill, remnant of a former lava lake, shows at marginal slopes inward dipping stratified palagonite tuffs of proximal facies with frequent blocks, some of them creating bomb-sags. Bogota and Strechov volcanoes initiated their evolution in a shallow marine environment and the tuff-cone represents a product of a Surtseyan type eruption.

The second one is close to the town Žiar nad Hronom in the Central Slovakia Volcanic Field (Figures 1-2a, 6) and is related to activity of basaltic andesites during the early Pannonian time [92, 93]. It is preserved only its northern part, the rest being eroded and covered by alluvial deposits of the river Hron. The tuff cone is formed of stratified and sorted fine to coarse palagonite tuffs with a variable proportion of juvenile basaltic bombs and lapilli as well as accidental lithic fragments predominantly from the shallow underlying lithologies (mostly pebbles of various andesites, quartzite, quartz, gneiss and limestone). The volcanic successions are characteristic of wet surge deposits alternating with fall deposits. Notable are the impact sags of bombs and large lithic blocks. This suggests phreatomagmatic eruptions initiated by interaction of ascending magma with water in the sand and gravel horizons of the Middle to late Sarmatian volcano-sedimentary complex of the Žiarska kotlina basin. The volcano evolved in two stages: the initial phreatomagmatic stage forming the tuff cone was followed by the effusive stage, generating lava flow and corresponding feeder-dikes.

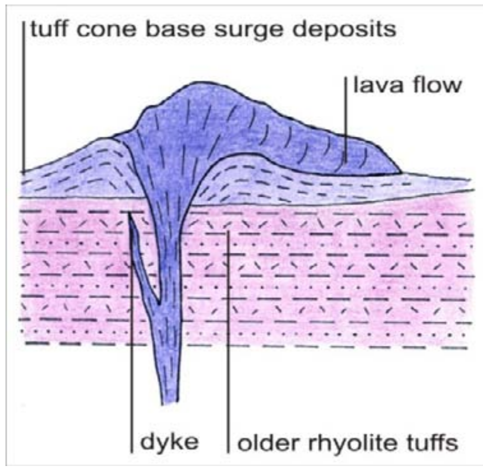

Figure 6. Paleovolcanic reconstruction of the calc-alkaline basalt monogenetic volcano Šibeničný vrch in the Central Slovakia Volcanic Field [219].

5.3. Scoria-cones with or without related lava flows

Scoria cones are rather simple and small-volume volcanic forms having an edifice volume of a cone around 0.05 km^3, a few hundred meters in basal diameter, and rarely exceeding 400 meter elevation from their base [94]. Scoria cones can be (1) as parasitic cones on larger polygenetic volcanoes or (2) as parts of monogenetic volcanic fields (Figure 3).

The “simple” scoria cone is characterized by a conical structure with a bowl-shaped crater at the top [95]. The scoria cone growth largely invokes models of emplacement and accumulation of pyroclasts through ballistic transportation (no-drag) [96]. The majority of the scoria cone growth models operate with the transportation and accumulation of few cm to few dm sized (scoria or cinder) pyroclastic fragments [95, 96] that behave in a granular fashion after deposition [97, 98].

The eruption mechanism of scoria cones is dominated by magmatic fragmentation which is generally governed by the speed of the rising magma, which determines the viscosity and the gas content of the magma [99, 100] producing Hawaiian [101, 102] and Strombolian-style eruptions [38, 103]. In general, the Strombolian-style magma fragmentation occurs at shallow depth (upper part of the magma column), when the over-pressurized and segregated gas-bubbles break and burst out from the rising, mostly mafic, magmas [38, 103–105]. In spite of this general trend of scoria cone eruption and growth, scoria cones are far more complex volcanoes [98], and evolve through various eruption styles from Hawaiian lava fountaining [106], violent Strombolian [107] to intermittent phreatomagmatic eruptions such as maar-forming [104, 108] can take place during their relatively short lived (days to years) eruption period [109]. As a result, the shape of the scoria cone will depend on these variations of the eruption style.

In addition, scoria cones can be active over years [38, 109–111], commonly forming fissure-controlled chains of cones [112, 113], and producing relatively large-volume complex volcanic edifices that clearly document polycyclic, or even polymagmatic eruptive nature [38].

5.3.1. Scoria-cones related to Na-alkalic rocks

Scoria cones in the CPR are moderately to strongly eroded volcanic forms [58, 114]. There is no scoria cone known to be preserved in its nearly original morphological features. They are heavily vegetated, covered, and in many cases their original morphology is obscured by post-volcanic mass movements that are not directly related to volcanic processes. There are a few relatively well-preserved scoria cones in the BBHVF [114] as well as in southern Slovakia (Figures 1-2e, 7) [75, 76, 115, 116] and in the Perşani Mts., Romania (Figure 1-7d) [117]. These cones still can be recognized, in spite of the significant slope angle reduction from the theoretical 30 degrees to about 15 degrees. These cones still retain some recognizable crater morphology [114]. The preserved volcanic volumes of the scoria cones of the CPR area are rather small, and the potential original size of the cones are estimated to fall in the small end of the cone volume spectrum.

Scoria cones are also point sources of extensive lava flows in many fields (Figures 7, 8) [118–122]. The longest preserved lava flow in the Southern Slovakia – Northern Hungary monogenetic volcanic field that were likely sourced from a small cone is less than 10 km [75]. With this lava flow length, this eruption would mark an average lava flow length, typical for many scoria cone-dominated volcanic fields. In spite of these preserved elongated lava flows, the erosion level due to the relatively old age of the cones prevent finding widespread lava flow fields composed of linear flow channels. In the northern Hungarian and western Pannonian Basin monogenetic volcanic fields, it has been inferred that many lava-capped buttes represent lava lakes that were confined between tephra rings, and that occasional overspill of lava initiated some lava flows [43]. These lava flows were then likely to form lava fields in the lowlands. While this model seems to be logical and expected in cone-dominated fields, there is little evidence to support such a scenario. Large lava dominated shield-like volcanoes from the BBHVF and Southern Slovakia are still relatively small in volume in comparison to other similar lava shields in central Mexico, Argentina, or northern Africa. At this stage there is not enough geological evidence to draw a better picture of the lava field evolution associated with the monogenetic fields in the CPR region.

Extensive lava fields, commonly associated with lava shield volcanoes in the CPR, preserved very little evidence on their original surface features. Lava plateau surfaces however occasionally exhibit ropy lava surface textures such as on the Fekete-hegy in the BBHVF (Figures 1-1d, 9). These lava flow surfaces indicate pahoe-hoe lava flow origin, however, they are not common. This either would mean that they are not preserved or the lava flows are dominantly not pahoe-hoe types. Lava flow surfaces in the BBHVF and in the northern Hungarian monogenetic volcanic fields resemble angular block-dominated, rugged surface morphology typical for aa lava fields, thus the lava flow fields associated with monogenetic volcanic fields in the CPRarea are likely rather aa or blocky types. The presence of scoria cones in the CPR area can also be inferred on the basis of preserved thin veneers of scoriaceous ash and lapilli. Lava capped buttes are commonly hosting pyroclastic rock units that are strongly welded, agglutinated, and gradually transformed into coarse grained scoria lapilli succession. This type of preserved rock is inferred to be an erosional remnant of a core of a scoria cone. It can be inferred that many of the CPR necks and plugs are deeply eroded cores of former scoria cones.

5.3.2. Scoria cones related to intermediate calc-alkaline rocks

In one case we have identified remnants of such a cone in the northern part of the Tokaj Mts. at the Hungary/Slovakia boundary (Figure 1-4a/b). The cone is now dissected by a creek and its internal parts are exposed to observation. It rests on shallow marine clays of the Sarmatian age and its lowermost part is formed of alternating fine to coarse grained tuff and lapilli tuff horizons formed by initial hydromagmatic explosive eruptions. The subsequent Strombolian type eruptions built a 100 m high scoria cone, being also the point-source of subsequent aa-to block-type lava flows.

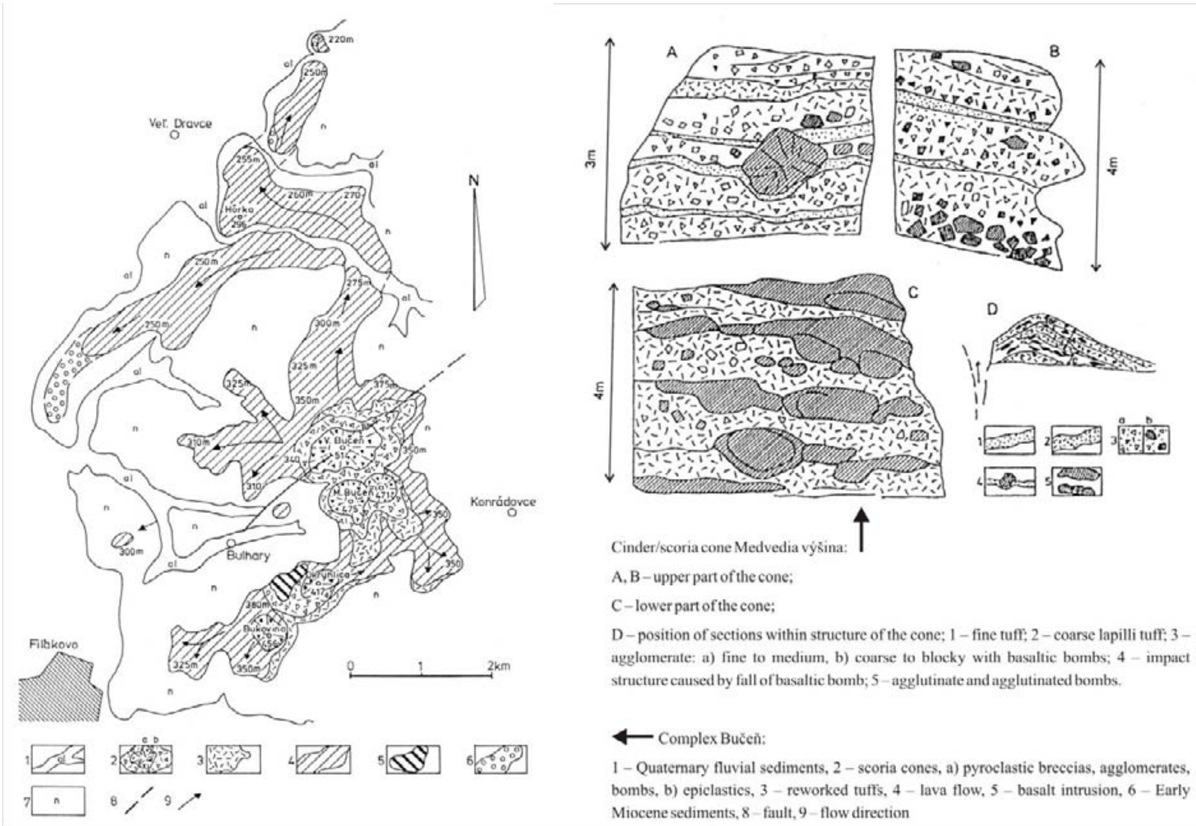

Figure 7. Left: complex of cinder/scoria cones Bučeň in southern Slovakia and related lava flows; right: details of the scoria cone Medvedia výšina at the Slovakia/Hungarian boundary [359].

5.4. Domes / Dome-flows

Lava domes are mounds of volcanic rocks that form and amass on the surface over a vent [123] with steep slopes extruded from and accumulated above the volcanic vent, while dome-flows are defined as domes that have experienced some flow away from the vent [124–127]. They form as viscous magma cools following its extrusion onto the surface. The form of lava domes varies with viscosity from tholoids and spines (upheaved plugs) surrounded by crumble breccias in the case of extreme viscosity to conical domes with spines of the Peléean type covered by crumble breccias in the case of high viscosity and to domes and dome-flows or coulees in the case of medium to lower viscosity (Figure 10) [36, 126, 128]. Their final shape is a function of lava viscosity, volume and time of extrusion. They may have a steady or episodic growth with emplacement times ranging from hours to many decades [125]. There are subaerial and subaqueous domes; in a subaqueous environment, the outer shell generates hyaloclastites, a term applied to glassy volcaniclastic rocks formed by non-explosive fragmentation in the presence of water (e.g. Batiza and White, [129]).

5.4.1. Dome/Dome-flows related to K-alkalic rocks

A few simple small-sized monogenetic volcanic forms displaying a flat dome morphology, with the aspect ratio (height/diameter ratio) different from common taller volcanic domes of felsic rocks with higher aspect ratio, occur in the South Harghita Mts. (Bixad and Luget)(Figure 1-7c) and in the southern part of the Apuseni Mts. (Uroiu) in Romania (Figure 1-5). They show a wide flat top without an obvious crater depression. The basal outline is quasi-circular or slightly elongated, suggesting the dome-flow type of volcanic form. They resulted from a single extrusion event of moderately viscous alkaline magma of shoshonitic composition and of similar age (Pleistocene). They apparently were controlled by an east-west striking South Transylvanian Fault system [130]. The lack of breccia envelops around the actual massive rock core in all these cases, leaving their nature open to alternative interpretations, namely that they might represent shallow unroofed intrusions. However, their frequent porous texture and the presence of fumarole-stage minerals in the

Figure 8. Volcanic forms of alkali basalt monogenetic volcanic field in Southern Slovakia – Northern Hungary [363]. Classification of volcanic forms in northern Hungary is not available yet.

pores are arguments for their extrusive origin.

5.4.2. Dome/Dome-flows related to intermediate calc-alkaline rocks

Isolated simple-shaped or compound lava domes occur in places in the Harghita Mts. range and at the north-eastern periphery of South Harghita Mts. (East Carpathians) as the expression of rare monogenetic activity of calc-alkaline magmas (Figures 1-7c, 11-7c). Şumuleu-Ciuc is one of the relevant examples. Its overall conical dome shape and almost circular map outline (ca. 1 km in diameter) is complicated by lateral bulges suggesting a multiphase extrusion history during its extrusion. It is composed of amphibole-biotite andesite. Remnants of a former breccia envelope or of a crumble/talus breccia apron are mapped sporadically at the lower peripheries of the dome.

A similar-sized and-shaped solitary dome – Murgul Mare – was emplaced at the south-western periphery of the

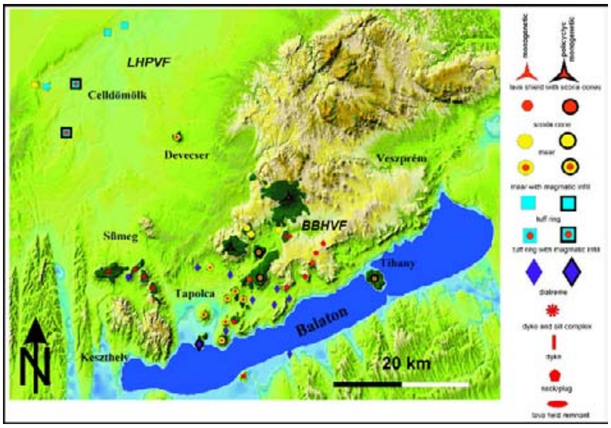

Figure 9. Volcanic forms of the Bakony-Balaton Highlands alkali basalt monogenetic volcanic field, Hungary.

South Harghita Mts. in the vicinity of two shoshonitic domes (Figure 1-7c) to which it strongly contrasts by its higher height/diameter ratio. Its simple conical shape suggests an extrusion event developed through the central vent during one single eruptive event. No breccia occurrence has been reported so far related to the dome suggesting a primary absence or erosion of the presumed clastic envelope of the dome. Its composition is high-K calc-alkaline amphibole-biotite dacite.

The two domes – Şumuleu-Ciuc and Murgul Mare – share their particular peripheral position with respect to the main volcanic range axis. They are the result of a small-volume, unique effusive eruption of viscous volatile-poor magma. None of these domes preserved any summit features suggesting crater depressions.

Solitary extrusive domes of andesite to rhyolite composition (Figure 12) occur in the Tokaj, Zemplín, Beregovo and Oaş areas (Figures 1, 11 – areas 4a, 4d and 4e) [10, 131–133]. Presence of unstable phenocrysts associations and frequent enclaves point to the involvement of magma mixing in the origin of the rocks. Individual extrusive domes may exceed 1 km in diameter forming recognizable rounded hills. Their internal parts are formed of massive rock with blocky jointing, often showing incipient flow-banding. Hematitization of mafic minerals and groundmass magnetite suggests interaction with exsolved steam during the final stage of crystallization. At some domes, domains of massive rock are separated by subvertical zones of brecciation implying relative movement of blocks during dome growth. At marginal parts of the domes, massive rock grades into blocky breccias.

Two solitary dome-flows of biotite-amphibole andesite composition occur in the central part of Kremnické vrchy Mts. in the Central Slovakia Volcanic Field (Figures 1-2a, 13) [134]. They rest on the eroded surface of an older effusive field infilling the Kremnica graben. In both cases the substrate slope has driven growing domes into a lateral outflow, creating lava flows up to 100 m thick with an extended cover of blocky breccia, up to 1 to 2 km distance. Domes themselves are deeply eroded and formed by massive andesite with blocky jointing, and glas close to contacts.

Several lava domes and associated volcaniclastic rocks that evolved in a shallow marine environment during the Early Badenian time in the southern part of the Central Slovakia Volcanic Field (Figures 1-2a, 13) also qualify as monogenetic volcanic forms. Extensive brecciation due to contact with water, occurrence of peperitic breccias and extensive deposits of submarine breccia flows (Figures 14) are characteristic features [135].

At the western foot of the Diel stratovolcano in eastern Slovakia (Figures 1, 11 – area 4c) a basaltic andesite lava dome represents a rare case of a monogenetic parasitic vent [136]; it is 300 m in diameter and is formed of massive rock with platy jointing, showing a fan-like internal structure.

5.4.3. Dome/Dome-flows related to felsic calc-alkaline rocks

Lava domes and dome-flows are the most frequent volcanic forms of felsic calc-alkaline composition. Rhyolite domes, dome-flows and lava flows have been analyzed and described in the Central Slovakia Volcanic field [137, 138]. They form solitary volcanic forms as well as a continuous field next at the eastern part of the Žiar volcano-tectonic depression (Figures 1-2a, 13). In analogy with basaltic monogenetic volcanic fields, they are monogenetic volcanoes: the products of a small volume subaerial single eruption occurring either as dispersed vents, or as individual vents. They are of the Middle to Late Sarmatian age (12.5-11.0 Ma) [137]. A moderate degree of erosion enables the study of their internal structure, as well as their paleovolcanic reconstruction. All of the domes are endogenous as they have been fed internally [28]. They represent upheaved plugs, Peléean type domes (most frequent), low domes and dome-flows [36, 126, 128] (Figure 15). In one case we have observed a regular 50 m thick lava flow extending for 5 km distance implying a relatively low viscosity. Internal parts of extrusive domes are formed of massive and/or spherulitic rhyolite, often with banded texture showing fan-like orientation. This implies growth by viscous injection of incoming lava into the central part of the dome and its lateral spreading. Dome flows may reach the length 1 to 2 km having thicknesses up to 100 m. Rhyolitic domes are usually glassy close to the margins and at the surface of the domes. There are extensive accu-

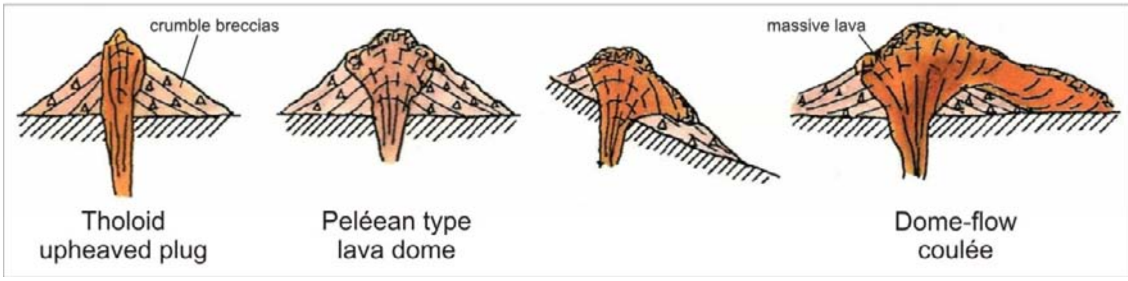

Figure 10. Essential types of andesite extrusive domes in the CPR.

mulations of glassy crumble breccias, some of them serving as perlite deposits (Lehôtka pod Brehmi and Jastrabá in the Central Slovakia Volcanic Field, Figures 1-2a and 13). Some of the domes have been eroded down to the level of feeding intrusive roots, showing funnel-like shape with glassy inward dipping margins at the contact with older rocks.

Extrusive domes and dome-flows in other sub-regions of the CPR (Tokaj, Zemplín, Beregovo, Oaş, Figure 1-4e) generally do not differ from those described in the Central Slovakia Volcanic Field [10, 132, 133, 139]. However, some of the domes were growing in a subaqueous environment. Pichler [140] described extreme brecciation of submarine lava domes due to contact with water at the island of Lipari, and he used the term acid hyaloclastites for the resulting glassy breccias. In the CPR extreme marginal brecciation of this type has been observed at the perlite deposit Pálháza in the northern part of the Tokaj Mts. in Hungary (Figures 1, 11 – area 4a) [141, 148]. Extensive accumulation of perlitic hyaloclastite breccias at the margin of a dome/cryptodome is in this case accompanied by zones of peperitic breccias at contacts with deformed soft sediments. Massive perlite/pitchstone with obscured inward dipping columnar jointing forms internal parts of the dome/cryptodome.

An isolated rhyolite dome-flow (Oraşu Nou) from the southern part of the Oaş Mts. (Figures 1-4e, 16) is another example of extensive quench fragmentation developed in an extrusive dome grown in a submarine environment. The strongly flow banded, vesicular glassy lavas are surrounded by both the in situ and resedimented hyaloclastites preserving the original lava flow foliation or showing abundant perlite spheroids [142]. The perlite spheroids expose a wide range of sizes and shapes, from mostly spherical 1 cm - 20 cm, to 1 m macro-perlite ovoid shapes.

Dome volcanoes

The growth of felsic extrusive domes and/or dome-flows is usually accompanied by explosive processes that, along with the lava extrusion, represent one eruptive cycle [18, 143, 144]. Products of such an eruptive cycle, along with contemporaneous reworked material, build up the monogenetic felsic extrusive volcanoes (Figure 15) developed in the Sarmatian Jastrabá Formation (Central Slovakia Volcanic Field, Figures 1-2a and 13) [145, 146]. The early stage in their evolution was either dominated by phreatomagmatic eruptions due to the contact of rising magma with groundwater or was dominated by vulcanian to sub-plinian type of eruptions in the case of dry bedrock. Phreatomagmatic eruptions led to the formation of tuff-rings in the central zone surrounded by a fine-grained pyroclastic rim originated from pyroclastic fall and base surge, and reworked tuffs in proximal and distal zones. Vulcanian and sub-plinian type eruptions and related pumice flows and pyroclastic surges created pumice-dominated tuff-cones in the central zone surrounded by pumiceous pyroclastic flow, surge and fall originated pyroclastic rock units and in the proximal and distal zones reworked pumiceous pyroclastic successions. Evolution of the volcanoes continued with extrusion of viscous lava. The growth of the dome was accompanied by extensive brecciation, accumulation of dome top breccias at its slopes and eventually also by the gravity driven block-and-ash flows. Periods of the vulcanian type explosive activity created horizons of pumice flow and fall deposits that alternate in the proximal zone with reworked pyroclastic rocks and epiclastic volcanic breccias. Reworked tuffs and epiclastic volcanic conglomerates and sandstones were deposited in the distal zone.

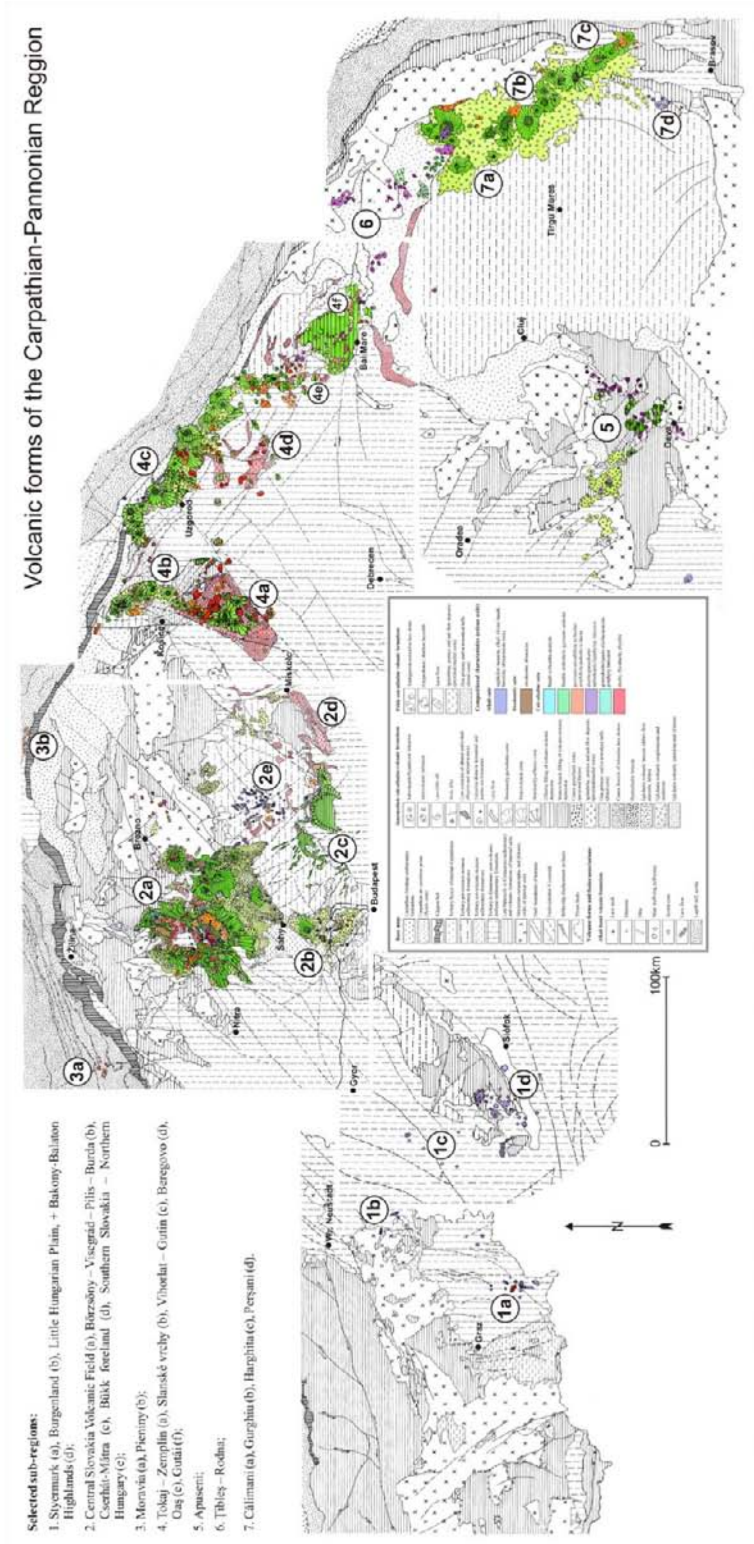

Figure 11. Volcanic forms in the selected sub/regions of the Carpathian/Pannonian Region. Compiled on the basis of schemes by authors of this paper and schemes provided by Fritz Ebner, Austria (Styrian Basin, Burgenland), Tibor Zelenka, Hungary (Czerhát, Mátra, Tokaj) and Bogdan Mackiv, Ukraine (Transcarpathia).

Cryptodomes

Cryptodome is a variety of lava dome or shallow laccolith that has grown just beneath the surface [123], commonly in sedimentary or pyroclastic rocks (Figure 17). Alternatively it has also been termed as "intrusive" dome [29, 147]. The growth of a cryptodome typically lifts up overlying rocks and the growing dome can eventually reach the surface. Internal structure of cryptodomes is often "onion-skin" concentric implying ballooning as the principal mechanism of growth [123].

In the CPR, cryptodomes have been identified in the Sarmatian Jastrabá Formation in Central Slovakia Volcanic Field (Figures 1-2a and 13) [134] and in Tokaj Mountains (Figure 1-4a) at Pálháza in NE Hungary where a complex association of Miocene rhyolitic shallow intrusions, cryptodomes and endogenous lava domes were emplaced into and onto soft, wet pelitic sediment in a shallow submarine environment [148]. In the Central Slovakia Volcanic Field the cryptodomes are up to 2 km in diameter, emplaced in a thick succession of rhyolite tuffs. Their internal part is massive, affected by interaction with exsolved fluids (oxidation, subsolidus recrystallization). Massive rhyolite grades at marginal parts into breccias showing textures typical for processes of hydraulic fracturing. Apparently, in this case brecciation is not a result of crumbling due to extension of solidified carapace but a result of fluid escape from a crystallizing cryptodome. Escaping fluids altered overlying tuffs into zeolites [149].

5.5. Shield volcanoes

A shield volcano is a low relief edifice made up of relatively fluid lava flows, having low angle flank slopes [32, 40]. Shield volcanoes are typically of effusive origin. Thin pahoe-hoe and aa-type lava flows dominate in their structure. They display typical upward-convex outer slopes and generally low slope angles due to low viscosity of lavas [32]. Most of the shield volcanoes are classified as polygenetic volcanic forms, however, small shields, referred to as scutulum, occurring in the framework of monogenetic volcanic fields are classified as monogenetic volcanic forms [40].

5.5.1. Shield volcanoes related to Na-alkalic and ultrapotassic rocks

In the CPR there are two relevant examples of these types of volcanic forms:

An ultrapotassic lamproite well-preserved shield volcano (198 m a.s.l.), revealing a sequence of vesicular lava flows possibly intercalated with some fallout scoria deposits was found in Banat region, south-western Romania (Figure 1). It is symmetric, with a 750 m diameter base and a craterial area of ca. 150 m wide (Figure 18). This isolated lamproite volcano, dated at 1.32 Ma, is situated at the southeastern margin of the Pannonian Basin on flat-lying Miocene sedimentary rocks which overlie older crystalline basement, along an important NE–SW fault system [150–152].

A slightly eroded alkali basaltic volcano, 2.4-2.6 Ma old, located along the South Transylvanian fault system southwest of the Apuseni Mountains (Figure 1-5) [130, 153] suggests a shield volcano. Presently, the volcano is covered by Quaternary deposits, exposed only in quarries. It has been quarried since the 17th century and is described as being generated in several eruptive episodes. Bedded tuffs overlie Pliocene sands and argillaceous sands in the northern part [154, 155]. The paroxismal activity corresponds to massive outpouring of lavas while the presence of rare scoria, volcanic bombs and scoriaceous lapilli at the top of volcano may suggest insignificant strombolian activity [156]. A detailed volcanological study of the volcano is still missing in order to determine if it is monogenetic or polygenetic.

5.5.2. Shield volcanoes related to intermediate calc-alkaline rocks

Some small-sized (commonly 3-5 km in diameter) effusive volcanic edifices occurring in the Călimani-Gurghiu-Harghita volcanic range (East Carpathians) (Figures 1, 11 – area 7) display a cone topography which can be characterized as shield morphology. One typical case is the 6.8 Ma amphibole andesite Borzont volcano in the southern Gurghiu Mts. (Figures 1, 11 – area 7b), located in between two large composite volcanoes (Seaca-Tătarca in the north and Șumuleu in the south) [1, 157, 158].

In the northernmost part of the North Harghita volcanic range (Figures 1, 11 – area 7c), another effusive center (Răchitiș), developed as a monogenetic volcanic event building up a flat shield-like cone, made of aphanitic amphibole andesites, 5.8 Ma in age [1, 157]. Although the outer slopes are not obviously upward-convex, as at Borzont, they are neither typically upward-concave as in the case of composite volcanoes; its lithology and effusive monogenetic nature inclined us to classify it among the shield forms. In the south-easternmost part of the Călimani volcanic area (Figure 1-7a) there is a rare, large (~10 km in diameter) ~8.5 Ma old calc-alkaline basalt shield volcano whose lavas have columnar jointing and marginal vesicular clinker (Sărmaș basalts) [159].

No crater remnants can be recognized at the mentioned volcanoes. In most cases autoclastic breccias which might have been associated with these effusive centers were presumably removed by erosional processes.

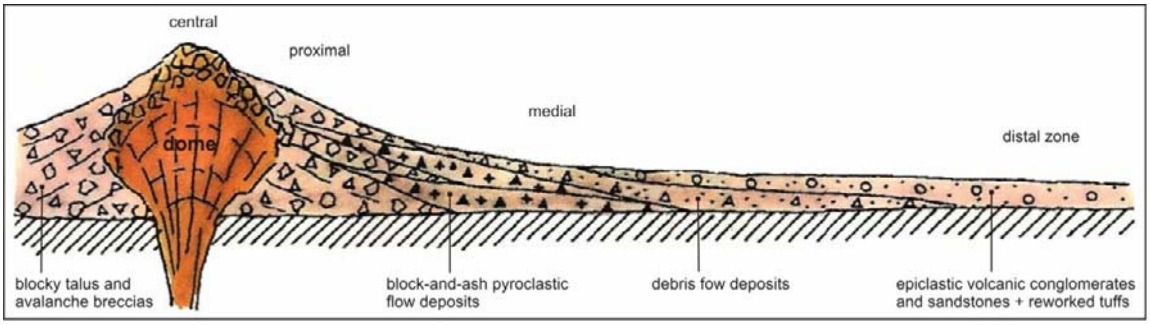

Figure 12. Scheme of andesite endogenous lava dome and related volcaniclastic rocks [219].

