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Mio/Pliocene Phreatomagmatic Volcanism in the Western Pannonian Basin

by

Ulrike Martin
and
Károly Németh

BUDAPEST, 2004

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DTP:
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Preface

On the occasion of the Second International Maar Conference held in Hungary in September 2004, the Geologica Hungarica, series Geologica *presents this monograph on the Mio/Pliocene (8-2 My) small volume intraplate alkaline volcanism in the western Pannonian Basin. The volcanic activity occurred in the Bakony – Balaton Highland Volcanic Field (BBHVF) just north of Lake Balaton, in the smaller Little Hungarian Plain Volcanic Field (LHPVF) just to the north, and in the Styrian Basin Volcanic Field barely reaching into westernmost Hungary. The western Pannonian Basin is underlain by Neogene siliciclastic sediments which overlie Mesozoic karstic limestones which in turn overlie crystalline basement rocks. As volcanism was active during and after deposition of the Neogene siliciclastic sediments, volcanicity was largely synsedimentary and consequently effected, more or less, by the unconsolidated water saturated sediments. The volcanic fields of the western Pannonian Basin will be visited during two identical volcanological field trips run prior and after the conference.*

In this monograph the present state of physical volcanological research (over the last 10 years) on the volcanism of the western Pannonian Basin is presented. The authors of the several papers present the relevant details and interpretations of the regional geology, of the volcanic fields and also of the many individual volcanoes and their various phreatomagmatic and magmatic eruption styles. In addition, the authors compare the western Pannonian volcanic fields with other volcanic fields in the world many of which they know from personal aquaintance and studies. The list of references contains not only very informative Hungarian publications but also a wealth of international publications relevant to the understanding of volcanological processes generating maars, tuff-rings, diatremes, scoria cones, lava lakes and their tephra and volcanic rocks, but also relevant to the understanding of volcanic fields.

The overview publications and detailed descriptions of the individual volcanoes of the three volcanic fields provide the reader with a state of the art view on the volcanicity of the Neogene volcanism of the western Pannonian Basin which for many decades has been almost unknown to volcanologists from many countries. The monograph has been organized in such a way that it contains both the overview publications but also the publications of the individual volcanoes and their outcrops so that the publications can easily be used as field guides for volcanological field trips of groups but also for individuals. The various levels of erosion of the many individual volcanoes make the volcanic fields of the western Pannonian Basin very informative in respect to research of maars, diatremes and lava lakes, and complimentary to other volcanic fields displaying monogenetic volcanoes exposed at different levels of erosion, as, e.g. the maar volcanoes visited on the occasion of the First International Maar Conference in the West Eifel Volcanic Field in 2000. This monograph is also complimentary to the field guide published on the volcanic field in southern Slovakia, also visited prior and after the Second International Maar Conference.

Würzburg, 15th of June 2004.

PROF. DR. VOLKER LORENZ

Institute for Geology University of Würzburg,
Würzburg, Germany

Volcanology and its importance in Hungary

Magmatism and volcanism played an important role in the geological development of Hungary. The formation of acidic rocks due to volcanic activity during the Variscan cycle can be found in the Velence–Balaton and the Mecsek area south-west of Hungary. Typical "Pietra verde" type pyroclastic rocks, displaying a distinctively alkaline-trachytic character, are intercalated in the Middle Triassic carbonat sequence in the South Bakony Mts north of Lake Balaton. Middle Triassic to Jurassic basic and ultrabasic suites in different tectonic units of NE Hungary are commonly intercalated with deep-sea sediments.

The Tisza Tectonic Megaunite south-east of the Hungary is related to basic-alkaline volcanism in the Early Cretaceous. Volcanic rocks are best exposed in the Mecsek Mts and they extend as far as to the basement of the Great Plain. In the Eocene andesitic-type volcanism occurred both in the Transdanubian Range and the Mecsek Mts.

Intermediate volcanism during the Miocene is of particular importance. In Hungary occurrences of Miocene volcanic rocks are distributed in a broad belt from Dunazug–Börzsöny Mts to the Zemplén Mts, basically from Budapest to the north-east tip of Hungary. Furthermore volcanic rocks are covered under thick Neogene to Quaternary sediments in the northern part of the Great Plain, which form all together the innermost zone of the volcanic belt of the Inner Carpathians. Particularly important are metallic ore deposits of Hungary that are associated with this Miocene volcanic belt. The latest volcanic activity, occurred in several regions of Hungary, and has been dated to be of Mio/Pliocene age. One of the most important volcanic regions in relation to the Second International Maar Meeting 2004 is the western Pannonian Basin, also because of the scenic impression of individual volcanic mountains in the Tapolca Basin, west of the Bakony – Balaton Highland area.

Volcanological research in Hungary has of long tradition, first of all because of the old mining of metallic ore deposits related with the Neogene volcanism. Professor József Szabó (1822–1894) is know well in the scientific community as "father of Hungarian geology", the first teaching geology in Hungarian language at the university. He established an internationally recognized petrographic system of trachytic rocks.

For long time after Szabó's activity the descriptive petrology prevailed in the Hungarian geology. In addition, the contribution of Lajos Lóczy sen. has a great importance. He studied the morphological development of the Balaton Highland, in particular, the basaltic volcanoes of the Tihany Peninsula.

Since the late 1970's, early 1980's new trends of volcanological studies are appeared in Hungary, focusing on the Neogene intraplate basalt volcanism. On one hand petrogenetic studies, based on detailed mineralogical, chemical and isotope geochemical investigations, have been carried out. On the other hand investigations and new theories tried to support interpretation of tectonic, geodynamic position and significance of the volcanism in the geological evolution of the area.

A new detailed morphogenetic approach and the theory of phreatomagmatic volcanism conducted by Ulrike Martin and Károly Németh gave way to a new view in interpreting basaltic volcanism of the western Pannonian Basin. We strongly believe that the IAVCEI–CVS–IAS Second International Maar Conference held in Hungary jointly organised with Slovakia and Germany will promote both the scientific and the educational activity in the field of volcanology in the region and worldwide.

Budapest, 14th of July 2004.

Dr. Károly Brezsnyánszky

Director
Geological Institute of Hungary,
Budapest, Hungary

Introduction

In the western Pannonian Basin there are three distinct areas where during Mio/Pliocene time alkaline basaltic volcanism took place (front inner cover). Volumetrically the largest volcanic field, the Bakony – Balaton Highland Volcanic Field (BBHVF) is located adjacent to the north shore of the Lake Balaton (front inner cover). There, the vent remnants are advanced in erosion down to their crater/vent zone and give a good opportunity to study the feeding systems of predominantly phreatomagmatic volcanoes. North of the BBHVF, in a more dispersed setting, less vents form the Little Hungarian Plain Volcanic Field (LHPVF — front inner cover). The erosion level of these volcanoes allows studying the crater zone of predominantly phreatomagmatic volcanoes. In the western margin of the Pannonian Basin, a cluster of strongly eroded alkaline basaltic volcanoes form the Styrian Basin Volcanic Field (SBVF), a term which is generally used for volcano clusters located in Burgenland, Styria (both in Austria and in Slovenia). In this book the present state of research in the field of physical volcanology will be presented predominantly focusing on the BBHVF and LHPVF. The book reflects a summary of the results and volcanological view of the authors on the basis of their past ten years research on understanding the eruptive mechanism of the Mio/Pliocene volcanoes in the western Pannonian Basin.

The book starts with a general review of the Neogene small-volume intraplate volcanism in the western Pannonian Basin, with respect to other fields in the Pannonian Basin system of the same age and to other fields world-wide. The general review put the volcanism in a tectonic and geochemical framework based on the relevant literature. This information allows to create a general volcanological model, which is based on the authors own results. The general review of the volcanism is followed by a summary of the geomorphological aspects of the study of the volcanic fields in this region. This is a "stand alone" summary of the authors view of the syn-volcanic morphology and the possible volcanic landforms on the basis of studies of the preserved volcanic rocks. The chapter is followed by systematic descriptions and interpretations of the BBHVF, the Keszthely Mts. region and the LHPVF with a brief description of a few sites from the SBVF. The reason to present the BBHVF and the Keszthely Mts region separately is, that the latter is predominantly characterised by preserved intrusive complexes. Their description therefore thematically could be well separated from the BBHVF. The specific chapters of the volcanic fields are followed by concluding remarks.

Alongside a monographic style of the book, the chapters and the listed sites are arranged in a way that they are in concert with the relevant field trips organised before and after the Second International Maar Conference (2IMC) 2004.

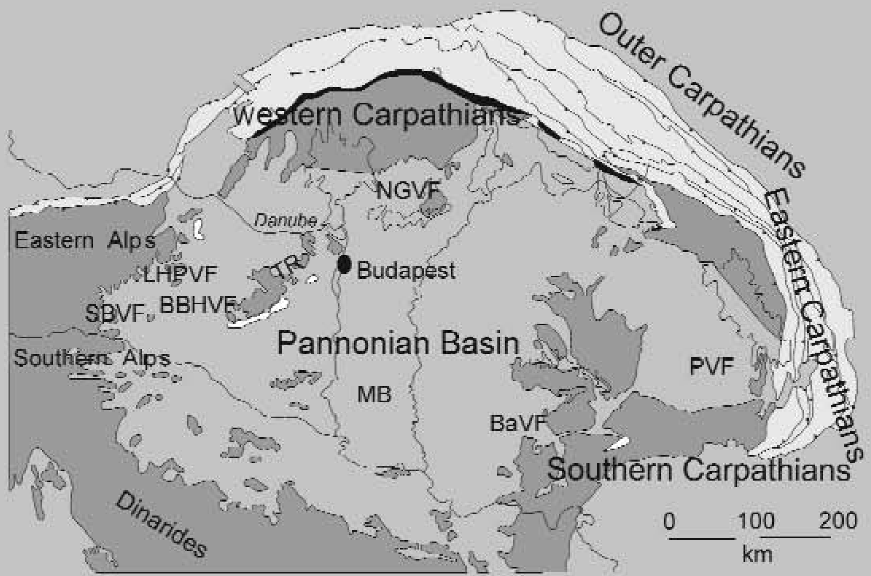
The 2IMC an international volcanological meeting under the auspices of the International Association of Volcanology and Chemistry of the Earth Interior (IAVCEI) and the International Association of Sedimentologists (IAS) and also supported by other scientific organisations such as the Society for Economy Geology (SEG), Geologische Vereinung (GV), Deutsche Geologische Geselschaft (DGG), Magyarhoni Földtani Társulat (MFT) hosted in Hungary in 2004 was a joint meeting of Hungary, Slovakia and Germany.

There is no direct reference to the certain field trip stops in the main text of the chapters to keep the scientific information of certain volcanic fields confined and relevant. The field trip programmes are presented in separate hard page (as a page marker plate) with references of the single stops to the related chapters. Enclosed at the back inner cover are 3D digital terrain models which highlight the field trip routes. Numbers on the maps correspond to page numbers of the related pages in the chapters of the monograph.

ULRIKE MARTIN and KÁROLY NÉMETH
Heidelberg–Balatonlelle

geologica hungarica
SERIES GEOLOGICA, TOMUS 26, pp. 12–56

Phreatomagmatic volcanic fields in a Mio/Pliocene fluvio-lacustrine basin, western Pannonian Basin, Hungary: a review



Contents

Abstract

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

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

Introduction

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

The BBHVF itself alone has approximately 50 basaltic volcanoes in a relatively small (around 3500 km^2) area (JUGOVICS 1969a), however, the number of vents maybe far more than 50 due to the existence of volcanic complexes and nested volcanoes (MARTIN et al. 2003). The volcanic erosional remnants form a more scattered distribution in the LHPVF, and the number of volcanoes is less (~10) than in the BBHVF (JUGOVICS 1915, 1916, 1972), however, volcanic erosional remnants buried under thick Quaternary deposits are known from the region (TÓTH 1994). Individual Neogene alkaline basaltic volcanic erosional remnants are located close to the triple border of Hungary, Austria and Slovenia as well as close to the eastern metamorphic core complexes in the Eastern Alps (JUGOVICS 1916, 1939, KRALJ 2000) together often referred as SBVF.

The volumetrically largest field of all is the BBHVF, often mentioned together with the LHPVF due to similarities in the age, timing and eruption mechanism. The BBHVF belongs to the Transdanubian Range unit, which is correlated with the Upper Austroalpine nappes of the east Alpine orogen (MAJOROS 1983, KÁZMÉR and KOVÁCS 1985, TARI 1991). The underlying basement of the volcanic fields consists of Palaeozoic rocks (Silurian schist, Permian red sandstone – CSÁSZÁR and LELKESNÉ-FELVÁRI 1999) and a thick Mesozoic carbonate sequence (BUDAI and VÖRÖS 1992, BUDAI and HAAS 1997, HAAS and BUDAI 1999, HAAS et al. 1999). This basement forms a large-scale anticline structure of Eoalpine origin (Figure 1.3) in the Transdanubian Range area and is locally covered by Tertiary sediments (TARI et al. 1992, 1999, HORVÁTH 1993, SACCHI and HORVÁTH 2002).

Tertiary sediments were deposited in local sedimentary basins on a regional erosional unconformity (MÜLLER and MAGYAR 1992, MÜLLER 1998, TARI and PAMIC 1998, JUHÁSZ et al. 1999, MÜLLER et al. 1999). In the Neogene, just shortly before the volcanism started, a large lake, the Pannonian Lake, occupied main parts of the Pannonian Basin (Figure 1.4.), which had a very colourful sedimentary environment as reflected in the irregular basin morphology

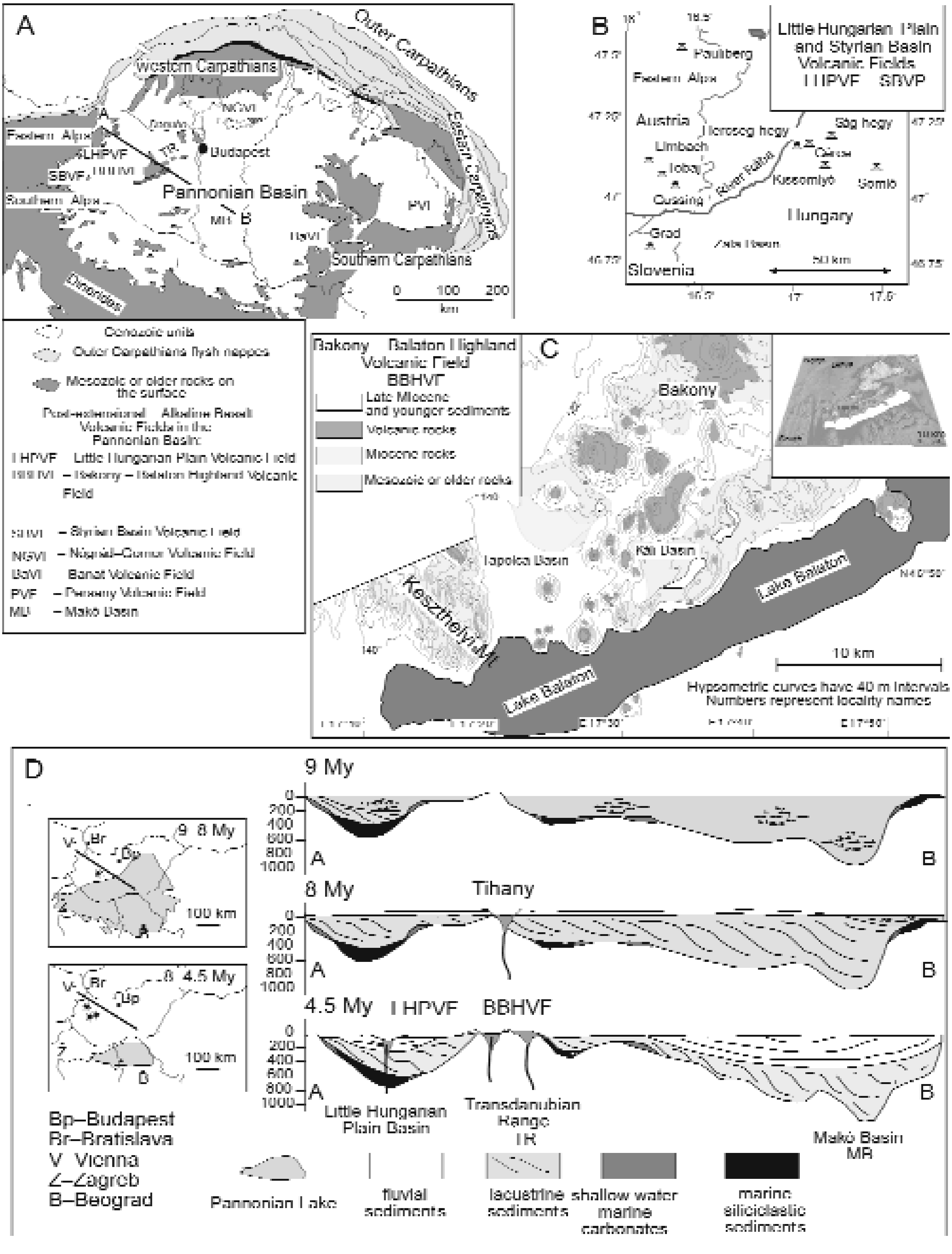


Figure 1.1. Structural elements of the Carpatho-Pannon region

(A) with respect to the Mio/Pliocene alkaline basaltic volcanic fields. The location of the volcanic erosional remnants of the Little Hungarian Plain and Styrian Basin Volcanic Fields (B) and the Bakony – Balaton Highland Volcanic Field (C) are shown in relevance to other rock rock units. The western Pannonian Basin was occupied by the Pannonian Lake around 9 My (D); that lake gradually vanished and had only small basins in the southern part of the Pannonian Basin 4.5 My ago (D). During onset of the volcanism, the region was occupied by an alluvial plain with a flat morpholohy, and potentially with large, but shallow water masses (D). Please note: The model on "D" is modified after MAGYAR et al. (1999)

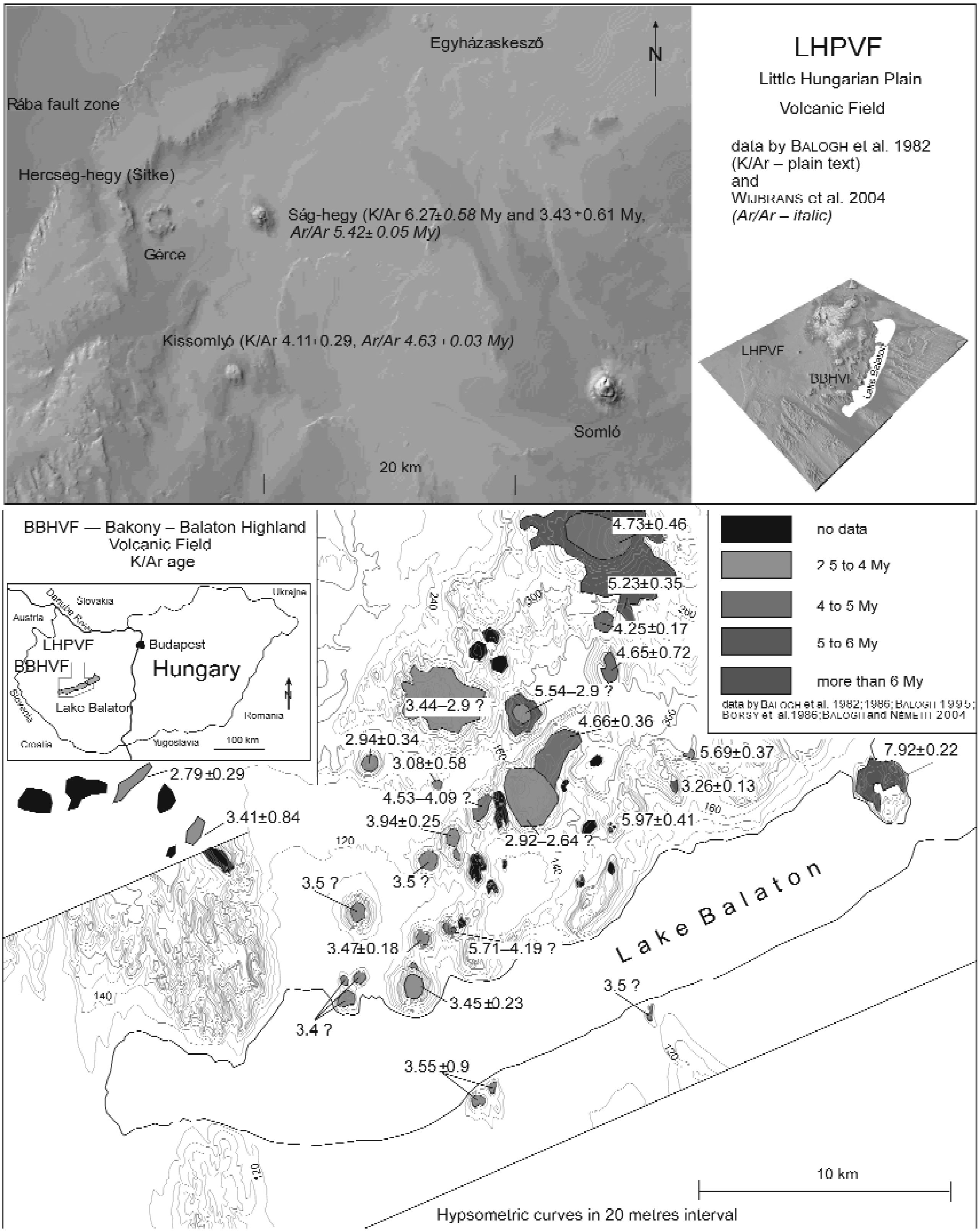


Figure 1.2. K/Ar age distribution map of the Little Hungarian Plain (A) and the Bakony – Balaton Highland (B) Volcanic Fields (after BALOGH et al. 1982, 1986, BORSY et al. 1986, and BALOGH and NÉMETH 2004)

(Kázmér 1990). The lacustrine sandstones, mudstones, marls of the brackish Pannonian Lake are widespread in the Pannonian Basin (Jámbor 1980, 1989, Kázmér 1990, Müller and Magyar 1992, Müller 1998, Gulyás 2001). Just before volcanism started the area in the western Pannonian Basin formed an alluvial plain with unconsolidated water-saturated sediments in significant spatial and temporal variation (Kázmér 1990).

