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Mio/Pliocene phreatomagmatic volcanism in the Bakony – Balaton Highland Volcanic Field, Hungary

Contents

Abstract

The Bakony – Balaton Highland Volcanic Field is a Mio/Pliocene alkaline basaltic volcanic field located in the western Pannonian Basin, just north of Lake Balaton, Hungary. Erosional remnants of volcanoes such as tuff rings and maars, often topped by scoria cones and/or extensive lava flows, are witness to phreatomagmatic activity in the past. In the eastern part of the volcanic field a somewhat older group of volcanoes exposes deep diatreme zones, in contrast to a series of younger volcanic edifices in the west where spatter-dominated scoria cones are still preserved. The eruption mechanism of the individual volcanoes was strongly controlled by the hydrogeologic character of the subsurface during volcanism. Wide, broad tuff rings developed over porous media aquifers and deep maar basins formed over fracture controlled aquifers. Volcanic vents are often aligned, and are located over major structural elements. Deeper levels of the individual volcanoes are exposed in the western part of the area relative to the central section of the volcanic field. In the central sector, volcanic edifices resemble skeleton structures of lava-spatter dominated scoria cones that often developed over lava flow fields (interpreted as lava lakes) that filled the craters of tuff rings and maars. The presence of sideromelane glass shards, of variable but mainly low vesicularity, as juvenile fragments plus a wide range of country rock fragments and/or mineral phases derived from such rocks, often filling craters cut into country rock, attest to the phreatomagmatic origin of the volcanoes in the Bakony – Balaton Highland Volcanic Field. This indicates that this volcanic field developed in an area which was near to the palaeo-ground water table and/or evolved in a fluvial basin with good surface water availability.

Keywords: Pannonian Basin, phreatomagmatic, scoria cone, maar, tuff ring, sideromelane, Gilbert-type delta, pyroclastic, scoria, base surge, explosive, intraplate, monogenetic, basalt, basanite

Tihany Maar Volcanic Complex

Introduction

In this section, a field description and their interpretation will be given for the oldest volcanic erosional remnants of the Bakony – Balaton Highland Volcanic Field (BBHVF) located on the Tihany Peninsula (Plate 3.1). Descriptions and interpretations are based on several published papers that deal with the details of stratigraphy, eruption mechanisms, and palaeogeography of the region (NÉMETH et al. 1999a, 2001, NÉMETH 2001).

The remnant of an unusual Late Miocene (7.92±0.22 My — BALOGH and NÉMETH 2004) maar volcanic complex (Tihany Maar Volcanic Complex — TMVC — Plate 3.1), consisting of several intra-plate eruptive centres, is preserved in the Pannonian Basin and belongs to the Bakony – Balaton Highland Volcanic Field (NÉMETH et al. 1999a, b, 2001). The TMVC formed about 8 My ago and represents the earliest manifestation of the alkali basalt volcanism in the BBHVF (BALOGH and NÉMETH 2004). TMVC volcanic glass compositions range between tephrite, phono-tephrite and trachy basalt (NÉMETH and MARTIN 1999a, b, c — Table 3.1). The TMVC consists of pyroclastic deposits formed by phreatomagmatic eruptions with tephra frequently re-deposited into maar lakes.

Initial base surge and fallout deposits were formed by phreatomagmatic explosions, caused by interaction between water-saturated sediments (Neogene sand) and rapidly ascending alkali basalt magma carrying lherzolite xenoliths as well as pyroxene and olivine megacrysts (NÉMETH et al. 1999a, 2001). Subsequently, the deep excavated maar functioned as a local sediment trap where inflows of reworked or remobilised scoriaceous tephra built up Gilbert-type delta sequences (NÉMETH 2001). The nature of local aquifers coupled with the syn-eruption development of the vent controlled the style of activity and type of deposits formed by the eruptions (NÉMETH et al.1999a, 2001). At Tihany, high poros-

Table 3.1. Composition of volcanic glass shards from the phreatomagmatic units at Tihany

Sample name	PH2	PH2	PH2	ML1-a	ML1-a	ML1-b	ML1-b	ML1-b
SiO_2	43.27	44.06	45.84	46.87	47.08	52.67	47.04	46.83
TiO_2	2.84	2.67	2.73	2.63	2.76	2.39	3.00	2.87
Al_2O_3	16.14	17.44	17.10	17.25	17.13	15.73	17.88	17.62
Fe_2O_3	2.26	2.30	2.19	2.20	2.22	2.03	2.26	2.29
FeO	7.53	7.65	7.28	7.34	7.4	6.76	7.54	7.62
MnO	0.13	0.20	0.16	0.28	0.20	0.15	0.15	0.18
MgO	4.09	3.24	4.52	3.66	4.24	3.41	3.97	3.89
CaO	9.71	8.76	10.66	9.98	10.02	9.81	10.31	10.11
Na_2O	4.49	4.27	4.33	4.82	4.47	4.24	2.53	2.55
K_2O	2.92	3.29	2.58	2.84	2.79	2.69	2.90	2.75
total	93.40	93.88	97.39	97.87	98.30	99.88	97.58	96.70
Rock type	tephrite, phono-tephrite	tephrite, phono-tephrite	tephrite	tephrite, phono-tephrite	tephrite, phono-tephrite	trachy andesite	trachy basalt	trachy basalt

Data derived from electron microprobe analyses on polished thin sections by JEOL 8600 Superprobe housed at the Geology Department of the University of Otago, Dunedin, New Zealand. 15 kV acceleration voltage, ZAF correction method and 5 to 50 µm electron beam diameter was used during measurements.

ity Neogene sand beds with low secondary permeability overlie karst and fractured lithologies characterised by low porosity and high secondary permeability created by solution-enhanced and tectonically generated fractures (BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999). Thus groundwater is stored in either porous media or fracture controlled aquifers depending on the lithology of the aquifer. At the onset of eruptions, magma began to interact with a moderate amount of groundwater in the water-saturated Neogene sand beds (a "soft rock" environment). The deposits of these early explosions are a series of base surge and phreatomagmatic fall out deposits. As the eruptions continued, the crater deepened by down-migration of the explosion locus and the phreatomagmatic blasts excavated the deeper-seated (harder, consolidated – "hard rock" environment) rock type allowing ingress of abundant water, trapped in karst fractures, into the explosion chamber (NÉMETH et al. 2001). At the surface, this "wet" eruption style led to the emplacement of massive tuff breccias around the vents by a combination of fall, surge, mudflows and other gravity flow deposition.

The nature of the TMVC maar eruptions and their deposits appear to be strongly dependent on the hydrologic condition of the fracture-controlled aquifer, which varies seasonally because of its dependence upon rainfall and spring runoff (NÉMETH et al. 2001). Phreatomagmatic explosions caused by the interaction of magma and water-saturated sediments formed the West and East Maar volcanoes of TMVC (Plate 3.1). The West Maar vents (termed "summer vents") represent low water input from fracture controlled aquifer karst lithologies (NÉMETH et al. 2001). The phreatomagmatic explosions that formed the unusual East Maar had a special combination of water sourced from both the porous media aquifer and fracture-controlled aquifer, with the latter being the dominant supply (NÉMETH et al. 2001). This situation most likely operated during spring, when the local water table is at its' highest levels; thus the vents are termed "spring vents" (NÉMETH et al. 2001).

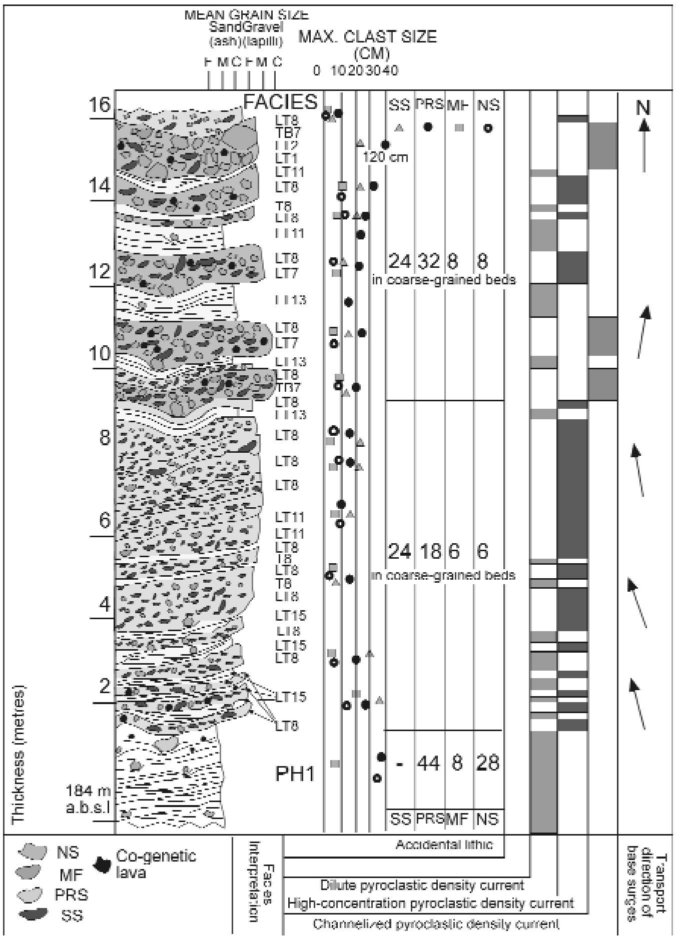
Geological setting at Tihany

At Tihany, as for other parts of the BBHVF, volcanic rocks are underlain by Palaeozoic to Neogene sedimentary sequences (BUDAI and CSILLAG 1998, 1999, BUDAI, et al. 1999). The Silurian Lovas Schist Formation (SS) is more than 1000 m thick and contains very low-grade metamorphosed interbedded psammite and pelite and is exposed ~15 km NE and ~20 km SW of Tihany (CSÁSZÁR and LELKESNÉ-FELVÁRI 1999). Dip and strike of the SS indicates that at Tihany it should be ~800 m below surface (LÁNG et al. 1970, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999). The SS Formation is overlain by the Permian Balatonfelvidék Red Sandstone (PRS) which is a 400–600 m thick alluvial formation (MAJOROS 1983, 1999). At Tihany, the top of the PRS is at least 300 m below surface (LÁNG et al. 1970), and the nearest outcrop is about 2-3 km N of Tihany (BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, 2002). The PRS features a well-developed tectonic fracture system (BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, DUDKO 1999). Mesozoic formations (MF) are represented by a range of Alpine-style Triassic limestones and dolomites, which vary in thickness from a few tens to hundred metres (BUDAI and VÖRÖS 1992, BUDAI and HAAS 1997, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, 2002, HAAS and BUDAI 1999, HAAS et al. 1999). The nearest outcrop is about 3-4 km N of Tihany (BUDAI et al. 1999). The Neogene sediments (NS) mainly consist of lacustrine to fluvial conglomerate, sandstone and mudstone formed in the brackish Pannonian Lake or in fluvial systems related to this lake, as well as older Miocene limestone formations (JÁMBOR 1980, 1989, JUHÁSZ et al. 1997, MÜLLER 1998, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, JUHÁSZ et al. 1999, GULYÁS 2001). The siliciclastic sediments crop out along the eastern and southern parts of the Tihany peninsula (MÜLLER and SZÓNOKY 1988, 1989), recently proposed as a type locality for the Transdanubian stage (ca. 9–7.4 My – SACCHI et al. 1999, SACCHI and HORVÁTH 2002). Beneath Tihany the NS siliciclastic beds are around 200 m thick in the eastern part of the field and at least 600 m thick in the western part (LÁNG et al. 1970, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999). The difference in thickness can be explained by a large strike slip NW–SE trending fault cutting through the middle of the peninsula that displaced the pre-volcanic sequence (LÁNG et al. 1970, DUDKO 1999, BUDAI et al. 2002). The older Miocene rock units are limestone formations with high porosity with a potencial karst-water storage capacity.

The shoreline of the Pannonian Lake must have been just a few tens of km away from Tihany during the onset of volcanism (SACCHI et al. 1999, NÉMETH et al. 2001, SACCHI and HORVÁTH 2002). The depth of this shallow lake fluctuated both in short and long term (JUHÁSZ 1993, 1994, JUHÁSZ et al. 1997) and the maar volcanoes very likely developed in the vadose groundwater zone of the lake itself (MAGYAR 1988, SACCHI et al. 1999). Thus, the Tihany maar volcanoes evolved near a large body of standing water with an associated ground water table that fluctuated in response to seasonal rainfall that likely affected the groundwater circulation of the region. In this respect, the Tihany maar volcanoes are inferred to have developed in a setting similar to other near-shore maar volcanic fields such as in the South Australian Volcanic Field, Victoria (GRIFFIN et al. 1984, PRICE et al. 1997 JONES, et al. 2001) or the Auckland Volcanic Field, New Zealand (ALLEN et al. 1996, SHANE and SMITH 2000, SANDIFORD et al. 2001, HAYWARD et al. 2002, SHANE and HOVERD 2002). Such maars are often flooded by the nearby water masses during lake or sea-level high stands, and the level of their crater lakes strongly depends on nearby water level changes (JONES et al. 2001, HAYWARD et al. 2002).

Table 3.2. Lithofacies discrimination diagramme on the basis of identified volcaniclastic lithofacies from Tihany (after NÉMETH et al. 2001)

Volcanism related facies	Tuff breccia (TB)	Lapilli tuff (LT)	Tuff (T)
Clast-supported			
Non-volcanic lithic-rich			
1 Massive	TB1	LT1	T1
2 Weakly-bedded	TB2	LT2	T2
Scoriaceous			
3 Massive	TB3	LT3	T3
4 Weakly-bedded	TB4	LT4	T4
5 Well-bedded	TB5	LT5	T5
Matrix-supported			
Non-volcanic lithic-rich			
6 Scour-fill bedded	TB6	LT6	T6
7 Channel-fill massive	TB7	LT7	T7
8 Unsorted massive	TB8	LT8	T8
9 Strongly lithified pisolitic massive	TB9	LT9	T9
10 Pisolitic	TB10	LT10	T10
11 Diffusely stratified	TB11	LT11	T11
12 Thinly bedded	TB12	LT12	T12
13 Cross-stratified	TB13	LT13	T13
14 Undulatory-bedded	TB14	LT14	T14
15 Dune-bedded	TB15	LT15	T15
Scoriaceous			
16 Scour-fill bedded	TB16	LT16	T16
17 Unsorted massive	TB17	LT17	T17
18 Thinly bedded	TB18	LT18	T18
19 Dune-bedded	TB19	LT19	T19
20 Inverse-to-normal graded	TB20	LT20	T20



Tihany volcanic succession

The primary and reworked volcaniclastic deposits of the peninsula can be divided into 3 stratigraphic units (PH = Phreatomagmatic units, M = Magmatic units, ML = Maar Lake units) based on the relative amount of lithic clasts, sedimentary structures of the deposits and their stratigraphic position (Tables 3.2 and 3.3). The identification of individual lithofacies has followed the methods used for tuff rings and cones at Cheju Island, Korea (SOHN and CHOUGH 1989, 1992, 1993, CHOUGH and SOHN 1990, SOHN 1995, 1996). Each stratigraphic unit contains lithofacies associations (Figures 3.1 and 3.2) which are subdivided into 30 separate facies (Tables 3.2 and 3.3). Detailed facies descriptions are in separate papers

Table 3.3. Lithofacies association diagram for the description of the volcaniclastic lithiofacies that have been recognised in Tihany (after NÉMETH et al. 2001)

Stratigraphic units	Lithofacies associations	Lithofacies	Interpretation
ML	ML2	ML2	Maar lake centre lacustrine sedimentation
	ML1	LT20, T20	Maar lake margin Gilbert-type delta fronts with volcaniclastic gravity flow deposition
M	MSH	TB3, LT4	Hawaiian lava fountaining with lava spatter deposition with occasional clastogenetic lava flow forming
	MS	LT4, LT5, LT17	Strombolian fall out deposition
PH	PH4	LT6, LT8, T10, LT12, T12, LT13, LT14, T14, LT15, T15	Shallow locus (?) "dry" phreatomagmatic explosion derived dilute pyroclastic density current and co-surge fall out deposition
	PH3	T10, LT12, LT15, LT16, LT17, LT18, LT19, T19	Shallow locus, "dry" phreatomagnetic explosion derived dilute pyroclastic density current deposition influenced by simutaneous Strombolian activity
	PH2	LT1, LT2, TB7, LT7, LT8, T8, LT11, LT13, LT15	Deep locus, "wet" phreatomagnetic explosion derived high concentrated pyroclastic density current and co-suge fall outdeposition
	PH1	LT6, LT7, LT8, LT9, T10, LT12, T12, LT14, T14, LT15, T15	Shallow locus, "wet" phreatomagnetic explosion derived dilute pyroclastic density current and co-ruge fall out deposition
	PHLD	TB8, T8, T14	Phreatomagmatic vent-filling lapilli tuff deposited by "en masse" fall back of collapsing phreatomagnetic eruption collumn, accompined by occasional the vent

Figure 3.1. Simplified stratigraphic log of Barátlakások (Tihany; NÉMETH et al. 2001) SS = Silurian schist, PRS = Permian red sandstone, MF = Mesozoic formations, NS = Neogene siliciclastic sediments

(NÉMETH et al. 1999a, 2001). In the following sections detailed descriptions and interpretations are given from the identified lithofacies associations of the Tihany Maar Volcanic Complex (NÉMETH et al. 2001).

PH1 lithofacies association

Description: This lithofacies association represents the lowest part of the volcanic sequence at Tihany (Plate 3.2, A). It discordantly overlies the NS sequences, and has a sharp contact with the overlying PH2 lithofacies association. It crops out in the north-eastern and western hill side of the study area (Plate 3.1) and consists of fine-grained, thinly (LT12, T12), undulatory (LT14, T14), dune bedded (LT15, T15), accretionary lapilli rich (T10) tuffs, lapilli tuffs with ballistic bombs (Figure 3.3), massive (LT8), and related compacted horizons (Plate 3.2, B – LT9), generally medium lapilli tuff units. Scour-fill bedded (LT6) and small channel fill massive (LT7) units are more frequent in the upper part of the lithofacies association. The lithofacies association generally contains about 80 vol.% (visual estimate) of lithic clasts, which are dominantly (more than 50 vol.%) fragments derived from NS sequences (Plate 3.2, C). Large (larger than 5 cm) lithic clasts are dominantly derived from PRS units. Trajectories of impact sags at Barátlakások indicate south to north transport, similar to the flow direction derived from density current deposits. At the western side the transport directions indicate north-east to south-west transport.

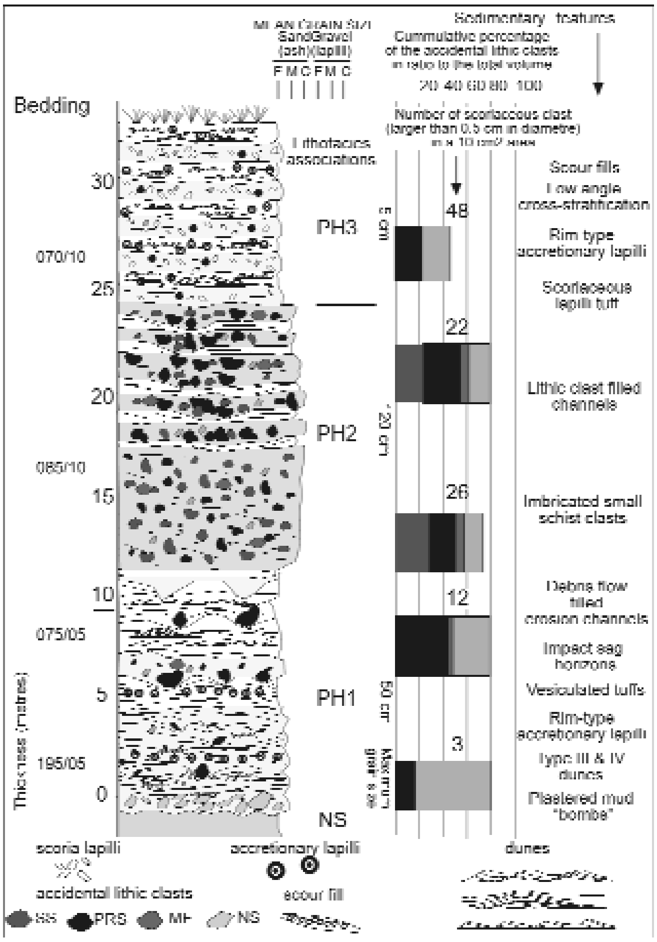


Figure 3.2. Composite stratigraphic log of the eastern outcrops (Barátlakások) at Tihany (Németh et al. 2001). Legend see on Figure 3.1.

Interpretation: Abundant dune structures, planar bedding and unsorted, fine-grained character suggest a complex pyroclastic density current (base surge), co-surge fall-out, fall-out and ballistic origin of PH1 (FISHER and WATERS 1970, CROWE and FISHER 1973, FISHER and SCHMINCKE 1984). The large amount of lithic clasts in similar deposits has been interpreted to record subsurface phreatomagmatic, maar-forming explosions that occurred during eruptions (FISHER and SCHMINCKE 1984, LORENZ 1987, WHITE 1991a, 1991b). Planar beds (LT12, T12) are inferred to record phreatomagmatic fall-out deposition, probably related to co-surge ash clouds (FISHER and SCHMINCKE 1984, SOHN and CHOUGH 1989). The high proportion of irregularly shaped NS fragments in the basal zone of the lithofacies association indicates that the first explosions occurred at a shallow level, in unconsolidated, water-saturated mud and sand (WHITE 1991b, ORT et al. 1998, HOUGHTON et al. 1999). In this lithofacies association, the increase in accidental lithic clasts of deeper-seated origin indicates that either the explosion focus down-migrated during the eruptive history (LORENZ 1986) or the vent progressively widened downward (LORENZ 1986, ORT et al. 1998, LORENZ 2000b). U-shaped channels filled with debris flow deposits (LT7) represent syn-volcanic reworking of the tephra (FISHER 1977). Accretionary lapilli beds (T10), vesiculated tuffs (part of T12), soft deformation under impact sags, mud-cracks and debris flow deposit filled erosion channels suggest a "wet" depositional environment (FISHER and WATERS 1970, WATERS and FISHER 1970, CROWE and FISHER 1973, SCHMINCKE et al. 1973, DELLINO et al. 1990).

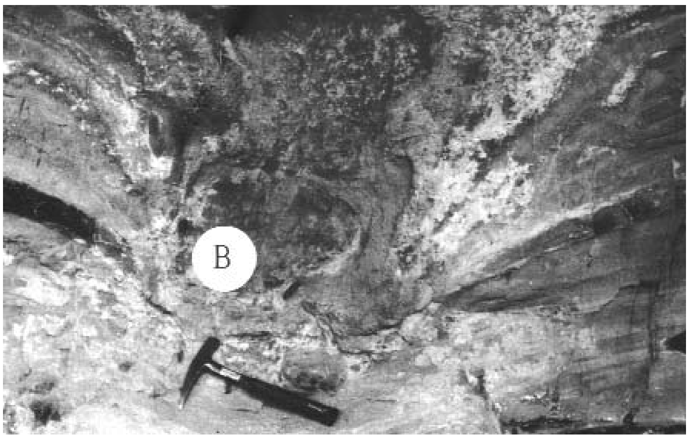


Figure 3.3. Impact sag (B) caused by a ballistically emplaced block from the Permian units, photo is taken from the Barátlakások outcrop

PH2 lithofacies association

Description: This lithofacies association represents the middle part of the volcanic sequence at Tihany (Figures 3.1 and 3.2). It discordantly overlies PH1, has a sharp contact with the overlying PH3 litho-

Figure 3.4. Unconformity (arrow) surface (arrow) between PH1 and PH2 at Barátlakások
For explanation of LT8 see Table 3.2

facies association and only occurs at the Barátakások locality (Figure 3.4). It consists of unsorted massive tuff breccia (TB8), medium- to coarse-grained lapilli tuff (LT8), and tuff (T8) interbedded with either clast-supported units (LT1, LT2) typically related to major channel-filling units (TB7, LT7) or less commonly, diffusely stratified (LT11), cross-stratified (LT13) or dune bedded (LT15) units. The beds of this lithofacies association generally contain ~60 to 80 vol.% (visual estimate) lithic clasts which are dominantly SS and PRS fragments (each ~30 vol.%. – Plate 3.2, D). The orientation of large channels and flow-direction derived from density current deposits indicate south to north transport.

Interpretation: The PH2 lithofacies association is a succession of phreatomagmatic base surge and fall-out deposits interbedded with debris flow deposits. The volumetrically dominant facies (LT8) and the channel-filling facies represent rapid, proximal sedimentation of multiple highly concentrated, laminar to turbulent, cohesive mass flows (SOHN and CHOUGH 1989). The near-vent origin is supported by the presence of ballistic blocks and bombs and the overall coarse-grained character of the beds. The fine-grained, stratified, dune-bedded facies are related to more dilute pyroclastic surges (SOHN and CHOUGH 1989). The abrupt increase of the SS and PRS fragments upward in the ejecta pile suggests that the explosion focus migrated downward and/or the vent widened progressively downward with time. The considerable thickness and uniformly lithic clast-rich nature of the lithofacies association reflect a steady supply of water to active eruption sites that supported explosive fuel-coolant interaction (NÉMETH et al. 2001).

It is probable that progressive deepening of the explosion focus coupled with an increased water influx when the base of the vents intercepted abundant water hosted in the fracture-controlled aquifer produced less effective phreatomagmatic explosions during deposition of PH2 (SHERIDAN and WOHLETZ 1983, WOHLETZ 1983, WOHLETZ and SHERIDAN 1983, WOHLETZ 1986) relative to the explosions that formed PH1, resulting in weakly fragmented, structureless lithic-rich deposits (PH2). The resulting low, dense, wet eruption clouds formed high-concentration pyroclastic mass flows that moved horizontally away from eruptive vents and deposited near-vent pyroclastic density current deposits (NÉMETH et al. 2001). The large amount of quartzofeldspathic fragments in the matrix of the beds and the plastic deformation of NS clasts indicate the water-saturated unconsolidated state of the Neogene siliciclastic unit-derived sand slurry in the vent. Mud-coated cauliflower bombs, commonly with sand inclusions, in PH2 beds also support the presence of unconsolidated muddy, sandy volcaniclastic slurry in the volcanic vent/conduit at this stage of eruption (e.g. LORENZ et al. 2002).

PH3 lithofacies association

Description: This is the highest unit of the volcanic sequence and is present at only one site in the north-eastern side of the study area (Plate 3.1). It discordantly overlies PH2 and consists of scoria-rich facies (LT16, LT17, LT18, T18, LT19, T19) that are the counterparts of the lithic-rich facies, PH1. It includes subordinate non-volcanic lithic-rich facies as well (T10, LT12, LT15). PH3 generally contains ~20 to 40 vol.% (visual estimate) lithic clasts, predominantly fragments derived from PRS and NS. Trajectories of impact sags indicate south to north transport, similar to the flow direction derived from density current deposits.

Interpretation: PH3 is composed of a series of "dry" pyroclastic surge (unsorted, dune-antidune bedded, cross-bedded, fine-grained beds) and fallout (moderately sorted, mantle bedded, accretionary lapilli-rich, fine-grained beds) deposits. The higher concentration of scoriaceous juvenile fragments implies a decrease in the water to magma ratio, most likely caused by an increase in the magma discharge (FISHER and SCHMINCKE 1984). The increasing role of magmatic explosivity in driving eruptions may imply a continuous drying of the system (HOUGHTON and SCHMINCKE 1989, HOUGHTON et al. 1996).

PH4 lithofacies association

Description: PH4 is exposed only on the western side of the study area (Plate 3.1) where it discordantly overlies PH1 and is overlain by travertine mounds interpreted to be maar lake deposits (NÉMETH et al. 1999a, NÉMETH 2001) although there is no exposed contact. It consists of similar facies to PH1 except that PH4 contains a minor proportion (less than 10 vol.%) of PRS-derived lithic clasts and exhibits features suggestive of higher temperature deposition (e.g. baked NS sandstone clasts). The facies grouped in PH4 are LT6, LT8, T10, LT12, T12, LT13, LT14, T14, LT15 and T15. Trajectories of impact sags at Barátlakások indicate north-east to south-west transport similar to the calculated flow direction derived from density current deposits.

Interpretation: This lithofacies is interpreted as having been generated by a similar eruptive mechanism to that which produced PH1 (i.e. predominantly low-concentration pyroclastic density currents and fallout). However, the lithic clast population differs from PH1 and also sedimentary structures (fractured beds under impact sags, no accretionary lapilli) indicate less water involvement in the phreatomagmatic processes. Several factors suggest that PH4 was erupted from a different vent with respect to PH1-PH3:

1. PH1, PH2 and PH3 lithofacies associations are (partly) capped by (or related to) maar lake deposits with bedding dip directions indicating a vent on the eastern side of the study area, whereas PH4 is capped by maar lake deposits with dip directions indicating a vent on the western side of the study area; and
2. PH1–PH3 are predominantly distributed on the eastern side of the study area whereas PH4 occurs on the western side.

PHLD lithofacies association

Description: PHLD occurs only in the middle part and on the western side of the study area (Plate 3.1). From drill hole Tht-2, a 200 m sequence of PHLD is also known. This lithofacies association consists dominantly of unsorted, massive tuff breccia (Figure 3.5 — TB8) and lapilli tuff (T8) with fluidisation structures interbedded with undulatory bedded tuff often contain accretionary lapilli horizons (Plate 3.3, A, B — T14). In the surface outcrops, beds of PHLD are overlain by maar-lake debris flow and turbidity current deposits and seem to be in steep contact with underlying NS beds (NÉMETH 2001).

Interpretation: The poor sorting and fluidisation structures within PHLD beds imply a vent filling position (i.e. lower diatreme deposit – term after WHITE 1991b) where the fall-back tephra is emplaced "en masse" at the base of the funnel-shaped vent (LORENZ 1971, 1973, 1975, 1985, 1986, 1987, 2000a, 2003b, WHITE 1991b). Coarse polymict breccia from the drill core is inferred to represent initial vent-clearing episodes. The thick, unsorted massive, matrix-supported characteristics of the majority of the core beds support "en masse" fall-back emplacement after discrete explosions. Finely laminated tuffs are probably inter-eruption suspension-deposited ash layers.

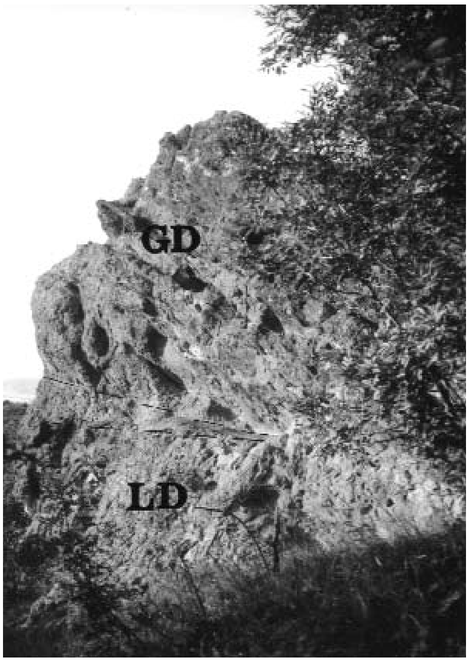


Figure 3.5. Massive, unsorted tuff breccia from PHLD of the basal zone Kiserdő-tető (LD)
The tuff breccia is rich in clasts of every known pre-volcanic rock unit. GD = is interpreted to be a Gilbert-type delta. (NÉMETH 2001)

MS and MSH lithofacies associations

Description: MS consists of facies LT4, LT5 and LT17. MS is scoriaceous, moderately to well-sorted, medium to coarse lapilli with a variable amount of matrix. MS occurs on the northern side of the study area and is known from the top 12 m of Tht-2 drill core. MSH consists of TB3 and LT4 facies and is predominantly spatter-rich with moderate matrix content.

