Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, western Pannonian Basin, Hungary: a review
Abstract
The volcanic erosion remnants in the western Pannonian Basin are grouped into volcanic fields such as the Bakony – Balaton Highland Volcanic Field (BBHVF), Little Hungarian Plain Volcanic Field (LHPVF), and Styrian Basin Volcanic Field (SBVF). These volcanic fields were active in Late Miocene to Late Pliocene times and are located in the territories of Hungary, Austria and Slovenia. They are formed by at least 100 alkaline basaltic eruptive centres such as variably eroded scoria cones, tuff rings, maars, maar volcanic complexes, shield volcanoes, mesa flows and shallow subsurface intrusive complexes (dykes and sills). In this paper the basic volcanologic characteristics of the volcanic fields in the western Pannonian Basin are presented as a review of past research activities as well as the result of ongoing physical volcanological research in the region. It is demonstrated that the distribution of volcanoes in the western Pannonian Basin is structurally controlled by old structural elements in the pre-volcanic rock units. The geographical distribution of different vent types, such as phreatomagmatic versus magmatic vents, are in relationship with the general hydrogeological characteristics of the pre-volcanic rock units. In areas where thick Pannonian sandstone beds build up the underlying strata the so-called “normal maar volcanic centres” have usually a thick late magmatic infill in the maar basins. The eruptive mechanism of these volcanoes was controlled by the unconsolidated, water-saturated porous media aquifer, that lead to the formation of “champagne glass” shaped volcanoes and flat tuff rings. In areas where relatively thin Pannonian sandstone beds cover the thick Mesozoic or Palaeozoic fracture controlled karstwater bearing aquifer, large maar volcanic sequences are common, classified as Tihany-type maar volcanoes. These maar volcanic centres are commonly filled with thick maar lake deposits, building up Gilbert-type gravelly, scoria rich deltas in the northern side of the maar basins, suggesting mostly north to south oriented palaeo-fluvial systems. In the elevated, northern part of the field erosional remnants of scoria cones and associated shield volcanoes indicate a minor impact of the ground and surface water that may have led to phreatic and phreatomagmatic explosive activity.

Keywords: phreatomagmatic, volcanic glass, maar, tuff ring, scoria, hydrogeology, erosion, Pannonian Basin, vent, aquifer, dyke, sill, monogenetic

Introduction

The Bakony – Balaton Highland (BBHVF), the Little Hungarian Plain (LHPVF) and the Styrian Basin Volcanic Fields (SBVF) are located in the western Pannonian Basin, Hungary (Figure 1.1). The largest number of Neogene volcanic erosional remnants is in the BBHVF, which includes the Keszhely Mts., where shallow subsurface sills and dykes associated with monogenetic volcanoes are exposed (Plate 1.1). The BBHVF is close to the Lake Balaton north shore. The volcanic centres of the BBHVF were active between 7.54 My and 2.8 My (Balogh et al. 1982, Borsy et al. 1986, Balogh and Pécskay 2001, Balogh and Németh 2004 – Figure 1.2) and produced mostly alkaline basaltic volcanic products (Szabó et al. 1992, Embeý-Isztin 1993). The volcanoes in the LHPVF are in the same age and compositional range as the BBHVF (Balogh, et al. 1982, 1983, 1986, Harangyi et al. 1994, 1995, Pécskay et al. 1995, Szabó et al. 1995, Balogh and Pécskay 2001). The eruptive centres of the BBHVF have a close relationship with the eruption centres of the LHPVF eruptive centres according to their composition, age and general eruption mechanisms (Jugovics 1969b, 1972, Harangi et al. 1994, Németh and Martin 1999c, Martin et al. 2003). The two volcanic fields operated simultaneously (Balogh and Pécskay 2001, Wübrans et al. 2004) but the general palaeoenvironment (Kázmer 1990) and their hydrogeology could have caused different styles of eruptive mechanism mainly in the explosive volcanic activity (Németh and Martin 1999c).

The BBHVF itself alone has approximately 50 basaltic volcanoes in a relatively small (around 3500 km²) area (Jugovics 1969a), however, the number of vents maybe far more than 50 due to the existence of volcanic complexes and nested volcanoes (Martin et al. 2003). The volcanic erosional remnants form a more scattered distribution in the LHPVF, and the number of volcanoes is less (~10) than in the BBHVF (Jugovics 1915, 1916, 1972); however, volcanic erosional remnants buried under thick Quaternary deposits are known from the region (Tóth 1994). Individual Neogene alkaline basaltic volcanic erosional remnants are located close to the triple border of Hungary, Austria and Slovenia as well as close to the eastern metamorphic core complexes in the Eastern Alps (Jugovics 1916, 1939, Kralj 2000) together often referred as SBVF.

The volumetrically largest field of all is the BBHVF, often mentioned together with the LHPVF due to similarities in the age, timing and eruption mechanism. The BBHVF belongs to the Transdanubian Range unit, which is correlated with the Upper Austroalpine nappes of the east Alpine orogen (Majoros 1983, Kázmer and Kovác 1985, Tari 1991). The underlying basement of the volcanic fields consists of Palaeozoic rocks (Silurian schist, Permian red sandstone – Császár and Lelkesné-Felvár 1999) and a thick Mesozoic carbonate sequence (Budai and Vörös 1992, Budai and Haas 1997, HAAS and Budai 1999, HAAS et al. 1999). This basement forms a large-scale anticline structure of Eoalpine origin (Figure 1.3.) in the Transdanubian Range area and is locally covered by Tertiary sediments (Tari et al. 1992, 1999, Horváth 1993, Sacchi and Horváth 2002).

Tertiary sediments were deposited in local sedimentary basins on a regional erosional unconformity (Müller and Magyar 1992, Müller 1998, Tari and Pamic 1998, Juhasz et al. 1999, Müller et al. 1999). In the Neogene, just shortly before the volcanism started, a large lake, the Pannonian Lake, occupied main parts of the Pannonian Basin (Figure 1.4.), which had a very colourful sedimentary environment as reflected in the irregular basin morphology.
Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, western Pannonian Basin, Hungary: a review

Figure 1.1. Structural elements of the Carpatho-Pannon region (A) with respect to the Mio/Pliocene alkaline basaltic volcanic fields. The location of the volcanic erosional remnants of the Little Hungarian Plain and Styrian Basin Volcanic Fields (B) and the Bakony – Balaton Highland Volcanic Field (C) are shown in relevance to other rock units. The western Pannonian Basin was occupied by the Pannonian Lake around 9 My (D); that lake gradually vanished and had only small basins in the southern part of the Pannonian Basin 4.5 My ago (D). During onset of the volcanism, the region was occupied by an alluvial plain with a flat morphology, and potentially with large, but shallow water masses (D). Please note: The model on “D” is modified after MAGYAR et al. (1999).
Figure 1.2. K/Ar age distribution map of the Little Hungarian Plain (A) and the Bakony – Balaton Highland (B) Volcanic Fields (after Balogh et al. 1982, 1986, Borsy et al. 1986, and Balogh and Németh 2004)

The alkaline basaltic volcanism in the western Pannonian Basin was of a predominantly subaerial, intracontinental type. However, large shallow water bodies may have been present during eruptions, which most likely led to the formation of emergent volcanoes (KOKELAAR 1983, 1986, WHITE 1996a, 2001, WHITE and HOUGHTON 2000). These volcanoes quickly breached the water table of these lakes (MARTIN and NÉMETH 2002a, 2004b). The distribution of volcanoes in the western Pannonian Basin is related to the distribution of palaeo-valleys, which formerly were occupied by streams with good water supply (MARTIN et al. 2003). These "wet" valleys are most likely related to the reactivation of pre-Neogene fracture zones similar to the zones of structural weakness in the Eifel Volcanic Fields, Germany (LORENZ and BÜCHEL 1980a, b, BÜCHEL and LORENZ 1982, HUCKENHOLZ and BÜCHEL 1988, 1993, BÜCHEL et al. 2000).

After volcanism ceased, fluvial/alluvial sedimentation was widespread in the western Pannonian Basin. Major erosion affected the region well after volcanism ceased (CSILLAG et al. 1994, JORDÁN et al. 2003, NÉMETH et al. 2003b). Lake Balaton, as one of the major landmarks of western Pannonia, is a recent landform and its history dates back only 17,000 to 15,000 years (CSERNY and CORRADA 1989, CSERNY 1993, CSERNY and NAGY-BODOR 2000, TULLNER and CSERNY 2003). However, pre-Lake Balaton lacustrine systems very likely existed in the region also throughout the Quaternary (TULLNER and CSERNY 2003).

All types of eroded volcanoes can be found in the western Pannonian Basin (Plate 1.2) which show similar characteristics as most other monogenetic intracontinental volcanic fields such as Hopi Buttes, Arizona (WHITE 1991b, ORT et al. 1998), Western Snake River, Idaho (GREELEY 1982, GODCHAUX et al. 1992, BRAND 2004, WOODS and CLEMENS 2004), Waipiata Volcanic Field, New Zealand (NÉMETH and WHITE 2003), or West Eifel, Germany (LORENZ 1984). The most prominent geomorphologic formations are the circular, lava capped buttes. These centres are usually related to underlying phreatomagmatic volcanoes such as maar structures and tuff rings. Individual maar structures without lava infill are less common and difficult to identify. Such volcanic structures are locally filled by post-maar lava flows that subsequently have been buried under thick lacustrine units. In the northern part of the BBHVF Strombolian scoria cone remnants and Hawaiian spatter cone deposits are common. However, they commonly consist of scoria beds, which are inter-bedded with phreatomagmatic tuffs and lapilli tuffs, suggesting simultaneous Strombolian and phreatomagmatic activity. Large lava flow fields in the western Pannonian Basin only exist...
in the northern part of the BBHVF and they form the areas of highest elevation today (Plate 1.1). The lava flows are eroded and their type is often hard to reconstruct due to the advanced erosion. Smaller lava flows are inferred to have filled valleys. The largest lava fields are in the Bakony Mountains and consist of eroded shield volcanoes such as Kab-hegy and Agár-tető (Plate 1.1). Lava plugs intruded into small vents and are preserved to their higher resistance to erosion (Hegyes-tű).

The BBHVF is of great volcanological and palaeo-geomorphological interest. The relatively long volcanic history of the area (7.54–2.8 My — BALOGH and PÉCSKAY 2001) and the adjacent lacustrine to fluviatile environment make the western Pannonian Basin an ideal area for studying phreatomagmatic volcanoes in relation to a changing lacustrine environment and the palaeo-geomorphological evolution of the Late Miocene to Pliocene landscape. There is a great potential in developing our knowledge about:

1. eruption mechanisms resulting from magma/water interactions with different magma/water-ratio,
2. the relationship of volcanic activity and the confining palaeo-environment,
3. the related palaeohydrology and petrophysical characteristics of the pre-volcanic units, and
4. interrelationships between tectonics and lithospheric rheology that control the magma ascent in various tectonic regimes.

**Tectonic framework of the western Pannonian Basin and its relevance to the Neogene intraplate volcanism**

The Pannonian region has been considered to be part of the Alpine belt, and it reveals the complexity of orogenic evolution (HORVÁTH and TARI 1999). Continental to oceanic rifting of the Tethyan realm followed the Variscan convergence, subduction and continental collision, all shaping the Palaeozoic to Mesozoic substrata of the region. Subsequently, two periods of basin formation and development occurred in a compressional-transpressional regime during the Late Cretaceous and Palaeogene (TARI et al. 1993, TARI 1994). From the earliest Miocene large-scale lateral displacement and block rotation took place in the internal domain of the orogen, simultaneously with the formation of the Pannonian Basin (HORVÁTH 1993, CSONTOS 1995, FODOR et al. 1999, BADA and HORVÁTH 2001). This has been characterised by lithospheric extension, interrupted by compressional events (HORVÁTH 1995). Gravitational collapse of the Intra-Carpathian domain, combined with subduction zone roll-back are thought to have been the driving mechanism of the Neogene back-arc extension (RATSCBACHER et al. 1991, FODOR et al. 1999, BADA and HORVÁTH 2001, HORVÁTH et al. 2004), which gave way to widespread volcanism in the basin. The modern Pannonian Basin is in an initial phase of positive structural inversion, the related structural features are not yet fully developed (HORVÁTH and CLOETINGH 1996, GERNER et al. 1999, BADA et al. 2001). The structure of the basin system is the result of distinct modes of Miocene through Pliocene extension exerting a profound effect on the lithospheric configuration. In summary, the Miocene through Quaternary evolution of the Pannonian Basin was characterised by considerable depth-dependent lithospheric stretching (TARI et al. 1999) as a consequence of the collapse of former orogenic terrains and the subduction rollback of the Carpathian arc (CSONTOS et al. 1992, KOVAČ et al. 1994, FODOR et al. 1999, BADA and HORVÁTH 2001, HORVÁTH et al. 2004). Therefore, the Pannonian basin was classified as a typical Neogene back-arc basin in the Mediterranean system (HORVÁTH and BERCKHEMER 1982). These plate-scale processes led to the formation of the early Inner Carpathian and the late East Carpathian volcanic arcs (SZABÓ et al. 1992, LEYA 1999).