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

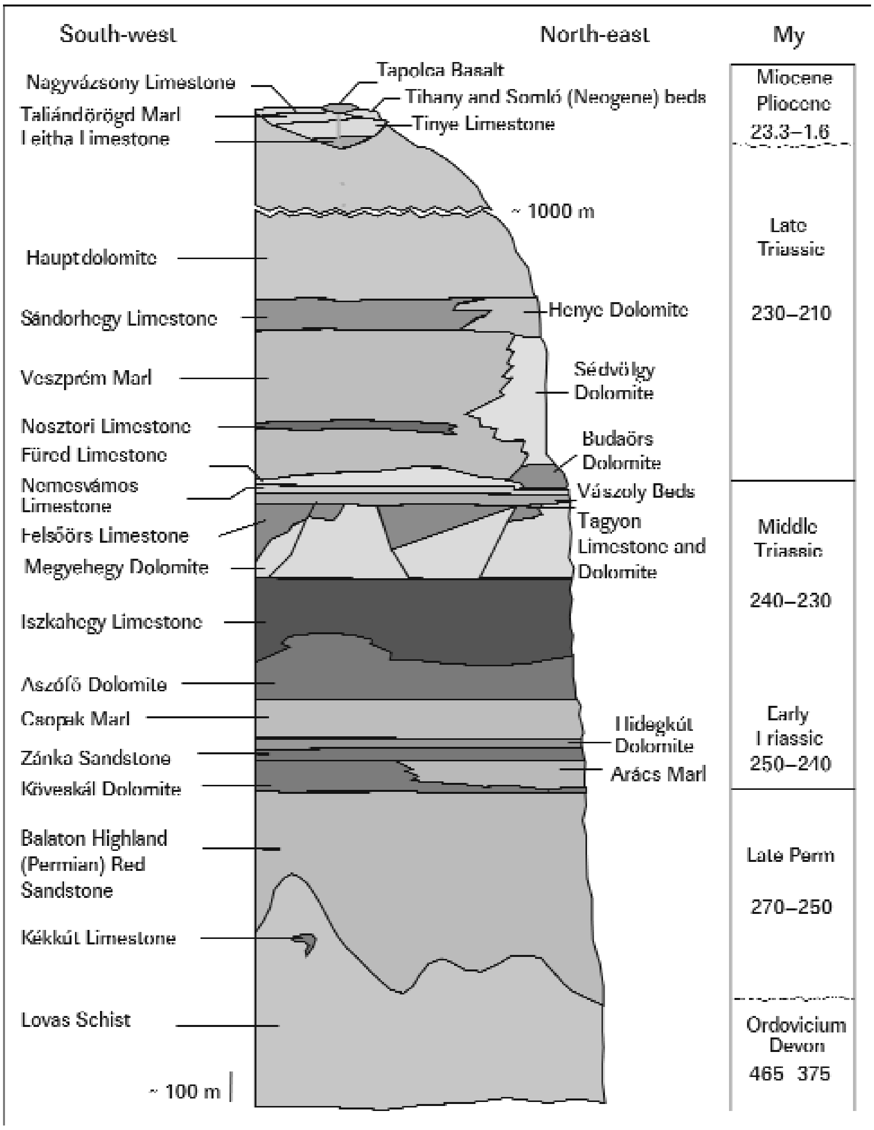


Figure 1.3. A general stratigraphy of the southern part of the Transdanubian Range (after Budai et al. 1999). The section shows all the identified rock formations and also gives a hint of the approximate location of certain rock formations

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

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

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

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

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

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

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

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

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

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

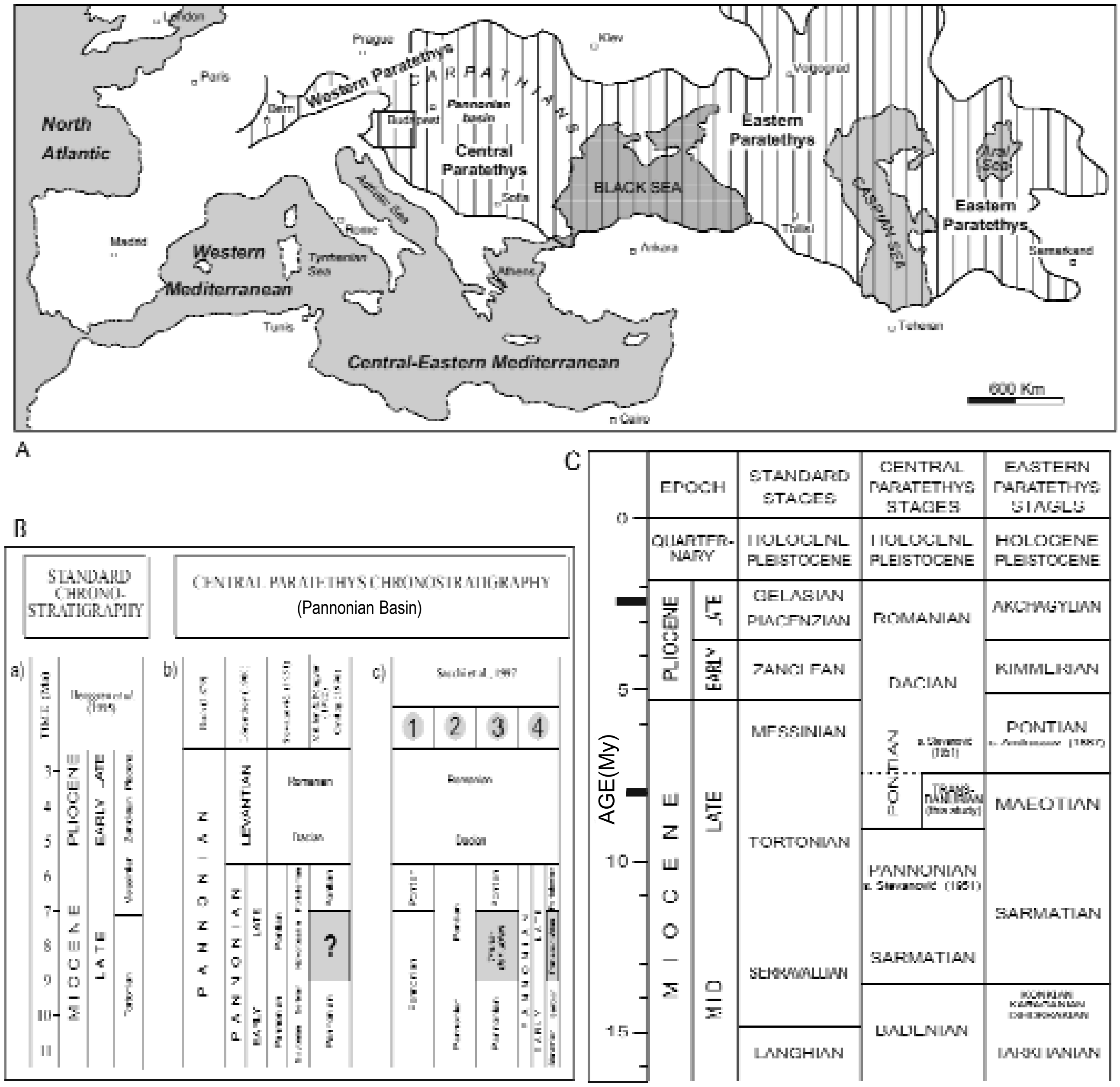


Figure 1.4.

(A) Outline of the Paratethys–Mediterranean region during Late Miocene (after Müller et al. 1999). Diagram is by the authors' permission.

(B) Correlation between the Late Neogene (a) Mediterranean and (b) Central Paratethys stages. Column (c) illustrates four possible redrawings of the Late Miocene chrono-stratigraphic framework of the Pannonian Basin (Central Paratethys – from Sacchi et al. 1997). Numbers on "c" represent possible solutions of correlation. Diagram is after Sacchi and Horváth 2002 with the authors' permission.

(C) Mediterranean and Paratethys stages for the last 15 My (after Rögl 1998, and Magyar et al. (1999). Transdanubian is referred after (Sacchi et al. 1999a) as an intermediate stage (or substage) between the Pannonian s.str. (s. Stevanovic1951) and the Pontian s.str. (Andrussov 1887). The Transdanubian substitutes the lower part of Pontian s. Stevanovic (1951). The interval of Neogene volcanism in the western Pannonian Basin is marked by two thick black lines on the left side of the diagram. The diagram is after Sacchi and Horváth 2002, with the authors' permission

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

Starting in the latest Miocene (Figure 1.4.), a considerable amount of basaltic magma erupted in the Transdanubian Range (TR) and Balaton Highland area in the Pliocene (Szabó et al. 1992, Németh and Martin 1999c, Németh et al. 2000). A total magma output has been estimated for the BBHVF on the basis of volume estimates of coherent lavas and dense rock equivalents of juvenile pyroclasts from phreatomagmatic and magmatic explosive eruptive products. This gave a minimum estimate of ~2.7 km^3 and a maximum estimate of ~6.5 km^3 total magma output (Németh et al. 2000). A realistic estimate of 4±0.5 km^3 of total magma output (Németh et al. 2000) over ca. 5.7 My time indicates that the BBHVF rather belongs to a volcanic field with low magma output ratio, which is typical for a strike-slip tectonic regime, or regions of moderate lithospheric extension such as the San Francisco Volcanic Field, Arizona. Approximately 90% of the total magma output is estimated to be basanitic in composition, leaving 10% for more differentiated rock types such as tephrite, phonotephrite, that are the major constituents of the phreatomagmatic pyroclastic deposits (NÉMETH et al. 2000). Estimating of the erupted volume of juvenile material from pyroclasts is more complicated due to the uncertainty of the eroded volumes. The tectonic role and style of magmatism is still under debate (SZABÓ et al. 1992, EMBEY-ISZTIN et al. 1993, HARANGI et al. 1995, KEMPTON et al. 1997). In general most of the workers agree that the Neogene alkali basalts from western and central Europe dominantly derived from asthenospheric partial melting (EMBEY-ISZTIN et al. 1993). However, there are growing evidences that in most cases they were modified by melt components from the enriched lithospheric mantle through which they have ascended (EMBEY-ISZTIN et al. 1993). A debate on the role of crustal contamination of the magmas still exists. Most of the models deal with some contamination from a subducted slab from former subduction in the region and/or some degree of metasomatozism (BALI et al. 2002). In general, the incompatible-element patterns of the lavas of the volcanic fields of the western Pannonian Basin show that these lavas are relatively homogeneous (EMBEY-ISZTIN et al. 1993) and are enriched in K, Rb, Ba, Sr, and Pb with respect to average ocean island basalt such as Hawaii, and resemble alkali basalts of Gough Island.

Heat flow values in the Pannonian Basin greatly vary (DÖVÉNYI and HORVÁTH 1988), and are up to 300 mW/m^2 in areas of volcanic fields that are inferred to be located above shallow magma chambers (SACHSENHOFER et al. 1997, 1998, 2001). Elevated heat flow values from the Austrian basins were obtained from areas related to the Eastern Alps, a consequence of rapid uplift of the whole orogen (SACHSENHOFER et al. 1997, 2001, SACHSENHOFER 2001).

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

Lithospheric structure and magma ascent of Neogene alkaline basaltic volcanic fields

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

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

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

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

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

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

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

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

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

1. Early Miocene mainly acidic explosive volcanism that led to the accumulation of extensive ignimbrite sheets (welded and non-welded type, predominantly rhyolitic in composition), block-and-ash flow deposits as well as their

reworked volcano-sedimentary units, often intercalated with normal marine to terrestrial sediments (LIFFA 1940, PANTÓ 1963, 1966, HÁMOR et al. 1980, SZÉKY-FUX and KOZÁK 1984, PÓKA 1988, CAPACCIONI et al. 1995, SZAKÁCS et al. 1998, LEXA 1999). The recent spatial distribution of this volcanic province exhibits a great separation which is inferred to be the result of a large right lateral displacement along the Mid-Hungarian shear zone (Figure 1.1) during the Early Miocene (STEGENA et al. 1975, BALLA 1980, 1981, ROYDEN et al. 1983a).

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

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

There is a general agreement regarding the petrogenesis of Neogene alkali basaltic rocks in the Pannonian Basin. The basaltic rocks were formed during the Late Cenozoic post-orogenic phase and their eruption was related to the evolution of the extensional Pannonian Basin following Eocene–Miocene subduction and its related calc-alkaline volcanism (SZABÓ et al. 1992, EMBEY-ISZTIN et al. 1993). The alkaline volcanic centres, dated by K/Ar methods are between ~12 and 1 My in age (PÉCSKAY et al. 1995), forming well-distinguishable volcanic fields. Some fields are near the western (Graz Basin, Burgenland, Slovenia), northern (Nógrád–Gömör/Gemer), and eastern (Eastern Transylvanian Volcanic Field) margins of the basin, but the majority are concentrated near the Transdanubian Range (Bakony – Balaton Highland and Little Hungarian Plain Volcanic Field). Coherent lavas range from slightly hy-normative transitional basalts through alkali basalts and basanites to olivine nephelinites. No highly evolved coherent lava (extrusive nor intrusive) compositions have been identified from any of the locations yet (SZABÓ et al. 1992, EMBEY-ISZTIN et al. 1993, HARANGI et al. 1995). This makes the Pannonian Neogene basaltic volcanic fields different from other European volcanic fields such as the Eifel, where e.g. phonolitic lava flows as well as ignimbrites are common (BOGAARD and SCHMINCKE 1985, FREUNDT and SCHMINCKE 1986, BEDNARZ and SCHMINCKE 1990, HARMS and SCHMINCKE 2000). The presence of mantle peridotite xenoliths, xenocrysts, and high-pressure megacrysts in coherent lavas, even in the slightly more evolved ones and in pyroclastic rocks, is inferred to indicate that differentiation took place within the upper mantle (DOWNES et al. 1992). However, the mantle source often is referred to be heterogeneous (DOBOSI 1989, DOBOSI et al. 1991, DOBOSI and FODOR 1992, SZABÓ and BODNAR 1995, 1998, DOBOSI et al. 2003). The study of peridotite xenoliths revealed a strong relationship between deformation and temperatures of peridotites, in as much as coarse-grained protogranular and poikilitic xenoliths had high temperatures (up to 1175 °C), whereas fine-grained equigranular and mosaic xenoliths had low temperatures (800–900 °C — EMBEY-ISZTIN et al. 2001). This picture suggests that diapiric uplift of hot mantle material into a cooler uppermost mantle has probably taken place (EMBEY-ISZTIN et al. 2001).

Isotope geochemistry

The Sr and Nd isotope ratios from the Neogene coherent lava flows of the Pannonian Basin span the range of Neogene alkali basalts from Western and Central Europe (DUDA and SCHMINCKE 1978, 1985, MERTES and SCHMINCKE 1985, BEDNARZ and SCHMINCKE 1990), and suggest that the magmas of the Pannonian Basin dominantly derived from asthenospheric partial melting. Pb isotope studies, however, indicate that in most cases the astenospheric melt composition was modified by melt components from the enriched lithospheric mantle through which the magma ascended (EMBEY-ISZTIN et al. 1993). Various metasomatic processes may have interacted with the uprising melts (BALI et al. 2002), similarly to other alkaline volcanic provinces in Central Europe (WITTEICKSCHEN et al. 1993, SHAW 1997, SACHS and HANSTEEN 2000, SHAW and EYZAGUIRRE 2000). Delta ^{18}O values indicate that the magmas have not been significantly contaminated with crustal material during ascent and isotopic and trace-element ratios therefore reflect mantle source characteristics (EMBEY-ISZTIN et al. 1993). The uniform oxygen isotope ratio in the phenocrysts suggests that the mantle source of the alkali basalts was also homogeneous with respect to its oxygen isotope composition, which is in contrast to the relatively wide variation of Sr, Nd and Pb isotope ratios in the source (DOBOSI et al. 1998). Variations in radiogenic isotope compositions in the basalts have been interpreted as result of the interaction of subduction-related fluids with the mantle source of the basalts. If this was the case, then the fluids, which caused significant changes in the Sr and Pb isotope ratios of the mantle source, did not noticeably modify its oxygen isotope composition. Incompatible-element patterns show that the basic lavas, which erupted in the Balaton area and Little Hungarian Plain, are relatively homogeneous and are enriched in K, Rb, Ba, Sr, and Pb with respect to average ocean island basalt, and resemble alkali basalts of Gough Island type (EMBEY-ISZTIN et al. 1993). In addition, $^{207}Pb/^{204}Pb$ is enriched relative to $^{206}Pb/^{204}Pb$. In these respects, the lavas of the Balaton area and the Little Hungarian Plain differ from those of other regions of Neogene alkaline magmatism of Europe (EMBEY-ISZTIN et al. 1993). This may be due to the introduction of marine sediments into the mantle during the earlier period of subduction and metasomatism of the lithosphere by slab-derived fluids rich in K, Rb, Ba, Pb, and Sr. Lavas erupted in the peripheral areas have incompatible-element patterns and isotopic characteristics different from those of the central areas of the basin, and more closely resemble Neogene alkaline lavas from areas of western Europe where recent subduction has not occurred (EMBEY-ISZTIN et al. 1993). However, in this respect there is no agreement yet. The alkaline volcanic activity that occurred in the Persani Mountains (eastern Transylvanian Basin) and Banat (eastern Pannonian Basin) regions of Romania between 2.5 My and 0.7 My (DOWNES et al. 1995b) produced coherent alkaline basaltic lavas that are primitive, silica-undersaturated alkali basalts and trachybasalts (7.8–12.3 wt.% MgO; 119–207 ppm Ni; 210–488 ppm Cr), which are LREE-enriched (DOWNES et al. 1995b). Mantle-normalised trace-element diagrams revealed an overall similarity to continental intraplate alkali basalts, but when compared to a global average of ocean island basalts (OIB), the Banat lavas are similar to average OIB, whereas the Persani Mts. basalts have higher Rb, Ba, K and Pb and lower Nb, Zr and Ti. These features slightly resemble those of subduction-related magmas, particularly those of a basaltic andesite related to the nearby older arc magmas (DOWNES et al. 1995b). With $^{87}Sr/^{86}Sr$ varying from 0.7035–0.7045 and $^{143}Nd/^{144}Nd$ from 0.51273–0.51289, the Romanian alkali basalts are indistinguishable (DOWNES et al. 1995b) from those of the western Pannonian Basin (Hungary and Austria — HARANGI et al. 1994, 1995, EMBEY-ISZTIN and KURAT 1997) and Neogene alkali basalts throughout Europe. It is inferred that, although the Romanian alkali basalts have a strong asthenospheric (i.e. OIB-type mantle source) component, their Pb isotopic characteristics were derived from mantle, which was affected by the earlier subduction (DOWNES et al. 1995b). It is in general agreement that Neogene alkaline basaltic rocks in the Pannonian Basin have some characteristics that represent some influence by former subduction in the region and associated metasomatic processes in the mantle.