Interpretation: The scoria lapilli-rich, bedded lapilli tuffs are interpreted as remnants of a Strombolian scoria cone. The vent site is not known but the presence of MSH at the northern side of the study area suggests that a larger vent was present there. This vent produced small-volume Hawaiian spatter deposits. Quartzofeldspathic sand in the vesicles is either Pannonian sand or both Pannonian sand and tephra, which suggest that sediment-charged slurry was probably present in the volcanic vent/conduit during magmatic explosive eruptions, or that sediment was disrupted from conduit and vent walls during explosions (HOUGHTON and NAIRN 1989). The common, dense volcanic blocks could be derived from degassed magma, which had a long residence in the vent, or from a disrupted sill or dyke (HOUGHTON and HACKETT 1984, BÜCHEL and LORENZ 1993).

ML1 and ML2 lithofacies associations

Description: ML1 and ML2 occur in the western and central part of the study area (Plate 3.1). ML1 discordantly overlies PH1, PH4 or PHLD, and is discordantly overlain by laminated fresh water limestone beds (ML2 — Figures 3.6 and 3.7). ML1 consists of scoria-rich inverse-to-normally graded lapilli tuff (Figures 3.7 and 3.8 and Plate 3.3, C) and tuff, facies LT20 and T20 respectively (NÉMETH 2001). The beds of ML1 dip steeply (>20°) toward local depressions (Figures 3.6 and 3.7). These pyroclastic rocks are calcite cemented, often inverse-graded and rich in broken phenocrysts and/or xenocrysts (Figure 3.8 and Plate 3.3, C). ML2 is a finely laminated silicified mound succession often truncated by vertical pipe structures (Plate 3.3, D) and soft sediment deformation features such as dish structures (Figure 3.9).

Interpretation: These lithofacies are interpreted as reworked tephra which were transported by grain flows (inverse-to-normal grained, coarse-grained beds) and/or turbidity currents (fine-grained, bedded, cross-bedded beds) into the maar lake, producing Gilbert-type delta fronts (WHITE 1992, NÉMETH 2000). The reworked origin is supported by the high amount of different types of volcanic glasses (tachylite and sideromelane), relatively well sorted texture, rounded

Figure 3.6. Panoramic view to the Kiserdő-tető inferred to represent a preserved succession of a Gilbert-type delta (arrow) has been built into a former maar basin located left from the hill (NÉMETH 2001)
LB = Lake Belső, LK = Lake Külső, dashed lines represent bedding planes

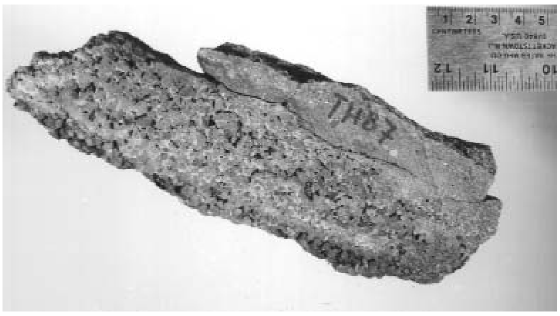


Figure 3.8. Close up of an inversely graded lapilli tuff, which is part of the Gilbert-type delta front from the Csúcs-hegy (western part of the Tihany Peninsula)

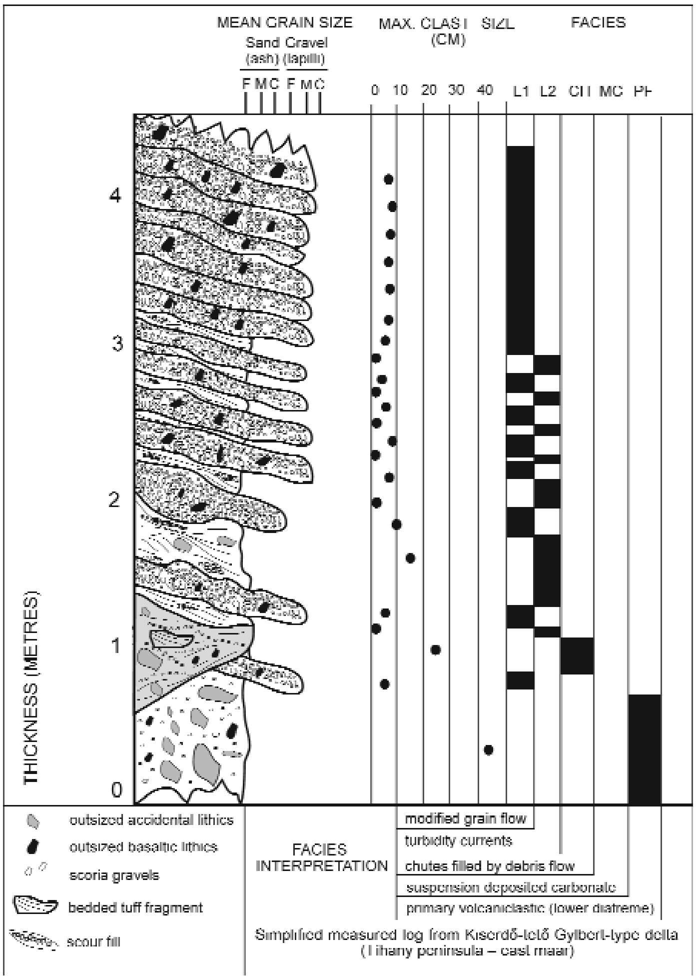


Figure 3.7. Simplified log from the Kiserdő-tető succession exposing the PHLD facies capped by volcaniclastic succession interpreted to be the result of a Gilbert-type delta in a former maar basin (NÉMETH 2001)

Figure 3.9. Soft sediment deformation features (arrow) in the silicified mound of Csúcs-hegy, interpreted to be the result of penecontemporaneous (probably volcanic) earthquake and/or vigorous degassing on the lake floor and/or post-maar volcanic resettling of the diatreme fill itself

clasts (often palagonite-rimmed), large amounts of calcite in the matrix, and the bedding structures and graded character. The dominance of scoriaceous fragments suggests reworking from late fragmented magmatic deposits with only minor reworking of the crater-rim phreatomagmatic tephra. The laminated carbonate with soft-sediment structures is interpreted as maar lake carbonate sediments with hot spring pipes (ML2). ML2 represents the latest volcanic-related depositional event in the field. Soft sediment deformation features in the silicified travertine mound of Csúcs-hegy, inferred to be a result of penecontemporenaeous (probably volcanic) earthquakes and/or vigorous degassing on the lake floor (ROGIE et al. 2000, ZHANG 2000, CHIODINI and FRONDINI 2001) similar to Lake Nyos (LEGUERN et al. 1992, COTEL 1999, FREETH and REX 2000), and/or post-maar volcanic resettling of the diatreme fill itself (LORENZ 2000a, 2003a, b) such as it has been reported from the Saxony, eastern Germany (LORENZ et al. 2003).

Volcanic centres in the TMVC

The occurrence of strong negative gravity anomalies (BENDERNÉ et al. 1965), locations of Gilbert-type delta fronts and the PHLD (lower diatreme) facies association point to at least three main maar/diatreme structures in the area (Plate 3.1). At two localities, maar-lake Gilbert-type deltafront deposits cover PHLD. This facies relationship is well preserved in both the East and West Maars. The central part of the peninsula (Lake Külső) is probably a third maar/diatreme structure that was buried by later scoria cones, which were subsequently eroded and the scoria re-deposited into the two open maar basins to the east and west. Maar lake carbonate deposits (ML2 often silicified by hot springs along the rims of maar craters, cover the large Gilbert-type delta fronts (ML1). Along the bedding planes of the steeply dipping scoriaceous reworked lapilli tuff beds, mineral-rich hot spring water was able to rise up to produce strong calcification and silicification filling the vesicles and pores of the originally open-work reworked tephra of debris flow/turbidity current deposits. The strong zonation of calcification of reworked tephra and their semi-circular aerial distribution above lower diatreme deposits (LÁNG et al. 1970), suggest the existence of three maar craters referred to as the East Maar, Central Maar and West Maar.

The best preserved maar-crater rim sequence is located at the north-east side of the East Maar (Barátlakások) and consists of PH1, PH2 and PH3. The inferred positions of the phreatomagmatic vents are supported by the measured transport directions of dune structures from the base surge deposits, imbrications of platy clasts, inferred orientations of ballistic trajectories from asymmetric bedding sags, and the distribution of bedforms (dune-bedded, plane parallel bedded, massive) from base surge deposits. The large size of ballistic blocks (25-120 cm) and their deep impact craters (20-45 cm) supports a near vent position for the phreatomagmatic lithofacies. This distribution of the maar lake beds (ML1) and their dip directions lend further support to reconstruction of three different phreatomagmatic vents in the area (West, Central and East Maar).

The position of the Strombolian scoria cone(s) is uncertain. Vent sites are inferred from the position of near-vent scoriaceous lapilli tuff (LT4, LT5), ribbon and spindle bomb-rich deposits (LT4) accompanied with spatter deposits (TB3).

The oldest maars are the Central and East Maars but there is no strong supporting evidence to determine which one developed first. The West Maar is probably the youngest because PH4 (derived from the West Maar) overlies PH1, which is inferred to come from the East Maar. Two lines of evidence indicate that most magmatic explosive activity postdates each maar forming process:

1. in the Central Maar, the lower diatreme units are covered by scoria beds; and
2. the maar lake deposits (LT20, T20) are scoria-rich and PH1-4 are relatively scoria-poor, thus scoria must have derived from scoria cones formed later than PH1–4 but earlier than ML. K/Ar ages of co-genetic lava fragments from PH1 (7.8±1.07 My, 7.56±0.5 My) and MSH (6.24±0.73 My, 6.64±0.71 My) show age differences that favour a younger age for MSH (BALOGH 1995, BALOGH and PÉCSKAY 2001, BALOGH and NÉMETH 2004), but the large error may change its significance.

Style of eruption of TMVC and its relation to the aquifers

It has been suggested that "passive" (simple collapse of conduit walls without significant involvement of explosive excavation) widening of the vent during down-migration of explosion locus alone might not be sufficient to explain the increase in accidental lithic clasts up-section in the pyroclastic succession described at Tihany, in addition the necessity of having the explosion focus in great depth needs to be better justified for the Tihany examples (NÉMETH et al. 2001). An alternative model was given recently that explains many of the unusual features of the Tihany maar-diatremes, and is explained briefly here (NÉMETH et al. 2001).

The large amount of deeply excavated lithic fragments in the volcanic succession of TMVC suggests phreatomagmatic explosions were driven by explosive interaction of magma with groundwater. The lithic clast population of pyroclastic rocks at Tihany indicates that the locus of the explosion must have been at some depth, and that the depth and duration of explosive activity must have remained relatively stable to generate the observed thick succession of indurated, predominantly "wet" phreatomagmatic pyroclastic deposits. The most common scenario in which surplus water is evident during phreatomagmatic explosive eruptions occurs when the explosion locus is near-surface and located in a standing water body. This eruption style generally results in formation of tuff cones. In contrast, at Tihany it is clear that the explosion locus must have been deep for the most of the lifetime of single vents, yet the pyroclasts must have been transported by water charged, relatively high clast density flows, which are inferred on the basis of the small spatial distribution of such deposits. In this respect, the situation at Tihany is very similar to that reported from the Joya Honda, Mexico (ARANDA-GOMEZ and LUHR 1996). At Joya Honda, the very "wet" phreatomagmatic eruption style was controlled by the hydrological characteristics of a fracture controlled aquifer (ARANDA-GOMEZ and LUHR 1996), similar to the deep rock units that underlie Tihany (e.g. karst-water bearing Mesozoic and/or older Miocene limestone units). However, in Tihany, the fracture controlled aquifer is covered by a few hundred metres of siliciclastic porous media aquifer, which acted to confine the explosions in the initial period of magma uprise, allowing the first magmas to rise through the water filled fractures in the deep (e.g karst-water) aquifers, without explosive magma-water interaction (NÉMETH et al. 2001). Because of it has low to moderate hydraulic conductivity, water in a porous media aquifer may not flow fast enough to the vent area despite the abundance of groundwater in the rest of the aquifer. Thus the conditions for a purely magmatic eruption may be reached and Strombolian-type eruptions may occur, and explosion locus will tend to migrate downward (LORENZ 1985, 1986). This process leads to eruption of the type of pyroclasts that forms the initial PH1 lithofacies association at Tihany (NÉMETH et al. 1999a, 2001). In fracture-controlled aquifers, in which secondary permeability may locally be very large, water flow is controlled by open channel systems and water flow velocity can be very high, in the order of kilometres per day in karst (compared to metres or centimetres per day in a porous media aquifer. – PADILLA and PULIDOBOSCH 1995, LAROCQUE et al. 1998). Therefore, once the fracture-controlled aquifer is breached and the water supply to explosion sites abruptly increases, the time between phreatomagmatic blasts will be considerably shorter than for explosions seated in the porous-media aquifer, and the total energy output of the volcanic event will be consumed through a closely-spaced series of approximately uniform explosions. The resulting pyroclastic deposits will be monotonous, with only minor breaks marked by subtle changes in the average grain size and/or the proportion of juvenile to lithic clasts, like the PH2 lithofacies association deposits in the East Maar. It is noteworthy that early geological descriptions from the Tihany (HOFFER 1943a, b) referred to the identified vents as mud volcanoes ("iszap" or "sár" volcano), which produced laterally moving, high density mud-charged, wet pyroclastic currents, which had not travelled far from their vents.

For the eruption history for the TMVC the following summary can be given (Figure 3.10 – NÉMETH et al. 2001). The TMVC aquifer was a unique combination of a porous media and a fracture-controlled aquifer. At the initiation of volcanism, the fracture-controlled aquifer was covered by ~200 m thick of porous media siliciclastic aquifer. The thickness of the porous media aquifer was enough to develop a cone of depression in the porous media aquifer and produce a clearly down-migrating explosion locus in the initial eruptive phase (Figure 3.10, A–B), which resulted in deposition of the lowest stratigraphic position beds of the PH1 lithofacies association (Figure 3.10, C). As the eruption progressed the fracture-controlled aquifer was disrupted (Figure 3.10, D) causing

1. an increase in the secondary permeability and major water influx into the system, and

2. further excavation of the early maar crater, decreased lithostatic pressure on the ascending magma and therefore increased vesiculation. At this stage the system became a partially open system and the energy of the explosions was consumed by evacuating the vent-filling, sediment-rich slurry from craters, and fracturing the wall rock. Because the explosive energy was used for ejecting the wet slurry (probably several 100 m thick), the resulting eruption column was probably not high, thus producing the horizontal high concentration mass flows that were deposited in the immediate vicinity of the vent site (PH2). This type of eruption mechanism is apparently more common in the BBHVF than has been previously documented. For example, there is evidence of dual aquifer involvement from Pula (see later in this chapter), where a "champagne glass" shape maar crater developed over the karst water-bearing Mesozoic basement. The shape of the vent, with an abrupt widening at the top of the karst aquifer, suggests that the explosion locus must have been stabilised its position on the unconformity between the Mesozoic fracture controlled and the Neogene porous media aquifers.

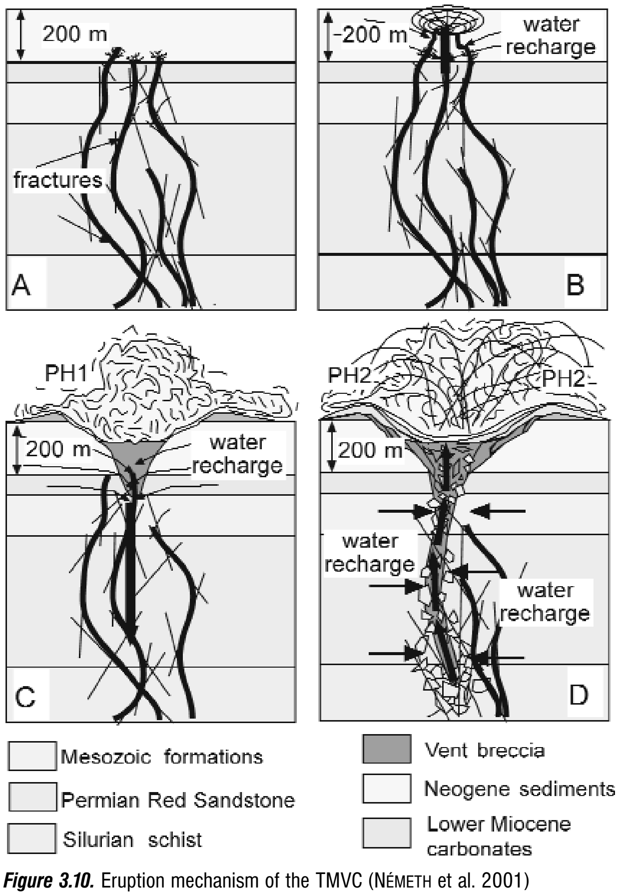


Figure 3.10. Eruption mechanism of the TMVC (NÉMETH et al. 2001)
A, B, C and D represent evolutional sequence of the eruption

The modern fracture-controlled aquifer in the Tihany area shows a strong seasonality linked to water supply by rainfall or spring run-off. It has been suggested previously, if this was the case historically, as seems likely, that the volcanic eruption style (either "wet" or "dry") might have been affected by the state of water saturation of both the porous media and/or the fracture controlled aquifers (NÉMETH et al. 2001). In this way, the term "*spring vent/volcano*" has been introduced for vents in those areas developed over aquifers with a dual recharge style that formed in the wet season, and "*summer vent/volcano*" for vents/volcanoes developed when the fracture-controlled aquifer was "empty", with subsequent formation of scoria cones rather than phreatomagmatic volcanoes (NÉMETH et al. 2001). The term "*Tihany-type maar*" is suggested for maars for which the eruptive mechanism is inferred to have been strongly influenced by two strikingly different sub-surface aquifers, after the described type locality (NÉMETH et al. 2001).

Hegyes-tű plug and surrounding eroded diatremes

Introduction

In the eastern part of the Kál Basin and its eastern border (Plate 3.4) 4 small diatremes have been identified. The diatremes have cut into the pre-volcanic, predominantly Mesozoic or older rock sequence (Plate 3.5, A). In this region, Neogene sediments are preserved only as thin veneers blanketing the surface of the Mesozoic basement, and lithic blocks of Neogene sedimentary rocks within the fill of the small diatremes are the only hints that those rock formations existed in the region during volcanism. The landscape shows characteriscs of an eroded landscape, which is cut through by the small diatreme pipes (NÉMETH et al. 2003 – Plate 3.5, B). The Hegyes-tű is a remnant of a basanitic coherent lava body, forming the highest elevated hill of the four erosional remnants discussed here. At Hegyes-tű, no pyroclastic rocks have been found. In contrast, the other three localities are diatreme remnants formed of pyroclastic material.

Hegyes-tű plug

Hegyes-tű is a 336 m high and about 200 m wide landmark in the Kál Basin, formed by a columnar-jointed basanite plug (Figure 3.11). The columns are predominantly vertical, having some bends in the marginal zone of the exposure. The columns are fairly regular, and range in diameter between 10 and 45 cm. There is no systematic distribution pattern of column diameter in the exposed section. In the upper part of the columnar jointed lava body, zones of slightly vesicular lava can be recognised, indicating entrapped water rich sediment, or water itself, in the melt during its eruption. The vertical jointing pattern indicates that cooling of the melt was along isotherms perpendicular to the jointing pattern (DEGRAFF and AYDIN 1987, BUDKEWITSCH and ROBIN 1994). This would imply that the preserved basanite outcrop at Hegyes-tű is related to a lava body that represents a horizontally emplaced melt. The present elevation of the inferred contact between the pre-volcanic rock units and the volcanic rocks is around 280–300 m and thus this level would mark a possible palaeosurface. However, the irregularities in the jointing pattern, especially in the north-eastern side of the outcrops, indicates that the lava body is not part of an extensive lava sheet, and instead represents a remnant of a lava that had a complex cooling history, as would be expected in a volcanic conduit/crater zone.

Figure 3.11. Columnar jointed basanitic lava plug of the Hegyes-tű

In the northern side of the Hegyes-tű plug, highly vesicular basanite clasts form a clastic rock unit that has a strongly palagonitized, mud-rich matrix (Plate 3.6, A). This clastic zone is surrounded by coherent lava, defining a well-localised structure, possibly a "bubble", that formed inside the still-liquid basanite. The highly vesicular clasts vary in size from ash to block and the outcrop may be interpreted as a tuff breccia (Plate 3.6, A). The blocky, rugged shape of the chilled vesicular lapilli exhibit irregularly shaped vesicles (Plate 3.6, B) characteristic of magma–water interaction (HEIKEN 1972, 1974, DELLINO and LIOTINO 2002). The presence of mud and siliciclastic fragments (quartz grains, and/or mud chunks) between the chilled lapilli as well as in a few of the vesicles indicates that the formation of these deposits was somehow related to magma-water interaction near the active vent. The strong palagonitization (Plate 3.6, B) of this preserved pyroclastic unit indicates a water-rich depositional environment and high temperatures (FARRAND and SINGER 1992, AUGUSTSSON 2001). The plug is similar in texture and size to the vent-filling of a scoria half-cone section exposed at the East Grants Ridge in New Mexico, often known as "the Plug" (CRUMPLER 2003).

Zánka, Vár-hegy diatreme

Just 2 km south of Hegyes-tű in the foothills of a 300 m high range around 1.7 km from the northern shoreline of Lake Balaton, a small outcrop of pyroclastic rocks, cut through the Palaeozoic–Mesozoic basement of the BBHVF (BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, NÉMETH et al. 2003), that forms a ~10 m high, 30-50 m wide by 250 m long ridge elongated toward the NNE–SSW. The very poorly exposed outcrop of yellow, brown lapilli shows a weakly stratified tuff (NÉMETH et al. 2003). This unsorted, weakly bedded lapilli tuff is rich in volcanic glass shards with variable amounts of elongated vesicles (NÉMETH et al. 2003 – Plate 3.6, C). The texture of the glass shards is defined by the distribution of microlites (Plate 3.6, D). The lapilli tuffs contain small fragments of mud, a high percentage of quartz grains and muscovite flakes inferred to have derived from Neogene siliciclastic units (NÉMETH et al. 2003).

Horog-hegy diatreme

Horog-hegy represents a small pyroclastic unit. It is located about 2 km west from the Hegyes-tű, at the eastern margin of the Kál Basin (Plate 3.4). The pyroclastic rocks form a circular shape zone in map view with ~100 m in diameter, rising above the surrounding agricultural land.

The recovered pyroclastic rock debris exhibits diverse textures that are indicative of magma–water interaction. The rock is rich in blocky sideromelane ash to lapilli size pyroclasts that are moderately microvesicular and rich in angular lithic fragments of Palaeozoic and Mesozoic rock units as well as in "exotic" xenoliths and megacrysts such as lherzolite, or amphibole aggregates (Plate 3.6, E). Mud chunks and irregularly shaped silt fragments characteristic of Neogene sedimentary rocks are common in these pyroclastic samples, as well as single grains of quartz and muscovite inferred to derive from these units. The composition of the volcanic glass shards, measured by electron microprobe method, from this location tends to be evolved phonotephrite to tephriphonolite.

Kis-Hegyes-tű (Lapos-Hegyes-tű) diatreme

About 4 km south-west of Hegyes-tű, Kis-Hegyes-tű (238 m) and Lapos-Hegyes-tű stand ~100 m above the floor of the Kál Basin (Plate 3.4). The double hill (each about 250 m long and 100 m wide) north to south elongated consists almost exclusively of steeply dipping bedded pyroclastic rocks. The pyroclastic units are apparently above the Neogene siliciclastic successions; these siliciclastic rocks are quarried in the northern side of the preserved pyroclastic hills. The contact between the pyroclastic rocks and the siliciclastic rocks is inferred to be steep and 60 metres below the top of the hill of Lapos-hegyes-tű (Plate 3.5, A). Palaeozoic rocks crop out 1 km to the north from the Kis-Hegyes-tű. Geophysical anomalies indicate of low density and/or fragmented rocks below this location. The identification of Neogene sediments on the top of the nearby Palaeozoic range (BUDAI and CSILLAG 1998, 1999) suggests that these pyroclastic hills are diatremes that cut through the pre-volcanic rocks.

The pyroclastic rocks of this region form alternating beds of coarse and fine tuff and lapilli tuff that are rich in volcanic glass shards. The volcanic glass shards are blocky and moderately vesicular with a small proportion of microlites (Plate 3.6, F). The pyroclastic rocks contain lithic clasts derived from every known type of pre-volcanic rock unit. These rocks are very rich in baked, subrounded Neogene silt- and sandstone clasts. The matrix is often replaced by calcite cement, and shows intense palagonitization. Gel palagonite is especially prominent along fractures in the glass shards, and along the rims of the glass shards.

Haláp maar/diatreme and the surrounding eroded diatremes

Introduction

About 5 km north from the centre of the city of Tapolca, is the location of an erosional remnant of a maar/tuff ring volcano called Haláp (Figure 3.12). This erosional remnant overlies Triassic and Middle-to-Upper Miocene carbonate beds. The units immediately underlying the pyroclastic rocks at Haláp belong to the Neogene shallow marine to fluvio-lacustrine siliciclastic succession (BUDAI et al. 1999), which are not preserved around the Véndek-hegy (see later in this chapter), a diatreme just about 3 km west of Haláp (Figure 3.13). The existence of the Neogene sedimentary units is also ambiguous at the Hegyesd (Figure 3.13) diatreme 5 km south-west of the Haláp volcano (BUDAI et al. 1999).

Haláp maar/tuff ring

At Haláp, quarrying has removed the central lava lake facies of the volcano, leaving behind a "castle-like", about 500 m across pyroclastic ejecta rampart and thin lava layers, which allows study of the contact zone of the lava lake and the former crater filling rocks and/or tuff ring (Plate 3.7). The coherent basanite lava is dated using both

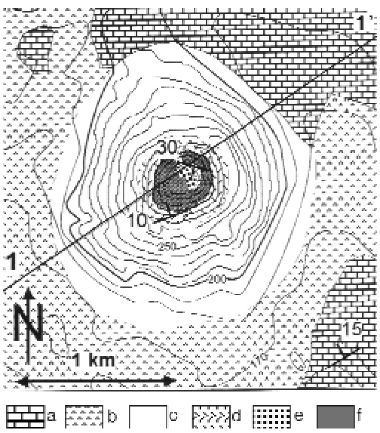


Figure 3.12. Simplified geological map of the region around Haláp maar/tuff ring volcano
a = Upper Triassic carbonates, b = Middle to Upper Miocene limestone, c = neogene siliciclastic units, d = bedded lapilli tuff (crater rim units) e = massive to bedded lapilli tuff (conduit filling facies), f = solidified basanite lava lake, 1–1' cross section shown on Plate 3.7, C

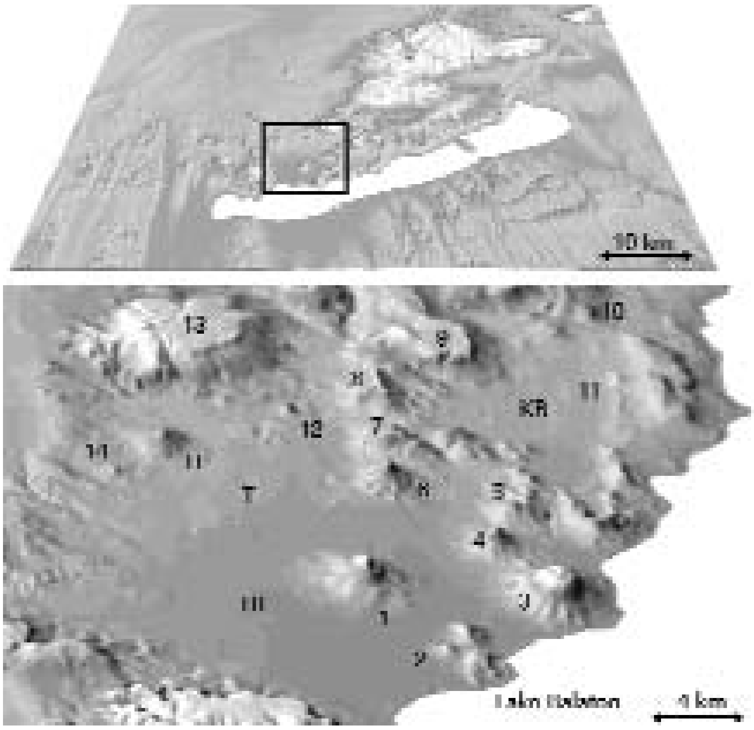


Figure 3.13. Digital Terrain Model of the Tapolca Basin looking from south-west to north-east
H – Haláp, 1 – Szent György-hegy, 2 – Szigliget group, 3 – Badacsony, 4 – Gulács, 5 – Tóti-hegy, 6 – Csobánc, 7 – Hajagos, 8 – Sátorma, 9 – Fekete-hegy, 10 – Hegyes-tű, 11 – Kis-hegyes-tű, 12 – Hegyesd, 13 – Agártető, 14 – Véndek-hegy, T = City of Tapolca, TB = Tapolca Basin

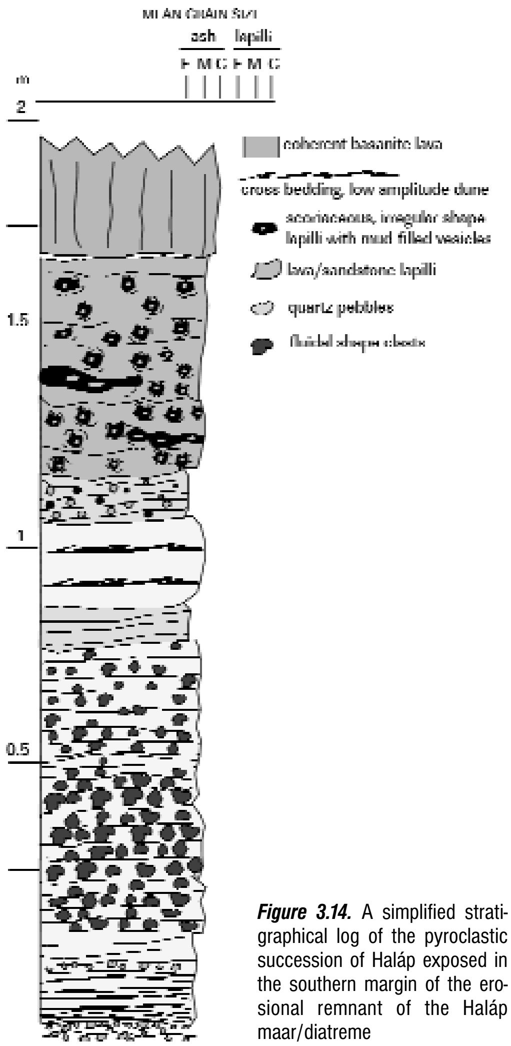


Figure 3.14. A simplified stratigraphical log of the pyroclastic succession of Haláp exposed in the southern margin of the erosional remnant of the Haláp maar/diatreme

K/Ar and $^{39}Ar/^{40}Ar$ techniques that resulted in more or less the same age of 3 My. The pyroclastic units (Figure 3.14 and Plate 3.8) at Haláp are yellow-brown, with abundant sideromelane glass shards (Plate 3.8, B). Single layers are well bedded, internally structureless or inversely graded and contain large volumes (~25 vol.%) of lithic clasts mainly derived from the underlying pre-volcanic fluvio-lacustrine sedimentary units (Late Miocene, Pannonian – Plate 3.8, C). The pyroclastic succession has been interpreted as being primary with locally remobilized material deposited from grain flows into the crater. The contact between lava and the pyroclastic units shows a peperitic structure in a zone of about one metre in thickness. The peperitic margin of the solidified lava lake comprises highly vesicular, irregularly shaped coherent basanite fragments (dm-scale size) through, which clay is dispersed (Plate 3.8, D). The clay in these fragments is inferred to have been derived from the tuff ring-forming tephra and/or the underlying pre-volcanic siliciclastic units mobilised by fluidisation caused by the emplacement of the hot lava into a dish-shaped, slurry-filled vent. The presence of these peperites at Haláp suggests a wet and unconsolidated state of the tephra prior to and during formation of the lava lake. In the center part of the Haláp maar/tuff ring a small stack of scoriaceous lapilli tuff is preserved (Plate 3.9, A, B) with steeply dipping beds. The clasts are well packed and a gradual upsection transition into more matrix-rich lapilli tuffs indicates a decreasing welding and increased cooling of the pile of hot pyroclasts grew during their deposition.