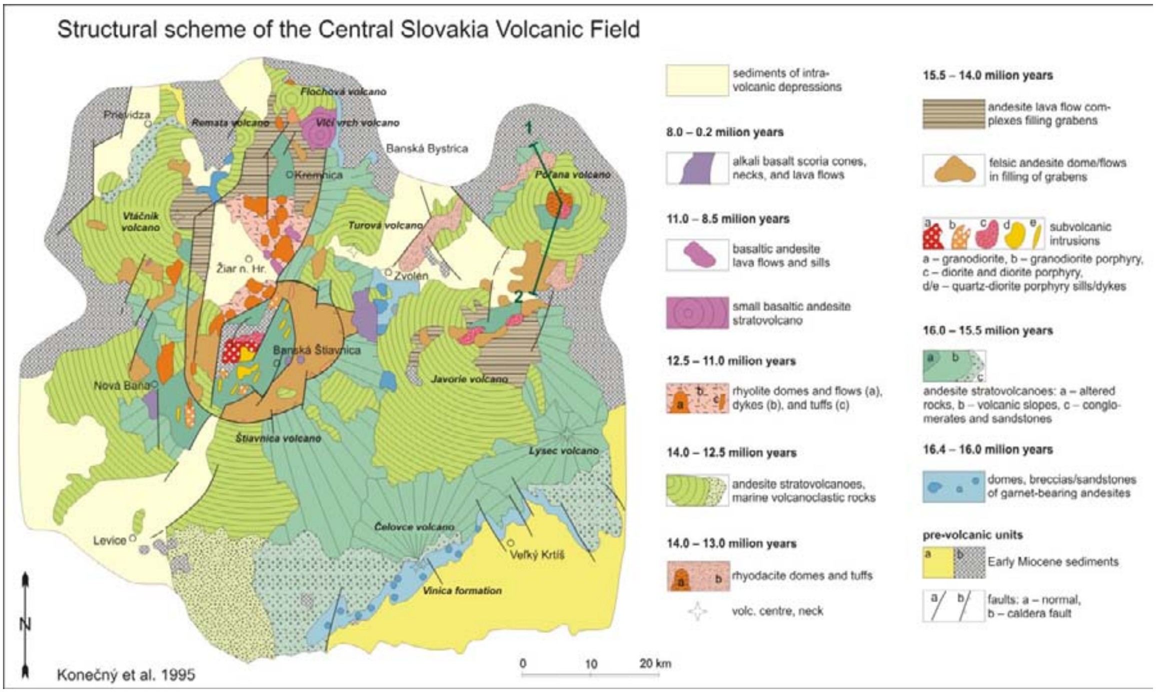

Figure 13. Structural scheme of the Central Slovakia Volcanic Field [145].

5.6. Intrusive forms

5.6.1. Intrusive forms related to Na-alkalic and ultra-potassic rocks

Alkali basaltic volcanic forms are associated with intrusive necks and plugs common in eroded monogenetic volcanic fields across the CPR [58, 73]. Some of them are well exposed, forming cliffs and/or lava capped hills (Figure 19). Others are covered and only in open pit quarries it is possible to closely examine the architecture of such volcanic remnants. By definition necks/plugs sensu stricto should show almost vertical contacts with the surrounding pre-volcanic basement rocks, while remnants of lava lakes should show inward dipping contacts with pyroclastic rocks of the tuff cone. While pointed hills with cliffs are characteristic forms of necks/plugs s.s. [160, 161], lava-capped buttes are characteristic of remnants of lava lakes [58].

Volcanic necks and plugs are commonly columnar-jointed (Figure 19) with polygonal section columns of three to eight, mostly five to six sides. The column width varies from 1 to 5 dm. The columns are normally regular by size and orientation, having growth-marks on their outer surfaces. Orientation of columns can be diverse from sub-horizontal to vertical arrays with gradual changes. The changing column orientation is related to the ther-

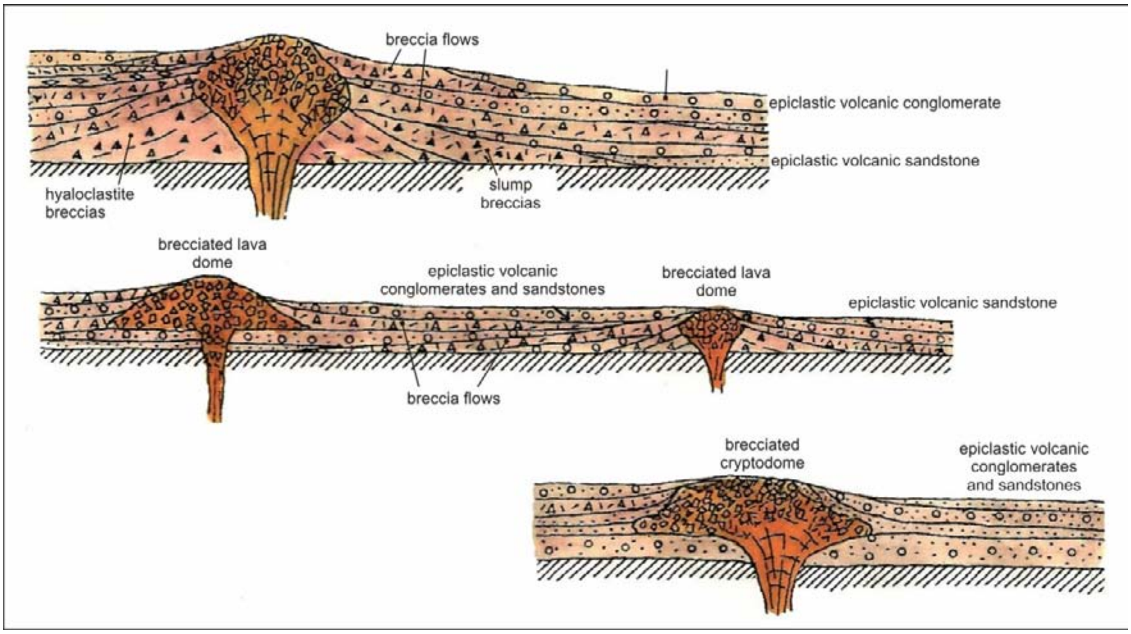

Figure 14. Paleovolcanic reconstruction of submarine andesite lava domes and related facies of volcaniclastic rocks in the southern part of the Central Slovakia Volcanic Field [219].

mal stress influencing the cooling of the trapped magma. In general, the cooling front is perpendicular to the column orientation [162–165]. Dikes associated with monogenetic alkalic volcanoes are commonly described from deeply eroded volcanic fields [38, 103, 166].

Extensive intrusive bodies are not normally expected to be associated with monogenetic volcanoes due to their short eruption duration and the limited magma volume involved [167]. Recently however, in many locations, a potential relationship between shallow magmatic intrusive processes and phreatomagmatic volcanism have been suggested on the basis of 3D architecture of volcanic rock facies exposed beneath the syn-eruptive paleosurface [103, 168]. In particular, monogenetic volcanic fields exhumed down to their deeper volcanic conduit zones showed volcanic facies architecture closely resembling shallow dike and sill complexes beneath the volcanic edifice at the surface [168]. Konečný and Lexa [169] have reported on a laccolith-like intrusion emplaced in the maar filling (Figure 20).

A diatreme is the subsurface part of a maar volcano [47, 170]. To express the close relationship between maar and diatreme, they are commonly referred to as maar-diatreme volcanoes [171]. Diatremes are vital parts of a maar-diatreme volcano (Figure 3), and can be envisioned as vertical pipes filled with volcanic debris located beneath the maar crater [171]. Diatremes "store" potentially the most valuable information on how the magma and water interacted and how the resulting fragmented rocks are transported away from their fragmentation site to the deposition sites surrounding the craters. The mode of transportation of pyroclast from the explosion site is much under debate and involves various processes generating fluidisation, debris jet formation and various vertically moving pyroclastic density current activities [78, 172–177].

Necks and plugs

The main problem with many necks and plugs in the CPR is that they are usually located in areas where no other independent volcanic stratigraphy data are available to establish the origin of this volcanic feature. Without drill core, shallow surface geophysical, volcanic stratigraphy and age data, the interpretation of volcanic erosion remnants forming butte-like hills covered by basaltic rocks can be very speculative. In locations in southern Slovakia and the western Pannonian Basin [58] lava-capped buttes have been associated with minor pyroclastic rocks, commonly massive lapilli tuffs with mixed pyroclasts suggesting their vent-filling origin (Figure 21). If such pyroclastic assemblages have been identified, the volcanic plug and neck origin can be confirmed. In any other cases the interpretation is rather speculative. Outward dipping columns would be a characteristic feature for remnants of lava lakes and such an orientation of columns has been observed at several localities of central and southern Slovakia (Fig-

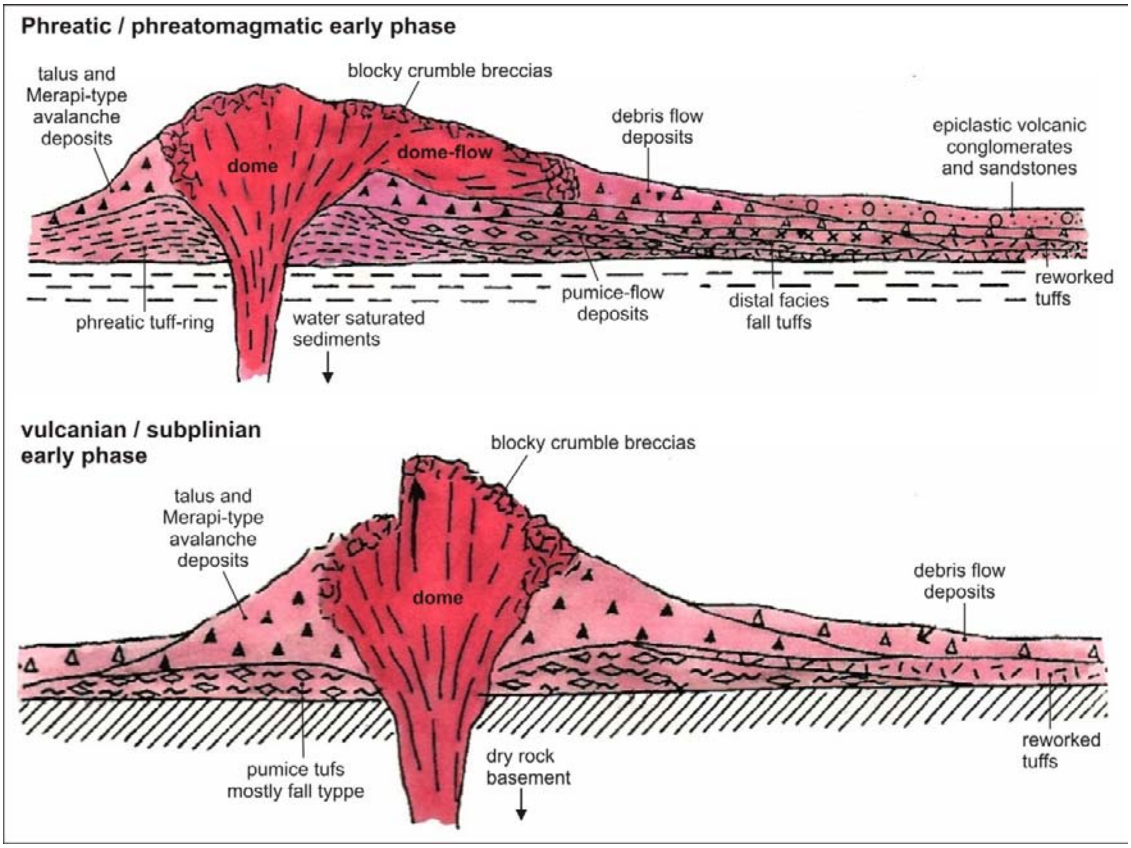

Figure 15. Synthetic paleovolcanic reconstruction of rhyolite extrusive dome monogenetic volcanoes in the CPR [219].

ure 19) [160] and Balaton Highland in Hungary [58].

Dikes

Erosional dike remnants are common in the CPR as cross cutting exposed volcanic conduits and vent facies (Figure 3). In spite of the abundance of dikes in the preserved cores of monogenetic volcanoes, relatively few dikes are known as an individual volcanic feature outside of the core of erosional remnants of monogenetic volcanoes in the CPR. Dikes suspected to represent magmatic bodies intruded along fissure networks of the country rocks are narrow basaltic rocks that commonly form chains of basalt-capped hills [178–180]. Due to the vegetation cover and the possible narrow thickness of these dikes [180], the lateral correlation and therefore the correct interpretation of these bodies in the CPR area is problematic [58, 179]. A laterally traceable dike swarm has been reported from the western margin of the BBHVF (Figures 1-1d and 9) [58, 179, 181, 182]. This dike swarm follows a NW-SE strike and a slightly curving line that connects to erosion remnants of phreatomagmatic monogenetic volcanoes [168]. The uncertainty of recognizing dikes as km-scale linear features within CPR monogenetic volcanic fields could either be the artifact of the advanced erosion and current vegetation cover, or be the sign that feeder dikes were rather narrow and short, and their preservation potential is poor over millions of years of further evolution of the region.

Individual dikes, however, are common in association with diatremes and other near-vent pyroclastic successions of monogenetic volcanoes [87, 183–186]. They commonly form swarms in a chaotic fashion. Their width rarely exceeds the dm-scale range. Dike margins commonly display peperites [187], where the coherent magmatic body interacts with the host wet pyroclastic succession [188, 189]. Blocky and globular peperites have been reported from the western Pannonian Basin [190] and in southern Slovakia [191, 192]. Dikes intruding Neogene siliciclastic sediments are common in the western edge of the BBHVF and form arrested dike networks commonly with peperitic margin, indicating the water-saturated and unconsolidated nature of the host sediments at the time of dike intrusion [193].

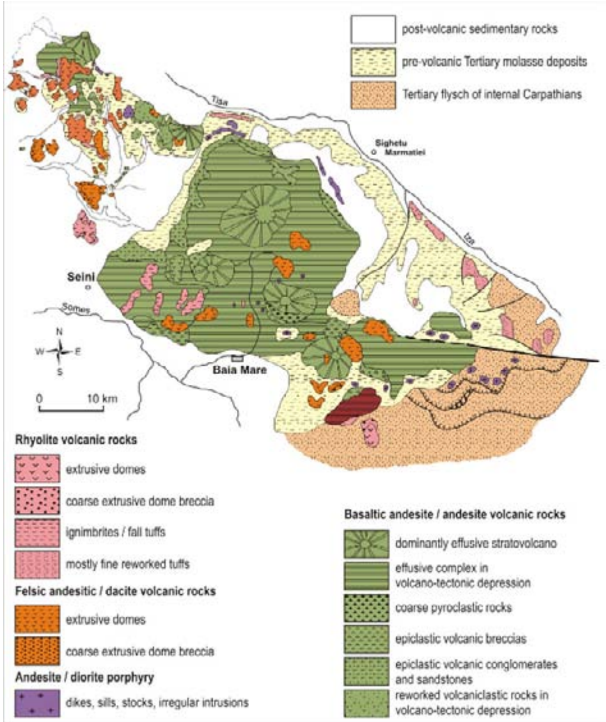

Figure 16. Volcanic forms of the Oaș – Gutâi sub-region, northern Romania.

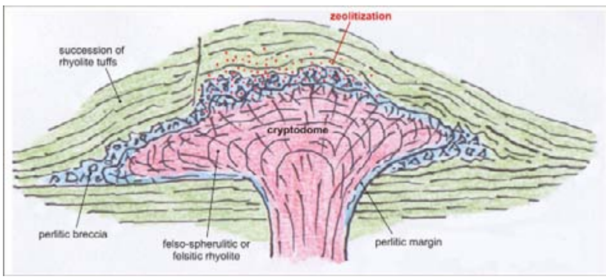

Figure 17. Scheme of a cryptodome (intrusive lava dome) emplaced in a succession of rhyolite tuffs in Central Slovakia Volcanic Field [137].

Dike-sill complexes and shallow intrusive bodies

In the western Pannonian Basin Neogene siliciclastic sedimentary basins have been penetrated by alkaline basaltic magma forming dike and sill complexes [168]. The erosion in these areas of exhumed dike and sills represent syn-eruptive depth of at least 100 meters beneath the syn-volcanic surface as supported by independent 3D correlation of volcanic units preserved. The basanitic sills are irregular in shape and their lateral extent is highly variable suggesting complex conditions upon their emplacement [168]. Individual sills can reach a thickness of a few tens of meters and they commonly form dome-like structures with rosette-like radial columnar joint patterns. The margins of sills and dikes are commonly irregular, peperitic, showing evidences of magma and wet, water-saturated sediment interaction [190]. Complex changes from various types of peperite to non-peperitic margins of intrusive bodies have been interpreted as results of inhomogeneities in water content and rheology of the siliciclastic deposits during intrusion.

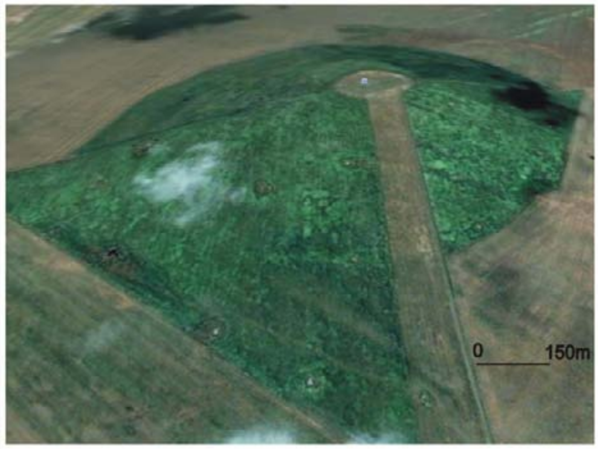

Figure 18. Google Earth image of the 1.32 Ma ultrapotassic lamproite Șumiga volcano, Banat, Romania. Presently the volcano is covered by Quaternary deposits and agricultural crops.

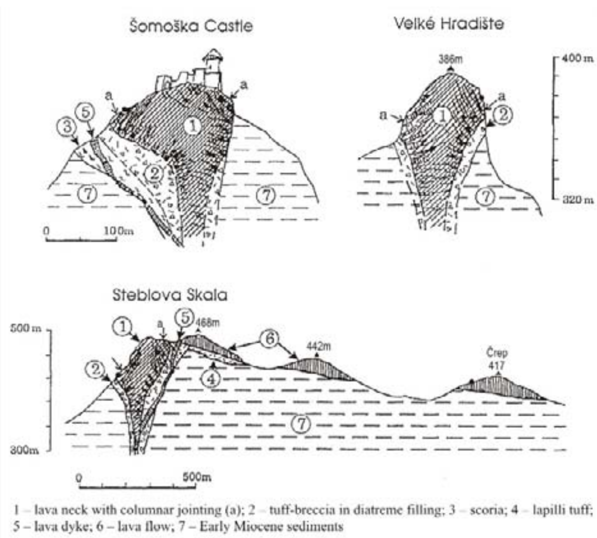

Figure 19. Selected lava necks/plugs in the Northern Hungary – Southern Slovakia alkali basalt monogenetic volcanic field [359]. Note that the lava body at the Šomoška castle hill represents a root of a lava lake.

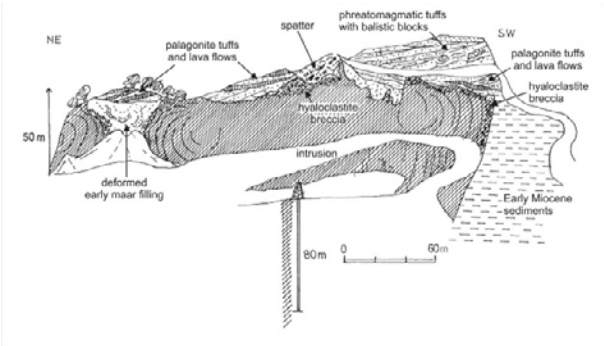

Figure 20. Bulhary maar in southern Slovakia – a bulbous laccolith-like intrusion emplaced in the lower part of maar filling [360].

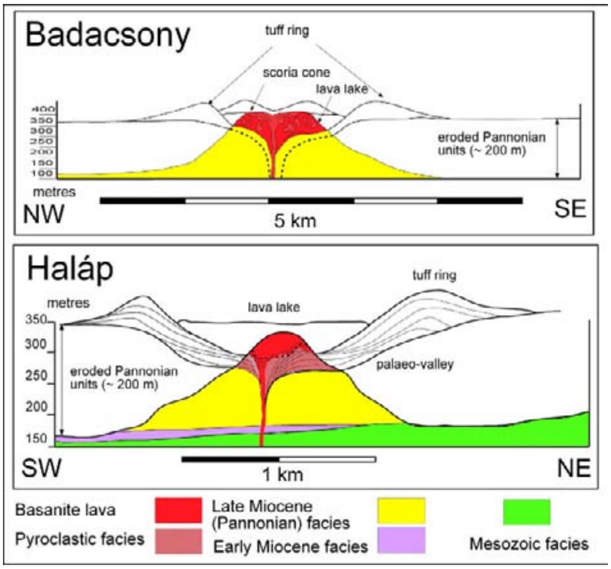

Figure 21. Architecture of lava buttes in the Bakony-Balaton Highlands Volcanic Field, Hungary.

Diatremes

In CPR a direct access to diatremes is possible since erosion is advanced and the diatremes are exposed (Figure 22), such as in many Miocene to Pliocene phreatomagmatic volcanic fields in the Pannonian Basin [58, 73]. Many of the Pannonian Basin diatremes are exposed cutting through a typical soft-substrate country rock environment dominated by sand, gravel and silt of the Neogene shallow marine to fluvio-lacustrine silicilastic succession [75, 194]. The meso- to micro-scale textural features of the exposed diatremes show evidence of the coarse mixing between the rising basaltic magma and the host water-saturated sediments [190]. Disrupted peperitic-domains are common, as well as plastically deformed chunks of mud and semi-baked silt in a dm-to-m scale size. Many quarries exposing diatreme fill made few of these diatremes world-class sites to understand the emplacement mechanism of mafic magma into water-saturated sediments forming peperitic feeder dikes [190]. The preserved rocks of the CPR diatremes are various lapilli tuffs and tuff breccias with a dominantly massive texture [195]. Large blocks of country rocks from underlying hard-substrate can be recognized. Cross-cutting dikes and bulbous sill-like magmatic bodies are present in large diatremes in southern Slovakia and western Hungary [168]. The diatreme filling pyroclastic rocks are rich in accidental lithic fragments, rounded basaltic lapilli, cored bombs and irregular-shaped mud chunks indicating active conduit processes, particle mixing and syn-volcanic mass reorganization typical for many diatremes worldwide. Diatremes in the CPR also contain exotic deep-seated crustal and upper mantle xenoliths indicating rapid ascent of magma from its heterogeneous and commonly metasomatized source [59, 196–198].

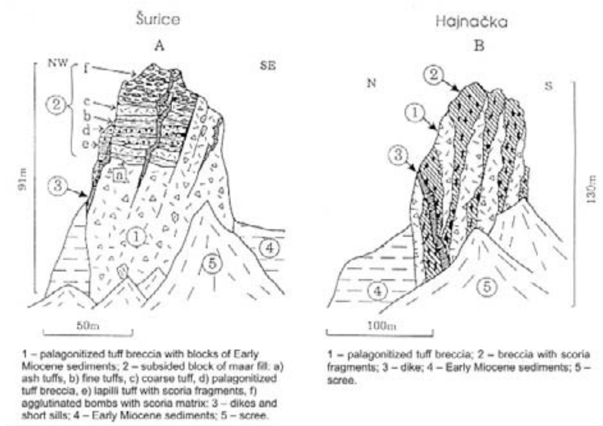

Figure 22. Diatremes Šurice and Hajnáčka in southern Slovakia [192]. Diatreme filling is exposed due to a deep erosion of surrounding soft Tertiary sedimentary tocks. Note that upper part of the Šurice diatreme represents a subsided maar complex.

5.6.2. Intrusive forms related to intermediate calc-alkaline rocks

Intermediate calc-alkaline intrusions associated with volcanic structures are usually andesitic or dioritic, depending on the intrusion level. They are either (I) “intravolcanic”: (Ia) closely related to volcanic core-complexes (calderas or craters) associated with hydrothermal alteration and mineralization processes [159, 199–202], (Ib) intrusions tectonically controlled in earlier volcanic structures [203–206] or (II) “hypabyssal”, emplaced in basement rocks [207]. Most of the intrusions associated with volcanic activity are buried and reached only through drilling or mining works. Unroofed intrusions are found only inside the calderas [158, 159, 201]. The larger subvolcanic intrusions (I) have irregular shapes that are difficult to be properly characterized, however fitting the volcanic edifice, mostly like stocks; other associated smaller size intrusions are represented by necks, sills and dikes.

Intermediate calc-alkaline subsurface intrusions show a large petrographic range from basalts (gabbro-diorite) to andesite (diorite) to rhyolites (granite) and a wide range of textures, commonly related to the depth of cooling. They usually form independent bodies piercing various basement lithologies (metamorphic or sedimentary) [207–210]. The independent intrusions are mostly outcropping in strongly eroded terranes [211]. Few complex studies involving K/Ar, Ar/Ar dating, and U-Pb dating on zircon crystals from successive mineralized and barren intrusions (e.g. Apuseni Mts.) established their age relationships with mineralization events in an interval of ~1.5 Ma between 12.5-11 Ma [212, 213].

Dikes

Dikes are discordant, commonly near vertical, sheet-like bodies of magmatic rocks, cutting discordantly through country rocks [214]. In the CPR they cut exposed volcanic vent area rocks and volcanic deposits outside of the vent areas, or they cut basement rocks [159, 200, 207]. Some of the dikes were feeders of volcanic, especially effusive eruptions. Their thickness varies from decimeters to tens of meters, length reaching in extreme case to kilometers (Cserhát Mts. in N Hungary, Figures 1, 11 – area 2c). Often they show chilled margins with columnar or platy jointing.

Irregular protrusions or plugs

Irregular protrusion or plug is a solidified lump of magmatic rock filling a volcanic vent or crater [214]. Intermediate composition plugs are rare in the CPR. As examples, we can mention dacitic plugs filling the diatreme-vent edifice of Roșia-Montana volcano in the Apuseni sub-region (Figures 1, 11 – area 5) [215] and the lava body at the top of the Vihorlat stratovolcano in eastern Slovakia (Figures 1, 11 – area 4c) [216].

Sills

A sill is a sheet-like magmatic body usually emplaced conformably in the stratified basement deposits [217]. In the CPR they mostly occur in basement rocks and rarely as intrusions in the vent or outside-vent areas [207, 210, 218]. In mature andesite stratovolcanoes they occur frequently along the volcanic/basement boundary [219] and/or take part in the structure of volcanic core complexes (an assemblage of volcanic and intrusive rocks in central zones of andesite stratovolcanoes affected by areal alteration)

Laccoliths

A laccolith is an intrusion of igneous rock in the Earth's crust which has a sub-horizontal base and spreads and forces the overlying strata into a dome [184, 220–223]. It is found in the CPR as an exposed subvolcanic intrusion in the sedimentary basement in the Visegrád area in Hungary (Figure 1-2b) [209], Salgótarján area at the Hungary/Slovakia boundary (Figure 1-2e) [224], the Lysá Stráž – Oblík zone in eastern Slovakia (Figure 1-4b) [225, 226], and in the "Subvolcanic zone" of the Eastern Carpathians (Figure 1-6) [207]. Laccoliths take part also in the structure of volcanic core complexes [219].

Stocks

A stock is a quasi cylindrical or irregular solidified subsurface magmatic body with no clear connection to a volcanic vent [214]. Such kinds of intrusions are known in the central part of composite volcanoes in CPR, however rarely exposed by erosion, and mostly found in mining works [90, 131, 159, 201, 205, 218, 227, 228]. Diorite porphyry stocks are responsible for alterations and intrusion-related mineralizations in the central zones of mature andesite stratovolcanoes. Several individual stocks have been described crosscutting the basement in the "Subvolcanic zone" of Eastern Carpathians (Figure 1, 11 – area 6) [207].

5.6.3. Intrusive forms related to felsic calc-alkaline rocks

Dikes

Felsic dikes in the CPR show a great variability. Relatively thick and short dykes associate with lava domes and apparently represent their roots in areas of deeper erosion. They are formed of massive, usually flow-banded rhyolite with blocky to columnar jointing (sub-horizontal), often glassy when close to contacts with surrounding rocks. Another type of dike is fault-controlled, relatively thin and long. Characteristic is the platy jointing parallel with the contacts. Frequently they occur at marginal faults of local horsts. The third type of dike associates with low sulfidation epithermal systems. They are emplaced at the same faults as epithermal veins and affected by silicification and adularization processes. Thicker dikes of this type show holocrystalline groundmass and should be classified as granite porphyry, rather than rhyolite.

6. Polygenetic volcanoes

In contrast to monogenetic volcanoes, polygenetic volcanoes have been created by longer-lasting, often recurrent and voluminous eruptions [35]. Among alkali basalt and ultrapotassic volcanic forms only a part of maar/diatreme volcanoes fulfill this definition. Only those are considered as polygenetic volcanoes, which show increased complexity and evidence for recurrent eruptive activity (see below).

Most polygenetic volcanoes are of intermediate composition and such the polygenetic volcanoes are the most widespread in the world [32] as well as in the CPR [2]. Besides alkali basalt monogenetic volcanic fields, polygenetic volcanoes of the stratovolcanic type dominate in all other sub-regions (Figure 1). Intermediate composition polygenetic volcanic forms are represented also by dome-flow fields and polygenetic lava fields with no obvious central volcanic features in volcanotectonic depressions – grabens. Subvolcanic and hypabyssal intrusive complexes represent subsurface equivalents of polygenetic volcanoes.

Felsic rocks commonly tend to occur as monogenetic volcanic forms (see above). However, they may build up polygenetic volcanic forms as well. We classify several (compound) dome-flow complexes as polygenetic. Felsic calderas and related ignimbrite deposits are other types of polygenetic volcanic forms occurring in the CPR. Felsic rocks also occur as an integral part of some calc-alkaline intermediate composite volcanoes.

Well-preserved felsic and intermediate calc-alkaline and alkali mafic polygenetic volcanoes are also found in the Pannonian basin buried by Miocene to Pliocene sedimentary rocks. Their presence is documented by geophysical data as well as by petrological data and K/Ar dating on cores [229].

6.1. Maar/diatreme volcanoes

Complex maar/diatreme volcanoes with recurrent eruptive activity that can be classified as polygenetic are associated only with Na-alkalic rocks in CPR. In general, maar-diatreme volcanoes are considered to be monogenetic phreatomagmatic volcanoes. During their short-lived eruption usually small magma volumes are involved producing small-scale volcanic landforms [38, 40]. In spite of this simple model, current research showed numerous evidences that volcanoes considered earlier as monogenetic, including maar-diatremes, can be complex both from a volcanic edifice and chemical point of view [103, 109]. This paradox can be solved through the introduction of an intermediary type in volcano classification where complex monogenetic volcanoes are somehow distinguished from *sensu stricto* monogenetic and purely polygenetic volcanic landforms [109]. This problem is critical while interpreting large maar-diatreme volcanoes that evidently represent complex volcanic features which cannot be explained to have been formed in a single eruptive episode [109].

The CPR is the home of many such complex maar-diatreme volcanoes that occur in the BBHVF (Figures 1-1d, 9) and in the Northern Hungary – Southern Slovakia (Nógrád-Gemer) volcanic field (Figures 1-2e, 8).

Complex maar-diatreme volcanoes would clearly be different from a typical polygenetic volcano, and would also be distinguishable from a *sensu stricto* monogenetic volcanic landform. (see Terminology). The CPR has numerous complex maar-diatreme volcanoes [43, 230, 231]. These are fairly large volcanic complexes, with multiple vents, commonly in nested array. Their eruption style usually show great variations besides a single eruptive episode, as fluctuating from phreatomagmatic to magmatic and vice versa. Styles may change as a response to the changing magma flux and external water availability. It seems that in the CPR such complex volcanoes formed over areas where larger magma volumes were generated and rising melts were captured in deep structural zones as potential pathways to the surface during eruption. They were "switched on and off" during the total time-span of an individual volcanic field. Good examples for such volcanism are the Fekete-hegy [43] and Bondoró in the central part of the Balaton Highland (Figures 1-1d and 9) or the Ostrá Hora and Bulhary (Figures 8,23) in southern Slovakia [232, 233]. In addition, the multiple groundwater sources and surface water availability provided good support to form diverse volcanic landforms. Also, given the fact that the Pannonian Basin was filled with thick Neogene water-saturated sediments, the rising magma likely followed a complex path in the uppermost hundreds of meters below the surface. Laterally shifting vents and/or rejuvenating vents more or less in the same location have created complex phreatomagmatic volcanoes over time.