Petrography of pyroclastic rocks

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

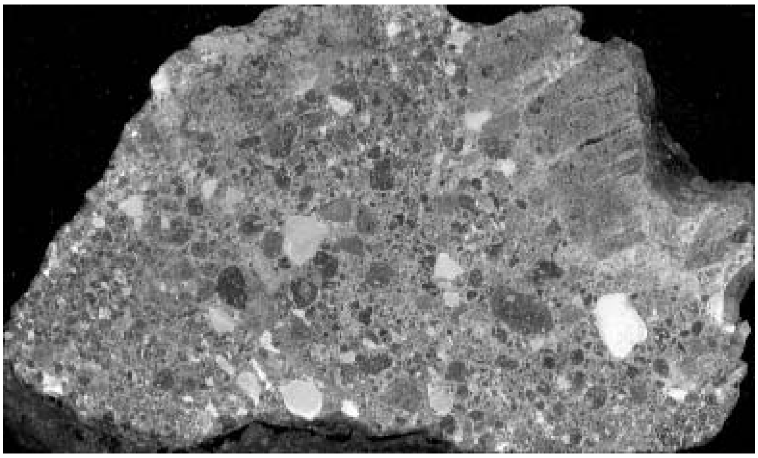


Figure 1.5. Typical phreatomagmatic accidental lithic rich (white angular clasts from the Triassic carbonates) lapilli tuff from Pula, BBHVF
The shorter side of the photo is 10 cm

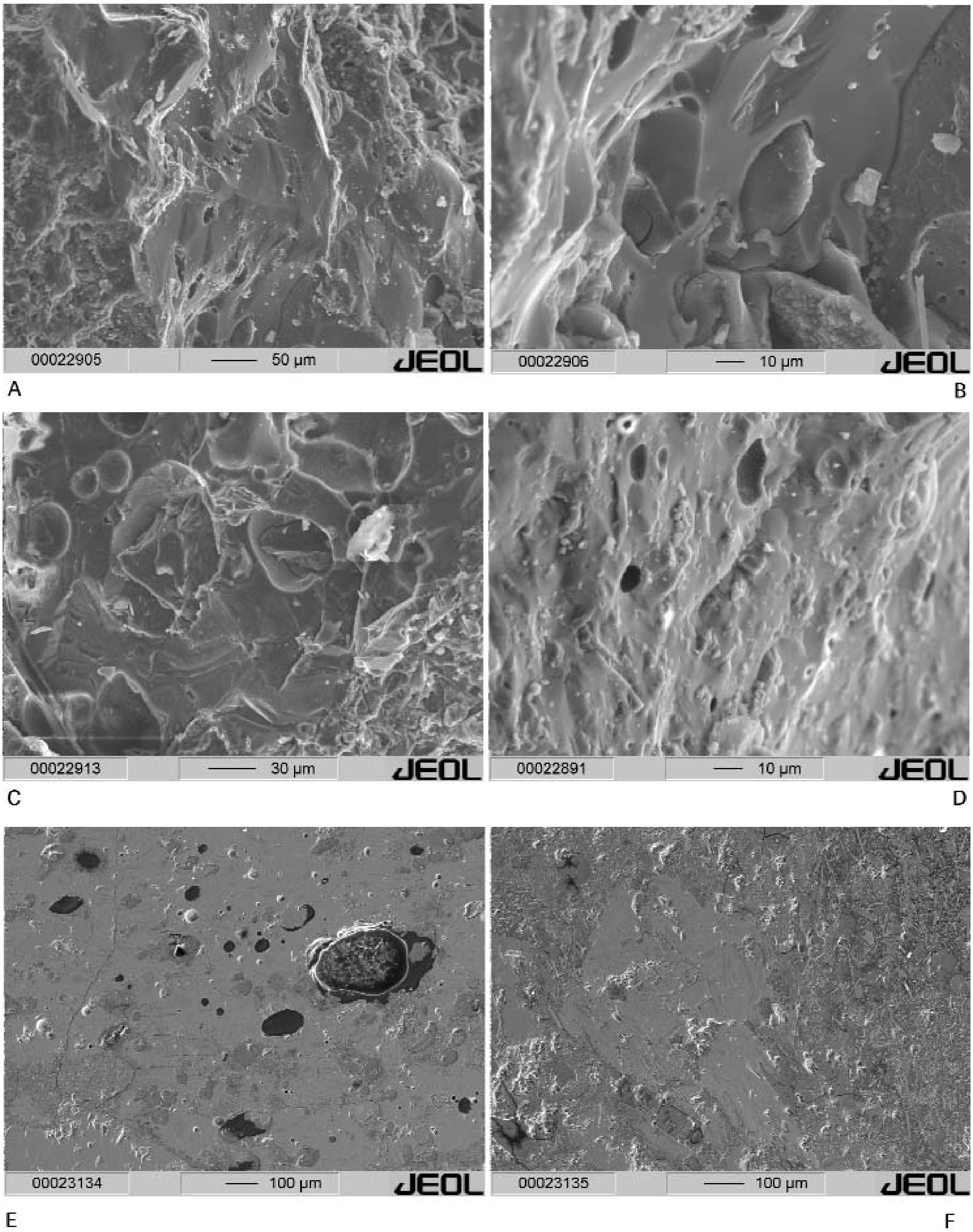


Figure 1.6. Scanning Electron Microscope photos of volcanic glass shards from the western Pannonian Basin made on JEOL 6740 Superprobe housed at the Geological Institute, TU-Bergakademie, Freiberg, Germany

A – Ság-hegy; phreatomagmatic lapilli tuff. Note the smooth surface and blocky shape of the glass shard. [on rock fragment]
B – Ság-hegy; phreatomagmatic lapilli tuff, a close-up of a vesicle. Note the angular limit of the vesicle, as well as the fracture, blocky shape of the shard. [on rock fragment]
C – Hajagos; phreatomagmatic lapilli tuff. Glass shards are moderately vesicular, with angular shape vesicles. [on rock fragment]
D – Pula; phreatomagmatic lapilli tuff with moderately vesicular phonotephritic lapilli. [on rock fragment]
E – Kis-Hegyes-tű; tephritic glass shard with low vesiculaity of a phreatomagmatic lapilli tuff. [on polished thin section]
F – Kis-Hegyes-tű; phreatomagmatic lapilli tuff with blocky, non-vesicular tephritic glass shards. Microlites shown on this image are slightly darker. [on polished thin section]

water interaction (Figure 1.6). There are abundant deep-seated xenoliths, 1 to 15 cm in size, in the uppermost beds of pyroclastic deposits at some of the volcanoes (Plate 1.4).

Major element variations in monogenetic volcanoes

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

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

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

Table 1.1. Composition of volcanic glass shards from pyroclastic rocks of erosion remnants of Neogene alkaline basaltic volcanoes of the western Pannonian Basin

Volcanic glass shard analyses from the BBHVF

sample	53	SZ3	SZV3	77	48	43	64	65	VD	52	K27	K28	K29	KH30	HJ27	HJ22
locality	20	8	8	24	5	5	14	14	10	19	15	15	18	22	14	14
deposit	P-bs	P-phf	P-bs	RW	P pf	P-phf	P-phf	P-phf	RW ?	RW ?	P-bs	P-bs	RW ?	P-bs	P-bs	P-bs
SiO_2	49.14	51.61	49.06	47.58	49.20	45.12	47.00	46.75	49.53	47.51	48.78	45.97	45.88	46.27	49.39	48.29
TiO_2	1.55	2.43	2.83	2.88	2.60	2.46	2.25	2.40	3.03	2.89	2.68	2.14	2.33	2.44	2.74	2.61
Al_2O_3	18.36	19.31	18.18	18.24	17.87	17.60	17.02	16.07	19.19	15.88	19.20	18.06	17.23	16.89	17.33	17.56
Fe_2O_3	2.04	1.81	2.03	2.51	2.01	2.22	2.13	1.95	2.12	2.43	2.34	2.10	1.71	1.98	2.19	2.44
FeO	6.81	8.00	6.77	8.38	6.60	7.39	7.12	6.48	7.02	8.09	7.81	7.00	5.71	6.62	7.29	8.14
MnO	0.17	0.17	0.13	0.16	0.14	0.16	0.12	0.12	0.13	0.20	0.16	0.21	0.10	0.18	0.06	0.20
MgO	2.95	2.64	3.26	3.41	3.62	3.53	3.54	3.57	3.79	3.43	3.07	3.12	4.13	2.74	3.82	3.73
CaO	7.43	6.47	9.66	8.88	9.16	8.90	8.83	9.11	8.80	8.70	8.21	7.88	10.25	9.88	9.35	9.06
Na_2O	6.01	1.97	3.60	2.40	2.14	4.64	4.87	4.53	2.17	4.96	2.23	5.27	5.04	4.92	2.37	2.46
K_2O	3.56	2.86	1.71	2.86	3.27	3.00	2.59	2.54	3.62	3.01	2.77	2.98	2.64	2.63	2.75	2.60
Total	98.02	95.27	97.43	97.28	96.70	95.02	95.47	94.12	99.60	97.10	97.25	94.73	95.02	96.53	97.29	97.09
norm ne	19.97	0.00	1.43	0.00	0.00	16.64	13.81	11.84	0.00	14.94	0.00	17.39	19.05	16.78	0.00	0.00
norm ol	4.34	0.00	4.52	7.11	0.94	4.59	4.20	3.20	5.45	3.17	0.00	4.86	1.61	1.83	0.15	4.56
	lph	bsn?	lbp	lbp	lbp	l	phl	phl	lbp	phl	lbp	phl	phl	phl	lbp	lbp

Volcanic glass shard analyses from two large Tihany type maar volcanic complex

sample	TIH85	TIH85	BIB	BIB	II/11	II/11	SzK31	SzK31	SzK31	SzK31	SzK31	SzK19	SzK12	SzK12	SzK7
locality	33	33	33	33	33	33	21	21	21	21	21	21	21	21	21
deposit	RW	RW	P-bs	P-bs	RW ?	RW ?	P-pf	P-pf	P-pf	P-pf	P-pf	P-bs	P-bs	P-bs	P-bs
SiO_2	47.04	52.67	44.17	44.06	46.87	46.68	47.20	47.44	66.39	48.61	46.73	44.18	47.24	50.68	49.50
TiO_2	3.00	2.39	2.85	2.67	2.63	2.68	2.33	2.36	1.20	2.26	2.30	2.23	1.73	1.82	2.58
Al_2O_3	17.88	15.73	20.43	17.44	17.25	17.29	17.82	17.94	15.60	17.93	17.82	18.23	17.50	20.38	18.03
Fe_2O_3	2.26	2.28	1.71	2.30	2.20	2.40	1.63	1.66	1.73	1.93	1.78	1.74	2.22	1.77	2.06
FeO	7.54	6.51	5.72	7.65	7.34	8.02	5.44	5.52	4.31	6.35	5.95	5.78	6.35	5.05	6.89
MnO	0.15	0.15	0.16	0.20	0.28	0.25	0.19	0.09	0.07	0.15	0.11	0.26	0.17	0.17	0.18
MgO	3.97	3.41	3.78	3.24	3.66	3.72	4.07	4.06	1.89	3.39	3.01	3.48	0.57	2.49	3.80
CaO	10.31	9.81	9.02	8.76	9.98	10.06	9.41	9.77	1.45	7.72	9.48	9.84	5.52	8.83	8.71
Na_2O	2.53	4.24	4.21	4.27	4.82	4.48	4.69	4.37	0.46	5.43	5.22	4.67	5.58	4.84	2.34
K_2O	2.90	2.69	2.39	3.29	2.84	2.83	3.08	2.90	1.88	3.48	3.10	3.11	3.72	2.74	3.08
Total	97.58	99.89	93.94	93.88	97.87	100.41	90.86	91.11	95.13	97.28	97.70	93.30	90.65	98.77	96.97
norm ne	3.65	4.46	14.51	16.23	17.06	13.20	21.50	19.14	0.00	16.86	17.75	19.43	18.96	9.41	0.00
norm ol	6.02	1.27	4.80	4.44	2.78	3.90	3.05	3.11	0.00	3.90	1.90	2.41	3.29	3.46	0.51
	tb-p	bt a	t	l	l	t	t	l	d	pht	phl	pht	phl	bt a	tb-p

lbp - potassic trachy basalt; bt a - basaltic trachy andesite; l - tephrite; d - dacite; phl - phono tephrite; tph - tephri-phonolite
P pf - pyroclastic (hydroclastic) flow; P-bs - base surge; P-phf - phreatomagmatic fall out; RW - reworked (Gilbert delta) or recycled (vent filling)

Sample abbreviations (first number refers to the sample number, the numbers in brackets refer to the location of the sample and correspond to the numbers shown on Figure 1.2: 53 (20) – Horog-hegy, Sz3 (8) – Szigliget, SzV3 (8) – Szigliget, 77 (24) – Pula, 48 (5) – Fekete-hegy south, 43 (5) – Fekete-hegy south, 64 (14) – Hajagos, 65 (14) – Hajagos, VD (10) – Boglár, 52 (19) – Öreg-hegy, K27 (15) – Kopasz-hegy , K28 (15) – Kopasz-hegy, K29 (18) – Kereki-domb, KH30 (22) – Kis-Hegyes-tű, HJ27 (14) – Hajagos, HJ22 (14) – Hajagos, TIH85 (33) – Tihany, BIB (33) – Tihany, II/11 (33) – Tihany, SzK31 (21) – Szentbékkálla, SzK19 (21) – Szentbékkálla, SzK12 (21) – Szentbékkálla, SzK7 (21) – Szentbékkálla have been recently described from the Transmexican Volcanic Belt (STRONG and WOLF 2003, SIEBE et al. 2004). However, scoria cones from the Transmexican Volcanic Field often form transition between monogenetic and composite volcanoes (MCKNIGHT and WILLIAMS 1997). In contrast, monogenetic volcanoes are formed by more or less direct eruption of magma from the mantle, with each volcano resulting from successful propagation of a small batch of magma to the surface along a new pathway (SPERA 1984, HASENAKA and CARMICHAEL 1985, HASENAKA and CARMICHAEL 1987, HASENAKA 1994, CONNOR and CONWAY 2000).

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

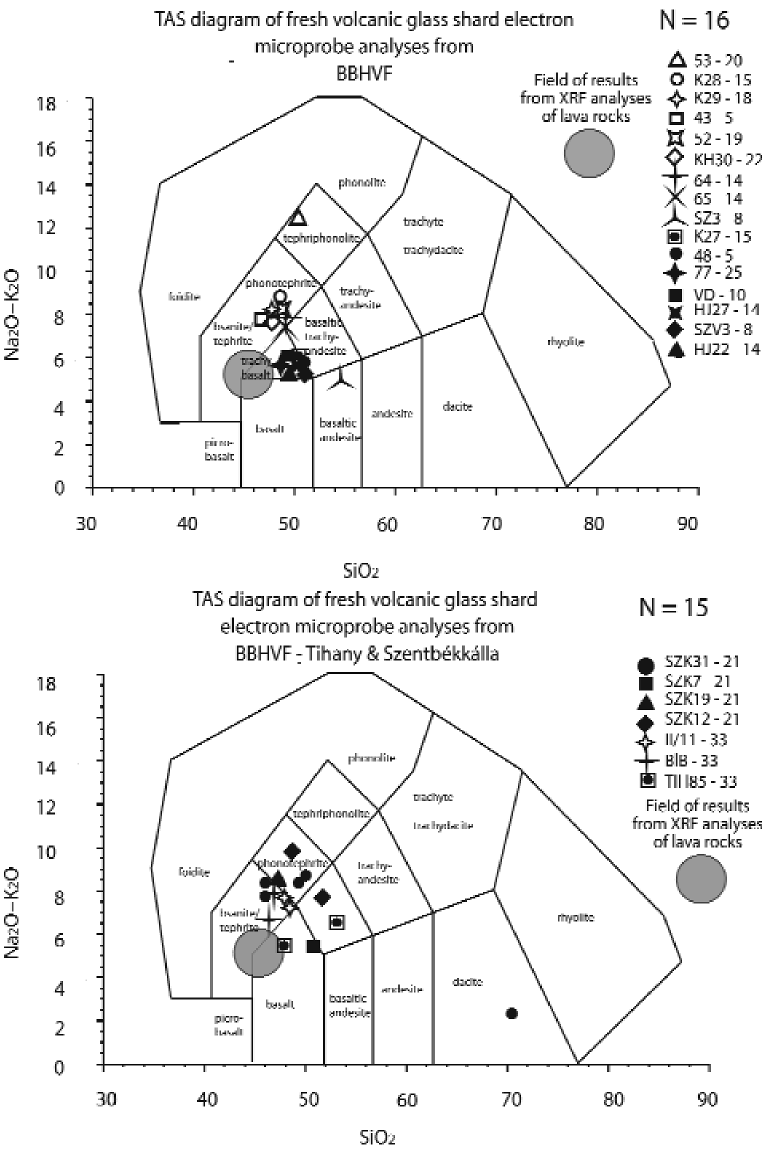


Figure 1.7. **TAS diagrams showing the composition of volcanic glass shards and subsequent lava flows from the BBHVF**

Sample abbreviations (first number refers to the sample number, the numbers in brackets refer to the location of the sample and correspond to the numbers shown on Plate 1.1): 53 – Horog-hegy (20), K28 – Kopasz-hegy (15), K29 – Kereki-domb (18), 43 – Fekete-hegy south (5), 52 – Öreg-hegy (19), KH30 – Kis-Hegyes-tű (22), 64 – Hajagos (14), 65 – Hajagos (14), Sz3 – Szigliget, Kamon-kő (8), K27 – Kopasz-hegy (15), 48 – Fekete-hegy south (5), 77 – Pula (25), VD – Boglár, Vár-domb (10), HJ27 – Hajagos (14), SzV3 – Szigliget, Vár-hegy (8), HJ22 – Hajagos (14), SzK31 – Szentbékkálla mafic pyroclastic flow (21), SzK19 and SzK12 – Szentbékkálla (21), II/11, BIB and TIH88 – Tihany Maar Volcanic Complex (33)

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

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

Age of the Neogene intraplate volcanoes in the western Pannonian Basin and their relationship to the immediate pre- and syn-volcanic sedimentation in the region

Intracontinental Mio/Pliocene volcanic fields of the western Pannonian Basin developed between 7.56 and 2.3 My (BALOGH et al. 1986, PÉCSKAY et al. 1995, BALOGH and NÉMETH 2004) across an area in size of about 50,000 km^2 (Figure 1.1). In the western Pannonian Basin, there are two closely related volcanic fields, that contain the largest number of volcanoes,

1. Bakony – Balaton Highland Volcanic Field (BBHVF) and
2. Little Hungarian Plain Volcanic Field (LHPVF – Plate 1.5).

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

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

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

Although there are analytical and sampling difficulties a great number of K/Ar age data are available from the Neogene basaltic volcanic rocks from the western Pannonian Basin. There is no apparent spatial distribution pattern among major age groups of volcanic rocks (Figure 1.2). The age

Figure 1.8. Aerial photograph of the Füzes-tó region, a 2.64 My ($^{39}Ar/^{40}Ar$ – WIJBRANS et al. 2004) old erosion remnant of a scoria cone

Note the tuff ring (line) in which the scoria cone formed and which is still well-preserved, (crater rim marked with dashed line). Note that the scoria cone must have been either open, or subsequently partly collapsed toward the north-east dates between 8 to 2.3 My seem to be randomly scattered in the area (Figure 1.2). There is a general centre point of ages at around 3.5–4 My BP, derived from volcanic remnants in the western part of the BBHVF (Figure 1.2). On the basis of new, laser induced step heating $^{39}Ar/^{40}Ar$ high precision ages, it seems, that one of the major part of the alkaline basaltic volcanism in the region, falls into the 3.8–4 My old period, which is in good concert with the previous K/Ar dates (Halom-hegy/3.78, 3.82, Hajagos/3.81, Hegyesd/3.9, Fekete-hegy lava field/3.81, Szigliget diatreme Várhegy pyroclastic sequence/4.08 – BALOGH et al. 1986, BORSY et al. 1986, WIJBRANS et al. 2004). An older age group of volcanoes can be identified on the basis of the $^{39}Ar/^{40}Ar$ ages at around 4.2 to 4.8 My (Szent György-hegy/4.22, Szigliget lava/4.53, Kis-Somlyó/4.63 and Tóti-hegy/4.74), however dates from Szigliget is likely to be in error, and they rather belong to the previous age group (WIJBRANS et al. 2004). These numbers also represent similar values than previous K/Ar dates from the same volcanoes (BALOGH et al. 1986, BORSY et al. 1986). The oldest known volcanic remnants are in Tihany, and their ages are fixed at around 8 My by repeated attempt to obtain isochron dates by the K/Ar method (BALOGH and NÉMETH 2004). The youngest volcanic edifices are erosion remnants of scoria cones topping the Haláp (3.08 $^{39}Ar/^{40}Ar$ – WIJBRANS et al. 2004), Agár-tető (2.9 K/Ar – BALOGH et al. 1986), Füzes-tó (2.64 $^{39}Ar/^{40}Ar$ – WIJBRANS et al. 2004) and Bondoró (2.3 K/Ar – BALOGH and PÉCSKAY 2001). Among these locations are the most well-preserved scoria cones in the western Pannonian Basin, which are still holding some primary morphology as well-defined crater rim (Figure 1.8).