Véndek-hegy diatreme

Véndek-hegy (255 m) is just 3 km west of the Haláp, forming small hills (the largest is about 200 m and NE–SW elongated) that rise less than 60 metres above the pre-dominantly Middle Miocene/Triassic carbonate country rocks. It consists of three hills each forming a semicircular morphology similar to the Kis-Hegyes-tű/Lapos-Hegyes-tű volcanic remnants. The pyroclastic rocks cut through Triassic dolomite, thin Middle Miocene limestone and thin veneer of gravel horizons from the oldest sequence of the Late Miocene siliciclastic succession (BENCE and PEREGI 1988, BUDAI et al. 1999). The immediate pre-volcanic rock units are Late Miocene gravel beds that reach a thickness of 30 m nearby, at the Véndek-hegy (BUDAI et al. 1999). Mapping of the pyroclastic units has confirmed that the triple hillside is a uniform succession of pyroclastic rocks. No coherent lava has been identified yet from this locality (NÉMETH et al. 2003).

The pyroclastic rocks from Véndek-hegy are yellow to brown, unsorted, and weakly stratified to massive lapilli tuffs (NÉMETH et al. 2003). The pyroclastic rocks are rich in microvesicular sideromelane glass shards (Plate 3.9, C). In the tachylite glass shards entrapped mud is often present, which suggests that a certain degree of recycling in a closed vent filled with muddy slurry must have occurred during the eruption of the Véndek-hegy.

Véndek-hegy is therefore interpreted to be an erosional remnant of a diatreme that formed as a result of phreatomagmatic eruptions (NÉMETH et al. 2003).

Agár-tető shield volcano and scoria cone

Agár-tető (Figure 3.13) is an extensive lava field (about 8 km across) inferred to have been derived from several possible fissure sources. The lava field is capped by a remnant of a scoria cone with preserved geometry of about 600 m across and 80 metres above the lava field. The original morphology of the scoria cone is preserved (Plate 3.10, A, B), and the cone consists of large number of spindle-shaped basanitic bombs (Plate 3.10, C) that often contain lherzolite xenoliths. K/Ar ages for the Agár-tető volcanic complex range from ~5 to 2.8 My (BALOGH et al. 1982, 1986), indicating that it was a long-lived volcanic complex tapping a stable melt source over time.

Hegyesd diatreme

Hegyesd (Figure 3.13) is located ~5 km south-east from Haláp, and represents a typical deeply eroded diatreme (about 150 m accross, slightly north to south elongated), which has been preserved by its capping basanite lava. The basanite plug on the top of the erosion remnant is columnar jointed, dark grey coherent lava that intruded into a pyroclastic succession. The age of the plug has been determined by the K/Ar method, giving an age range from 3.43 My to 4.77 My; however, the widely accepted age of the diatreme is 3.70±0.28 My (BALOGH et al. 1982, 1986). New isochron $^{39}Ar/^{40}Ar$ dates give more-or-less the same range, at 3.91±0.19 My (WIJBRANS et al. 2004).

The pyroclastic succession of Hegyesd consists of weakly bedded lapilli tuff rich in volcanic glass shards. The volcanic glass shards are slightly vesicular, moderately microlite-rich to microlite-free, and tephritic in composition according to EMP measurements. The lithic componenet of the lapilli tuff (Plate 3.10, D) is made up of fragmented silt- and sandstones as well as silt to sand sized limestone and dolomite clasts from the pre-volcanic Triassic units.

Hajagos maar/tuff ring and surrounding eroded volcanoes

In this section erosion remnants from the eastern margin of the Tapolca Basin will be described (Plate 3.11). The immediate pre-volcanic rock units in this region are 50–100 m of Neogene siliciclastic units. The deeper pre-volcanic units consist of a Triassic carbonate succession plus Palaeozoic terrestrial and metamorphic units, as have been described earlier.

Hajagos maar/tuff ring

Hajagos maar/tuff ring is a prominent landmark at the eastern margin of the Tapolca Basin, forming a north to south elongated flat hill (~ 300 m high, and about 150 m above the surrounding with about 800 m across) with a well-defined, semi-circular distribution of a negative Bouguer-anomaly and positive magnetic anomaly. Due to active quarrying in the past decades a large quantity of basanitic lava has been removed and the deep crater zone of a phreatomagmatic volcano has been exposed with a great variety of peperites. A smaller hill (Láz-tető) at the southern margin of the major part of the Hajagos is 344 m high and is considered to be part of the same volcanic erosion remnant. The age of the volcano has been determined both by K/Ar and $^{39}Ar/^{40}Ar$ techniques giving a range of isochron ages of 3.94±0.25 My by the K/Ar method (BALOGH et al. 1986) and 3.72±0.05 My by $^{39}Ar/^{40}Ar$ methods (BALOGH et al. 1986). Hajagos hill is predominantly built up by basanite lava; however, in dissected outcrops lapilli tuff and tuff successions have been identified in a collar-like setting in the northern sector of the erosional remnant (Plate 3.12).

The bulk of the volcanic rocks and surrounding sedimentary formations are not well exposed at Hajagos-hegy (Plate 3.12). The best exposure in the northern side of the hill reveals tens of metres thick unit that underlies capping basanite lava flows. This volcaniclastic sequence dips 15° towards the centre of the hill. The volcaniclastic beds consist mostly of coarse-grained (lapilli size) and fine-grained (ash size) bed couplets. Coarse-grained lapilli tuff beds display normal grading. The basic, slightly micro-vesicular, predominantly tachylite lapilli are semi-rounded and display rims comprised of altered glassy material. A few small, angular and blocky grains of sideromelane show slightly oriented microlites. The lapilli are interpreted as clasts derived from pre-existing cemented volcaniclastic deposits, with the altered rims around larger volcanic clasts representing remnants of the former matrix of a lapilli tuff. The lapilli tuff beds are interpreted as reworked volcaniclastic rocks deposited in a volcanic depression such as a maar or the crater lake of a tuff ring (MARTIN and NÉMETH 2000). The crater zone of the Hajagos-hegy volcano is inferred to have been occupied by a volcaniclastic slurry that contained large (metre-scale) chunks of irregular-shaped siliciclastic sediments derived from rocks immediately underlying the volcanic sequence (Plate 3.13, A). Fine-grained beds largely consist of lapilli and ash of sideromelane and/or palagonite showing low-angle cross bedding, cross-lamination and often fining upward. The presence of overlying pillow basalts, the well-bedded character of these beds together with low angle cross-stratification suggest deposition of the fine-grained beds from turbidity currents, that require deposition by both traction and suspension (LOWE 1982).

The well-defined, semi-circular distribution of a negative Bouguer-anomaly, positive magnetic anomaly, and the semi-circular, gentle inward dip direction of the juvenile shard-rich beds suggest that the Hajagos-hegy is an erosional remnant of a low-profile phreatomagmatic volcano, a maar/tuff ring.

Subsequent invasion of the crater/conduit-filling sediments by dykes formed both globular and blocky peperite (term after BUSBY-SPERA and WHITE 1987), indicating that the host sediment was wet and unconsolidated during intrusion (MARTIN and NÉMETH 2000, 2004). Blocky peperite at Hajagos-hegy comprises basanite mingled with quartz sand and tephra host sediment (Plate 3.13, B). Blocky peperites at Hajagos-hegy appear at the deepest exposed level of the feeder dykes. This type of peperite consists of a fine-grained, brown to light brown, sandy matrix with angular basanite fragments (Plate 3.13, B). The siliciclastic host sediment most closely resembles sandstone from the Neogene siliclastic units that underlies the volcanic formations. The siliciclastic host sediments are not uniform beds but instead form well-defined zones, with single sandstone and siltstone "chunks" blokked in the vent/conduit zone of the phreatomagmatic volcano (Plate 3.13, A). The volcaniclastic host sediments of blocky peperites at Hajagos are coarse grained lapilli tuffs. Basanite fragments within the peperites show very thin chilled margins (max. 0.1 cm). Small (and a few large) basanite fragments (cm-scale) form a jig saw-fit structure. Some of the basanite blocks are located up to several metres from the margin of intrusions. Commonly the original bedding of the host sediment has been destroyed, probably by fluidisation (Plate 3.13, C). The juvenile clast/host sediment ratio decreases quickly within a few metres of intrusive contacts.

Fluidal peperite is also common at Hajagos-hegy, where the host is either lapilli tuff and tuff (Plate 3.13, D) or less commonly, quartz sand and silt (Plate 3.13, E). The volcaniclastic host closely resembles the maar-forming volcaniclastic and/or vent-filling pyroclastic deposits in its composition and texture. Fluidisation of the fine matrix of the volcaniclastic host is recorded in oriented crystals, glass shards and quartz grains preserved along the margins of larger (cm-scale) magmatic clasts at both the micro- and meso-scale (MARTIN and NÉMETH 2000, 2004).

Another type of globular peperite has been identified at Hajagos-hegy. It is developed at the base of lava flows (Plate 3.14), and forms pillows that commonly occur at least 10-15 m above feeder dykes with blocky peperite. The pillow fragments forming the basal zone of the lava flow are fairly regular in size and shape, and are commonly detached from the main pillow lava body and mingled with the host quartz sand. Piles of the basanite pillows are up to 1 m thick and closely packed.

The lava-foot peperite at Hajagos-hegy may have formed by a combination of two processes.

1. Basanite magma may have flowed over a swampy area or shallow ponds, where a large amount of steam locally formed "mega bubbles" in the lava flow. In this case the lava flows may be interpreted as having been emplaced into a surrounding swampy area after overflowing the former tuff ring rim.

2. Alternatively, the peperite formed subaqueously, while lava flows erupted from the vent zone. Between eruptions, thin sedimentary layers were deposited on the previously emplaced solidified lava. The next lava flow was trapped in the shallow water and captured the thin lacustrine siliciclastic sediments. The shallow water vaporised and caused the formation of the tumuli due to hydrostatic uplift.

In either case, the lava flow with its associated pillowed foot zone can be interpreted as lava entering a wet environment similar to lava-fed deltas, with associated passage zones (SCHMINCKE et al. 1997, SKILLING 2002).

The basanite dykes that intrude various host sediments are interpreted as feeder conduits to a lava lake, which infilled the crater (MARTIN and NÉMETH 2000, 2004).

Bondoró maar/tuff ring

Bondoró is located north-east of Hajagos and forms in a map view a lens-shaped volcanic remnant (about 2 km diameter), capped by a coherent lava flow (Plate 3.11). The contact between the lava flow and the underlying pyroclastic succession is about at 260–280 m elevation. The topmost surface of the lava flow is uniform and forms a plateau at an elevation of 300–320 m. A small north to south elongated crater-like feature (about 900 m in longer axis) rises ~100 m above the lava plateau. These morphological features consist of scoriaceous lapilli stone, and red scoria lapilli, intercalated with spattery lava fragments and lava with irregular morphologies, all indicating that this hill is an eroded scoria cone. K/Ar age determinations for the coherent lava from Bondoró give an age of ~2.3 My, indicating that this volcanic remnant is among the youngest landforms in the BBHVF (BALOGH and PÉCSKAY 2001). The young age of the coherent lava measured at Bondoró is in good concert with the observed morphological features of the hill, which indicate only moderate modification of its volcanic edifice. This is remarkable because erosion of scoria cones with an age over a million years usually results in complete modification of the volcanic edifice, leaving behind only a mound-like architecture (WOOD 1980a, b, HASENAKA and CARMICHAEL 1985, DOHRENWEND et al. 1986, INBAR et al. 1994, HOOPER and SHERIDAN 1998, INBAR and RISSO 2001). However, most of the cinder cone degradation models are based on cones with loose medium lapilli-rich scoria cones. The cone itself rests on a lava field covering a phreatomagmatic pyroclastic succession, best exposed along the eastern margin of the hill, right next to the village of Kapolcs. The pyroclastic succession is estimated to be at least 40 m thick, consisting of alternating lapilli tuff and tuff beds all rich in fine quartzofeldspathic grains typical of the underlying Neogene siliciclastic sedimentary succession (Plate 3.15, A). The fine tuff beds are rich in rim-type accretionary lapilli up to 0.5 cm in diameter. Volcanic glass shards within the volcaniclastic rocks are blocky, moderately vesicular, and microlite-poor, all indicative of sudden chilling of melt upon contact with water. The glass shards are commonly palagonitized, or have palagonite rims. The lapilli tuff and tuff beds also contain limestone and dolomite fragments from the Mesozoic strata, as well as broken phenocrysts and/or xenocrysts of olivine, spinel and pyroxene. The presence of the phreatomagmatic pyroclastic rocks at the base of the Bondoró suggests initial phreatomagmatic explosive activity that changed into lava effusion, with late lavas filling the crater (presumably a shallow maar). In the final stage a spatter-dominated cone built in the crater, which has retained its morphology after 2.3 My.

Csobánc diatreme

Csobánc hill (376 m) is located on the western margin of the Tabolca Basin, and is clearly visible from the top of Hajagos (Plate 3.11). The hill stands ~250 m above the Neogene siliciclastic sediment-filled basin, forming about a 300 m wide important landmark in the region (Plate 3.15, B). The hill is capped by lava spatter that has been intruded by basanite feeder dykes, today preserved as columnar jointed basanite.

The capping volcanic rock units at Csobánc are inferred to represent a welded lava spatter and scoriaceous lapilli succession (e.g. THORDARSON and SELF 1993, WOLFF and SUMNER 2000). Most of the lava exposures on the top of the hill still exhibit recognisable clast outlines of scoriae that suggest that mafic lava fountaining at Csobánc was a dominant eruption style, probably in the final stage of the eruption(s).

The lower section of the erosional remnant is a phreatomagmatic lapilli tuff succession that is only very poorly exposed. The preserved pyroclastic rocks crop out in the northern sector of the hill at ~250 m elevation. The basal pyroclastic rocks are uniformly poorly bedded, moderately sorted lapilli tuff and lapilli stone. These pyroclastic rocks are rich in strongly palagonitized volcanic glass, attesting to their phreatomagmatic origin. The large volume of juvenile clasts in these pyroclastic rocks indicates near-surface fragmentation, probably driven by interaction of magma with water in shallow level and/or surface reservoirs (WHITE 1991a). The pyroclastic rocks are often clast supported, and calcite cemented, indicating that the matrix of the rocks has either been washed out by secondary processes, and/or that the tephra deposits were originally fines-poor. The presence of siliciclastic silt and sand in the matrix, as well as a few lapilli-sized mud chunks, supports the first interpretation.

Pula maar and surrounding eroded volcanoes

Introduction

Pula is located in the central part of the BBHVF (Plate 3.15, C, D) and forms a small basin (about 800 m across) between the Kab-hegy shield volcano and the Tálodi-erdő lava field. Volcanic rocks around the small depression are widespread and all textural feature point to a phreatomagmatic origin. In the western margin of the Pula region, there are small hills that are very likely diatreme remnants; however, they have never been studied form a volcanological point of view.

Pula maar

Pula maar is a Pliocene eroded, phreatomagmatic volcano, and forms part of the Mio/Pliocene Bakony – Balaton Highland Volcanic Field. The remnant of the maar consists of a

1. distinct depression with a thick alginate (oil shale), lacustrine unit infill interbedded with coarse grained lapilli tuff,
2. a narrow belt of a primary pyroclastic unit along the margin of the depression (inferred to be the erosion remnant of the tuff ring) and,
3. a reworked coarse-grained volcaniclastic unit in the marginal zone. Palaeo-earthquakes associated with ongoing nearby volcanic eruptions and/or large volume debris flows initiated by crater wall collapses into the maar crater lake are inferred to have been responsible for the soft sediment deformation evident in fine-grained volcaniclastic sediments.

From the BBHVF, alginite (oil-shale) studies in the past decades have characterised laminated sediments formed in closed crater lakes such as Pula or Hercseg-hegy near Gérce (JÁMBOR and SOLTI 1975, 1976, BENCE et al. 1978, JÁMBOR et al. 1981, SOLTI 1986, FISCHER and HABLY 1991, PÁPAY 2001). However, only recently the importance of studies that describe and interpret the sedimentary processes involved in the formation of these maar pitted basins has been recognised (NÉMETH et al. 2002).

Volcaniclastic succession of Pula

The Pula crater is a north–south elongated depression, currently forming a max. 50 m deep basin (Figure 3.15). The volcanic-related rocks have been grouped into four major lithofacies on the basis of their bedding, sorting, grading and compositional characteristics. The central part of the volcanic depression is filled by finely bedded, laminated, normally graded, fine-grained volcanic silt and sandstone with angular quartz and minor (up to 20 vol.%) non-to-weakly vesicular, non-abraded tephrite to phonotephrite glass shards (facies 1 – central laminated). Such deposits are often used for palaeoclimatic reconstructions (VOS et al. 1997, ZOLITSCHKA et al. 2000, DIMITRIADIS and CRANSTON 2001, HOEK 2001). In Pula a 124,000-year periodicity has been revealed in terrestrial vegetation changes during the Late Pliocene epoch (WILLIS et al. 1999) in a diatom-dominated (Figure 3.16) maar lake (HAJÓS 1976). The normal grading and the well-bedded characteristics of these beds indicate sedimentation from turbidity currents, a common process in modern maar lakes (WALKER 1992, DROHMANN and NEGENDANK 1993, MINGRAM 1998, GOTH and SUHR 2000, KULBE et al. 2000, LEROY et al. 2000). Facies 2 consists of thicker bedded, coarse-grained lapilli tuff beds that are predominantly inverse-graded and indicate grain flow deposition (facies 2 – central juvenile-rich facies – WHITE 1992). Tephrite/phonotephrite glass shards in beds of facies 2 are weakly vesicular, microlite poor and blocky (Plate 3.16, A) suggesting formation during phreatomagmatic explosions (HEIKEN 1974). These glass shards were derived from the crater rim and are inferred to be sourced from slumping and collapse of part of the loose phreatomagmatic tephra surrounding the crater lake, as has been observed in young maar volcanoes (BÜCHEL and LORENZ 1993, DROHMANN and NEGENDANK 1993, FISHER et al. 2000, SCHARF et al. 2001). However, the source of the volcanic glass shards accumulated in the Pula maar basin is not yet fully understood. The nearby volcanic eruptions very likely contributed sediment to the Pula maar basin fill, and distal phreatomagmatic falls can also act to trigger small turbidity currents in the maar lake floor. Such processes are well-documented in young maar volcanic fields, and may be used to reconstruct recurrence and periodicity of distal explosive volcanic events (SIEBE 1986, ZOLITSCHKA et al. 1995, SHANE and SMITH 2000, SHANE and HOVERD 2002). It is planned to distinguish and fingerprint the tephra in the Pula maar as part of a research project in the near future. The coarse-grained beds often truncate underlying laminae, with the contact marked by dewatering structures, soft sediment deformation and development of dish structures (Plate 3.16, B). All these features suggest active syn-sedimentary slumping and shaking, and are interpreted as the result of debris flow and/or turbidity current emplacement from the crater rim accompanied by palaeo-earthquakes as it is suggested elsewhere (e.g. PIRRUNG et al. 2003).

The marginal area of the depression is made up of a narrow belt of phreatomagmatic lapilli tuff and tuff beds (facies 3 – tuff ring facies; ~30 m thick). This lithofacies consists of rim-type accretionary lapilli-bearing (Plate 3.16, C –

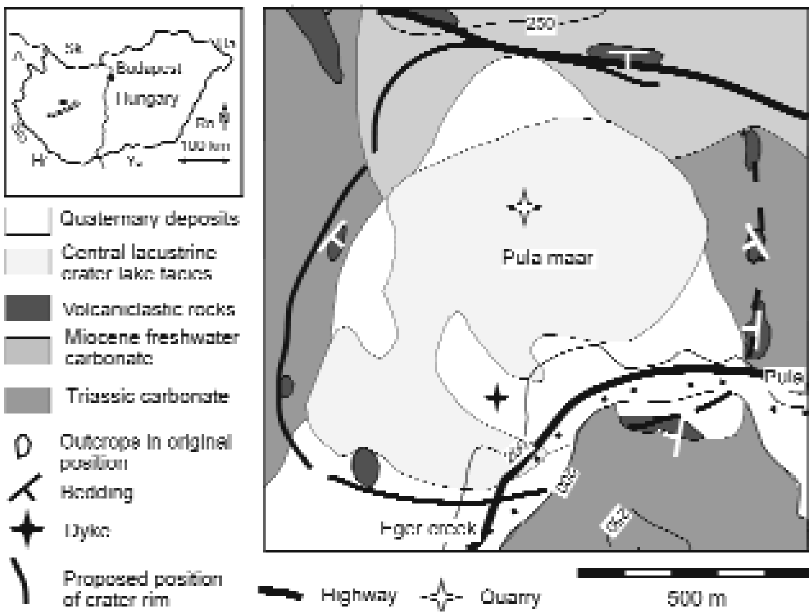


Figure 3.15. Simplified geological map of the Pula maar region

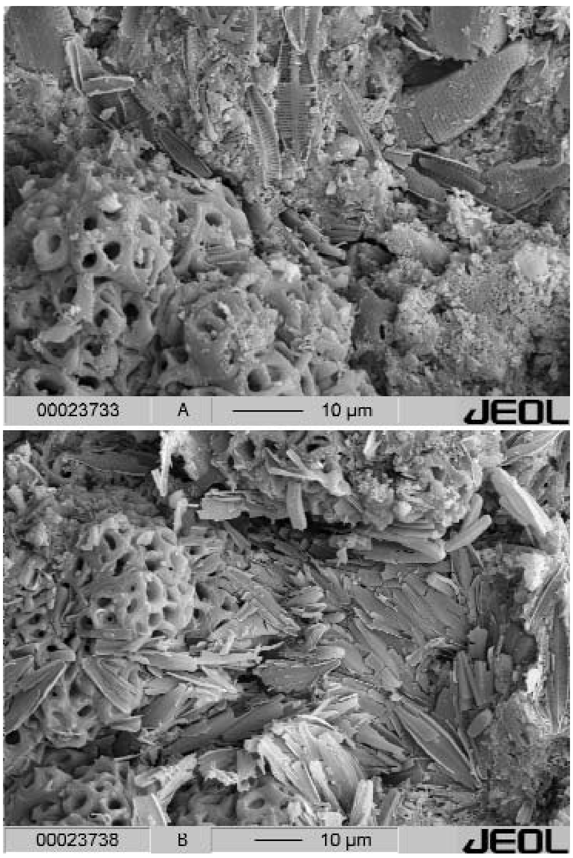


Figure 3.16.
A) Alginite from the Pula maar lake, B) Botryococcus colonies and diatom frustules on SEM images [photo courtesy of KURT GOTH (2004)]

SCHUMACHER and SCHMINCKE 1995), lithic-rich, cross- and dune bedded lapilli tuff and tuff, which are inferred to be primary of origin (BULL and CAS 2000). They dip toward the centre of the basin or are sub-horizontal. Flow indicators suggest that their source was in the centre of the depression. The fourth facies (facies 4 – volcaniclastic debris flow facies), (~20 m) is related to the marginal primary pyroclastic facies (facies 3), which dips at 20-30° towards the centre of the depression. Sedimentary features of the volcaniclastic beds of facies 4, such as

1. the presence of large (dm-scale), semi-rounded lapilli tuff fragments in the volcaniclastic beds (Plate 3.16, D),
2. a high percentage of carbonate cement,
3. a larger proportion of broken, angular phenocrysts and/or xenocrysts (mostly olivine and clinopyroxene) relative to inferred primary beds, and
4. the absence of primary origin indicators (e.g. lack of accretionary lapilli) suggest a reworked origin by debris flows, which were generated on the inner wall of the crater.

The common presence of abraded pyroclastic fragments in these beds shows that some of the pyroclastic rocks were partially consolidated and cemented prior to their disruption, however, their origin is inconclusive and either could represent

1. pre-existing pyroclastic rocks disrupted by the phreatomagmatic eruption of the Pula maar and incorporated into its tephra as lithic fragments or
2. cemented, lithified parts of the Pula maar's own tephra ring that was eroded into the maar crater. The large number of coherent lava clasts in reworked volcaniclastic beds (facies 4), their diverse shape and textural characteristics (microcrystalline to aphanitic) indicate that older lava units were disrupted by the phreatomagmatic volcanic eruption(s) of Pula and subsequently reworked by debris flows that developed on the collapsing inner wall of the growing phreatomagmatic crater.

At Pula maar sedimentological evidence indicates that monogenetic volcanism in the BBHVF had different phases as well as a significant time span (in comparison to the lifetime of a more typical monogenetic phreatomagmatic volcano – days to years), with the time between eruptions allowing solidification of early lava flows and lithification of pyroclastic units before their disruption.

Kab-hegy shield volcano and Tálodi-erdő lava flow

The largest accumulation (covering an area about 12 km across) of volcanic rocks by volume in the BBHVF is the mainly coherent basanitic lava flows that build up the Kab-hegy, a shield volcano similar in size to Rangitoto in New Zealand (ROUT et al. 1993, SPÖRLI and EASTWOOD 1997). The lava flows of the Kab-hegy region have been dated using the K/Ar method; however, the dates obtained vary between ~5 and 2.8 My (BALOGH et al. 1982, 1986), reflecting some difficulty in obtaining good measurements (BALOGH, et al. 1985, BALOGH, et al. 1996). The range in ages could point to the existence of a multiple generations of lava forming the Kab-hegy complex during repeated recurrence of volcanic activity in the region, as suggested earlier by basic geological mapping (VITÁLIS 1934). This detailed mapping of the Kab-hegy region identified soil horizons between lava units (VÖRÖS 1962, 1966), and the co-existence of basaltoid rocks forming lava flows with different textural characteristics (JUGOVICS 1971, KORPÁS 1983). Lapilli tuffs with vesicular scoriaceous lapilli have been reported, as well as pyroclastic rocks below the lava plateau of Kab-hegy (VÖRÖS 1966), however, their existence and spatial relationship with the Kab-hegy vent(s) have not been investigated yet in detail.

The Tálodi-erdő is an elliptical, flat topped hill about 3 km in its longer (north to south) axis rising about 100 metres above the surrounding landscape, located south of the Kab-hegy massif. It is capped by a coarse grained, coherent porphyric basaltoid lava flow, similar to those exposed in the northern side of the Pula maar, indicating that the maar disrupted a once much larger lava field. This is supported by the fact that basaltoid fragments similar to the rocks that make up the lava flows are widespread in the pyroclastic succession of the Pula maar, reaching metre-size. The age of the flows north of the Pula maar and the Tálodi-erdő are similar, also indicating their former connection (BALOGH et al. 1982, 1986).

Fekete-hegy maar volcanic complex and associated rocks

Introduction

The region of the Fekete-hegy is in the geometrical centre of the BBHVF (Plate 3.17, A). Vent distribution analyses of the BBHVF have revealed that the highest identified vent density is located around the Fekete-hegy. By erupted volume, the Fekete-hegy lava field, which covers significant thicknesses of phreatomagmatic pyroclastic deposits, is also among the largest in the region. The most evolved volcanic glass shards (tephriphonolite) known from the BBHVF have been found in phreatomagmatic tuffs and lapilli tuffs from this region, as well as large (dm-scale) mantle-derived xenoliths (from pyroclastic and coherent lava flow units) of great compositional diversity that have made this region a centre of geochemical research and work on models of the mantle and lithosphere in the Pannonian region (EMBEY-ISZTIN et al. 1989, Downes et al. 1992, Demény and Embey-Isztin 1998, Dobosi et al. 1998, Embey-Isztin et al. 2001, Dobosi et al. 2003, Török et al. 2003). Here a description and interpretation of pyroclastic rocks crop out in the vicinity of the Fekete-hegy will be presented from the northern Kál Basin (e.g. Szentbékkálla) to Kapolcs (Plate 3.17).

Mafic phreatomagmatic pyroclastic flow deposits at Szentbékkálla

Pyroclastic succession near Szentbékkálla

Mapping of the area north of Szentbékkálla village (Plate 3.17, A) reveals small-volume pyroclastic flow deposits inferred to be a result of phreatomagmatic explosive eruptions, previously referred to as hydroclastic flow deposits to describe their unusual textural characteristics (Németh and Martin 1999b).

The massive, unsorted coarse grained lapilli tuff beds alternate with cross-bedded, matrix rich, block bearing lapilli tuff beds, and mantle bedded tuff layers (Plate 3.17, B). The main body of the pyroclastic sequences consists of grey, massive, compact lapilli tuff beds (Plate 3.17, C). There is neither any evidence of grading or well-developed sedimentary structures nor welding in this unit. The lapilli tuff contains a high proportion of semi-rounded to rounded gravel-like ultramafic xenoliths, broken olivines and pyroxene megacrysts without any systematic accumulation pattern. The beds contain a high proportion of fragments derived from the entire known thickness of the pre-volcanic rock units. The basal massive lapilli tuff unit has a non-erosional contact with the underlying gravel beds. The contact zone of the lapilli tuff contains lithic clasts picked up from the gravel beds (Plate 3.18, A). The main body of the massive lapilli tuff unit contains several well-developed, metres long sub-vertical and cm-to-dm wide curvilinear segregation pipes, which are filled by lithic lapilli (Plate 3.18, B).