There is a large number of models describing the evolution of the Pannonian Basin which is summarized in BADA and HORVÁTH (2001), such as:

1. asthenospheric dome-triggered active rifting (STEGENA 1967),
2. active rifting that has been initiated by a subduction generated mantle updoming (HORVÁTH et al. 1975, STEGENA et al. 1975),
3. a hinge retreat of the subduction of the European margin driven by the negative buoyancy of the slab that induces trench suction forces and hence, passive rifting in the overriding plate (ROYDEN et al. 1983a, 1983b, ROYDEN and KARNER 1984), and
4. a similar hinge retreat is inferred to be sustained by an eastward mantle flow pushing against the downgoing slab (DOGLIONI et al. 1999) similarly to the inferred present situation in the Etna region (DOGLIONI et al. 2001).

In summary, these basin evolution models generally reflect two major views, i.e. active versus passive rifting.

The volcanism in the Pannonian Basin is marked by submarine pyroclastic and coherent trachyandesitic lava of Karpathian age (~17.5~16.2 My), Badenian (~16.2–3 My) pyroclastic and coherent basaltic, andesitic, dacitic and rhyolitic, and Late Miocene to Pliocene (~12~2 My in the western Pannonian Basin) alkali basaltic rocks, which are interlayered with coeval sedimentary rocks (SZABÓ et al. 1992). Structural interpretation of reflection seismic profiles reveals distinct modes of upper crustal extension in the Middle Miocene – Recent Pannonian Basin (TARI et al. 1992).
Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, western Pannonian Basin, Hungary: a review

While some sub-basins in the system show little extension (planar rotational normal faults), others are characterised by a large magnitude of extension (detachment faults, metamorphic core complexes — TARI et al. 1992). Seismic stratigraphic interpretations indicate that the non-marine post-rift sedimentary fill of the Pannonian Basin can be described in terms of sequence stratigraphy (VAKARCS et al. 1994).

Starting in the latest Miocene (Figure 1.4.), a considerable amount of basaltic magma erupted in the Transdanubian Range (TR) and Balaton Highland area in the Pliocene (SZABÓ et al. 1992, NÉMETH and MARTIN 1999c, NÉMETH et al. 2000). A total magma output has been estimated for the BBHVF on the basis of volume estimates of coherent lavas and dense rock equivalents of juvenile pyroclasts from phreatomagmatic and magmatic explosive eruptive products. This gave a minimum estimate of ~2.7 km³ and a maximum estimate of ~6.5 km³ total magma output (NÉMETH et al. 2000). A realistic estimate of 4±0.5 km³ of total magma output (NÉMETH et al. 2000).
over ca. 5.7 My time indicates that the BBHVF rather belongs to a volcanic field with low magma output ratio, which is typical for a strike-slip tectonic regime, or regions of moderate lithospheric extension such as the San Francisco Volcanic Field, Arizona. Approximately 90% of the total magma output is estimated to be basanitic in composition, leaving 10% for more differentiated rock types such as tephrite, phonolite, that are the major constituents of the phreatomagmatic pyroclastic deposits (NÉMETH et al. 2000). Estimating of the erupted volume of juvenile material from pyroclasts is more complicated due to the uncertainty of the eroded volumes. The tectonic role and style of magmatism is still under debate (SZABÓ et al. 1992, EMBEY-ISZTIN et al. 1993, HARANGI et al. 1995, KEMPTON et al. 1997). In general most of the workers agree that the Neogene alkaline basaltic volcanism from western and central Europe dominantly derived from asthenospheric partial melting (EMBEY-ISZTIN et al. 1993). However, there are growing evidences that in most cases they were modified by melt components from the enriched lithospheric mantle through which they have ascended (EMBEY-ISZTIN et al. 1993). A debate on the role of crustal contamination of the magmas still exists. Most of the models deal with some contamination from a subducted slab from former subduction in the region and/or some degree of metasomatozism (BALI et al. 2002). In general, the incompatible-element patterns of the lavas of the volcanic fields of the western Pannonian Basin show that these lavas are relatively homogeneous (EMBEY-ISZTIN et al. 1993) and are enriched in K, Rb, Ba, Sr, and Pb with respect to average ocean island basalt such as Hawaii, and resemble alkali basalts of Gough Island.

Heat flow values in the Pannonian Basin greatly vary (DÖVÉNYI and HORVÁTH 1988), and are up to 300 mW/m² in areas of volcanic fields that are inferred to be located above shallow magma chambers (SACHSENHOFER et al. 1997, 1998, 2001). Elevated heat flow values from the Austrian basins were obtained from areas related to the Eastern Alps, a consequence of rapid uplift of the whole orogen (SACHSENHOFER et al. 1997, 2001, SAChSENHOFER 2001).

In spite of the large number of data available from tectonic and geochemical studies, the question of the relationship between volcanism and the Neogene tectonic processes in the Pannonian Basin has not been addressed adequately yet. One possible explanation, which is in agreement with other field-based observations such as geomorphological evolution and facies relationships, could be that magmatism is related to the latest stage of post-rift faulting (FODOR et al. 1999). In this course, basaltic volcanism may have occurred after the cessation of post-rift sedimentation and preserved incipient denudation surfaces (NÉMETH and MARTIN 1999d, NÉMETH et al. 2003b). The Pliocene basaltic volcanism may also belong to the late-stage inversion of the Pannonian Basin, which was generally associated with uplift and denudation (HORVÁTH 1995, HORVÁTH and CLOETINGH 1996).

Lithospheric structure and magma ascent of Neogene alkaline basaltic volcanic fields

Continental monogenetic volcanic fields are subject to the same physical constraints as other volcanic systems. Dense mantle-derived magmas are prone to pond near their levels of neutral buoyancy, at depths of 25–30 km, in the upper mantle/continental crust boundary and/or in rheological and density contrast zones between the brittle/ductile transition in mid-crustal levels (RYAN 1987a, b, WALKER 1989, LISTER 1991, LISTER and KERR 1991, WATANABE et al. 1999, 2002). Eruption of such magmas in small volumes requires substantial injected volumes of which only a small proportion reaches the surface, and/or specific stress conditions within the transected lithosphere (LISTER 1991, WATANABE et al. 1999).

Various mechanical considerations of fluid-filled crack propagation (e.g. LISTER 1991) through a lithosphere that is under tectonic stress conclude that either extension (e.g. ascend of mantle material and then upflow of magma along “open” (extensional) fractures) or tectonic inversion (e.g. mantle material ponds at the Moho and other density and/or rheology contrast zones in the lithosphere and then magma is expelled by tectonic forces) seem to be a viable mechanism for volcanic activity of the Neogene alkaline basaltic volcanism in the western Pannonian Basin.

During pure extension, when the vector of the maximum compressional stress is vertical and the minimum compressional stress is in horizontal position (predominantly normal faulting) vertical dyke propagation is favoured (LISTER 1991, WATANABE et al. 1999). Magma can reach the surface and predominantly form monogenetic, deep-rooted volcanoes (WATANABE et al. 1999). When the maximum compressional stress is in horizontal and the minimum is in vertical position, melt intrusion is only possible when this configuration temporarily switches into either pure extension or to a period when the maximum and minimum compressional stress orientation change place (WATANABE et al. 1999). If the switching period is short, magma can be trapped and form sill-like reservoirs. Further movement toward the surface is possible during a new switching period, which process must lead to a multiple level magma “pocket” build up through a series of so called “failed eruptions”. When the maximum compressional stress is in vertical and the minimum in horizontal position, but their differential stress is small (e.g. strike slip system) magma gradually can reach the
surface through multiple “failed eruptions” (WATANABE et al. 1999). In this situation a large magma supply rate is necessary in general for the magma to reach the surface. In this condition the generation of polygenetic volcanic systems is favoured if the differential stress is generally small in absolute value and does not change significantly with depth. In conditions when the differential stress is variable according to the depth (e.g. depending on the changeable physical conditions of the crust — temperature, lithology distribution etc.) both polygenetic volcanoes and monogenetic volcanic fields can form.

Locations of highly differentiated silicic igneous bodies such as in the axis of the Little Hungarian Plain sub-basin, which is filled by about 3000 m thick low density Miocene to recent siliciclastic sediments (TAIRI 1994) and cap a metamorphic complex, suggest that these thick, low-density rock-filled basins may have functioned as a density trap (WALKER 1989) and led to magma chamber formation where alkaline basaltic magma fractionated to trachyte, trachyandesite in the otherwise predominantly alkaline basaltic volcanic products. Similar density traps have been reported in the Yucca Mts area (Nevada, USA), where low density thick (up to 6 km) ignimbrites that filled a basin functioned as density trap against the ascent of the otherwise buoyant, hot basaltic melt (CONNOR et al. 2000). The inferred timing (~12–9 My) of the formation of such otherwise bimodal (trachyte – basalt — SCHLEDER and HARGÁNYI 2000) polygenetic volcano in the axis Little Hungarian Plain region is in good concert with the predominantly strike-slip controlled tectonic regime in the region in this time (TAIRI et al. 1992, TAIRI 1994). Perhaps the alkaline basaltic volcanism in the western Pannonian Basin post-dates this complex volcano and is coincident with the general tectonic inversion that most of the workers accept but the reason of this volcanism is not fully understood. From a tectonic point of view, magma may reach the surface in the general compressional regime when it may switch for various time length to be pure extensional or strike-slip dominated. The general complexity of the Neogene “so called” monogenetic volcanoes of the western Pannonian Basin from both a geochemical (see later) and volcanological point of view, and the newly identified dyke and sill complexes associated with these volcanoes suggest that a temporal switch from a compressional tectonic regime to a more strike-slip dominated regime may be a sensible reason for the melt to reach the surface. However, the general agreement on fast uprise of the basaltoid melt that carry large mantle nodules somehow indicate more pure extensional periods temporarily during the general compressional regime. Because up to now there is no clear sign of temporal and/or spatial distribution of volcanic features indicative for these two major scenarios, we suggest that in a generally “unstable” compressional regime, the tectonic stress field may have switched to either pure extension or strike-slip regimes according to other controlling factors such as changes of heat flow or position of the mantle anomaly. For a working hypothesis, if there is any temporal change in tectonic regime (e.g. gradual transition from pure extension to pure compression as a result of the tectonic inversion),

1. monogenetic volcanoes (e.g. short-lived volcanoes with simple architecture) that issued lavas and/or pyroclastic rocks carried large mantle nodules are expected to be older and

2. younger volcanic edifices may be more complex, often associated with sill complexes. There are sill and dyke complexes that are relatively young (~3 My) and volcanoes erupted large mantle nodules that are old, but it is too early to draw any broader conclusion on the basis of the very limited research that has been carried out in this respect.

In summary, it can be concluded that the lithosphere in the Pannonian Basin has a very complex structure. The crust as well as the lithosphere is strongly attenuated and high heat flow prevails (hottest basin in continental Europe). Rock units with variable density, rheology and heat conductivity could have facilitated magma ponding that might have led to further magma evolution via fractionation in these magma storage places (chambers). Such lithospheric architecture could temporarily trap otherwise “fast” uprising basaltoid melt of mantle origin when the regional stress field is switching to a more compressive regime for a short time interval. Such a scenario is very likely during basin evolution dominated by strike slip faulting, such as is inferred for the western Pannonian Basin during the onset of volcanism in the Late Miocene through Pliocene.