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

Distribution of volcanic erosion remnants

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

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

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

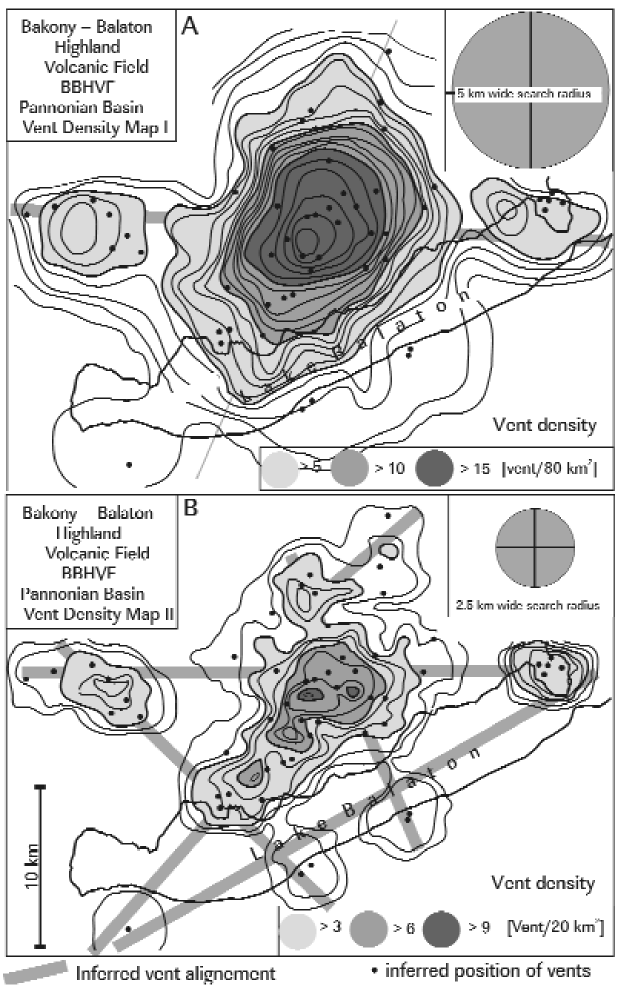


Figure 1.9. Vent density maps of the Bakony – Balaton Highland Volcanic Field (BBHVF – MARTIN et al. 2003)

The distribution of vents is analysed by the density contours drawn manually on the basis of rectangular grids with the spacing of 2 km and search radius of 5 km on figure A, and with a spacing of 1 km and search radius of 2.5 km on figure B maar system (Fekete-hegy — MARTIN et al. 2003). This location represents mafic volcaniclastic flow deposits (referred also as hydroclastic flow — NÉMETH and MARTIN 1999b), with large amounts of dm-size lherzolite xenoliths, mantle and/or deep crustal nodules (TÖRÖK and DE VIVO 1995, TÖRÖK et al. 2003), and pyroclastic deposits indicative of high-energy phreatomagmatic explosive eruptions (MARTIN et al. 2003). Further individual vent clusters are shown on a vent density map on the basis of a 1-km rectangular grid and 2.5 km search radius (Figure 1.9, B), mimicking major known crustal structural zones orientated mainly NE–SW and NW–SE (TARI 1991, BUDAI et al. 1999, BUDAI and CSILLAG 1999, DUDKO 1999). In general, using larger search radius and larger steps on grid, it is inferred that vent-distribution features are related to deep subsurface features such as the geometry of the melting anomaly (CONNOR 1990). In contrast, smaller search radius and steps on the grid give information on the surface structure of the pre-volcanic system (CONNOR 1990). It has been thus inferred that the Mio-Pliocene volcanism in the BBHVF was related to a characteristic melting anomaly from where the magma intruded into shallow subsurface crustal inhomogeneities, such as fault lines. Along with the vent clustering and alignments the north–south elongation of individual vents is likely to be related to valley pattern and inherited structural elements of the basement rocks (Figure 1.10). Analysis of the present morphology of the central part of BBHVF has revealed that north-south oriented textural pattern exist in this region, either representing palaeovalleys and/or surface expressions of old structural elements (JORDÁN et al. 2003). In the LHPVF, however, the vents are scattered. It is also noteworthy that major volcanic complexes such as Ság-hegy, Somló, Kab-hegy and Tihany (Plate 1.5) fall on a straight line that has no obvious surface expression in the form of faults or other structural elements (JUGOVICS 1969b, JÁMBOR et al. 1981).

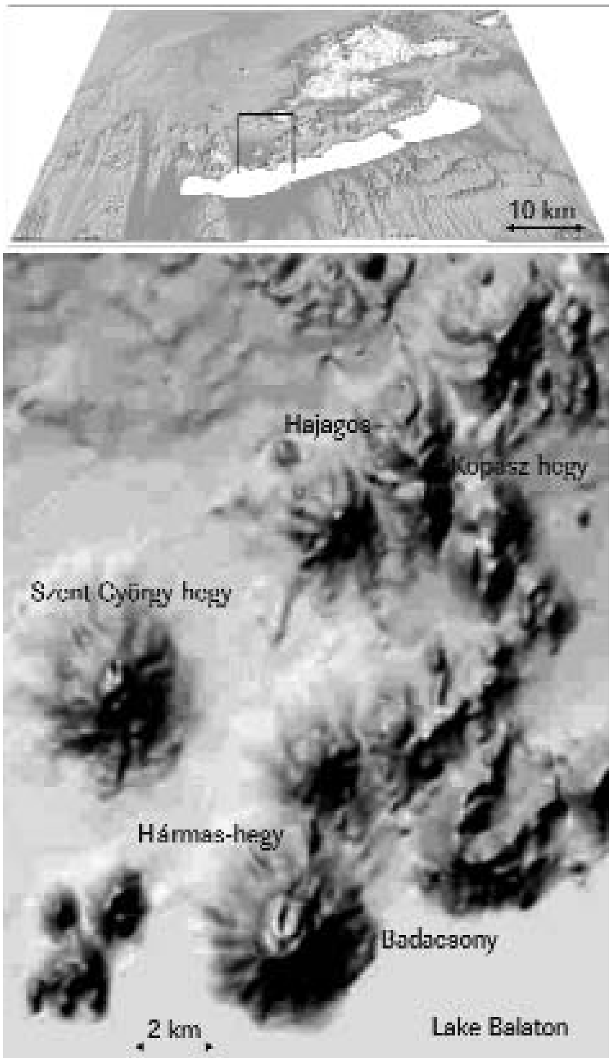


Figure 1.10. North-south elongated volcanic erosion remnants from the BBHVF on digital terrain model
The elongated volcanic remnants are named on the map

Lavaflows, scoria and spatter cones

There is a clear evidence for the presence of preserved fissure-vent systems in the western Pannonian Basin. In a small area around the northern part of the Keszthely Mts (e.g. Sümegprága — Plates 1.1 and 1.5) NE/EW elongated coherent lava rock outcrops occur. The extent of the lava rocks is strongly related to shallow subsurface intrusions such as sill and dyke systems and adjacent small lava flows, plugs. The linear alignment, of the lava rock outcrops suggests a fissure related origin, however, their surface exposures either have been eroded already, or never existed. The age distribution of the different eruptive centres also shows an alignment which probably is related to older structural zones in the basement, probably performed by the fluvial systems during the volcanism. Elongated structures of individual centres, or eruption complexes, especially in the middle part of the BBHVF (Hajagos-hegy, Sátorma-hegy, Fekete-hegy) also suggest longitudinal orientation of individual vents (Figure 1.10). The principal source of the lava flows appears to have been elongated, north to south, north-east to south-west trending former lava lakes. The best example for this is the Hajagos-hegy with a north to south trending lava lake, which overflowed southwards (Kő-hegy — Plate 1.6). The original lava lake was probably 700–1000 m long and 800 m wide. The buttes of Badacsony, Szent György-hegy, Csobánc, Haláp (Plate 1.1) show slightly elongated north–south shape. In the middle part of the BBHVF, there is a large volcanic complex, called Fekete-hegy volcano (Plates 1.1 and 1.5), which consists of smaller eruptive centres with different intercalated lava layers (MARTIN et al. 2002). The lava filled individual centres that also show a NE-SW-trend. The Fekete-hegy is interpreted as a complex lava channel, spatter cone and scoria cone system with several large intercalated lava lakes as well as lava flows fed by scoria cones in the surrounding phreatomagmatic tuff ring(s – MARTIN et al. 2002). The southernmost outcrops clearly show irregular shapes of the former lava lake and pyroclastic beds of a former tuff ring (MARTIN et al. 2002).

Shield volcanoes are common and give the major sources of lavas in intraplate provinces (WALKER 1993, 2000, CONNOR and CONWAY 2000). Eruptions of large Hawaiian-type volcanic centres are commonly related to fissure-vent systems, but in a small plain-basalt province eruptions are related to central vent systems. However, there are several examples where shield volcanoes developed along basement fissure systems (JOHNSON 1989). In the BBHVF there are two shield volcanic complexes probably associated with a large number of eruption vents (KORPÁS 1983, NÉMETH and MARTIN 1999d). The larger one (Kab-hegy – Plate 1.1) represents the highest topographic point in the Transdanubian Range. Individual lava flows tend to be around 5 to 8 km long and cover around 50 km^2 in area (JUGOVICS 1971, KORPÁS 1983, KORPÁS and SZALAY MÁRTON 1985). The total thickness of the lava cover reaches several tens of metres (JUGOVICS 1971, KORPÁS 1983, KORPÁS and SZALAY MÁRTON 1985). The lava field around Kab-hegy is a thick accumulation of various lavas that have been recognised already in the year 1934 by VITÁLIS. The lava flow units are often separated by thick palaeosoil layers as well as intensive alteration horizons of the basalt itself, indicating time gaps between effusion of lava flows and suggesting a complex eruptive history of this region (VITÁLIS 1934, VÖRÖS 1962, 1966, 1967). The top of the lava field is inferred to be an eroded scoria cone preserved as a plug. Adjacent to the lava fields of Kab-hegy small Strombolian scoria cone remnants and Hawaiian spatter deposits are common. The other large shield volcanic complex is the Agár-tető south-west of the Kab-hegy with a present elevation of 499 m. There is a small remnant of a scoria cone sitting over the lava plateau of the Agár-tető. The wide range of measured K\Ar age (5.25–2.8 My – BALOGH et al. 1986), the different lava flow units, the slightly different petrographic characteristics of the lava flow(s) and the lava inter-beds on the flank of the topping scoria cone suggest a long-lived volcanic activity, which might be related to stabile melt sources over structural weakness zone. Such time sequence is well known from several small to medium size shield volcanoes such as the Rangitoto Island (Auckland Volcanic Field, New Zealand – JOHNSON 1989, ENDBROOKE 2001) or large scoria cone examples from the Eifel (Germany) region (HOUGHTON and SCHMINCKE 1989). Moreover, Rangitototo (New Zealand) has a well-developed capping scoria cone as well as eroded satellite vents on the flank of the main edifice of the shield. The small spatter deposits on the top zone of the Agár-tető represent former summit craters of small (50–100 m) scoria and associated spatter cones. The preserved deposits represent small vent zones of this former explosion centres. Small remnants of lava cone structures are traceable everywhere where large lava lakes preserved. These areas usually are small (few tens of metres) irregularities in the large lava fields. They are inter-bedded with lava, and consist of spatter deposits (Badacsony, Szent György-hegy, Fekete-hegy, Sátorma-hegy – Figure 1.10).

Figure 1.11. Welded lava spatter-rich deposit from Ság-hegy

The magmatic explosive and effusive volcanic activity which produced large volumes of eruptive products as well as the presence of the elevated Mesozoic basement under this volcanic zone suggest that there was no magma-water interaction during the eruptive history, thus this area is inferred to have been already a higher elevated area in the Pliocene. These shield volcanoes in comparison with eastern-Australian examples, are relatively small with less than 1 km^3 volume of lava products (7 km^3 in eastern-Australia, JOHNSON 1989).

Large spatter and scoria cones (Figure 1.11) are strongly eroded in the BBHVF. They remained only as erosional remnants. Only the summit craters of the Agár-tető, Bondoró and Füzes-tó (Plate 1.5) are preserved in a recognisable morphology. Strombolian scoria cone remnants are however often preserved only as scoria mounds on top of larger volcanic erosion remnants such as the Boncsos-tető on top of the Fekete-hegy (MARTIN et al. 2002). Highly vesicular scoriaceous deposits often accumulated between lapilli tuffs rich in accidental lithic clasts indicating that eruption styles may have changed especially in the late stage of the eruptions of individual vents according to the available water to sustain magma–water interaction (HOUGHTON and HACKETT 1984, HOUGHTON and SCHMINCKE 1986, HOUGHTON et al. 1999). Scoriaceous deposits are often rich in irregular shaped mud chunks which are preserved between scoriaceous lava spatters often in significant thickness (tens of metres: e.g. Ság-hegy – Plate 1.1), indicating an active quarrying of an unstable volcanic conduit and/or presence of water-rich slurry in the vent zones during more magmatic fragmentation of magma (KOKELAAR 1986, WHITE 1991b, ORT et al. 1998, MCCLINTOCK and WHITE 2000). Scoria cone remnants are usually preserved in western Pannonian Basin as near vent strongly baked, red, slightly bedded sequences with large spindle or highly vesicular fluidal bombs. Welding of scoriaceous lapilli is common (Plate 1.6). Strombolian scoria and spatter deposits are common in relation with maar volcanoes. Such beds often reflect irregularities in the magma/water interaction having clear magmatic and phreatomagmatic fragmentation styles such as it was observed during the eruption of the Ukinrek Maar, Alaska (KIENLE et al. 1980, SELF et al. 1980, ORT et al. 2000). Scoria cones often develop in maar basins such as the La Breña Maar in Mexico (Plate 1.7, A), which is a recent analogy how e.g. the Pula maar (Plate 1.1) may have looked like prior to the water filled its basin. Scoria cones are inferred to have grown in maar basins in the BBHVF such as inferred for Uzsa maar (Plates 1.1 and 1.5). Such scoria cones (e.g. Uzsa) may have collapsed into the maar basin, feeding extensive volcanic debris avalanches and associated debris flows, which tend to accumulate in the maar basin. A remnant of a Strombolian scoria cone in the Füzes-tó region preserves near vent scoriaceous volcaniclastic breccia with muddy matrix that is inferred to represent remnants of the water saturated slurry in the vent during the Strombolian activity. The uprise of magma at Füzes-tó is inferred to have involved turbulent jets that did not generate shockwaves as suggested in theoretical considerations (MASTIN 2004).

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

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

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

Hyaloclastite and peperite

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

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

Phreatic and phreatomagmatic eruptive centres

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

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

Phreatomagmatic activity

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

The majority of the eruptive centres of the volcanic fields in the western Pannonian Basin have a phreatomagmatic history. Subaerial phreatomagmatic explosions usually produce low rimed tuff rings (WATERS and FISHER 1970, HEIKEN 1971, KELLER 1973, WOHLETZ and SHERIDAN 1983, SOHN and CHOUGH 1989, GODCHAUX et al. 1992, ALLEN et al. 1996, MASTROLORENZO 1994, SOHN 1996, VESPERMANN and SCHMINCKE 2000). The water source of the phreatomagmatic centres in the Western Pannonian region is predominantly a combination of ground water and some water from shallow lakes and/or fluvial systems (NÉMETH and MARTIN 1999c, NÉMETH et al. 1999b, 2001, MARTIN et al. 2003). The identification of magma-water interaction is based on the common presence of

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

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

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

Pyroclastic rocks from the LHPVF and the western part of the BBHVF are like the first type. In contrast pyroclastic rocks from vent remnants in the eastern BBHVF are more like the second type. This relationship is inferred to be related with the explosion locus, and the sub-surface architecture of the different regions, which is likely to determine the palaeohydrogeology of the regions (MARTIN and NÉMETH 2003). The depth of the phreatomagmatic fragmentation at the volcanoes of the western Pannonian Basin varies greatly according to the accidental lithic clast population of the phreatomagmatic pyroclastic rocks (NÉMETH and MARTIN 1999c). In the central part of the BBHVF a few locations preserve pyroclastic rocks, which are rich in accidental lithic fragments from every known lithology from the upper crust, and therefore the fragmentation level could have been in a range of kilometers from the syn-volcanic palaeosurface. Because aquifer pore pressure increases with depth, many researchers have assumed that once pressure exceeds water's critical pressure, magma/water interaction ceases to be explosive because steam is not formed (e.g. CAS and WRIGHT 1988). This assumption is suggested to be incorrect according to the latest results of theoretical experimental work, which show the potential for dynamic interaction at confining pressures up to and perhaps exceeding 100 MPa (WOHLETZ 2004). These results suggest that interaction can initiate at maximum depth >4 km, reaching depth perhaps exceeding 10 km as suggested by WOHLETZ (2004). Indeed, there is evidence of fluid (H_2O) inclusion study on volcanic glass from pyroclastic rocks associated with phreatomagmatic successions of the BBHVF (e.g. Szigliget), that their entrappment may have occurred at arround a few km (~10 km) depth. This fluids could have been responsible for the magma/water interaction (e.g. BALI et al. 2002, TÖRÖK et al. 2003).

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

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

Maars and tuff rings

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

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

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

Types of phreatomagmatic volcanoes

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

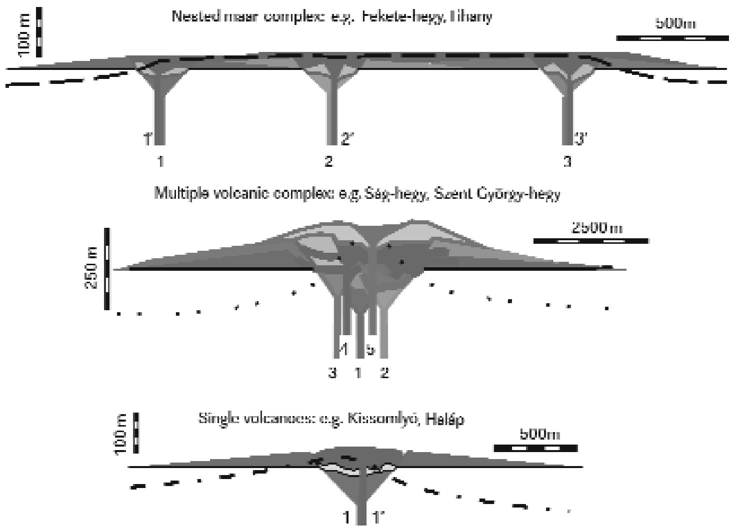


Figure 1.12. Volcanic types have been identified from the western Pannonian Basin, according to the preserved type of pyroclastic rocks, the relative ratios of juvenile versus accidental lithic clasts, the 3D architecture of the preserved volcanoes in relevance to the reconstructed craters, and the amount of effusive products associated with such volcanoes. Numbers represent time sequence of eruptions. Dashed lines show present surface

Complex multiple volcanoes (Figure 1.12 – MARTIN et al. 2003) with solidified large volume lava lakes are characteristic volcanic remnants especially in the western part of the BBHVF, such as Badacsony (Plate 1.5), one of the largest lava-capped buttes in the BBHVF (Figure 1.13). Thick (>50 m), black, strongly chilled, aphanitic basanitic lava overlies a coarse-grained, unsorted yellow lapilli tuff (MARTIN and NÉMETH 2002b). The lapilli tuff consists of finely dispersed accidental lithic fragments of quartz or quartzofeldspathic sandstone, and blocky, weakly to highly vesicular microlite-poor sideromelane (tephrite, phonotephrite), indicative of phreatomagmatic origin, near-surface vesiculation and possible excavation of pre-volcanic country rocks (MARTIN and NÉMETH 2002b). In addition, xenocrysts of olivine and pyroxene may reach 5 vol.%. The lava lake at Badacsony exhibits irregular lower contacts with the pyroclastic units, often displaying peperite structures (MARTIN and NÉMETH 2002b). The peperite encloses highly vesicular scoriaceous lava spatter clasts, with vesicles filled by clay, calcite or quartzofeldspatic assemblage suggesting rejuvenation or longevity of volcanic vents at the same site (MARTIN and NÉMETH 2002b).

Figure 1.13. Panoramic view to Badacsony from north

Figure 1.14. Small diatreme remnant of the Kereki-domb in the central part of the Kál Basin

Single phreatomagmatic volcanoes (Figure 1.12 – MARTIN et al. 2003) are also widespread, such as Kis-Somlyó, (MARTIN and NÉMETH 2002a, 2004b) Kereki-domb (Figure 1.14) or Vár-hegy of Zánka (NÉMETH et al. 2003a). They consist of moderately to strongly eroded pyroclastic mounds, forming small hills and exposing phreatomagmatic lapilli tuff beds, often in chaotic setting, which are commonly covered by lava flows. The presence of chilled, angular, moderately vesicular sideromelane fragments and the predominantly Neogene sediment-derived accidental lithic clast population in the exposed pyroclastic rocks allow to infer that these volcanic remnants represent near vent to vent-fill pyroclastic units of former phreatomagmatic volcanoes, such as maars surrounded by tuff rings, which were topped by lava flows in the last stage of volcanic activity.