The juvenile fragments of the lapilli tuffs and tuffs from Szentbékkálla are usually micro-vesicular and slightly palagonitized (Plate 3.18, C). Their composition, according to electron microprobe analyses, range from tephrite through phono-tephrite to tephriphonolite (Németh and Martin 1999b). Small altered, light coloured glass shards with 62–69 wt.% SiO_2 (88-95 wt.% total) show dacite/trachydacite and basaltic andesite compositions (Németh and Martin 1999b). These glass shards are inferred to have been picked up from early explosive volcanic products (Németh and Martin 1999b).

The pyroclastic rocks near Szent-békkálla have been divided into two facies [Figure 3.17, 1. a valley filling facies (PFVF) and 2. overbank facies (PFOB)].

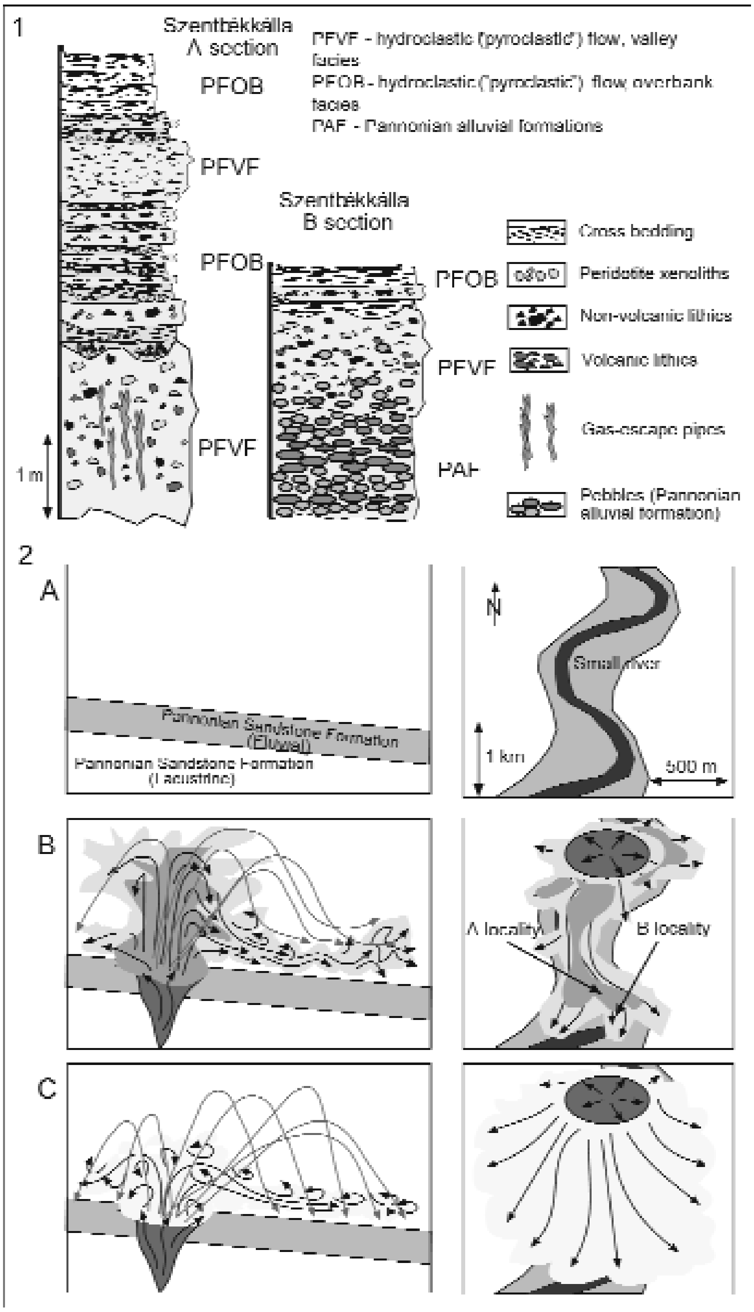


Figure 3.17. Simplified stratigraphic columns (1) from the Szentbékkálla section showing vertical relationship between massive and dune-bedded pyroclastic facies that largely correspond with valley filling and overbank facies of a phreatomagmatic pyroclastic flow (Németh and Martin 1999). Localities A and B are shown on the map of Plate 3.17, A. A simplified eruption model for the mafic pyroclastic flows of Szentbékkálla is given on the part 2 after Németh and Martin (1999)

The lower part of the Szentbékkálla open-air theatre outcrop shows a minimum 2.5 m thick succession of grey, polymict volcaniclastic breccia and block bearing lapilli tuff (PFVF). The lower part of the sequence is massive but in higher stratigraphic positions faint clast alignments give a crudely bedded impression. The massive volcaniclastic beds are compact and locally show crude jointing. The matrix of the lithofacies comprises siliciclastic sand or silt and vitric shards. Large clasts are dominantly lithics (min. 85 vol.% of total) with a wide range of lithologies from the pre-volcanic strata. The most common lithics are Mesozoic carbonates (limestones, dolomites, and marls, up to 70 vol.% of the total large "accidental lithics" (term after FISHER and SCHMINCKE 1984), up to 25 cm in diameter with an average of 2–5 cm. There is also a small amount of Palaeozoic schist and quartzite (15 vol.% of total large accidental lithics, up to 5 cm in diameter, average 0.5 cm), and occasionally larger Pannonian sandstone fragments (5 vol.% of total large accidental lithics, up to 35 cm in diameter, average 2 cm). Clasts are not oriented or stretched. An *Echinoidea* fossil from the pre-volcanic Triassic beds was found without any thermal effect on its rim. Crystalline igneous rock fragments differ from known basaltic lava rocks occurring at the surface in this area and are probably disrupted fragments from the sub-volcanic region. There are many clasts picked up from the underlying pebble beds at the Szentbékkálla "B" locality on the bottom of the flow body. This pebble concentration decreases upward in the section but is still present around 3–4 m above the base at the Szentbékkálla "B" locality. In general there is no sorting or gradational texture in the entire massive unit. The large clasts did not cause any impact or scour fill structures.

The upper part of the Szentbékkálla open air theatre is made up of a crudely bedded, cross bedded lithofacies (PVOB) and is compositionally similar to the lower PFVF lithofacies, but the smaller maximum clast size and the bedding distinguishes this unit from the basal one. The relative ratio among the lithics is also different compared to the lower unit. The Mesozoic limestone and dolomite clasts (up to 50 cm in diameter) are more abundant and larger than in the PFVF. Schist fragments and Permian red sandstone fragments are less common and their grain size is smaller (up to 5 cm in diameter). The carbonate clasts are angular, broken and have less thermal effects on their surface. Large Pannonian sandstone fragments are not so frequent as in the PFVF facies but quartzofeldspathic grains derived from the Pannonian sandstone in the matrix are more common. Scoriaceous particle concentration zones are common behind the large, mainly angular clasts. Trains of scoriaceous clasts, forming 1–5 cm thick, 50–100 cm long lenses with upward concave bases and slightly upward convex tops, are also common. The scoriaceous fragments are less than 1 cm in diameter.

The present ridges nearby Szentbékkálla are interpreted to represent former river-valleys occupied by the horizontally moving pyroclastic high density currents that were triggered by phreatomagmatic explosive eruption(s) (NÉMETH and MARTIN 1999b). North to south transport directions are indicated by horizontal transport features (dune, antidune, scour fillings – NÉMETH and MARTIN 1999b). The presence of gas segregation pipes in the unit suggests that this unit represents a distal facies (BOGAARD and SCHMINCKE 1984, FREUNDT and SCHMINCKE 1985a, b, 1986) of a high concentration pyroclastic density current (NÉMETH and MARTIN 1999b). However, the low juvenile to lithic clast ratio (compared to pyroclastic flows) plus the predominantly chilled glassy volcanic nature of the lithic clasts, as well as that of the volcanic glass shards leads to a conclusion that the pyroclastic rocks near Szentbékkálla resulted from an eruption of a "hydroclastic flow" (NÉMETH and MARTIN 1999b). The term "hydroclastic flow" is designed to emphasize the difference between real pyroclastic flows and flows generated by the collapse of the marginal parts of phreatomagmatic eruption clouds (NÉMETH and MARTIN 1999b). Recent developments in volcanic terminology, however, suggest that the term "hydroclast" is misleading, because it literally means clasts composed of free (condensed) water. Such droplets perhaps exist during magma-water triggered explosive eruptions in the eruption cloud, as well as the horizontal moving currents, and term "hydroclast" should refer to such "clasts" only. According to the original meaning of "pyroclast" from FISHER and SCHMINCKE (1984, p. 89) "*fragments are produced by many processes connected with volcanic eruptions. They are particles expelled through volcanic vents without reference to the causes of eruption or origin of the particles*". In this sense, 2 types of pyroclasts can be defined:

1. juvenile clasts directly derived from the magma involved in explosion, which if they are a result of magma-water interaction may be chilled glassy pyroclasts (e.g. glass shards); and

2. lithic clasts which may have been disrupted from pre-volcanic rock units. Because in this argument lithic clasts are also pyroclasts since they are expelled by explosive processes directly related to the volcanic explosion itself, the pyroclastic rock unit identified at Szentbékkálla is best reconstructed as a result of pyroclastic flow eruption through a phreatomagmatic vent. The term *phreatomagmatic pyroclastic flow* has been introduced recently to describe processes inferred from very similar deposits/rock units as the pyroclastic succession identified near Szentbékkálla. Phreatomagmatic pyroclastic flows have been recently interpreted as major transporting and depositing agents of tephra, e.g. in Central Italy (DE RITA et al. 2002, GIORDANO et al. 2002, WATKINS et al. 2002).

Eruption mechanism

On the basis of the observed facies characteristics of the pyroclastic rock units (NÉMETH and MARTIN 1999b), the following eruptive history is given for the volcanic history of the Szentbékkálla area.

A) Stream valley(s) cut into former Pannonian lacustrine sediment were filled by gravelly, fluvial beds (probably north to south transportation).

B) Initial phreatomagmatic explosions occurred at shallow levels due to the water content of stream valley sediments (sideromelane clasts, large amount of lithics derived from subsurface strata). The explosion locus (due to the drying process of a porous media aquifer) migrated downward at high speed, following the model of LORENZ (1986). The explosion locus probably reached the fracture controlled aquifer quickly (presence of the large number of Mesozoic carbonate fragments), where interaction between magma and abundant karst water could have fuelled the phreatomagmatic processes. The magma supply was probably steady (even increasing) producing further efficient phreatomagmatic interactions between magma and (at this stage) probably karst water (Tihany-type maar volcano - NÉMETH et al. 2001). Subsequent explosions produced a high particle-concentration eruption column as a result of the continuous (even increasing) input of disrupted material, which became heavy. Thus the column margin collapsed and produced small-volume pyroclastic flow units, which travelled downward following the palaeotopography (north to south transportation direction according to the PHOB lithofacies features). During flow, water from the streams was ingested into the body of the flow and clastic material (pebbles) was entrained from the base of the flow.

C) With decreasing magmatic supply (or sudden cut off of the water supply) the efficiency of the phreatomagmatic process decreased. At this stage dry base surge and fall-out processes occurred (normal base surge and fall out beds in the top of PHOB lithofacies at Szentbékkálla).

Fekete-hegy maar volcanic complex — Nested phreatomagmatic volcanic system, Fekete-hegy

Different pyroclastic rock outcrops at Fekete-hegy (Plate 3.17, A) may represent more distal or proximal sites in relation to their volcanic source, depending on the erosional stage of the volcanic butte (MARTIN et al. 2002). In the basal pyroclastic units, large basaltic bombs, lherzolite nodules (<70 cm) or blocks from the basement (<40 cm), as well as large (dm-scale) flattened and softly deformed unconsolidated sediment rags often occur in only crudely stratified or massive beds. Collectively, these rocks represent a near vent facies. Some coarse-grained lapilli tuff beds contain fragments of well-preserved tree trunks (cm-scale) indicating a forested area surrounded the vents (MARTIN et al. 2002). Other pyroclastic deposits are very thinly bedded and cross-bedded, with cross-beds dipping at low-angles (<10°) and showing dune structures of low amplitude (cm-scale) and long wave length (m-scale — MARTIN et al. 2002). Varying contents (~25–90 vol.%) of lithic clast types, as well as different kinds of pyroclastic deposits in respect of bedding characteristics, grain size or juvenile to lithic clast ratio (depending on more or less intensive fragmentation, water content, depositional mechanism and other primary factors of the system) show a complex eruptive history in an area of ~15 km^2 (Figure 3.18).

The basal pyroclastic deposits of the nested maar system of Fekete-hegy were formed by pyroclastic density currents (base surges), fall-out, and volcaniclastic mass flows generated by synvolcanic reworking, as inferred from their grain-size, bedding characteristics, volcanic textures and km-scale field relationships (e.g. DRUITT 1998, WHITE and SCHMINCKE 1999, DELLINO 2000, DELLINO and LA VOLPE 2000). The presence of a large amount (up to 90 vol.%) of lithic fragments in the pyroclastic rocks of Fekete-hegy, with the majority derived from the immediate underlying fluvio-lacustrine, Late Miocene (Pannonian) sedimentary units (Plate 3.19, A), indicate that interaction of the ascending magma with water occurred in water-saturated Late Miocene shallow marine to fluvio-lacustrine siliciclastic sediments, as well as with water in aquifers which have been part of a wide-spread and multilevel karst system in Mesozoic carbonate rocks (e.g. LORENZ 1986, 2000b, GEVREK and KAZANCI 2000, NÉMETH et al. 2000b). In this respect, this nested maar complex is similar to the other well-characterised nested maar system at BBHVF, the Tihany Maar Volcanic Complex (NÉMETH et al. 2001).

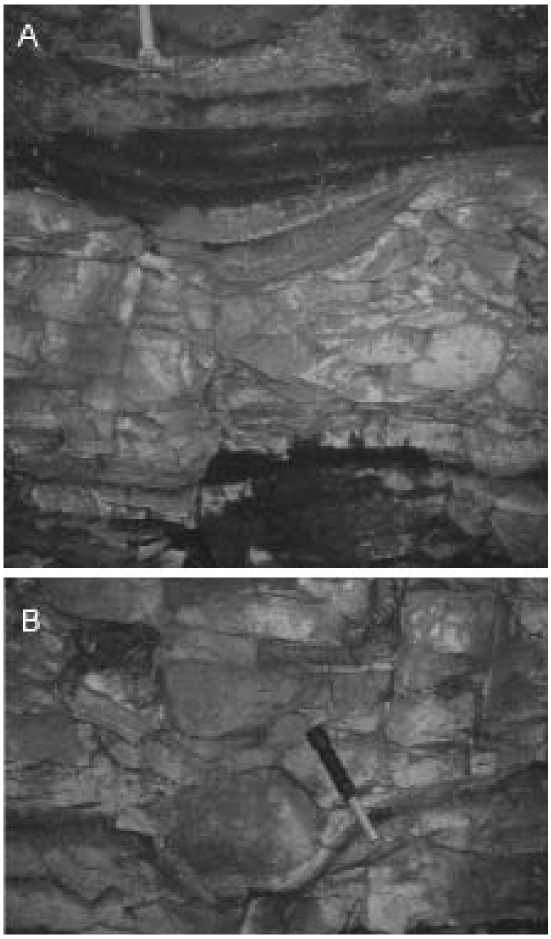


Figure 3.18. Alternating tuff and lapilli tuff (A) of phreatomagmatic origin from the Fekete-hegy southern flank (Vaskapu-árok)
The succession is commonly built up by finely bedded accretionary lapilli rich units that alternate with coarser grained dune bedded lapilli tuff beds (A). The pyroclastic succession is commonly truncated by large impact sags (B) caused by country rocks from the entire known pre-volcanic strata of the region, including mantle derived, angular shaped lherzolite blocks up to 75 cm in diameter (ANDREAS AUER pers. com. 2002)

The final phase of activity in the maar/tuff ring complex at Fekete-hegy is represented by effusive eruptions of lava flows and lava lakes that fill and cover the pyroclastic deposits in an area of about 10 km^2. The change in eruptive style presumably reflects the termination of water supply in the basement, or sealing of conduits by chilled melt that prevented influx of external water to the system as it is suggested from other similar volcanoes elswhere (e.g. HOUGHTON and SCHMINCKE 1986, 1989, WHITE 1989, WHITE 1991a, HOUGHTON et al. 1999). The latest pyroclastic deposits are red, dark brown scoria agglomerate and tuff breccia interpreted on the basis of field relationships, bedding characteristics and scoria clast flattening to be the remnants of at least two scoria cones. Eruptions at both scoria cones also started with a short period of phreatomagmatic activity, as it is indicated by a thin veneer of lithic clast-rich basal units underlying the scoriaceous volcanic piles.

Recently obtained new K/Ar age data (BALOGH pers. com. 2004) support the conclusion that the volcanic system at Fekete-hegy was long lived. Samples from the capping dark, fresh, aphanitic plateau lava flows (Fekete-hegy, Vaskapu) give an age of a 3.98±0.34 My. Samples from the capping scoria cone and associated young lava field (Fekete-hegy, Boncsos-tető) gave a younger age of 3.36±0,20 My, consistent with its higher stratigraphic position in comparison to the major lava plateau of the Fekete-hegy.

Fekete-hegy is interpreted to be an erosional remnant of a phreatomagmatic volcanic complex (Fekete-hegy Maar Volcanic Complex — FMVC) made up of several closely spaced vents, and predates the volcanoes developed west of it. The more-or-less north-south trending chain of identified volcanic edifices suggests that the Fekete-hegy maar complex either developed in a north–south valley system and/or is associated with pre-existing structural elements with the same orientation. Such structural patterns in the region have recently been recognised on the basis of digital terrain models and other image analysis methods (JORDÁN et al. 2003). The development of such a long (10 km-scale) phreatomagmatic vent chain in the geometrical centre of the BBHVF is significant. Its existence points toward three important conclusions;

1. long structural elements and/or valley systems must have existed during the initiation of eruptions in the Fekete-hegy volcanic complex,

2. these valley systems respectively underlying hydrologically active zones of structural weekness that provided substantial water sources to sustain phreatomagmatic volcanism, producing a large volume of phreatomagmatic tephra and

3. that in the final stage of the eruption of the Fekete-hegy vent system pure magmatic fragmentation of the alkali basaltic magma produced Strombolian-type eruptions building up at least two scoria cones and associated lava flows, with lava flows confined by relief between the rims of tuff rings and the palaeo-valleys.

Kereki-hegy diatreme

Kereki-hegy (171 m) is located 1 km south-east from Mindszentkálla and stands as a small, ~150 m north to south elongated hill about 60 metres above the floor of the Kál Basin. Lower Triassic carbonate formations crop out on the surface nearby (BUDAI and CSILLAG 1998, BUDAI et al. 1999). In the immediate vicinity of the hill, a few metres thick Neogene siliciclastic units have been mapped (BUDAI et al. 1999); however, their existence is subject of a debate. In spite of that, former geological mapping described columnar jointed basalt in this location (VITÁLIS 1911), although new mapping attempts have not been able to confirm this information (BUDAI et al. 1999, NÉMETH et al. 2003). The pyroclastic beds are steeply bedded (60°) and dip toward the east, which is significantly different from other known tectonic trends in the region as well as from sub-horizontal bedding of the immediate pre-volcanic rock units that the diatreme cuts through (BUDAI et al. 1999). The pyroclastic rocks are well bedded and have scour fillings and slight undulations on upper and lower bedding surfaces; however, beds are persistent, and traceable on a metre scale. At the base of the hill, large blocks of massive, structureless lapilli tuff fragments have been recovered. The pyroclastic rocks of this location are rich in volcanic glass shards (Plate 3.19, B, C) of tephrite composition (NÉMETH et al. 2003), that are elongate, fluidal to blocky in shape and moderately microvesicular. In combination, these features suggest that magma–water interaction was the dominant fragmentation style which formed them (NÉMETH et al. 2003). From a textural and compositional point of view the bedded and massive parts of the pyroclastic rocks do not differ from each other (NÉMETH et al. 2003). Kereki-hegy is interpreted as a deeply eroded diatreme, in which the deep levels of the diatreme are exposed. The presence of fragments is inferred to have been derived from the immediate pre-volcanic siliciclastic units suggest that the Neogene sedimentary cover was still intact during the eruption of the Kereki-hegy (NÉMETH et al. 2003).

Harasztos-hegy diatreme

Harasztos-hegy near Kékkút village (Plate 3.17, A) is located in the western margin of the Kál Basin and comprises four small hills each of them less than 100 m in diameter (the highest being 212 m a.s.l.). This group of hills is about 80 metres above the surrounding basin. The basement is Permian to Lower Triassic rocks, which are blanketed by the Neogene siliciclastic succession (50-70 m thick). A geomagnetic study undertaken to delimit the lateral extent of a basalt dyke revealed that the exposed section of the dyke, which forms a ridge a few tens of metres long, pinches out quickly. This dyke is inferred to be a feeder dyke of the former diatreme, now deeply eroded (Bence et al. 1988). The dyke shows rosette-like joints, supporting its origin as a feeder structure. The 212 m high central hill is composed of pyroclastic rocks that are weakly bedded to massive and rich in juvenile lapilli with trachytic texture (Plate 3.19, D), that are brown, yellow or red in colour (Németh et al. 2003). The pyroclastic rocks are intruded by a basaltoid dyke that has an irregular contact with its host rocks. The pyroclastic rocks are thermally altered, red, and slightly welded close to the dyke. The texture of the host sediment quickly changes from normal lapilli tuff to lapilli stone within a few metres of the dyke, while the lapilli tuff further away from the dyke is grey, unsorted, and weakly bedded with a quartzofeldspathic sand/silt-rich matrix. Lapilli and ash size carbonate fragments, as well as rocks derived from Palaeozoic units are common as lithic fragments; however, their total volume is not more than 30%.

The textural characteristics of the pyroclastic rocks at Harasztos-hegy suggest some degree of magma–water interaction during fragmentation of the rising basanitoid melt (Németh et al. 2003). The vent zone of this phreatomagmatic volcano has been invaded subsequently by dykes that may have fed scoria cones at the surface.

Eroded phreatomagmatic volcanoes in the Tapolca Basin

Eroded small-volume intraplate volcanoes form a cluster in the western part of the BBHVF (Budai et al. 1999, Németh and Csillag 1999), i.e. in the Tapolca Basin (Figure 3.19). Szigliget is a triple hill on the southern margin of the basin that consists of hills covered by pyroclastic rocks that forms a small peninsula on the northern shoreline of Lake Balaton (Figure 3.19). In the centre of the Tapolca Basin, two large buttes form the two highest areas (Badacsony and Szent György-hegy), both covered by columnar-jointed basanite and spattery scoria units. The Tapolca Basin generally has a stratigraphy typical of the rest of the BBHVF: Silurian schist (very low-grade metamorphised psammitic, pelitic beds Budai et al. 1999, Császár and Lelkesné-Felvári 1999) and Permian red sandstone (continental alluvial facies Budai et al. 1999, Majoros 1999), overlain by Mesozoic predominantly carbonate sequences (Budai and Vörös 1992, Budai and Csillag 1998, 1999, Budai et al. 1999, Haas and Budai 1999). These beds are covered by a thick sequence of Neogene gravels, sandstones and mudstones deposited in the Late Miocene Pannonian Lake and related fluvial systems (Budai et al. 1999). Along the axis of the Tapolca Basin, the basement rocks are in progressively deeper positions toward the west following a series of north–south trending normal fault-displaced blocks that bound the elongate Tapolca Basin (Budai et al. 1999, Dudko 1999). It is estimated that the Ordovician/Silurian schist beds are few hundreds of metres below the surface (Budai et al. 1999, Dudko 1999).

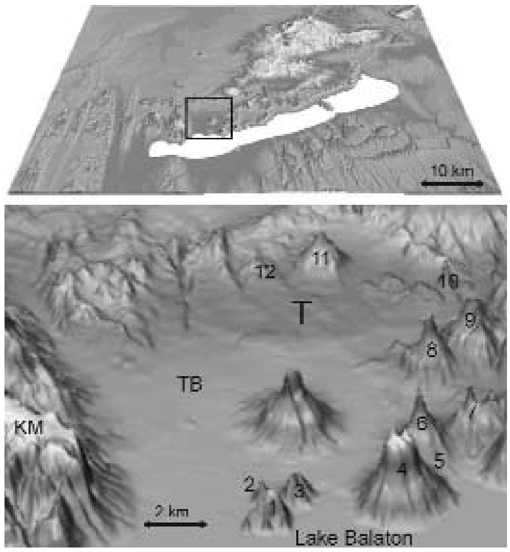


Figure 3.19. DTM model of the Tabolca Basin, showing eroded remnants of small-volume intraplate volcanoes forming small hills up to 350 metres above the basin floor

1 – Kamon-kő (Szigliget), 2 – Vár-hegy (Szigliget), 3 – Antall-hegy (Szigliget), 4 – Badacsony, 5 – Hármas-hegy, 6 – Gulács, 7 – Tóti-hegy group, 8 – Csobánc, 9 – Hajagos, 10 – Hegyesd, 11 – Haláp, 12 – Véndek-hegy, T = City of Tapolca, TB = Tapolca Basin, KM = Keszthely Mts

Hármas-hegy diatreme

The Hármas-hegy (Triple hill) is located in the south-eastern margin of the Tapolca Basin, just between two lava capped buttes (Badacsony and Gulács – Figure 3.19). It is a 180–210 m high elongated chain of 550 m long and 80–100 m wide hills that are 40–60 metres above the basement (Figure 3.20). The Hármas-hegy is entirely composed of pyroclastic rocks. The pyroclastic rocks are completely surrounded by the pre-volcanic Neogene siliciclastic rock beds (Somló Formation – Budai et al. 1999). The exposed pyroclastic rocks are massive to weakly stratified.

The pyroclastic rocks studied at the Hármas-hegy (Németh et al. 2003) are uniform from a volcanic textural point of view (Plate 3.20, A). The matrix of the pyroclastic rocks from this locality is rich in mud, silt and minerals derived

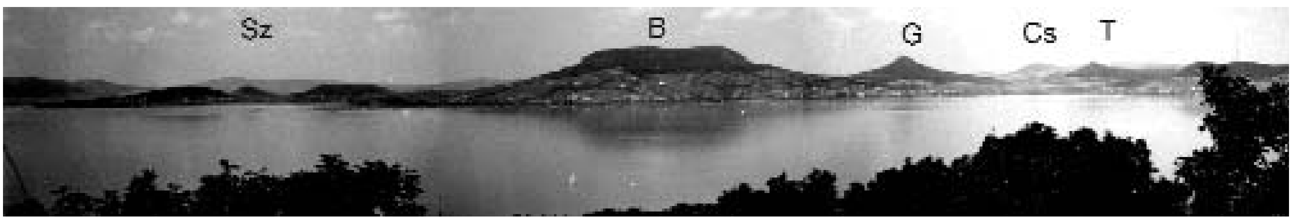


Figure 3.20. View toward Szigliget (Sz), Badacsony (B), Gulács (G), Csobánc (Cs) and Tóti-hegy group (T) from the southern shoreline of the Lake Balaton

from rocks characteristic of the Neogene succession in this region. Phenocrysts and xenocrysts are common in the matrix as broken lapilli (NÉMETH et al. 2003). The pyroclastic rocks are rich in volcanic glass shards, of tephritic to phonotephritic composition (NÉMETH et al. 2003). These glass shards show variable vesicularity, vesicle shape, and proportion of microlites (Plate 3.20, B – NÉMETH et al. 2003). The glass shards are both blocky and fluidal, and are often mud coated (NÉMETH et al. 2003). Palagonitisation is moderate, and is especially pronounced on the rim of the shards (NÉMETH et al. 2003). Dark, tachylite glass shards often exhibit entrapped siliciclastic and/or volcaniclastic mud/ash (Plate 3.20, C) indicating a premixing and possible recycling of pyroclasts during repeated eruptions through the vent of a "wet" volcano (HOUGHTON and SMITH 1993, WHITE 1996).

The Hármas-hegy is interpreted to be an individual erosion remnant of a diatreme, which was the conduit of a maar volcano (NÉMETH et al. 2003).

Tóti-hegy and Gulács diatreme

Tóti-hegy and Gulács are two volcanic erosional remnants with similar architecture (BUDAI et al. 1999). Both hills are about a km in diameter capped by basanite lava and form a slightly asymmetric butte, steeper in the southern side, with low angle flanks toward north (Figure 3.19).

Gulács is 398 m high, standing about 250 m above the surrounding landscape, similar to Tóti-hegy which is slightly lower (346 m). Both erosional remnants are located at the western margin of the Tapolca Basin (BUDAI et al. 1999), cutting through a thick succession of Neogene sediments (BUDAI et al. 1999). Palaeozoic rock formations crop out along the southern margin of the Tóti-hegy, and form an elongate east-west trending ridge, reaching 311 m elevation. Silt and sand of the Neogene sedimentary units crop out just below both hills (the Gulács and the Tóti-hegy), and are traceable up to about 250 metres elevation. These units are sub-horizontal and are apparently undisturbed (BUDAI and CSILLAG 1999).

At Tóti-hegy, rosette-like columnar-jointed basanite lava overlies a massive succession of lapilli tuffs. The basanite at Tóti-hegy appears to form two units. A lower unit, crops out in the north-western side of the hill about at 200 metres elevation, and is a planar, vertically and/or platy jointed relatively coarse grained, porphyric basanite that is inferred to be intrusive in the Neogene succession. The age of this rock has been determined by the $^{39}Ar/^{40}Ar$ method, and the rock has an age of 4.74±0.03 My (WIJBRANS et al. 2004), which is similar to the previous K/Ar age estimate of 4.71±0.34 My (BORSY et al. 1986). With this age, the Tóti-hegy belongs to the older volcanic phase of the BBHVF. Due to the poor outcrop no further textural characteristics of this rocks have been described. The pyroclastic rocks are rich in volcanic glass shards and are similar to lapilli tuffs from nearby phreatomagmatic vent remnants. Around Tóti-hegy a chain of small hills have been identified forming a NE–SW trending line, where pyroclastic rocks crop out. All the hills are heavily covered by vegetation, and the rocks are poorly exposed. The significance of this region is, that each of the hills is composed of pyroclastic rocks that are rich in volcanic glass shards, suggesting a phreatomagmatic origin. Moreover, pyroclastic rocks from the region (e.g. Sabar) are rich in exotic nodules (up to dm size) derived from mantle or lower crustal regions. A preliminary reconstruction of the region can be drawn, that interprets these hills as the erosional remnants of small diatremes of individual phreatomagmatic volcanoes, very similar to the hills of Szigliget (see later).