Geochemistry and petrogenesis of eruptive products of the Miocene to Pliocene volcanism in the western Pannonian Basin

Alkaline basaltic volcanism throughout the Neogene was widespread in the Pannonian Basin, leading to the formation of distinct volcanic fields. Volcanism, in general, since the Miocene is associated with the tectonic development of the Carpatho-Pannonian region and is connected with the formation of the Pannonian Basin (SZABÓ et al. 1992). Volcanic activity has previously been divided into three main genetic types (SZABÓ et al. 1992) according to the common composition, eruption style, and location of the volcanic centres:

1. Early Miocene mainly acidic explosive volcanism that led to the accumulation of extensive ignimbrite sheets (welded and non-welded type, predominantly rhyolitic in composition), block-and-ash flow deposits as well as their
reworked volcano-sedimentary units, often intercalated with normal marine to terrestrial sediments (Liffa 1940, Pantó 1963, 1966, Hámor et al. 1980, Szeky-Flux and Kozák 1984, Póka 1988, Capaccioni et al. 1995, Szakács et al. 1998, Leva 1999). The recent spatial distribution of this volcanic province exhibits a great separation which is inferred to be the result of a large right lateral displacement along the Mid-Hungarian shear zone (Figure 1.1) during the Early Miocene (Stegen et al. 1975, Balla 1980, 1981, Royden et al. 1983a).

2. Middle Miocene – Pliocene calc-alkaline, mainly intermediate stratovolcanic complexes in the Inner Western Carpathians and in the East Carpathians, related to a subducted oceanic slab (Konečný et al. 1999b). Their geochemical compositions show a transitional character between active continental margin and island arc type magmatic rocks (Downes et al. 1995a, b). The thickness of the crust increased with time and from west to east beneath this volcanic arc. The small-to-medium sized block-and-ash flow dominated lava dome fields (Zelenka 1960, Balla and Korpás 1980, Korpás and Lang 1993, Karátszon et al. 2000) and their associated reworked volcanlastic successions often formed thick accumulations of volcanlastic mass flows (Karátszon and Németh 2001). Such volcanlastic successions can be traced in the entire Carpathian Volcanic Chain and often bear significant information on basin evolution.

3. Pliocene–Pleistocene alkali basaltic volcanism in the Pannonian Basin is considered to have been related to an up-welled, then cooled asthenospheric dome (Szabó et al. 1992). This is thought to have induced the thermal regime from which magmatic melts ascended (Szabó et al. 1992). Various geochemical studies of the alkaline basalts suggest mantle up-welling as a major driving force of the alkaline basaltic Neogene volcanism (Szabó et al. 1992). The compositional difference in space and time are inferred to reflect the existence of local individual small-sized diapiric bodies as well as several processes (e.g. fractional crystallisation, mixing), which modified the original magma (Szabó et al. 1992). Such diapiric bodies are often referred to as common Central European mantle up-welling feeding volcanic fields across Europe in the Neogene (Duda and Schmincke 1978, Wilson and Bianchi 1999). The eruption of basaltic melts was temporally associated with the final phase of the development of the Pannonian Basin, however, it often has been considered to post-date the cessation of the post-rift sedimentation. The accumulation of volcanic debris on an erosional surface and the volcanism itself is coeval with the start of the basin inversion as was pointed out earlier (Horváth 1995, Fodor et al. 1999, Németh and Martin 1999d, Martin et al. 2003).

There is a general agreement regarding the petrogenesis of Neogene alkali basaltic rocks in the Pannonian Basin. The basalitic rocks were formed during the Late Cenozoic post-orogenic phase and their eruption was related to the evolution of the extensional Pannonian Basin following Eocene–Miocene subduction and its related calc-alkaline volcanism (Szabó et al. 1992, Embeý-Isztin et al. 1993). The alkaline volcanic centres, dated by K/Ar methods are between ~12 and 1 My in age (Peckskay et al. 1995), forming well-distinguishable volcanic fields. Some fields are near the western (Graz Basin, Burgenland, Slovenia), northern (Nógrád–Gömör/Gemer), and eastern (Eastern Transylvanian Volcanic Field) margins of the basin, but the majority are concentrated near the Transdanubian Range (Bakony – Balaton Highland and Little Hungarian Plain Volcanic Field). Coherent lavas range from slightly hy-normative transitional basalts through alkal basalts and basanites to olivine nephelinites. No highly evolved coherent lava (extrusive or intrusive) compositions have been identified from any of the locations yet (Szabó et al. 1992, Embeý-Isztin et al. 1993, Harangi et al. 1995). This makes the Pannonian Neogene basaltic volcanic fields different from other European volcanic regions such as the Eifel, where e.g. phonolitic lava flows as well as ignimbrites are common (Bogard and Schmincke 1985, Freundt and Schmincke 1986, Bednarz and Schmincke 1990, Harms and Schmincke 2000). The presence of mantle peridotite xenoliths, xenocrystals, and high-pressure megacrysts in coherent lavas, even in the slightly more evolved ones and in pyroclastic rocks, is inferred to indicate that differentiation took place within the upper mantle (Downes et al. 1992). However, the mantle source often is referred to be heterogeneous (Dobosi 1989, Dobosi et al. 1991, Dobosi and Fodor 1992, Szabó and Bodnar 1995, 1998, Dobosi et al. 2003). The study of peridotite xenoliths revealed a strong relationship between deformation and temperatures of peridotites, in as much as coarse-grained protogranular and poikilitic xenoliths had high temperatures (up to 1197 °C), whereas fine-grained equigranular and mosaic xenoliths had low temperatures (800–900 °C – Embeý-Isztin et al. 2001). This picture suggests that diapiric uplift of hot mantle material into a cooler uppermost mantle has probably taken place (Embeý-Isztin et al. 2001).

Isotope geochemistry

The Sr and Nd isotope ratios from the Neogene coherent lava flows of the Pannonian Basin span the range of Neogene alkali basalts from Western and Central Europe (Duda and Schmincke 1978, 1985, Mertes and Schmincke 1985, Bednarz and Schmincke 1990), and suggest that the magmas of the Pannonian Basin dominantly derived from asthenospheric partial melting. Pb isotope studies, however, indicate that in most cases the asthenospheric melt composition was modified by melt components from the enriched lithospheric mantle through which the magma ascended (Embeý-Isztin et al. 1993). Various metasomatic processes may have interacted with the uprising melts (Bali et al. 2002), similarly to other alkaline volcanic provinces in Central Europe (Witteickschen et al. 1993, Shaw 1997, Sachs
and HANSTEEN 2000, SHAW and EYZAGUIRRE 2000). Delta $^{18}$O values indicate that the magmas have not been significantly contaminated with crustal material during ascent and isotopic and trace-element ratios therefore reflect mantle source characteristics (EMBEY-ISZTIN et al. 1993). The uniform oxygen isotope ratio in the phenocrysts suggests that the mantle source of the alkali basalts was also homogeneous with respect to its oxygen isotope composition, which is in contrast to the relatively wide variation of Sr, Nd and Pb isotope ratios in the source (DOBOSI et al. 1998). Variations in radiogenic isotope compositions in the basalts have been interpreted as result of the interaction of subduction-related fluids with the mantle source of the basalts. If this was the case, then the fluids, which caused significant changes in the Sr and Pb isotope ratios of the mantle source, did not noticeably modify its oxygen isotope composition.

Incompatible-element patterns show that the basic lavas, which erupted in the Balaton area and Little Hungarian Plain, are relatively homogeneous and are enriched in K, Rb, Ba, Sr, and Pb with respect to average ocean island basalt, and resemble alkali basalts of Gough Island type (EMBEY-ISZTIN et al. 1993). In addition, $^{207}$Pb/$^{204}$Pb is enriched relative to $^{206}$Pb/$^{204}$Pb. In these respects, the lavas of the Balaton area and the Little Hungarian Plain differ from those of other regions of Neogene alkaline magmatism of Europe (EMBEY-ISZTIN et al. 1993). This may be due to the introduction of marine sediments into the mantle during the earlier period of subduction and metasomatism of the lithosphere by slab-derived fluids rich in K, Rb, Ba, Pb, and Sr. Lavas erupted in the peripheral areas have incompatible-element patterns and isotopic characteristics different from those of the central areas of the basin, and more closely resemble Neogene alkaline lavas from areas of western Europe where recent subduction has not occurred (EMBEY-ISZTIN et al. 1993). However, in this respect there is no agreement yet. The alkaline volcanic activity that occurred in the Persani Mountains (eastern Transylvanian Basin) and Banat (eastern Pannonian Basin) regions of Romania between 2.5 My and 0.7 My (DOWNES et al. 1995b) produced coherent alkaline basaltic lavas that are primitive, silica-undersaturated alkali basalts and trachybasalts (7.8–12.3 wt.% MgO; 119–207 ppm Ni; 210–488 ppm Cr), which are LREE-enriched (DOWNES et al. 1995b). Mantle-normalised trace-element diagrams revealed an overall similarity to continental intraplate alkali basalts, but when compared to a global average of ocean island basalts (OIB), the Banat lavas are similar to average OIB, whereas the Persani Mts. basalts have higher Rb, Ba, K and Pb and lower Nb, Zr and Ti. These features slightly resemble those of subduction-related magmas, particularly those of a basaltic andesite related to the nearby older arc magmas (DOWNES et al. 1995b). With $^{87}$Sr/$^{86}$Sr varying from 0.7035–0.7045 and $^{143}$Nd/$^{144}$Nd from 0.51273–0.51289, the Romanian alkali basalts are indistinguishable (DOWNES et al. 1995b) from those of the western Pannonian Basin (Hungary and Austria — HARANGI et al. 1994, 1995, EMBEY-ISZTIN and KURAT 1997) and Neogene alkali basalts throughout Europe. It is inferred that, although the Romanian alkali basalts have a strong asthenospheric (i.e. OIB-type mantle source) component, their Pb isotopic characteristics were derived from mantle, which was affected by the earlier subduction (DOWNES et al. 1995b). It is in general agreement that Neogene alkaline basaltic rocks in the Pannonian Basin have some characteristics that represent some influence by former subduction in the region and associated metasomatic processes in the mantle.

Petrography of pyroclastic rocks

The western Pannonian volcanic fields also consistently comprise basal vitric pyroclastic units overlain by lavas (NÉMETH and MARTIN 1999c, MARTIN et al. 2003). The pyroclastic rocks of the volcanic fields contain various proportions of country rock clasts, which apparently represent vent-filling assemblages. Locally there are well-bedded tuff ring deposits preserved (Figure 1.5). Dykes and lava flows have sub-planar to highly irregular, locally pephritic (MARTIN and NÉMETH 2000), contacts with pyroclastic rocks, suggesting intrusion shortly after emplacement of the tuffs and tuff breccias while they were still unconsolidated. The pyroclastic rocks typically have aphyric or sparsely feldspar-phylitic juvenile clasts (sideromelane glass shards), whereas the slightly younger dykes and lavas are characterized by abundant pyroxene±kaersutite phenocrysts. The volcanic glass shards are variously shaped from blocky to strongly stretched being microvesicular and/or containing abundant microlites and/or microphenocryst (Plate 1.3). The vesicle morphology of the glass shards exhibits features characteristic for both magmatic degassing and sudden collapse due to cooling of the melt by magma.
water interaction (Figure 1.6). There are abundant deep-seated xenoliths, 1 to 15 cm in size, in the uppermost beds of pyroclastic deposits at some of the volcanoes (Plate 1.4).
Major element variations in monogenetic volcanoes

Compositional variations among eruptive products of individual volcanoes just recently have been studied in detail from the Pannonian region (MARTIN et al. 2003, NÉMETH et al. 2003c). In former studies it is assumed that monogenetic volcanoes are small to very small volcanoes such as scoria cones, tuff cones and rings, and maars, which formed by single, typically brief eruptions (WALKER 1993). Monogenetic volcanoes might form in 2 distinct settings:

1. as isolated fields of volcanoes on continental lithosphere, ranging from thinned lithosphere (<30 km) resulting from stretching and extension (e.g. Ethiopia, Basin and Range — BARBERI and VARET 1970, ARANDA-GOMEZ et al. 1992) to normal or thick lithosphere (e.g. San Francisco field, Hopi Buttes etc. — CONWAY et al. 1997, 1998), and

2. as “parasitic” vents along the rift zones or flanks of large polygenetic central volcanoes (e.g. Tolbachik (Russia), La Palma (Spain), Mauna Loa (Hawaii, USA), Tavenui (Fiji), Sawai (Western Samoa) etc. — FLEROV and BOGOYAVLENSKAYA 1983, DOUBIK and HILL 1999). Some single eruptions forming monogenetic volcanoes atop large central volcanoes are known to have produced petrologically variable magmas (KLÜGEL et al. 1999, 2000) that reflect the presence of magma reservoirs within the large volcano. Such a variation has not been demonstrated in detail in single (small volume) monogenetic volcanoes of continental fields, which are thought to lack stable magma-storage zones. However, a general trend of compositionally more evolved eruption products in higher stratigraphic level in the volcanic units of complex phreatomagmatic-to-magmatic volcanoes from the Eifel (Germany) region have been reported (DUDA and SCHMINCKE 1978, HOUGHTON and SCHMINCKE 1989). Compositional variations among scoria cones in volcanic fields in a single cone

Table 1.1. Composition of volcanic glass shards from pyroclastic rocks of erosion remnants of Neogene alkaline basaltic volcanoes of the western Pannonian Basin
have been recently described from the Transmexican Volcanic Belt (STRONG and WOLF 2003, SIEBE et al. 2004). However, scoria cones from the Transmexican Volcanic Field often form transition between monogenetic and composite volcanoes (McKNIght and WILLIAMS 1997). In contrast, monogenetic volcanoes are formed by more or less direct eruption of magma from the mantle, with each volcano resulting from successful propagation of a small batch of magma to the surface along a new pathway (SPEERA 1984, HASENAKA and CARMICHAEL 1985, HASENAKA and CARMICHAEL 1987, HASENAKA 1994, CONNOR and CONWAY 2000).