6.2. Stratovolcanoes

Polygenetic volcanoes of the stratovolcano type – or simply stratovolcanoes – are relatively large, long-lived, mostly constructional volcanic edifices with upward-concave outer slopes, comprising lava and volcaniclastic products erupted from one or several more closely-spaced vents, including their reworked medial to distal equivalents [35]. They can be (1) simple, corresponding to one evolutionary stage, showing a high degree of symmetry, (2) composite, corresponding to several evolutionary stages separated by periods of erosion, but still with a moderate degree of symmetry, or (3) compound (multiple) representing coalesced edifices of several closely-spaced simple

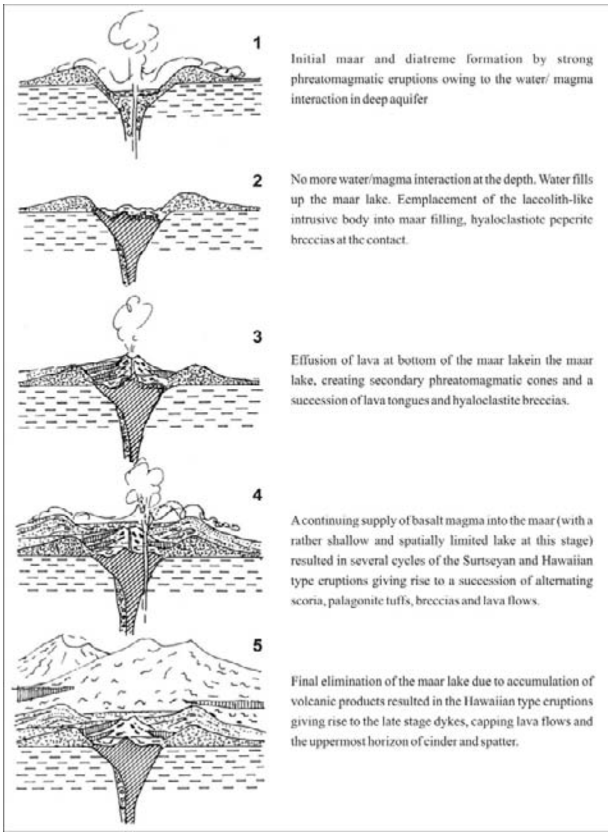

Figure 23. Evolution of the Bulhary polygenetic maar/diatreme volcano, southern Slovakia [360].

and/or composite volcanoes [28, 32, 34, 35]. The life-span of stratovolcanoes varies with increasing complexity from thousands of years to hundreds of thousand years [35]. Radiometric dating of the most complex compound stratovolcanoes in the CPR, involving several periods of erosion, however, points to life-spans up to ~3 Ma [2].

Stratovolcanoes and their erosional remnants show a pronounced facies zonality that allows for their recognition and paleovolcanic reconstruction even in instances when deep erosion has eliminated their characteristic morphological aspects. Traditionally, the central, proximal and distal zones [29] or central, proximal, medial and distal zones [234] are recognized. Davidson and DeSilva [35] have replaced these terms with terms based on facies associations: (a) main vent association, (b) cone-building association and (c) ring plain association, corresponding more or less to the central, proximal and medial zones of Mathisen and McPherson [234]. Facies associations prograde outward with increasing size of stratovolcanoes.

Stratovolcanoes in the CPR show a great variability in size and complexity from simple cones of the dominantly pyroclastic, stratovolcanic or effusive type to large composite and compound stratovolcanoes involving variably incorporated monogenetic volcanic forms, calderas, domes and dome-flows of differentiated rocks and intravolcanic and/or subvolcanic intrusive complexes. Most of the stratovolcanoes evolved in terrestrial conditions and their distal zones are represented by fluvial, lacustrine facies. Some of the stratovolcanoes grew close to the coast of the coeval Miocene Parathethys Sea, hence they grade into shallow marine facies at the distal zone.

Stratovolcanoes are formed dominantly of intermediate calc-alkaline rocks of andesitic composition. However, mafic and felsic rocks may occur as minor constituents. Konečný and Lexa [235] have classified andesite stratovolcanoes of the Central Slovakia Volcanic Field on the basis of their lithology and complexity. Their classification (Figure 24) with minor changes is applicable to andesite stratovolcanoes in the whole CPR.

6.2.1. Pyroclastic stratovolcanoes

Dominantly pyroclastic stratovolcanoes of pyroxene to amphibole pyroxene andesites in their central zone display cone building facies – vent breccias, coarse pyroclastic breccias, agglomerates, rare tuffs and horizons of scree with primary dips of 20 – 30° (Figure 24). In the proximal zones with primary dips around 15 – 20° these rocks are accompanied by block-and-ash-flow deposits and rare coarse debris flow deposits. Block-and-ash-flow and associated debris flow and mudflow deposits dominate in the medial zone, with primary dips around 10°. Mudflow deposits, reworked tuffs and epiclastic volcanic conglomerates and sandstones are characteristic for the distal zone ring-plain facies. Lithology of pyroclastic rocks points to explosive activity dominantly of the strombolian and vulcanian types. Block-and-ash flow deposits imply a temporal growth of extrusive domes. Čelovce volcano in the southern part of the Central Slovakia Volcanic Field (Figures 1, 11 – area 2a, Figure 13) is of this type [135, 200]. Only the central zone facies association of mostly coarse pyroclastic breccias and agglomerates is preserved of the Kamienka volcano in the Vihorlat Mts. in Eastern Slovakia (Figures 1, 11 – area 4c) [216].

At some of the pyroclastic stratovolcanoes, differentiation of magma has lead to the late stage emplacement of tholoids and/or lava domes of more silicic lava accompanied by block-and-ash-pyroclastic flows (Figure 24). Lysec volcano in the southern part of the Central Slovakia Volcanic Field is of this type (Figures 1, 11 – area 2a, Figure 13) [135, 200].

An early transition to summit dome growth results in stratovolcanoes whose final form is determined by extrusive domes and related pyroclastic deposits including coarse

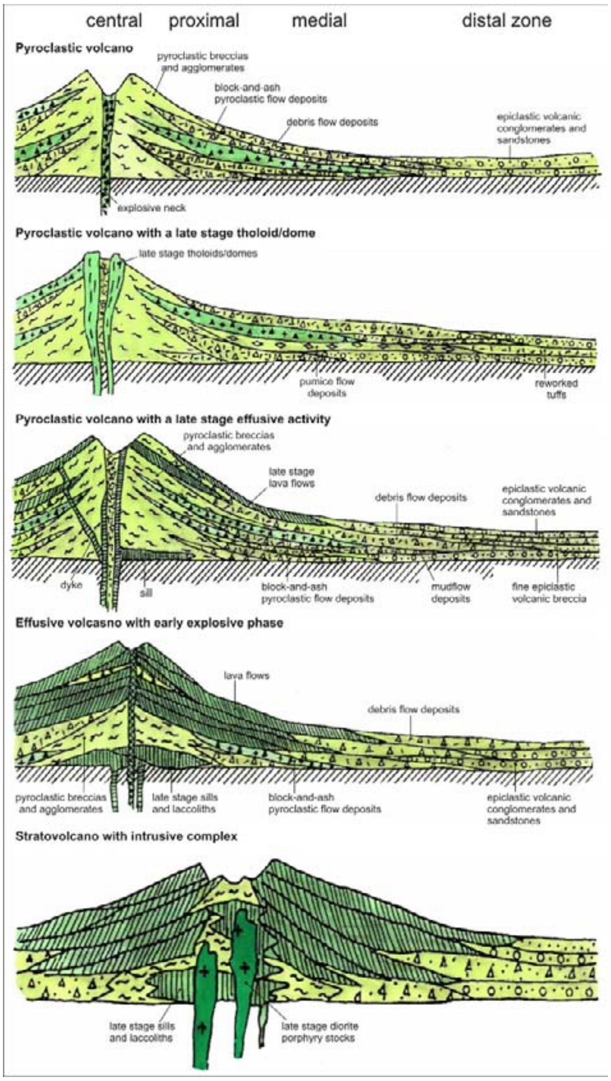

Figure 24. Integrated schemes of andesite stratovolcano types in the CPR [219].

block-and-ash-flow deposit dominated fans that grade outward into a typical volcanic ring-plain acting as an accumulation zone of volcanic material of distal primary pyroclastic flows, and their reworked counterparts – debris flows, mud flows, hyperconcentrated mud flows and normal stream flow deposits. Remnants of such the stratovolcanoes are exposed at ridges of the Börzsöny Mts. [236–238] and Visegrád Mts. [239] in Hungary (Figures 1, 11 – area 2b).

6.2.2. Stratovolcanoes sensu stricto

Stratovolcanoes sensu stricto (limited meaning of the term) are those volcanic cones that are composed of alternating pyroclastic rocks and lava flows [40]. It is a common feature of longer lasting volcanic eruptions that during their early phase lava is richer in volatiles and volcanic eruptions are dominantly explosive [28]. Corresponding stratovolcanoes, formed dominantly of pyroxene and amphibole-pyroxene andesites, have been mostly explosive during their early stage of evolution, showing facies associations corresponding to the strombolian and vulcanian, rarely sub-plinian types of eruptions. However, diminishing explosivity has lead later to alternating explosive and effusive activities and finally to dominantly effusive activity during the late stage of the volcanoes that created a cover of lava flows (Figure 24) similar to mostly effusive volcanoes described below. At this stage agglomerates, agglutinates and thin lava flows with dips of 20-25° accumulate in the central and proximal zones, while thicker lava flows and coarse epiclastic volcanic breccias with primary dips of 10-15° are characteristic of the medial zone. Epiclastic volcanic breccias, conglomerates and sandstones make up the distal zone sequences [200]. Individual stratovolcanoes differ by relative proportion of the explosive and effusive stages products. An initial hydromagmatic explosive stage has been observed at volcanoes that have started their activity in the shallow marine environment. In more eroded volcanoes, dikes, sills and laccoliths were observed in their central zones that represent intrusive counterparts to the late stage effusive parts of the stratovolcano.

Stratovolcanoes of this type are the most frequent in the CPR region. Poľana and Remata stratovolcanoes in the Central Slovakia Volcanic Field (Figures 1, 11 – area 2a, Figure 13) [200] and the majority of stratovolcanoes in the Slanské vrchy and Vihorlat mountain ranges of Eastern Slovakia (Figures 1, 11 – areas 4a,b) [91], Gutin Mts. of Transcarpathia in Ukraine (Figures 1, 11 – area 4c) [10, 132] and the Călimani-Gurghiu-Harghita range in Romania (Figures 1, 11 – area 7) [202] belong to this type.

6.2.3. Effusive stratovolcanoes

Stratovolcanoes of this type differ from the previous ones in relative proportion of the explosive and effusive stages (Figure 24). Explosive activity of the strombolian and/or vulcanian type plays a substantial role only at the early stage of the stratovolcano evolution. Later on effusive activity dominates – effusive rocks dominate in their structure and morphology. This aspect is reflected in lower primary dips in the central and proximal zones and in a higher resistance to erosion due to thick accumulation of lava flows in their proximal zone. However, they still preserve the upward-concave morphology due to syngenetic erosion and accumulation of epiclastic volcanic rocks in the medial and distal zones. That makes them different from shield-like composite volcanoes described below.

Primary dips of effusive stratovolcanoes reflect viscosity of lava flows and in turn their composition [35]. More silicic lavas create steeper cones. Tokaj Hill volcano in NE Hungary represents an example. Here, dacitic lava created an effusive cone with unusually steep slopes.

Stratovolcanoes of this type are almost as frequent as the preceding ones in the CPR region. Flochová and Vtáčnik stratovolcanoes in the Central Slovakia Volcanic Field (Figures 1, 11 – area 2a, Figure 13) [200] and a large part of stratovolcanoes in the Tokaj Mts. of NE Hungary (Figure 1-4a) [139], in the Slanské vrchy and Vihorlat mountain ranges of Eastern Slovakia (Figure 1-4b) [91], Gutin Mts. of Transcarpathia in Ukraine (Figure 1-4c) [10] and the Călimani-Gurghiu-Harghita range in Romania (Figure 1-7) [202] belong to this type.

6.2.4. Stratovolcanoes with late-stage intrusive activity

A characteristic feature of large and mature andesite stratovolcanoes is the late-stage emplacement of intravolcanic and subvolcanic intrusions (Figure 24). Intravolcanic sills and laccoliths have intruded along and above the basement-volcanic complex boundary due to the density contrast (basement density is higher and volcanic complex density is lower than density of andesite magma) [200]. Emplacement of subvolcanic sills and laccoliths is limited to stratovolcanoes with sedimentary rock basement. An important event in the evolution of andesite stratovolcanoes represents the emplacement of subvolcanic/intravolcanic stocks of diorite porphyry. Emplacement of high-level stocks initiates hydrothermal processes that alter the internal parts of volcanoes and create epithermal mineralizations [240]. Extensive alteration enables fast erosion in the central zone of the volcano and can be a cause of its sector collapse (see below).

Only few andesite stratovolcanoes in the CPR have been eroded deeply enough to expose their intravolcanic/subvolcanic intrusive complex and associated alteration/mineralization. These are Poľana and Javorie stratovolcanoes in the Central Slovakia Volcanic Field (Figures 1, 11 – area 2a, Figure 13) [241, 242], Zlatá Baňa stratovolcano in the Slanské vrchy mountain range in Eastern Slovakia (Figures 1, 11 – area 4b) [90] and Popričny stratovolcano at the Slovakia/Ukraine state border (Figures 1, 11 – area 4c) [10]. At less eroded stratovolcanoes their intrusive complexes are not exposed. However, their presence in depth is inferred on the basis of alterations, drilling or interpretation of geophysical data. In this way, intrusive complexes have been confirmed at many of the above mentioned stratovolcanoes in the Slanské vrchy and Vihorlat mountain ranges in Eastern Slovakia, Gutin Mts. in Ukraine and in the Călimani-Gurghiu-Harghita Mts. in Romania (Figures 1, 11 – areas 4b, 4c and 7 respectively).

6.2.5. Collapsed stratovolcanoes

Volcano instability has been recently recognized as a significant catastrophic landform modification process [243–245]. Current research has recognized that nearly every volcanic setting shows results of volcano failure and their associated large volume volcanic debris avalanche deposits and typical hummocky terrains [246–251]. Volcano instability has not only been recognized on terrestrial volcanoes, but increasing amount of evidence indicates its importance of landscape evolution of volcanic islands around the globe [252–254].

Edifice failure in the form of lateral collapse has been pointed out at two large stratovolcanoes in the Călimani-Gurghiu-Harghita volcanic range (East Carpathians), Rusca-Tihu (Călimani Mts.) (Figures 1, 11 – area 7a) and Vârghiş (North Harghita Mts.) (Figures 1, 11 – area 7c) [8]. These destructive events resulted in the generation of large-volume debris avalanche deposits spread out over large territories and of horseshoe-shaped depressions left behind at the edifice center. They strongly marked both the further evolution of the volcanoes and the topography and hydrography of their environs. The main characteristics of the two debris avalanche events are summarized in Table 1. The Rusca-Tihu volcano sector collapse occurred at ca. 8 Ma b.p. as a consequence of gravitational instability resulting from rapid growth of the stratovolcanic edifice through high-frequency eruptions of basaltic andesite magma. It was followed by intensification of volcanic activity which "healed" the debris avalanche depression. The failure event was directed sectorially (sector collapse) toward the west, south-west and south, probably due to tectonic uplift of the basement beneath the northern flank of the volcano. In contrast, the volcano collapse event at Vârghiş volcano at ca. 5 Ma cannot be related to a particular eruptive event, nor was it followed by significant post-collapse activity, so that its horseshoe-shaped depression is still recognizable in the present-day volcano topography. The resulting debris avalanche deposit covers large surfaces to the south and south-west of the volcano, including areas later covered by products of the younger South Harghita volcanoes (Figure 1-7c). The direction of the debris avalanche transport as well as of the collapse itself was determined by the NNE-striking horst-and-graben structure whose extension beneath the Harghita volcanoes was documented geophysically.

The recent identification of debris avalanche deposits (DADs) originating from the southern edge of the Igniş volcano (1306 m, highest of the Gutâi Mts., Figure 1-4f, 16) is closely connected to an important strike-slip fault system [255]. The flank failure event has left an

Table 1. Summary of characteristics of the debris avalanche deposits in the Călimani-Gurghiu-Harghita volcanic range (acc. to Szakács and Seghedi, [8]).

Volcano	Surface covered (km^2)	Volume (km^3)	Runout distance (km)	Drop (km)
Rusca-Tihu	870	26	55	1.4
Vârghiș	520	13	45	1.3

E-W-oriented horseshoe-shaped scar with an estimated volume of material removed of at least 0.35 km^3 and an estimated area covered of at least 4,345 km^2. The deposit is a mega breccia with a variable amount of coarse matrix with jigsaw-fractured blocks, large boulders, and several southward-elongated hummocks up to 1800 m distance from the scar. Between 720-850 m altitudes the DADs contain megablocks of 5-12 m thick and up to 100 m long of layered fine-grained poorly consolidated pyroclastic materials of interlayered ash and lapillistone of fallout origin, and clay beds rich in vegetation remnants (known as the "Chiuzbaia flora" of similar age as the surrounding lava flows, i.e. ca. 10-7 Ma) and diatoms.

6.2.6. Stratovolcanoes with caldera or volcanic graben

Evolution of large mature stratovolcanoes is eventually interrupted by evolution of calderas or volcanic grabens. Calderas are large volcanic depressions, more or less circular in form, the diameter of which is many times greater than included vents [45]. They generally associate with some kind of a roof collapse of an underlying shallow magma chamber [256] that can be caused by a voluminous explosion of the Plinian type or a large scale outflow of lava. Volcanic grabens (or sector grabens, [256]) represent caldera-like volcanic depressions heavily influenced by regional extension tectonics. Volcanic grabens hosted by a single composite stratovolcano should be distinguished from volcano-tectonic grabens or depressions that in turn usually host products of more than one individual volcano. Calderas and volcanic grabens associate especially with those stratovolcanoes whose magmatic system has created a shallow magma chamber in which differentiation has taken place towards felsic magmas of lesser density, enriched in volatiles. Evolved rocks form dome-flow complexes or thick units of ignimbrites in caldera/graben filling and take part in the emplacement of coeval intravolcanic/subvolcanic intrusions. Formation of a caldera or volcanic graben may, or may not be the last event in the evolution of the stratovolcano. Post-caldera or post-graben volcanic activity creates new volcanic cones, often with vents at marginal faults that cover the caldera/graben rim. Subsidence of large calderas associated with silicic magma chambers is often followed by resurgent updoming in their central part caused by emplacement of silicic magma underneath the dome. Such calderas are termed resurgent calderas [45, 257–259].

The Călimani caldera structure (Figures 1, 11 – area 7a, Figure 25), ca. 9 km in diameter, represents the final major volcanic episode of the Călimani stratovolcano [159], being the largest in the Călimani-Gurghiu-Harghita range, Romania. Its last products, mainly lava flows, partially cover the row of NNE-trending older stratocones in the west, the "Drăgoiasa dacite Formation" and the volcaniclastic deposits of the older Rusca-Tihu volcano and the "Lomaș Formation" in the east and south. The pre-caldera Călimani edifice consists of large-volume andesitic lavas. Flow directions of the lavas were dependent on local topography, with south- and east-directed slopes, originating from at least four independent vents. A huge volume of lava, estimated to be in excess of 10 km^3, was erupted in a relatively short period of time (ca. 300 Ka), as constrained by K-Ar data [157]. The caldera, with a current summit rim altitude of ca. 2000 m a.s.l., was a result of the collapse initiated by this large-volume effusive eruption. The horseshoe shape of the caldera is inferred to be related to a half-block tilting downward the southeastern part from a NE-SW oriented hinge, resulting in a trap-door type caldera. Post-caldera volcanism is represented by few andesitic stratovolcanic cones in the interior of the caldera followed by extensive hydrothermal alteration. A large monzodioritic-dioritic resurgent subvolcanic intrusion is exposed in an area of about 11 km^2 in the central part of the caldera. Three dacitic domes located on the caldera rim and outer slopes are also late post-caldera features.

Fâncel/Lăpușna in the northern part of Gurghiu Mts. (Figures 1, 11 – area 7b) is the second caldera of the Călimani-Gurghiu-Harghita range (Romania), formed at ca. 8 Ma on top of a former typical stratovolcanic cone during a large-volume Plinian-type explosive eruption of amphibole andesite composition. Its pumice fall and pumice-and-ash-flow deposits and their reworked counterparts are largely exposed to the north and east of the caldera [158]. Post-caldera activity includes intrusions in the southward-open semicircular caldera interior. The Bacta dome complex at its south-eastern periphery (see below, section 6.2.3.) can also be viewed (in one version of possible interpretations) as a result of the post-caldera volcanism; its age range is consistent with such a view.

An extensive caldera associates with a mostly effusive composite volcano in Mátra Mts. in Hungary (Figures 1, 11 – area 2c) [260]. The caldera,10 km in diameter, hosts a diorite porphyry intrusive complex with associated base-metal epithermal mineralization and two late stage rhyolite lava domes. A complex of andesite lava flows fill-

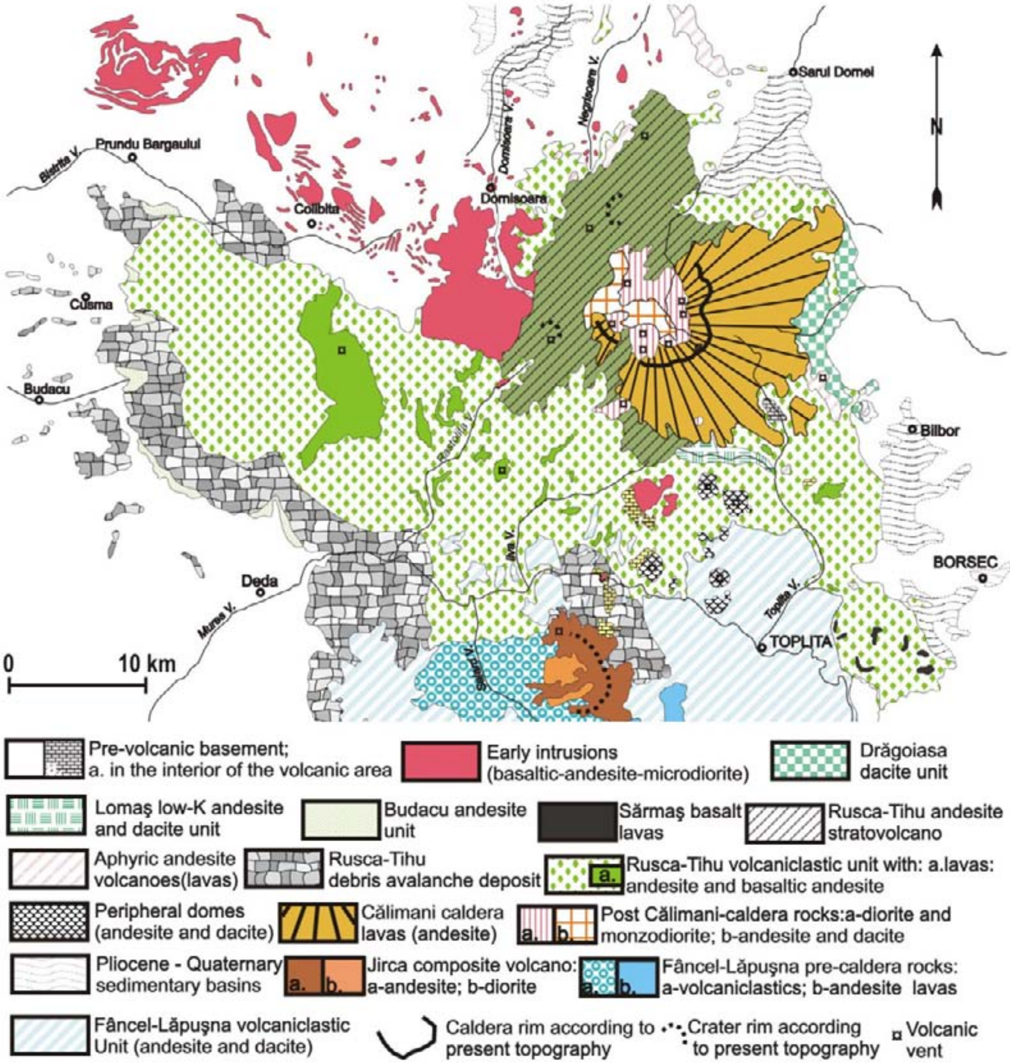

Figure 25. Simplified volcanological sketch of the Călimani and northernmost Gurghiu Mts. (according to Seghedi et al. [159]).

ing the caldera is affected by propylitic alteration.

The Poľana stratovolcano of the Sarmatian age (around 12.5 Ma? [241]) in the eastern part of the Central Slovakia Volcanic Field (Figures 1-2a, 13) shows relatively well preserved morphology with extensive caldera and central intrusive complex. No collapse features and associated volcanic activity has been observed. In this case the caldera is a result of extensive erosion. The stratovolcano rests on remnants of an older denudated andesite stratovolcano that hosts a small caldera associated with rhyodacite extrusive and explosive volcanic activity. Diameter of the caldera is about 3 km. Together they form a compound polygenetic volcano [241].

The Javorie stratovolcano in the eastern part of the Central Slovakia Volcanic Field (Figures 1-2a, 13) represents a stratovolcano with volcanic graben [242]. It is a large compound andesite stratovolcano 30 km in diameter with partially preserved morphology and a sectorially developed volcanic graben in its central and northeastern parts (Figures 13, 26). The volcano evolved in five stages that took place in the time interval of 14.4-12.3 Ma (Konečný and Pécskay, unpublished data): (1) formation of a large, dominantly effusive pyroxene and amphibole pyroxene andesite stratovolcano; (2) initial subsidence of the graben accompanied by effusive activity of mafic andesites giving rise to a complex of lava flows and hyaloclastite breccias; (3) continuing subsidence of the graben accompanied by emplacement of pyroxene-amphibole andesite to dacite extrusive domes and related pyroclastic/epiclastic breccias; (4) emplacement of diorite porphyry to monzodiorite stocks hosting advanced argillic alterations and Au-porphyry mineralization; and (5) post-graben activity giving rise to dominantly effusive stratovolcanic complex grading outward into accumulations of epiclastic volcanic breccias in the medial zone and epiclastic volcanic conglomerates and sandstones in the distal zone.

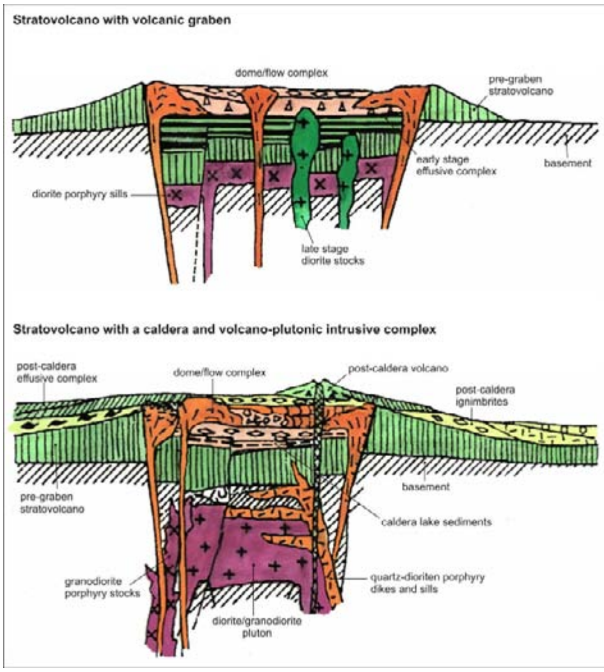

Figure 26. Schematic paleovolcanic reconstruction of composite stratovolcanoes with volcanic graben resp. caldera in the Central Slovakia Volcanic Field [219]. The Javorie stratovolcano (the upper one) is shown before the post-graben andesite stage. The Štiavnica stratovolcano (the lower one) is shown before the resurgent horst uplift and associated rhyolite activity.

The Štiavnica stratovolcano in the Central Slovakia Volcanic Field with its 50 km diameter and area of over 2000 km^2 is the largest volcano in the CPR (Figures 1-2a, 13, 26, 27). However, its primary morphology is not preserved due to extensive erosion and differential uplift and subsidence. The stratovolcano hosts a resurgent caldera 20 km in diameter and extensive subvolcanic intrusive complex. The Štiavnica stratovolcano evolved in five stages [200, 262, 263] during the time interval 15.9-10.7 Ma [2]: (1) formation of a large pyroxene and amphibole pyroxene andesite compound polygenetic volcano whose stratovolcanic complex in the proximal zone grades outward into accumulations of epiclastic volcanic breccias in the medial zone and shallow marine epiclastic volcanic conglomerates and sandstones in the distal zone; (2) break in volcanic activity and extensive denudation of the volcano, contemporaneous emplacement of subvolcanic intrusive complex of diorite, granodiorite, granodiorite porphyry and quartz-diorite porphyry, dominantly by the mechanism of underground cauldron subsidence; subhorizontal discontinuities among crystalline basement, Late Paleozoic and Mesozoic sedimentary rocks and volcanic complex were used as primary zones of detachment; the early emplacement of diorite in the northern part of the intrusive complex was soon followed by the emplacement of granodiorite pluton extending over the area of 100 km^2; subsequent emplacement of granodiorite to quartz diorite porphyry stocks and dike clusters took place around the pluton; emplacement of the extensive system of quartz-diorite ring dikes and sills reaching into the lower part of volcanic complex concluded its evolution; (3) subsidence and formation of a caldera 20 km across accompanied by voluminous extrusion of differentiated biotite-amphibole andesites giving rise to a dome-flow complex filling the caldera in thickness up to 500 m; (4) renewed explosive and effusive activity of undifferentiated andesites in the caldera and along its marginal faults giving rise to several post-caldera volcanoes of smaller size; (5) uplift of a resurgent horst in the central part of the caldera accompanied by emplacement of rhyolite dikes and epithermal veins at its faults. Thanks to the uplift of the resurgent horst, erosion has reached basement and exposed the subvolcanic intrusive complex (Figure 27). Association of the caldera with the pluton-size subvolcanic intrusion allows for a use of the term "volcano-plutonic complex".

6.3. Polygenetic volcanoes with shield morphology

This category of volcanic forms include those simple polygenetic volcanoes (as defined by Francis, [32]) that resemble shield volcanoes in their external shape while their rock composition and internal make-up do not differ from those of lava-dominated stratovolcanoes. MacDonald [28] has used for this type of polygenetic volcanoes the term "lava cones". A number of intermediate composition calc-alkaline volcanoes in the Călimani-Gurghiu-Harghita range (Figures 1, 11 – area 7) recorded as stratovolcanoes [202] display unusual topographic features. Instead of upward-concave outer slopes typical for stratovolcanoes, they have rather upward-convex profiles characteristic for shield volcanoes. Relevant examples are from North Harghita (Ivo-Cocoizaș) and South Harghita (Luci-Lazu) (Figures 1, 11 – area 7c). Both of them are lava-dominated in their cone facies portion, as their neighboring edifices display the typical upward concave stratovolcano morphology. Considering similarities between

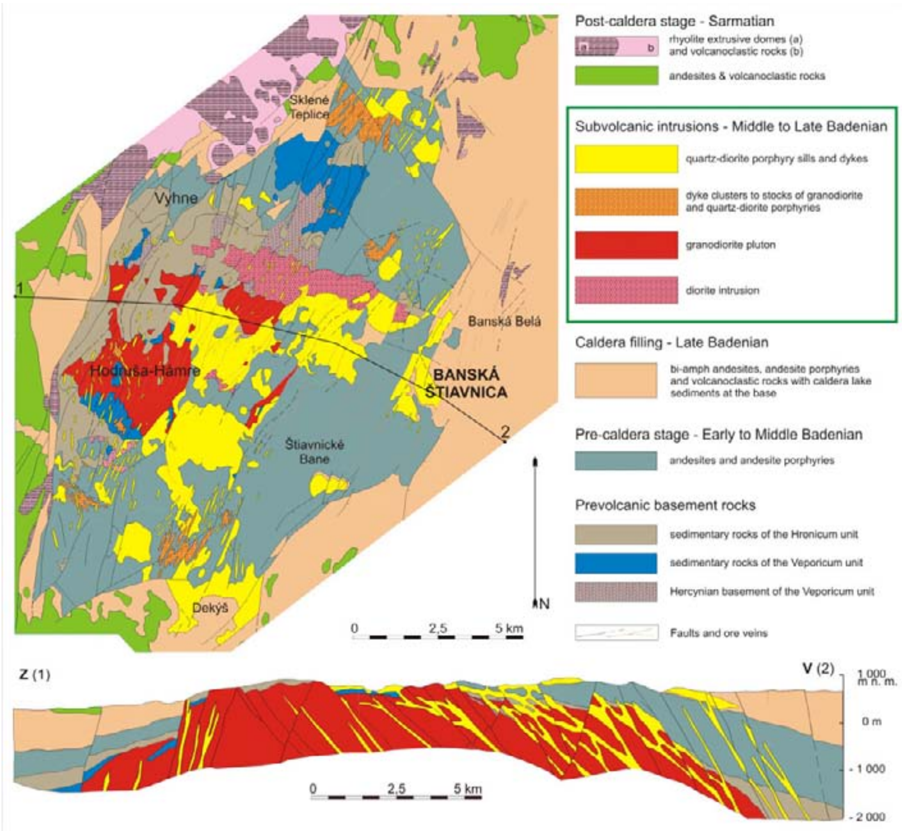

Figure 27. Volcano-plutonic complex of the Štiavnica stratovolcano in the Central Slovakia Volcanic Field. Modified after Lexa et al. [201].

the eruptive products of these composite volcanoes their contrasting morphology (stratovolcano versus shield morphology, respectively) still needs to be explained.