Gulács is just east of the Tóti-hegy but very poorly exposed. The walls of a former quarry in the hillside are now heavily vegetated and the quarry walls have become recultivated. Lava flows exposed in the former quarry comprise a thick unit (few tens of metres) of compound lava flow units, each showing vertical columnar and/or platy joints. The lava flow first crops out at about 220 metres, and seems to show similar textures to the basal flow units of the Tóti-hegy. Above these dissected lava outcrops, a massive succession of thick basanite lava crops out, which is apparently embedded in the pre-volcanic Neogene rock units. The top of the Gulács is formed by a rosette-like columnar-jointed basanite. Between the columns coarse grained lapilli tuff and welded scoriaceous, massive pyroclastic rocks can be recognised, indicative of lava fountain origin. Pyroclastic remnants can only be found in debris flanks as small (dm-size) pieces of lapilli tuff that is rich in volcanic glass shards and epiclastic mud- and siltstone fragments. A small number of Permian red sandstone fragments and highly vesicular, red to black scoriaceous lapilli are also present. These lapilli tuffs suggest a partly phreatomagmatic origin for the Gulács, which may represent the eroded diatreme of a phreatomagmatic volcano, that was subsequently invaded by basanite feeder dykes. These feeder dykes have acted to armour the Gulács and reduce the speed of erosion, thus preserving the complex.

Badacsony maar/tuff ring

Badacsony, one of the largest lava capped buttes in the BBHVF (Figure 3.19) is made up of thick (>50 m) black, strongly chilled, aphanitic basanite lava overlying a coarse grained, unsorted yellow lapilli tuff. Despite Badacsony being among the volumetrically largest volcanic remnants of the BBHVF with a current elevation of 438 m and a ~1 km in diameter lava cap, little has been published concerning its geological framework (CSERNY et al. 1981) or eruptive history (HOFMANN 1875–1878, LÓCZY SEN. 1913, 1920).

Pyroclastic rocks have been sparsely reported from the region earlier, and earlier work mainly focused on an elongate outcrop of lapilli tuff in the northern margin of the area (Hármas-hegy. – HOFMANN 1875-1878, NÉMETH et al. 2003). Here, recent observations of the pyroclastic successions of Badacsony are summarised, based on an extended abstract (MARTIN and NÉMETH 2002).

The lapilli tuff from Badacsony crops out in a thickness of approximately 250–300 m, and consists of finely dispersed quartz or quartzofeldspathic sandstone, xenocrysts of olivine and pyroxene, as well as blocky, weakly to highly vesicular, microlite poor sideromelane glass shards (tephrite, phonotephrite – Plate 3.20, D, E). In combination, these features indicate phreatomagmatic fragmentation, near-surface vesiculation and excavation of pre-volcanic country rocks. The pyroclastic beds at Badacsony are poorly exposed, and are covered by a thick debris flank of rock falls from the capping lava unit. In dissected outcrops (metre-scale) pyroclastic beds are exposed in a collar-like distribution and exhibit gentle dips (~10°) toward the centre of the butte, at each locality. In exposures along the southern and the north-western margins of the butte, the pyroclastic rocks are slightly bedded, with cm-to-dm thick beds that are poorly sorted and non-graded.

The 50 m thick coherent lava forming the plateau on the top of the Badacsony butte has been dated by the K/Ar method repeatedly, and has an age of about 3.5 My (BORSY et al. 1986). On the north-western and eastern side of the butte, two large quarries into the lava cap of Badacsony show that the lava has irregular lower contacts with the pyroclastic units, commonly showing tumuli structures (bubble-like features that are similar to those described at Hajagos earlier). The tumuli enclose highly vesicular scoriaceous lava spatter clasts, with vesicles filled by clay, calcite or quartzofeldspatic fragments, as well as strongly palagonitized, often red blocky volcanic glass shards (Plate 3.20, F). The irregular shape of the tumuli and their irregular geographical distribution indicate that they formed when the basanite lava came into contact with wet unconsolidated tephra along the inner tuff ring wall (MARTIN and NÉMETH 2002). The capping lava units of Badacsony are made up of multiple flows and at least two major flow units have been identified. Between these major lava flow units a thin sedimentary veneer of volcaniclastic origin is present (MARTIN and NÉMETH 2002). The tumuli are inferred to have been associated with these thin inter-lava flow sedimentary veneers and suggest a short-lived episode of lava emplacement into the wet, water-filled Badacsony crater. The presence of irregular bubble-shaped, clay-rich vesicular zones (tumuli) in the dense, coherent lava body of the lava lake, which developed over initial lava units, as well as parts of the tuff ring, suggest that lava emplacement may have occurred intermittently during the formation of the Badacsony tuff ring, allowing time for water to fill the crater and/or for some wet volcaniclastic deposits to accumulate on top of the first lava flow. Partial filling of the crater by water was probably accomplished by water inflow from the groundwater table, which must have been relatively high to allow rapid filling from a porous media aquifer like the Neogene immediate pre-volcanic silt and sand beds at Badacsony.

The topmost structure of the Badacsony is built up by a semicircular feature that is open toward the north. This rim-like structure consists of lava spatter invaded by rosette-like columnar jointed basanite in the western side, and a mound-like red scoria-rich unit in the northern side (BUDAI et al. 1999, NÉMETH and CSILLAG 1999). This later structure forms the highest point of the Badacsony today.

Szigliget diatremes

Szigliget is a small peninsula in the southern end of the Tapolca Basin which forms three major group of hills build up predominantly by pyroclastic rocks. The largest goup of hills is about 800 m across. The smalles pyroclastic hill is about 150 m across. The peninsula was an island during high stands of Lake Balaton during its 17,000 year history (CSERNY 1993, CSERNY and NAGY-BODOR 2000, TULLNER and CSERNY 2003). The youngest pre-volcanic rocks at Szigliget belong to the Neogene siliciclastic formations known from other parts of the BBHVF, and are traceable to an elevation of about 175–200 m (BUDAI et al. 1999); however, the hills surrounding Szigliget are built up Quaternary deposits derived from Lake Balaton (TULLNER and CSERNY 2003). Bedding in the pre-volcanic Neogene sequences is sub-horizontal, in contrast to the often steep bedding in the pyroclastic successions making up the three hills, which gen-

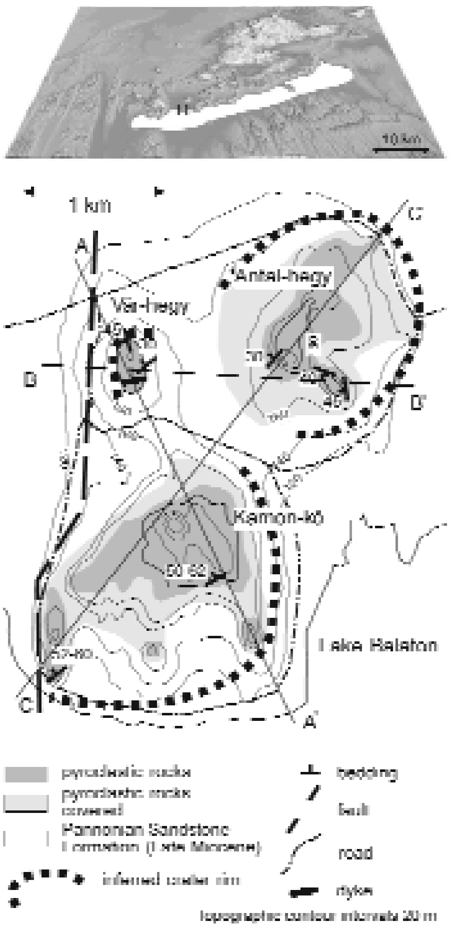


Figure 3.21. Simplified geological map of the Szigliget Peninsula

erally dips toward the north-westward (Figure 3.21). Beds in each hillside show a similar north-westward dip direction and similar textural and compositional characteristics, suggestive of a complex but closely related volcanic system in the area (Figure 3.21 and Plate 3.2, A). The Szigliget volcanic rocks, based on K/Ar age determinations of a rosette-like columnar jointed coherent lava unit inferred to be a dyke that intrudes the pyroclastic succession of the Vár-hegy, give an isochron age of 3.30±0.28 My (BORSY et al. 1986). New $^{39}Ar/^{40}Ar$ age datings give ages of 4.33±0.18 My for the lava flow and 4.08±0.05 My for cauliflower bombs from the Vár-hegy pyroclastic succession (WIJBRANS et al. 2004), which highlights some ambiguities in age determinations for these rocks.

Pyroclastic succession of Szigliget

The pyroclastic rocks of Szigliget have been grouped into three units according to their textural, compositional and stratigraphic characteristics (NÉMETH et al. 2000a).

Unit 1 crops out in the southern side of the study area. It consists of coarse-grained, matrix-supported massive to weakly bedded, lithic-rich, block-bearing lapilli tuffs/tuff breccias rich in deep-seated lithic and lherzolite clasts. Beds of this unit crop out in the southern part of the study area, close to the recent shoreline of Lake Balaton. These pyroclastic rocks are in the lowest topographic position in the BBHVF, and their stratigraphic position has been subject of debate for a long time (JUGOVICS 1969, BORSY et al. 1986). There are no exposed contacts between other units or the pre-volcanic strata. Unit 1 consists of thickly bedded, massive, unsorted, matrix-supported lapilli tuff and/or tuff breccia. It is very rich in lithic clasts, especially large lherzolite and amphibolite clasts (EMBEY-ISZTIN 1976). The occurrence of unit 1 coincides with a large negative gravity anomaly in the area.

Unit 2 crops out in the southern and north-easth hilltops and makes up the volumetrically largest amount of pyroclastic rocks at Szigliget. It consists of coarse-grained, unsorted lithic-rich, normally graded, bedded, vitric lapilli tuff and tuff beds (Plate 3.2, B). Bedding surface is regularly sharp, impact sags are just occasionally present and they are usually shallow and symmetric. Deep-seated lithic clasts are common with an average size of a few cm in diameter (Plate 3.21, C), but large lherzolite fragments are relatively rare. The lapilli tuff is occasionally clast-supported and calcite-cemented; in contrast, the tuff beds are well-bedded and commonly undulate or show cross-bedding with low amplitude (few cm-to-dm) and long wavelength (metres) dunes. The matrix of this unit is rich in fragments of Neogene sand, silt and mud. Large fragments of Neogene sandstone often have thermally affected rims (i.e. radial cracks). Juvenile clasts are weakly to highly vesicular, comprising weakly to moderately microcrystalline sideromelane ash and lapilli of tephritic or phono-tephritic composition.

Unit 3 crops out in the north-western side of the hill. It consists of fine-to-coarse grained, bedded, accidental lithic-rich, vitric lapilli tuff and tuff (Plate 3.21, D). Deep-seated lithic clasts as well as lherzolite fragments are rare. Accidental lithic clasts represent a dominant proportion of the pyroclastic rocks of this unit derived from shallow pre-volcanic strata (Neogene sediments – Plate 3.22, A). The quartz grains are delicately dispersed in the unsorted matrix of the lapilli tuff and tuff. Accretionary lapilli beds have not been described yet, but quartz grain clots are very common both in a mm and cm scale. Larger Neogene sediment fragments are surrounded by a thermally affected rim (Plate 3.22, B) having very often an irregular shape. Intercalated, thin, dune-bedded sequences are more common in the middle section of the unit.

The uppermost juvenile rich pyroclastic units form steep (up to 40 degrees) well-bedded cliffs (Plate 3.22, C). Laterally continuous lenticular, grain-supported, often calcite cemented lapillistone form inversely graded subunits (Plate 3.22, D). Individual lensoidal units are a few dm in length showing a gradual coarsening downward. In these lenses abraded lapilli tuff clasts are common (Plate 3.22, E). These lapillistone units often completely lack an ash matrix and show characteristic vertical inverse grading, tongue-like geometry and positive primary relief.

In each unit the volcanic glass shards are angular, non- to highly vesicular and tephritic to phono-tephritic in composition.

Interpretation of pyroclastic units of Szigliget

The pyroclastic rocks of Szigliget are interpreted as part of former phreatomagmatic volcanes. The presence of sideromelane glass shards and the large amount of accidental lithic clasts in beds from each unit indicate sub-surface phreatomagmatic explosive processes during formation of pyroclastic deposits at Szigliget.

Unit 1 is interpreted to be a lower diatreme deposit on the basis of the presence of matrix-supported, unsorted characteristics of its deposits without any well-developed bedding that suggests “en masse” deposition of a collapsing phreatomagmatic eruption column (LORENZ 1986, 2000a, WHITE 1991a, b, ORT, et al. 1998). The presence of large amount of deep-seated accidental lithic clasts of these deposits suggests very active vent dynamics during explosive processes with possible repeated vent/conduit collapse. The presence of clasts from the deepest known pre-volcanic stratigraphic units (i. e. Silurian schist) indicates that the explosion locus in this stage of the eruption must have been several hundred metres below syn-volcanic surface (at least ~700 m plus erosion since the volcanism) and the conduit must have been in a semi-sealed (not clear) state.

The presence of delicately mixed angular sideromelane glass shards mixed with finely dispersed accidental lithic clasts, the unsorted texture of beds and the mostly high energy bed-forms of the pyroclastic rocks of **Unit 2** are suggestive of a primary, phreatomagmatic explosive eruption generated origin of these beds. The presence of the large amount of accidental lithic clasts indicates subsurface explosions forming tuff rings and/or maars (LORENZ 2003b). The larger amount of juvenile fragments and the proportionally smaller amount of deeper-seated accidental lithic fragments in the beds indicate that the erupting vent must have been in a clearer stage and/or by this time the conduit wall may have been stabilised (LORENZ 2003b). The higher vesicularity of juvenile fragments supports this conclusion as well (HOUGHTON and WILSON 1989, HOUGHTON et al. 1999). The fine-grained, thinly, and/or dune-bedded lapilli tuffs and tuffs are deposited by turbulent and possible low-concentration pyroclastic density currents (CHOUGH and SOHN 1990, SOHN 1996, SOHN et al. 2003), whereas the coarse-grained lapilli tuff beds more likely represent fallout deposits from a phreatomagmatic eruption column.

In **Unit 3** the large amount of accidental lithic clasts from shallow depths (Neogene sandstone) suggests shallow sub-surface phreatomagmatic explosive origin of these beds. The presence of baked margins around larger sand- and mudstone fragments especially up-section is indicative of higher temperature/lower water content of these disrupted strata allowing occasional baking of the disrupted sand fragments. In contrast, the fluidal shape of large silt clasts, and the clot-like distribution of the quartzofeldspatic sand grains are more indicative of wet conditions of these strata at the moment of magma/sediment contact (WHITE 1991b). These conditions could be reached during a high magma discharge period, when a large amount of magma had sudden contact with wet, unconsolidated sediment thus from time to time larger amounts of magma batch could have contacted partially dry parts of the sub-surface sand beds. The undulatory, dune- or parallel bedding indicates deposition by-low concentration pyroclastic density currents and associated co-surge fallout (SOHN and CHOUGH 1989).

The textural characteristics of the steeply dipping beds of the pyroclastic succession of the Vár-hegy suggest emplacement by grain flows.

It can be concluded that the steeply inclined pyroclastic units at Szigliget represent either

1. original steep bedding surfaces and/or
2. post-eruptive reorganisation of tephra in a phreatomagmatic volcanic conduit/crater, and they are not subsequently tilted blocks (e.g. regional tectonism).

Szent György-hegy maar/diatreme

Szent György-hegy is a km wide large lava capped butte, similar to the nearby Badacsony (Figure 3.19). Szent György-hegy (415 m) is located in the axis of the Tapolca Basin, and has a similar slightly north-to-south elongate structure as Badacsony (Plate 3.23, A, B, C). The K/Ar age dating on the black, aphanitic fresh coherent lava rocks gave a whole rock age ranging from 3.48 My to 3.21 My (BORSY et al. 1986). Recently obtained $^{39}Ar/^{40}Ar$ ages from the lowermost coherent lava flows gave an age of 4.26±0.12 My in isochron age. The difference between the ages is under discussion (WIJBRANS et al. 2004).

The pre-volcanic sedimentary units can be traced up to ~300 m. Above this unit, in sporadic distribution a yellow, light grey fine grained lapilli tuff crops out, but often can only be collected as in situ debris. The textural characteristics of the lapilli tuff are very similar to those ones that have been recovered from the Badacsony and the nearby Hármas-hegy. The lapilli tuff is matrix supported that is rich in mud, silt, and or mineral phases derived from the immediate underlying Neogene siliciclastic rock units (Plate 3.23, D). The lapilli tuff is bedded, unsorted, and rich in blocky to slightly fluidal shaped volcanic glass shards. The glass shards are tephrite to phonotephrite in composition and are moderately microvesicular, with generally low microlite content. The existence of this pyroclastic unit in the basal zone of the erosional remnant of the Szent György-hegy suggests, that this location has been formed by phreatomagmatic explosive eruption, and build up a tuff ring that has been broad, and low rimmed, similar to those that have been reported from Oregon (HEIKEN 1971).

As a capping unit, scoria lapilli rich, lava spatter inter-bedded succession forms a castle like architecture of the Szent György-hegy. This pyroclastic unit is truncated by feeder dykes and minor lava flows, inferred to have fed former lava lakes in the centre of the former volcano.

Szent György-hegy is interpreted to be a phreatomagmatic volcano, that quickly evolved to be a magmatic vent, that built up a scoria cone on the crater floor of a tuff ring (VESPERMANN and SCHMINCKE 2000) similar to many examples world-wide (LORENZ et al. 1970) such in Oregon (HEIKEN 1971), Arizona (HACK 1942, WENRICH 1989) or New Zealand (HOUGHTON et al. 1999, AFFLECK et al. 2001).

Boglár diatreme

Volcanic rocks next to Balatonboglár township represent ~3.5 My old (BORSY et al. 1986) small, eroded volcanic centres located on the southern shore of Lake Balaton and are genetically related to the BBHVF (Figure 3.19). In a relatively small area (500 m times 500 m) pyroclastic rocks crop out in three hills (Figure 3.22). The immediate pre-volcanic rocks are the same Neogene siliciclastic successions that form the immediate pre-volcanic rock units in the BBHVF. At Boglár, similarly to Szigliget, the Neogene siliciclastic sequences are traceable up to variable elevations of the flank of the preserved

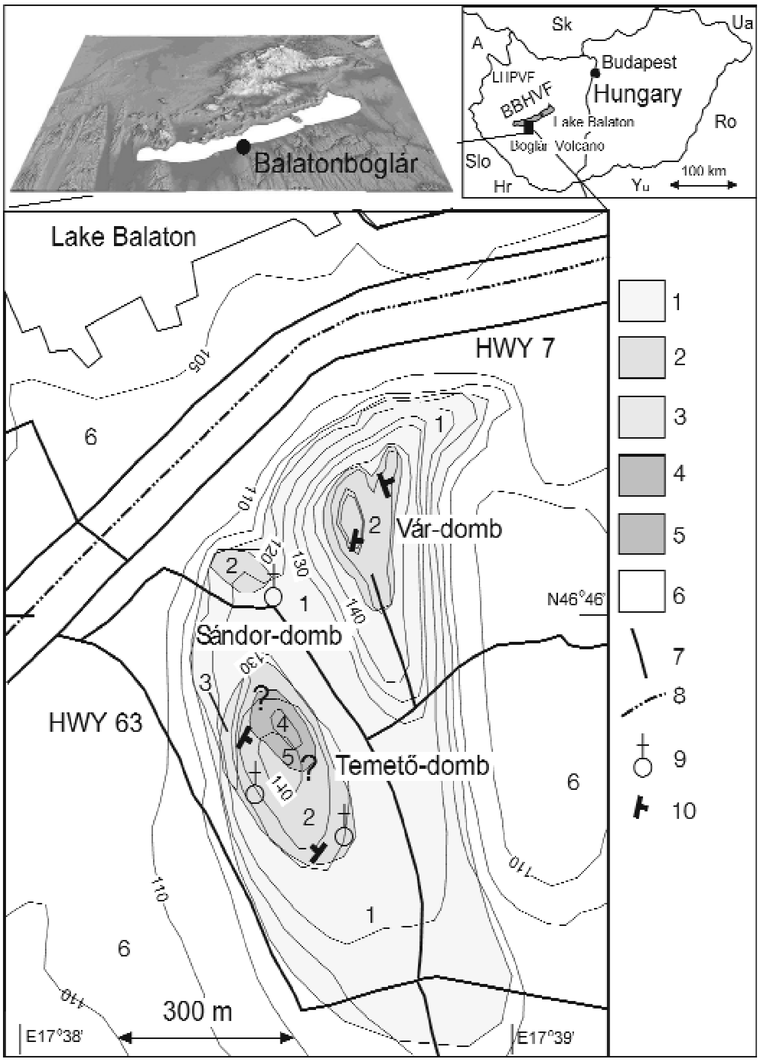


Figure 3.22. Simplified geological map of the Boglár diatreme region (after NÉMETH et al. 1999b)
1 – Neogene rock units, 2 – Lapilli tuff, 3 – Lapilli tuff with fossil tree trunks, 4 – Maar lake sediments, 5 – Peperite, 6 – Quaternary deposits, 7 – road, 8 – Railway, 9 – Church, 10 – Bedding

hill sides indicating an angular unconformity to the pyroclastic rocks exposed in Boglár. The surrounding of the hills is occupied by Quaternary swamp, lake and river sediments. The volcanic rocks of the Boglár region are entirely pyroclastic (NÉMETH et al. 1999b).

The Boglár Volcano is genetically part of the BBHVF but is geographically separated from it by Lake Balaton. Pliocene volcanic rocks on the surface are known only from two localities in the southern shoreline of the Lake Balaton, Fonyód and Boglár.

Fonyód and Boglár represent volcanic erosional remnants. Fonyód in the west is inferred as being an amphibole basalt plug on the basis of the identification of in situ basalt debris from the currently 233 m high hill (NÉMETH and CSILLAG 1999). In spite of the in situ debris that has similar petrographical characteristics to the lava flows of Badacsony, no outcrops have been identified so far from Fonyód (NÉMETH and CSILLAG 1999).

At Boglár, only pyroclastic rocks have been recovered, both in situ debris and in outcrops. Boglár consists of two main hills, Vár-domb (Castle Hill – up to 165 m) and the smaller Temető-domb (Graveyard Hill – up to 145 m), however, a third hill next to the Temető-domb, called Sándor-domb (Alexan-der Hill – up to 128 m) is often referred to in older literature (Figure 3.22 – LÓCZY 1913). K/Ar age determination of rock samples from Boglár give an age of around 3.5 My (BALOGH, pers.com.). Early geological maps show this region as *"explosion breccia-buried hills with small-scale lava flows"* (LÓCZY SEN. 1920).

Pyroclastic succession of Boglár

The volcaniclastic rocks from Boglár have been divided into two lithofacies associations (NÉMETH et al. 1999b). The largest amount of the exposed pyroclastic rocks is located in the central part of the local hills in elevated position. These rocks are rich in blocky, angular, slightly elongated, microvesicular, fresh sideromelane glass shards, commonly bearing amphibole crystals. The matrix of these pyroclastic rocks is rich in accidental lithic fragments, predominantly mud, silt, sand and minerals that have been derived from these rock units. Olivine megacryst are also present, as well as fractured, small pieces of peridotite lherzolite clasts in mm-scale. The lapilli tuff beds are unsorted, none or weakly bedded and generally chaotic in structure. They have an angular contact with the immediate pre-volcanic rock formations that have been documented well in one of the first description of the volcanic rocks of this area (LÓCZY SEN. 1913). LÓCZY SEN. (1913) found exposed zones of vertical contact between Neogene siliciclastic sediments and lapilli tuff, that was interpreted to be an explosion breccia cut through the pre-volcanic succession (LÓCZY SEN. 1913, p. 343), and thus could be interpreted as a diatreme (NÉMETH et al. 1999b). These slightly bedded, brown lapilli tuffs alongside the large volume of Neogene siliciclastic debris, contain fragments from crystalline basements from unknown origin (NÉMETH et al. 1999b). Where the bedding is relatively well developed (Vár-domb) measurements are always very diverse without any well-defined characteristic orientation. The dip is always steep, around 25°. In the bedded part of the sequence there are no impact sags, cross bedding or well defined scour fill structures. The individual beds are usually undulating with diffuse upper and lower contact. The sideromelane glass shards (Plate 3.24, A, B) have been measured by electron microprobe method to be tephrite, tephriphonolite, and/or minor trachybasalt (NÉMETH et al. 1999b). Trachybasalt composition is more common among oriented trachytic texture pyroclasts that are darker in colour and interpreted to be tachylite (NÉMETH et al. 1999b).

The general massive character of these rocks, the angular contact with pre-volcanic sediments, the presence of sideromelane glass shards as well as mud and silt and other but low proportion of lithic fragments from deep-seated sources suggest that it has been a result of magma–water interaction, probably in a shallow level. The produced phreatomagmatic volcanoes have been eroded back to their root zones, and now small diatremes left over, that cut into the pre-volcanic succession. Boglár is interpreted to be one of the deepest levels of exposure of a group of diatremes in the BBHVF. The major source of water to fuel the phreatomagmatic explosions is inferred to be the porous media aquifer of the Neogene sequences. After exhaustion of the water source a more magmatic explosive eruption may have taken place that has been resulted in more scoria rich lapilli tuffs in the centre of the Vár-domb of Boglár.

Figure 3.23. Silicified *Abies* sp. tree trunk (T) from volcaniclastic debris flow deposited succession, a result of a lahar on a phreatomagmatic volcano of Boglár

The second lithofacies association that has been identified in Boglár forms the basal zone of the hills (NÉMETH et al. 1999b). This pyroclastic unit composed of pyroclastic beds that are well- but thickly bedded to non-bedded, massive units (Plate 3.24, C, D). The matrix of the lithofacies is weakly cemented, and friable. The pyroclastic rocks are unsorted, and rich in yellowish rounded mud and silt "balls" (NÉMETH et al. 1999b). These rounded cm-to-dm size clasts are often strongly diagenised, and have a dark brownish crust and radial joint pattern in their interiour. The fine grained matrix of the lapilli tuff beds are a delicate mixture of silt, mud and altered, palagonitized sideromelane glass shards, that are diverse in shape, colour, vesicularity and degree of alteration (NÉMETH et al. 1999b). This succession is subhorizontally bedded and its stratigraphical position is seemingly "inside" the pre-volcanic Neogene succession (NÉMETH et al. 1999b). Due to poor outcrop availability it is not possible to better constrain the 3D relationship between the pre-volcanic and volcanic units. The uniqueness of this succession is that it contains an unusually large amount of silicified tree trunks (Figure 3.23), which are identified as *Abies* species (NÉMETH et al. 1999b). Just above of this volcanic unit, in a small area volcanogenic sandstone with minor volume of volcanic detritus and rounded quartz have been mapped in the western hills just above the fossil tree trunk bearing units (NÉMETH et al. 1999b). On the hilltops from in situ debris volcaniclastic rock with large (cm-to-dm-size) coherent vesicular basalt clasts in a volcaniclastic matrix have been recovered, that is indicative that feeder dyke with peperitic margin may have been cropped out in this area (NÉMETH et al. 1999b). However, to establish that peperite exists in this site needs a clear demonstration of intrusive and irregular contacts between host sediment and dyke (WHITE et al. 2000, SKILLING et al. 2002), which is due to the poor outcrop availability is very unlikely to be possible.

The chaotic structure of different lithologies in a very altered sandy to muddy matrix, and the presence of altered larger tuff fragments rimmed by pyroclastic rock, strongly suggests reworking processes and possible destructive events on a history of a "wet" phreatomagmatic volcano, that initiated volcanic debris flows (lahars. – PIERSON and SCOTT 1985, SMITH and LOWE 1991, SCHMINCKE et al. 1999, VALLANCE 2000, ELLIOT and HANSON 2001), very likely associated with intra-crater reorganisation of wet volcanic debris (COLE et al. 1999, WHITE and MCCLINTOCK 2001). The large amount of sandy matrix and matrix supported large clast bearing character is interpreted as cohesive debris-flow deposit in which the massive, matrix-supported pebbly mudstone (thermally affected Pannonian sandstone fragments) and tree trunk fragments were suspended in and supported by the matrix (FISHER and SCHMINCKE 1994).