Volcanic rocks from the western Pannonian Basin were subject of whole-rock analyses that gave systematically basaltic composition (EMBEY-ISZTIN 1993). In spite this, electron microprobe analyses on volcanic glass shards from associated, phreatomagmatic pyroclastic rocks (JEOL 8600 Superprobe, housed in the University of Otago, Geology Department, 15 kV acceleration voltage, 5–20 µm electron beam diameter, OXIDE9 standard, and ZAF correction method) systematically gave a more evolved tephritic, phono-tephritic composition (MARTIN et al. 2003, NÉMETH et al. 2003c — Table 1.1). The composition of the erupted magmas in the studied areas falls to the alkali basalt field, with the dominant magma type being basanitic (Table 1.1 and Plate 1.4). The pyroclastic rocks are commonly more evolved than the lava flows from the same sites. Volcanic glass shards from all sites are predominantly tephritic, phonotephritic in composition with a minor proportion of tephriphonolitic or trachybasaltic glass shards (Table 1.1 and Figure 1.7). Compositional variations of the initial pyroclastic sequences and subsequent lava flows and/or dykes suggest a complex magma evolution within a relatively short period of time (hours to weeks). This compositional bimodality of tuff ring formation and lava flow sequences can be explained in two different ways:

1. by the presence of "readily" evolved tephritic–phono-tephritic melt at upper crustal level, which — after a short period of residence (days to weeks) — continued its way to the surface and interacted explosively with external water or water-saturated sediments. Shortly after emptying these shallow-level "micro" magma storage places, a deep-sourced basanitic melt reached the surface and generated scoria cones and/or subsequent lava flows and lava lakes, which were commonly involved in peperite-forming processes at each locality (MARTIN and NÉMETH 2000, 2004c). This model is similar to that described from the Canary Islands (KLÜGEL et al. 2000). Alternatively,

2. the ascending melt evolved during its way to the surface, producing individual chemically zoned magma batches with evolved top levels and less evolved bottom parts, as suggested for the Rothenberg volcano in the German Eifel (HOUGHTON and SCHMINCKE 1989). The top level of each initial magma batch interacted with external water causing phreatomagmatic explosions. After exhausting the external water supply, a lower magma batch which was less evolved (basanite) managed to reach the surface without phreatomagmatic interaction, filling the craters and experiencing intensive interaction with the unconsolidated water-rich slurry that occupied the vent zones leading to peperite-forming processes (MARTIN and NEMETH 2002a,b, 2004b).

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**Figure 1.7.** TAS diagrams showing the composition of volcanic glass shards and subsequent lava flows from the BBHVF

Sample abbreviations (first number refers to the sample number, the numbers in brackets refer to the location of the sample and correspond to the numbers shown on Plate 1.1): 53 — Horog-hegy (20), K28 — Kopasz-hegy (15), K29 — Kereki-domb (18), 43 — Fekete-hegy south (5), 52 — Öreg-hegy (19), KH30 — Kis-Hegyes-tú (22), 64 — Hajagos (14), 65 — Hajagos (14), Sz3 — Szigliget, Kamon-kő (8), K27 — Kopasz-hegy (15), 48 — Fekete-hegy south (5), 77 — Pula (25), VD — Boglár, Vár-domb (10), HJ27 — Hajagos (14), SzV3 — Szigliget, Vár-hegy (8), UJ22 — Hajagos (14), SzK31 — Szentbékkálla mafic pyroclastic flow (21), SzK19 and SzK12 — Szentbékkálla (21), II/11, BlB and TH08 — Tihany Maar Volcanic Complex (33)
Age of the Neogene intraplate volcanoes in the western Pannonian Basin
and their relationship to the immediate pre- and syn-volcanic sedimentation in the region

Intracontinental Mio/Pliocene volcanic fields of the western Pannonian Basin developed between 7.56 and 2.3 My (BAŁOGH et al. 1986, PECSKAY et al. 1995, BALOGH and NÉMETH 2004) across an area in size of about 50,000 km² (Figure 1.1). In the western Pannonian Basin, there are two closely related volcanic fields, that contain the largest number of volcanoes,
1. Bakony – Balaton Highland Volcanic Field (BBHVF) and
2. Little Hungarian Plain Volcanic Field (LHPVF — Plate 1.5).

Phreatomagmatic volcanoes in the northern LHPVF tend to comprise broader, lensoid landforms and in their crater/vent volcanic facies peperites are common (MARTIN and NÉMETH 2004a, c). The depth of magma-water interaction in these volcanoes is inferred to have been shallow (MARTIN and NÉMETH 2004b). The presence of peperites indicates that the host sediment (both siliciclastic and pyroclastic) into which the magma intruded or on which the lava erupted was water saturated (MARTIN and NÉMETH 2000). In the BBHVF, especially in the central and eastern part, a large number of volcanic remnants exhibit features that are characteristic for magma-water interaction in deeper zones (e.g. karst water) of the pre-volcanic sedimentary sequence (NÉMETH et al. 2001).

Shallow lakes may have existed in an alluvial plain during onset of volcanism (MAGYAR et al. 1999), which may have led to shallow subaqueous-to-emergent volcanism (Figures 1.1 and 1.4). Textures of pyroclastic rock units as well as the common occurrence of peperites prove this (MARTIN and NEMETH 2004b). Shallow lacustrine siliciclastic sedimentary units that deposited in these shallow lakes represent the immediate pre-volcanic rock units of the volcanic facies in the western Pannonian Basin. On the basis of unconformity-bounded sedimentary units in the Neogene sequence of the continental sedimentary record of the western Pannonian Basin, three major maximum flooding surfaces have been identified and dated by magnetostratigraphic correlation to be 9.0 My, 7.3 My and around 5.8 My (LANTOS et al. 1992, SACCHI et al. 1999, SACCHI and HORVÁTH 2002). The first maximum flooding event correlates with the Congeria czjzeki open lacustrine beds (LŐRENTHEY 1900, MÜLLER and MAGYAR 1992, MAGYAR 1995, MAGYAR et al. 1999), which marks the Lower Pannonian stage of LŐRENTHEY (1900). After the flooding event, a significant base level drop and subaerial erosion took place around 8.7 My (MÜLLER and MAGYAR 1992, SACCHI et al. 1999). The second maximum flooding event took place around 7.3 My ago and is represented by the appearance of Congeria rhomboidea beds (MÜLLER and MAGYAR 1992, MAGYAR et al. 1999, SACCHI et al. 1999). General low-stand and subaerial conditions in the marginal areas are estimated to have occurred around 6 My ago (SACCHI et al. 1999), which was followed by the last known flooding around 5.3 My ago. However, this flooding event has not affected the region of the western Pannonian volcanic fields. It has reached only the southern margin of the basin (MAGYAR et al. 1999).

The intensive geochronological research carried out in the past decades on young alkaline basaltic rocks from the Pannonian Basin has confirmed that K/Ar data on these rocks give the reasonable geological ages and the most frequent error is caused by the presence of excess Ar (BAŁOGH et al. 1996). In spite of the presence of excess Ar detected from the Neogene basaltic rocks of the Pannonian Basin the geological age of these rocks has been obtained by applying the isochron methods (MCDOUGALL et al. 1969, HARPER 1970, MCDOUGALL and COOMBS 1973, SHAFIQULLAH and DAMON 1974, HAYATSU and CARMICHAEL 1977, MCDOUGALL and DUNCAN 1980, MCDOUGALL et al. 2001).

Although there are analytical and sampling difficulties a great number of K/Ar age data are available from the Neogene basaltic volcanic rocks from the western Pannonian Basin. There is no apparent spatial distribution pattern among major age groups of volcanic rocks (Figure 1.2). The age...
dates between 8 to 2.3 My seem to be randomly scattered in the area (Figure 1.2). There is a general centre point of ages at around 3.5–4 My BP, derived from volcanic remnants in the western part of the BBHVF (Figure 1.2). On the basis of new, laser induced step heating \(^{39}\text{Ar} / {^{40}\text{Ar}} \) high precision ages, it seems, that one of the major part of the alkaline basaltic volcanism in the region, falls into the 3.8–4 My old period, which is in good concert with the previous K/Ar dates (Halom-hegy/3.78, 3.82, Hajagos/3.81, Hegyesd/3.9, Fekete-hegy lava field/3.81, Szilisiglet diatreme Várhegy pyroclastic sequence/4.08 — BALOGH et al. 1986, BORSY et al. 1986, WIJBRANS et al. 2004). An older age group of volcanoes can be identified on the basis of the \(^{39}\text{Ar} / {^{40}\text{Ar}} \) ages at around 4.2 to 4.8 My (Szent György-hegy/4.22, Szilisiglet lava/4.53, Kis-Somlyó/4.63 and Tőti-hegy/4.74), however dates from Szilisiglet is likely to be in error, and they rather belong to the previous age group (WIJBRANS et al. 2004). These numbers also represent similar values than previous K/Ar dates from the same volcanoes (BALOGH et al. 1986, BORSY et al. 1986). The oldest known volcanic remnants are in Tihany, and their ages are fixed at around 8 My by repeated attempt to obtain isochron dates by the K/Ar method (BALOGH and NÉMETH 2004). The youngest volcanic edifices are erosion remnants of scoria cones topping the Haláp (3.08 \(^{39}\text{Ar} / {^{40}\text{Ar}} \) — WIJBRANS et al. 2004), Agár-tető (2.9 K/Ar — BALOGH et al. 1986), Füzes-tó (2.64 \(^{39}\text{Ar} / {^{40}\text{Ar}} \) — WIJBRANS et al. 2004) and Bondoró (2.3 K/Ar — BALOGH and PÉCSKAY 2001). Among these locations are the most well-preserved scoria cones in the western Pannonian Basin, which are still holding some primary morphology as well-defined crater rim (Figure 1.8).

Overall it can be summarised that the Neogene basaltic volcanism was active in the western Pannonian Basin between ~ 8 and 2.3 My, having a total duration of 5.3 My.

**Distribution of volcanic erosion remnants**

Studies of vent distribution in a volcanic field are very useful to establish the relationship between volcanism and tectonism and give some conclusions on the relationship between structural elements and the location of volcanic edifices (CONNOR and CONWAY 2000). Such studies have been successfully applied on various volcanic fields. Major tendencies of vent migration and tectonic events have been found elsewhere (CONNOR 1990, TOPRAK 1998, CONNOR et al. 2000, ROWLAND and SIBSON 2001).

Volcanic erosion remnants in the western Pannonian Basin are clustered into three major well-distinguished volcanic fields. The Styrian area (in Austria and Slovenia) is well-separated from the Bakony – Balaton Highland and Little Hungarian Plain Volcanic Fields (both are in Hungary), and features only largely separated vents. Volcanic erosional remnants from the BBHVF and LHPVF are not clearly separated from each other and the transition between these two fields is more continuous (Plate 1.5). The significant difference between these two fields is that the apparent vent density in the BBHVF is larger, and vent clustering is more prominent. Vent alignment is more characteristic in the LHPVF, where vents seemingly follow the Rába Fault Zone (Plate 1.5).