6.4. Dome/flow complexes

A dome/flow complex is a special type of compound polygenetic volcano. It is an assemblage of nested lava domes, dome-flows and related pyroclastic and epiclastic volcanic rocks spaced so closely in space and time that they are considered rather as one polygenetic volcano instead of a group of monogenetic volcanoes. Dome/flow complexes with relatively closely spaced vents and longer lasting recurrent explosive and/or extrusive activity fulfill the definition of the polygenetic volcanic form as defined by Francis [32], Davidson and De Silva [35] and Németh and Martin [22]. Dome/flow complexes are formed by viscous lavas of intermediate or felsic calc-alkaline composition.

6.4.1. Intermediate composition dome/flow complexes

Intermediate composition dome/flow complexes in the CPR occur as autonomous polygenetic volcanoes (Ciomadul volcano in South Hargita Mts., Figsures 1, 11 – area 7c) or as parts of compound andesite stratovolcanoes showing morphological individualization; that is the case of the dome/flow complex in the volcanic graben of the Javorie stratovolcano and dome/flow complex filling the Štiavnica caldera in the Central Slovakia Volcanic Field (Figures 1-2a and 13).

The 7.0-7.5 Ma old dome complex (Bacta) showing no crater remnants in its present day topography is located at the south-eastern periphery of the Fâncel-Lăpușna caldera (Gurghiu Mts., Figures 1, 11 – area 7b) [158]. It is questionable whether it represents an individual volcanic form or if it is part of the edifice of the larger Fâncel-Lăpușna volcano. Its polygenetic nature (interpretation favored by the authors) is suggested by the K/Ar age data ranging ca. 0.5 Ma which is larger than any possible monogenetic activity. However, given the limited resolution of the K/Ar dating method at such ages, the monogenetic nature of the dome complex cannot be ruled out.

Ciomadul (at the southern tip of South Harghita Mts., East Carpathians, Figures 1, 11 – area 7c), the volcano with the most recent eruption in the whole CPR, is a typical example of a dome complex resulted from extrusion of viscous (high-K dacitic) magma. Following a quite low frequency dome-forming eruption stage spread out over ca. 1 Ma, during which a central tightly-packed dome assemblage formed surrounded by isolated peripheral domes, two late-stage explosive events shaped the fresh-looking present day volcano topography whose prominent features are the two craters: the older Mohoş (filled with a peat-bog) and the younger Sf. Ana (hosting a crater lake). The oldest domes (Bálványos, Puturosul, Haramul Mic) are strongly eroded while those of more recent ages (e.g. Dealul Cetăţii, Haramul Mare, Ciomadul Mic) show their initial dome morphology intact. Talus breccias are present at Dealul Cetăţii. The larger and shallower Mohoş crater resulted from a phreatomagmatic eruption of unknown age (probably close to that of the Sf. Ana crater). The most recent eruption of (sub) plinian/phreatoplinian type occurred through the Sf. Ana crater sometime in the interval of 42-11 Ka, according to radiocarbon ages [264, 265]. The pyroclastic deposits of these explosive eruptions partly cover the previous dome rocks, while reworked tephra was transported farther away around the dome complex as debris flow deposits or redeposited in local lacustrine environments.

The dome/flow complex in the volcanotectonic graben of the Javorie stratovolcano (Figure 13, 26) is composed of amphibole-pyroxene and pyroxene-amphibole plugs, lava domes and dome-flows [242]. Marginal parts of domes grade into thick accumulations of crumble breccias. Among lava domes there are accumulations of block-and-ash pyroclastic flow deposits and mostly coarse epiclastic volcanic breccias.

The dome/flow complex filling the Štiavnica caldera (Figure 13, 26) is as extensive as the caldera, i.e. 20 km in diameter with thickness of up to 500 m. Extensive dome-flows, domes and rare plugs of biotite-amphibole andesites are accompanied by thick zones of crumble breccias, rare block-and-ash-flow deposits and mostly coarse epiclastic volcanic breccias [262]. Pumice flow deposits and reworked tuffs occur locally. Primary morphology of lava domes is obscured by extensive erosion.

The complex of Early Badenian garnet-bearing andesite extrusive domes and related volcaniclastic rocks in the Börzsöny Mts., Visegrád Mts. in Hungary and Burda hills in Slovakia (Figures 1, 11 – area 2b) evolved in a shallow marine environment [238, 239, 266]. Contact with water is reflected in extensive, often almost complete, brecciation of lava domes. Gravitationally unstable accumulations of crumble breccias enabled mobilization of slumps and subaqueous breccia flows whose deposits dominate in the complex, partially in the form of relatively low angle aprons around lava domes (compare Figure 14). In the distal zone, breccias wedge out in a succession of conglomerates and sandstones.

6.4.2. Felsic dome/flow complexes

Felsic dome/flow complexes may associate with calderas and may represent the central zones of silicic explosive volcanoes (ignimbrite shields) [45]. The dome/flow complex is not composed of domes and dome-flows only. Aprons of blocky breccias, block-and-ash pyroclastic flow deposits, debris avalanche/flow deposits, phreatomagmatic and phreatoplinian type pyroclastic flow, surge and fall deposits variably take part in their structure too (Figure 28). Mutual overlapping of edifices is a characteristic feature.

The dome/flow complex of the Middle to Late Sarmatian Jastrabá Formation in the Central Slovakia Volcanic Field (Figures 1-2a and 13) is associated with a subsidence of the Žiarska kotlina volcano-tectonic depression. A complex of overlapping edifices is up to 400 m thick and includes [134, 146]: (1) early rhyodacite domes with associated pumice flow/fall deposits and debris flow deposits; (2) a succession of phreatomagmatic/phreatoplinian tuffs and pumiceous tuffs; (3) overlapping edifices of extrusive volcanoes as defined in the part 1.3.2 above; (4) cryptodomes emplaced in the succession of tuffs; (5) a horizon of reworked tuffs, tuffaceous sediments and lacustrine siliciclastic deposits corresponding to a break in volcanic activity; (6) the youngest pumiceous fall-out tuff and overlying lava flow.

Silicic dome-flow fields showing recurrent activity occur in the Central Slovakia Volcanic Field (Figures 1-2a and 13) and Tokaj Mts. in northeastern Hungary (Figures 1, 11 – area 4a). Dome/flow fields in Tokaj Mts. are quite extensive [139]. Clusters of extrusive domes and dome-flows associate with aprons of breccias, often perlitic, reworked breccias and extensive accumulations of tuffs in the surrounding area. Only few modern volcanological analyses have been carried out in this region [148].

6.5. Polygenetic lava fields

In the CPR there are several occurrences of polygenetic lava fields of the intermediate calc-alkaline composition that do not show relationship with a polygenetic volcano of the stratovolcanic type. With few exceptions they take part in filling of volcano-tectonic depressions and we assume that they have been created by fissure eruptions along marginal faults of such depressions. As they occur in areas of subsidence, lava flows often associate with hyaloclastite breccias, autochtonous or reworked and rare hydromagmatic pyroclastic rocks. Effusive complexes in

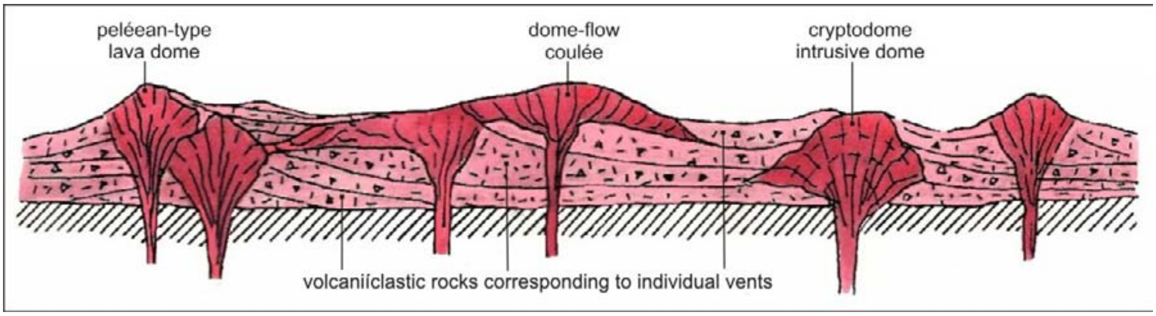

Figure 28. Scheme of a felsic dome/flow complex. Based on the structure of the Jastrabá Formation in the Central Slovakia Volcanic Field [219].

the filling of the Javorie stratovolcano volcanic graben and Kremnica volcano-tectonic graben in the Central Slovakia Volcanic Field (Figure 13) are described elsewhere. Rare necks, feeder dikes and extensive pyroxene andesite lava fields accompanied by hyaloclastite breccias make up the volcanic complex in Cserhát Mts. and in the eastern part of Mátra Mts. in Hungary (Figures 1, 11 – area 2c) [260, 267]. Quite extensive andesite lava fields with hyaloclastite breccias have been observed also in the northern part of Tokaj Mts. (Figures 1, 11 – area 4a). Individual lava flows are 15-50 m thick, coarse hyaloclastite breccias representing anywhere between 30 to 90% of their volume. An origin in the shallow marine environment is proven by interfingering with fauna-bearing marine sediments and peperitic breccias at lava/sediment contacts.

A lava field of this type extends over an area of ca. 80 km^2 in the northern part of Gutâi Mts., Romania (Figures 1, 11 – area 4f), where it represents the youngest part of a volcano-tectonic graben filling. The thickness of the effusive complex is variable, reaching 500 m in the center of the graben. The effusive complex evolved in a terrestrial environment. However, hyaloclastite breccias and phreatomagmatic pyroclastic rocks at its base indicate that its evolution started in a shallow subaqueous environment.

6.6. Felsic calc-alkaline caldera volcanoes and related ignimbrite fields

Calderas are large volcanic depressions, more or less circular in form with a diameter many times greater than that of the included vents. Their formation is related to some form of roof collapse over an underlying shallow magma reservoir related to large-volume eruptive events [45]. Calderas may occur on top of large composite or compound volcanoes or as volcanic forms independent of the central-type pre-existing volcanic form. The latter are called caldera volcanoes given the fact that their overall topography is dominated by the large (up to 20-40 km across) collapse-related depression. As a rule caldera volcanoes include an extensive ignimbrite field surrounding the caldera.

A felsic caldera volcano with associated ignimbrites occurs west of Gutâi Mts. in northern Romania (Figure 1, 11 – area 4e). The 15.4 Ma caldera-related rhyolite ignimbrites belong to the group of eroded volcanic forms with remnants of original morphology preserved [268]. The unidirectional caldera outflow can be traced at the surface and underground from the West to the East along the southern part of the Gutâi Mts. for 20 km. The original wedge morphology shows upwards and downwards displacements along the major faults crosscutting the southern part of the mountains (Figure 29). The distalmost outcrop of the ignimbrites is located towards the East, near the Firiza Lake. The incomplete caldera ring (with a diameter of approximately 10 km) was outlined by the south-western corner of the Gutâi Mts. and the eastern complex system of faults from the Ilba valley. Despite the lack of the visible caldera morphology, the caldera fill can be spatially located on the basis of numerous and reliable drill core data. It consists of ca. 350 m thick densely welded ignimbrites topped by several meters of ash fall tuffs and a ca. 350 m thick succession of reworked pyroclastic rocks, mostly volcaniclastic conglomerates and sandstones as well as pumice-rich layers interbedded with deep water sediments (Figure 29). The buildup of this syn-volcanic succession was triggered by caldera subsidence and the outcrops from all over the area show complex mass movements entraining unconsolidated pyroclastic rocks.

Large-volume eruptions of felsic calc-alkaline magma at the early stages of Neogene volcanic evolution in the CPR generated widespread ignimbrite fields in the intra-Carpathian regions. They are mostly covered by younger sedimentary rocks of the Pannonian Basin and are exposed in only limited areas at the basin margins, such as in the Bükk Foreland (Northern Hungary) (Figures 1, 11

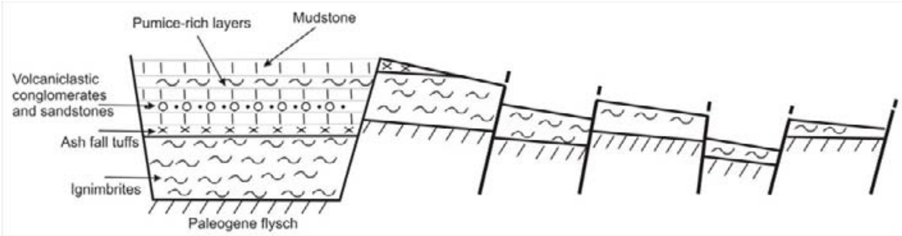

Figure 29. A model of the felsic caldera west of Gutâi Mts., Romania.

– area 2d), but encountered in many boreholes. These rhyolitic to dacitic ignimbrite sheets consist of numerous flow units and their reworked counterparts (e.g. debris flow deposits), as well as thin Plinian ash-fall deposits located commonly at the base of the ash-flow deposits. Both welded and non-welded ignimbrites (ash-flow tuffs) are recorded. Traditionally, three main phases of explosive volcanic activity resulting in three tuff complexes are distinguished within the Pannonian Basin in the Lower to Middle Miocene time interval. Their large volumes and widespread occurrence strongly suggest that they resulted from caldera-forming eruptions. Two caldera locations have tentatively been identified by using transport direction indicators and geophysical data in an area south of the Bükk Foreland [269]. Due to the complexity of the ignimbrite piles, it is very likely that the caldera-type volcanic centers were long-lived. Being buried under younger sediments, their detailed investigation is strongly limited; hence their actual volcanic forms and features of the presumed caldera volcanoes are as yet largely unknown.

A widespread tuff complex, traditionally named „the Dej Tuff", occurs throughout the whole Transylvanian Basin (between Apuseni Mts. at the west and the Călimani-Gurghiu-Harghita range at the east in Romania), encountered in outcrops along its rims and in boreholes in the basin interior. Most of the ca. 15 Ma old tuff complex up to 116 m thick consists of numerous submarine-deposited tuff layers of rhyolite composition resulting from reworked loose tephra deposited as subaqueous volcanic debris flow deposits and high- to low-concentration volcaniclastic turbidite deposits [270]. In only few outcrops primary slightly welded ignimbrites are found, suggesting the origin of the large-volume rhyolitic tephra in the Plinian type eruptions, related probably to a caldera formation. Detailed study of chemical sequences demonstrated the occurrence of at least three large eruptive events, while a regional grain-size distribution study revealed the position of the caldera-type eruptive centers outside the Transylvanian Basin in the NW direction at a distance of at most 110-120 km [270]. The caldera structure suggested a source area for the ignimbrites occurring in the western Gutâi Mts. [268] is a likely candidate for the ‚Dej Tuff" centers as well. In such a hypothesis, the west-Gutâi ignimbrites are the primary proximal land-deposited facies, while the „Dej Tuff" is the reworked distal submarine facies of the same large caldera-type polygenetic volcanic edifice.

6.7. Polygenetic intrusive forms – intrusive complexes

Closely spaced, usually co-magmatic intrusions form intrusive complexes. Individual intrusions are of the uniform or variable composition, mostly intermediate calc-alkaline, and in a lesser extent mafic and felsic calc-alkaline composition. Degree of their crystallinity depends on the cooling rate that in turn depends on the size of intrusion and depth of its emplacement. Rocks vary from basalts/andesites/dacites/rhyolites with glassy groundmass in the case of minor intrusions through basalt/andesite/dacite/rhyolite porphyry and gabbro/diorite/granodiorite/granite porphyry to equigranular gabbros/diorites/granodiorites/granites in the case of large stocks and plutons. Respecting their relationship with comagmatic volcanic activity we distinguish: 1) deeply eroded hypabyssal intrusive complexes with questionable relationship to volcanic activity; 2) subvolcanic intrusive complexes related most often to mature composite andesite stratovolcanoes; 3) volcano-plutonic intrusive complexes uniting pluton-size subvolcanic intrusive complex with a highly evolved polygenetic volcano into one magmatic system.

6.7.1. Hypabyssal intrusive complexes

Intrusions that should be considered as hypabyssal are represented by various size bodies. They are mainly laccoliths and/or associated stocks, sills and dikes (Figure 30). Columnar jointing of decimeter size is characteristic for the dikes and sills while the larger intrusions show irregular jointing. Most of the laccoliths are marginally surrounded by a complex sill and dike system. These large intrusions, reaching up to 10 km across, could be a result of a single phase or of multiple phases of intrusion. Since in the CPR the stock or laccolith intrusions usually pierce various basement lithologies, their timing could be so far established only using K/Ar dating for each area in the CPR (e.g. Pécskay et al, [2] and references therein). K/Ar age determinations on different phases suggest a 1-2 million year life-span for emplacement of a major intrusive complex. Hypabyssal intrusive complexes occur in the CPR primarily in flysch-type sedimentary rocks along the internal margin of the Carpathian accretion prism (Figure 1, 11 – areas 3 and 6), including crystalline basement in the Rodna massive of northern Romania (Figure 1-6), due to extreme uplift during the final stage of the Carpathian arc formation.

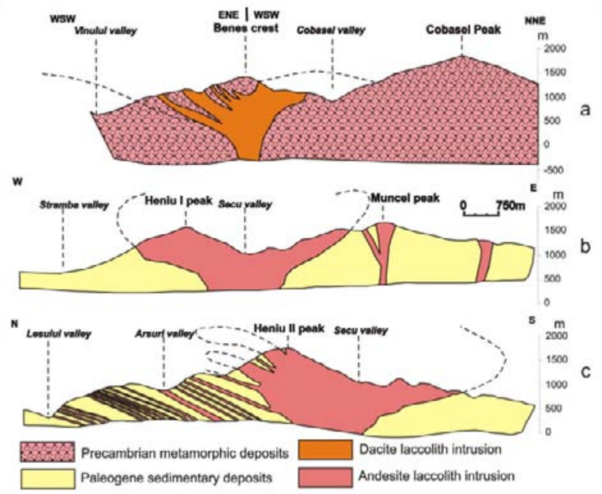

Figure 30. Profiles in several laccolithic intrusions in Rodna - Bârgău region: a. Dacite laccolith on the Cobașel valley in the Rodna Mts. (after Kräutner et al., [361]); b and c. Heniu andesite laccolith in the Bârgău Mts. (after Teodoru et al, [362]).

6.7.2. *Subvolcanic intrusive complexes*

Most of the stock or laccolith intrusions associated with volcanic activity are buried, only being reached by drilling or mining works in central zones of mature andesite stratovolcanoes. Unroofed subvolcanic intrusions are found only inside several calderas [158, 159, 271], erosionally deepened central volcanic depressions (e.g. Vârghiș, East Carpathians Figure 1-7c) and areas of larger post-volcanic uplift (e.g. Tisovec intrusive complex east of the Central Slovakia Volcanic Field, Figure 1-2a). Further comments are given in the section 6.2.4.

6.7.3. *Volcano-plutonic intrusive complexes*

The concept of a volcano-plutonic intrusive complex was developed by Cloos [272] who united for the first time the volcano and underlying pluton into one magmatic system. Nowadays we consider as volcano-plutonic those subvolcanic intrusive complexes that include pluton-size intrusions. Their relationship with a major polygenetic volcano is obvious. Volcano-plutonic intrusive complexes are usually complex, including dikes, sills, stocks and plutons of variable petrographic composition. While radially oriented dikes and cone sheets (curvilinear inward dipping dikes) tend to be mafic, ring dikes and central plutons show more silicic composition [273]. These compositional differences are explained by mechanism of their emplacement in brittle crust [274]. A space for cone sheets and radial dikes is made by uplift of the central block due to overpressure of incoming undifferentiated magma. A space for ring dikes and central plutons is made by underground cauldron subsidence of a cylindrical crustal block with outward dipping walls into underlying magma chamber filled by differentiated magma of lesser density. This mechanism is common with the subsidence of collapse calderas and that is a reason why volcano-plutonic complexes frequently associate with calderas.

The sole polygenetic volcano in the CPR that hosts a volcano-plutonic intrusive complex in the subvolcanic level is the Štiavnica stratovolcano in the Central Slovakia Volcanic Field (Figure 1-2a and 13). For this stratovolcano there are characteristic also a large caldera and a late stage resurgent horst and details are given in the relevant section 6.2.6.

6.8. Volcanic complexes in volcano-tectonic depressions/grabens

One of the specific features of intermediate composition volcanic complexes and/or fields in the CPR is their spatial and temporal coincidence with back-arc extension [23, 93, 213, 275, 276]. Due to this coincidence, extension-controlled grabens in volcanic areas show a spatial relationship with major volcanic centers and are filled by volcanic fields, hence the term "volcano-tectonic depressions/grabens". They are large, bigger than major volcanoes, hosting overlapping volcanic forms of variable composition related to several vents; such grabens include buried central and proximal zones of pre-graben volcanic forms along the graben axis [45], volcanic complexes related to dispersed volcanic vents and/or fissure eruptions and intrusive complexes; and creation of temporary lakes, reflected in the presence of hyaloclastites, hydromagmatic volcanic products, volcanisedimentary rocks and/or sedimentary rocks. There are many such buried volcanic forms in the Pannonian Basin that extend to a much larger area than outcropping volcanoes [1, 2, 229].

A N-S trending Kremnica volcano-tectonic graben is situated in the northern part of the Central Slovakia Volcanic Field (Figures 1-2a and 13). It is up to 15 km wide and amplitude of relative subsidence reaches 1,500 m [277]. Subsidence of the graben took place during the Late Badenian to Early Sarmatian time (14.0-12.5 Ma, [134]). A pre-graben volcanic complex represents denudation remnants of a large andesite stratovolcano with a diorite and andesite porphyry intrusive complex. The lower part of the graben filling, in thickness 400-500 m, is formed by a succession of mafic andesite lava flows with hyaloclastite breccias and hydromagmatic pyroclastic rocks. An effusive complex of thick amphibole-pyroxene andesite lava flows in thickness 300-500 m and two biotite-amphibole andesite dome-flows makes up the upper part of the graben fill. Centers of the post-graben Sarmatian andesite stratovolcanoes were situated at its marginal faults. The southern part of the graben was later a site of rhyolite volcanic activity giving rise to extensive dome-flow complex and of basaltic andesite volcanic activity giving rise to a tuff cone and several dikes and sills [134].

Despite the recent positive relief, volcanic forms in Tokaj Mts. of NE Hungary (Figures 1, 11 – area 4a) evolved in a major N–S trending graben [139]. Alternating andesite, dacite and rhyolite volcanic activity during the Upper Badenian and Sarmatian time (13.5-11.0 Ma, [2]), mostly from dispersed volcanic centers, has created a very variable volcanic complex. Remnants of andesite stratovolcanoes and effusive complexes with hyaloclastite breccias associate with rare dacite lava domes and lava cones, several rhyolite dome/flow complexes, extensive rhyolite ignimbrites and subordinate horizons of marine sediments.

Volcanic forms in the Apuseni Mts. (Figures 1, 11 – area 5) are spatially limited to NW-SE trending volcanotectonic grabens [278]. At the northwest, remnants of andesite stratovolcanoes and related volcaniclastic and volcano-sedimentary rocks fill the grabens [150]. At the southeast volcanic core complexes of altered andesites, intravolcanic/subvolcanic diorite porphyry intrusions and rare dacite lava domes hosting porphyry and/or epithermal mineralizations occur [213, 278].

The northern part of the Gutâi Mts. (Figures 1, 11 – area 4f) extending over an area of 35×25 km consists of a thick pile of volcanic products of the Pannonian age (10.6-9.0 Ma, [2]) filling a graben (Figure 1-4f). Drilling has confirmed thickness of a volcanic complex more than 1000 m above the Paleogene basement (Figure 31). Lava flows consisting of basaltic andesites and andesites seem to be predominantly on top, reaching up to 500 m thickness in the centre of the area and thinning laterally. Volcano-sedimentary succession underneath lava flows consists of volcaniclastic rocks, sedimentary deposits and subordinate lava flows. Autoclastic and hyaloclastite deposits represent lateral facies of coherent lavas. Thick and coarse heterolithic breccias consisting of phreatomagmatic deposits besides lavas and hyaloclastites suggest complex reworking processes associated with major erosional events. Among them, the debris flow deposits seem to be predominant. Fine-grained volcaniclastic rocks occur frequently in the succession, mostly interbedded with sedimentary deposits (sometimes with fossil flora content). They are interpreted as distal facies attributed to unidentified sources. The wide range of volcaniclastic deposits interbedded with sedimentary rocks and a lack of any evidence of volcanic centers suggest that fissure-fed eruptions in a submarine environment have been responsible for the origin of the complex succession of lavas and volcaniclastic rocks interbedded with marine sedimentary rocks. The thick lava plateaus at the top of the graben filling may be attributed to subaerial volcanism. No volcanic sources have been identified with the exception of a subaerial crater [279] and two inferred "calderas" tentatively outlined exclusively on a morphology basis.

7. Characteristics of volcanic fields / sub-regions

To discuss volcanic evolution and volcanic form distribution in volcanic fields of the CPR we have divided the region into seven sub-regions identified in the Figure 1 by corresponding numbers:

1. Styermark (a) + Burgenland (b) + Little Hungarian Plain (c) + Bakony-Balaton Highlands (d);
2. Central Slovakia Volcanic Field (a) + Börzsöny – Visegrád – Pilis – Burda (b) + Cserhát-Mátra (c) + Bükk foreland (d) + Southern Slovakia – Northern Hungary (e);
3. Moravia (a) + Pieniny (b);
4. Tokaj – Zemplín (a) + Slanské vrchy (b) + Vihorlat – Gutin (c) + Beregovo (d) + Oaş (e) + Gutâi (f);
5. Apuseni;
6. Ţibleş – Rodna;
7. Călimani (a) + Gurghiu (b) + Harghita (c) + Perşani (d).

Their spatial distribution and structure, including volcanic forms is given in the Figure 11.

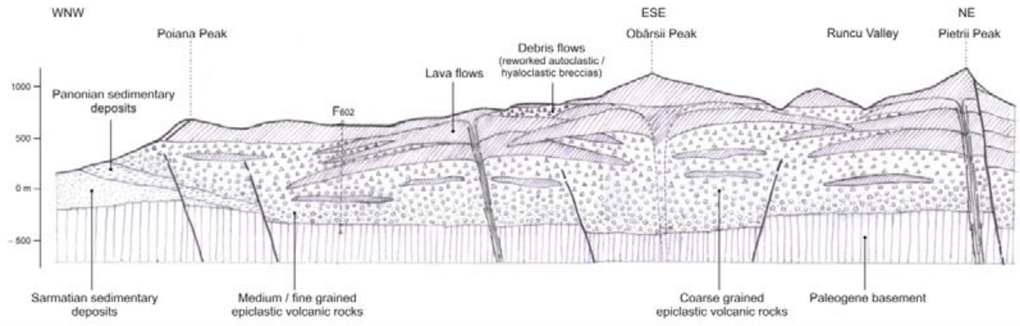

Figure 31. Section through the volcano-tectonic graben in the northern part of Gutâi Mts., Romania.

7.1. (a) Styermark + (b) Burgenland + (c) Little Hungarian Plain + (d) Bakony-Balaton Highlands

Volcanic fields in the western Pannonian Basin are Late Miocene to Pleistocene alkaline basaltic intracontinental fields [230, 280–282]. They consist of erosional remnants of maars, tuff rings, scoria cones, lava flows and fields [58]. Four individual fields have been separated on the basis of their location: Bakony-Balaton Highland Volcanic Field (BBHVF) in Hungary (Figures 1, 11 – sub-region 1d, Figure 9), Little Hungarian Plain Volcanic Field (LHPVF) in Hungary (Figures 1, 11 – sub-region 1c), Burgenland - Styria Volcanic Fields (BSVF) in Austria (Figures 1, 11 – sub-regions 1a, b) and Northern Slovenian Volcanic Field (NSVF) in Slovenia. Research on the volcanic history of the BBHVF and LHPVF has concluded their extensively phreatomagmatic origin [58]. In contrast, we know very little about the eruption mechanism, eruptive environment and style in the case of the Austrian and Slovenian volcanic fields (the wider Grad area in the Gorićko region of the Mura depression). Preliminary research however, indicates their similarity to those volcanic fields located in western Hungary [58]. Time and space distribution of monogenetic volcanism in the western Pannonian Basin seems to show a generally random pattern [1, 58].

The earliest known volcanic rocks of the Miocene series are located in Weitendorf (14.0±0.7 Ma) and at Gossendorf (13.0±1.0 Ma). Both of them are trachyandesites (absarokites?). Their ages are supported by the biostratigraphic record. The most reliable age however was determined on the Na-alkalic rocks of Pauliberg, from Burgenland, which is clearly part of the alkali basaltic volcanism in the western margin of the Pannonian Basin with an "isochrone age" of 11.5±0.7 Ma marking the onset of alkali basaltic volcanism about 10-13 Ma ago [281, 283].

The onset of volcanism is well established in the BBHVF about 8 Ma ago [282]. The main phase of the volcanism in the BBHVF falls in the 4 to 3 Ma time period, while in the LHPVF the peak of activity took place slightly earlier about 5 Ma ago [280, 284]. The volcanism ceased in the western Pannonian Basin in the Pliocene, however newly obtained ages suggest that the volcanism must have been active until the early Pleistocene, around 2.3 Ma ago or even less [230, 284].

At the onset of the eruption of monogenetic volcanoes in the western Hungarian fields, magma began to interact with a moderate amount of groundwater in the water-saturated Neogene fluvio-lacustrine sand beds [58]. As the eruptions continued, the craters grew both vertically and laterally and the repeated phreatomagmatic blasts fractured the deeper hard rock substrate around the explosion locus, giving the karst water (or water of any fracture-controlled aquifer) free access to the explosion chamber. The appearance of maar volcanoes and their deposits in the western Pannonian Basin are inferred to be strongly dependent on the paleo-hydrological conditions of the near-surface porous media as well as the deep fracture-controlled aquifer [58]. The seasonal variability of water-saturation of the karstic systems as well as the climatic influence on surface and/or near sub-surface water are recently considered as potentially controlling parameters of the style of explosive volcanism on the evolution of individual volcanic fields.

Shallow but broad maar volcanoes are inferred to have been formed due to phreatomagmatic explosions of magma mixing with water-saturated clastic sediments in areas where thick Neogene siliciclastic units build up shallow pre-volcanic strata (such as in the LHPVF, [88, 285]. Such volcanoes have often formed late magmatic features in their maar basins such as scoria cones and lava lakes. Today these volcanoes are preserved as lens-shaped volcanic successions usually capped by solidified lava lakes [88]. Pyroclastic successions of phreatomagmatic volcanoes of this type are rich in sand, silt and mud from sediments filling Neogene siliciclastic basins [88]. Deep-seated xenoliths are rare. In areas where the Neogene sedimentary cover was thin, deep maar crater formation has been inferred on the basis of the present day 3D architecture of phreatomagmatic pyroclastic rock facies. The abundance of sand and silt in the matrix of lapilli tuff and tuff breccia units with high proportion of angular accidental lithic fragments from deep-seated hard rock units suggests that these volcanoes must have had deeper fragmentation sites that allowed excavation of country rocks at deeper regions. The presence of abundant deep-seated xenoliths in such volcanic erosional remnants suggests that water must have been available in those zones in fractures. This type of maar volcano is interpreted to be developed in areas where relatively thin Neogene fluvio-lacustrine units rested on the Mesozoic or Paleozoic fracture-controlled (e.g. karst water-bearing) aquifer [231]. Most likely they were generated during times when the fractures were filled with water (e.g. spring).

In the western Hungarian monogenetic volcanic fields, each of the identified volcanic eruptive centers represents a proximal zone of a former volcano. Nearly all of the known volcanoes had at least a short period of phreatomagmatic activity in their vent opening stage as recorded in the preserved fine grained, accretionary lapilli bearing, massive to mega-ripple bedded pyroclastic rocks formed by pyroclasts deposited by base surges and phreatomagmatic falls associated with intermittent initial vent breccias [58].

In the central part of the BBHVF erosion remnants of scoria cones and shield volcanoes give evidence of a smaller impact of the groundwater and surface water in controlling volcanic activity [286]. The age distribution of scoria cone remnants suggest a peak in their formation about 3 Ma ago, which coincides well with a dryer period of the environmental history of the region suggesting a potential link of the large-scale climatic changes and eruption style variations over a long time period in this region [286].

Exposed diatreme-filling rocks in the erosion remnants in the western Pannonian Basin are rich in sedimentary grains as well as mineral phases that derived from Neogene shallow marine to fluvio-lacustrine sedimentary units in areas where such units are not preserved anymore suggesting their existence and a near intact sedimentary cover over the basement in syn-volcanic time [194, 287]. The general abundance of such clasts in the pyroclastic rocks also indicates the importance of soft substrate environment to where phreatomagmatic volcanoes were erupted commonly forming "champagne-glass" shaped maar/diatremes especially in the LHPVF. The presence of intra-vent peperite, subaqueous dome and/or cryptodome, shallow intrusions as well as hyaloclastite facies in craters indicate that maar/tuff ring volcanoes have been quickly flooded by ground and/or surface water, suggesting that they were erupted close to the level of the paleo-groundwater table.