Maar crater-fill sediments, Gilbert-type deltas

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

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

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

Maar lake carbonates

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

Today, the carbonate sequences, especially in Tihany case, are in the highest elevated areas. They must represent the former lake bottom thus they are useful for calculating local erosion rates.

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

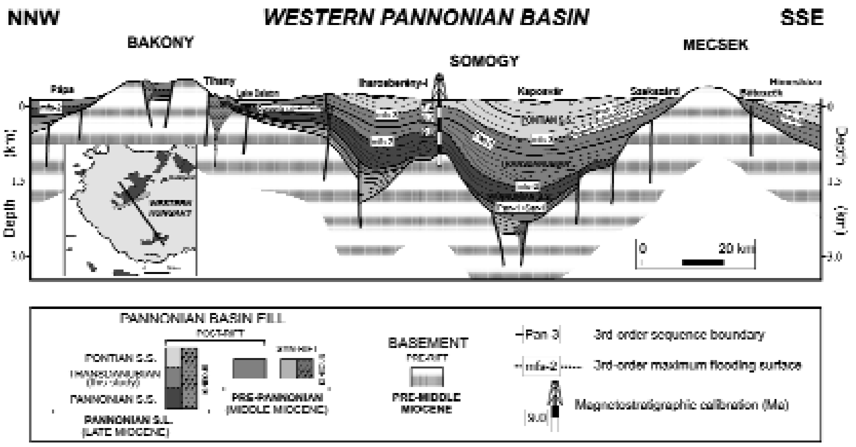


Figure 1.15. Correlation of units identified by seismic sections through the Lake Balaton (SACCHI et al. 1999, SACCHI and HORVÁTH 2002)

The stratigraphic unit between maximum flooding surfaces mfs-2 and mfs-3 (ca. 9.0–7.4 My) represent the newly introduced stage called Transdanubian (SACCHI and HORVÁTH 2002). The new K/Ar dating of the Tihany Maar Volcanic Complex give ~ 8 My for the eruption of the maars at Tihany, which correlates well with the regional stratigraphic situation of the Neogene sediments nearby (BALOGH and NÉMETH 2004). Figure used with permission of the authors of original figure

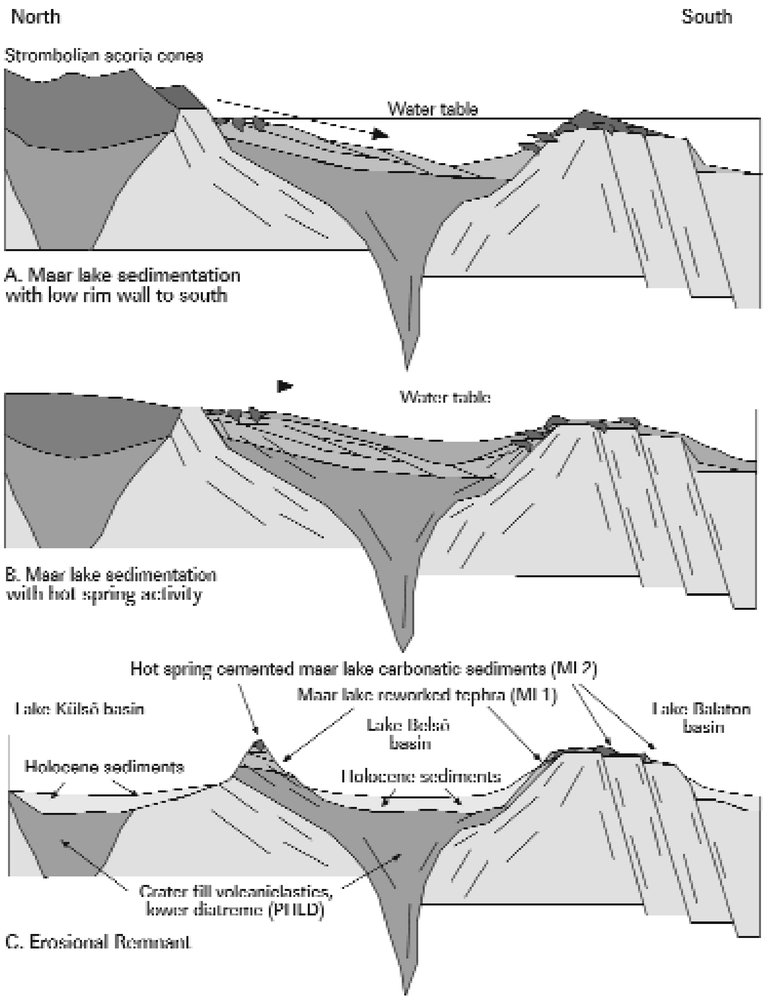


Figure 1.16. Model of the evolution of the Tihany Maar Volcanic Complex, with open maar basins, which may have been connected with normal lacustrine basins nearby (NÉMETH et al. 1999a)

General features of the western Pannonian Basin

The Neogene alkaline basaltic volcanic erosional remnants of the western Pannonian Basin are exposed from former subsurface to surface levels but they are commonly covered by the Quaternary erosional talus flanks. The outcrop availability strongly controls the identification of the facies relationships. The deepest levels of exposures are located in the western and the southern part of the area. The most strongly eroded regions are those where no subsequent lava caps sheltered the volcaniclastic sequences. The Balatonboglár, Szigliget, Tihany Peninsula allow the study of the deepest level of the volcanic centres, exposing diatreme facies (Plates 2.1 and 2.5). Probably the eruptive centres of Balatonboglár (Boglár Volcano), Kereki-domb, Vár-hegy of Zánka, Hármas-hegy and Véndek-hegy (Plates 2.1 and 2.5) represent the deepest exposed level of the phreatomagmatic eruptive centres (NÉMETH and MARTIN 1999c, NÉMETH et al. 2003a) and are inferred to be exposed zones of lower diatremes (terminology after WHITE 1991b). Apart from this deep level of exposures, there are no exposed irregular shapes, fragmented wall rock rich dykes, like those are widely reported from other monogenetic maar-diatreme volcanic fields as e.g. Hopi Buttes or southern Africa (WHITE 1991a, b, KURSZLAUKIS and LORENZ 1997, LORENZ and KURSZLAUKIS 1997, LORENZ 2000). Such levels of exposures and preserved outcrops are more common in the northern part of the Pannonian Basin, in southern Slovakia and northern Hungary (KONEČNÝ et al. 1999a, KONEČNÝ and LEXA 2000). In the BBHVF, the Tóti-hegy and Hegyesd are the good examples where such relationship between feeder dyke and host diatreme filling pyroclastic rocks may exist, however, their contact is not exposed (Plates 2.1 and 2.5). The best example for individual plugs as an exposed sub-volcanic feeder dyke is Hegyes-tű where a small remnant of vent filling mixture of volcanic and siliciclastic debris is preserved and intruded by the plug, indicating, that explosive fragmentation preceded the formation of the basanite plug (Plates 2.1 and 2.5).

Upper diatremes (WHITE 1991b) represent scoria cones and associated lava plugs built on the basal phreatomagmatic volcano. However, they do not necessarily represent volcanic landforms grown over the syn-volcanic landscape. After erosion, a significant part of the volcanic edifice could be eliminated, and in the eroded phreatomagmatic volcanic field the identification of such remnants should be dealt with precaution to establish the syn-volcanic palaeo-surface, to estimate the erosion. Most of the BBHVF buttes represent upper diatremes, similarly to the Hopi Buttes (WHITE 1991b) or Western Snake river subaerial volcanic centres (GODCHAUX et al. 1992). Surface volcanic edifices are preserved in areas of low erosion, and include volcanoes produced by both phreatomagmatic and magmatic eruptions. The vents characterized by magmatic explosivity are concentrated in the northern part of the area of the BBHVF (Kab-hegy, Agár-tető, Haláp, Hegyesd). The preserved maar diameters of the western Pannonian region ranges from few hundreds of metres up to 5 km in diameter (Fekete-hegy – 5 km; Tihany – 4 km; Bondoró – 2.5 km; Badacsony – 2.5 km), however, the largest centres probably represent maar volcanic complexes with connected large basins (MARTIN et al. 2002). The average maar basins are inferred to have been 1–1.5 km wide originally, which is within the range of most maars worldwide (LORENZ et al. 1970, HEIKEN 1971, LORENZ 1973, SCHMINCKE et al. 1983, FISHER and SCHMINCKE 1984, LORENZ 1986, CAS and WRIGHT 1988). The maar vents from the western Pannonian region are reconstructed as hybrids of phreatomagmatic and magmatic volcanic edifices, formed by initial maar or tuff ring forming events. The gradual exhaustion of water source to fuel the magma/water interaction led to "drier" phreatomagmatic, then pure magmatic fragmentation of the uprising melt, often building large scoria cones inside of the phreatomagmatic volcanoes similarly to several examples from Eifel (HOUGHTON and SCHMINCKE 1986, 1989) or from Mexico (ARANDA-GOMEZ et al. 1992). The typical types of such volcanoes are located in the southwestern site of the BBHVF in the Tapolca Basin (Badacsony, Szent György-hegy, Hajagos-hegy, Fekete-hegy – Plate 1.1).

Proximal base surge beds commonly contain abundant accidental lithics. In the eastern side (Tihany maar volcanic complex) the major accidental fragments are permian red sandstone, silurian schist (e.g. Lovas Schist Formation) mesozoic carbonates (dolomites, marls, limestones), and pannonian sandstone. Commonly the large (up to 75 cm in diameter) fragments are silurian schist or permian red sandstone. The matrix of the surge beds contains a high proportion of sand from the Pannonian sandstone beds. In the western side of the BBHVF and in the LHPVF the main accidental lithic fragments are from the Pannonian siliciclastic units. Both the large fragments (up to 25 cm in diameter) and the matrix are rich in sandstone fragments. In smaller proportion (less than 20 vol.% of total accidental lithics), there are schist fragments and small carbonate fragments. In the middle part of the area (Fekete-hegy, Bondoró, Pipa-hegy – Plate 1.1) the Mesozoic carbonates are the major part of the accidental lithics (min. 85 vol.% of total accidental lithics). In distal facies the base surge beds become finer-grained. Clear distal facies of surge beds are visible from the Tihany Peninsula. Characteristics surge features such as sandwaves, dunes, impact sags, U-shaped valleys are common (Tihany Peninsula, Fekete-hegy, Bondoró, Szigliget – Figure 1.2). Fall deposits associated with surge beds are also common (Fekete-hegy).

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

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

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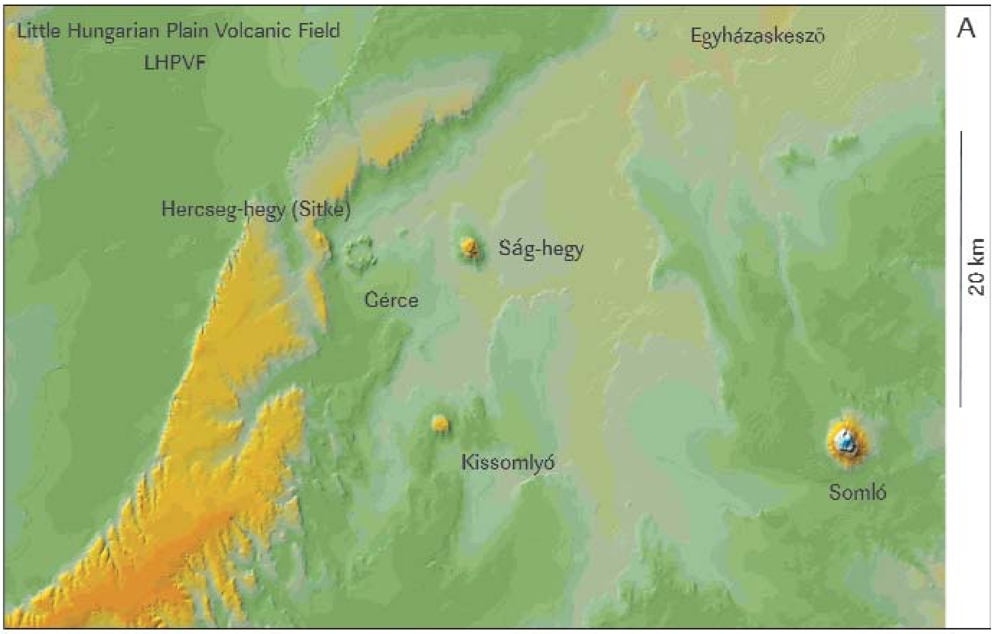
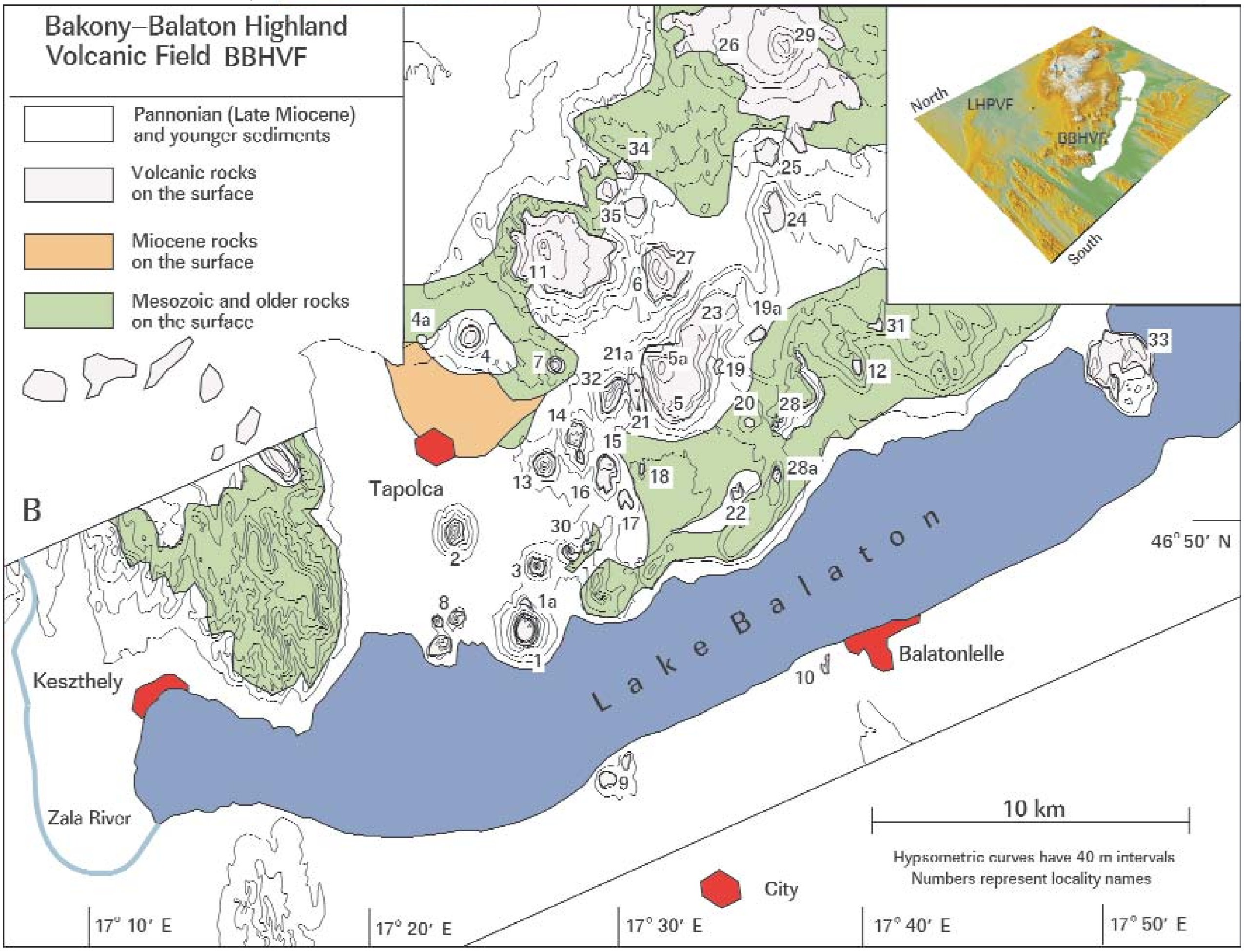
**Shaded relief model of the Little Hungarian Plain Volcanic Field (excluding the Styrian Basin Voclanic Field) (A).
Simplified geological map of the Bakony – Balaton Highland Volcanic Field (B) with the localities of volcanic erosion remnants**

1. Badacsony, 1a. Hármas-hegy, 2. Szent György-hegy, 3. Gulács, 4. Haláp, 4a. Véndek-hegy, 5. Fekete-hegy south, 5a. Boncsos-tető, 6. Bondoró west, 7. Hegyesd, 8. Szigliget, 9. Fonyód, 10. Boglár, 11. Agár-tető, 12. Tagyon, 13. Csobánc, 14. Hajagos, 15. Kopasz-hegy, 16. Pipa-hegy, 17. Kékkút (Harasztos-hegy), 18. Kereki-domb, 19. Öreg-hegy, 19a. Balatonhenye, 20. Horog-hegy, 21. Szentbékkálla, 21a. Füzes-tó (Kopácsi-hegy), 22. Kis-Hegyes-tű/Lapos-Hegyes-tű, 23. Kapolcs, 24. Tálodi-erdő, 25. Pula, 26. Kab-hegy west, 27. Bondoró east, 28. Hegyes-tű, 29. Kab-hegy peak, 30. Tóti-hegy group, 31. Halom-hegy, 32. Sátorma, 33. Tihany, 34. Taliándörögd hills, 35. Öcs hill ↓





Typical volcanic landforms associated with phreatomagmatic volcanic fields

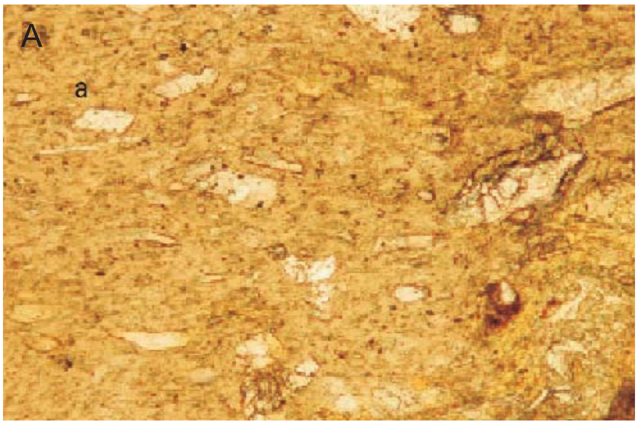
Crater Elegante (Mexico), maar

Diatremes of Hopi Butte, Arizona

Cerro Colorado (Mexico), tuff ring

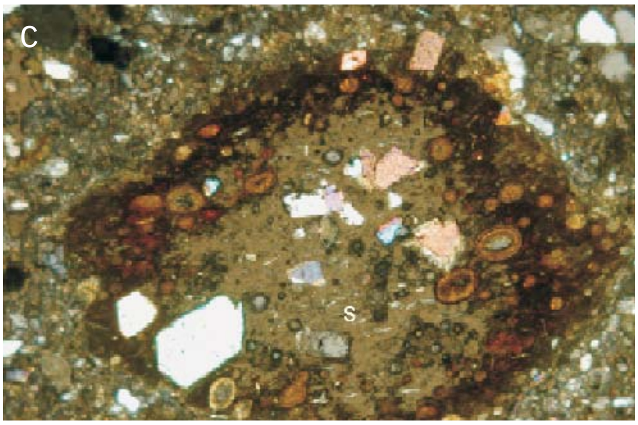
Cinder (scoria) cones
of the Durango Volcanic Field (Mexico)

Volcanic glass shards on photomicrographs from the western Pannonian Basin

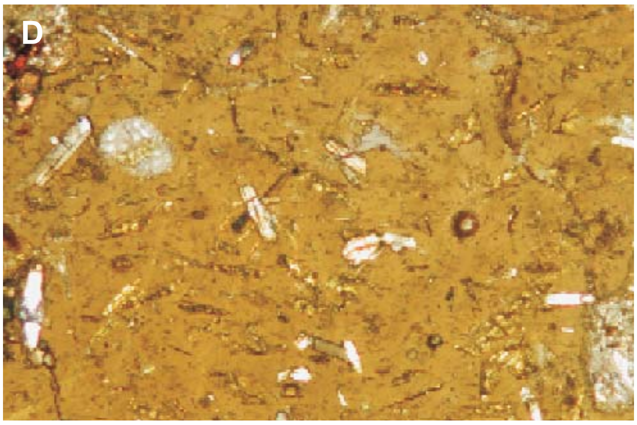


Boglár diatreme; oriented microlite-rich glass shard. The shards often contain amphybole crystals (a). The shorter side of the photo is 1 mm [parallel plane light]