The distribution of identified volcanic erosion remnants in the BBHVF is represented by contouring vent density on the basis of a rectangular grid with uniform spacing of 2 km and a search radius of 5 km (Figure 1.9, A). On this map, the BBHVF is characterised by one major vent cluster in the geometrical centre of the field and by two additional clusters, one in the east and one in the west, all together forming a more or less east–west-trending alignment (Figure 1.9, A). The highest vent density reaches 20 vents in an area of 80 km\(^2\) (0.25 vents/km\(^2\)), centred around a nested
Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, western Pannonian Basin, Hungary: a review

maar system (Fekete-hegy — MARTIN et al. 2003). This location represents mafic volcaniclastic flow deposits (referred also as hydroclastic flow — NEMETH and MARTIN 1999b), with large amounts of dm-size lherzolite xenoliths, mantle and/or deep crustal nodules (TÖRÖK and DE VIVO 1995, TÖRÖK et al. 2003), and pyroclastic deposits indicative of high-energy phreatomagmatic explosive eruptions (MARTIN et al. 2003). Further individual vent clusters are shown on a vent density map on the basis of a 1-km rectangular grid and 2.5 km search radius (Figure 1.9, B), mimicking major known crustal structural zones orientated mainly NE–SW and NW–SE (TARI 1991, BUDAI et al. 1999, BUDAI and CSILLAG 1999, DUDKO 1999). In general, using larger search radius and larger steps on grid, it is inferred that vent-distribution features are related to deep subsurface features such as the geometry of the melting anomaly (CONNOR 1990). In contrast, smaller search radius and steps on the grid give information on the surface structure of the pre-volcanic system (CONNOR 1990). It has been thus inferred that the Mio-Pliocene volcanism in the BBHV was related to a characteristic melting anomaly from where the magma intruded into shallow subsurface crustal inhomogeneities, such as fault lines. Along with the vent clustering and alignments the north–south elongation of individual vents is likely to be related to valley pattern and inherited structural elements of the basement rocks (Figure 1.10). Analysis of the present morphology of the central part of BBHV has revealed that north-south oriented textural pattern exist in this region, either representing palaeovalleys and/or surface expressions of old structural elements (JORDÁN et al. 2003). In the LHPVF, however, the vents are scattered. It is also noteworthy that major volcanic complexes such as Ság-hegy, Somló, Kab-hegy and Tihany (Plate 1.5) fall on a straight line that has no obvious surface expression in the form of faults or other structural elements (JUGOVICS 1969b, JÁMBOR et al. 1981).

Lavaflows, scoria and spatter cones

There is a clear evidence for the presence of preserved fissure-vent systems in the western Pannonian Basin. In a small area around the northern part of the Keszthely Mts (e.g. Sümegprága — Plates 1.1 and 1.5) NE/EW elongated coherent lava rock outcrops occur. The extent of the lava rocks is strongly related to shallow subsurface intrusions such as sill and dyke systems and adjacent small lava flows, plugs. The linear alignment, of the lava rock outcrops suggests a fissure related origin, however, their surface exposures either have been eroded already, or never existed. The age distribution of the different eruptive centres also shows an alignment which probably is related to older structural zones in the basement, probably performed by the fluvial systems during the volcanism. Elongated structures of individual centres, or eruption complexes, especially in the middle part of BBHV (Hajagos-hegy, Sátorma-hegy, Fekete-hegy) also suggest longitudinal orientation of individual vents (Figure 1.10). The principal source of the lava flows appears to have been elongated, north to south, north-east to south-west trending former lava lakes. The best example for this is the Hajagos-hegy with a north to south trending lava lake, which overflowed southwards (Kő-hegy — Plate 1.6). The original lava lake was probably 700–1000 m long and 800 m wide. The buttes of Badacsony, Szent György-hegy, Csobánc, Haláp (Plate 1.1) show slightly elongated north–south shape. In the middle part of BBHV, there is a large volcanic complex, called Fekete-hegy volcano (Plates 1.1 and 1.5), which consists of smaller eruptive centres with different intercalated lava layers (MARTIN et al. 2002). The lava filled individual centres that also show a NE-SW-trend. The Fekete-hegy is interpreted as a complex lava channel, spatter cone and scoria cone system with several large intercalated lava lakes as well as lava flows fed by scoria cones in the surrounding phreatomagmatic tuff.
Shield volcanoes are common and give the major sources of lavas in intraplate provinces (Walker 1993, 2000, Connor and Conway 2000). Eruptions of large Hawaiian-type volcanic centres are commonly related to fissure-vent systems, but in a small plain-basalt province eruptions are related to central vent systems. However, there are several examples where shield volcanoes developed along basement fissure systems (Johnson 1989). In the BBHVF there are two shield volcanic complexes probably associated with a large number of eruption vents (Korpás 1983, Németh and Martin 1999d). The larger one (Kab-hegy—Plate 1.1) represents the highest topographic point in the Transdanubian Range. Individual lava flows tend to be around 5 to 8 km long and cover around 50 km² in area (Jugovics 1971, Korpás 1983, Korpás and Szalay Márton 1985). The total thickness of the lava cover reaches several tens of metres (Jugovics 1971, Korpás 1983, Korpás and Szalay Márton 1985). The lava field around Kab-hegy is a thick accumulation of various lavas that have been recognised already in the year 1934 by Vitalis. The lava flow units are often separated by thick palaeosoil layers as well as intensive alteration horizons of the basalt itself, indicating time gaps between effusion of lava flows and suggesting a complex eruptive history of this region (Vitalis 1934, Vörös 1962, 1966, 1967). The top of the lava field is inferred to be an eroded scoria cone preserved as a plug. Adjacent to the lava fields of Kab-hegy small Strombolian scoria cone remnants and Hawaiian spatter deposits are common. The other large shield volcanic complex is the Agár-tető south-west of the Kab-hegy with a present elevation of 499 m. There is a small remnant of a scoria cone sitting over the lava plateau of the Agár-tető. The wide range of measured K/Ar age (5.25–2.8 My—Balogh et al. 1986), the different lava flow units, the slightly different petrographic characteristics of the lava flow(s) and the lava inter-beds on the flank of the topping scoria cone suggest a long-lived volcanic activity, which might be related to stable melt sources over structural weakness zone. Such time sequence is well known from several small to medium size shield volcanoes such as the Rangitoto Island (Auckland Volcanic Field, New Zealand—Johnson 1989, Endbrooke 2001) or large scoria cone examples from the Eifel (Germany) region (Houghton and Schmincke 1989). Moreover, Rangitototo (New Zealand) has a well-developed capping scoria cone as well as eroded satellite vents on the flank of the main edifice of the shield. The small spatter deposits on the top zone of the Agár-tető represent former summit craters of small (50–100 m) scoria and associated spatter cones. The preserved deposits represent small vent zones of this former explosion centres. Small remnants of lava cone structures are traceable everywhere where large lava lakes preserved. These areas usually are small (few tens of metres) irregularities in the large lava fields. They are inter-bedded with lava, and consist of spatter deposits (Badacsony, Szent György-hegy, Fekete-hegy, Sátorma-hegy—Figure 1.10).

The magmatic explosive and effusive volcanic activity which produced large volumes of eruptive products as well as the presence of the elevated Mesozoic basement under this volcanic zone suggest that there was no magma-water interaction during the eruptive history, thus this area is inferred to have been already a higher elevated area in the Pliocene. These shield volcanoes in comparison with eastern-Australian examples, are relatively small with less than 1 km³ volume of lava products (7 km³ in eastern-Australia, Johnson 1989).

Large spatter and scoria cones (Figure 1.11) are strongly eroded in the BBHVF. They remained only as erosional remnants. Only the summit craters of the Agár-tető, Bondoró and Füzes-tó (Plate 1.5) are preserved in a recognisable morphology. Strombolian scoria cone remnants are however often preserved only as scoria mounds on top of larger volcanic erosion remnants such as the Boncos-tető on top of the Fekete-hegy (Martin et al. 2002). Highly vesicular scoriaceous deposits often accumulated between lapilli tuffs rich in accidental lithic clasts indicating that eruption styles may have changed especially in the late stage of the eruptions of individual vents according to the available water to sustain magma-water interaction (Houghton and Hackett 1984, Houghton and Schmincke 1986, Houghton et al. 1999). Scoriaceous deposits are often rich in irregular shaped mud chunks which are preserved between scoriaceous lava spatters often in significant thickness (tens of metres: e.g. Ság-hegy—Plate 1.1), indicating an active quarrying of an unstable volcanic conduit and/or presence of

![Figure 1.11. Welded lava spatter-rich deposit from Ság-hegy](image-url)
water-rich slurry in the vent zones during more magmatic fragmentation of magma (KOKELAAR 1986, WHITE 1991b, ORT et al. 1998, MCLINTOCK and WHITE 2000). Scoria cone remnants are usually preserved in western Pannonian Basin as near vent strongly baked, red, slightly bedded sequences with large spindle or highly vesicular fluidal bombs. Welding of scoriaeous lapilli is common (Plate 1.6). Strombolian scoria and spatter deposits are common in relation with maar volcanoes. Such beds often reflect irregularities in the magma/water interaction having clear magmatic and phreatomagmatic fragmentation styles such as it was observed during the eruption of the Ukinrek Maar, Alaska (KENILE et al. 1980, SELF et al. 1980, ORT et al. 2000). Scoria cones often develop in maar basins such as the La Breña Maar in Mexico (Plate 1.7, A), which is a recent analogy how e.g. the Pula maar (Plate 1.1) may have looked like prior to the water filled its basin. Scoria cones are inferred to have grown in maar basins in the BHVF such as inferred for Uzska maar (Plates 1.1 and 1.5). Such scoria cones (e.g. Uzska) may have collapsed into the maar basin, feeding extensive volcanic debris avalanches and associated debris flows, which tend to accumulate in the maar basin. A remnant of a Strombolian scoria cone in the Füzes-tó region preserves near vent scoriaeous volcaniclastic breccia with muddy matrix that is inferred to represent remnants of the water saturated slurry in the vent during the Strombolian activity. The uprise of magma at Füzes-tó is inferred to have involved turbulent jets that did not generate shockwaves as suggested in theoretical considerations (MASTIN 2004).

Pyroclastic cone growth modelling focuses on the role of ballistic (no-drag) ejection that often are referred as Strombolian activity as a result of weak-intensity, strongly intermittent activity observed to be associated with bursting of large gas bubbles extending across much of the vent, and producing ballistic emplacement of clasts larger than 10 cm (e.g. MCGETCHIN et al. 1974, JAUPART and VERGNIOLLE 1997, VERGNIOLLE and MANGAN 2000). Particles with a size less than 10 cm are normally unable to follow ballistic trajectories instead depositing from eruption clouds with characteristic jet dynamics (e.g. subplinian — e.g. SPARKS et al. 1997). The grain size pattern observed from scoria cone remnants of western Hungary suggests other than the classic ballistic (no-drag) model of cone growth (e.g. subplinian) and indicates not only Strombolian dynamics of the eruptions as it suggested on the basis of experimental studies (e.g. RIEDEL et al. 2003).

No pahoehoe or aa lava formations in the western Pannonian volcanic fields have been preserved. The blocky appearance of the preserved lava flow remnants indicate that they were originally predominantly aa type flows. Direct surface remnants of tumuli, hornitos, pressure ridges, lava tubes, caves or channels are not known from the BBHVVF, but several surface irregularities from the Kab-hegy suggest their existence, however, further research needs to constrain this conclusion. On the top of the Fekete-hegy, micro-pahoehoe surfaces on red, rugged lava flows, which are covered by vegetation, are inferred to be the youngest lava flow surfaces in the western Pannonian Basin.

Columnar jointing, that is a product of the progressive cooling of lavas or intrusions (SPRY 1962, DEGRAFF and AYDIN 1987, BUDKEWITSCH and ROBIN 1994, LYLE 2000), is widespread in the coherent lavas from the western Pannonian Region. Usually the simple thin sheet-like lava bodies produce simple, upright joints. Thicker lava bodies have two or multi-tiered layering with lower, well developed upright joints (colonnade) and irregular, semi or wholly radial small joint systems in the top (entablature — LYLE 2000), among which the best exposed is in the Hajagos (Plate 1.7, B). In thick (~10 m) lava flow remnants in the BBHVVF. The joint package reaches several metre thicknesses. Other examples comprise Badacsony and Hegyes-tü. Radiating, rosette-like joints in thick lava flows may represent former individual lava channels or feeder dykes that are especially common in the Sümegprága shallow subsurface sill and dyke complex (Plate 1.7, C). Rosette-like joints related to lava tubes are present in the upper level of the Hajagos-hegy basalt quarry, and the Badacsony basalt quarry (MARTIN and NEMETH 2002b).