The Neogene alkaline basaltic volcanic erosional remnants of the western Pannonian Basin are exposed from former subsurface to surface levels of maar-diatreme volcanoes. Today the lower parts of the exhumed diatremes are commonly covered by Quaternary talus flanks. The deepest levels of exposures are located in the western and the southern part of the area. In the western part of the BBHVF (Figures 1-1d and 9) the Neogene sedimentary cover was easily eroded, exhuming the diatremes. The most deeply eroded regions are those where no subsequent lava caps sheltered the volcaniclastic sequences. The Balatonboglár, Szigliget, Tihany Peninsula allow the study of the deepest level of the volcanic centers, exposing diatreme facies. Probably the eruptive centers of Balatonboglár (Boglár Volcano), Kereki-domb, Vár-hegy of Zánka, Hármas-hegy and Véndek-hegy represent the deepest exposed level of the phreatomagmatic eruptive centers [58] and are inferred to be exposed lower diatreme levels. In this area there are no exposed irregular shaped, wall-rock fragment rich dikes [78], like those widely reported from other monogenetic maar-diatreme volcanic fields such as e.g. Hopi Buttes in the southwest US [288]. Such deeper levels of exposures and preserved outcrops are more common in the northern part of the Pannonian Basin, in southern Slovakia and northern Hungary [191]. In the BBHVF, the Tóti-hegy and Hegyesd are good examples where such relationships between feeder dike and host diatreme filling pyroclastic rocks do exist, although their contact is not exposed. The best example of individual plug as exposed sub-volcanic feeder dike is Hegyes-tű where a small remnant of vent filling mixture of volcanic and siliciclastic debris is preserved and intruded by the plug, indicating, that explosive fragmentation preceded the formation of the basanite plug [281].

Most of the BBHVF buttes represent upper diatremes, similar to the Hopi Buttes [288] or Western Snake River subaerial volcanic centers [188, 289] in the state of Idaho, USA. Surface volcanic edifices are preserved in areas of low erosion, and include volcanoes produced by both phreatomagmatic and magmatic eruptions. These vents resulted from eruptions triggered by magmatic explosive fragmentation. Vent remnants of these volcanoes are concentrated in the northern and central part of the BBHVF (Kab-hegy, Agár-tető, Haláp, Hegyesd, Figure 9).

Erosion remnants of scoria cones are commonly strongly modified after erosion [94, 290], and their original volcanic landform can only hardly be recognized [95, 97, 291]. In spite of the general assumption of the fast erosion of scoria cones, there are remarkable well-preserved scoria cones known from the central part of the BBHVF [58, 114]. These scoria cone remnants are about 3-2.3 Ma old and still retain their original crater morphology and some part of their constructional edifice [114].

The preserved phreatomagmatic pyroclastic successions and their distributional pattern suggests that the original maar sizes in the western Pannonian region range from a few hundreds of meters up to 5 km in diameter (Fekete-hegy - 5 km; Tihany - 4 km; Bondoró - 2.5 km; Badacsony - 2.5 km). However, the largest centers probably represent maar volcanic complexes with interconnected large basins similar to some maars known from South-Australia and Victoria [292, 293]. The average maar basins are inferred to have been 1-1.5 km wide originally, which is within the range of most maars worldwide [51, 294].

The maar vents from the western Pannonian region appear as hybrids of phreatomagmatic and magmatic volcanic edifices and formed by initial maar or tuff ring forming events. The gradual exhaustion of water source to fuel the magma/water interaction led to "drier" phreatomagmatic, then pure magmatic fragmentation of the uprising melt, often building large scoria cones inside of the phreatomagmatic volcanoes similar to several examples from Eifel in Germany [295], Auckland Volcanic Field in New Zealand [296] or from Durango, Mexico [297]. The typical examples of such volcanoes are located in the southwestern part of the BBHVF, in the Tapolca Basin (Badacsony, Szent György-hegy, Hajagos-hegy, Fekete-hegy, Figures 9, 21).

Proximal base surge beds commonly contain abundant accidental lithics. In the eastern side (Tihany maar volcanic complex) the major accidental fragments are Permian red sandstone, Silurian schist (e.g. Lovas Schist Formation) Mesozoic carbonates (dolomites, marls, limestones), and Pannonian sandstone [58]. Commonly the large (up to 75 cm in diameter) fragments are Silurian schist or Permian red sandstone fragments. The matrix of the surge beds contains a high proportion of sand from the Pannonian sandstone beds. In the western side of the BBHVF and in the LHPVF the main accidental lithic fragments are from the Pannonian siliciclastic units. Both the large fragments (up to 25 cm in diameter) and the matrix are rich in sandstone fragments. In smaller proportion (less than 20 vol% of total accidental lithics), there are schist fragments and small carbonate fragments. In the middle part of the area (Fekete-hegy, Bondoró, Pipa-hegy) the Mesozoic carbonates are the major part of the accidental lithics (min. 85 vol% of total accidental lithics). In distal facies the base surge beds become finer-grained. Clear distal facies of surge beds are visible from the Tihany Peninsula. Characteristic surge features such as sandwaves, dunes, impact sags, and u-shaped valleys are common (Tihany Peninsula, Fekete-hegy, Bondoró, Szigliget, Figure 9). Fall deposits associated with surge beds are also common (Fekete-hegy).

Current research in the western Pannonian Basin's phreatomagmatic volcanoes has documented clear evidence of polycyclic monogenetic volcanism at Fekete-hegy [43], Bondoró [230] and Tihany [231]. These volcanoes are complex phreatomagmatic to magmatic volcanic edifices, erupted over a long time, leaving behind pyroclastic successions separated by well-marked unconformities.

The general features of the volcanic fields of the western Pannonian Basin are very similar to other eroded volcanic fields where magma erupted into wet environments such as Fort Rock Christmas Valley, Oregon [298], Snake River Plain, Idaho ...[87, 189, 291], Hopi Buttes, Arizona [288, 299–301], and Saar-Nahe (Germany) [184].

7.2. (a) Central Slovakia Volcanic Field + (b) Börzsöny – Pilis – Visegrád + (c) Cserhát – Mátra + (d) Bükk foreland + (e) Southern Slovakia – Northern Hungary

Evolution of volcanic activity and inventory of volcanic forms in individual sub-regions differ. The first three subregions represent products of intermediate composition; in the sub-region 2a felsic and rare alkali basalt volcanic activity also took place. In the sub-region 2d felsic volcanic forms are exposed and the sub-region 2e represents an alkali basalt monogenetic volcanic field.

The Central Slovakia Volcanic Field (CSVF) (Figures 1, 11 – sub-region 2a, Figure 13) is of the Badenian through Pannonian age (16.5-8.5 Ma) [2, 200]. It has evolved in the area of pronounced horst and graben structure. Volcanic rocks belong to the high-K orogenic volcanic suite. Scattered andesite (often garnet-bearing) extrusive domes and related extrusive and reworked breccias represent initial Early Badenian andesite volcanic activity (16.5-16.3 Ma). Two andesite pyroclastic volcanoes were subsequently formed in the SE part of the CSVF, while large composite andesite stratovolcanoes were built in the remaining parts of the CSVF during the Middle to Late Badenian time (15.5-14.0 Ma). During the Late Badenian time (14.0-13.0 Ma) evolution of the stratovolcanoes continued by the subsidence of grabens and caldera associated with activity of both – relatively mafic undifferentiated basaltic andesites/pyroxene andesites and differentiated volcanic rocks – hornblende and biotite-bearing andesites to dacites. Granodiorite, diorite and various porphyries were emplaced at deeper levels and basement. Rhyodacite domes and related pumice tuffs of the Early Sarmatian age occasionally occur. Renewed activity of less differentiated andesites during the Sarmatian time (13.5-12.5 Ma) formed at the south discontinuous complexes on the slopes of older stratovolcanoes, while in the northern part of the CSVF it formed new volcanoes, with centers situated on marginal faults of grabens (Figure 13). An extensive Middle to Late Sarmatian rhyolite volcanic activity (12-5-11.0 Ma) gave rise to a dome/flow complex and related volcaniclastic rocks in the western part of the CSVF along the N-S to NE-SW trending fault system. A small Pannonian age basalt /basaltic andesite stratovolcano at the North and scattered basalt flows and intrusions in the central part of the CSVF are the latest products of the calc-alkaline volcanism. Scarce nepheline basanite / trachybasalt volcanic activity during the Pliocene created two necks and a small lava plateau. A sole Late Pleistocene nepheline basanite scoria cone with related lava flows represents the youngest volcano in Slovakia (0.11 Ma, [302])

Volcanic forms of the Börzsöny – Pilis – Visegrád sub-region (Figures 1, 11 – sub-region 2b) evolved in three stages during the Early Badenian to Late Badenian time, around 16.0, 16.0-14.5 and 14.5-13.5 Ma respectively [239]. Phreatomagmatic explosive activity and dacitic extrusive domes of the first stage evolved in a shallow marine environment [266]. Subsequent andesite extrusive activity took place in the shallow marine to terrestrial environment. Edifices of individual vents mutually overlap. Extensively brecciated extrusive domes associate with aprons of coarse breccias and thick units of subaqueous breccia flow deposits, respectively block-and-ash pyroclastic flow deposits in the proximal-medial zone that grade outward into fine epiclastic volcanic breccias, conglomerates and sandstones in the distal zone. Horizons of reworked tuffs among units of breccias and epiclastic volcanic rocks imply periods of vulcanianc and/or phreatoplinian explosive eruptions. Strongly eroded remnants of pyroxene to amphibole-pyroxene andesite stratovolcano, overlying the extrusive dome complex in Börzsöny Mts., represent the third stage. Lava flows and block-and-ash pyroclastic flow deposits (implying a temporary growth of extrusive domes) and coarse epiclastic volcanic breccias of the proximal/medial zone grade outward into a succession of fine epiclastic volcanic breccias, conglomerates and sandstones in the distal zone. Deep erosion in the central zone exposed a core (vent) complex of altered andesites, andesite porphyry intravolcanic intrusions and diorite porphyry stocks [303].

Two stages of volcanic activity have been recognized also in the Cserhát – Mátra sub-region (Figures 1, 11 – sub-region 2c) [2, 267]. During the Early Badenian time (15.9-14.8 Ma), effusive activity of pyroxene andesites created in the Cserhát Mts. a complex of lava flows and hyaloclastite breccias including feeder dikes and rare necks. In Mátra Mts., andesitic volcanic activity started already in the late Lower Miocene time by emplacement of submarine lava flows and hyaloclastite breccias [304] among sedimentary rocks. Following deposition of distal facies ignimbrites at the Lower/Upper Miocene boundary (16.5 Ma, [2]), evolution of an extensive, dominantly effusive stratovolcano took place during the Lower Badenian time whose activity was concluded by subsidence of a caldera, emplacement of subvolcanic intrusions accompanied by hydrothermal processes and extrusions of rare rhyolite lava domes [260]. Sarmatian pyroxene andesite effusive activity (13.2-11.8 Ma, [260]) created a post-caldera effusive complex capping ridges of Mátra Mts.

A felsic ignimbrite field is exposed in the Bükk Foreland (Figures 1, 11 – sub-region 2d). It represents only a small part of widespread horizons covered by younger sedimentary rocks of the Pannonian Basin [269]. Ignimbrites represent 3 stratigraphic horizons with ages 18.5-17.9, 17.5-16.0 and 14.5-13.5 Ma respectively, corresponding to the lower, middle and upper rhyolite tuff horizons recognized traditionally in the Pannonian Basin [305, 306]. While rhyolite ignimbrites of up to 750 m in thickness (as recorded in boreholes) at inferred near-source areas represent the lower horizon, overall dacitic pyroclastic rocks up to ca. 500 m in thickness (as recorded in boreholes) represent the middle horizon, while the patchy outcropping and much thinner rhyolite ignimbrites and tuffs up to ca. 220 m in thickness (as recorded in boreholes) represent the upper horizon. Their large volumes and widespread occurrence strongly suggest that they resulted from caldera-forming eruptions. Two caldera locations have tentatively been identified by using transport direction indicators and geophysical data in an area south of the Bükk Foreland [269].

The northern Hungary – Southern Slovakia sub-region (Figures 1, 11 – sub-region 2e), sometimes referred to also as Nógrád-Gömer Volcanic Field, represents an alkali basalt monogenetic volcanic field (Figure 8). Beside alkali basalt volcanoes, garnet-bearing andesite laccoliths emplaced in the Early Miocene sedimentary complex are present in this sub-region. Alkali basalt volcanic activity took place during the Pliocene and Pleistocene time in 6 stages dated to 8.0-6.4, 5.4-3.7, 2.9-2.6, 2.3-1.6, 1.5-1.1 and 0.6-0.4 Ma respectively [160]. Two lava flows and two maars in the western part of the field belong to the oldest phase. Maar sediments are represented by diatomaceous clays resp. alginites, reflecting oxidative resp. eutrophic conditions in maar lakes [307]. Volcanic activity of subsequent phases took place in the terrestrial environment of gradually uplifted highlands [160, 233, 308]. The uplift contemporaneous with volcanic activity was responsible for the observed occurrence of older volcanic products on ridges (an inversion of relief) and younger volcanic products in valleys, as well as for generally radial orientation of lava flows following existing valleys. Lava flows are formed dominantly by massive lava with platy, blocky and/or columnar jointing. Their uppermost part is formed of vesicular lava of the aa or pahoe-hoe type. Spatter and/or scoria cones formed of agglutinates, agglomerates and lapilli tuffs are in the source areas of many lava flows. Other volcanic centers and/or sources of lava flows in more eroded parts of the region are represented by dikes and lava necks cutting underlying sedimentary rocks. Maars and tuff-cones are situated at a lower elevation and are generally younger (the older ones were eroded to the level of underlying diatremes). Early phreatomagmatic eruptions caused by contact of uprising magma with water saturated Early Miocene sedimentary rocks in the depth around 700-1000 m created maars surrounded by tuff-rings filled by temporary lakes [232]. Subsequently, due to a decreasing water/magma ratio at the level of the deep aquifer with time, a transition towards the Surtseyan type of hydromagmatic eruptions took place as lava interacted with water in maar lakes (instead of the deep aquifer), building up tuff-cones. Especially their inward dipping tuff successions have survived the subsequent erosion. The late stage in evolution of maars was marked by transition towards mixed Surtseyan-strombolian eruptions (mixed strombolian bombs with palagonite tuff in proximal base surge deposits), and finally towards Hawaiian eruptions creating small spatter cones, lava flows and/or lava lakes forming, nowadays, conspicuous cliffs with columnar jointing.

Diatremes represent exposed conduits of former maars (now eroded). Their filling corresponds to the above mentioned phases in maar evolution [232]. The early phreatic phase is represented by mega-breccias of the Miocene sedimentary rocks cemented by a sandy matrix with tuffaceous admixture. The following phreatomagmatic phase is represented by palagonite tuffs with sandy admixture and fragments of the Miocene sedimentary rocks, vesiculated chilled basalt, and maar sediments. The late phase transition towards strombolian eruptions was marked by increasing content of scoriaceous basalt fragments. The youngest basalt dikes cutting diatreme filling are feeders of closing Hawaiian-type eruptions. At the shallow level of diatremes there are usually crudely stratified and sorted subsided parts of the maar filling. The volcanic field includes also polygenetic maar-diatreme volcanoes. One of them, the Bulhary maar-diatreme volcano has been analyzed by Konečný and Lexa [169]. The volcano evolved in 5 stages: (1) maar formation by phreatomagmatic eruptions; (2) emplacement of a laccolith-like intrusive body into maar/diatreme filling; (3) creation of hyaloclastite breccias and phreatomagmatic tuffs with spatter due to a direct contact of magma with water in the maar lake; (4) several cycles of the Surtseyan and Hawaiian type eruptions; (5) final Hawaiian type eruptions giving rise to the capping horizon of cinder and spatter (Figures 20, 23).

7.3. (a) Moravia + (b) Pieniny

Eastern Moravia and Pieniny in Southern Poland (Figures 1, 11 – sub-regions 3a, b) are two areas where numerous small andesite intrusions occur as hypabyssal intrusive complexes in the environment of deformed flysch deposits that are a part of the Carpathian accretion prism. Emplacement of intrusions was post-tectonic. Intrusions occur in deeply eroded areas. It is probable that they have never been connected with surficial volcanic activity.
In eastern Moravia (Figures 1, 11 – 3a) high-K biotite-pyroxene-amphibole basalt to andesite (shoshonitic trachybasalt to trachyandesite) intrusions occur as sills, dykes and irregular intrusive bodies [309]. Their age is 13.5-11.0 Ma [2].
In the Pieniny area (Figures 1, 11 – 3b) pyroxene-amphibole andesite small- to moderate-size dikes and sills form a belt extending for 20 km [210]. Two phases of emplacement have been recognized: The first phase dike swarm is parallel with the Carpathian arc and segmented by transversal faults that have been used for emplacement of the second phase intrusions. At contacts with surrounding sedimentary rocks some of the dikes and sills show extensive brecciation of the hyaloclastite type, however peperitic breccias have not been observed due to already solidified nature of sedimentary rocks. Age of the intrusions falls in the interval of 13.3-10.8 Ma [210].

7.4. (a) Tokaj – Zemplín + (b) Slanské vrchy + (c) Vihorlat – Gutin + (d) Beregovo + (e) Oaș + (f) Gutâi

Four volcanic alignments parallel with the Carpathian arc are distinguished in the area of NE Hungary, E Slovakia and Transcarpathia (SW Ukraine) [10, 90, 91, 131, 132, 139, 216]:
(1) Bimodal andesite – rhyolite volcanic forms of the Late Badenian to Early Pannonian age (14.5-9.5 Ma, [2]) associated with a system of horsts and grabens Tokaj – Zemplín – Beregovo – Oaș south of the Transcarpathian Basin (Figures 1, 11 – sub-regions 4a, b, d, e). Volcanic forms evolved in the terrestrial and shallow marine environment. Characteristically, there are small andesite volcanic cones and effusive complexes with hyaloclastite breccias and hydromagmatic volcanic products alternating with extensive

rhyodacitic and rhyolitic pumiceous tuff horizons, less frequent dacite, rhyodacite and rhyolite domes, dome-flows and dome/flow complexes and horizons of marine sedimentary rocks. A larger andesite stratovolcano with a caldera and exposed volcanic core complex including a hydrothermal system occurs in the southern part of Tokaj Mts. [310].

In the Telkibánya area (northern part of Tokaj Mts.) volcanic activity started in the marine environment with deposition of clays by emplacement of dacitic to rhyolitic pumiceous density current deposits in thickness up to 400 m and emplacement of dacitic and andesitic subvolcanic intrusions [310]. Volcanic activity continued during the Sarmatian time by emplacement of rhyolitic pumiceous density current deposits alternating with tuffaceous clays and sandstones and andesite lava flows associated with hyaloclastite and peperite breccias. Subsequently an andesite to dacite composition stratovolcano formed including a caldera hosting late stage rhyodacite and rhyolite extrusive domes.

(2) Pyroxene andesite stratovolcanoes of the Early to Late Sarmatian age in the northern part of the Slanské vrchy mountain range (Figures 1, 11 –sub-region 4b) and southeastward buried under younger sedimentary rocks of the Transcarpathian Basin. Stratovolcanoes are of the dominantly effusive type with early explosive activity and core complexes of altered rocks due to emplacement of late stage subvolcanic intrusions. The largest one, Zlatá Baňa stratovolcano at the north includes diorite porphyry stocks associated with epithermal mineralization. Volcanoes at the southern part of Slanské vrchy started their activity in the shallow marine environment. At the base of volcanoes there are hydromagmatic pyroclastic rocks and their reworked equivalents.

(3) Mostly hornblende-pyroxene andesite extrusive domes and andesite/diorite porphyry shallow intrusions of the Middle Sarmatian age (around 12.5 Ma, [2]) in the alignment Kapušany – Vinné – Mukatchevo (Figures 1, 11 – sub-regions 4b, c). Diameter of individual bodies reaches up to 2 km. Extrusive lava domes associate with aprons of breccia and debris flow deposits in their surroundings. Intrusions are emplaced in flysch-type sedimentary rocks. Their form is close to laccoliths. A possibility that some of the intrusions represent cryptodomes cannot be excluded.

(4) Mostly pyroxene andesite stratovolcanoes of the Middle Sarmatian to Early Pannonian age (12.5-9.1 Ma, [2]) forming the conspicuous alignment of the Vihorlat and Gutin mountain ranges northeast of the Transcarpathian Basin (Figures 1, 11 – sub-region 4c). Volcanoes are of the stratovolcanic and dominantly effusive types, with a variable proportion of pyroclastic rocks, especially during the early stage of their evolution. The largest ones include core complexes of altered rocks and minor intrusions in the central zone. Some of the stratovolcanoes in the eastern part of the Gutin Mts. show late stage effusive complexes of mafic andesites. Dikes, sills and small irregular intrusions of andesite to rhyodacite composition occur in flysch along the northern limit of the volcanic range.

The various types of igneous rocks building up the Oaş and Gutâi Mts. (Figures 1, 11 – sub-regions 4e, f) were attributed to the felsic, extensional/back-arc type volcanism (ca. 15.4 Ma) and to the intermediate, arc type volcanism (13.4-7.0 Ma), respectively [311]. The volcanism was initiated on a basement consisting of several overthrust units composed of Paleogene flysch deposits belonging to the Pienides [312]. It developed simultaneously with the evolution of the Oaş and Baia Mare basins (located at the NE part of the Pannonian Basin) during the Badenian, Sarmatian and Pannonian ages, volcanic products being accordingly in contact relationships with coeval sedimentary deposits or components of widespread volcano-sedimentary successions.

The felsic volcanism is represented by the rhyolitic ignimbrites known as a marker horizon in the south-western part of the Gutâi Mts. The lithology and structure of the ignimbrite sequence suggest the caldera-related onset of volcanism in Gutâi Mts. The caldera was located in the south-western extremity of the mountains, where the thickest ignimbrite deposits are topped by a thick succession of reworked pyroclastic rocks of mass flow origin related to the ignimbrite volcanism, interbedded with deep water sediments [268, 313]. The whole succession described as the caldera fill counts ca. 700 m. A similar succession of rhyolitic volcaniclastic rocks of pyroclastic origin interbedded with deep water sediments was identified in drill cores in the central northern part of Oaş Mts. [314] lacking the primary pyroclastic deposits. The geological data were not sufficient to account for a certain type of volcanic source.

The intermediate volcanism started in Sarmatian (13.4-12.1 Ma, [315]) in two southern areas located in the East and West parts of the Gutâi Mts. It spread afterwards, during the Pannonian (12.0-7.0 Ma, [2]) towards the North, North–West and North–East. Mostly extrusive, it shows a calc-alkaline, medium K character, with typical subduction-related signatures and crustal contamination as the major petrogenetic processes [316]. Most of the volcanic forms identified in the Oaş and Gutâi Mts. belong to this intermediate volcanism although most of the volcanic deposits are largely overlapping or are buried and therefore, difficult to be attributed to a certain volcanic source or form. Intrusive forms have also been identified at the surface, in boreholes and mining works.

Common volcanic forms in both the Oaş and Gutâi Mts. (Figures 1, 11 – areas 4e,f) are the volcano-tectonic depression fillings (Figures 16, 31) [317]. They show extensive lava sheet/lava plateau usually interbedded with thick sequences of volcaniclastic and sedimentary deposits. Flat-topped, they suggest a fissure-control of the volcanism and an active subsidence developed throughout most of the time interval of the volcanic activity, triggering the subaqueous emplacement of part of the thick pile of volcanic rocks. The volcaniclastic deposits identified within these successions show evidence of magmatic and phreatomagmatic explosive phases associated with the non-explosive fragmentation, as well as with extensive reworking. Extrusive domes can also be identified on the basis of morphology, internal structure and associated volcaniclastic and sedimentary deposits. The volcanic forms in the Oaş Mts. show typical dome morphologies, ranging from small, flat-topped monogenetic domes or dome-flow to composite domes surrounded by sedimentary deposits, all of them accounting for subaqueous emplacement [133, 318]. Some of the volcanic structures show proximal fragmental deposits such as hyaloclastite lavas and hydromagmatic pyroclastic rocks passing to complex reworking successions interbedded with sedimentary deposits, mostly in a medial to distal position with respect to the eruptive centers. Similar volcanic forms also occur in the Gutâi Mts., most of them suggesting the same subaqueous emplacement. Along the southern border of the mountains there is an alignment of monogenetic and polygenetic domes showing complex relationships with the surrounding deposits [142]. Most of them are spatially connected with primary volcaniclastic rocks such as hyaloclastite and pyroclastic deposits and reworked volcaniclastic deposits of mass flow origin consisting of both explosive and non-explosive volcanic debris. Debris flow deposits are largely represented and probably connected to other mass movements (such as debris avalanches) related to major destructive phases of the volcanic structures.

Effusive cones are sparse volcanic forms in the Oaş and Gutâi Mts.: e.g. Rotundu volcanic edifice, in the northern part and Mogoşa volcano in the southern part of Gutâi Mts. [317]. They consist of lava flows passing laterally into in situ and reworked autoclastic and hyaloclastite deposits. Mapping the volcanic products belonging to these structures is often difficult because they overlap with other products belonging to unknown volcanic centers.

Subvolcanic and shallow–level intravolcanic intrusive rocks of irregular shapes and various sizes (from tens of meters up to 6 km long) developed on over 3000 m vertical extent (based on drill core data). Their morphologies suggest dykes, sills and apophysis of laccolites crosscutting the Paleogene flysch-type basement, the Neogene sedimentary deposits, as well as the volcanic suite. Hundreds of small intrusions crop out mostly in the south-eastern part of the Gutâi Mts. The main intrusive magmatism developed contemporaneously with the paroxysm of the volcanism in Oaş and Gutâi Mts. (11.8-9.2 Ma). Important epithermal ore deposits are associated with the intrusions in the central part of the Oaş Mts. and in the southern part of the Gutâi Mts., respectively.

7.5. Apuseni

Most of Neogene calc-alkaline magmatic rocks in the Apuseni Mountains (Figures 1, 11 – sub-region 5) crop out in a WN-ESE graben-like intra-mountain basins, of which the ca. 100 km long Zărand-Brad Basin is the longest. Its evolution is related to the brittle tectonics during the Paleogene-Neogene interval [319, 320] as a consequence of its behavior as a solitary rigid lithospheric block (Tisia block). The horst/graben structures are mostly visible along the western edges of the Apuseni Mountains (Figure 1) as a consequence of the Neogene extensional development of the neighboring Pannonian Basins [321, 322], as well as of the translational and clockwise rotational movements of the Tisia block [323–325]. Development of the Neogene volcanism was closely related to this extensional evolutionary stage of the Apuseni Mountains, as part of the Tisia block [5, 213, 275, 276]. The Békés Basin represents a prolongation of the Zărand basin to the west beneath the Pannonian Basin and it records buried igneous rocks as well (Figure 11 – sub-region 5) [326–328]. Dominant NW-SE oriented development of the igneous activity is in connection with coeval Miocene sedimentation in the extension-controlled basins [278, 329].

The knowledge of the volcanic forms in the Apuseni Mountains was closely related to mining and exploration works of porphyry-copper, base-metal and gold-silver ore deposits since pre-Roman times [330, 331]. The volcanic rocks are associated with sedimentary deposits, mainly Sarmatian, within the NW-trending tectonic graben-type basins, widening toward NW along normal and listric fault systems [323]. The outcrops of the oldest volcanic rocks are aligned along the western margins of the largest basin (Zărand Basin) and are represented by isolated lava domes and various volcaniclastic deposits dominantly characteristic for medial or distal ring plain facies. Association of reworked fine volcaniclastic deposits, sometimes inferred as fallout tuffs, with terrigeneous deposits also generated volcano-sedimentary deposits [329]. The first obvious volcanic structure is a composite volcano situated at the edge of the basin [150]. This partly eroded composite volcano was firstly described by Berbeleac et al. [332, 333] as a collapse-type caldera and was named after a town situated inside the eroded edifice. The edifice is quasi-symmetric (16-18 km in diameter) with a central eroded vent area (9×4 km). The volcano was built up in two sub-aerial phases (14.0-12.5 Ma and 11.0-10.0 Ma) from successive eruptions of normal calc-alkaline pyroxene and then amphibole pyroxene andesite lavas and pyroclastic rocks. In the initial phase, scattered individual volcanic lava domes associated marginally with lava flows and/or pyroclastic block-and-ash flows were formed. The second phase attests a succession of strombolian to sub-plinian explosive eruptions and extrusion of volcanic domes from a central vent area, associated with secondary debris flow deposits at the periphery. Several andesitic-dioritic bodies and associated hydrothermal and mineralization processes occur in the volcano central core complex area [199, 333]. Around the volcano at its western and eastern parts the ring plain facies consists of volcanic epiclastic, terrestrial detritic deposits and coal.

Towards the south-east of the largest Zărand basin, another eroded volcanic structure (Caraciu -composite volcano?) bordering the south-west flank of the basin was recognized. Its central part is a dome-like extrusion of amphibole andesite, well preserved morphologically, that is surrounded by older lava flows and volcaniclastic deposits cut by dikes and irregularly shaped intrusive bodies or breccia [334].

The series of several eroded basins at the eastern periphery of the Apuseni Mts. are shallower than the Zărand – Bekes basin, and are aligned from NNE to SSE. They are associated with four calc-alkaline and adakite-like calc-alkaline volcano-intrusive areas (Figures 1, 11 – subregion 5) The northernmost and southernmost areas are characterized mostly by individual volcanic centers and intrusive bodies, while two central areas show multiple vent volcano-intrusive complexes. The volcano-intrusive relationships are well described based on numerous mining works in the area [218, 227], as well the age relationships. The volcanic rocks erupted in successive events between 14.7-7.4 Ma [157, 213, 278]. Several calderas or craters have been described here in earlier works [203, 218] although they are not properly justified. With few exceptions modern volcanological studies are missing. Recent intense exploration of the Roșia Montana Au-Ag deposit (northernmost basin) evidenced an eroded volcanic diatreme of dacite composition. The funnel shaped edifice formed by phreatomagmatic fragmentation, subsequently followed by a couple of plug protrusions. During the last phase, various intrusive hydrothermal breccias cut all the assemblage followed by extensive hydrothermal alteration and mineralization processes [215, 335, 336].

Lavas are typical for calc-alkaline andesites and extrusive dome flows for the adakite-like calc-alkaline andesites and dacites. Such kinds of lava domes cover volcaniclastic deposits and are better preserved in the morphology (e.g. Cetraș-1081 m, Buha, 902 m in the south-central basin) [218] suggesting that they were much resistant to the erosion than the prior lavas and volcaniclastic deposits (Figure 32). The most mafic basaltic andesites at Detunata (northernmost basin) shows two pipe-like columnar-jointed sub-circular necks that stand ca. 100 m above the surrounding topographic surface [203]. The intrusive bodies are various: dikes, domes, sills, micro-laccoliths or intrusive breccias; only some of them represent the root area of the volcanoes.

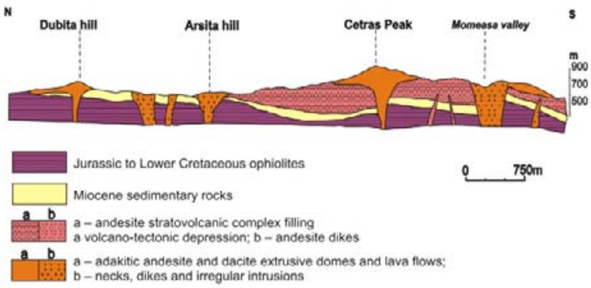

Figure 32. N-S profile in the eroded Sacarâmb basin (after Berbeleac [364] with modifications).

7.6. Ţibleș – Rodna

The metamorphic and sedimentary basement was pierced by shallow exhumed calc-alkaline hypabyssal intrusions showing various compositions as basalts-basaltic andesites-andesites (some garnet- bearing)-dacites and rhyolites in the 11.5-8.0 Ma interval [207, 337, 338]. Their emplacement was controlled by sinistral transtensional stress regimes at 12-10 Ma along an transcurrent fault system and oblique convergence of Tisia–Dacia with the NW–SE striking European margin [339]. The spatial distribution of intrusive bodies however is not along the main trace of the transcurrent fault. They were controlled by secondary conjugate extensional faults (NW-SE and NE-SW) located both to the North (Toroiaga) and South (Poiana Botizei, Ţibleș, Rodna-Bârgău) of the main fault trace (Figures 1, 11 – area 6). Even a detailed assessment on the estimation of intrusion depths is still missing, however the recent integrated thermochronological (fission track and (U–Th)/He analysis) data suggests an important amount of exhumation of ca 5-7 km in areas with subvolcanic bodies between 13-7 Ma in a similar time interval with dated rocks [211]. A possible relationship with volcanism especially of the larger bodies (∼10 km across) cannot be entirely excluded. Such bodies may represent magma chambers to feed volcanism on the surface, as proven also by the large amount of cognate enclaves in the large basaltic andesite - andesite intrusions that indicate clinopyroxene and amphibole fractionation typical for crustal magmatic chambers [340]. Volcanic deposits possibly emplaced on the surface may have been eroded away completely due to the strong uplift. The shape of the bodies is laccolith–like, no matter if they are intruding in metamorphic or Tertiary sedimentary deposits (Figure 30). They present laterally a swarm of sills and dikes. Some large intrusions are pierced by younger dikes of different composition [207].