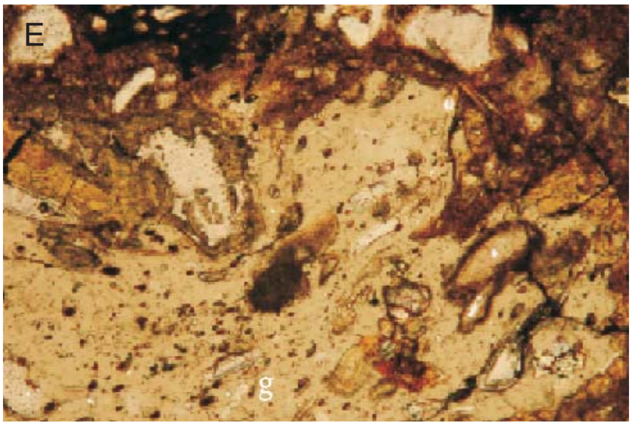
Pula; moderately vesicular, blocky volcanic glass shard (g) from a glass shard rich layer in the maar lake lacustrine succession. The short side of the photo is 1 mm [parallel plane light]



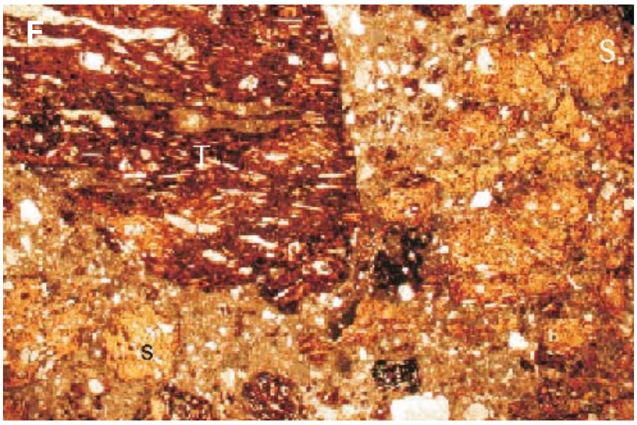
Egyházaskesző; blocky, moderately vesicular sideromelane glass shard (s) with palagonite rim (dark rim) from a lapilli tuff. Note the non-oriented, microphenocryst bearing texture of the glass shard. The short side of the photo is 1 mm [parallel plane light]



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



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

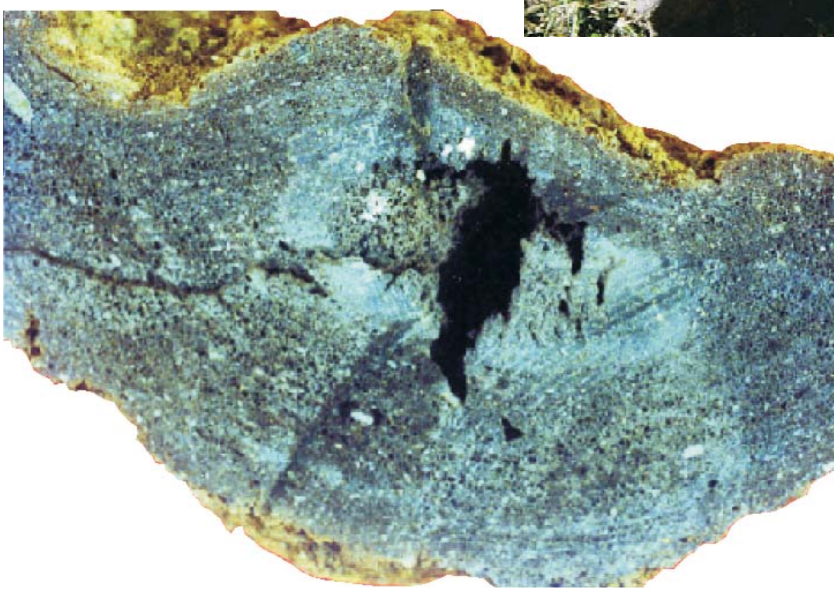


Kékkút diatreme; glass shard with trachytic texture (T). Note the smaller, lighter coloured sideromelane glass shards (s). The short side of the photo is 1 mm [parallel plane light]

Lherzolite xenolith (s) from a spindle bomb ~30 cm across (A) from the Füzes-tó. A great variety of mantle nodules can be found in dense (B) as well as highly vesicular spindle bombs (C – bomb is 40 cm across) from this region

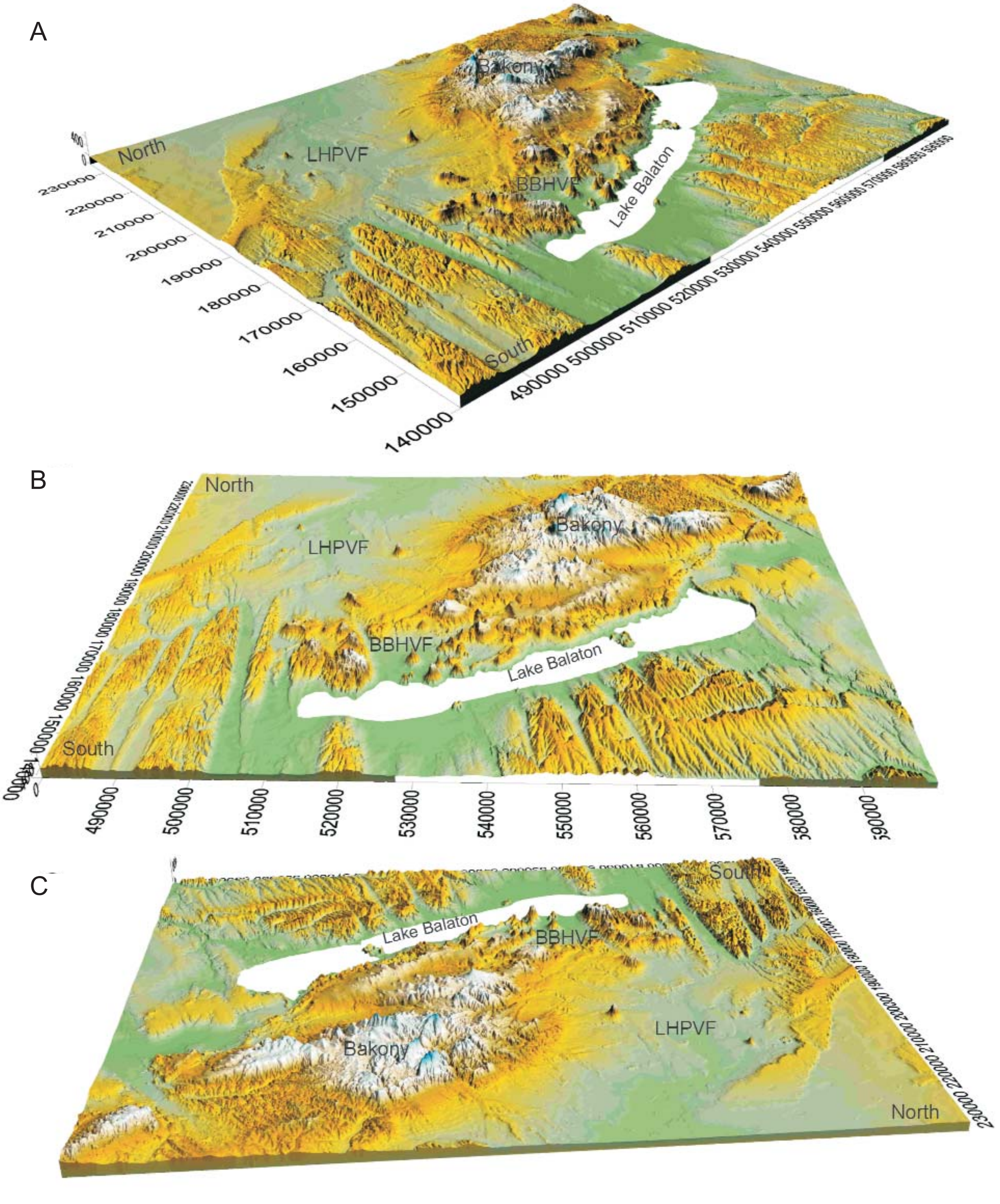
A

B

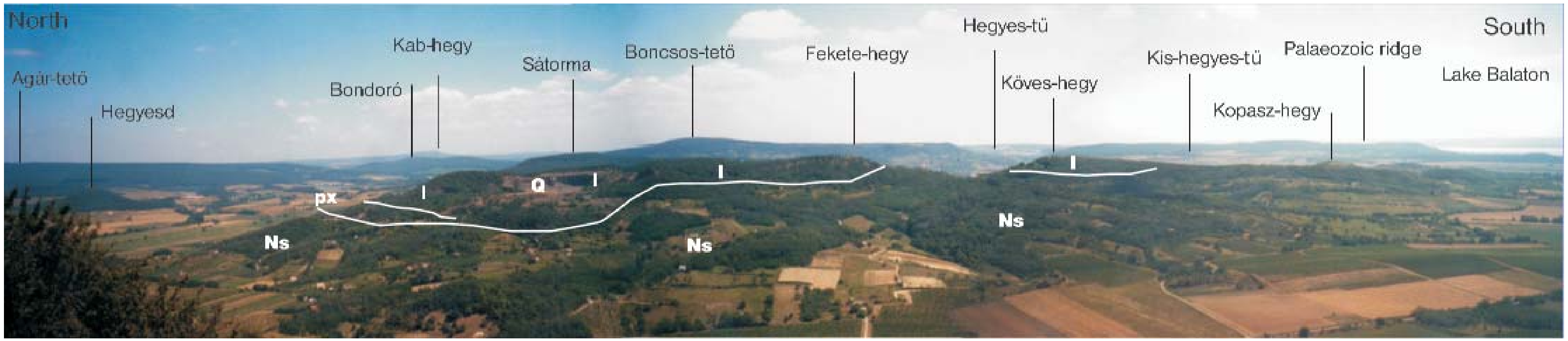


C

Oblique 3D views of the Bakony – Balaton Highland and the Little Hungarian Plain Volcanic Fields



North–south elongated hill of Hajagos, looking from the Csobánc (from west). The lines mark the limit of volcanic rocks mapped.
Lava flow – l, Pyroclasic rock units – px, Neogene siliciclastic units – Ns, Quarry – Q



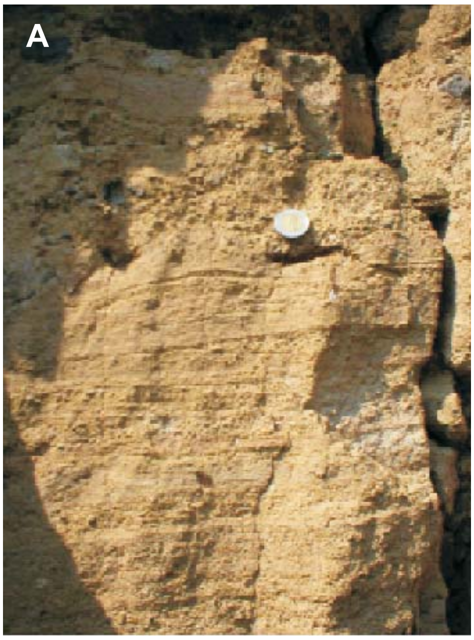
Scoria cone and adjacent lava flow field in the La Breňa Maar, Mexico. This maar is in good analogy with maar remnants of the BBHVF such as the Pula Maar prior to flooding by water

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

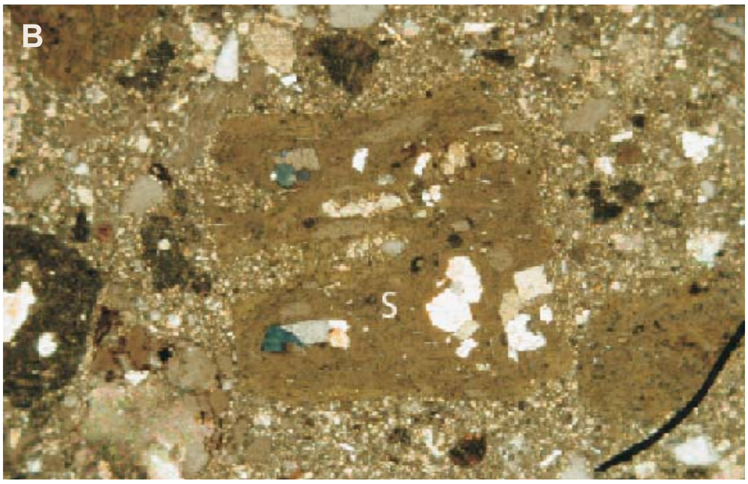
Rosette-like joints in the Sümegprága shallow sill and dyke complex

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

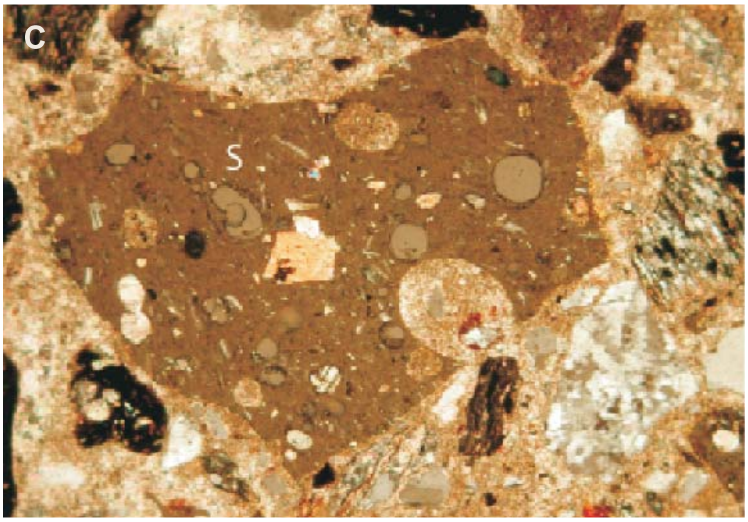
Peperitic sill that intruded into the tuff ring sequence of the Ság-hegy



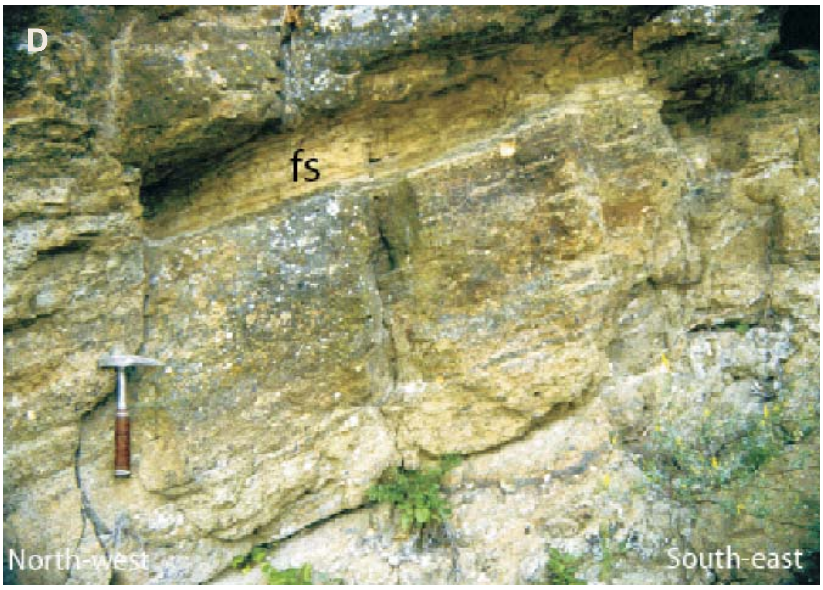
Accidental lithic rich tuff from Ság-hegy. The dominant proportion of the clasts are sand or quartz derived from Neogene clastic succession, all suggested earlier that the deposit is rather a siliciclastic deposit, and thereforefore volcanism and siliciclastic deposition from the Pannonian Lake is coeval.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



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



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



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

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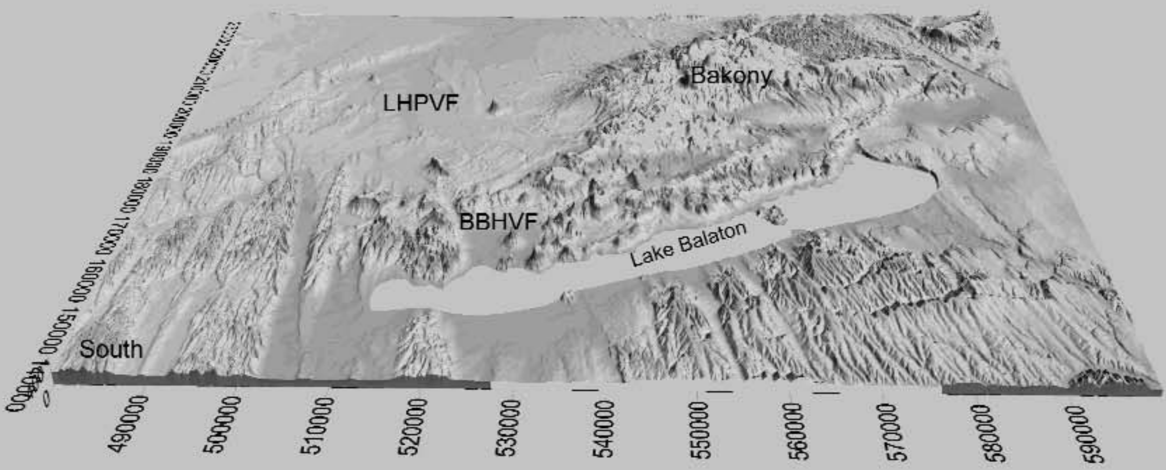
Late Miocene to Pliocene palaeogeomorphology of the western Pannonian Basin based on studies of volcanic erosion remnants of small-volume intraplate volcanoes



Contents

Abstract

Neogene volcanic fields in the western Pannonian Basin are eroded clusters of predominantly phreatomagmatic volcanoes that erupted from the Late Miocene (~8 My) and repeatedly became active until the Pliocene (~2.3 My). The volumetrically largest accumulation of volcanic erosion remnants is located in the Bakony – Balaton Highland Volcanic Field (BBHVF). In contrast more randomly distributed vent remnants occur in the Little Hungarian Plain Volcanic Field (LHPVF), and are aligned along major faults. The alkaline basaltic eruptive centres are eroded scoria cones, tuff rings, maar volcanic complexes and shield volcanoes. The amount of erosion has been calculated using volcanic lithofacies-distribution of these volcanic erosional remnants. Erosion remnants of the BBHVF are considered as being diatremes that undercut the syn-volcanic palaeosurfaces and therefore their present elevation represents various exposure levels below the syn-volcanic palaeosurface. At the LHPVF, volcanism is inferred to have developed on a flat lying land dominated by alluvial sedimentation, and volcanic landforms that very likely have been buried under younger (Late Pliocene to Quaternary) sediments. Volcanic remnants of the LHPVF are inferred to have been exhumed recently. Preserved tuff ring structures are still recognisable on old landforms. Erosion rates from the BBHVF are calculated based on estimation of the depth of exposure level of the preserved diatreme facies. Considering a phreatomagmatic origin of the studied volcanoes and a characteristic north–south lineament of the remnants of the volcanic edifices, it is inferred that they erupted in hydrogeologically active zones (valleys), which are likely to be controlled by faults. The calculated amount of 100–300 m erosion since volcanism ceased is interpreted as minimum value for the erosion of the predominantly Neogene siliciclastic sedimentary units. Erosion between the end of the Neogene shallow marine to fluvio-lacustrine sedimentation and the start of volcanism (~8 My) is calculated in two ways; 1. using the same erosion rates before volcanism, which was calculated from the time since the volcanism terminated, or 2. using a uniform average erosion rate for the pre-volcanic time based on field evidences (e.g. 10 to 100 m/My). These calculations result in a total thickness of Neogene sedimentary cover in the BBHVF region of 250 m up to 900 m before erosion. The erosion calculations based on volcanological evidences and the estimation of the total thickness of the immediate pre-volcanic Neogene sedimentary cover at ~8 My support the conclusion that Neogene sedimentary cover buried most of the western Pannonian Basin including the Transdanubian Range. The erosion rate of the BBHVF is estimated to vary between 100 m/My and 20 m/My.