Hyaloclastite and peperite

Hyaloclastite forms by quench fragmentation of magma in contact with water. It can form when lava flows erupt into water or flow from land into water, or where magma intrudes wet unconsolidated sediments (RITTMANN 1958, 1962, 1973, McPHEE et al. 1993). Such hyaloclastite deposits were found in several places in the western Pannonian region and are inferred to represent lava flows entering a maar lake such as the Uzsa maar. At Hajagos, there are volcaniclastic breccias from in situ debris flanks that consist of large, strongly chilled, micro-vesicular, black, angular basaltic fragments in micro-crystalline carbonate cemented matrix. The fragments in several places contain fine-grained, brown, silty, muddy fragments in the vesicles. Another locality close to the Hegyes-tü columnar jointed basalt outcrop shows hyaloclastic volcaniclastic sediment that contains large pillowed lava fragments indicating lava and water interaction in a water-rich vent.

Peperite forms when lava intrudes into wet, unconsolidated sediment. Peperite can be described on the basis of juvenile clast morphology being blocky or fluidal (globular — BUSBY-SPERA and WHITE 1987) but other shapes occur and mixtures of different clast shapes are also found (SKILLING et al. 2002). Magma is dominantly fragmented by quenching, phreatomagmatic explosions, magma-sediment density contrasts, and mechanical stress as a consequence of inflation.
or movement of magma or lava (DOYLE 2000, SKILLING et al. 2002, ZIMANOWSKI and BÜTTNER 2002, WOHLETZ 2002). According to the observations of BUSBY-SPERA and WHITE (1987) blocky peperite forms usually during interaction of coarse grained, water saturated sediment and melt. In contrast globular peperite forms when melt intrudes into fine grained sediment (BUSBY-SPERA and WHITE 1987). This tight relationship between host sediment grain size distribution and the peperite type seems not always to be the case (DOYLE 2000, DADD and VAN WAGONER 2002, Hooten and ORT 2002, MARTIN 2002). Variations are clearly demonstrated from the western Pannonian Basin (MARTIN and NÉMETH 2000). Peperites are common in the western part of the BBHVF and on the LHPVF. Blocky peperite identified from the Hajagos-hegy (Plate 1.1) basalt quarry (lower level — MARTIN and NÉMETH 2000) is related to the feeder dykes that invaded fine grained host sediment and lava lake margin that developed in the volcanic depression caused by the phreatomagmatic eruptions (MARTIN and NÉMETH 2000). In near vent position of the lower part of the lava flow at Hajagos there is a lava foot breccia, where small, pillowed lava fragments mix with yellowish sandstone fragments. The peperite formed when a lava flow entered water-saturated sediment, probably in a swampy area. In several localities large (2–3 m wide, 3–4 m high) peperitic bubble structures formed in the lava flow units. Inside the bubbles highly vesicular, closely packed pillowed lava formed in sandy matrix. The lava flows are inferred to have formed as tumuli by the vaporisation of the swamp water, during the flow movement. This kind of tumuli structure is common in the lower level of the Badacsony (Plate 1.1) lava flows and in the Hajagos-hegy southern region. Peperites are also described from Balatonboglár, Temető-domb (Plate 1.1 — NÉMETH et al. 1999b). Large black, red scoria fragments in a fluidised sandy matrix represent magma and water-saturated sediment interaction in near vent position. Peperitic lava lake margins have been described from the Ság-hegy, where a lava lake fed small sills (Plate 1.7, E) that intruded into the wet irregular shape tuff ring (MARTIN and NÉMETH 2004c). A crater lake of a small tuff ring at Kis-Somlyó has been flooded by a basanite lava flow and developed pillow lava, as well as delicate mixture of basanite melt and silt forming peperite (MARTIN and NÉMETH 2004b).

**Phreatic and phreatomagmatic eruptive centres**

Explosive volcanic eruption could be a result of elevated heat of pore water due to dyke (phreatic explosion) or cryptodome emplacement or direct contact between hot magma and various aquifers or standing water body (phreatomagmatic explosion — CAS and WRIGHT 1988).

Phreatic explosions are steam generated and do not involve the ejection of fresh magma (CAS and WRIGHT 1988). Phreatic explosions resulting in steep, deep and often wide craters such as formed by the USU 2000 (Hokkaido) eruption due to the heat of emplaced shallow subsurface magma body (OHBA et al. 2002). Clear phreatic explosion centres and their products are not described yet from the western Pannonian Basin, however, a large amount of the pyroclastic beds associated with volcanic remnants of the region is commonly very rich (90 vol.%.) in accidental lithic rock fragments derived from various pre-volcanic rock units (Plate 1.8, A). This very high percentage of non-volcanic country rock fragments in the accumulated deposits around vents led to the conclusion that volcanism in the western Pannonian region might be synsedimentary with the siliciclastic sedimentation in the Pannonian Lake (JUGÓVICS 1937, KULCSÁR and GUCYZNÉ SOMOGYI 1962, JÁMBOR and SOLTI 1975, JÁMBOR et al. 1981).

**Phreatomagmatic activity**

Phreatomagmatic explosions involve dynamic explosive interaction between magma and external water source such as groundwater, or a surface body of water such as a lake or the sea, and the ejection of a significant juvenile magmatic component (WOHLETZ 1983, FISHER and SCHMİNCKE 1984, WO HLETZ and McQUEEN 1984, WO HLETZ 1986, CAS and WRIGHT 1988, ZIMANOWSKI et al. 1991, WHITE and HOUGHTON 2000). The term “phreatomagmatic eruption” is predominantly used for terrestrial magma-water interaction driven processes. Eruptions initiated in standing water bodies are often referred to as Surteyyan style eruptions (KOKELAAR 1983, KOKELAAR and DURANT 1983, KOKELAAR 1986, WHITE and HOUGHTON 2000). They are characterised by eruption cloud that breach the water surface in advance of the eruption (emergent volcanism). In case the eruption is fully subaqueous with no subsequent water surface breach, the magma water interaction lead to explosive eruption which fed subaqueous pyroclastic density currents mantling the sea/lake floor and move radially, leading to an accumulation of a pyroclastic mound (WHITE 1996a, 2000, 2001, WHITE and HOUGHTON 2000, MARTIN and WHITE 2001). Volcanic edifices resulted from both of these volcanic eruptions (emergent and fully subaqueous) often have a similar basal setting, exhibiting pyroclastic rocks rich in juvenile chilled fragments and only a few accidental lithics or minerals derived from the synsedimentary non-volcanic units (e.g. sea floor sediments — WHITE 1996a, BELOUSOV and BELOUSOVA 2001, MARTIN and WHITE 2001). In case of emergence, the vent temporarily could be blocked from the open water, and more or less subaerial conditions may be reached, resulting in similar eruption styles than in other terrestrial
phreatomagmatic explosions or even lava fountaining as it has been reported from Surtsey (KOKELAAR 1983, HOUGHTON and NAIRN 1991). During emergent volcanism, tuff cones often build up to levels above the water surface, which consists of steeply dipping juvenile clast-rich pyroclastic units (SOHN and CHOUGH 1992, 1993). In case of magma-water interaction in terrestrial setting, the volcanic landform and the erupted products largely depend on the depth of explosion locus and the type of bed rocks (hard rock versus soft rock — LORENZ 1986, 2002, 2003). In case of an unstable volcanic conduit wall, the recycling of erupted pyroclasts as well as the sediment laden slurry in the vent could play an important role in the course of the eruption and determine the type of deposit that may accumulate around the vent (HOUGHTON and SMITH 1993, WHITE 1996b).


1. chilled juvenile lithic clasts,
2. the angular, blocky, moderately vesicular volcanic glass shards and the
3. variable amount of fragments from the disrupted pre-volcanic rock units.

The blocky shape of the volcanic ash and a low vesicularity attest to the sudden chilling, as well as the high confining pressure during the magma-water interaction (NÉMETH and MARTIN 1999a) as it has been concluded from other similar volcanic fields (HEIKEN 1972, 1974, HEIKEN and WOHLETZ 1986, ZIMANOWSKI 1986, 1997, 1995, 1998, DELLINO and LAVOLOPIE 1995, BÜTTNER and ZIMANOWSKI 1998, DELLINO 2000, DELLINO et al. 2001, DELLINO and LIOTINO 2002). Generally the phreatomagmatic products from the western Pannonian Basin are rich in chilled semi-angular juvenile volcanic lithic fragments and fresh to moderately palagonitized volcanic glass shards typical for fragmentation driven by magma/water interaction (e.g. FRÖCHLICH et al. 1993, BÜTTNER et al. 1999, 2002 — Plate 1.3). The sideromelane glass shards are light brown to yellow (Plate 1.3). The glass shards are slightly to strongly palagonitized, having a palagonite rim and/or palagonite bands along microfractures (Plate 1.8). The glass shards commonly contain a few elongated microvesicles, which are filled with secondary minerals, especially if the glass shard itself shows advanced stage of palagonitization (Plate 1.3 and 1.8). The vesicles are slightly stretched (Plate 1.3 and 1.8). The phreatomagmatic rock units are characteristically rich in mud, silt and sand derived from the immediate pre-volcanic Neogene shallow marine to fluvo-lacustrine sedimentary sequences (NÉMETH and CSILLAG 1999, NÉMETH and MARTIN 1999c). Two major types of pyroclastic rock could be defined,

1. juvenile clast and siliciclastic sediment grain (from Neogene units) rich and
2. which are more enriched in accidental lithic clasts from deep seated pre-volcanic rock units (NÉMETH and MARTIN 1999c, NÉMETH et al. 2001).

Pyroclastic rocks from the LHPVF and the western part of the BBHVF are like the first type. In contrast pyroclastic rocks from vent remnants in the eastern BBHV are more like the second type. This relationship is inferred to be related with the explosion locus, and the sub-surface architecture of the different regions, which is likely to determine the palaeo-hydrogeology of the regions (MARTIN and NÉMETH 2003). The depth of the phreatomagmatic fragmentation at the volcanoes of the western Pannonian Basin varies greatly according to the accidental lithic clast population of the phreatomagmatic pyroclastic rocks (NÉMETH and MARTIN 1999c). In the central part of the BBHVF a few locations preserve pyroclastic rocks, which are rich in accidental lithic fragments from every known lithology from the upper crust, and therefore the fragmentation level could have been in a range of kilometers from the syn-volcanic palaeosurface. Because aquifer pore pressure increases with depth, many researchers have assumed that once pressure exceeds water’s critical pressure, magma/water interaction ceases to be explosive because steam is not formed (e.g. CAS and WRIGHT 1988). This assumption is suggested to be incorrect according to the latest results of theoretical experimental work, which show the potential for dynamic interaction at confining pressures up to and perhaps exceeding 100 MPa (WOHLETZ 2004). These results suggest that interaction can initiate at maximum depth >4 km, reaching depth perhaps exceeding 10 km as suggested by WOHLETZ (2004). Indeed, there is evidence of fluid (H₂O) inclusion study on volcanic glass from pyroclastic rocks associated with phreatomagmatic successions of the BBHV (e.g. Sziliget), that their entrainment may have occurred at around a few km (~10 km) depth. These fluids could have been responsible for the magma/water interaction (e.g. BALU et al. 2002, TÖRÖK et al. 2003).

The eruptions inferred to produce deeply excavated maar/diatreme structures occur in areas with commonly karst water bearing, fracture controlled aquifers and which are covered by relatively thin Neogene soft rocks (NÉMETH et al. 2001, MARTIN and NÉMETH 2003). This scenario is equivalent to LORENZ (2002) “hard rock” model, where the conduit wall is able to be stable during the eruption. With the substantial water supply from the fracture controlled aquifer (e.g. karst water) the eruption could continue for a longer time, however, the eruptive products may only accumulate near-
by the vent (NÉMETH et al. 2001). These deep maar basins later functioned as sedimentary traps and give way to accumulation of thick maar lake deposits (e.g. alginite) and intercalated debris flow and turbidity current deposits (NÉMETH 2001) similarly to other deep maars (WHITE 1992, DROHMANN and NEGENDANK 1993, FISHER et al. 2000, VASS et al. 2000, BULLWINKEL and RIEGEL 2001, PIRRUNG et al. 2003).