7.7. Călimani + Gurghiu + Harghita + Perşani

The Călimani-Gurghiu-Harghita volcanic range (CGH – Figures 1, 11 – sub-region 7) represents the south-eastern segment of the Carpathian Neogene mostly calc-alkaline volcanic arc. It consists of a SE-striking row of closely spaced stratovolcanoes developed in the 11- <0.05 Ma time interval. Its peculiar time-space evolution pattern particularizes CGH with respect to all other volcanic areas in CPR by displaying an obvious along the range migration of volcanic activity trough time from NW to SE (Figure 2). The overall gradually decreasing height, width and volume of volcanic edifices in the direction of migration, enhanced along the terminal South Harghita sub-segment, points to a progressively diminishing volcanism involving decreasing magma volumes and output rates [341–344]. The volcanic edifices, most of which are typical stratovolcanoes, a few polygenetic composite volcanoes with shield topography, some peripheral isolated monogenetic domes or polygenetic dome complexes, show a general facies distribution in a concentric yet asymmetrical arrangement with a central core-complex facies (including shallow, in places unroofed intrusions and related hydrothermal alteration areas), a lava-dominated proximal cone facies and peripheral medial to distal volcaniclastic ring-plain and fluvial facies (Figure 11 – sub-regions 7a,b,c). The asymmetrical facies distribution (the volcaniclastic ring-plain facies is much more developed westward) is strongly influenced by eastward buttressing by the Carpathian fold-and-thrust belt and by axial buttressing by neighboring volcanoes. Because the tight spatial packing of composed volcanoes, medial facies volcaniclastic deposits are merged or interfingered at neighboring volcanoes and often cannot be distinguished on the maps. Cone-facies-ringplain facies boundaries are prograding during volcano edifice growth; hence frequently higher-position lava flow units overlap lower-position volcaniclastic sequences at cone peripheries causing earlier misinterpretation of the overall volcanic structure [202].

Long-lived large-volume volcanic edifices became unstable during volcano evolution and buildup. In two cases (Călimani caldera and Fâncel-Lăpuşna volcanoes) the instability was resolved by caldera formation (Figures 1, 11 – sub-regions 7a, 7b and Figure 25) [158, 159]. Lateral edifice failure and related debris avalanche formation operated as an alternative re-equilibration process, as seen at two other volcanoes (Rusca-Tihu in the Călimani Mts. - Figure 25 and Vârghiş in the North Harghita Mts., Figures 1, 11 – sub-region 7c) [8, 159]. Still other volcanoes reequilibrated by interacting with their basement, which contained plastically deformable rocks. Volcano spreading, a particular type of volcano-basement interaction, has been pointed out for a large group of volcanoes in the Gurghiu and North Harghita sub segments of CGH. This process consists of subsidence of the central part of the edifice, lateral spreading and peripheral uplift as the main volcano deformation features. Basement deformation include squeezing out of the plastic layer (Middle Badenian salt in our case) from below, its lateral displacement and consequent enhanced salt diapirism, formation of detachment surfaces and lateral displacement of tilted basement blocks along it, as well as outward-verging reverse fault swarms at periphery [345]. Due to the buttressing effect to the east, north and south, as well as due to the different basement rock rheologies (only brittle lithologies in the east, and brittle plus ductile lithologies in the west) volcano spreading was sectorial and asymmetrical with preferential westward orientation (toward the Transylvanian Basin). Volcano spreading has not been reported as yet from other volcanic areas in CPR. In the Perşani Mts. (Figures 1, 11 – sub-region 7d) the volcanism took place in two stages (1.2-1.5 Ma, and ca. 0.65 Ma) building up small monogenetic volcanoes or fields with maars, scoria cones and lava flows in an area of ca. 22×8 km [117, 346]. The first stage opened with phreatomagmatic explosive activity followed by a less energetic explosive strombolian or effusive phase [158]. Thinly bedded pyroclastic deposits with plan-parallel, ondulatory or cross lamination and frequent bomb-sags are testimony of near-vent phreatomagmatic explosion-derived dilute density currents and co-surge fall-out deposition resulting from the interaction of ascending magma with a shallow aquifer. A number of maar and tuff-ring volcanoes were generated [158]. The second stage was initially phreatomagmatic but shortly turned to dominantly strombolian, combined with lava outpourings. Lava spatter (agglutinate) deposits and lava flows are topping various first-stage deposits of the largest maar [346]. Thin lake sediments with occasional tephra intercalations or palesoil separate the above mentioned sequences. Abundant ultramafic xenoliths [347] and the presence of amphibole megacrysts are obvious features of the Perşani alkali basaltic field. Alkalic basaltic rocks from Perşani Mts. are similar to other continental intraplate alkali basalts and plot in the trachybasaltic field [348], but have distinctively higher La/Nb and lower Ce/Pb ratios than the other alkaline basalts in the Pannonian Basin, suggesting a subduction-related component in their source region [23, 348, 349].

8. Discussion

Volcanic forms in the Carpathian-Pannonian Region show a great variability in the space and time that reflects such essential aspects as geotectonic setting, magma composition, rate of magma production, magma volume, environment of volcanic activity, age of volcanic activity and degree of erosion. Some of these aspects are interrelated, e.g. geotectonic setting with magma composition, degree of erosion with age and geotectonic setting, etc. Based on information presented in the previous sections of the paper we shall discuss these aspects.

Geotectonic setting and its changes with time were fundamental factors governing magma composition, volume and rate of production and, indirectly, volcanic forms and their associations in volcanic fields. Based on published reviews [3, 5, 24, 25, 350–353] volcanic activity took place in the arc type and back-arc type geotectonic settings, both of them related to the Neogene subduction in front of the advancing Carpathian arc. The back-arc geotectonic setting changed its parameters with time. During the Lower Miocene time (up to the early Middle Miocene time in the eastern part of the CPR) the initial stage back-arc extension took place in the area with a thick continental crust and magmas reaching the surface in large volumes were of felsic calc-alkaline composition. Naturally, corresponding volcanic forms were caldera volcanoes with associated extensive ignimbrite fields and related dome/flow complexes. During the Middle Miocene time the advanced stage of back-arc extension took place causing thinning of the crust. Magmas reaching the surface were mostly of intermediate calc-alkaline composition (andesites and minor basaltic andesites). However, differentiation of these magmas in crustal magma chambers by AFC processes led to magmas of dacitic composition and related crustal anatexis even to smaller volumes of felsic magmas. Corresponding volcanic forms were dome/flow complexes and extensive composite andesite stratovolcanoes forming volcanic fields associated with the basin & range (horst/graben) type structures (Central Slovakia Volcanic Field, Börzsöny-Visegrád, Cserhát-Mátra and Apuseni sub-regions, Figure 1, 11 – sub regions 2a, 2b, 2c, 5). A long-lasting supply of magma created compound polygenetic volcanoes with multiple evolutionary stages involving formation of calderas, emplacement of subvolcanic intrusive complexes, evolution of resurgent domes and evolution of dome/flow complexes of intermediate to felsic composition. Owing to relationship with back-arc extension, volcanoes do not show linear alignment. Their spatial distribution is controlled by major extension faults and volcano-tectonic depressions. Since the Late Miocene time (in the west) to Pleistocene time (in the east) extension in the Pannonian Basin was governed by continuing subduction in the Eastern Carpathians. An extension environment created conditions for the generation of local asthenospheric plumes responsible for generation of magmas of the Na-alkalic and ultrapotassic composition. Low magma production rates caused the evolution of monogenetic volcanic fields. Interestingly enough, there is the Perşani alkali basalt monogenetic field whose activity was contemporaneous with neighboring calc-alkaline andesitic volcanic activity in the Southern Hargita mountain range (compare sub-regions 7c and 7d in the Figure 11).

The arc-type geotectonic setting was also variable in space and time [1, 3, 5, 24, 25, 284, 350–354]. First of all, subduction involving roll-back and verticalization before the final slab break-off migrated along the Carpathian arc from the west eastward. While in the west it started at the early Lower Miocene and finished at Middle Miocene, in the east it started in the Middle Miocene and has not finished completely yet. This is reflected in the progressive age pattern of volcanic activity (Figure 2). Arc-type magma generation took place only for a very short time during the final slab break-off with the exception of the central segment (Slanské vrchy and Tokaj Mts. at the west to Gutâi Mts. at the east), where magma production depth had been reached by the subducting slab before its verticalization and final break-off. Volume and composition of magmas along the arc are variable. In the western segment volumes of magma produced were almost negligible and volcanic arc extends over the internal part of the accretion prism with a strong uplift. Corresponding "volcanic" forms in this case are relatively small hypabyssal intrusive complexes of high-K andesites in eastern Moravia (Figures 1, 11 – sub region 3a) and Pieniny area in southern Poland (Figure 1, 11 – sub-region 3b) [210, 309]. A similar geotectonic setting is characteristic of the so called "subvolcanic zone" between the Gutâi and Călimani mountain ranges in NE Romania (Figures 1, 11 – sub-region 6). However, production of magma was higher and hypabyssal intrusive complexes are quite extensive, including dikes, sills, laccoliths and subordinate irregular intrusions [337, 338]. Evolution of high-K calc-alkaline magma in shallower magma chambers lead also to compositional/petrographic variably from basalts to rhyolites (gabbro porphyry to granite porphyry). Due to a considerable post-volcanic uplift no volcanic rocks are preserved and their original presence is doubtful. As mentioned above, in the central segment, the subducting slab reached the magma-generation depth before its verticalization and final break-off. This is reflected in migration of the volcanic arc towards the subduction zone with time and separation of volcanic arcs by the Transcarpathian inter-arc basin. Alignment of volcanoes is mostly parallel with the Carpathian arc. A conspicuous N-S apparent alignment of volcanoes in the Slanské vrchy and Tokaj mountain regions (Figures 1, 11 – sub-regions 4a and 4b) follows the eastern margin of the ALCAPA continental microplate, however, volcanoes are of different age and have their counterparts buried under younger sedimentary rocks of the Transcarpathian basin. For the southernmost alignment Tokaj – Beregovo – Oaş (Figsures 1, 11 – sub-regions 4a, 4d, 4e) at the boundary of the arc and back-arc domains there is a characteristic bimodal suite of andesitic and rhyolitic volcanism with dacites as products of differentiation and mixing. Due to the evolution of volcano-tectonic grabens, a part of volcanic activity took place in the shallow marine environment. Corresponding volcanic forms of andesitic volcanic activity are mostly small, dominantly effusive stratovolcanoes, effusive complexes associated with hyaloclastites and hydromagmatic products and minor andesite/diorite porphyry intrusions. Dacitic volcanic activity created a number of extrusive domes with associated breccias. Rhyolitic volcanic activity created dome/flow complexes and solitary extrusive domes and cryptodomes and extensive ignimbrites. For the shallow marine environment there are characteristic pumiceous density flow deposits and reworked pumiceous tuffs [10, 133, 310, 317, 318]. Larger volumes of mostly mafic medium-K magmas are characteristic for the younger volcanic alignments northeastward (Slanské vrchy – sub-region 4b and Vihorlat-Gutin – sub region 4c). Corresponding volcanic forms are stratovolcanoes sensu stricto and dominantly effusive stratovolcanoes, often with late stage emplacement of subvolcanic intrusions. Dacite and rhyolite extrusive domes are sporadic with an exception of the alignment of Kapušany-Vinné-Mukatchevo (Figure 11 – between sub-regions 4b and 4c). Minor intrusions occur also in underlying flysch deposits. A large volume of mostly mafic magmas was characteristic also for the eastern segment of the volcanic arc – alignment of volcanoes Călimani-Ghurghiu-Hargita. Volcanic activity at this segment shows a pronounced migration with time (Figure 2) [1, 2] and relatively short duration at each part – what is explained by magma generation stimulated by a progressive slab break-off starting at the north and still going on at the south (the Vranca seismic zone). Corresponding volcanic forms are polygenetic stratovolcanoes with calderas and late stage intrusions at the north that gradually lose complexity and volume going southward [338, 342, 343]. A more silicic magma of the youngest volcanic activity at the southern tip of the volcanic alignment resulted in the formation of a dome complex.

An interesting aspect of andesitic volcanoes in the arc-type geotectonic setting is their density along the arc. While there are segments with very low, respectively low density of "volcanoes" (hypabyssal intrusions) at the west and between the central and eastern segments of the Carpathian volcanic arc, the central and especially eastern segments show high density of closely spaced, often overlapping volcanoes, implying temporary high magma production rate. The average spacing of volcanoes in the Vihorlat-Gutin volcanic chain (sub-region 4c in the Figure 11) is 12 km and in the Călimani-Ghurghiu-Hargita volcanic chain (sub-region 7 in the Figure 11) only 10 km. Spacing of volcanoes in the most active recent island arcs and continental margins varies in the range 20 to 35 km. We explain the unusually high magma production rate by acceleration of magma generation processes due to the slab break-off (compensation of its subsidence by faster upflow of asthenospheric material stimulating its decompression partial melting).

Intermediate calc-alkaline polygenetic volcanoes in the CPR evolved in two types of geotectonic setting – the arc-type as well as the back-arc type setting. It is interesting to analyze their differences. We shall base our discussion on the comparison of the central Slovakia Volcanic Field (sub-region 2a in the Figure 11) in the back-arc setting and Vihorlat-Gutin and Călimani-Ghurghiu-Hargita volcanic chains (sub-regions 4c and 7 in the Figure 11) in the arc setting. For composite/compound andesite stratovolcanoes in the back-arc setting there is characteristic relatively large size, long lasting evolution at several stages, higher degree of explosivity (larger proportion of pyroclastic rocks), higher proportion of differentiated rocks including late stage rhyolites due to evolution of magmas in shallow crustal magma chambers, presence of calderas and volcanic grabens, extensive subvolcanic intrusive complexes and related hydrothermal processes. Activity of the largest Štiavnica stratovolcano including a caldera and volcano-plutonic complex lasted for over 4 million years and involved around 2000 km^3 of magma [262]. Composite andesite stratovolcanoes in the arc setting show generally more mafic composition and dominantly effusive nature, mostly limited extent of differentiation and corresponding limited extent or absence of extrusive domes and explosive activity and less complex internal structure. Subvolcanic intrusive complexes are of limited extent and mostly dioritic composition. Hydrothermal processes are rare and often of the high-sulfidation type, related to the emplacement of diorite porphyry subvolcanic stocks. Calderas are rare and associate only with the largest volcanoes. They are mostly small to medium size and their activity lasted for a shorter time [2].

The CPR volcanic forms are erosion modified volcanic landforms. Due to age differences and variable geotectonic setting, degree of erosion of primary volcanic forms shows extreme variability. Owing to their prominent relief, stratovolcanoes are subject to degradation by erosion, already during their long-lasting activity. Davidson and De Silva [35] distinguish 5 stages in erosional history of stratovolcanoes: (1) fresh young cones with pristine lava flows and visible summit crater; (2) young cones with small gullies on flanks and lava flows still visible and summit crater degraded; (3) cones with well established gullies, barely visible lava flows, summit crater obliterated and constructional surfaces disappearing; (4) cones deeply incised by valleys with considerable relief, no visible lava flows, large planezes, little original surface left; (5) barely recognizable cones of lower relief with radial symmetry as the only clue to volcanic origin. In the CPR the degree of erosion of the youngest volcanoes in Southern Hargita (sub-region 7c in the Figures 1 and 11) corresponds to stage 3, and older volcanoes in the arc-setting correspond to the stage 4. To the same stage is assigned also the Poľana Stratovolcano in the Central Slovakia Volcanic Field (Figure 13). Degree of erosion of other stratovolcanoes in the back-arc setting corresponds to stage 5. Many of the older (16.5-15 Ma), deeply eroded stratovolcanoes in the CPR do not show even that radial symmetry. In the extreme case, remnants of volcanic forms in grabens and the presence of subvolcanic intrusive complexes remain the only clues to the former existence of stratovolcanoes. In many cases local basins acted as depocenters for volcanic detritus and they are the only messengers for the type of volcanism during a specific time period. Degree of erosion of volcanic forms in the CPR was strongly influenced by post-volcanic uplift, respectively, subsidence. The most uplifted parts along the internal side of the accretion prism have been eroded down to the hypabyssal level. On the other side, many of the volcanoes and volcanic products situated in the Pannonian basin were buried under a thick cover of younger sedimentary rocks. Volume of buried volcanic rocks may be even equal to the exposed ones. A stronger uplift of Apuseni Mts. (sub-region 5 in Figure 11) stands behind a much deeper erosion of volcanic forms down to the subvolcanic level (they are preserved in volcano-tectonic grabens only) if compared with the Central Slovakia Volcanic Field (sub-region 2a in Figure 11).

Overall morphology of stratovolcanoes is strongly influenced by climatic conditions. In dry and cold conditions the rate of erosion is relatively low, aggradation/degradation ratio is high and the resulting stratovolcano edifice shows a relatively large cone with a surrounding ring-plane of limited extent. In contrast, in humid and temperate climatic conditions the rate of erosion is high and the resulting stratovolcano shows a relatively small cone with an extensive ring plane around it. During the life-span of stratovolcanoes in the CPR, climatic conditions have changed from warm humid during the Badenian time to temperate, dryer and seasonal during the Sarmatian time, semiarid during the Late Sarmatian, temperate humid during the Pannonian time, temperate semi arid during the Early Pliocene, warm humid during the Late Pliocene and to alternating cold semiarid and temperate humid during the Quaternary time [355, 356]. The observed difference between older Badenian stratovolcanoes with extensive aprons of epiclastic volcanic rocks and younger Sarmatian stratovolcanoes with a lesser extent of these aprons probably reflects the observed climatic changes.

9. Conclusion

This work is a result of geological mapping and a great variety of investigations over recent decades that improved understanding of very complex volcanism associated with Tertiary geotectonic evolution of the Carpathian arc and Pannonian Basin. However, many uncertainties and problems still exist which have to be solved in the future. We expect that upcoming research will focus not only on new petrological, geochemical and geophysical investigations, but also on detailed fieldwork as a key to paleovolcanic reconstruction. In this respect we consider this work as our heritage to the next generation, as a starting point for the further investigations.

Acknowledgments

This paper is to be presented in the XIX Congress of the Carpathian Balkan Geological Association in Thessaloniki on September 23–26, 2010. To attend the XIX CBGA congress, JL was invited by the Organizing Committee of the Congress, KN was supported by the Massey University Leave & Ancillary Appointments Committee granted fund (LAAC10/37). IS thanks the Institute of Geodynamics, Romanian Academy and the Organizing Committee of the Congress for support. ZP acknowledge the financial support provided by the Hungarian national Scientific Fund (OTKA No. K 68153). JL acknowledges support of the VEGA grant 2/0171/08. Constructive comments by anonymous journal reviewers and Managing Editor, Katarzyna Cyran helped to improve the quality of the manuscript.

References

[1] Pécskay Z., Lexa J., Szakács A., Balogh K., Seghedi I., Konečný V., Kovács M., Márton E., et al., Space and time distribution of Neogene-Quaternary volcanism in the Carpatho-Pannonian region, Acta Vulcanologica, 1995, 7, 15-28

[2] Pécskay Z., Lexa J., Szakács A., Seghedi I., Balogh K., Konečný V., Zelenka T., Kovacs M., et al., Geochronology of Neogene-Quaternary magmatism in the Carpathian arc and Intra-Carpathian area: a Review, Geol. Carpath., 2006, 57, 511-530

[3] Csontos L., Nagymarosy A., Horvath F., Kovac M., Tertiary evolution of the Intra-Carpathian area: a model, Tectonophysics, 1992, 208, 221-241

[4] Konečný V., Kováč M., Lexa J., Šefara J., Neogene evolution of the Carpatho-Pannonian region: an interplay of subduction and back-arc diapiric uprise in the mantle, European Gophysical Society Special Publication Series, 2002, 1, 165-194

[5] Seghedi I., Downes H., Szakacs, A., Mason, P.R.D., Thirlwall, M.F., Rosu, E., Pecskay, Z., Marton, E., Panaiotu, C., Neogene-Quaternary magmatism and geodynamics in the Carpathian-Pannonian region: a synthesis, Lithos, 2004, 72, 117-146

[6] Harangi S., Downes H., Seghedi I., Tertiary-Quaternary subduction processes and related magmatism in Europe. In: Gee D.G., Stephenson R.A. (Eds.), European Lithosphere Dynamics, Geological Society of London, London, 2006, 67-190

[7] Harangi S., Lenkey L., Genesis of the Neogene to Quaternary volcanism in the Carpathian - Pannonian region: Role of subduction, extension, and mantle plume, Geol. S. Am. S., 2007, 418, 67-90

[8] Szakács A., Seghedi I., Large volume volcanic debris avalanche in the East Carpathians, Romania. In: Leyrit H., Montenat C. (Eds.), Volcaniclastic rocks, from magmas to sediments, Gordon Breach Science Publishers, Amsterdam, The Netherlands, 2000, 131-151

[9] Maleev E.F., On the method of geological mapping of pyroclastic and effusive rocks, Razvedka Nedr, 1951 (in Russian)

[10] Maleev E.F., Neogene volcanism of Transcarpathia, Moscow, 1964, (in Russian)

[11] Konečný V., Lexa J., Planderová E., Stratigrafické členenie neovulkanitov stredného Slovenska (Stratigraphy of the Central Slovakia Volcanic Field), Zapad. Karpaty, Ser. Geol., 1983, 9, 1-203 (in Slovak)

[12] Konečný V., Lexa J., Geological map of the Central Slovakia Neogene Volcanic Field, Geologický ústav D. Štúra, Bratislava, 1984

[13] Maleev E.F., Volcaniclastic rocks, Moscow, 1963 (in Russian)

[14] Fisher R.V., Proposed classification of volcaniclastic sediments and rocks, Geol. Soc. Am. Bull., 1961, 72, 1395-1408

[15] Fisher R., Rocks composed of volcanic fragments, Earth Sci. Rev., 1966, 1, 287-298

[16] Fisher R.V., Schmincke H.U., Pyroclastic Rocks, Springer, Heidelberg, 1984

[17] Cas R., Wright J., Volcanic Successions, Modern and Ancient, Allen and Unwin, London Boston Sydney Wellington, 1987

[18] Wohletz K., Heiken G., Volcanology and geothermal energy, University of California Press, Berkeley, 1992

[19] Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J., (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000

[20] McPhie J., Doyle M., Allen R., Volcanic Textures. A guide to the interpretation of textures in volcanic rocks, Tasmanian Government Printing Office, Tasmania, 1993

[21] Schmincke H.U., Volcanism, Springer Verlag, Heidelberg, 2004

[22] Németh K., Martin U., Practical Volcanology, Occasional Papers of the Geological Institute of Hungary, Budapest, 2007, 207, 1-221

[23] Seghedi I., Downes H., Vaselli O., Szakács A., Balogh K., Pécskay Z., Postcollisional Tertiary-Quaternary mafic alkalic magmatism in the Carpathian-Pannonian region: a review, Tectonophysics, 2004, 393, 43-62

[24] Harangi S., Neogene to Quaternary volcanism of the Carpathian-Pannonian Region - a review, Acta Geologica Academiae Scientiarum Hungaricae, 2001, 44, 223-258

[25] Harangi S., Neogene magmatism in the Alpine-Pannonian Transition Zone - a model for melt generation in a complex geodynamic setting, Acta Vulcanologica, 2001, 13, 25-39

[26] Kovács I., Csontos L., Szabó C., Bali E., Falus G., Benedek K., Zajácz Z., Paleogene-early Miocene igneous rocks and geodynamics of the Alpine-Carpathian-Pannonian-Dinaric region: an integrated approach. In: Beccaluva L., Bianchini G., Wilson M., (Eds.), Cenozoic Volcanism in the Mediterranean Area, 2007, 93-112

[27] Szabó C., Falus G., Zajácz Z., Kovács I., Bali, E., Composition and evolution of lithosphere beneath the Carpathian-Pannonian Region: a review,

Tectonophysics, 2004, 393, 119-137

[28] MacDonald G.A., Volcanoes, Prentice-Hall, Englewoods Cliffs, N.J, 1972

[29] Williams H., McBirney A.R., Volcanology, Freeman, Cooper & Co., San Francisco, 1979

[30] Wright J.V., Smith A.L., Self S., A working terminology of pyroclastic deposits, J. Volcanol. Geoth. Res., 1980, 8, 315-336

[31] Schmid R., Descriptive nomenclature and classification of pyroclastic deposits aand fragments: recommendations of the IUGS Subcommission on the Systematics of Igneous rock. Geology, 1981, 9, 41-43

[32] Francis P., Volcanoes, a planetary perspective, Oxford University Press, Oxford, 1993

[33] White J.D.L., Houghton B.F., Primary volcaniclastic rocks, Geology, 2006, 34, 677-680

[34] Cotton C.A., Volcanoes as landscape forms, Whitcombe and Tombs Ilmited, Wellington, New Zealand, 1944

[35] Davidson J., de Silva S., Composite volcanoes. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 665-681

[36] Francis P., Volcanoes, a planetary perspective, Oxford University Press, Oxford, 1993

[37] Connor C.B., Conway F.M., Basaltic volcanic fields. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 331-343

[38] Valentine G.A., Gregg T.K.P., Continental basaltic volcanoes - Processes and problems. J. Volcanol. Geoth. Res., 2008, 177, 857-873

[39] Németh K., Monogenetic volcanic fields; their origin, sedimentary record, relationship with polygenetic volcanism, and how monogenetic they are. In: Canon-Tapia E., Szakacs A., (Eds.) What is a volcano? Geological Society of America, 2010 (in press)

[40] Walker G.P.L., Basaltic volcanoes and volcanic systems, In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 283-290

[41] Rose W.I., Jr., Stoiber R.E., The 1966 eruption of Izalco volcano, El Salvador, J. Geophys. Res., 1969, 74, 3119-3130

[42] Carr M.J., Pontier N.K., Evolution of a young parasitic cone towards a mature central vent: Izalco and Santa Ana volcanoes in El Salvador, Central America, J. Volcanol. Geoth. Res., 1981, 11, 277-292

[43] Auer A., Martin U., Németh K., The Fekete-hegy (Balaton Highland Hungary) "soft-substrate" and "hard-substrate" maar volcanoes in an aligned volcanic complex - Implications for vent geometry, subsurface stratigraphy and the palaeoenvironmental setting, J. Volcanol. Geoth. Res., 2007, 159, 225-245

[44] Németh K., Cronin S.J., Haller M.J., Brenna M., Csillag G., Modern analogues for Miocene/Pliocene alkali basaltic phreatomagmatic fields in the Pannonian Basin: "soft-substrate" aquifer controlled phreatomagmatism in intraplate volcanic fields, Cent. Eur. J. Geosci., 2010, in press

[45] Lipman P.W., Calderas. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 643-662

[46] Lorenz V., Some Aspects of the Eruption Mechanism of the Big Hole Maar, Central Oregon, Geol. Soc. Am. Bull., 1970, 81, 1823-1830

[47] Lorenz V., McBirney A.R., Williams H., An investigation of volcanic depressions. Part III. Maars, tuff-rings, tuff-cones and diatremes. NASA Progress Report (NGR - 38-003,012). Clearinghouse for Federal Scientific and Technical Information, Springfield, Houston, Texas, 1970

[48] Lorenz V., On the formation of Maars, Bulletin of Volcanology, 1973, 37, 183-204

[49] Lorenz V., Collapse structures in the Permian of the Saar-Nahe area, Southwest Germany, Geologische Rundschau, 1971, 60, 924-948

[50] Lorenz V., Formation of phreatomagmatic maar-diatreme volcanoes and its relevance to kimberlite diatremes, Phys. Chem. Earth, 1975, 9, 17-27

[51] Lorenz V., On the growth of maars and diatremes and its relevance to the formation of tuff rings, Bulletin of Volcanology, 1986, 48, 265-274

[52] Zimanowski B., Buttner R., Lorenz V., Hafele H.G., Fragmentation of basaltic melt in the course of explosive volcanism, J. Geophys. Res. Sol. Ea., 1997, 102(B1), 803-814

[53] Kurszlaukis S., Buttner R., Zimanowski B., Lorenz V., On the first experimental phreatomagmatic explosion of a kimberlite melt, J. Volcanol. Geoth. Res., 1998, 80(3-4), 323-326

[54] Lorenz V., Syn- and posteruptive hazards of maar-diatreme volcanoes, J. Volcanol. Geoth. Res., 2007, 159, 285-312

[55] Suhr P., Goth K., Lorenz V., Long lasting subsidence and deformation in and above maar-diatreme volcanoes - a never ending story, Zeitschrift der Deutschen Gesellschaft für Geowissenschaften, 2006, 157, 491-511

[56] Manville V., Németh K., Kano K., Source to sink: A review of three decades of progress in the understanding of volcaniclastic processes, deposits, and hazards. Sediment. Geol., 2009, 220, 136-161

[57] Pirrung M., Fischer C., Buchel G., Gaupp R., Lutz H., Neuffer F.O., Lithofacies succession of maar crater deposits in the Eifel area (Germany), Terra Nova, 2003, 15, 125-132

[58] Martin U., Németh K., Mio/Pliocene phreatomagmatic volcanism in the western Pannonian Basin, Geologica Hungarica Series Geologica, Geological Institute of Hungary, Budapest, 2004

[59] Falus G., Szabó C., Vaselli O., Mantle upwelling within the Pannonian Basin: evidence from xenolith lithology and mineral chemistry, Terra Nova, 2000, 12, 295-302

[60] Magyar I., Geary D.H., Muller P., Paleogeographic evolution of the Late Miocene Lake Pannon in Central Europe, Palaeogeogr. Palaeocl., 1999, 147, 151-167

[61] Sacchi M., Horváth F., Magyari O., Role of unconformity-bounded units in the stratigraphy of the continental cord; a case study from the late Miocene of the western Pannonian Basin, Hungary. In: Durand B., Jolivet L., Horváth F., Ranne M. (Eds), The Mediterranean basins; Tertiary extension within the Alpine Orogen., Geological Society of London, United Kingdom, London, 1999, 357-390

[62] Horváth F., Bada G., Szafián P., Tari G., Ádám A., Cloetingh S., Formation and deformation of the Pannonian basin: constraints from observational data. In: Gee D.G., Stephenson R.A. (Eds.), European Lithosphere Dynamics, Geological Society, London, Memoirs, 2006, 32, 191-206

[63] Sacchi M., Horváth F., Towards a new time scale for the Upper Miocene continental series of the Pannonian basin (Central Paratethys), Stephan Mueller Special Publication Series, 2002, 3, 79 - 94

[64] Juhász E., Phillips L., Müller P., Ricketts B., Tóth M.A., Lantos M., Kovács L., Late Neogene sedimentary facies and sequences in the Pannonian Basin, Hungary. In: Durand B., Jolivet L., Horvath F., Ranne, M. (Eds.), The Mediterranean basins; Tertiary extension within the Alpine Orogen., Geological Society of London, London, United Kingdom, 1999, 335-356

[65] Müller P., Stratigraphy of the Pannonian sediments. In: Bérczi I., Jámbor Á. (Eds.), Stratigraphy of geological units of Hungary, Budapest, 1998, 485-493 (in Hungarian)

[66] Budai T., Haas J., Triassic sequence stratigraphy of the Balaton Highland, Hungary, Acta Geologica Hungarica, Budapest, 1997, 40, 307-335

[67] Juhasz A., Toth T.M., Ramseyer K., Matter A., Connected fluid evolution in fractured crystalline basement and overlying sediments, Pannonian Basin, SE Hungary, Chem. Geol., 2002, 182, 91-120

[68] Haas J., Mioc P., Pamic J., Tomljenovic B., Arkai P., Berczi-Makk A., Koroknai B., Kovacs S., et al, Complex structural pattern of the Alpine-Dinaridic-Pannonian triple junction, Int. J. Earth Sci., 2000, 89, 377-389

[69] Fodor L., Csillag G., Németh K., Budai T., Cserny T., Martin U., Brezsnyánszky K., Dewey J., Tectonic development, morphotectonics and volcanism of the Transdanubian Range and Balaton Highland: a field guide to the workshop on "Application of GPS in plate tectonics, in research on fossil energy resources and in earthquake hazard assessment", Budapest, 24-08-01-09/2002, Annual Reports of the Geological Institute of Hungary - 2003, 2004 (in press)

[70] Németh K., Martin U., Large hydrovolcanic field in the Pannonian Basin: general characteristics of the Bakony- Balaton Highland Volcanic Field, Hungary. Acta Vulcanologica, 1999, 11, 271-282

[71] Vass D., Konečný V., Tunyi I., Dolinský P., Balogh K., Hudáčková N., Kováčová-Slamková M., Beláček B., Origin of the Pliocene vertebrate bone accumulation at Hajnáčka, southern Slovakia, Geol. Carpath., 2000, 51, 69-82

[72] Fejfar O., Heinrich W.D., Importance of two sites of fossil vertebrates, Ivanovce and Hajnáčka, for mammalian paleontology in European Pliocene and Early Pleistocene: present stage of knowledge and problems, Vestnik Ustredniho Ustavu Geologického, 1985, 60, 213-224

[73] Konečný V., Lexa J., Pliocene to Pleistocene alkali basalt diatremes and maars of Southern Slovakia: a common model for their evolution, Terra Nostra, 2000, 6, 220-232

[74] Vass D., Elecko M., Konečný V., Alginite; a raw material for environmental control, Geology Today, 1997, 13, 149-153

[75] Konečný V., Lexa J., Balogh K., Konečný P., Alkali basalt volcanism in southern Slovakia; volcanic forms and time evolution, Acta Vulcanologica, 1995, 7, 167-171

[76] Konečný V., Balogh K., Orlicky O., Lexa J., Vass D., Anonymous, Evolution of the Neogene-Quaternary alkali basalt volcanism in central and southern Slovakia (West Carpathians), Special Publications of the Geological Society of Greece, 1995, 4, 533-538

[77] Martin U., Németh K., Lorenz V., White J.D.L., Introduction: Maar-diatreme volcanism. J. Volcanol. Geoth. Res., 2007, 159, 1-3