Keywords: maar, diatreme, monogenetic, erosion, alluvial, fluvial, basalt, exhumation, erosion rate

Introduction

The alkaline basaltic, intracontinental, monogenetic volcanic fields in the western Pannonian Basin such as the Bakony – Balaton Highland Volcanic Field (BBHVF) and the Little Hungarian Plain Volcanic Fields (LHPVF) had been active during the Mio/Pliocene (~8 to 2.3 My – BALOGH et al. 1982, 1986, BORSY et al. 1986, BALOGH and PÉCSKAY 2001, BALOGH and NÉMETH 2004, WIJBRANS et al. 2004 – Plate 2.1). They largely comprise variably eroded tuff rings, maars, scoria cones and lava flows. The basement of the volcanic fields consists of two major groups of rocks,

1. a deeper seated hard rock succession including major karst aquifers, and
2. topmost unconsolidated "soft" sedimentary succession of Neogene shallow marine to fluvio-lacustrine silt, sand, and gravel units deposited either from the Pannonian Lake or fluvio-lacustrine systems that occupied the region after this lake disappeared (KÁZMÉR 1990, MÜLLER 1998, MAGYAR et al. 1999). The erosion style and rate after the end of the Neogene shallow marine to fluvio-lacustrine sedimentation as well as the initial thickness and the maximum extent of the sediments from this depositional cycle have been the key issues of the geological research in Western Hungary for a long time (JÁMBOR 1989, MÜLLER and MAGYAR 1992, JUHÁSZ 1994, JUHÁSZ et al. 1997, MÜLLER 1998, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999, MAGYAR et al. 1999). With the results of volcanic facies analysis of the eroded remnants of the western Pannonian Basin an estimation of the erosional level of the volcanoes and the thickness of the eroded Neogene siliciclastic units of the small-volume intraplate volcanoes was possible. The level of exposure of pyroclastic rocks of the diatreme facies and the location of crater rim deposits and their relationship with lava flows and/or pre-volcanic rock units show that the erosion rate since the end of volcanism has not exceeded a few tens of metres per million years. This suggests that the total thickness of the Neogene sedimentary cover in the BBHVF region did not exceed 450 m. The preserved intact volcanic landforms in the LHPVF suggest that those volcanic landforms might have been buried, and were therefore well-preserved against erosion. Such an interpretation is supported by geophysical data that indicate the existence of volcanic structures below Quaternary sediments in the LHPVF region (TÓTH 1994). There are clear evidences that intrusive processes also were taking place resulted in form of sill and dyke intrusion into the Neogene sediments (e.g. Sümegprága). Because gravimetry and geomagnetic methods are not able to distinguish intrusive and extrusive bodies that are covered by younger sedimentary units, the reconstruction of the palaeosurfaces defined by coherent lava flow units are predominantly based on the surface exposures of coherent lava bodies that are inferred to be extrusive. However, Pliocene volcanic rocks in the western Pannonian region that are covered by younger siliciclastic (e.g. not related to deposition in volcanic depressions) sediments are only known from the LHPVF and the hills north of the Keszthely Mts. It is inferred that very recent fluvial processes played a role in the exhumation of tuff rings in the LHPVF. The palaeogeomorphological reconstruction that is made predominantly for the BBHVF is an attempt to calculate the relative pre-volcanic surface elevations for each volcanic centres on the basis of identified volcanic lithofacies relationships (NÉMETH et al. 2003b). Identified pyroclastic rocks in respect to their acci-

dental lithic fragment contents and types, the texture of the volcanic glass shards, the micro and macro texture of the preserved pyroclastic rocks as well as the presence or lack of coherent lava facies helped to establish negative and positive landforms such as maars and scoria cones with lava lakes. According to their original morphology the position of the pre-volcanic surface of a certain age was established (NÉMETH and MARTIN 1999b, NÉMETH et al. 2003b). Here a brief summary is given to present a few results that show the usefulness of such studies in respect to the estimation of the morphology during syn-volcanic time and the estimation of the thickness of eroded rock units.

Calculation method of erosion on the basis of eroded monogenetic volcanoes

The calculation of erosion rates can be separated into two steps;

1. estimation of erosion since the start of volcanism, and

2. estimation of erosion between the end of the Neogene shallow marine to fluvio-lacustrine sedimentation and the start of volcanism (Plate 2.2). This calculation is based on the assumption that the created volcanic landforms have not been covered by young sediments after their formation, and they have been stand against subaerial erosion following their eruption. This assumption is well-constrained in most of the locations in the BBHVF. However, the fact that some covered volcanic rocks (pyroclastic and coherent lava flows) exist in the LHPVF indicates that syn- and post-volcanic lacustrine sedimentation took place there and therefore these proposed two step erosion calculation model cannot be fully applied for the LHPVF.

It is generally accepted based on lithofacies and field relationships of lacustrine beds, seismic sections, and palaeontological evidence (MAGYAR et al. 1999, SACCHI et al. 1999, SACCHI and HORVÁTH 2002) that sedimentation in the Pannonian Lake terminated at 8 My ago in the BBHVF region. On the basis of seismic sections as well as facies relationships of Pliocene siliciclastic sediments it is inferred that local shallow lakes and wide fluvial systems existed well after the Pannonian Lake vanished in the LHPVF. There is also a general agreement that the water depth of the Pannonian Lake in western Pannonian region prior the volcanism was not more than 50 m (MÜLLER 1998, MÜLLER et al. 1999). There is evidence that tuff rings in the LHPVF were filled by thick lacustrine sediments (BENCE et al. 1978, SOLTI 1986, FISCHER and HABLY 1991, BRUKNER-WEIN et al. 2000), indicating a general availability of water through porous media aquifers and suggesting some extent of surface water involvement in their crater-lake formation. Moreover, pyroclastic successions of the Kis-Somlyó tuff ring (Plates 2.1 and 2.2) exhibit features suggesting that they were deposited in shallow water (MARTIN and NÉMETH 2002, 2004). Volcanic remnants of the LHPVF are generally flat, lensoid in plan view (pyroclastic mound like) with low angle bedding dip directions of the juvenile clast-rich lapilli tuffs. The dip direction of the pyroclastic beds commonly point to a former centre of the respective volcano suggesting a very broad volcanic edifice with low rims that easily have been flooded by water from a braided river system or from a shallow (e.g. few m) lake (Plate 2.1). This process could have been responsible for a short burial time and preservation of volcanic edifices under younger deposits. Such buried tuff rings are well known from the region (TÓTH 1994) and suggest, that volcanic landforms of the LHPVF are exhumed rather recently as can be seen by the relatively intact tuff rings such as the Gérce–Sitke system (Plates 2.1 and 2.2 and Figures 2.1 and 2.2). The implication of this observation of the LHPVF is that the LHPVF is rather erosion than an accumulation surface, which developed due to inversion from accumulation to sudden erosion very recently.

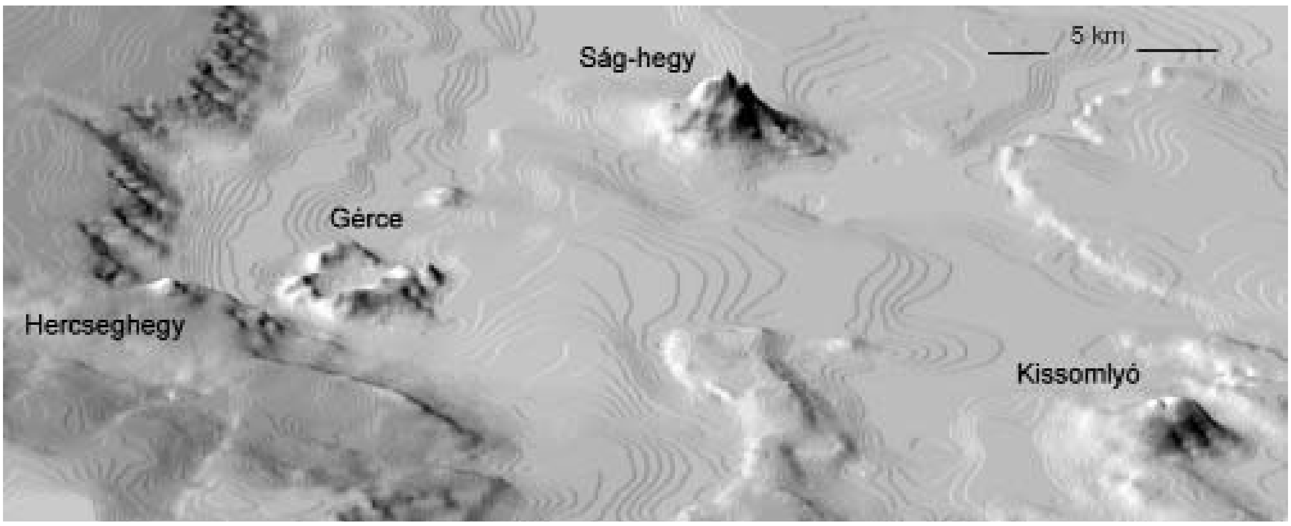


Figure 2.1. Digital terrain model for the LHPVF showing exhumed, intact tuff rings (such as the Gérce) that have been excavated recently

K/Ar as well as the recent $^{39}Ar/^{40}Ar$ ages of volcanic rocks in the BBHVF and LHPVF suggest a long-lasting activity (~8 to 2.3 My – BALOGH et al. 1982, 1986, BORSY et al. 1986, BALOGH and PÉCSKAY 2001, BALOGH and NÉMETH 2004, WIJBRANS et al. 2004). The age distribution of the measured eruptive centres strongly varies. The position of the eruptive centres in the different age groups from the BBHVF shows a slight westward shift of eruptive centres along a NE–SW line through time. The dominant part of the eruptive centres is grouped into the 4 to 2.8 My age group and show a more pronounced westward shift of the NE–SW trending line vents (Plate 2.1). A shield volcano in the north (Plate 2.1) was formed under "dry" eruptive conditions, suggesting little surface and groundwater availability during its eruption. The Hajagos-hegy, Gulács and Badacsony (Plate 2.1) represent large wide tuff ring with maar depressions, which were filled by thick (up to 50 m) lava flows or lava lakes. The eruptive centres show a slightly N–S elongated form. Clearly visible, especially in the Hajagos-hegy (Plate 2.1), is a lava flow, which cut through the former crater rim and flowed onto the pre-volcanic swampy area. These, N–S and NE–SW aligned centres suggest that explosion occurred in a N–S; NE–SW streams valley system. The ellipsoid shape of the volcanic centres may reflect an influence of the syn-volcanic stress field in the area as suggested from many volcanic fields (NAKAMURA 1977, CONNOR et al. 1992, 2000, CONNOR and CONWAY 2000).

Post-volcanic erosion

To estimate post-volcanic erosion, volcanic lithofacies were studied at each erosional remnant of the BBHVF (NÉMETH and MARTIN 1999b, NÉMETH et al. 2003a). Identification of different volcanic facies allowed the establishment of the exhumation level of the crater-filling deposits, vent zones or deep subsurface structures of individual volcanoes such as maars, tuff rings or scoria cones (Plate 2.3). With the relative proportions of different pyroclastic facies as well as the dip direction of the bedding planes, the position of the present exposures below the syn-volcanic palaeosurface was estimated (NÉMETH and MARTIN 1999b). As indicated by the presence of angular sideromelane glass shards and a large proportion of accidental lithic clasts in the pyroclastic rocks, most of the volcanic remnants were produced by subsurface phreatomagmatic explosive eruptions (NÉMETH and MARTIN 1999a, b, MARTIN et al. 2003). Most of the original landforms are interpreted as negative forms such as maars and/or near syn-volcanic palaeo-surface forms such as tuff ring craters; however, most of these landforms were subsequently filled by Strombolian scoria cones and/or lava flows (MARTIN et al. 2003). Using geometrical relationships between crater depth and width as well as thickness of crater rim deposit (LORENZ 1985, 1986), the size of the original volcanic landform and the present level of erosion was estimated (NÉMETH and MARTIN 1999b). According to the physical volcanological observations based on detailed mapping around the individual volcanic erosional remnant the total erosion was estimated for each volcano (NÉMETH and MARTIN 1999b). The major eruptive centres of the BBHVF with large negative Bouguer-anomalies, strong positive geomagnetic anomalies and occurrence of primary phreatomagmatic (and occasional reworked maar crater fill sediments) products suggest that these are maar structures that undercut the syn-volcanic palaeo-surface and often were filled with coherent lava bodies. In areas where mostly scoriaceous pyroclastic rocks were found with no negative Bouguer-anomaly the original landforms are interpreted as having been formed on a syn-volcanic palaeo-surface which is represented by the contact elevation between the volcanic and pre-volcanic rock units (NÉMETH and MARTIN 1999b). Where lava capped buttes are located above areas with significant negative Bouguer-anomalies and phreato-magmatic deposits are common in the pyroclastic succession, the original eruptive centres were maars (NÉMETH and MARTIN 1999b). The measured diameter of the remnant of the lava filled maar basin is inferred being almost equal in

Figure 2.2. Detail of an aerial photograph (Hungarian Military Collection) from the tuff ring east of Gérce village showing a more or less perfectly preserved crater rim, which very likely was buried similarly to other still covered volcanic edifices in the LHPVF

size as the original maar basin diameter (d – Figures 2.2 and 2.3). The maximum crater depth (D) was calculated according to LORENZ (1986);

$$D = d/5$$

wherein d = measured/estimated crater diameter, which is based on statistical studies of relatively young maars (Figure 2.4). Three basic type of erosional remnants have been identified and dealt with separately (NÉMETH and MARTIN 1999b):

1. maar basin filled by late magmatic scoria cones, lava lakes, which were then eroded to different levels;
2. maar basins developed and filled by late reworked tephra deposits, fresh water carbonates, which were then eroded to different levels;
3. scoria cones, shield volcanoes, commonly in association with lava lakes, spatter cones, which were then eroded.

The first are those centres, which have a thick lava cap overlying maar basins. The syn-volcanic palaeo-surface in relation to the present morphology is calculated by

$$Hp = Hc - w$$

where Hc = the recent plateau surface on the top of the buttes and w = thickness of the crater rim of the tuff ring.

The thickness of the crater rim deposits (w) is estimated to range from a few tens of metres to up to 200 metres according to observations of different types of maars (CAS and WRIGHT 1988). However, the crater rim thickness (w) is strongly controlled by the geometrical size of the maar crater.

The calculation of the erosion on the basis of erosion remnants without lava caps is strongly dependent on the calculated crater depth (D). The crater depth has been calculated after LORENZ (1986). Depending on the facies distribution around the maar complexes, erosion has been calculated by adding the relative amount of eroded material to the measured recent plateau elevation of the pyroclastic rock units (±10; 1/3D; 1/2D; 2/3D – NÉMETH and MARTIN 1999b).

On the basis of this very simple but systematic study an estimate is given for each location, which represents the possible syn-volcanic palaeosurface elevation (Table 2.1). The formula from LORENZ (1986) is based on empirical description of geometrical parameters of maar volcanoes, and it is generalised. It is evident, that in areas where maar volcanoes developed in "soft rock" environment

Figure 2.3. View of the tuff ring east of Gérce exhibiting primary morphological features of its crater rim

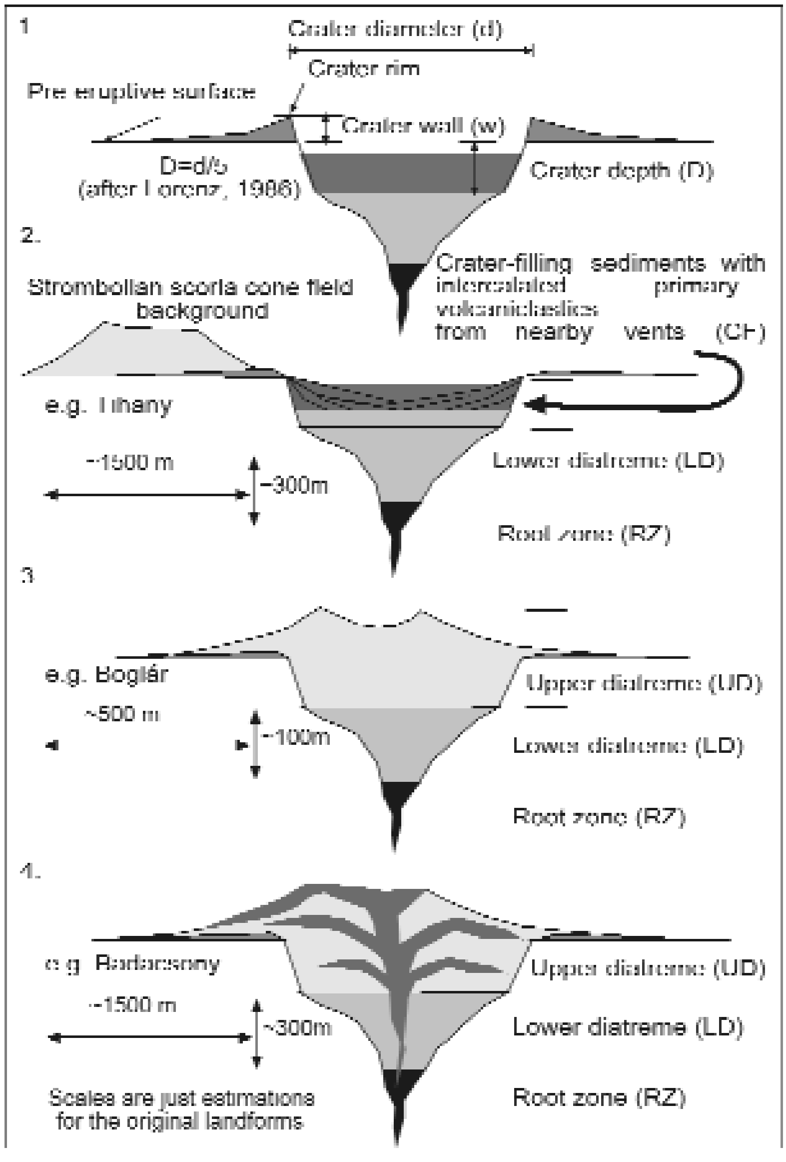


Figure 2.4. Calculation method to estimate the erosion of a monogenetic volcanic landform (NÉMETH and MARTIN 1999b)

Table 2.1. Volcanic erosional remnants of the BBHVF

Locality	Age [My]	Ev [m]	EvR [m/My]	T [My]	Ep1 [m]	Et1 [m]	Ep2 [m]	Et2 [m]	C [m]	P [m]
1. Badacsony	3.45	**270**	78	4.55	**355**	*625*	**46**	*316*	+100	725/416
2. Szt.György-hegy	2.80	**270**	96	5.2	**499**	*769*	**52**	*322*	+100	869/422
3. Gulács	3.47	**260**	75	4.53	**340**	*600*	**45**	*305*	+100	700/405
4. Haláp	2.94	**208**	71	5.06	**359**	*567*	**51**	*259*	-50	517/209
5. Fekete-hegy	2.92	**210**	72	5.08	**367**	*577*	**51**	*261*	-100	477/161
6. Bondoró-hegy	3.00	**180**	60	5.00	**300**	*480*	**50**	*230*	-50	430/180
7. Hegyesd	3.08	**180**	58	4.92	**285**	*465*	**49**	*229*	-100	365/129
8. Szigliget	3.4	**222**	65	4.6	**299**	*521*	**46**	*268*	+150	671/418
9. Fonyód	3.55	**142**	40	4.45	**178**	*320*	**45**	*187*	+150	470/327
10. Boglár	3.5	**130**	37	4.5	**166**	*296*	**45**	*175*	+150	446/225
11. Agártető	3.44	**150**	44	4.56	**201**	*351*	**46**	*196*	-100	251/96
12. Tagyon	3.26	**?**	?	4.74	**?**	*?*	**?**	*?*	?	?
13. Csobánc	3.5	**225**	64	4.5	**288**	*513*	**45**	*270*	+50	563/320
14. Hajagos-hegy	3.94	**180**	46	4.06	**187**	*367*	**41**	*221*	+100	467/321
15. Kopasz-hegy	3.5 ?	**160**	46	4.5	**207**	*367*	**45**	*205*	- 50	317/165
16. Pipa-hegy	3.5 ?	**160**	46	4.5	**207**	*367*	**45**	*205*	- 50	317/165
17. Kékkút	3.5 ?	**160**	46	4.5	**207**	*367*	**45**	*205*	- 100	267/105
18. Kerekimajor	3.5 ?	**80**	23	4.5	**104**	*184*	**45**	*125*	- 50	134/75
19. Öreg-hegy	3.5 ?	**90**	26	4.5	**117**	*207*	**45**	*135*	+ 100	307/235
20. Horog-hegy	3.5 ?	**192**	55	4.5	**248**	*440*	**45**	*235*	- 100	340/135
21. Füzes-tó	3.5 ?	**160**	46	4.5	**207**	*367*	**45**	*205*	- 50	317/165
22. Kishegyestű	3.5 ?	**178**	51	4.5	**230**	*408*	**45**	*226*	+ 50	458/276
23. Fekete-hegy N	4.66	**180**	39	3.34	**130**	*310*	**33**	*213*	- 100	210/113
24. Tálodi-erdő	4.65	**82**	18	3.35	**60**	*142*	**34**	*116*	+150	292/266
25. Pula	4.25	**120**	28	3.75	**105**	*225*	**36**	*156*	+100	325/256
26. Kabhegy	4.93	**200**	41	3.07	**126**	*326*	**31**	*231*	- 100	226/131
27. Bondoró	5.54	**180**	32	2.46	**79**	*259*	**25**	*205*	- 50	209/165
28. Hegyestű	5.97	**166**	28	2.03	**57**	*223*	**20**	*186*	- 100	123/86
29. Kabhegy (old)	5.23	**-**	-	2.77	**-**	-	**-**	-	-	-
30. Tóti-hegy	5.71	**256**	45	2.29	**104**	*360*	**23**	*279*	+ 100	460/379
31. Tagyon2	5.69	**229**	40	2.31	**92**	*321*	**23**	*252*	- 100	221/152
32. Sátorma	4.53	**184**	41	3.47	**142**	*326*	**35**	*219*	- 50	276/169
33. Tihany	7.54	**232**	31	0.46	**14**	*246*	**5**	*237*	+ 150	396/287
34. T.dörögd	4.5	**-**	-	3.5	**?**	*?*	**?**	*?*	?	?
35. T.dörögd	4.5	**-**	-	3.5	**?**	*?*	**?**	*?*	?	?
36. Zánka/Várhegy	6 ?	**160**	27	2	**54**	*214*	**20**	*180*	-100	114/80
37. Véndeg-hegy	3 ?	**140**	47	5	**235**	*375*	**50**	*190*	-50	325/140
38. Hármashegy	3.5 ?	**220**	63	4.5	**284**	*504*	**45**	*265*	+100	604/365

Numbers correspond to locations on Plate 1.1, B. Age data from BALOGH et al. (1982), BORSY et al. (1986) and BALOGH (1995). Abbreviations: Ev = post-volcanic erosion, EvR = post-volcanic erosion rate, T = time between end of Pannonian sedimentation and start of volcanism, Ep1 = erosion between end of Pannonian sedimentation and start of volcanism calculated with the same erosion rate estimated in post-volcanic time, Et1 = total erosion using Ep1 and Ev, Ep2 erosion between end of Pannonian sedimentation and start of volcanism calculated with low erosion rates (10 m/My), Et2 = total erosion using Ep2 and Ev, C = correction value, P = total thickness of Pannonian sediments

the maar basin will be rather broad, and shallow, in contrast to "hard rock" environment, where the maars could be very deep (LORENZ 2000, 2003a, b). Maar volcanoes often show transition toward tuff ring, and their geometrical parameters could differ from LORENZ (1986) empirical formula, such as it is known from maar and tuff ring volcanoes from Oregon where ratio of crater depth to diameter often is 1 to 10 (HEIKEN 1971). Deep maar volcanoes very likely need hard rock environment such as the Crater Elegante, Mexico (cut in lava flow succession – d, 270 – D, 1650 – GUTMANN 1976) or Joya Honda, Mexico (cut into limestone units – d, 300, – D,1200 – ARANDA-GOMEZ and LUHR 1996) to maintain the LORENZ (1986) average 1 to 5 ratio between crater depth to diameter. This implies that the calculations from the western part of the BBHVF where the phreatomagmatic volcanoes cut into soft rock (Figure 2.3), is in overestimate (e.g. few tens of metres), and the calculated erosion values should be viewed as maximum values (Table 2.1).