During phreatomagmatic eruptions not all of the volume of the magma can be simultaneously in interaction with water, therefore parts of the eruption column can be phreatomagmatic and other parts can purely be magmatic as it has been observed and documented from the Ukinrek maar eruption (SELF et al. 1980, BÜCHEL and LORENZ 1993, ORT et al. 2000). Similar involvement of different types of fragmentation history in the same vent zone is well known from Surteyean eruptions and the study of the deposits of this type of eruptions revealed a complex intercalation between deposits derived from strikingly different fragmentation histories (THORARINSSON et al. 1964, THORARINSSON 1965, 1967, LORENZ 1974, KOKELAARR 1983, WHITE and HOUGHTON 2000, COLE et al. 2001). Moreover, in case of nearby active vents (maybe in the same crater), intercalation of deposits from magmatic and phreatomagmatic fragmentation history are reported to be very common, such as reported from the White Island (New Zealand — HOUGHTON and NAIRN 1991), Eifel (Germany — HOUGHTON and SCHMINCKE 1986), Crater Hill (New Zealand — HOUGHTON et al. 1996) or Ohakune Crater (New Zealand — HOUGHTON and HACKETT 1984). Pyroclastic deposits with Strombolian magmatic and phreatomagmatic fragmentation histories in intercalated settings have been documented from the Tihany Peninsula and are inferred to be the result of the instability of magma/water ratio during the late stage of the eruption of the Tihany Maar Volcanic Complex (NÉMETH et al. 2001).

Maars and tuff rings

The common high amount of accidental lithics in the phreatomagmatic pyroclastic rocks from the western Pannonian Basin, the systematic gravity and geomagnetic anomaly associated with the location of such pyroclastic rock units suggest the presence of low density clastic material filled “holes” in the pre-volcanic rocks such as excavated maar/diatreme structures. In general, maar volcanoes (sensu lato) are low volcanic cones with bowl shaped craters that are wide relative to rim height (FISHER and SCHMINCKE 1984, LORENZ 1986). They range from craters cut into country rock below ground level (maar sensu stricto), to craters with low rims composed of phreatic, phreatomagmatic, and magmatic debris (tuff ring — FISHER and SCHMINCKE 1984). During the eruptive process the explosion locus migrates down, (LORENZ 1973, FISHER and SCHMINCKE 1984, LORENZ 1986) excavating deeper and deeper seated accidental lithics in time and resulting in an enrichment of such deeper fragments upwards in the stratigraphic tephra column (LORENZ 1986). This model has been tested in several locations worldwide, however, there are field evidences that in special cases magmatic fragmentation history of the magma may predate the phreatomagmatic fragmentation such as it has been documented from the Pinacate in Mexico (GUTMANN 2002). Systematic sampling and testing to determine the explosion locus migration has not been performed yet from the Neogene volcanics of the western Pannonian Basin and it is a subject of current research. However, in most of the studied places from the western Pannonian region, an abundance of shallow seated accidental lithics in the initial phreatomagmatic pyroclastic units, are results of vent opening eruptions, which has been documented from many places (NÉMETH and CSILLAG 1999, NÉMETH et al. 1999a, 2001, MARTIN et al. 2002, 2003). In spite of this initial enrichment of the pyroclastic successions in clast derived from the uppermost immediate pre-volcanic units, the majority of the volcanic successions do not show systematic clast population trends.

In the pyroclastic successions of the Neogene western Pannonian Basin base surge and fall out beds form the basal units in most of the places. A general trend shows that most of the surge beds of the eruptive centres are wet surge beds, however, subsequent accumulation of dry surges are characteristic. The textural characteristics that are more prominent in wet surge deposits are more common in locations in the eastern part of the BBHVF such as the Tihany Peninsula and the Kő Basin (north — NÉMETH et al. 2001). In the Tihany Peninsula (Plate 1.1), the pyroclastic succession has a thick (tens of metres) accumulation of wet surge beds that are rich in large (dm-scale), deeply excavated country rocks (NÉMETH et al. 2001). The formation of the phreatomagmatic volcanoes at Tihany is inferred to be related with magma and karst water interaction from fracture controlled aquifers combined with water from the porous media aquifer on the near surface region (NÉMETH et al. 2001). The products of this kind of eruptions, which had two very different type of water source to fuel the magma-water interaction, are highly “indurated” beds of pyroclastic breccias interbedded with tuffs and lapilli tuffs originated from wet base surges and phreatomagmatic fall out, combined with ballistic bomb and block shower during major conduit collapsing phases (NÉMETH et al. 2001). The deposits of this type of eruptions are thick and carry evidences of high water/magma ratio during the eruption and accumulation of deposits from high concentration and water charged pyroclastic mass flows (NÉMETH et al. 2001). This type of phreatomagmatic eruption termed as Tihany-type maar volcanic eruption, highlights the speciality of such events due to the interaction between two different water sources during the eruption (NÉMETH et al. 2001). Similar type of eruptions likely occurred in the centre part of the BBHVF, where fracture
filling water bearing units are covered by Neogene soft rocks that have likely been water saturated during the volcanism (MARTIN et al. 2002).

Dry surge beds in pyroclastic successions and evidence of the major role in controlling the magma-water interaction by the porous media aquifer are more common in the western part of the BBHVF and in few sites from the LHPVF. In these sites the pyroclastic beds show less indurated characteristics, baking of silt and mud as well as scoria are more prominent, and the relative amount of deep seated Mesozoic and Palaeozoic fragments is significantly less (Plate 2.8, D). There are strongly eroded remnants of this kind of centres from Szígliget, Balatonboglár, probably Szent György-hegy and Csobánc (Plate 1.5).

**Types of phreatomagmatic volcanoes**

The wide range of types of magma-water interaction led to the formation of different types of volcanic edifices, including maars, tuff rings and scoria cones with a great variety of depositional features as preserved in their pyroclastic units in the western Pannonian Basin (MARTIN et al. 2003). Different types of volcanoes were identified on the basis of volcanic textures, shape and composition of their pyroclasts, sedimentary structures, and the presence of pillow basalts and peperites. Base surge and fallout tephra were deposited around maars and tuff rings by phreatomagmatic explosions, caused by interactions between water-saturated sediments and alkali basalt magma locally carrying peridotite and pyroxenite xenoliths as well as pyroxene megacrysts, which are well exposed in nested maar complexes (Figure 1.12) such as Tihany (NÉMETH et al. 2001) and Fekete-hegy (MARTIN et al. 2002). Fekete-hegy is volumetrically one of the largest in the BBHVF (MARTIN et al. 2002), it is a representative example demonstrating the architecture of complex multivent systems with chains and/or groups of predominantly phreatomagmatic vents. It forms a lava-capped butte in the central part of the BBHVF with basaltic lava flows overlying ~50 m of pyroclastic unit (MARTIN et al. 2002).

**Complex multiple volcanoes** (Figure 1.12 – MARTIN et al. 2003) with solidified large volume lava lakes are characteristic volcanic remnants especially in the western part of the BBHVF, such as Badacsony (Plate 1.5), one of the largest lava-capped buttes in the BBHVF (Figure 1.13). Thick (>50 m), black, strongly chilled, aphanitic basanitic lava overlies a coarse-grained, unsorted yellow lapilli tuff (MARTIN and NÉMETH 2002b). The lapilli tuff consists of finely dispersed accidental lithic fragments of quartz or quartzofeldspathic sandstone, and blocky, weakly to highly vesicular microlite-poor sideromelane (tephrite, phonotephrite), indicative of phreatomagmatic origin, near-surface vesiculation and possible excavation of pre-volcanic country rocks (MARTIN and NÉMETH 2002b). In addition, xenocrysts of olivine and pyroxene may reach 5 vol.%. The lava lake at Badacsony exhibits irregular lower contacts with the pyroclastic units, often displaying peperite structures (MARTIN and NÉMETH 2002b). The peperite encloses highly vesicular scoriaceous lava spatter clasts, with vesicles filled by clay, calcite or quartzofeldspathic assemblage suggesting rejuvenation or longevity of volcanic vents at the same site (MARTIN and NÉMETH 2002b).
Single phreatomagmatic volcanoes (Figure 1.12 — MARTIN et al. 2003) are also widespread, such as Kis-Somlyó, (MARTIN and NÉMETH 2002a, 2004b) Kereki-domb (Figure 1.14) or Vár-hegy of Zánka (NÉMETH et al. 2003a). They consist of moderately to strongly eroded pyroclastic mounds, forming small hills and exposing phreatomagmatic lapilli tuff beds, often in chaotic setting, which are commonly covered by lava flows. The presence of chilled, angular, moderately vesicular sideromelane fragments and the predominantly Neogene sediment-derived accidental lithic clast population in the exposed pyroclastic rocks allow to infer that these volcanic remnants represent near vent to vent-fill pyroclastic units of former phreatomagmatic volcanoes, such as maars surrounded by tuff rings, which were topped by lava flows in the last stage of volcanic activity.

Maar crater-fill sediments, Gilbert-type deltas

In the BBHVF there are eroded remnants of maar crater fill volcaniclastic deposits (NÉMETH 2001). Commonly they represent the higher stratigraphic position in the phreatomagmatic eruptive centres. They are steeply bedded volcaniclastic units forming a radial dip pattern around a former crater (NÉMETH 2001). Their volcaniclastic successions are commonly rich in scoriaceous lapilli and broken pyrogenic minerals and/or xenocrysts. These high-level deposits that have been identified from the Tihany Peninsula are steeply dipping (about 30-35°) beds of reworked tuff and lapilli tuff on the former maar crater edge and interpreted to be erosional remnants of Gilbert-type of deltaic rim deposits (NÉMETH 2001). The dip of the beds mostly shows the original slope of the former maar craters, where the unconsolidated tephra were mobilised and moved down into the maar crater basin. Similar delta sequences have been identified from well exposed settings from the Hopi Buttes, (WHITE 1989, 1990, 1991b, 1992) where such deposits occur in dry maar settings after disappearance of water and moderate erosion. Similarly, steeply inward dipping pyroclastic units on the crater floor of the Crater Elegante maar (Mexico) forming ramp like structures have been interpreted to be result of delta building into the former maar lake, which is exposed today, due to the disappearance of the lake (GUTMANN 1976). The bedding structures of these sequences at Tihany are inferred to be the result of grain flow and turbulent sediment gravity flows on the steep inner slopes of the maar craters (NÉMETH 2001). Two major lithofacies association have been identified:

1. A coarse-grained, 0.5 to 1 m thick, juvenile, scoriaceous lapilli-rich lithofacies association with inverse to normal graded beds. The lapilli are rounded, black, tachylitic glass with small amount of accidental lithics. There is commonly a relatively high amount (5 vol.%) of large broken crystals of pyroxene, olivine, or amphibole with lapillus grain size. The grains in several places are algae-coated, spatic calcite cemented. Between the large grains there are micritic calcite, and occasionally fine grained muddy sediment (altered glass and/or sediment).

2. A fine grained, cross bedded, channelized tuff lithofacies association interbedded with the coarse grained units. This lithofacies association varies in thickness and in several places represents just a few cm thick unit or is even missing.

At Tihany, the large proportion of scoriaceous lapilli in these units indicates that the source region of the delta must have consisted of eroded scoria cones, or scoria-rich pyroclastic units topping the basal phreatomagmatic successions (NÉMETH 2001). It is therefore inferred that the deltas were fed either by small streams (either run off or creeks from the elevated hills of former Bakony Mts) or by passive collapse of such scoria lapilli rich tephra blocks onto the maar lakes of Tihany (NÉMETH 2001). In other part of the western Pannonian Basin, pyroclastic rocks with textures characteristic for reworking indicate that destructive events into maar crater lakes may have been important events, and the resulting pyroclastic rock units could be identified.