[78] Lorenz V., Kurszlaukis S., Root zone processes in the phreatomagmatic pipe emplacement model and

consequences for the evolution of maar-diatreme volcanoes, J. Volcanol. Geoth. Res., 2007, 159, 4-32

[79] Németh K., Cronin S.J., Haller M.J., Brenna M., Csillag G., Modern analogues for Miocene/Pliocene alkali basaltic phreatomagmatic fields in the Pannonian Basin: "soft-substrate" aquifer controlled phreatomagmatism in intraplate volcanic fields, Cent. Eur. J. Geosci., 2010, (in press)

[80] Sheridan M.F., Wohletz K.H., Hydrovolcanism - Basic considerations and review, J. Volcanol. Geoth. Res., 1983, 17, 1-29

[81] Wohletz K.H., Explosive magma-water interactions: Thermodynamics, explosion mechanisms, and field studies, B. Volcanol., 1986, 48, 245 - 264

[82] White J.D.L., Pre-emergent construction of a lacustrine basaltic volcano, Pahvant Butte, Utah (USA), B. Volcanol., 1996, 58, 249-262

[83] White J.D.L., Eruption and reshaping of Pahvant Butte volcano in Pleistocene Lake Bonneville. In: White J.D.L., Riggs N.R. (Eds.), Volcaniclastic sedimentation in lacustrine settings, Balckwell Sciences, Oxford, 2001, 61-80

[84] Sohn Y.K., Hydrovolcanic processes forming basaltic tuff rings and cones on Cheju Island, Korea, Geol. Soc. Am. Bull., 1996, 108, 1199-1211

[85] Sohn Y.K., Chough S.K., The Ilchulbong tuff cone, Cheju Island, South-Korea - Depositional processes and evolution of an emergent, surtseyan- type tuff cone, Sedimentology, 1992, 39, 523-544

[86] Sohn Y.K., Chough S.K., The Udo Tuff Cone, Cheju Island, South-Korea - Transformation of pyroclastic fall into debris fall and grain flow on a steep volcanic cone slope, Sedimentology, 1993, 40, 769-786

[87] Brand B.D., White C.M., Origin and stratigraphy of phreatomagmatic deposits at the Pleistocene Sinker Butte Volcano, Western Snake River Plain, Idaho, J. Volcanol. Geoth. Res., 2007, 160, 319-339

[88] Martin U., Németh K., Eruptive and depositional history of a Pliocene tuff ring that developed in a fluviolacustrine basin: Kissomlyo Volcano (Western Hungary), J. Volcanol. Geoth. Res., 2005, 147, 342-356

[89] Konečný V., Lexa J., Balogh K., Neogene-Quaternary alkali basalt volcanism of Slovakia; review of volcanic forms and evolution, Geol. Carpath., 1999, 50, 112-114

[90] Kaličiak M., Baňacký V., Jacko S., Janočko J., Karoli S., Molnár J., Petro L., Priechodská Z., et al, Explanatory notes to the geological map of Slanské vrchy Mts. and Košická depression - northern part, 1:50 000, Geologická služba SR., Bratislava, 1991 (in Slovak)

[91] Kaličiak M., Žec B., Review of Neogene volcanism of Eastern Slovakia, Acta Volcanologica, 1995, 7, 87 - 96

[92] Šimon L., Lexa J., Konečný V., Pannonian basalt volcano Šibeničný vrch, Central Slovakia Geol. Carpath., 2002, 53, (CD-version)

[93] Lexa J., Konečný V., Geodynamic aspects of the Neogene to Quaternary volcanism. In: Rakús M. (Ed.), Geodynamic development of the Western Carpathians, Geologická služba SR, Bratislava, 1998, 219-240

[94] Vespermann D., Schmincke H.U., Scoria cones and tuff rings. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 683-694

[95] Wood C.A., Morphometric evolution of cinder cones, J. Volcanol. Geoth. Res., 1980, 7, 387-413

[96] McGetchin T.R., Settle M., Chouet B.A., Cinder cone growth modeled after Northeast Crater, Mount-Etna, Sicily, J. Geophys. Res., 1974, 79, 3257-3272

[97] Hooper D.M., Sheridan M.F., Computer-simulation models of scoria cone degradation, J. Volcanol. Geoth. Res., 1998, 83, 241-267

[98] Riedel C., Ernst G.G.J., Riley M., Controls on the growth and geometry of pyroclastic constructs, J. Volcanol. Geoth. Res., 2003, 127, 121-152

[99] Parfitt E.A., A discussion of the mechanisms of explosive basaltic eruptions, J. Volcanol. Geoth. Res., 2004, 134, 77-107

[100] Parfitt E.A., Wilson L., Neal C.A., Factors influencing the height of Hawaiian lava fountains: Implications for the use of fountain height as an indicator of magma gas content, Bull. Volcanol., 1995, 57, 440-450

[101] Wilson L., Parfitt E.A., Head J.W., Explosive Volcanic-Eruptions .8. The Role of Magma Recycling in Controlling the Behavior of Hawaiian-Style Lava Fountain, Geophys. J. Int., 1995, 121, 215-225

[102] Parfitt E.A., A study of clast size distribution, ash deposition and fragmentation in a Hawaiian-style volcanic eruption, J. Volcanol. Geoth. Res., 1998, 84, 197-208

[103] Keating G.N., Valentine G.A., Krier D.J., Perry F.V., Shallow plumbing systems for small-volume basaltic volcanoes, Bull. Volcanol., 2008, 70, 563-582

[104] Martin U., Németh K., How Strombolian is a "Strombolian" scoria cone? Some irregularities in scoria cone architecture from the Transmexican Volcanic Belt, near Volcan Ceboruco, (Mexico) and Al Haruj (Libya)., J. Volcanol. Geoth. Res., 2006, 155, 104-118

[105] Blackburn E.A., Sparks R.S.J., Mechanism and dynamics of strombolian activity. J. Geol. Soc. London,

1976, 132, 429-440

[106] Di Traglia F., Cimarelli C., De Rita D., Gimeno Torrente D., Changing eruptive styles in basaltic explosive volcanism: Examples from Croscat complex scoria cone, Garrotxa Volcanic Field (NE Iberian Peninsula), J. Volcanol. Geoth. Res., 2009, 180, 89-109

[107] Pioli L., Erlund E., Johnson E., Cashman K., Wallace R., Rosi M., Granados H.D., Explosive dynamics of violent Strombolian eruptions: The eruption of Paricutin Volcano 1943-1952 (Mexico), Earth Planet. Sc. Lett., 2008, 271, 359-368

[108] Weinstein Y., A transition from Strombolian to phreatomagmatic activity induced by a lava flow damming water in a valley, J. Volcanol. Geoth. Res., 2007, 159, 267-284

[109] Valentine G.A., Perry F.V., Krier D., Keating G.N., Kelley R.E., Coghill A.H., Small-volume basaltic volcanoes: Eruptive products and processes, and posteruptive geomorphic evolution in Crater Flat (Pleistocene), southern Nevada, Geol. Soc. Am. Bull., 2006, 118, 1313-1330

[110] Foshag W.F., Gonzalez R.J., Birth and development of Paricutin volcano, Mexico, United States Geological Survey Bulletin, 1956, 965-D, 355-489

[111] Luhr J.F., Simkin T., Paricutin. The volcano born in a Mexican cornfield, Geosciences Press, Phoenix, 1993

[112] Carracedo J.C., Rodriguez Badiola E., Soler V., The 1730-1736 eruption of Lanzarote, Canary slands: a long, high-magnitude basaltic fissure eruption, J. Volcanol. Geoth. Res., 1992, 53, 239-250

[113] Thordarson T., Self S., The Laki (Skaftar-Fires) and Grimsvotn Eruptions in 1783-1785, Bull. Volcanol., 1993, 55, 233-263

[114] Kereszturi G., Németh K., Structural and morphometrical irregularities of eroded scoria cones at Bakony–Balaton Highland Volcanic Field, Hungary. Geomorphology, 2010, (in press)

[115] Dobosi G., Fodor R.V., Goldberg S.A., Late-Cenozoic alkalic basalt magmatism in northern Hungary and Slovakia; petrology, source compositions and relationship to tectonics, Acta Vulcanologica, 1995, 7, 199-207

[116] Szabó C., Bodnar R.J., Changing magma ascent rates in the Nograd-Gomor volcanic field northern Hungary/southern Slovakia; evidence from CO (sub 2) -rich fluid inclusions in metasomatized upper mantle xenoliths, Petrologiya, 1996, 4, 240-249

[117] Seghedi I., Szakács A., The Upper Pliocene-Pleistocene effusive and explosive basaltic volcanism from the Perşani Mountains, Romanian Journal of Petrology (Bucureşti), 1994, 76, 101-107

[118] Hare A.G., Cas R.A.F., Volcanology and evolution of the Werribee Plains intraplate, basaltic lava flowfield, Newer Volcanics Province, southeast Australia, Aust. J. Earth Sci., 2005, 52, 59-78

[119] Németh K., Suwesi K.S., Peregi Z., Gulácsi Z., Ujszászi J., Plio/Pleistocene flood basalt related scoria and spatter cones, rootless lava flows, and pit craters, Al Haruj Al Abyad, Libya, Geolines, 2003, 98-103

[120] Risso C., Nemeth K., Combina A.M., Nullo F., Drosina M., The role of phreatomagmatism in a Plio-Pleistocene high-density scoria cone field: Llancanelo Volcanic Field (Mendoza), Argentina, J. Volcanol. Geoth. Res., 2008, 169, 61-86

[121] Stephenson P.J., Burch-Johnson A.T., Stanton D., Long lava flows in North Queensland: context, characteristics, emplacement. In: Chapman Conf. Long Lava Flows, Townsville, Australia, 1996, 86-87

[122] Stephenson P.J., Burch-Johnston A.T., Stanton D., Whitehead P.W., Three long lava flows in north Queensland, J. Geophys. Res. Sol. Ea., 1998, 103(B11), 27359-27370

[123] Fink H.J., Anderson S.W., Lava domes and coulees. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 307-319

[124] MacDonald G.A., Volcanoes. Prentice-Hall, Englewoods Cliffs, N.J., 1972

[125] Delmelle P., Stix J., Volcanic gases. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 803-815

[126] Williams H., The history and character of volcanic domes, Univ. Calif. Berkeley Pub. Bull., Dept. of Geoll Sci.„ 1932, 21, 51-146

[127] Leyden R., Staukuppen und verwandte Bildungen, Zeitschrift für Vulkanologie, 1936, 16, 225-247

[128] Blake S., Viscoplastic models of lava domes. In: Fink J.H. (Ed.), IAVCEI Proceedings in Volcanology, Springer Verlag, Heidelberg. 1989, 88-126

[129] Batiza R., White J.D.L., Submarine lavas and hyaloclastite. In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 361-381

[130] Szakács A., Seghedi I., Volcaniclastic sequences around andesite stratovolcanoes, East Carpathians, Romania. Workshop Guide, Romanian Journal of Petrology (Bucureşti), 1996, 77, 1-55

[131] Kaličiak M., Baňacký V., Bodnár J., Dubéciová A., Jacko S., Janocko J., Jetel J., Karoli S., et al, Explanatory notes to the geological map of Slanské mountains and Košice depression - southern part, 1:

50 000, Geol Surv Slovak Rep., Bratislava, 1996 (in Slovak)
[132] Pécskay Z., Seghedi I., Downes H., Prychodko M., Mackiv B., K/Ar dating of Neogene calc-alkaline volcanic rocks from Transcarpathian Ukraine, Geol. Carpath. 2000, 51, 83-89
[133] Kovacs M., Fülöp A., Neogene volcanism in Oaş Mts., Eastern Carpathians, Romania, Geol. Carpath., 2002, 53, 208-210
[134] Lexa J., Halouzka R., Havrila M., Hanzel V., Kubeš P., Liščák P., Hojstričová V., Explanation to the geological map of the Kremnické vrchy Mts. 1 : 50 000, Geol Surv Slovak Rep., Bratislava, 1998
[135] Vass D., Konečný V., Šefara J., Pristaš J., Škvarka L., Geologická stavba Ipeľskej kotliny a Krupinskej planiny, Geologický ústav D. Štúra, Bratislava, 1979, (in Slovak with English summary)
[136] Kaliciak N., Pécskay Z., Konečný V., Lexa J., Anonymous, Evolution of the Vihorlat Neogene volcanic range based on K/Ar dating, Special Publications of the Geological Society of Greece, 1995, 4, 523-526
[137] Lexa J., Halouzka R., Havrila M., Hanzel V., Kubeš P., Liščák P., Hojstričová V., Explanation to the geological map of the Kremnické vrchy Mts; 1 : 50 000, Geol Surv Slovak Rep., Bratislava, 1998
[138] Lexa J., Forms of rhyolite bodies in the surroundings of Žiar nad Hronom, Geologicke Práce, Správy, 1971, 56, 67-80 (in Slovak)
[139] Gyarmati P., A Tokaj-hegység intermedier vulkanizmusa, Magyar Álami Földtani Intezét, Budapest, 1977, 58, 1-195 (in Hungarian)
[140] Pichler H., Acid hyaloclastites, Bull. Volcanol., 1965, 28, 293-310
[141] Zelenka T., Geology of the perlites in the Pálháza region, Nemzetközi Perlit Konferencia kiadványa, 2008, 51-76
[142] Fülöp A., Kovacs M., Pannonian acid volcanism in Gutâi Mts. (East Carpathians, Romania): volcanological features, magmatological and tectonical significance, In: Popov P. (Ed.), Plate tectonic aspects of the alpine metallogeny in the Carpatho-Balkan region, Proceedings of the annual meeting UNESCO-IGCP Project 356, Sofia, 1996, vol. 2, 57-67
[143] Heiken G., Wohletz K., Tephra deposits associated with silicic domes and lava flows. Geol. Soc. Am. Spec. Pap., 1987, 212, 55-76
[144] Cas R.A.F., Allen R.L., Bull S.W., Clifford B.A., Wright J.V., Subaqueous, rhyolitic dome-top tuff cones: a model based on the Devonian Bunga Beds, southeastern Australia and a modern analogue, Bull. Volcanol., 1990, 52, 159-174
[145] Konečný V., Lexa J., Hojstričová V., The Central Slovakian volcanic Field: a review. , Acta Vulcanologica, 1995, 7, 63-78
[146] Bezák V., Lexa J., Genetic types of rhyolite volcaniclastic rocks in the surroundings of Žiar nad Hronom, Geologicke Práce, Správy, 1982, 79, 83-112 (in Slovak)
[147] Williams H., The history and character of volcanic domes. University California Berkeley Publication, Bulletin of Department of Geological Sciences, 1932, 21, 51-146
[148] Németh K., Pécskay,Z., Martin U., Gméling K., Molnár F., Cronin S.J., Hyaloclastites, peperites and soft-sediment deformation textures of a shallow subaqueous Miocene rhyolitic dome-cryptodome complex, Pálháza, Hungary. In: Thomson K., Petford N., (Eds.), Structure and Emplacement of High-Level Magmatic Systems, The Geological Society of London, Bath, UK., 2008, 61-83
[149] Lexa J., Koděra P., Biroň A., Kraus I., Mikušová J., Fallick A.E., Gregor M., Rhyolite volcanites of the Žiar tectonis depression as a potential source of mineral raw materials. In: Beránek M., Pástor J., (Eds.), Zborník prednášok z medzinárodnej konferencie Nerastné suroviny Slovenska a ich využívanie. Slovenská banícka spoločnosť, Bratislava, 2009, 104-106
[150] Seghedi I., Szakács A., Roşu E., Pécskay Z., Gméling K., Note on the evolution of a Miocene composite volcano in an extensional setting, Zărand Basin (Apuseni Mts., Romania), Cent. Eur. J. Geosci., 2010, in press
[151] Seghedi I., Maţenco L., Downes H., Mason P.R.D., Szakács A., Pécskay Z., Tectonic significance of changes in post-subduction Pliocene-Quaternary magmatism in the south east part of the Carpathian-Pannonian Region, Tectonophysics, 2010, http://dx.doi.org/10.1016/j.tecto.2009.12.003.
[152] Seghedi I., Ntaflos T., Pécskay Z., The Gătaia Pleistocene lamproite: a new occurrence at the southeastern edge of the Pannonian Basin, Romania, Geological Society London Special Publication, 2008, 293, 83-100, DOI:10.1144/SP293.5
[153] Downes H., Seghedi I., Szakacs A., Dobosi G., James D.E., Vaselli O., Rigby I.J., Ingram G.A., et al., Petrology And Geochemistry Of Late Tertiary Quaternary Mafic Alkaline Volcanism In Romania, Lithos, 1995, 35, 65-81
[154] Adda K., Lukarecz és vidékének geológiai viszonyai. (Az 1896. évi részi. fölvételekről.) (Geology of Lukaretz and its environs (according to detailed

mapping in 1986)), Budapest, 1897

[155] Mateescu Ş., Quaternary volcanism in the north-western part of Banat, Rev. Muzeul de Geolgie şi Mineralogie, Universitatea Babes-Bolyai, Cluj-Napoca, 1937, 6, 72-97 (in Romanian)

[156] Ilinca G., Marincea S., Russo-Sandulescu D., Iancu V., Seghedi I., Mineral occurences in the south-western Banat, Romania. Excursion Guide. Second Symposium on Mineralogy, Timisoara, Romanian Journal of Mineralogy, Bucuresti, 1993, 76, suppl. 2

[157] Pécskay Z., Edelstein O., Seghedi I., Szakács A., Kovács M., Crihan M., Bernad A., K-Ar datings of Neogene-Quaternary calc-alkaline volcanic rocks in Romania, Acta Vulcanologica, 1995, 7, 53-61

[158] Seghedi I., Szakacs A., Snelling N.J., Pecskay Z., Evolution of the Neogene Gurghiu Mountains volcanic range (Eastern Carpathians, Romania), based on K-Ar geochronology, Geol. Carpath., 2004, 55, 325-332

[159] Seghedi I., Szakacs A., Pecskay Z., Mason P.R.D., Eruptive history and age of magmatic processes in the Călimani volcanic structure (Romania), Geol. Carpath., 2005, 56, 67-75

[160] Konečný V., Lexa J., Balogh K., Neogene-Quaternary alkali basalt volcanism of Slovakia; review of volcanic forms and evolution, Geol. Carpath., 1999, 50, 112-114

[161] Konečný V., Lexa J., Paleovolcanic reconstruction of Neogene volcanoes in the central Slovakia volcanic field, Geol. Carpath., 1999, 50, 109-112

[162] Budkewitsch P., Robin P.-Y., Modelling the evolution of columnar joints. J. Volcanol. Geotherm. Res., 1994, 59, 219-239

[163] DeGraff J.M., Aydin A., Surface morphology of columnar joints and its significance to mechanics and direction of joint growth, Geol. Soc. Am. Bull., 1987, 99, 605-617

[164] Lyle P., The eruption environment of multi-tiered columnar basalt lava flows, J. Geol. Soc. London, 2000, 157, 715-722

[165] Spry A.H., The origin of columnar jointing particularly in basalt flows. Journal of the Geological Society of Australia, 1962, 8, 191-216

[166] Valentine G.A., Keating G.N., Eruptive styles and inferences about plumbing systems at Hidden Cone and Little Black Peak scoria cone volcanoes (Nevada, USA), Bull. Volcanol., 2007, 70, 105-113

[167] Walker G.P.L., Basaltic-volcano systems. In: Prichard H.M., Alabaster T., Harris N.B.W., Nearly C.R., (Eds.), Magmatic Processes and Plate Tectonics, Geological Society of London Special Publications, 1993, 76, 3-38

[168] Németh K., Martin U., Shallow sill and dyke complex in western Hungary as a possible feeding system of phreatomagmatic volcanoes in "soft-rock" environment, J. Volcanol. Geoth. Res., 2007, 159, 138-152

[169] Konečný V., Lexa J., Evolution of the phreatomagmatic/extrusive/intrusive complex of the Bulhary maar-diatreme volcano in Southern Slovakia, Geolines, 2003, 15, 47-51

[170] Hearn B.C.J., Diatremes with kimberlitic affinities in North-Central Montana, Science, 1968, 159, 622-625

[171] Lorenz V., Formation of maar-diatreme volcanoes, Terra Nostra, 2000, 6, 284-291

[172] Gernon T.M., Gilbertson M.A., Sparks R.S.J., Field M. Gas-fluidisation in an experimental tapered bed: insights into processes in diverging volcanic conduits, J. Volcanol. Geoth. Res., 2008, 174, 49–56

[173] Gernon T.M., Gilbertson M.A., Sparks R.S.J., Field M., Gas-fluidisation in an experimental tapered bed: Insights into processes in diverging volcanic conduits, J. Volcanol. Geoth. Res., 2008, 174, 49-56

[174] Walters A.L., Phillips J.C., Brown R.J., Field M., Gernon T., Stripp G., Sparks R.S.J., The role of fluidisation in the formation of volcaniclastic kimberlite: Grain size observations and experimental investigation, J. Volcanol. Geoth. Res., 2006, 155, 119-137

[175] Ross P.S., White J.D.L., Zimanowski B., Buttner R., Rapid injection of particles and gas into non-fluidized granular material, and some volcanological implications, Bull. Volcanol., 2008, 70, 1151-1168

[176] Ross P.S., White J.D.L., Zimanowski B., Buttner R., Multiphase flow above explosion sites in debris-filled volcanic vents: Insights from analogue experiments, J. Volcanol. Geoth. Res., 2008, 178, 104-112

[177] Ross P.S., White J.D.L., Debris jets in continental phreatomagmatic volcanoes: A field study of their subterranean deposits in the Coombs Hills vent complex, Antarctica, J. Volcanol. Geoth. Res., 2006, 149, 62-84

[178] Bence G., Bernhardt B., Bihari D., Bálint C., Császár G., Gyalog L., Haas J., Horváth I., et al., A Bakony hegység földtani képződményei. Magyarázó a Bakony hegység fedetlen földtani térképéhez 1:50 000. Földt. Int. alkalmi kiadv., 1990 (in Hungarian)

[179] Budai T., Csillag G. (Eds.), Geology of the Balaton Highland: explanatory booklet for the geology map of the Balaton Highland, scale 1:50 000, Occasional Papers of the Geological Institute of Hungary, Geological Institute of Hungary, Budapest, 1999

[180] Bence G., Bihari D., Lantos M., Geomagnetic measurements to detect basaltic volcanic vents in the

Balaton Highland, Annual Report of the Hungarian Geological Institute, 1988, 1988, 363-369

[181] Fodor L., Csillag G., Németh K., Budai T., Cserny T., Martin U., Brezsnyanszky K., Dewey J., Tectonic development, morphotectonics and volcanism of the Transdanubian Range: a field guide, Occasional Papers of the Geological Institute of Hungary, 2005, 204, 59-86

[182] Budai T., Csillag G., Dudko A., Koloszár L., Geological map of the Balaton Highland, Magyar Állami Földtani Intézet, Budapest, 1999

[183] Németh K., Martin U., Haller M.J., Alric V.L., Cenozoic diatreme field in Chubut (Argentina) as evidence of phreatomagmatic volcanism accompanied with extensive Patagonian plateau basalt volcanism? Episodes, 2007, 30, 217-223

[184] Lorenz V., Haneke J., Relationship between diatremes, dykes, sills, laccoliths, intrusive-extrusive domes, lava flows, and tephra deposits with unconsolidated water-saturated sediments in the late Variscan intermontane Saar-Nahe Basin, SW Germany. In: Breitkreuz C., Petford N. (Eds.), Physical Geology of Subvolcanic Systems - Laccoliths, Sills, and Dykes, Blackwell Sciences, Oxford, 2004, 75-124

[185] Cassidy J., France S.J., Locke C.A., Gravity and magnetic investigation of maar volcanoes, Auckland volcanic field, New Zealand, J. Volcanol. Geoth. Res., 2007, 159, 153-163

[186] Ort M.H., Dallege T.A., Vazquez J.A., White J.D.L., Volcanism and sedimentation in the Mio-Pliocene Bidahochi Formation, Navajo Nation, NE AZ. In: Duebendorfer E. (Ed.), Geologic Excursions in northern and Central Arizona. Field trip Guidebook for Geological Society of America Rocky Mountain Section Meeting, Arizona, Geological Society of America, Flagstaff, Arizona, 1998, 35-57

[187] Skilling I.P., White J.D.L., McPhie J., Peperite: a review of magma-sediment mingling, J. Volcanol. Geoth. Res., 2002, 114, 1-17

[188] Godchaux M., Bonnichsen B., Syneruptive magma-water and posteruptive lava-water interactions in the Western Snake River Plain, Idaho, during the past 12 million years. In: Bonnichsen B., White C.M., McCurry M. (Eds.), Tectonic and magmatic evolution of the Snake River Plain Volcanic Province, Idaho Geological Survey, University of Idaho, Moscow, Idaho. 2002, 387-435

[189] Németh K., White C.M., Intra-vent peperites related to the phreatomagmatic 71 Gulch Volcano, western Snake River Plain volcanic field, Idaho (USA), J. Volcanol. Geoth. Res., 2009, 183, 30-41

[190] Martin U., Németh K., Blocky versus fluidal peperite textures developed in volcanic conduits, vents and crater lakes of phreatomagmatic volcanoes in Mio/Pliocene volcanic fields of Western Hungary, J. Volcanol. Geoth. Res., 2007, 159, 164-178

[191] Konečný V., Lexa J., Konečný P., Volcanic and subvolcanic lava bodies-Structural aspects, Geolines, 2004, 17, 121-127

[192] Konečný V., Lexa J., Konečný P., A guidebook to the southern Slovakia alkali basalt volcanic field. Field Guide - Second International Maar Conference (Hungary - Slovakia - Germany - 15-29 September, 2004), Geological Survey of the Slovak Republic Bratislava, Slovak Republic, 2004

[193] Németh K., Martin U., Shallow sub-surface sill and dyke complexes in Western Hungary as a possible feeding system of phreatomagmatic volcanoes in "soft rock" environment, J. Volcanol. Geoth. Res., 159, 2006, 138-152

[194] Németh K., Martin U., Csillag G., Calculation of erosion rates based on remnants of monogenetic alkaline basaltic volcanoes in the Bakony–Balaton Highland Volcanic Field (Western Hungary) of Mio/Pliocene age, Geolines, 2003, 15, 93-97

[195] Németh K., Martin U., Csillag G., Eroded maar/diatrema structures from the Bakony-Balaton Highland Volcanic Field, Annual Report of the Geological Institute of Hungary, 2003, 83-99 (in Hungarian)

[196] Kovacs I., Zajacz Z., Szabo C., Type-II xenoliths and related metasomatism from the Nograd-Gomor Volcanic Field, Carpathian-Pannonian region (northern Hungary-southern Slovakia), Tectonophysics, 2004, 393, 139-161

[197] Zajácz Z., Szabó C., Origin of sulfide inclusions in cumulate xenoliths from Nograd-Gomor volcanic field, Pannonian Basin (North Hungary/South Slovakia), Chem. Geol., 2003, 194, 105-117

[198] Bali E., Szabó C., Vaselli O., Török K., Significance of silicate melt pockets in upper mantle xenoliths from the Bakony-Balaton Highland Volcanic Field, Western Hungary, Lithos, 2002, 61, 79-102

[199] Berbeleac I., Iliescu D., Andrei J., Ciuculescu, O., Ciuculescu, R., Relationships between alterations, porphyry copper-gold and base metal-gold hydrothermal vein mineralizations in Tertiary intrusions, Talagiu area, Zărand Mountains, Romanian Journal of Mineral Deposits, 1995, 76, 31-39

[200] Konečný V., Lexa J., Hojsričová, V., The Central Slovakian volcanic Field: a review, Acta Vulcanologica, 1995, 7, 63-78

[201] Lexa J., Stohl J., Konečný V., The Banska Stiavnica

ore district: relationship between metallogenetic processes and the geological evolution of a stratovolcano, Mineralium Deposita, 1999, 34, 639-654

[202] Szakács A., Seghedi I., The Călimani-Gurghiu-Harghita volcanic chain, East Carpathians, Romania: Volcanological features, Acta Vulcanologica, 1995, 7, 145-153

[203] Ianovici V., Giuşca D., Ghiţulescu T.P., Borcoş M., Lupu M., Bleahu M., Savu H., Geological evolution of the Metaliferi Mountains, Editura Academia Republica Socialistă România, Bucharest, 1969 (in Romanian)

[204] Boştinescu S., Porphyry copper systems in the south Apuseni Mountains, Romania, Annuaire de l'Institut de Géologie et Géophysique, 1984, 64, 163-174

[205] Borcoş M., Lang B., Boştinescu S., Gheorghiţa J., Neogene hydrothermal ore deposits in the volcanic Gutai Mountains, part III, Révue Roumain de Géologie et Géographie, série Géologie, 1975, 19, 21-35

[206] Kovacs M., Edelstein O., Gabor M., Bonhomme M., Z., P., Neogene magmatism and metallogeny in Oaş-Gutâi-Ţibleş Mts.; a new approach based on radiometric datings, Romanian Journal of Mineral Deposits, 1997, 78, 35-45

[207] Pécskay Z., Seghedi I., Kovacs M., Szakács A., Fülöp A., Geochronology of the Neogene calc-alkaline intrusive magmatism in the "Subvolcanic Zone" of the Eastern Carpathians (Romania), Geol. Carpath., 2009, 60, 181-190

[208] Udubaşa G., Edelstein O., Pop N., Răduţ M., Iştvan D., Kovacs M., Pop V., Stan D., et al., The Tibles neogene igneous complex of North Romania: some petrologic and metallogenetic aspects, Annales Institutul Geologie şi Geofizica, Bucharest, 1983, 59, 285-295

[209] Bendö Z., Korpás L., How much time is needed for laccolith formation? A new approach based on a case study from Csódi-hegy, Dunabogdány, Hungary, Acta Geologica Hungarica, 2005, 48, 299-319

[210] Birkenmajer K., Pécskay Z., K-Ar dating of the Miocene andesite intrusions, Pieniny Mts, West Carpathians, Bulletin of Polish Academy of Sciences, Earth Science, 1999, 47, 155-169

[211] Gröger H.R., Fügenschuh T.B.M., Schmid S.M., Foeken J.P.T., Tertiary cooling and exhumation history in the Maramureş area (internal eastern Carpathians, northern Romania): thermochronology and structural data. In: Siegesmund S., Fügenschuh B., Froitzheim N. (Eds.), Tectonic Aspects of the Alpine-Dinaride- Carpathian System, Geological Society London, London, 2008, 169-195

[212] Kouzmanov K., von Quadt A., Peycheva I., Heinrich C.A., Pettke T., Roşu E., Geochemical and time constraints on porphyry ore formation in Barza magmatic complex, Apuseni Mountains Romania. In: Cook N.J., Ozgenc I., Oyman T. (Eds.), Au-Ag-Te-Se deposits. Proceedings of the 2006 Field Workshop, IGCP Project 486 (24-29 September, 2006), Izmir, Turkey. 2006, 96-102

[213] Roşu E., Seghedi I., Downes H., Alderton D.H.M., Szakács A., Pécskay Z., Panaiotu C., Panaiotu C.E. et al, Extension-related Miocene calc-alkaline magmatism in the Apuseni Mountains, Romania: origin of magmas, Swiss Bulletin of Mineralogy and Petrology, 2004, 84, 153-172

[214] Marsh B.D., Magma chambers., In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 191-206

[215] O'Connor G.V., Minut A., Leary S.F., The geology and exploration of the Roşia Montana gold deposit, Roşia montana, Transylvania, Romania. In: Ediopoulos et al. (Eds.), Mineral Exploration and sustainable development, Millpress: Rotterdam, 2003, 1213-1216

[216] Žec B., Kaličiak M., Konečný V., Lexa J., Jacko S.M., Baňacký V., Karoli S., Potfaj M., et al., Vysvetlivky ku geologickej mape Vihorlatských a Humenských vrchov 1 : 50 000, Geol Surv Slovak Rep., Bratislava, 1997, (in Slovak with English summary)

[217] Holmes A., Pinciples of Physical Geology, Van Nostrand Reinhold (UK) Co.Ltd., Wokingham, Berks, 1978

[218] Berbeleac I., Petrographic and metallogenetic study of Vălişoara (Porcurea) Metaliferi Mountains, Annales Institutul Geologie şi Geofizica, Bucharest, 1975, 46, 1-191 (in Romanian)

[219] Konečný V., Lexa J., Structure of the neogene stratovolcanoes in central Slovakia. Proceedings, XVth CBGA congress, Athens, Geolological Socirty of Greece Special Publications, 1995, 4, 539-544

[220] Dixon J.M., Simpson D.G., Centrifuge Modeling Of Laccolith Intrusion, J. Struct. Geol., 1987, 9, 87-103

[221] Mock A., Ehling B.C., Breitkreuz C., Anatomy of a laccolith complex - geometry and texture of porphyritic rhyolites in the Permocarboniferous Halle Volcanic Complex (Germany), Neues Jahrbuch Fur Geologie Und Palaontologie-Abhandlungen, 2005, 237, 211-271

[222] Dini A., Innocenti F., Rocchi S., Tonarini S., Westerman D.S., The magmatic evolution of the late Miocene laccolith-pluton- dyke granitic complex of Elba Island, Italy. Geol. Mag., 2002, 139, 257-279