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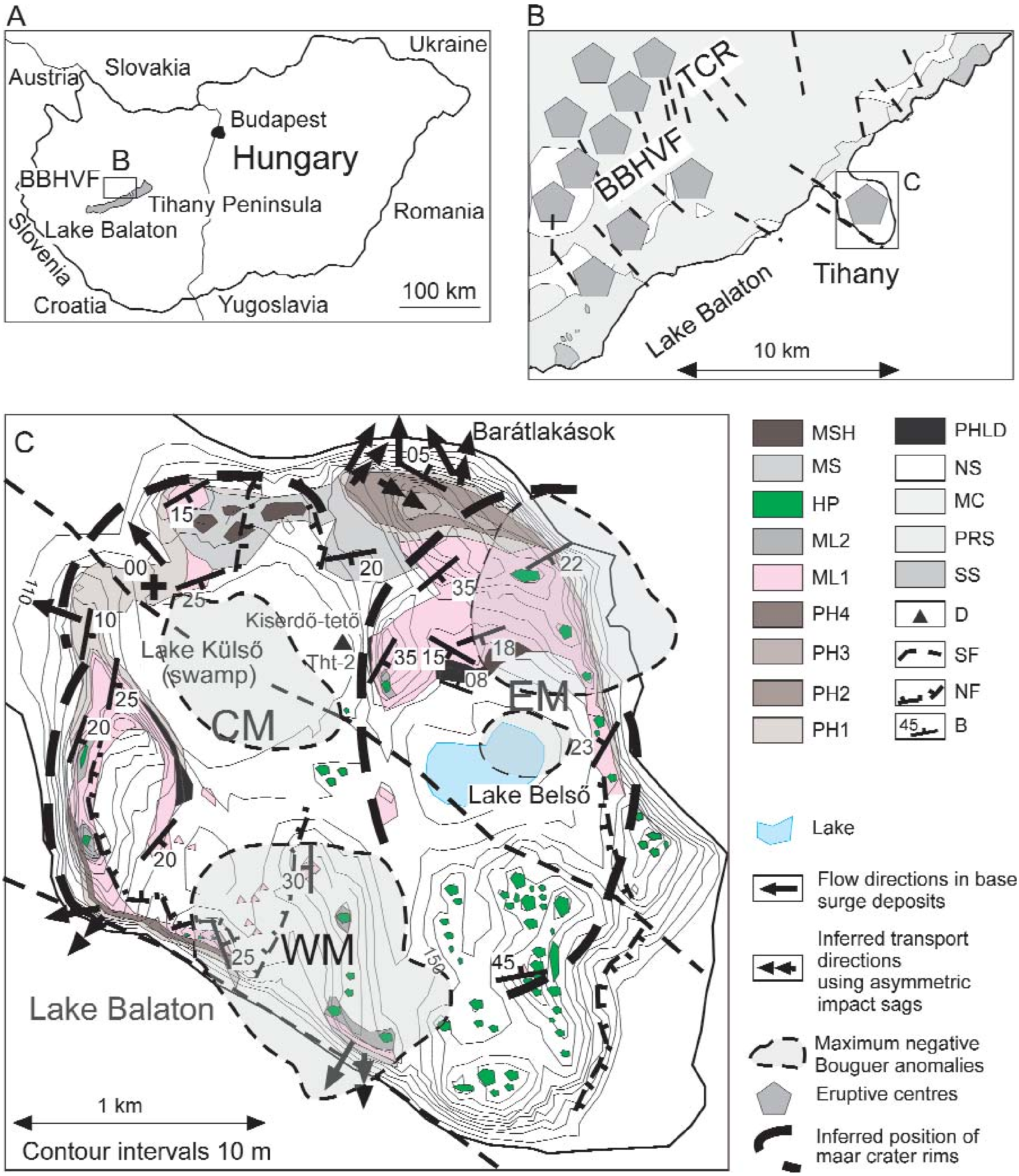
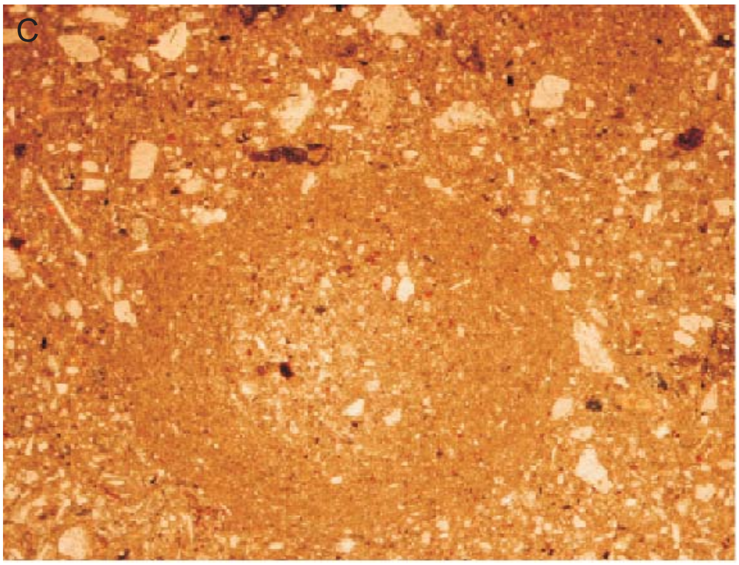
Simplified geological map of the Tihany Peninsula (after NÉMETH et al. 1999, 2001)



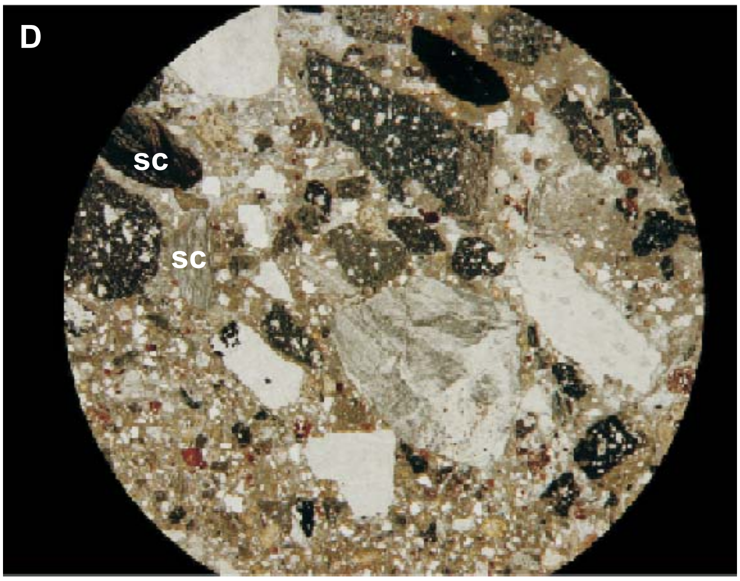
B = bedding
NF = normal fault
SF = strike slip fault
D = drill core
PHLD = phreatomagmatic lower diatreme
NS = Neogene siliciclastic sediments
MC = Mesozoic carbonates
PRS = Permian red sandstone
SS = Silurian schist
MSH = magmatic Strombolian/Hawaiian units
MS = magmatic Strombolian units
HP = hot spring pipes
ML1–ML2 = maar lake sediments
PH1–PH4 = phreatomagmatic lithofacies associations

Outcrop photo of the lowermost pyroclastic sequence at Tihany (Barátlakások) exhibiting beds of the PH1 lithofacies association

Hard tuff horizon (arrow), rich in accretionary lapilli in the PH1 lithofacies association



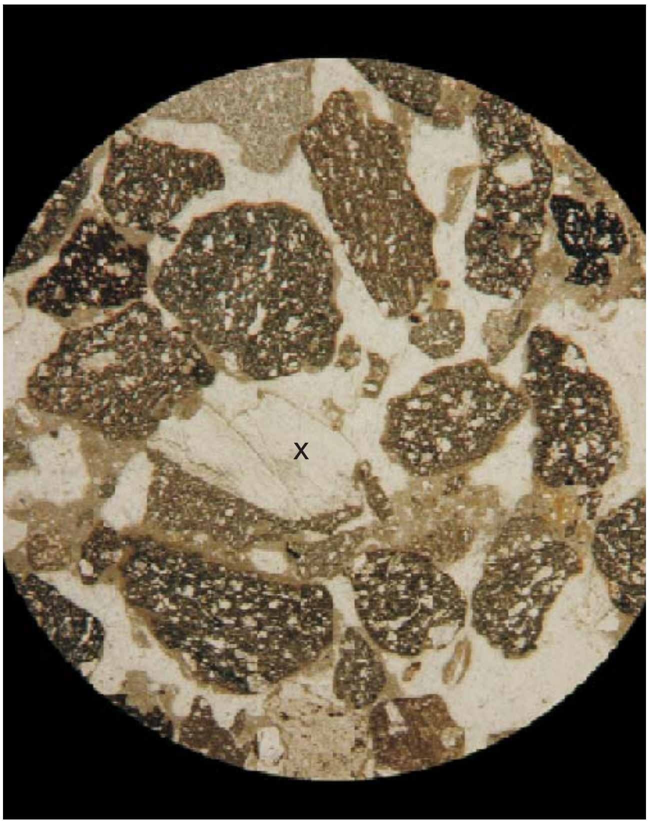
Photomicrographs of a fine-grained tuff predominantly composed of fragments derived from the Neogene siliciclastic units. Photo is 1cm across (paralell polarised light)



Photomicrograph of a Silurian schist-rich (sc) lapilli tuff from the PH2 at Barátlakások. Photo is about 2 cm across (parallel polarized light)

Accretionary lapilli bed from the LHLD of Kiserdő-tető. White lines point to the accretionary lapilli

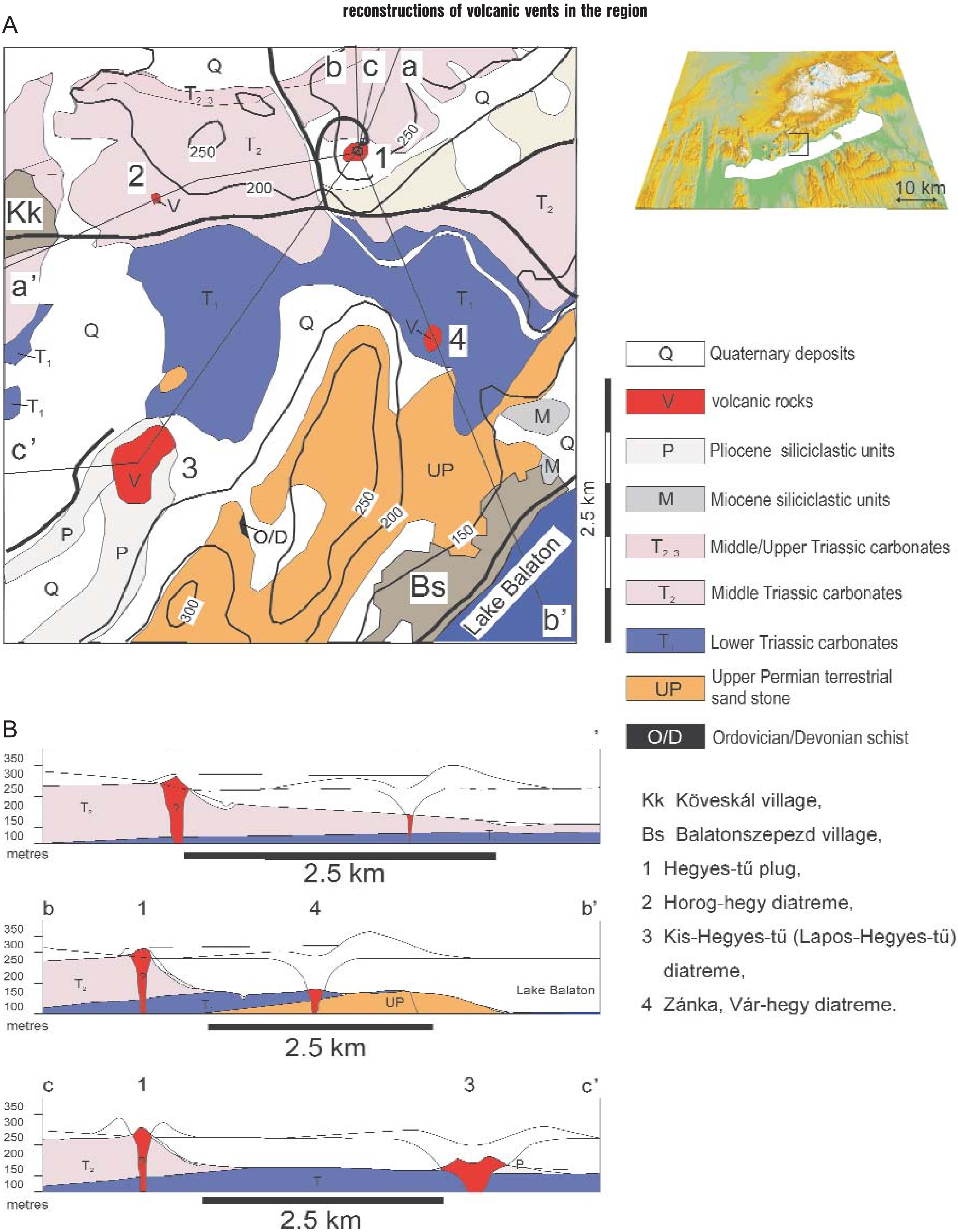
The massive pyroclastic breccia of Tihany (LHLD) contains a diverse variety of country rocks from the entire known pre-volcanic rock units, and even intact fossils of *Congeria* shells (white line) from the immediate underlying Neogene siliciclastic units have been identified



Photomicrographs of a lapilli tuff inferred to be part of the Gilbert-type delta front. Note the large amount of rounded lapilli and broken pyrogenic and xenocryst (x), cemented by calcite. The photo is 2 cm across

Vertical pipe that cut through the travertine mound on top of Csúcs-hegy, interpreted to be a hot spring pipe in a maar crater floor (Budai et al. 2002)

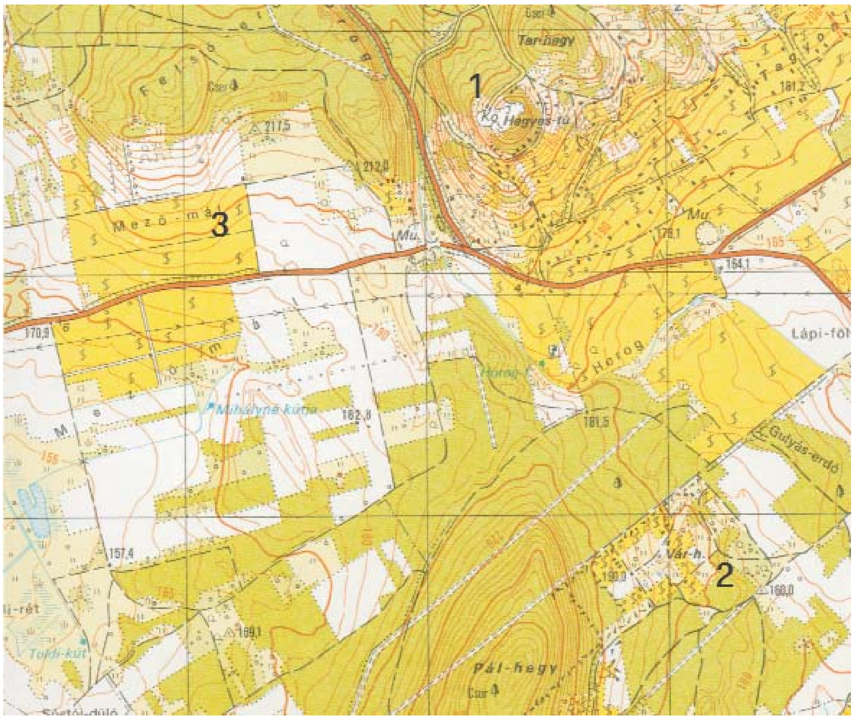
A) Simplified geological map of the eastern part of the Kál Basin. B) Cross sections through the eastern part of the Kál Basin show possible reconstructions of volcanic vents in the region



Topography of the eastern Kál Basin and its relationship with the location of deeply eroded diatremes

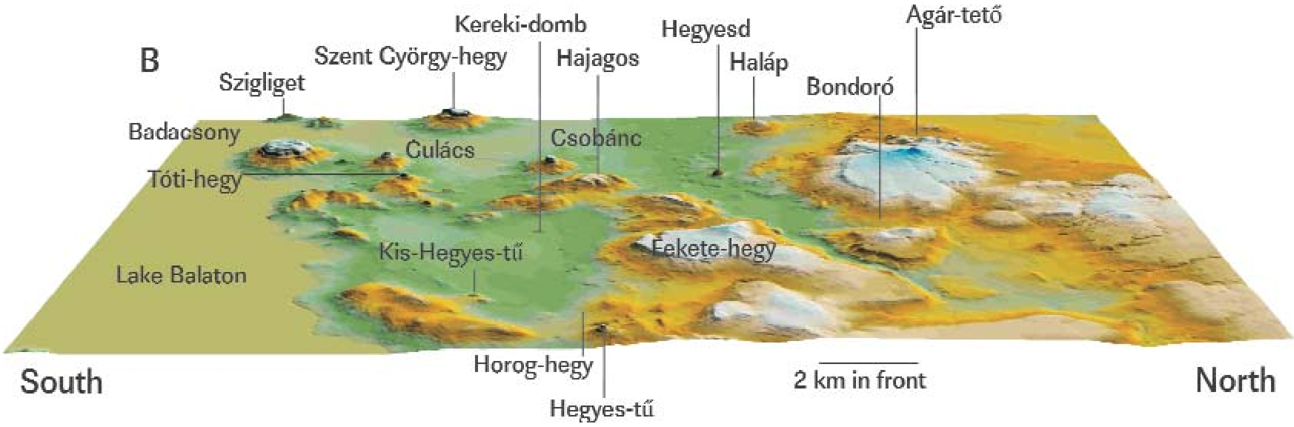
1 – Hegyes-tű, 2 – Zánka, Vár-hegy, 3 – Horog-hegy. Rectangular grid spacing on the map is 1 km. Map is a detail from the 1 to 25,000 scale topography map of Hungary

A

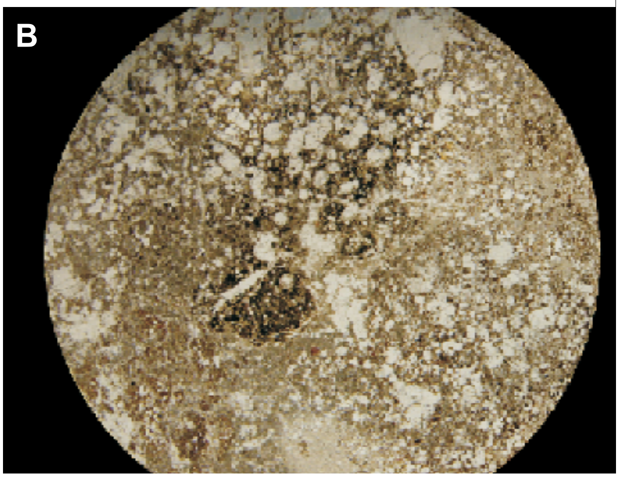


Panoramic view toward the Kál Basin on DTM from the Hegyes-tű. In the centre of the basin are small diatreme remnants (labeled) identified as small morphological irregularities

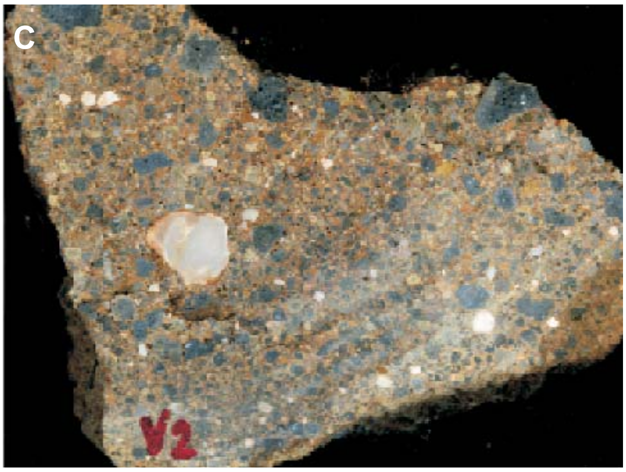
B



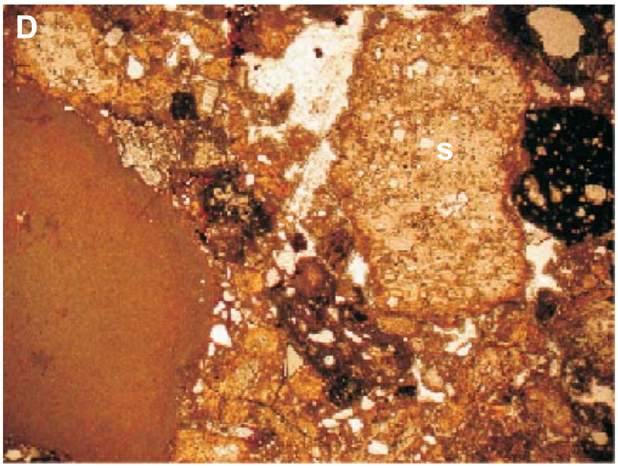
Highly vesicular basanitic lava clasts in a strongly palagonitized, mud-rich matrix adjacent to the coherent lava body of Hegyes-tű. The clastic zone is somehow surrounded by coherent lava indicating its well-localized structure, perhaps a bubble, that formed inside of the still liquid basanitic lava



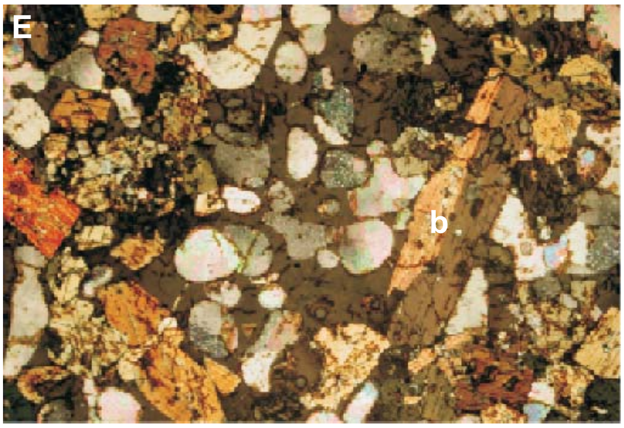
Photomicrograph of the pyroclastic breccia near the Hegyes-tű basanite plug. Note the irregular shape of tachylitic lapilli (black). The picture is ~2 cm across.



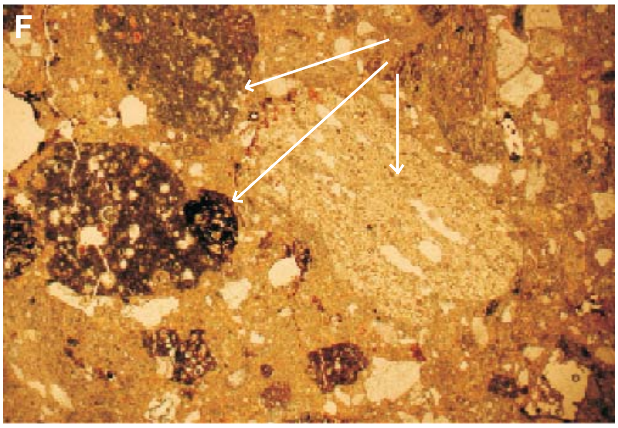
Handspecimen of a lapilli tuff of the zánkai Vár-hegy diatreme. Note the limestone lapilli (white) derived from Mesozoic rock units. The shorter side of the photo is about 15 cm



Photomicrograph of a lapilli tuff of the Zánka Vár-hegy diatréme, rich in sideromelane glass shards (s) with oriented vesicles that are hosted in a yellowish muddy matrix. The short side of the photo is ~4 mm



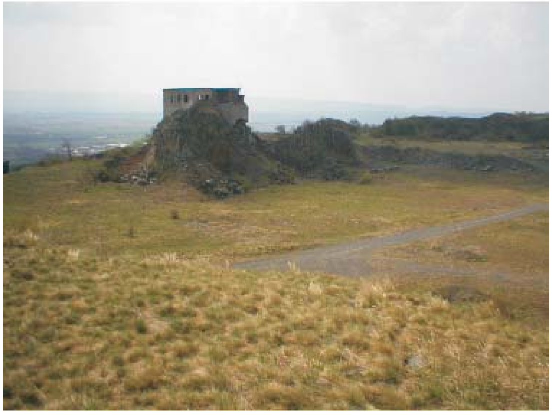
Photomicrograph of a tephriphonolitic glass hosted amphibole aggregate as a clast in a pyroclastic rock recovered from in situ debris from the Horog-hegy, near Hegyes-tű. Short side of the photo is ~2 mm



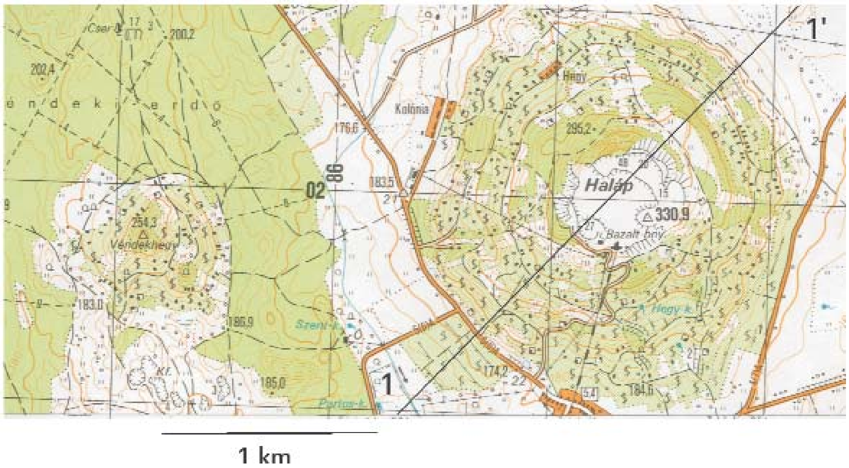
Photomicrograph of volcanic glass shards (arrows) from the Kis-Hegyes-tű. Shorter side of the picture is ~4 mm

A

The "castle-like" architecture of Haláp [the log on Figure 3.14 represents a section just below the abandoned building to the left of the photo]

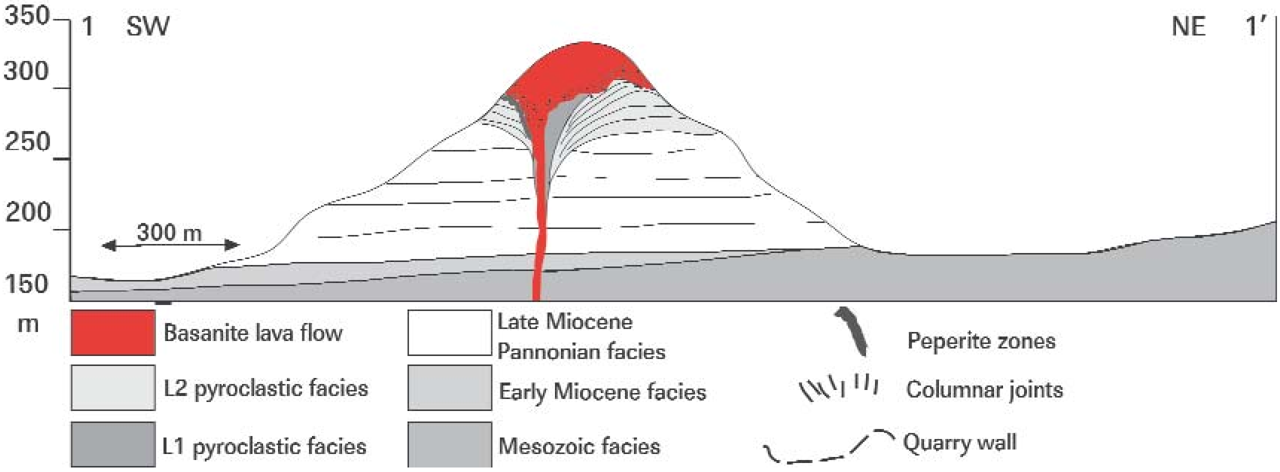


B

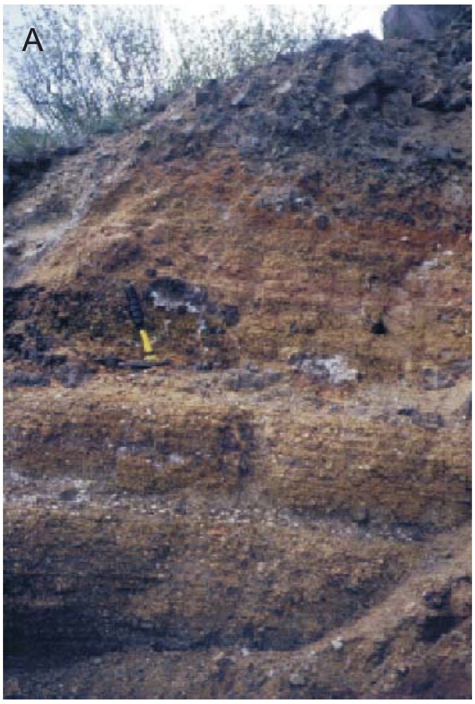


Location of the Haláp, and Véndek-hegy volcanic remnant [a detail from the 1 to 25,000 scale topographic maps of Hungary]. Active quarrying removed the former coherent basanitic lava, leaving behind collar of pyroclastic rocks. 1–1' line represents the line along the cross section is made on "C"

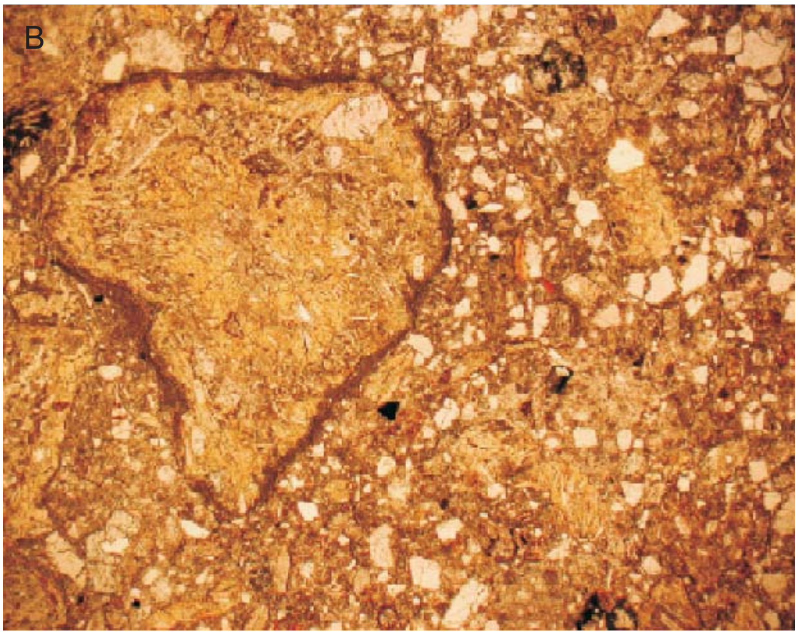
C



A cross section through Haláp shows the relationship between pyroclastic, coherent lava and pre-volcanic rock units



A bedded pyroclastic succesion in the southern margin of the outcrops at Haláp. The photo corresponds to the stratigraphic log presented on Figure 3.14



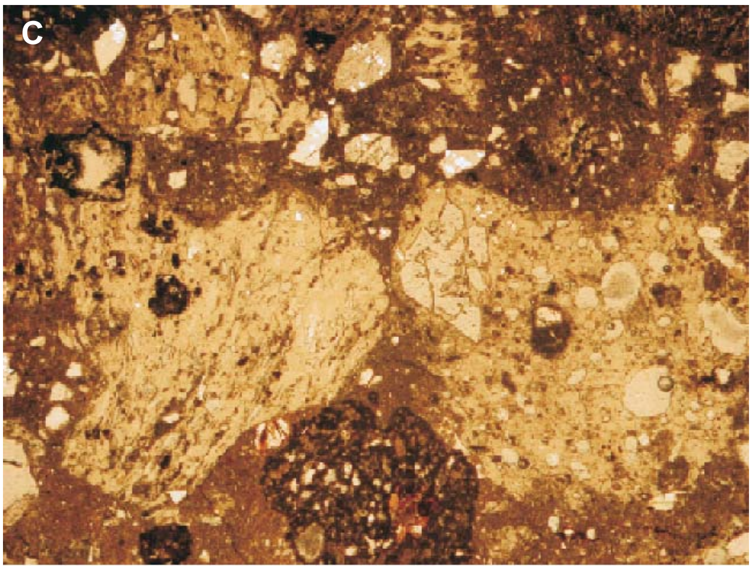
Photomicrograph of strongly palagonitised sideromelane glass shard from a lapilli tuff of the central part of the Haláp maar/tuff ring remnant

A close up of pebbles (middle of picture) in a bedded lapilli tuff at Haláp, which derived from the immediate pre-volcanic siliciclastic units

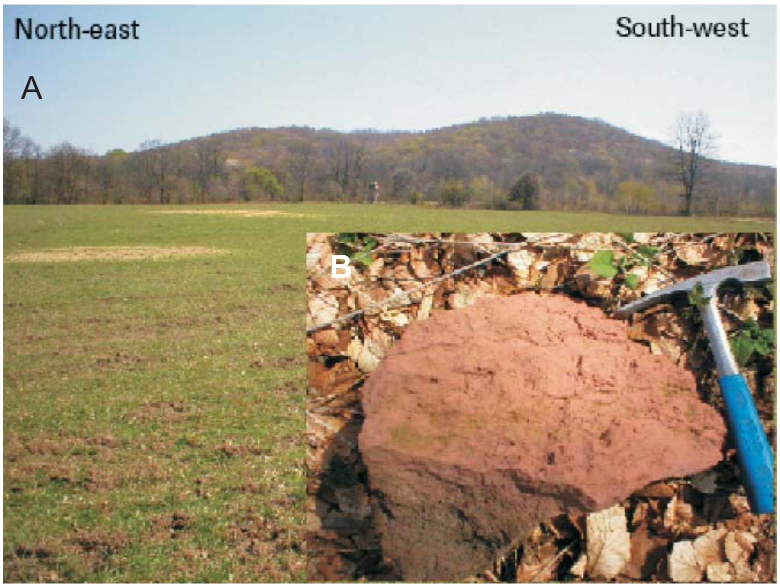
Mud filled vesicles of a scoriaceous lava spatter in a phreatomagmatic lapilli tuff of Haláp near the contact of pyroclastic succession and coherent lava