Maar lake carbonates

In several areas widespread carbonate sedimentation is inferred to have followed the maar volcanism and to have formed sedimentary basins and trapped reworked tephra from the former crater rims (NÉMETH and MARTIN 1999c). At the Tihany Peninsula at least 15 m thick fresh water carbonate unit is preserved capping the pyroclastic and crater lake successions. Fresh water, carbonate-rich interbeds are also known from Pula and from in situ debris from Balatonboglár and Fekete-hegy (Plate1.1). At the Tihany Peninsula the carbonate succession contains a high amount of soft sediment deformation structures that are located nearly vertical pipe-like structures inferred to be hot spring pipes. Therefore, these deformations are inferred to be related to hot spring activity and/or earthquakes caused by nearby explosive volcanism. From the thickness of 0.2–0.7 mm of single lamina built sequence with 15 m of total thickness at Tihany, a quite lacustrine sedimentation of 50 000 years has been calculated. This long term undisturbed period in the lake life suggest that the general relief of the area was smooth shortly after the erosion of scoria cones and phreatomagmatic rim deposits.
Today, the carbonate sequences, especially in Tihany case, are in the highest elevated areas. They must represent the former lake bottom thus they are useful for calculating local erosion rates.

Recent studies of seismic profiles through the Lake Balaton, east of the Tihany Peninsula (Figure 1.15) revealed hard surfaces, which may be correlates with the fresh water carbonate units topping the pyroclastic succession of Tihany (SACCHI et al. 1999, SACCHI and HORVÁTH 2002). These units have been interpreted as silicified travertine mounds developed at warm/hot springs and correlated with the PAN2 regional unconformity surface representing a maximum flooding event (9 My) of the Lake Pannon as a result of significant base level lowering as well as the volcanism in the region (SACCHI and HORVÁTH 2002). From a volcanological point of view, these travertine deposits clearly overly unconformably the pyroclastic successions of Tihany (NÉMETH et al. 2001) that have been dated to be about 8 My in age (BALOGH and NÉMETH 2004). These travertines at Tihany are seemingly correlated with the mound structures identified on the seismic profiles (SACCHI et al. 1999). There is a disagreement between timing of events that suggests that the origin of these rocks and their relationship with the volcanism and the general stratigraphy is far from resolved. Today these travertine bed remnants can be traced in a uniform elevation and cover the immediate Neogene siliciclastic units in the southern part of the peninsula, indicating that if they have been associated with the maar basins of Tihany, those basins must have been open towards the south (NÉMETH et al. 1999a) and probably have been part of a larger standing water body system (Figure 1.16). Such cases are described from maar fields near sea level, or large lakes (HAYWARD et al. 2002).

General features of the western Pannonian Basin

The Neogene alkaline basaltic volcanic erosional remnants of the western Pannonian Basin are exposed from former subsurface to surface levels but they are commonly covered by the Quaternary erosional talus flanks. The outcrop availability strongly controls the identification of the facies relationships. The deepest levels of exposures are located in the western and the southern part of the area. The most strongly eroded
regions are those where no subsequent lava caps sheltered the pyroclastic sequences. The Balatonboglár, Szigliget, Tihany Peninsula allow the study of the deepest level of the volcanic centres, exposing diatreme facies (Plates 2.1 and 2.5). Probably the eruptive centres of Balatonboglár (Bögöl Volcano), Kerek-domb, Vár-hegy of Zánka, Hármashegy and Vendeke-hegy (Plates 2.1 and 2.5) represent the deepest exposed level of the phreatomagmatic eruptive centres (Németh and Martin 1999c, Németh et al. 2003a) and are inferred to be exposed zones of lower diatremes (terminology after White 1991b). Apart from this deep level of exposures, there are no exposed irregular shapes, fragmented wall rock rich dykes, like those are widely reported from other monogenetic maar-diatreme volcanic fields as e.g. Hopi Buttes or southern Africa (White 1991a, b, Kurszlaukis and Lorenz 1997, Lorenz and Kurszlaukis 1997, Lorenz 2000). Such levels of exposures and preserved outcrops are more common in the northern part of the Pannonian Basin, in southern Slovakia and northern Hungary (Konécný et al. 1999a, Konécný and Lexa 1984, Lorenz 1986, Cas and Wright 1988). The maar vents from the western Pannonian region are served in areas of low erosion, and include volcanoes produced by both phreatomagmatic and magmatic eruptions. The vents characterized by magmatic explosivity are concentrated in the northern part of the area of the BBHVF (Káb-hegy, Agár-tető, Hálap, Hegyesd). The preserved maar diameters of the western Pannonian region ranges from few hundreds of metres up to 5 km in diameter (Fekete-hegy — 5 km; Tihany — 4 km; Bondoró — 2.5 km; Badacsony — 2.5 km), however, the largest centres probably represent maar volcanic complexes with connected large basins (Martin et al. 2002). The average maar basins are inferred to have been 1-1.5 km wide originally, which is within the range of most maars worldwide (Lorenz et al. 1970, Heiken 1971, Lorenz 1973, Schmincke et al. 1983, Fisher and Schmincke 1984, Lorenz 1986, Cas and Wright 1988). The maar vents from the western Pannonian region are reconstructed as hybrids of phreatomagmatic and magmatic volcanic edifices, 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 similarly to several examples from Eifel (Houghton and Schmincke 1986, 1989) or from Mexico (Aranda-Gómez et al. 1992). The typical types of such volcanoes are located in the southwestern site of the BBHVF in the Tapolca Basin (Badacsony, Szent György-hegy, Hajagos-hegy, Fekete-hegy — Plate 1.1).

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. Commonly the large (up to 75 cm in diameter) fragments are silurian schist or permian red sandstone. 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 silicilastic 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 — Plate 1.1) 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. Characteristics surge features such as sandwaves, dunes, impact sags, U-shaped valleys are common (Tihany Peninsula, Fekete-hegy, Bondoró, Szigliget — Figure 1.2). Fall deposits associated with surge beds are also common (Fekete-hegy).

At the onset of the eruption, magma began to interact with a moderate amount of groundwater in the water-saturated Neogene fluvo-lacustrine sand beds. As the eruptions continued, the craters grew and the phreatomagmatic blasts fractured the deeper (harder, consolidated) rock facies around the downward migrating explosion locus, giving the karst (or any fracture controlled aquifer-stored) water free access to the explosion chamber. The appearance of maar volcanoes and their deposits of western Hungary are strongly dependent on the palaeo-hydrological conditions of the fracture-controlled aquifer, which vary seasonally due to the wide range of water supply from rainfall or spring runoff. Maar volcanoes formed due to phreatomagmatic explosions of mixing magma with water saturated clastic sediments in areas where thick Neogene silicilastic units build up the immediate pre-volcanic strata. Such volcanoes have often formed late magmatic infill in their maar basins. These vents, named summer vents, represent low water input from the
lower karst level. Unusual maars (Tihany-type maar) had a special combination of water source from both the porous media aquifer and fracture-controlled aquifer, with the latter probably have been the dominant supplier. Such maars developed in areas, where relatively thin Neogene fluvial-lacustrine units rested on the Mesozoic or Palaeozoic fracture-controlled, e.g. karst water-bearing aquifer. These maars most likely were generated during springtime, thus the vents are named spring vents. In the northern part of the volcanic field former scoria cones and shield volcanoes give evidence for a smaller impact of the ground and surface water in control of the volcanic eruptions. Exposed diatreme-filling rocks with sedimentary grains as well as mineral phases that derived from already eroded Neogene shallow marine to fluvial-lacustrine sedimentary units are the evidence that such a sedimentary cover was intact in syn-volcanic time. The general abundance of such clasts in the pyroclastic rocks also indicates the importance of soft rock environment to where phreatomagmatic volcanoes were erupted forming “champagne-glass” shaped maar/diatremes. The presence of intravent 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 palaeoground water table.

The general features of the volcanic fields of the western Pannonian Basin are very similar to other eroded volcanic fields which erupted into wet environments such as Fort Rock Christmas Valley, Oregon (Heiken 1971), Snake River Plain, Idaho (GODCHAUX et al. 1992), Hopi Buttes, Arizona (White 1989, 1990, 1991b), Saar-Nahe, Germany (Lorenz 1971).

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References


Phreatomagmatic volcanic fields in a Miocene fluvo-lacustrine basin, western Pannonian Basin, Hungary: a review


DELLINO, P. and LIOTINO, G. 2002: The fractal and multifractal dimension of volcanic ash particles contour: a test study on the utili-


DELLINO, P. and LIOTINO, G. 2002: The fractal and multifractal dimension of volcanic ash particles contour: a test study on the utility-


DELLINO, P. and LIOTINO, G. 2002: The fractal and multifractal dimension of volcanic ash particles contour: a test study on the utility-


DELLINO, P. and LIOTINO, G. 2002: The fractal and multifractal dimension of volcanic ash particles contour: a test study on the utility-


RITTMANN, A. 1958: Il meccanismo di formazione delle lave a pillows e dei cosiddetti tufi palagonitici. —

SHAW, C. S. J. and EYZAGUIRRE, J. 2000: Origin of megacrysts in the mafic alkaline lavas of the West Eifel volcanic field, Germany.


Shaded relief model of the Little Hungarian Plain Volcanic Field (excluding the Styrian Basin Volcanic Field) (A).

Simplified geological map of the Bakony – Balaton Highland Volcanic Field (B) with the localities of volcanic erosion remnants.

Typical volcanic landforms associated with phreatomagmatic volcanic fields

A. Crater Elegante (Mexico), maar

B. Diatremes of Hopi Butte, Arizona

C. Cerro Colorado (Mexico), tuff ring

D. Cinder (scoria) cones of the Durango Volcanic Field (Mexico)
Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, Western Pannonian Basin, Hungary: a review

Volcanic glass shards on photomicrographs from the western Pannonian Basin

Boglár diatreme; oriented microlite-rich glass shard. The shards often contain amphibole crystals (a). The shorter side of the photo is 1 mm [parallel plane light]

Pula; moderately vesicular, blocky volcanic glass shard (g) from a glass shard rich layer in the maar lake lacustrine succession. The short side of the photo is 1 mm [parallel plane light]

Egyházaskesző; blocky, moderately vesicular sideromelane glass shard (s) with palagonite rim (dark rim) from a lapilli tuff. Note the non-oriented, microphenocryst bearing texture of the glass shard. The short side of the photo is 1 mm [parallel plane light]

Uzsá; vesicle-free, non-oriented microphenocryst bearing glass shard. The short side of the view is 0.5 mm [parallel plane light]

Hármashegy; moderately vesicular, microlite-poor tephritic glass shard (g) from a diatreme. The short side of the photo is 0.5 mm [parallel plane light]

Kékkút diatreme; glass shard with trachytic texture (T). Note the smaller, lighter coloured sideromelane glass shards (s). The short side of the photo is 1 mm [parallel plane light]
Lherzolite xenolith (s) from a spindle bomb ~30 cm across (A) from the Füzes-tó. A great variety of mantle nodules can be found in dense (B) as well as highly vesicular spindle bombs (C — bomb is 40 cm across) from this region.
Oblique 3D views of the Bakony – Balaton Highland and the Little Hungarian Plain Volcanic Fields
North–south elongated hill of Hajagos, looking from the Csobánc (from west). The lines mark the limit of volcanic rocks mapped.

Lava flow — l, Pyroclasic rock units — px, Neogene siliciclastic units — Ns, Quarry — Q.
Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, Western Pannonian Basin, Hungary: a review

Scoria cone and adjacent lava flow field in the La Breña Maar, Mexico. This maar is in good analogy with maar remnants of the BBHVF such as the Pula Maar prior to flooding by water.

Columnar jointed basanitic lava exposing variable oriented jointing pattern (Haláp)

Rosette-like joints in the Sümegprága shallow sill and dyke complex

Blocky peperite from Hajagos developed due to interaction of basanite melt and fine grained, wet, siliciclastic sediment

Peperitic sill that intruded into the tuff ring sequence of the Ság-hegy
Accidental lithic rich tuff from Sáp-hegy. The dominant proportion of the clasts are sand or quartz derived from Neogene clastic succession, all suggested earlier that the deposit is rather a siliciclastic deposit, and therefore volcanism and siliciclastic deposition from the Pannonian Lake is coeval. After careful studies it is concluded that most of these samples are base surge deposits from energetic explosions in the unconsolidated sand beds.

Photomicrograph of blocky volcanic glass shard (S) from Fekete-hegy, BBHVF (half cross polarized light). The shorter side of the picture is about 2 mm.

Photomicrograph of blocky moderately vesicular glass shard (S) from Tihany, BBHVF (half cross polarized light). The shorter side of the picture is about 2 mm.

Dry surge bed from Szigliget, exhibiting a larger amount of glassy shards as well as thermally affected clasts and minerals from the immediate pre-volcanic siliciclastic rock units.