[223] Friedman J.D., Huffman A.C.J. (Eds.), Laccolith com-

plexes of southeastern Utah; time of emplacement and tectonic setting; workshop proceedings, US Geological Survey Bulletin, 2158, 1998, 1-229

[224] Konečný V., Lexa J., Konečný P., Volcanic and subvolcanic lava bodies - Structural aspects, Geolines, 2004, 17, 121-127

[225] Kaličiak M., Baňacký V., Jacko S., Janocko J., Karoli S., Molnár J., Petro L., Priechodská, Z., Syčev A. et al., Explanatory notes to the geological map of Slanské vrchy Mts. and Košická depression - northern part, 1:50 000, Geologická služba, Bratislava, 1991, (in Slovak)

[226] Kaličiak M., Žec B., Review of Neogene volcanism of Eastern Slovakia, Acta Volcanologica, 1995, 7, 87-96

[227] Borcoş M., Volcanicity/metallogeny in the South Apuseni Mts. (Metaliferi Mts). In: Borcoş M., Vlad S. (Eds.), Field trip guide-Plate tectonics and metallogeny in the East Carpathians and Apuseni Mts., Geological Institute of Romania, Bucharest, 1994, 32-43

[228] Konečný V., Lexa J., Konečný P., Volcanic and subvolcanic lava bodies - Structural aspects, Geolines, 2004, 17, 121-127

[229] Zelenka T., Balázs E., Balogh K., Kiss J., Kozak M., Nemesi L., Pécskay Z., Püspöky Z. et al., Buried Neogene volcanic structures in Hungary, Foldtani Kozlony, 2004, 47, 177-219

[230] Kereszturi G., Csillag G., Németh K., Sebe K., Balogh K., Jáger V., Volcanic architecture, eruption mechanism and landform evolution of a Pliocene intracontinental basaltic polycyclic monogenetic volcano from the Bakony - Balaton Highland Volcanic Field, Hungary, Centr. Eur. J. Geosci., 2010, in press

[231] Németh K., Martin U., Harangi S., Miocene phreatomagmatic volcanism at Tihany (Pannonian Basin, Hungary), J. Volcanol. Geoth. Res., 2001, 111, 111-135

[232] Konečný V., Lexa J., Pliocene to Pleistocene alkali basalt diatremes and maars of Southern Slovakia: a common model for their evolution, Terra Nostra, 2000, 6, 220-232

[233] Konečný V., Lexa J., Konečný P., A guidebook to the southern Slovakia alkali basalt volcanic field. Field Guide - Second International Maar Conference (Hungary - Slovakia - Germany - 15-29 September, 2004) Bratislava, Slovak Republic, Geological Survey of the Slovak Republic, Bratislava, 2004

[234] Mathisen M.E., McPherson J.G., Volcaniclastic deposits: implications for hydrocarbon exploration. In: Fisher R.V., Smith G.A. (Eds.), Sedimentation in Volcanic Settings, Tulsa, Oklahoma, USA, 1991, 27-36

[235] Konečný V., Lexa J., Structure of the neogene stratovolcanoes in central Slovakia. Proceedings, XVth CBGA congress, Athens, Geological Society of Greece Special Publication, 1995, 4, 539-544

[236] Karátson D., Németh K., Székely B., Ruszkiczay-Rüdiger Z., Pécskay Z., Incision of a river curvature due to exhumed Miocene volcanic landforms: Danube Bend, Hungary. Int. J. Earth Sci., 2006, 95, 929-944

[237] Székely B., Karátson D., DEM-based morphometry as a tool for reconstructing primary volcanic landforms: examples from the Borzsony Mountains, Hungary, Geomorphology, 2004, 63, 25-37

[238] Karátson D., Németh K., Lithofacies associations of an emerging volcaniclastic apron in a Miocene volcanic complex: an example from the Borzsony Mountains, Hungary, Int. J. Earth Sci., 2001, 90, 776-794

[239] Karátson D., Oláh I., Pécskay Z., Márton E., Harangi S., Dulai A., Zelenka T., Kósik S., Miocene volcanism in the Visegrad Mountains (Hungary): an integrated approach to regional volcanic stratigraphy, Geol. Carpath., 2007, 58, 541-563

[240] Hedenquist J.W., Mineralization associated with volcanic-related hydrothermal systems in the circum-Pacific basin. In: Horn M.K. (Ed.), Trans. of the 4th Circum-Pacific Energy and Mineral Resources Conf., Singapore, 1987, 513-524

[241] Dublan L., Bezák V., Biely A., Bujnovský A., Halouzka R., Hraško L., Köhlerová M., Marcin D., et al, Vysvetlivky ku geologickej mape Pol̆any 1 : 50 000, Geol Surv Slovak Rep., Bratislava, Bratislava, 1997 (in Slovak with English summary)

[242] Konečný V., Bezák V., Halouzka R., P., Miháliková A., Marcin D., Iglárová L̆., Panáček A., et al., Vysvetlivky ku geologickej mape Javoria, Geol Surv Slovak Rep., Bratislava, Bratislava, 1998

[243] Francis P., Self S., Collapsing Volcanos, Sci. Am., 1987, 256, 91-97

[244] Francis P.W., Wells G.L., Landsat Thematic Mapper observations of debris avalanche deposits in the Central Andes, Bull. Volcanol., 1988, 50, 258-278

[245] McGuire W.J., Volcano instability: a review of contemporary themes. In: McGuire W.J., Jones A. (Eds.), Volcano Instability on the Earth and Other Planets, Geological Society of London, London, 1996

[246] Reid M.E., Sisson T.W., Brien D.L., Volcano collapse promoted by hydrothermal alteration and edifice shape, Mount Rainier, Washington, Geology, 2001, 29, 779-782

[247] Siebert L., Large volcanic debris avalanches: Characteristics of source areas, deposits, and associated eruptions, J. Volcanol. Geoth. Res., 1984, 22, 163-197

[248] Blong R.J., Pleistocene Volcanic Debris Avalanche From Mount Hagen, Papua-New-Guinea, Australian Journal of Earth Sciences, 1986, 33, 287-294

[249] Glicken H., Sedimentary architecture of large volcanic-debris avalanches. In: Fisher R.V., Smith G.A. (Eds.) Sedimentation in Volcanic Settings, Society for Sedimentary geology, Tulsa, Oklahoma, 1991, 99-106

[250] Belousov A., Belousova M., Voight B., Multiple edifice failures, debris avalanches and associated eruptions in the Holocene history of Shiveluch volcano, Kamchatka, Russia, Bull. Volcanol., 1999, 61, 324-342

[251] Ui T., Takarada S., Yoshimoto M., Debris avalanches., In: Sigurdsson H., Houghton B.F., McNutt S.R., Rymer H., Stix J. (Eds.), Encyclopedia of Volcanoes, Academic Press, San Diego, 2000, 617-626

[252] Bachelery P., Labazuy P., Lenat J.F., Submarine and subaerial debris avalanches deposits in Reunion island. C R Acad. Sci. II A, 1996, 323, 475-482

[253] Cantagrel J.M., Arnaud N.O., Ancochea E., Fuster J.M., Huertas M.J., Repeated debris avalanches on Tenerife and genesis of Las Canadas caldera wall (Canary Islands), Geology, 1999, 27, 739-742

[254] Hurlimann M., Ledesma A., Marti J., Conditions favouring catastrophic landslides on Tenerife (Canary Islands), Terra Nova, 1999, 11, 106-111

[255] Seghedi I. A. F., Late Miocene debris-avalanche deposit at the Gutâi shield volcano, NW Romania. Reevaluation of geological mapping and mineral deposits. AGU Joint Assembly, Toronto, Canada May 25-28, 2009, V32A-06

[256] Williams H., Calderas and their origin, University of Calififornia Publication, Department of Geological Sciiences, 1941, 25, 239-436

[257] Smith R.L., Bailey R.A., Resurgent cauldrons. Geol. Soc. Am. Mem., 1968, 116, 83-104

[258] Acocella V., Cifelli F., Funiciello R., Analogue models of collapse calderas and resurgent domes, J. Volcanol. Geoth. Res., 2000, 104, 81-96

[259] Acocella V., Cifelli F., Funiciello R., The control of overburden thickness on resurgent domes: insights from analogue models, J. Volcanol. Geoth. Res., 2001, 111, 137-153

[260] Zelenka T., Pécskay Z., Kiss J., Miocene vulcanism of the Mátra Mountains (N-Hungary). Pancardi, 2, Abstracts, 2001, 16

[261] Dublan L., Bezák V., Biely A., Bujnovský A., Halouzka R., Hraško L., Köhlerová M., Marcin D., Onačila D., Scherer S., Vozárová A., Vozár J., Žáková E., Vysvetlivky ku geologickej mape Poľany 1 : 50 000 (Explanatory notes to geological map of Poľana 1 : 50 000), Geol Surv Slovak Rep., Bratislava, 1997, 1-238 (in Slovak with English summary)

[262] Konečný V., Lexa J., Halouzka R., Hók J., Vozár J., Dublan L., Nagy A., Šimon L., Het al., Vysvetlivky ku geologickej mape Štiavnických vrchov a Pohronského Inovca (Štiavnický stratovulkán, Geol Surv Slovak Rep., Bratislava, 1998, (in Slovak with English summary)

[263] Konečný V., Evolutionary stages of Banská Štiavnica Caldera and its post-volcanic structures, Bull. Volcanol., 1971, 35, 95-116

[264] Juvigne E., Gewelt M., Gilot E., Hurtgen C., Seghedi I., Szakacs A., Gabris G., Hadnagy A., et al., An Eruption at About 10,7000 Yr Bp (C-14) in the Eastern Carpathian Mountains (Romania), C R Acad. Sci. II A, 1994, 318, 1233-1238

[265] Moriya I., Okuno M., Nakamura T., Ono K., Szakács A. I. S., Radicarbon ages of charcoal fragments from the pumice flow deposit of the last eruption of Ciomadul volcano, Romania, Summaries of Researches Using AMS at Nagoya University, 1996, 7, 252-255 (in Japanese)

[266] Konečný V., Lexa J., Processes and products of shallow submarine andesite volcanic activity. Abstracts, International Volcanological Congress, Ankara, 1994

[267] Póka T., Seghedi I., Márton E., Zelenka T., Pécskay Z., Miocene volcanism of the Cserhát Mts (N Hungary): Integrated volcano-tectonic, geochronologic and petrochemical study, Acta Geologica Hungarica, 2004, 47, 221-246

[268] Fülöp A., The onset of the volcanism in Gutâi Mountains. Paleovolcanological and paleosedimentological reconstructions, 2003 (in Romanian)

[269] Szakács A., Zelenka T., Marton E., Pécskay Z., Póka T., Seghedi I., Miocene acidic explosive volcanism in the Bükk Foreland, Hungary; identifying eruptive sequences and searching for source locations, Acta Geologica Hungarica, 1998, 41, 413-435

[270] Szakács A., Petrological and tephrological study of the Lower Badenian volcanic tuffs from the northwestern Transylvanian Basin. PhD Thesis, University of Bucharest, 2000 (in Romanian)

[271] Lexa J., Post-Caldera epithermal veins of the Stiavnica Stratovolcano, Guidebook Series - Society of Economic Geologists, 1999, 31, 249-257

[272] Cloos H., Hebung - spaltung - vulkanismus, Geologische Rundshau, 1939, 30, 405-527

[273] Richie J. E., Scotland, the Tertiary volcanic districts. 3rd edition - revised volume by MacGregor A.G. and Anderson F.W., Geol. Survey and Museum Edinburgh, Edingburgh, 1961

[274] Anderson E.M., The dynamics of the formation of cone sheets, ring dykes and cauldron subsidenced. Royal Society of Edinburg Proceedings, 1936, 56, 128-157

[275] Seghedi I., Balintoni I., Szakacs A., Interplay of tectonics and neogene post-collisional magmatism in the Intracarpathian region, Lithos, 1998, 45, 483-497

[276] Seghedi I., Bojar A.V., Downes H., Roşu E., Tonarini S., Mason P.R.D., Generation of normal and adakite-like calc-alkaline magmas in a non-subductional environment: A Sr-O-H isotopic study of the Apuseni Mountains Neogene magmatic province, Romania, Chem. Geol., 2007, 245, 70-88

[277] Nemčok M., Lexa J., Evolution of the basin and range structure around the Žiar mountain range, Geol. Carpat., 1990, 41, 229-250

[278] Roşu E., Pécskay Z., Stefan A., Popescu G., Panaiotu C., Panaiotu C.E., The evolution of the Neogene volcanism in the Apuseni Mountains (Romania): constraints from new K/Ar data, Geol. Carpath., 1997, 48, 353-359

[279] Karátson D., Bernád E., Geomorphological and geological evidences for one of the oldest crater remnant known in the Carpathians: Mt. Rotunda, Gutii Mts. Revue Roumaine de Géologie, Géophysique et Géographie – ser. de Géographie, Bucharest, 1994, 38, 115-121

[280] Wijbrans J., Nemeth K., Martin U., Balogh K., Ar-40/Ar-39 geochronology of Neogene phreatomagmatic volcanism in the western Pannonian Basin, Hungary, J. Volcanol. Geoth. Res., 2007, 164, 193-204

[281] Balogh K., Németh K., Itaya T., Molnár F., Stewart R.B., Thanh N.X., Hyodo H., Daróczi L., Loss of 40Ar(rad) from leucite-bearing basanite at low temperature: implications on K/Ar dating, Cent. Eur. J. Geosci., 2010, in press

[282] Balogh K., Németh K., Evidence for the neogene small-volume intracontinental volcanism in western Hungary: K/Ar geochronology of the Tihany Maar volcanic complex, Geol. Carpath., 2005, 56, 91-99

[283] Balogh K., Lobitzer H., Pécskay Z., Ravasz C., Solti G., Kelet-stájerországi és burgenlandi tercier vulkanitok K/Ar kora, Annual Report of the Hungarian Geological Institute, 1988, 1988, 451-468

[284] Balogh K., Pécskay Z., K/Ar and Ar/Ar geochronological studies in the Pannonian-Carpathians-Dinarides (PANCARDI) region, Acta Geologica Academiae Scientiarum Hungaricae, 2001, 44, 281-301

[285] Németh K., Martin U., Magyar I., Field guide to Pliocene phreatomagmatic volcanoes along the Raba Fault Zone, Western Hungary. International Union of Geological Sciences – Subcommission on Neogene Stratigraphy: Regional Committee on Mediterranean Neogene Stratigraphy - "Patterns and Processes in the Neogene of the Mediterranean Region", 2005

[286] Kereszturi G., Németh K., Csillag G., Kovács J., Balogh K., The role of external environmental factors in changing eruption styles of monogenetic volcanoes in a Pliocene continental volcanic field in western Hungary, J. Volcanol. Geoth. Res., 2010 in press

[287] Németh K., Martin U., Late Miocene paleo-geomorphology of the Bakony-Balaton Highland Volcanic Field (Hungary) using physical volcanology data, Zeitschrift für Geomorphologie, 1999, 43, 417-438

[288] White J.D.L., Maar-diatreme phreatomagmatism at Hopi Buttes, Navajo Nation (Arizona), USA, Bull. Volcanol., 1991, 53, 239-258

[289] Godchaux M.M., Bonnichsen B., Jenks M.D., Types of phreatomagmatic volcanoes in the Western Snake River Plain, Idaho, USA, J. Volcanol. Geoth. Res., 1992, 52, 1-25

[290] Valentine G.A., Krier D.J., Perry F.V., Heiken G., Eruptive and geomorphic processes at the Lathrop Wells scoria cone volcano, J. Volcanol. Geoth. Res., 2007, 161, 57-80

[291] Awdankiewicz M., Reconstructing an eroded scoria cone: the Miocene Sosnica Hill volcano (Lower Silesia, SW Poland), Geol. Q., 2005, 49, 439-448

[292] Griffin W.L., Wass S.Y., Hollis J.D., Ultramafic xenoliths from Bullenmerri and Gnotuk maars, Victoria, Australia: Petrology of a sub-continental crust-mantle transition, J. Petrol., 1984, 25, 53-87

[293] Jones R.N., McMahon T., Bowler J.M., Modelling historical lake levels and recent climate change at three closed lakes, Western Victoria, Australia (c.1840-1990), J. Hydrol., 2001, 246, 159-180

[294] Buchel G., Lorenz V., Schmincke H.U., Zimanowski B., Quaternary Volcanic Fields Of The Eifel, Fortschritte Der Mineralogie, 1986, 64, 97-141

[295] Houghton B.F., Schmincke H.U., Mixed deposits of simultaneous strombolian and phreatomagmatic volcanism; Rothenberg Volcano, East Eifel volcanic field, J. Volcanol. Geoth. Res., 1986, 30, 117-130

[296] Houghton B.F., Wilson C.J.N., Rosenberg M.D., Smith I.E.M., Parker R.J., Mixed deposits of complex magmatic and phreatomagmatic volcanism: An example from Crater Hill, Auckland, New Zealand, Bull. Volcanol., 1996, 58, 59-66

[297] Aranda-Gomez J.J., Luhr J.F., Pier J.G., The La-

Brena-El-Jaguey-Maar Complex, Durango, Mexico, Geological Evolution, Bull. Volcanol., 1992, 54, 393-404

[298] Heiken G.H., Tuff rings: examples from the Fort Rock-Christmas Lake Valley Basin, South-Central Oregon, J. Geophys. Res., 1971, 76, 5615-5626

[299] Vazquez J.A., Ort M.H., Facies variation of eruption units produced by the passage of single pyroclastic surge currents, Hopi Buttes volcanic field, USA, J. Volcanol. Geoth. Res., 2006, 154, 222-236

[300] White J.D.L., Depositional architecture of a maar-pitted playa - sedimentation in the Hopi Buttes Volcanic Field, Northeastern Arizona, USA, Sediment. Geol., 1990, 67, 55-84

[301] White J.D.L., Basic elements of maar-crater deposits in the Hopi Buttes volcanic field, Northeastern Arizona, USA, J. Geol., 1989, 97, 117-125

[302] Šimon L., Maglay J., Datovanie sedimentov podložia lávového prúdu vulkánu Putikov vrch metódou opticky stimulovanej luminiscencie (Dating of ediments underlying a lava flow of the Putikov vrch volcano by the method of opticaly stimulated luminiscence), Mineralia Slovaca, 2005, 37, 279-281

[303] Korpás L., Lang B., Timing of volcanism and metallogenesis in the Börzsöny Mountains, northern Hungary, Ore Geol. Rev., 1993, 8, 477-501

[304] Zelenka T., Pécskay Z., Kiss J., Miocene vulcanism of the Mátra Mountains (N-Hungary), Pancardi, 2, Abstracts, 2001

[305] Pantó G., Ignimbrites of Hungary with regard to their genetics and classification, Bull. Volcanol., 1963, 25, 175-181

[306] Pantó G., The role of ignimbrites in the volcanism of Hungary, Acta Geologica Academiae Scientiarum Hungaricae, 1962, 6, 307

[307] Vass D., Ogujanova-Rumenova N., Konečný V., Sedimentology and paleoecology of Southern Slovakian basalt maar lakes. In: Németh K., Martin U., Goth K., Lexa J. (Eds.), Abstract Volume of the Second International Maar Conference, Occasional Papers of the Geological Institute of Hungary, Budapest, 2004

[308] Konečný V., Lexa J., Balogh K., Konečný P., Alkali basalt volcanism in southern Slovakia; volcanic forms and time evolution, Acta Vulcanologica, 1995, 7, 167-171

[309] Prichystal A., Badenian potassium trachyandesites at the contact of the Bohemian Massif and West Carpathians. In: Magmatism and rift basin evolution. Czech Geological Institute, Praha, 1998, 85-85

[310] Zelenka T., Gyarmati P., Kiss J., Vértessy L., Horváth I., Márton E., Pécskay Z., Paleovolcanic reconstrucion in the Tokaj Mountains. Geologica Hungarica ser. Geologica, Magyar Állami Fëldtani Intézet, Budapest, 2010, (in press)

[311] Kovacs M., Fülöp A., Neogene volcanism in Gutâi Mts. (Eastern Carpathians). A review. Studia Universitatis Babeş-Bolyai, seria Geologia, 2003, 48, 3-16

[312] Săndulescu M., Geotectonics of Romania. (In Romanian). Bucureşti Edit, Tehnică, 1984

[313] Fülöp A., Geometry of 15.4 Ma rhyolitic ignimbrites from Gutai Mts., Eastern Carpathians, Romania, Romanian Journal of Mineral Deposits, 2002, 80, 38-40

[314] Fülöp A., Crihan M., Badenian and Sarmaţian acidic volcaniclastics of pyroclastic origin from Oaş Mts., Eastern Carpathians (Northwestern Romania), Révue Roumaine de Geologie, 2002, 46, 61-72

[315] Edelstein O., Bernad A., Kovacs M., Crihan M., Pécskay Z., Preliminary data regarding the K-Ar ages of some eruptive rocks from the Baia Mare neogene volcanic zone, Révue Roumaine de Geologie, 1992, 36, 45-60

[316] Kovacs M., Petrogenesis of the subduction-related igneous rocks from central- southeastern area of the Gutâi Mts, Cluj-Napoca Edit, Dacia, 2001 (In Romanian)

[317] Kovacs M., Fülöp A., Pécskay Z., Neogene calc-alkaline magmatism in Oaş-Gutai Mts., Eastern Carpathians, Romania. Time and space evolution. Proceedings XVIIIth Congress of the Carpatho-Balkan Geological Association, Belgrade, 2006, 290-294

[318] Fülöp A., Kovacs M., Features of domes from Oaş - Gutâi Mts., Romanian East Carpathians and their implications. European Science Foundation EUROPROBE TESZ, PANCARDI and GEORIFT Projects Joint Meeting, Romanian Journal Tectonics and Regional Geology, 1999, 77, 37

[319] Royden L.H., Horvath F. (Eds.), The Pannonian Basin; a study in basin evolution, AAPG Memoir, 45, 1988

[320] Săndulescu M., Cenozoic tectonic history of the Carpathians. In: Royden L.H., Horváth F. (Eds.), The Pannonian Basin: A study in basin evolution, AAPG Memoir, 45, 1988, 17-25

[321] Tari G., Dövényi P., Dunkl I., Horváth F., Lenkey L., Sefanescu M., Szafián P., Tóth T., Lithospheric structure of the Pannonian Basin derived from seismic, gravity and geothermal data. In: Durand B., Jolivet L., Horváth F., Ranne M., (Eds.), The Mediterranean basins; Tertiary extension within the Alpine Orogen., Geological Society of London: London, United Kingdom, 1999, 215-250

[322] Fodor, L., Csontos, L., Bada, G., Gyorfi, I., Benkovics,

L., Tertiary tectonic evolution of the Pannonian Basin system and neighbouring ogens; a new synthesis of palaeostress data. In: Durand B., Jolivet L., Horváth F., Ranne M., (Eds.), The Mediterranean basins; Tertiary extension within the Alpine Orogen., Geological Society of London: London, United Kingdom, 1999, 295-334

[323] Csontos L., Benkovics L., Bergerat F., Mansy J.L., Worum G., Tertiary deformation history from seismic section study and fault analysis in a former European Tethyan margin (the Mecsek-Villany area, SW Hungary), Tectonophysics, 2002, 357, 81-102

[324] Panaiotu C., Paleomagnetic constraints on the geodynamic history of Romania, Reports on Geodesy, 1998, 7, 205- 216

[325] Pătraşcu S., Panaiotu C., Şeclaman M., Panaiotu C.E., Timing and rotational motions of Apuseni Mountains, Romania: palaeomagnetic data from Tertiary magmatic rocks, Tectonophysics, 1994, 233, 163- 176

[326] Balla Z., Neogene volcanic rocks and their significance in geodynamic reconstructions in the Carpathian region, Geotectonics, 1981, 15, 255-265

[327] Györfi I., Csontos L., Structural evolution of SE Hungary and Neogene basins of the Apuseni mountains (Romania), Romanian Journal of Tectonics and Regional Geology, Bucharest, 1994, 75, 19-20

[328] Takács E., Szalay I., Bodoky T., Hegedüs E., Kántor J.I., Timár Z., Varga G., Berczi I., et al., International deep reflection survey along the Hungarian Geotraverse, Geophys. Trans., 1996, 40, 1-44

[329] Istocescu D., Geological study of the western part of the Crişul Alb valley and Codru and Highiş border, Studii Tehnicul şi Economice, Series J, Bucharest, 1971, 8, 1-201 (in Romanian)

[330] Udubaşa G., Roşu E., Seghedi I., Ivăşcanu P.M., The "Golden Quadrangle" in the Metaliferi Mts., Romania: What does it mean? Romanian Journal of Mineral Deposits, 2001, 79, 24-34

[331] Neubauer F., Lips A., Kouzmanov K., Lexa J., Ivascanu P.M., Subduction, slab detachment and mineralization: The Neogene in the Apuseni Mountains and Carpathians. Ore Geol. Rev., 2005, 27, 13-44

[332] Berbeleac I., David M., Zămârcă A., Petrological and petrochemical data on the Tertiary volcanics from the Eastern part of the Zărand Mountains, D.S. Inst. Geol. Geofiz. (please expand), 1984, 48, 27-46

[333] Berbeleac I., Neacşu V., Zămârcă A., Bratosin I., Geochemistry and mineralogy of altered rocks and pyrites associated with the porphyry copper-gold mineralization of the Tertiary subvolcanic from Tălagiu, Zărand Mountains, Romanian Journal of Mineral Deposits, 1992, 75, 55-64

[334] Jude R., Tabacu-Borcea M., Ionescu O., Geologic and petrographic study of the igneous rocks belonging to the Caraciu volcano area (Metaliferi Mountains), Annales Institutul Geologie şi Geofizica, Bucharest, 1973, 40, 7-71 (in Romanian)

[335] Kouzmanov K., von Quadt A., Peycheva I., Harris R.C., Heinrich C.A., Roşu E., Ivascanu P.M., Miocene magmatism and ore formations in the South Apuseni Mountains, Romania: New genetic and timing constraints. In: Andrew C.J. et al. (Eds.) Proceedings of the Ninth Biennal SGA Meeting, Dublin, 2007, 865-868

[336] Wallier S., Ray R., Kouzmanov K., Petke T., Heinrich C., Leary S., O'Connor G., Tămaş C., et al., Magmatic fluids in the breccia-hosted epithermal Au-Ag deposit of Roşia Montana, Romania, Econ. Geol., 2006, 101, 923-954

[337] Papp D.P., Ureche I., Seghedi I., Downes H., Dallai L., Petrogenesis of convergent margin calc-alkaline rocks and the significance of low-isotope oxygen ratio: the Rodna-Bârgău Neogene subvolcanic area (Eastern Carpathians), Geol. Carpath. 2005, 56, 77-90

[338] Seghedi I., Szakács A., Mason P.R.D., Petrogenesis and magmatic evolutions in the East Carpathians Neogene volcanic arc (Romania), Acta Volcanologica 1995, 7, 135-145

[339] Tischler M., Gröger H.R., Fügenschuh, B., Schmid, S.M., Miocene tectonics of the Maramureş area (Northern Romania): implications for the Mid-Hungarian fault zone. Int. J. Earth. Sci., 2006, 96, 473-496 DOI:10.1007/s00531-006-0110-x

[340] Niţoi E., Munteanu M., Marincea Ş., Paraschivoiu V., Magma-enclaves interaction in the East Carpathians subvolcanic zone, Romania: petrogenetic implications. J. Volcanol. Geoth. Res., 2002, 118, 229-259

[341] Szakács A., Seghedi I., Pécskay Z., Peculiarities of South Harghita Mts as terminal segment of the Carpathian Neogene to Quaternary volcanic chain. Revue Roumaine de Geologie, Geophysique et Geographie, Serie de Geologie, 1993, 37, 21-36

[342] Szakács A., Seghedi I., Ioane D., Rogobete M., Time-space evolution and magma production rates in the Călimani-Gurghiu-Harghita volcanic chain (East Carpathians), Romanian Journal of Stratigraphy, 1995, 76, 55-56

[343] Szakács A., Ioane D., Seghedi I., Rogobete M., Pécskay Z., Rates of migration of volcanic activity and magma output along the Călimani-Gurghiu-Harghita volcanic range, East Carpathians, Romania, Przeglad Geologiczny, 1997, 45, 1106

[344] Karátson D., Timár G., Comparative volumetric calculations of two segments of the Carpathian Neogene/Quaternary volcanic chain using SRTM elevation data: implications for erosion and magma output rates. Zeitschrift für Geomorphologie N. F., 2005, 140, 19-35

[345] Szakács A., Krézsek C., Volcano-basement interaction in the East Carpathians: Explaining unusual tectonic features in the Eastern Transylvanian Basin, Romania, J. Volcanol. Geoth. Res., 2010, in press

[346] Panaiotu C.G., Pécskay Z., Hambach U., Seghedi I., Panaiotu C.C., Itaya C.E.T., Orleanu M., Szakács A., Short-lived Quaternary volcanism in the Perşani Mountains (Romania) revealed by combined K-Ar and paleomagnetic data, Geol. Carpath., 2004, 55, 333-339

[347] Vaselli O., Downes H., Thirlwall M.F., Dobosi G., Coradossi N., Seghedi I., Szakács A., Vannucci R., Ultramafic xenoliths in Plio-Pleistocene alkali basalts from the eastern Transylvanian Basin; depleted mantle enriched by vein metasomatism, J. Petrol., 1995, 36, 23-53

[348] Downes H., Seghedi I., Szakacs A., Dobosi G., James D.E., Vaselli O., Rigby I.J., Ingram G.A., et al., Petrology and geochemistry of late Tertiary/ Quaternary mafic alkaline volcanism in Romania, Lithos, 1995, 35, 65-81

[349] Embey-Isztin A., Dobosi G., Mantle source characteristics for Miocene-Pleistocene alkali basalts, Carpathian-Pannonian region; a review of trace elements and isotopic composition, Acta Vulcanologica, 1995, 7, 155-166

[350] Szabó C., Harangi S., Csontos L., Review of Neogene and Quaternary Volcanism of the Carpathian Pannonian Region, Tectonophysics, 1992, 208, 243-256

[351] Konečný V., Kováč M., Lexa J., Šefara J., Neogene evolution of the Carpatho-Pannonian region: an interplay of subduction and back-arc diapiric uprise in the mantle, EGS expand Special Publication Series, 2002, 1, 165-194

[352] Harangi S., Downes H., Seghedi I., Tertiary-Quaternary subduction processes and related magmatism in the Alpine-Mediterranean region. In: Gee D.G., Stephenson R.A., (Eds.), European Lithosphere Dynamics, 2006, Geological Society, London, 167-190

[353] Harangi S., Lenkey L., Genesis of the Neogene to Quaternary volcanism in the Carpathian - Pannonian region: Role of subduction, extension, and mantle plume, Geol. S. Am. S., 2007, 418, 67-90

[354] Pecskay Z., Lexa J., Szakacs A., Seghedi I., Balogh K., Konečný V., Zelenka T., Kovacs M., et al., V., Geochronology of Neogene magmatism in the Carpathian arc and intra-Carpathian area, Geol. Carpath., 2006, 57, 511-530

[355] Kaiser M., A geomorphic evolution of the Transdanubian Mountains, Hungary, Zeitschrift für Geomorphologie, 1997, 110, 1-14

[356] Moreno G.J., Fauquette S., Suc J.P., Vegetation, climate and palaeoaltitude reconstructions of the Eastern Alps during the Miocene based on pollen records from Austria, Central Europe, J. Biogeogr., 2008, 35, 1638-1649

[357] Vass D., Konečný V., Elečko M., Kozač J., Molnár A., Zakovič M., Diatomite deposit in the Jelšovec basalt maar and possibility of its exploitation, Mineralia Slovaca, 1998, 30, 333-356 (in Slovak)

[358] Vass D., Konečný V., Elecko M., Kozac J., Molnar A., Zakovic M., Lozisko diatomitu v bazaltovom maare pri Jelsovci a moznosti jeho vyuzitia, Mineralia Slovaca, 1998, 30, 333-356 (in Slovak)

[359] Konečný V., Lexa J., Konečný P., A guidebook to the southern Slovakia alkali basalt volcanic field. Field Guide - Second International Maar Conference (Hungary - Slovakia - Germany - 15-29 September, 2004) Bratislava, Slovak Republic, Geological Survey of the Slovak Republic, Bratislava, 2004

[360] Konečný V., Lexa J., Evolution of the phreatomagmatic/extrusive/intrusive complex of the Bulhary maar-diatreme volcano in Southern Slovakia, Geolines, 2003, 15, 47-51

[361] Kräutner H.G., Kräutner F., Szasz L., Udubaşa G., Istrate G., Geological map scale 1 :50000, Rodna Veche, Geological Institute of Romania, Bucharest, 1978

[362] Teodoru I., Teodoru C., Popescu-Tismana A., Geological and petrographic studies in South Bârgău Mountains, Dari de Seama ale Sedintelor, Institutul Geologie şi Geofizica, Bucharest, 1973, 59, 155-173 (in Romanian)

[363] Konečný V., Lexa J., Balogh K., Neogene - Quaternary Alkali Basalt volcanism in Central and Southern Slovakia (Western Carpathians), Geolines, 1999, 9, 67-75

[364] Berbeleac I., Iliescu D., Andrei J., Ciuculescu O., Ciuculescu R., Relationships between alterations, porphyry copper-gold and base metal-gold hydrothermal vein mineralizations in Tertiary intrusions, Talagiu area, Zărand Mountains, Romanian Journal of Mineral Deposits, 1995, 76, 31-39