A stack of scoriaceous lapilli tuff and tuff breccia indicating subsequent Strombolian-style explosive activity in the maar/tuff ring basin

Transition to more matrix supported tuff breccias that have silty matrix between lava spatters

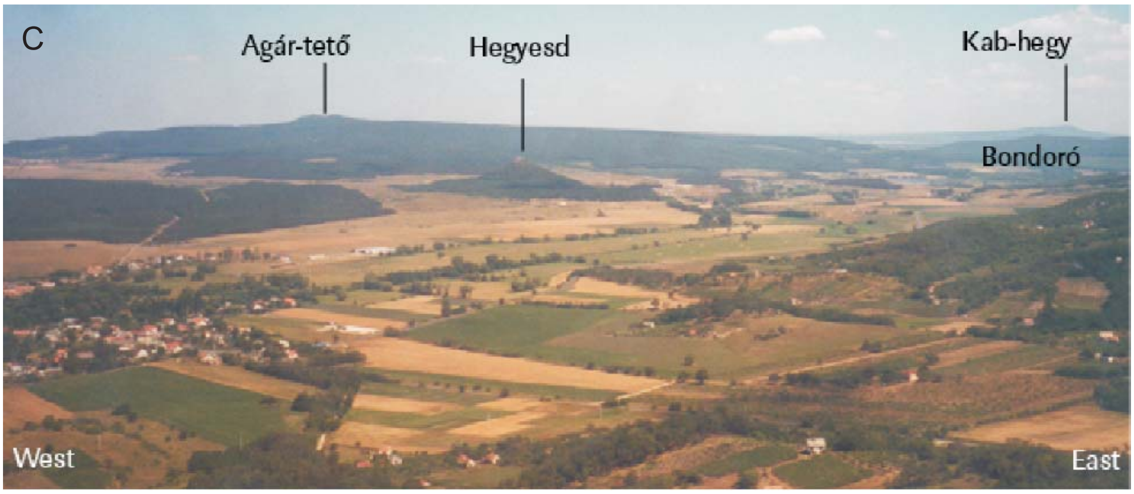


Photomicrograph of a volcanic glass shards from the Véndek-hegy pyroclastic succession. The short side of the photo is ~4 mm

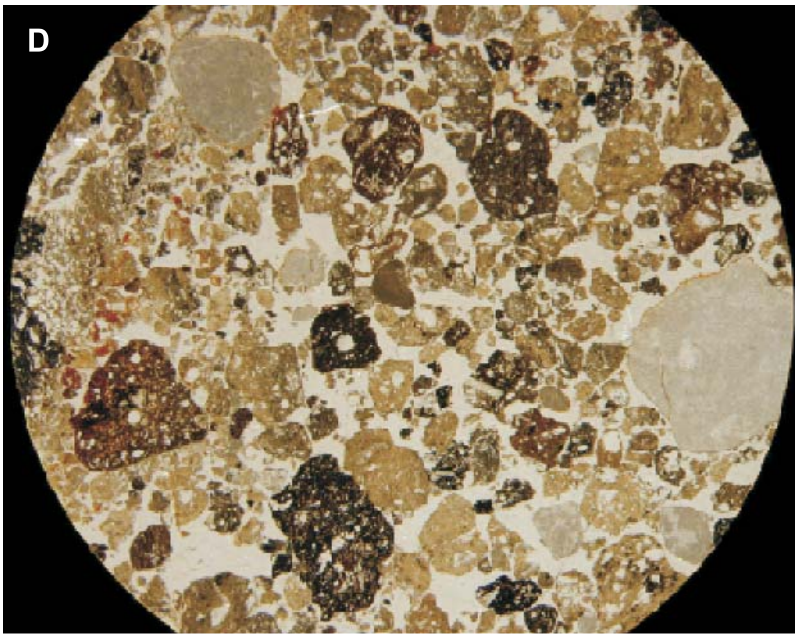


Erosional remnant of scoria cone at the top of the Agár-tető lava shield still retaining its original volcanic morphology, suggestive for young age

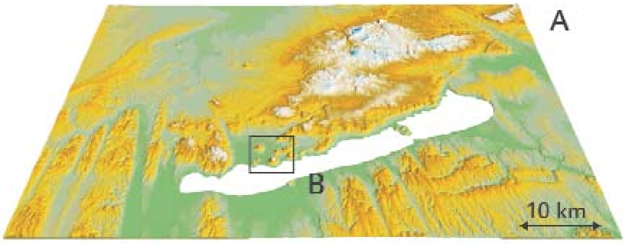
On the flank of the scoria cone remnant of Agár-tető a large number of fragments from red spindle bombs are present



The scoria cone cap is clearly visible (from south to north) on the Agár-tető volcanic complex



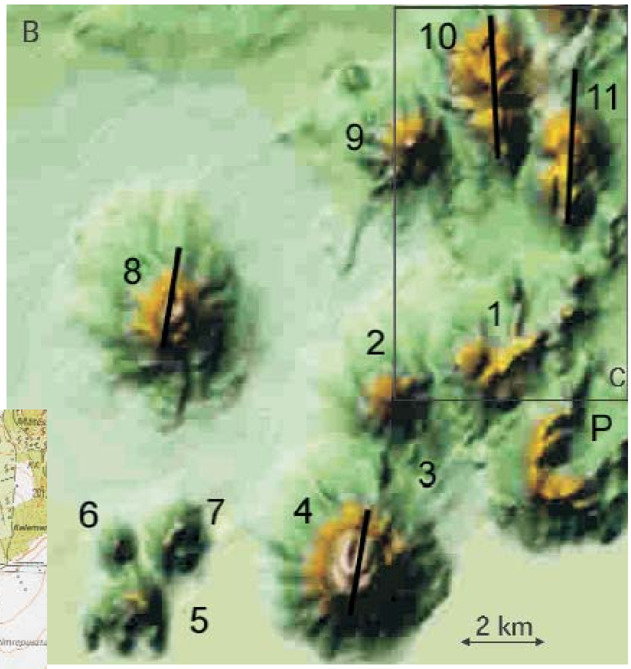
Photomicrograph of a pyroclastic rock recovered from Quaternary debris flank around Hegyesd exhibiting volcanic textures suggesting some degree of reworking. Note the calcite cement around rounded, abraded glassy pyroclasts and intact pyroclastic lapilli in the texture. The photo is ~2 cm across

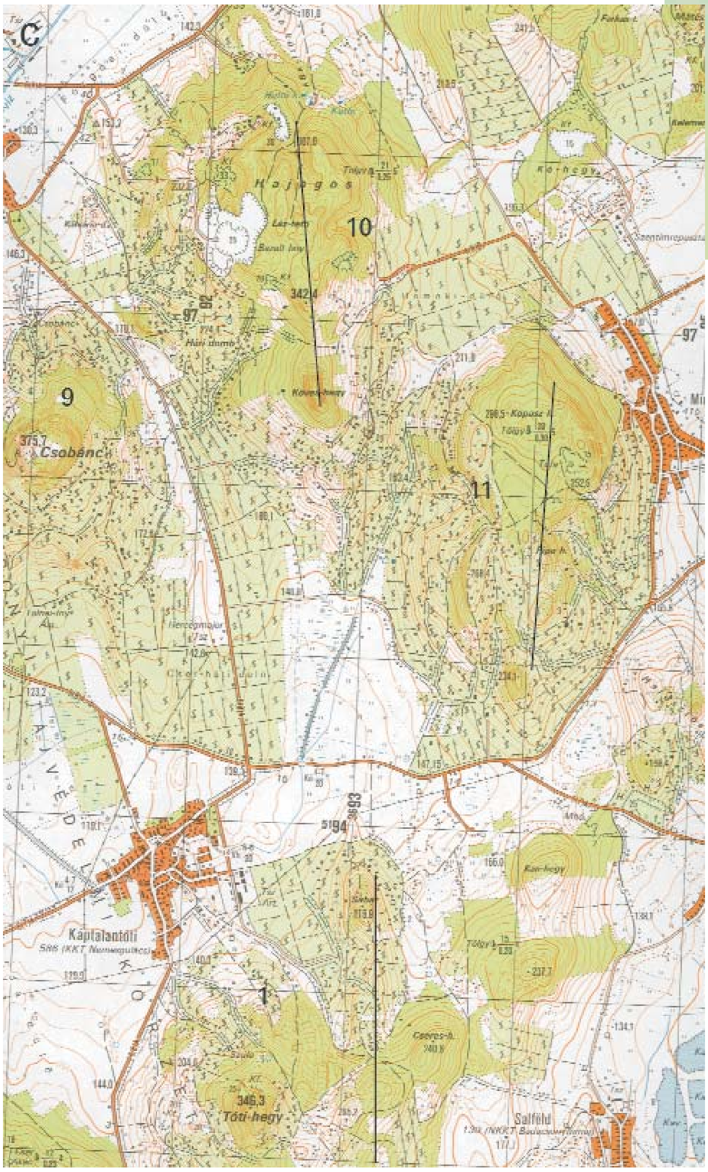


An overview DTM map shows the location of the Tapolca Basin.

DTM of the eastern margin of the Tapolca Basin, showing the location of the presented volcanic erosion remnants. Lines over the vent remnants indicate the elongation direction of vent remnants. Rectangular field corespondes to topographical map on "C"

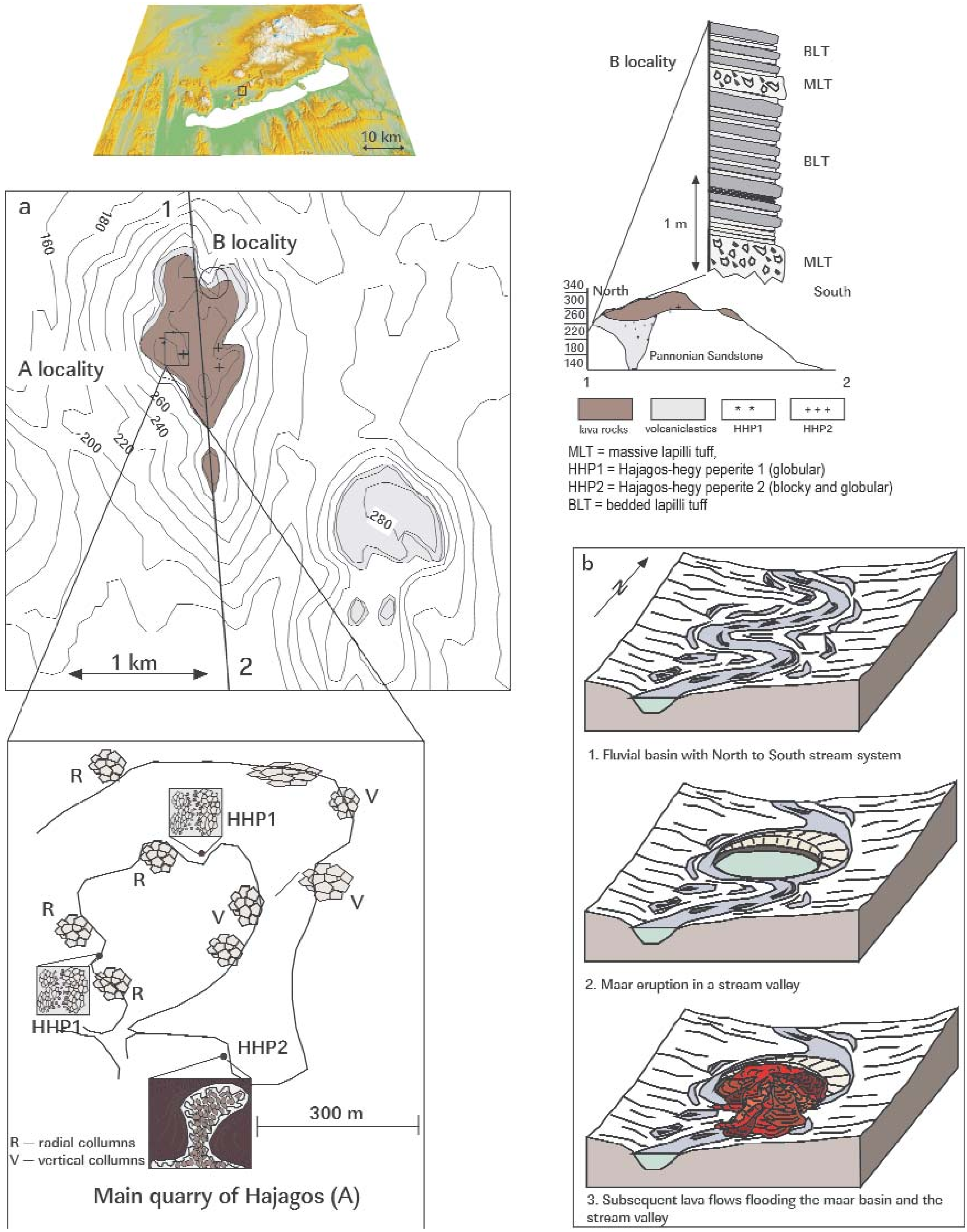
1 — Tóti-hegy group, 2 — Gulács, 3 — Hármas-hegy, 4 — Badacsony, 5 — Kamonkő (Szigliget), 6 — Várhegy (Szigliget), 7 — Antal-hegy (Szigliget), 8 — Szent György-hegy, 9 — Csobánc, 10 — Hajagos, 11 — Kopasz-hegy group, P = ridge of Palaeozoic unit





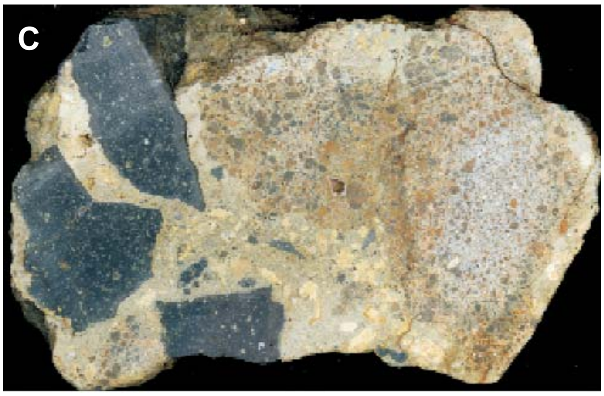
A detailed topographic map shows the fine morphology of the pyroclastic hills of the eastern Tapolca Basin. Numbers represent the same localities introduced on the "B" figure. The map source is a 1 to 25,000 scale topographic map series of Hungary. The rectangular grid on the map has 1 km spacing

Simplified geological map of the Hajagos erosional remnant (a) and a proposed evolution of the Hajagos volcano (b).



Siliciclastic sediment between the coherent lava body at Hajagos suggesting that the vent zone that subsequently was filled by a lava lake was occupied by muddy slurry

Blocky peperite that formed due to interaction of basanitic melt and siliciclastic host sediment at Hajagos



A blocky peperite in handspecimen from Hajagos, showing fluidization textures through a host lapilli tuff caused by the intrusion of basanitic melt

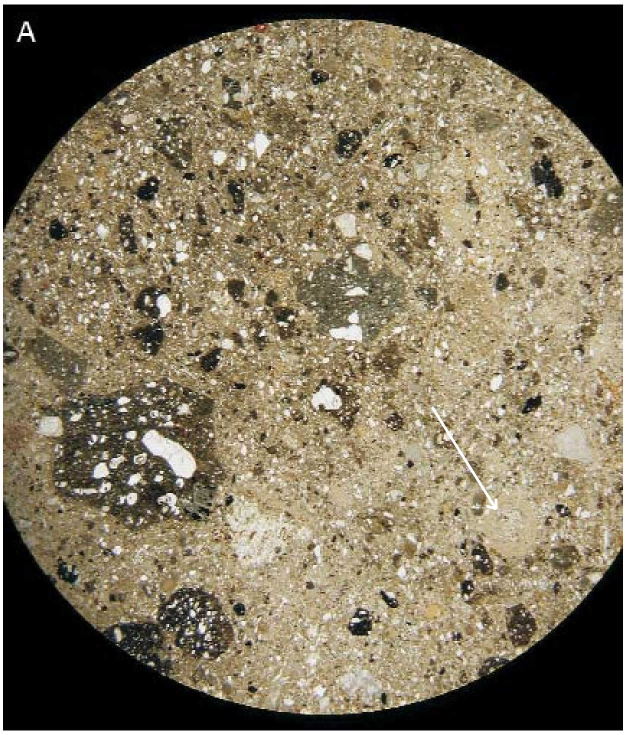
Fluidal peperite (globular) in lapilli tuff host at Hajagos. Note the horizontal fingerlike protrusions of basanitic lava that truncating the host pyroclastic deposit

Fluidal peperite developed due to intrusion of basanitic melt into siliciclastic silt at Hajagos

Peperitic zone along the foot of a lava flow at Hajagos. The lava flow movement is inferred to be from right (north) to left (south)

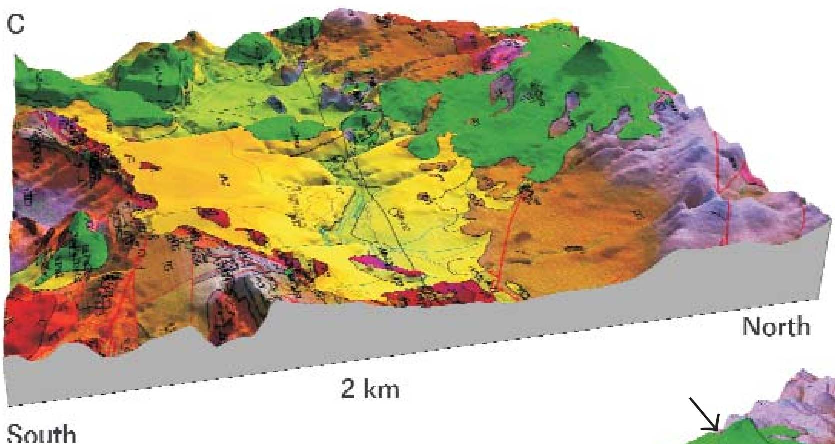
The contact zone of the lava flow at Hajagos is irregular, and vesicular pillowed lava with baked silt is the main textural feature of this unit

Vesicular lava flow foot texture is known from the lower flow unit of the lava flows from the southern margin of the Hajagos and from the Köves-hegy, just a few hundreds metres toward south of Hajagos

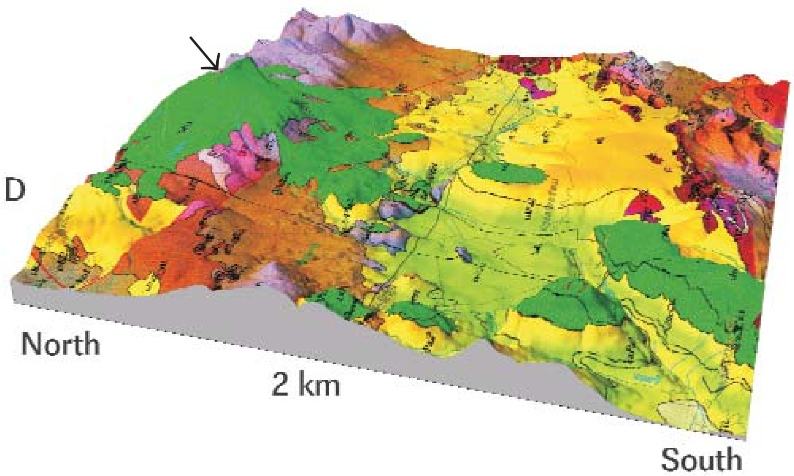


Photomicrograph of a phreatomagmatic tuff of the eastern sector of the tuff ring remnant of Bondoró. Note the palagonitized volcanic glass shards and accretionari lapilli (arrow)

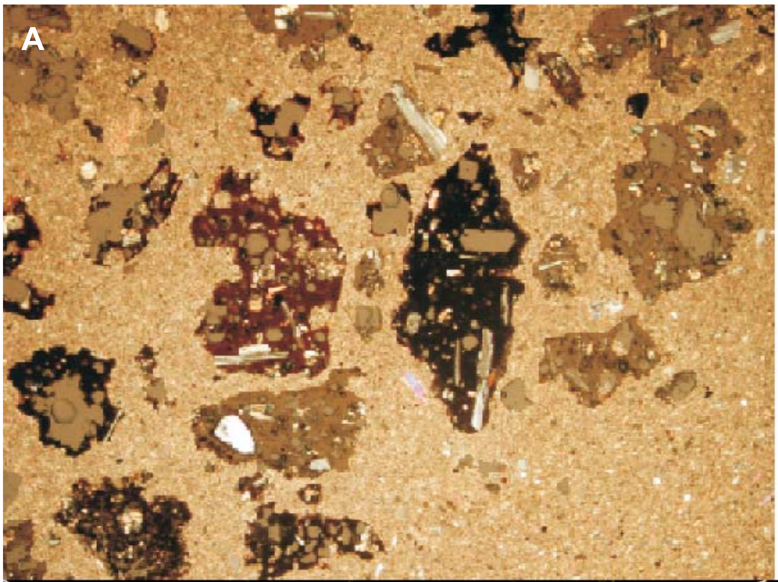
Panoramic view to the Csobánc (arrow) from Hajagos. The hill is capped by lava spatter that has been intruded by basanitic feeder dykes, today preserved as columnar jointed basanite. The lower section of the erosional remanant is formed by a phreatomagmatic lapilli tuff succession, which is only poorly exposed



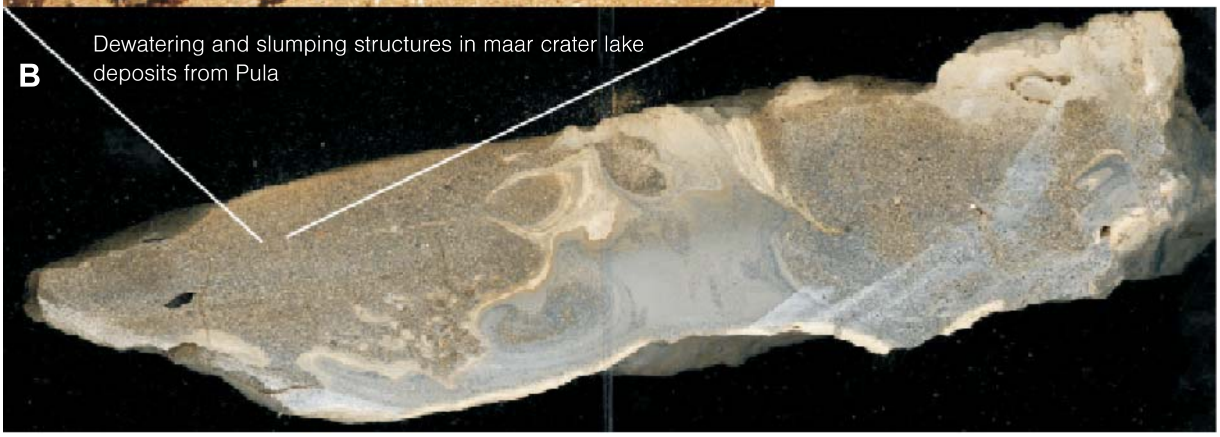
DTM model for the Pula region looking toward west



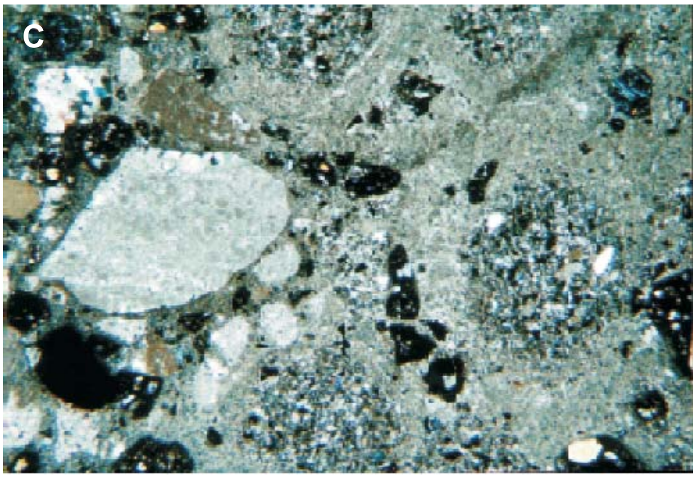
DTM model of the Pula region fitted with the 1:50,000 scale geological map of the BBHVF (BUDAI et al., 1999). Pula is a small depression in the centre of the 3D model between the background of the Kab-hegy shield volcano (arrow) and the foreground Tálodi-erdő lava field



Photomicrograph [half closed Nicols] of angular volcanic glass shards from deposits accumulated in the maar crater of Pula.
The short side of the photo is 4 mm

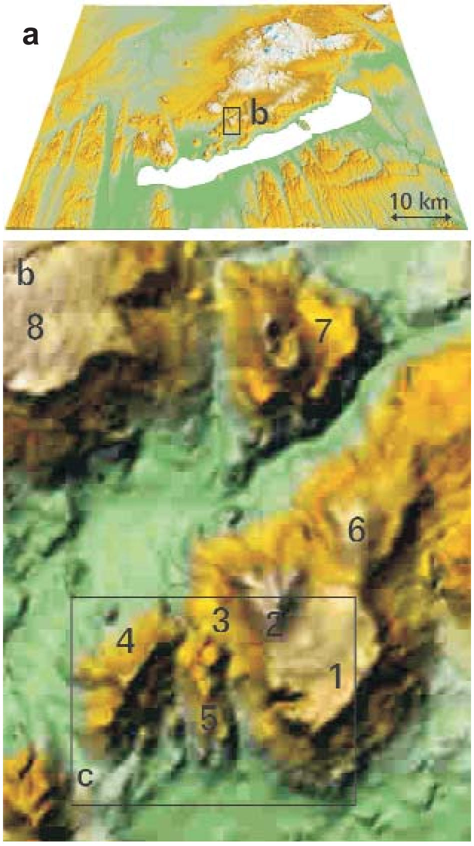


Dewatering and slumping structures in maar crater lake deposits from Pula



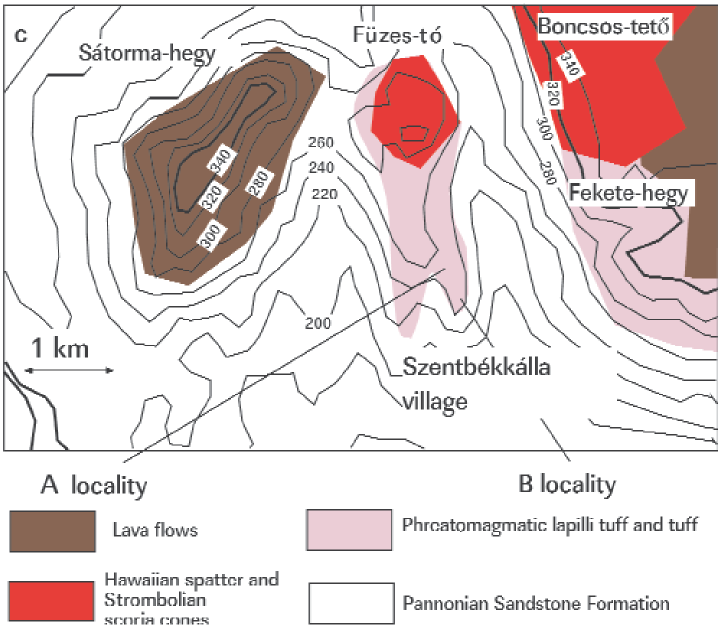
Photomicrograph of rim-type accretionary lapilli from a phreatomagmatic crater rim sequence of Pula

Rounded lapilli tuff as a clast (arrow) in the volcaniclastic succession of facies 4 at Pula, interpreted as reworked volcaniclastic mass flow deposit, interbedded with primary pyroclastic units, indicating syn-eruptive reworking of the already deposited eruptive products



DTM map of the area around Fekete-hegy ("a" and "b").

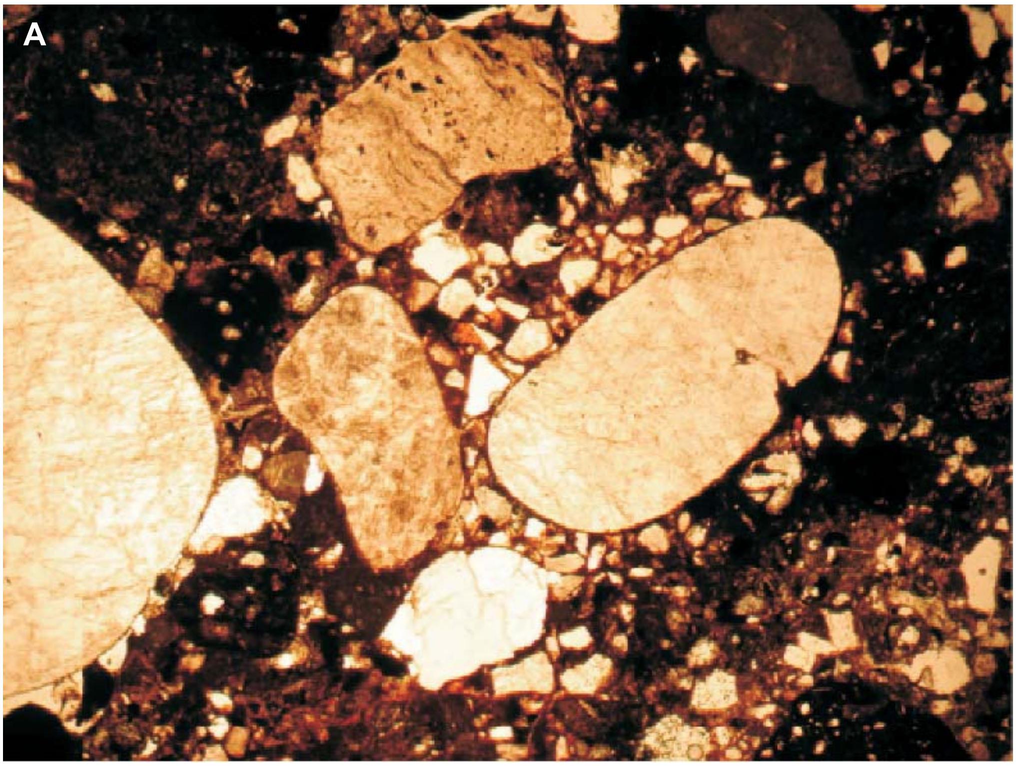
1 — Fekete-hegy, 2 — Boncsos-tető, 3 — Füzes-tó (Kopácsi-hegy), 4 — Sátorma-hegy, 5 — Szentbékkálla mafic pyroclastic flow, 6 — Kapolcs diatreme, 7 — Bondoró, 8 — Agártető lava field



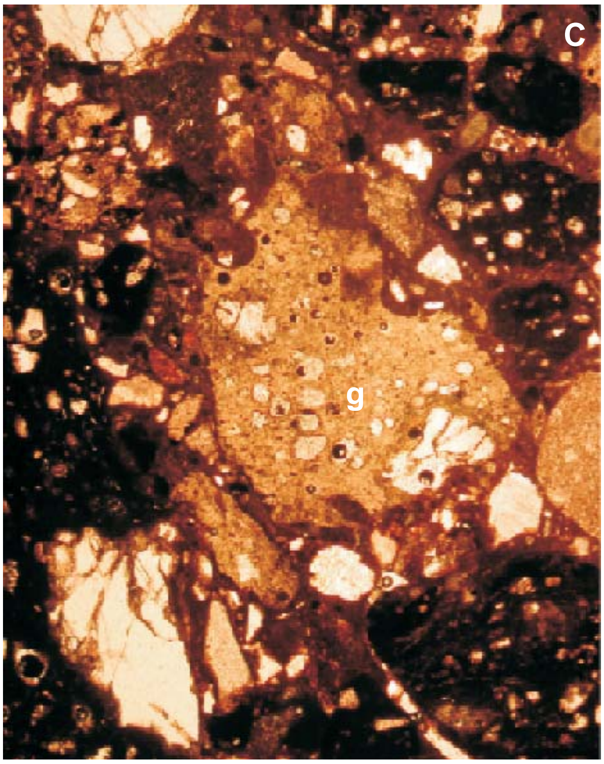
A simplified geological map shows the relationship between different volcanic facies nearby Szentbékkálla village. Localities A and B refer to two section that have been identified and described

Overview of the Szentbékkálla outcrop exhibiting a phreatomagmatic pyroclastic flow unit (lower part) overlain by dilute base surge deposits

Close up of the massive lower part of the section shown on Figure 3.17. Note the irregular shaped accretionary lapilli bearing tuff as fragment in the massive lapilli tuff (arrow)

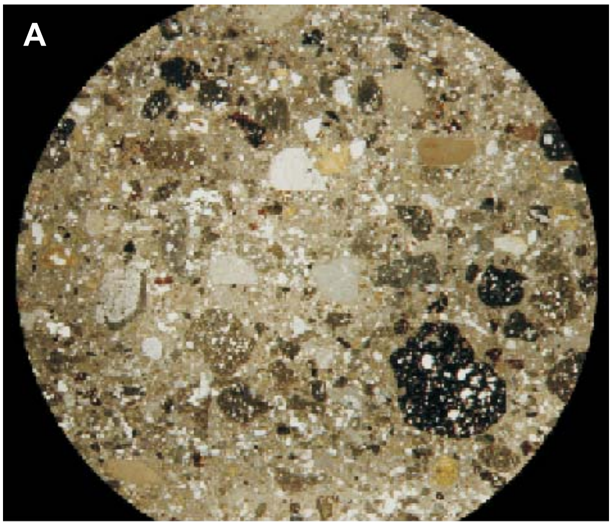


Photomicrograph of picked up pebble from the basal zone of the pyroclastic succession near Szentbékkálla, which is in contact to the underlying gravel beds indicating, that the gravel was still unconsolidated and easy to be picked up en route from horizontally moving pyroclastic density current

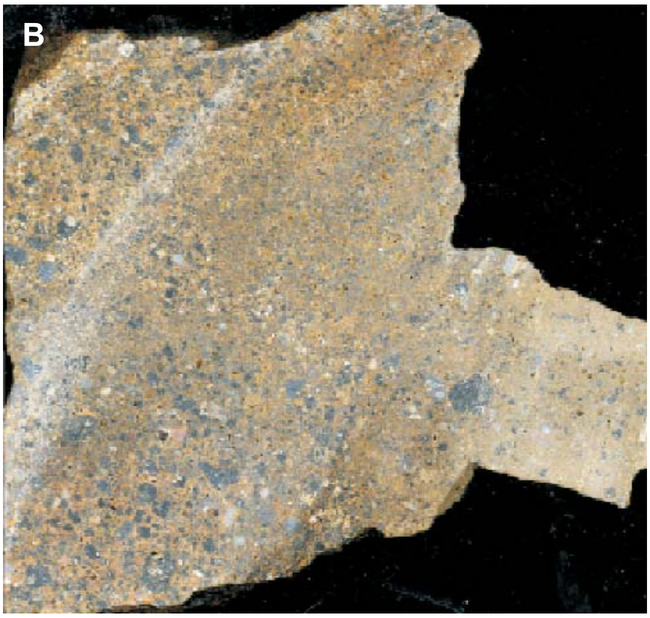


Photomicrograph of volcanic glass shards (g) in a fine siliclastic fragment-charged matrix of the massive lapilli tuff from the Szentbékkálla loacality. Short side of the photo is about 1 mm

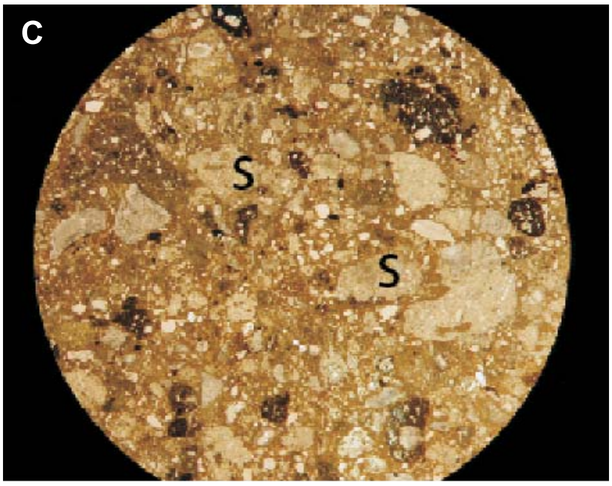
Gas-segregation pipes (arrows) in the basal zone of the massive lapilli tuff of Szentbékkálla filled with angular to subrounded predominantly accidental lithic fragments



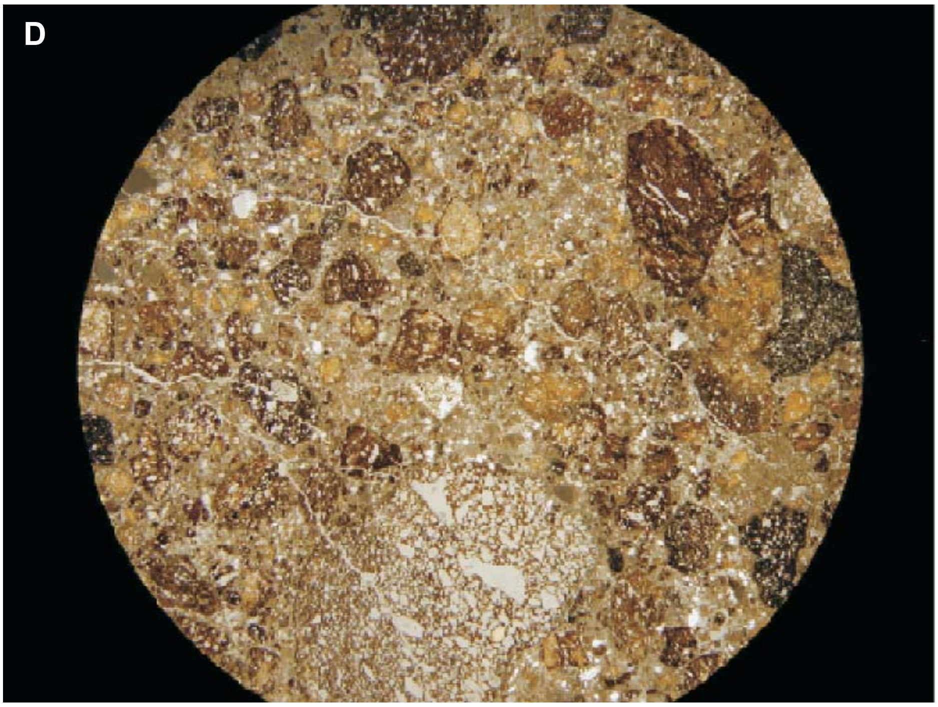
Photomicrograph of a phreatomagmatic lapilli tuff rich in quartzo-feldspatic sediment-derived mineral phases indicating the water saturated nature of the sediments that played an important role during the eruption at Fekete-hegy. The photo is ~2 cm across



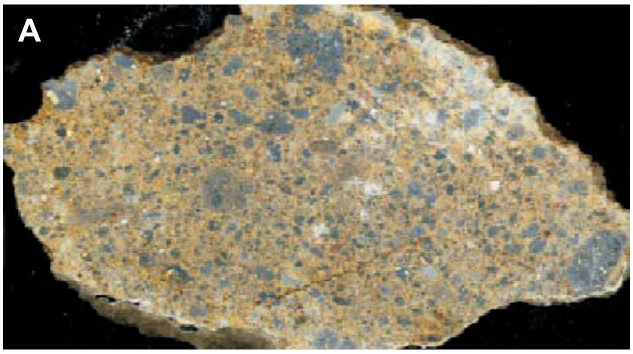
Handspecimen (short side of the photo is 10 cm) of a lapilli tuff from Kereki-hegy



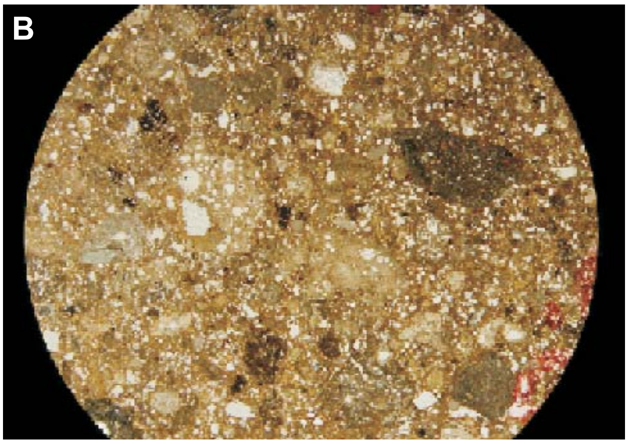
Photomicrograph of a lapilli tuff from Kereki-hegy, that is rich in fragments derived from silici-clastic sediment and blocky glass shards (s). The photomicrograph is ~1 cm across



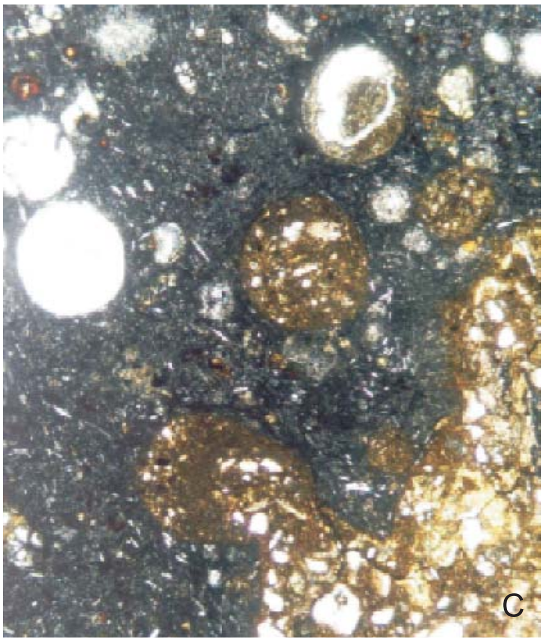
Photomicrograph of a juvenile lapilli with trachitic texture from the Harasztos-hegy pyroclastic succession. The photo is ~2 cm across



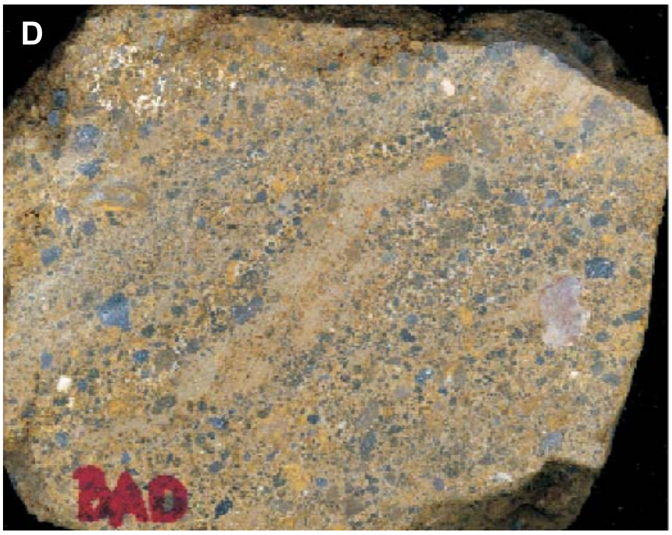
Handspecimen of a lapilli tuff from Hármas-hegy, characteristically yellow, to light brown in colour and rich in volcanic glass lapilli in a quartzofeldspathic matrix. The shorter side of the picture is about 10 cm



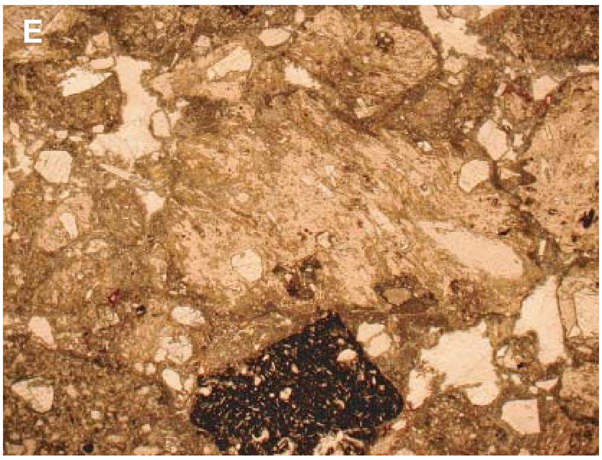
Photomicrograph [plan parallel polarized light] of a lapilli tuff recovered from the Hármas-hegy rich in volcanic glass shards with variable microvesicularity. The photo is ~ 2 cm across



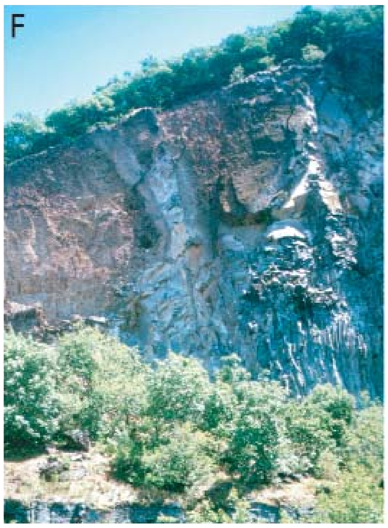
Photomicrograph (parallel polarized light) of a lapilli tuff from the Hármas-hegy near Badacsony that shows a tachylite glass shard (black vesicular clast in the left hand side), that entrapped siliciclastic and/or volcaniclastic mud/ash indicating a premixing and possible recycling of pyroclast through repeated eruptions from the same vent of a "wet" volcano. The short side of the photo is about 2 mm



Handspecimen of a lapilli tuff from Badacsony that is unsorted, matrix-rich and having volcanic glass shards with different shape and vesicularity. The matrix is rich in quartzofeldspatic-origin minerals and rock fragments. The short side of the photo is about 10 cm



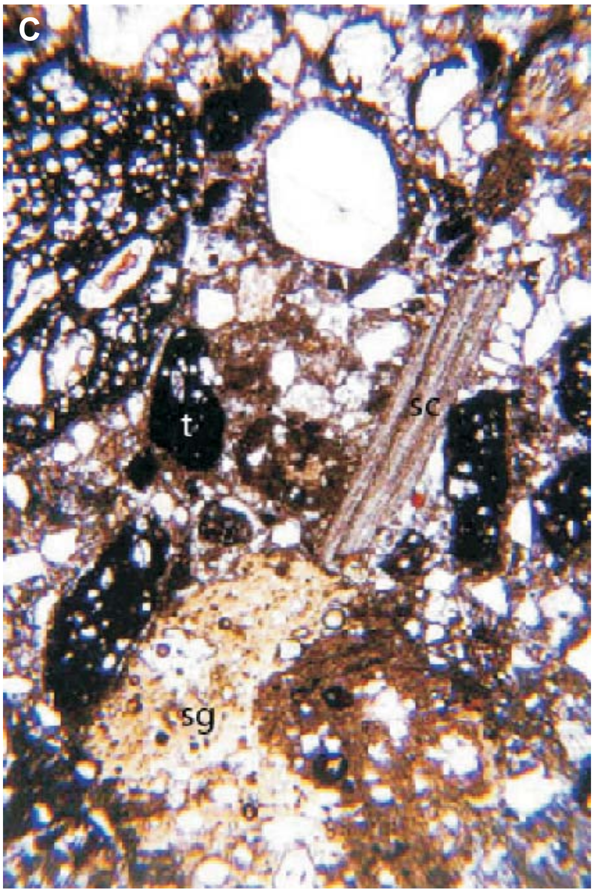
Photomicrograph of a lapilli tuff from Badacsony, containing elongated, blocky, moderately vesicular volcanic glass shards. Broken pyrogenic and xenocrysts are characteristic constituents of the pyroclastic rocks at Badacsony. The short side of the photo is 2 mm



Irregular shaped scoriaceous units that have been intruded by dykes from the Badacsony lava flows showing irregular margins

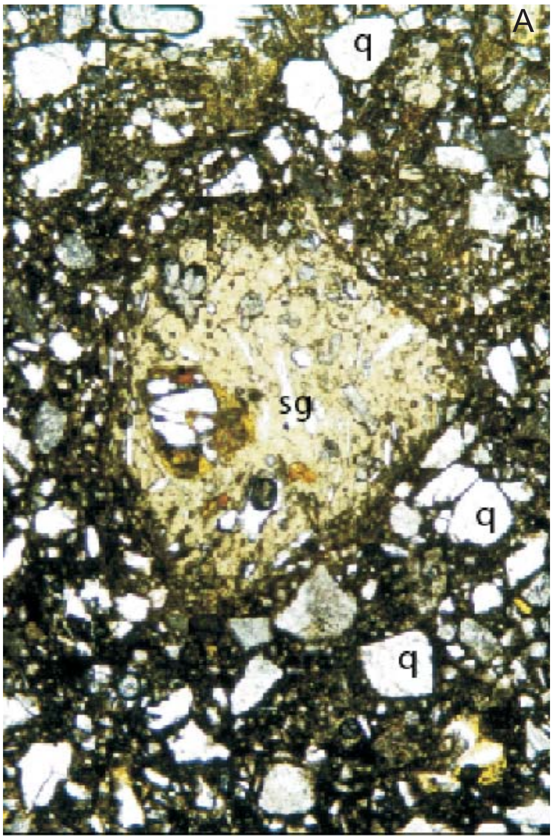
Steep bedded pyroclastic succession capping the Vár-hegy in the north-western margin of the Szigliget area, p = pyroclastic rock units, d = basanite dyke, n = Neogene subhorizontal siliciclastic rock units

Pyroclastic succession of Unit 2 in the southern hill of Szigliget (Kamon-kő). The unit is composed of bedded, unsorted, accidental lithic rich lapilli tuff that contain a large amount of clasts from deep regions such as schist, meta-volcanites, red sandstones and clasts from other lithologies of unknown origin

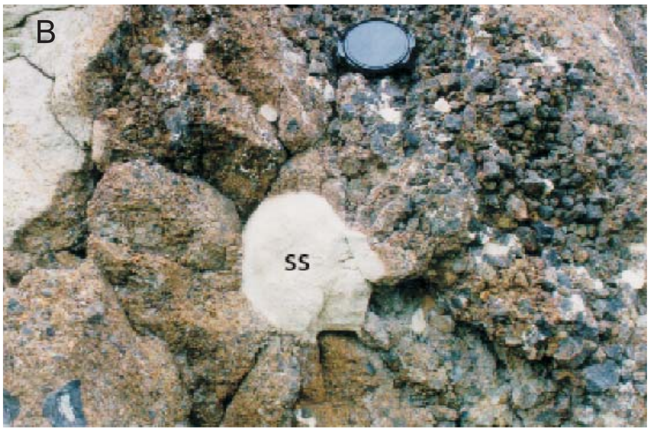


Photomicrograph of a lapilli tuff from the southern hill of Szigliget (Kamon-kő), that contains schist (elongated angular clast in the centre of the picture – sc) and blocky volcanic glass shard (in the left side of the picture – sg) and tachylite (t) cemented by calcite. The short side of the photo is about 4 mm

Bedded-to-massive lapilli tuff and tuff from Unit 3 of Szigliget in the top section of the north-western Vár-hegy. Between bedded lapilli tuff units there are massive lapilli tuff beds that contain irregular shape quarztofeldspathic clasts up to dm-size (light colour fragments in centre of view – marked by lines)



Photomicrograph of a lapilli tuff from Szigliget Unit 3 in the topmost section of the Vár-hegy. The lapilli tuff is rich in blocky (centre part of the picture – sg) to fluidally shaped, moderately vesicular volcanic glass shards hosted in a quartz-rich (bright angular clasts – q) matrix. The short side of the photo is 1 mm

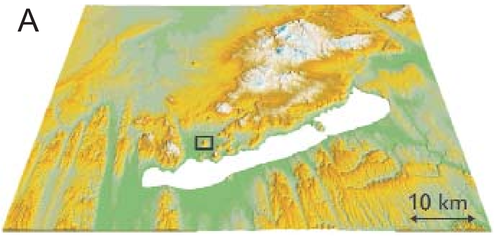


Large (dm-size), often rounded Neogene sedimentary clasts (creamy coloured clast in the centre of the photo – ss) of Unit 3 in the topmost part of the pyroclastic succession of the Vár-hegy

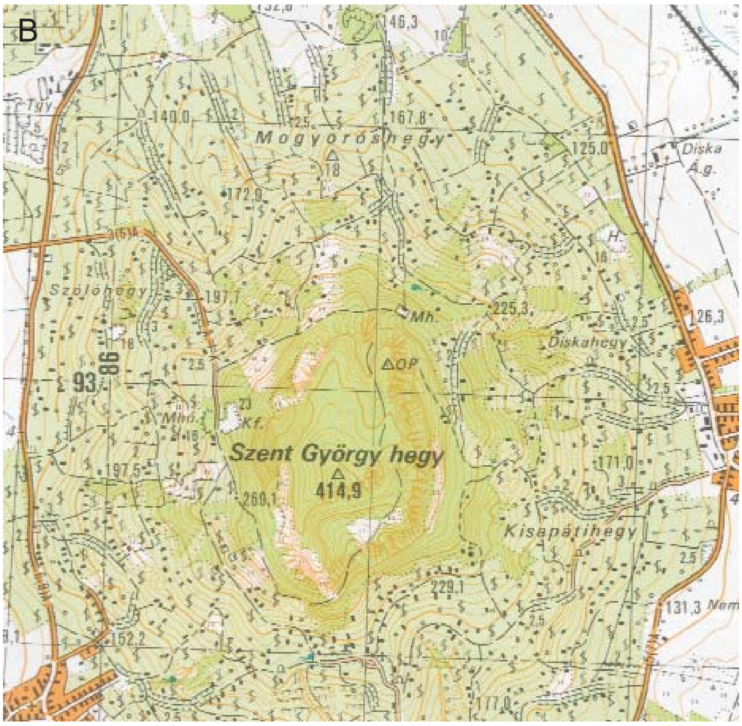
Steep dipping pyroclastic beds dip toward north-west in the eastern cliff of the Vár-hegy of Szigliget, suggesting syn-eruptive remobilization of tephra on a steep flanks (possible inner) of a phreatomagmatic volcano

Inverse graded, lenticular lapilli stone bed cemented by calcite from the uppermost outcrops of the Vár-hegy of Szigliget

Abraded lapilli tuff fragment as a clast (outlined) in a lensoid steep bed from Vár-hegy, indicates some degree of reworking and remobilization of tephra upon formation of this succession

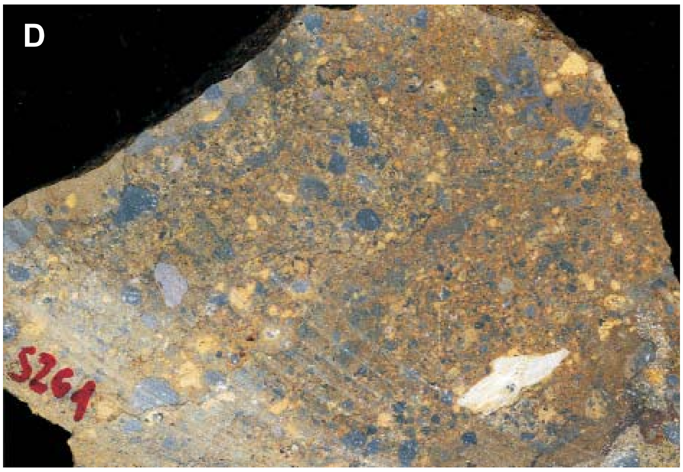


Location of the Szent György-hegy on DTM

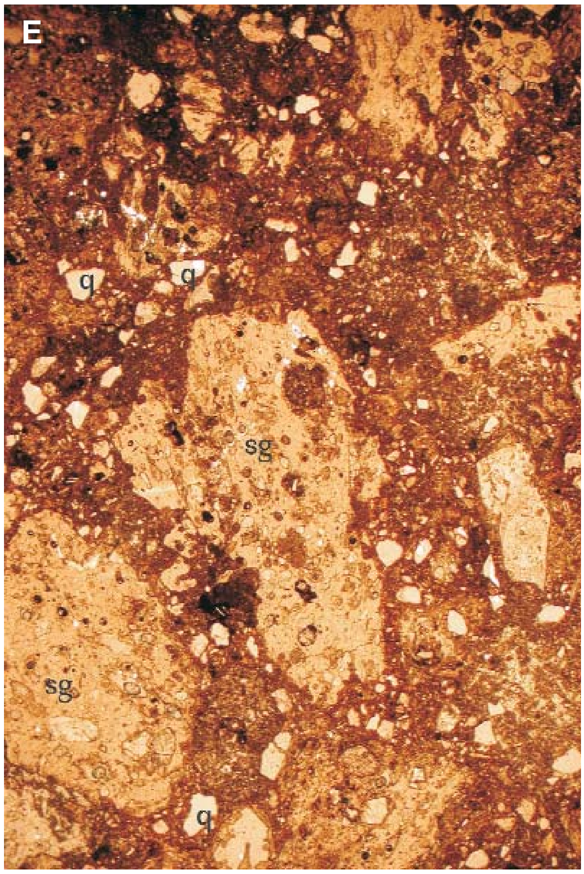


View to Szent György-hegy from south. Note the capping irregular shape structure of the hill above the grape yards. This zone is a spatter rich scoriaceous succession that was invaded by rosette-like columnar jointed basanite

Detailed topographic map of the Szent György-hegy erosional remnant. Note the north to south elongated erosional remnant. Map data is from the 1 to 25,000 scale topographic map series of Hungary. Rectangular grid has 1 km spacing

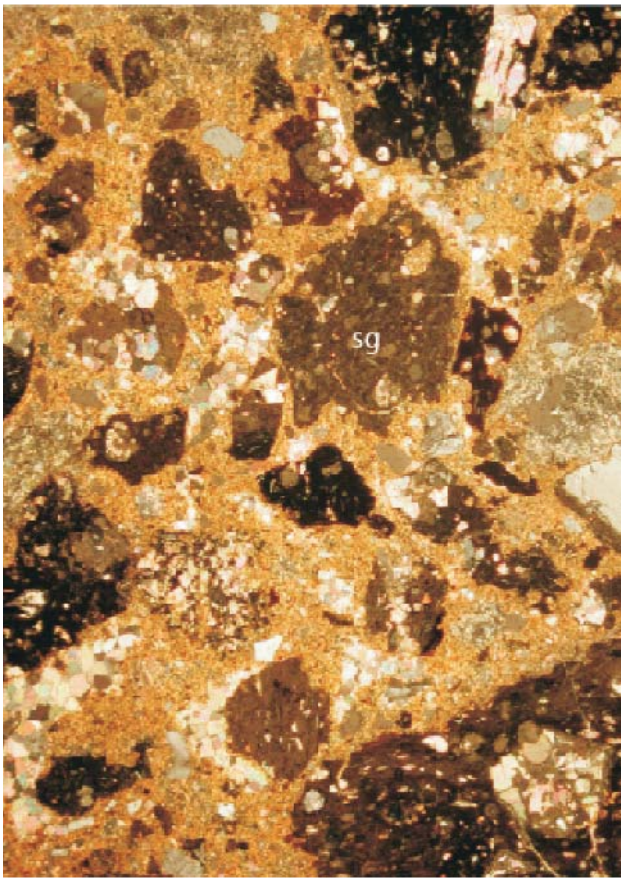


Handspecimen of a phreatomagmatic lapilli tuff from Szent György-hegy. The black lapilli are chilled pyroclasts. The short side of the photo is 10 cm

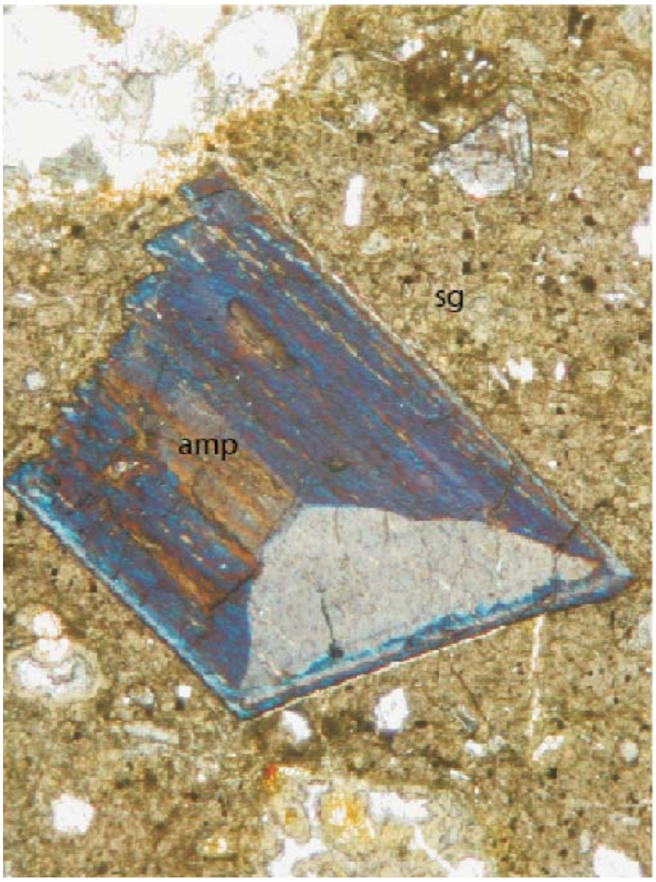


A photomicrograph of the same lapilli tuff shown on Plate 3.23, D from Szent György-hegy show a matrix supported texture with blocky, slightly vesicular volcanic glass shards (sg). Note the large amount of quartz fragments (q). The short side of the photo is about 4 mm

Plate 24 / Chapter 3 Second International Maar Conference — Hungary–Slovakia–Germany



Photomicrograph of moderately vesicular sideromelane glass shards (dark angular shards – sg) from diatreme filling massive lapilli tuff of Vár-hegy, Balatonboglár. The lapilli tuff is calcite cemented. The short side of the photo is about 6 mm. [cross polarized light]



The volcanic glass shards (sg) of the lapilli tuff from Boglár often show a trachytic texture and contain amphiboles (amp). The short side of the photo is about 1 mm. [parallel polarized light]

Lapilli tuff succession with silicified wood fragment (circle) in the western basal layer of the Temető-domb at Boglár

The massive lapilli tuff bearing fossil wood fragments is rich in rounded, dense silt- and sandstone clasts (circle)