Detailed analyses on thin-sections and hand specimens from the pyroclastic rock units revealed quartz, quartzo-feldspatic aggregates, plastically deformed mud chunks (mm to cm scale), and muscovite from all of the studied pyroclastic outcrops from the BBHVF (NÉMETH et al. 2003a). All of these components are characteristic for the Neogene siliciclastic sediments. The presence of lithic fragments or exotic minerals derived from these units in the phreatomagmatic pyroclastic rocks attest that these clasts must have been disrupted by the phreatomagmatic explosions, recycled in the phreatomagmatic vents to become part of various lithofacies in and around the vents (NÉMETH et al. 2003b). It can be concluded that post-volcanic (Pliocene) erosion rates vary between 18 and 96 m/My; however, mostly 50 m/My has been calculated (NÉMETH and MARTIN 1999b, NÉMETH et al. 2003b). Re-establishing the syn-volcanic palaeosurfaces in the region, a uniform surface below 100 m with relative elevation variation may be visible (Plate 2.4).

Erosion rate between end of Pannonian sedimentation and start of volcanism — pre-volcanic erosion

Calculation of the pre-volcanic erosion rate in the area is problematic. Few direct field evidence is available to help reconstructing palaeosurfaces for times prior to 8 My. The use of the same erosion rates for pre-volcanic times as for post-volcanic times would imply that the erosional potential was the same ever since the end of Pannonian Lake sedimentation. Based on this simplification, the amount of erosion since the end of the Pannonian Lake sedimentation would be more than twice as large as the amount of erosion since the end of volcanic activity. The total thickness of accumulated sediments in the area would have exceeded 800 m in basins with an estimated erosion of more than 600 m. However, field evidence such as lava flows over weakly dissected morphology and the general uniform elevation of lava flow contact to pre-volcanic rock units indicates that erosion rates must have been lower before the volcanism than after the end of volcanic activity (NÉMETH et al. 2003b). The youngest deposits of Pannonian age in the BBHVF are lacustrine limestones (Nagyvázsony Mésző Formation — NMF). These are intercalated with strongly altered basaltic tuffs (BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999) that are inferred to have been deposited from phreatomagmatic eruptions of the oldest volcanoes in the region (Tihany, 8 My). The limestone tends to form an extensive plateau, outlining the extent of a closed laguna that developed in the final stage of Pannonian Lake sedimentation (BUDAI et al. 1999). Extensive lava flows derived from the volumetrically largest volcano of the BBHVF (Kab-hegy, shield volcano) were observed to overlie the youngest units inferred to have been deposited from the Pannonian Lake (NMF) at uniform elevation (~300 m a.s.l. — BUDAI et al. 1999). As indicated by the results of K/Ar age determinations, textural and compositional similarities, mineral orientation and drill core data, lava sheets of the major shield volcano (Kab-hegy) can be correlated with dissected lava sheets located south of this centre (Tálodi-erdő — JÁMBOR et al. 1981). The large (10 km-scale) extent of lava on a uniform elevation and the undisturbed contact between lava and limestone suggest that the palaeosurface on which the lava erupted was that with insignificant fluvial incision and poorly developed valleys. The age difference between these lavas (~5 My) and the cessation of Pannonian sedimentation (~8 My) suggests that the interval of ~3 My was not long enough to develop a significant valley system. A low erosion rate of e.g. ~10 m/My was inferred for the region in pre-volcanic times. Based on these calculations, the total erosion after the end of the Pannonian Lake sedimentation can be maximalized to ~300 m, while the total thickness of the Pannonian sediments accumulated in the region was about at ~450 m (NÉMETH et al. 2003b).

Conclusion

Based on the analyses of the erosional remnants of monogenetic volcanoes from the BBHVF, the following conclusions can be drawn:

— the post-Pannonian erosion must be separated into two stages;
  — prior to the start of Mio/Pliocene volcanism, and
  — between the Mio/Pliocene volcanism and present;

— post-volcanic erosion rates vary between ~20 and 100 m/My, resulting in a total erosion of 80 to 270 m of predominantly Pannonian sediments;

— in reconstructing the Pliocene syn-volcanic palaeosurface, a remarkable flat landscape can be constructed with a total geomorphic relief of less than ~100 m;

— considering that most of the volcanoes had at least an initial phreatomagmatic eruptive phase, their position marks local lowlands;

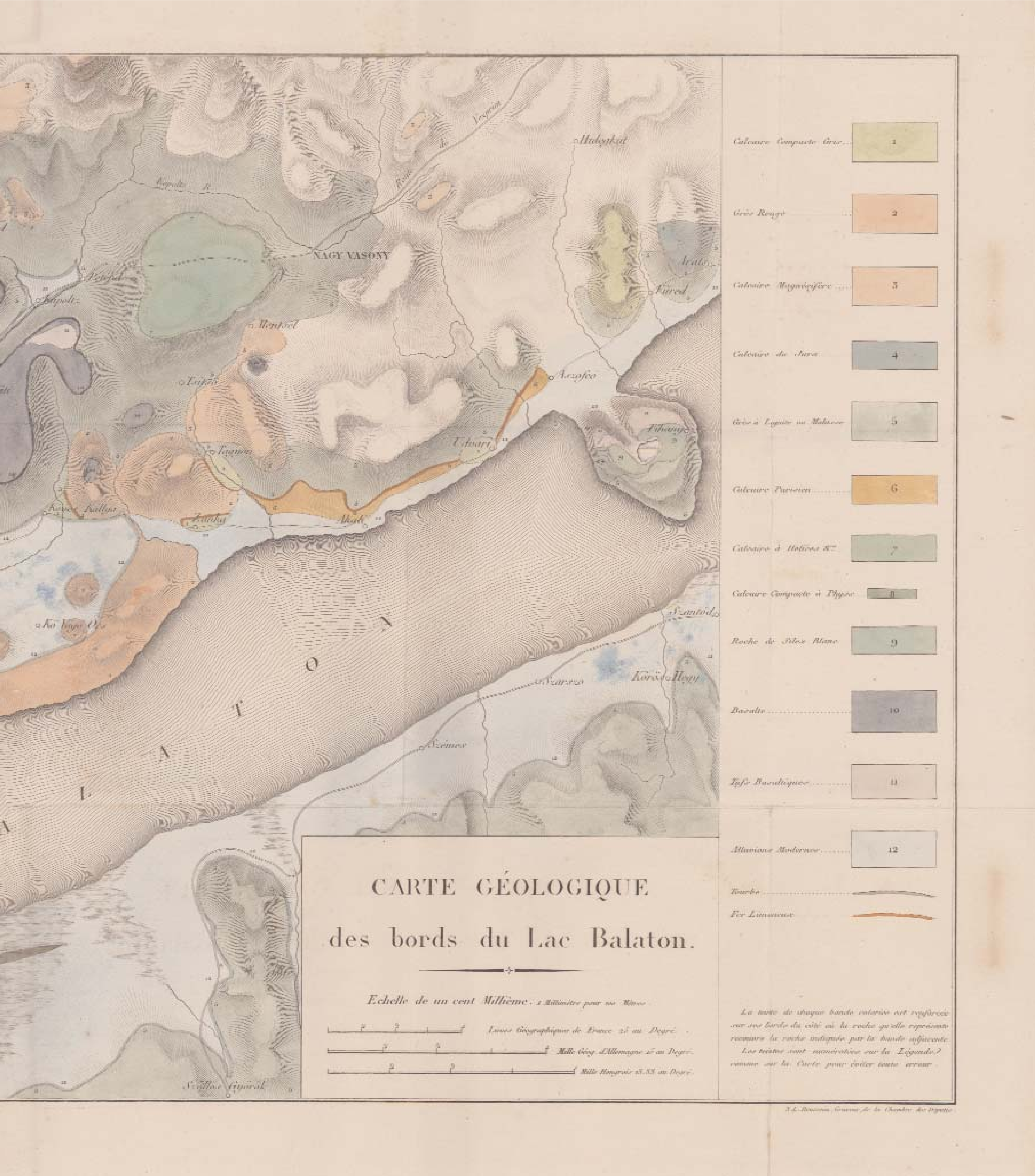
— the total thickness of the Pannonian sedimentary cover is estimated to have been 250 to 900 metres prior to erosion starting at ~8 My, most realistically being not more than 450 m;

— Pannonian sedimentary cover must have been still widespread in the region before volcanism started. Pannonian sediments were only stripped away from elevated ridges.

For the geomorphologic development of the BBHVF two major models have been considered on the basis of the erosion remnants of the Neogene basaltic volcanoes. Originally, the first observations suggested that volcanic erosional remnants represent volcanoes that gradually developed on an erosion surface (LÓCZY 1913, 1920, CHOLNOKY 1918). This would imply that the younger a volcano the deeper was its position in comparison to the elevation of stratigraphy markers (Figure 2.5). The other model postulates that the volcanoes evolved in a sedimentary basin with a fast accumulation rate, and they were buried quickly. Therefore the erosion gradually excavated older volcanic structures creating a "layer-cake" age relationship, having the older volcanic remnants in deeper level (which is represented by their present

Figure 2.5. One of the first geological map from the BBHVF made by a French geologists Beudant F.S. in 1822 shows in a very graphic way the uniform top level of the volcanic erosion remnants
This map was one of the first in Hungary where they used the metric system during mapping

topographic elevation – JÁMBOR and SOLTI 1976, JÁMBOR 1980, 1989, JÁMBOR et al. 1981). This conclusion is supported by some apparent intercalation of Neogene siliciclastic and pyroclastic sediments on the basis of drill core data (JÁMBOR and SOLTI 1975, JÁMBOR 1989). However, recent studies confirmed that most of the sequences intercalated between volcanic and siliciclastic units are near continuous accumulation of accidental lithic rich primary pyroclastic rocks (NÉMETH et al. 2001). Moreover, the pyroclastic rocks in the BBHVF are interpreted to represent often diatreme



filling rocks that cut into a syn-volcanic surface, thus their stratigraphy position is hard to establish (NÉMETH et al. 2003a). The present morphology however is remarkably uniform (Figures 2.6 and 2.7) in spite of the apparent age differences among volcanic erosion remnants. This means that volcanoes erupted in a very gentle morphology with no or underdeveloped drainage systems. Pyroclastic rocks that form erosional remnants of small intraplate volcanoes have been inter-

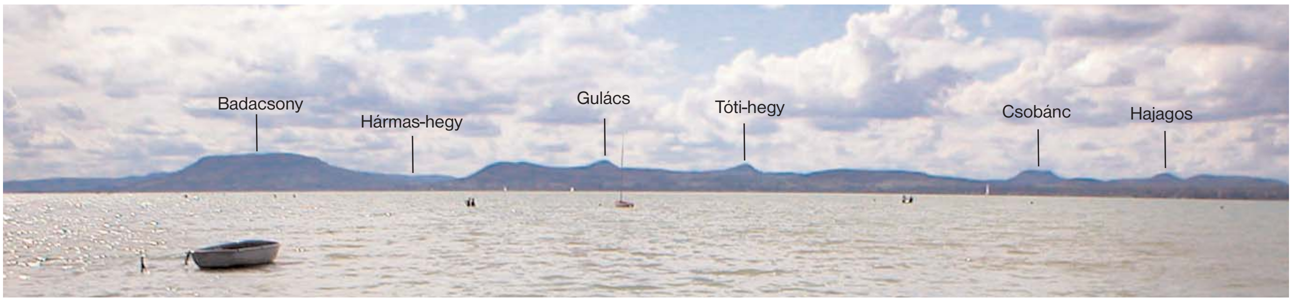


Figure 2.6. Uniform top level of the volcanic erosion remnants (named hills) of the central part of the BBHVF as it looks like from the southern shore of the Lake Balaton

Figure 2.7. Uniform top level of the volcanic erosion remnants (named hills) of the Tapolca Basin from the southern shore of the Lake Balaton. Note the more or less same elevation of the large volcanic butte (Badacsony) in the left hand side of the picture and the Palaeozoic ridge in the right hand side suggesting a very gentle relief in syn-volcanic time on that the volcanoes developed

preted to be exposed diatreme rocks long time ago from the Styrian Basin in Austria (e.g. WINKLER 1925, STRAUSZ1943). This interpretation is based entirely on the stratigraphic and spatial relationships between volcanic, pre-volcanic and post-volcanic rock units (e.g. WINKLER 1925). Regardless of these preliminary interpretations the general geomorphological view of the erosional remants as they are exposed diatreme rocks is new and recently supported and presented here in this summary by textural data of the volcanic rocks.

On the basis of the textural characteristics, as well as the sedimentological features of the preserved pyroclastic rocks of the BBHVF, it is more realistic, that these volcanoes erupted through the pre-volcanic Neogene siliciclastic units in terrestrial conditions, where their erosion immediately modified their shape giving way to develop younger volcanoes on an already eroded landscape. The picture on the LHPVF might be somehow different, however, more detailed studies need to establish temporal and spatial relationships among volcanoes and sedimentation versus erosion across these two volcanic fields.

The estimated erosion rates are in the range of few tens of metres per millions of years. This value is in good agreement with calculations based on other methods from Central Europe (BULLA 1965, MOLNÁR and ZELENKA 1995, KARÁTSON 1996, BAJNÓCZI et al. 2000).

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Acknowledgements: This review is a result of the past ten years research of the authors embedded in a framework of the general knowledge of the tectonic and magmatic development of the Neogene volcanism in the western Pannonian Basin. During this time various organisations supported this research such as, the Hungarian Science Foundation OTKA F 043346, OTKA T 032866, Magyary Zoltán Post-doctorate Fellowship 2003–2004, DAAD German Hungarian Academic Exchange Program 2002/2003 and the DFG (Ma 2440), many thanks to all of these organisations. Constructive suggestions and review by Ferenc Molnár (Eötvös University, Budapest) lifted significantly the quality of this summary and are greatfully acknowledged. General review by Volker Lorenz (Würzburg University, Germany and Tamás Budai (MÁFI, Budapest) helped to find a reader-friendly presentation style of this work. Great consultations about geomorphological processes in the western Pannonian region with Gábor Csillag (MÁFI, Budapest) are gratefully acknowledged.

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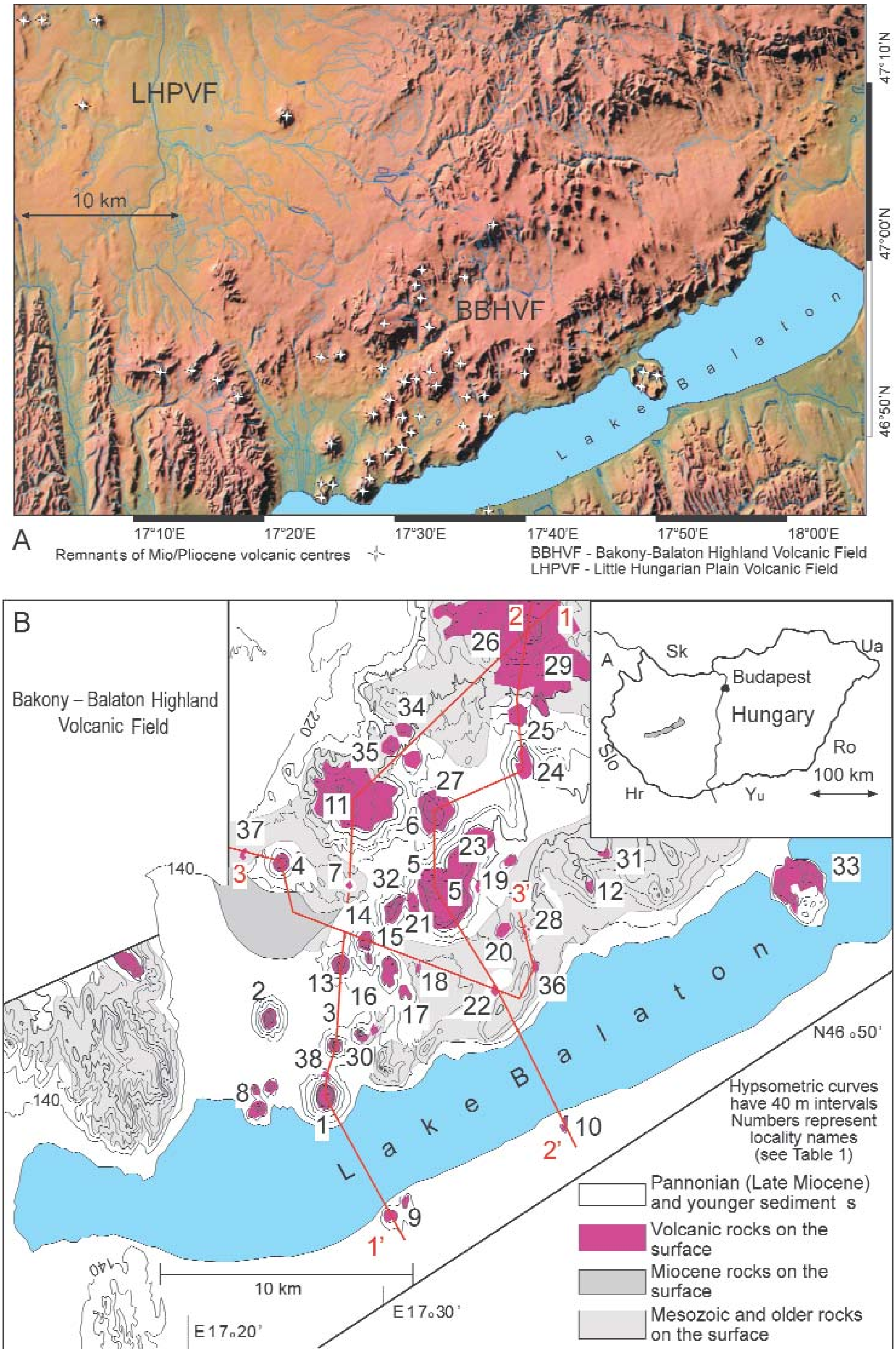
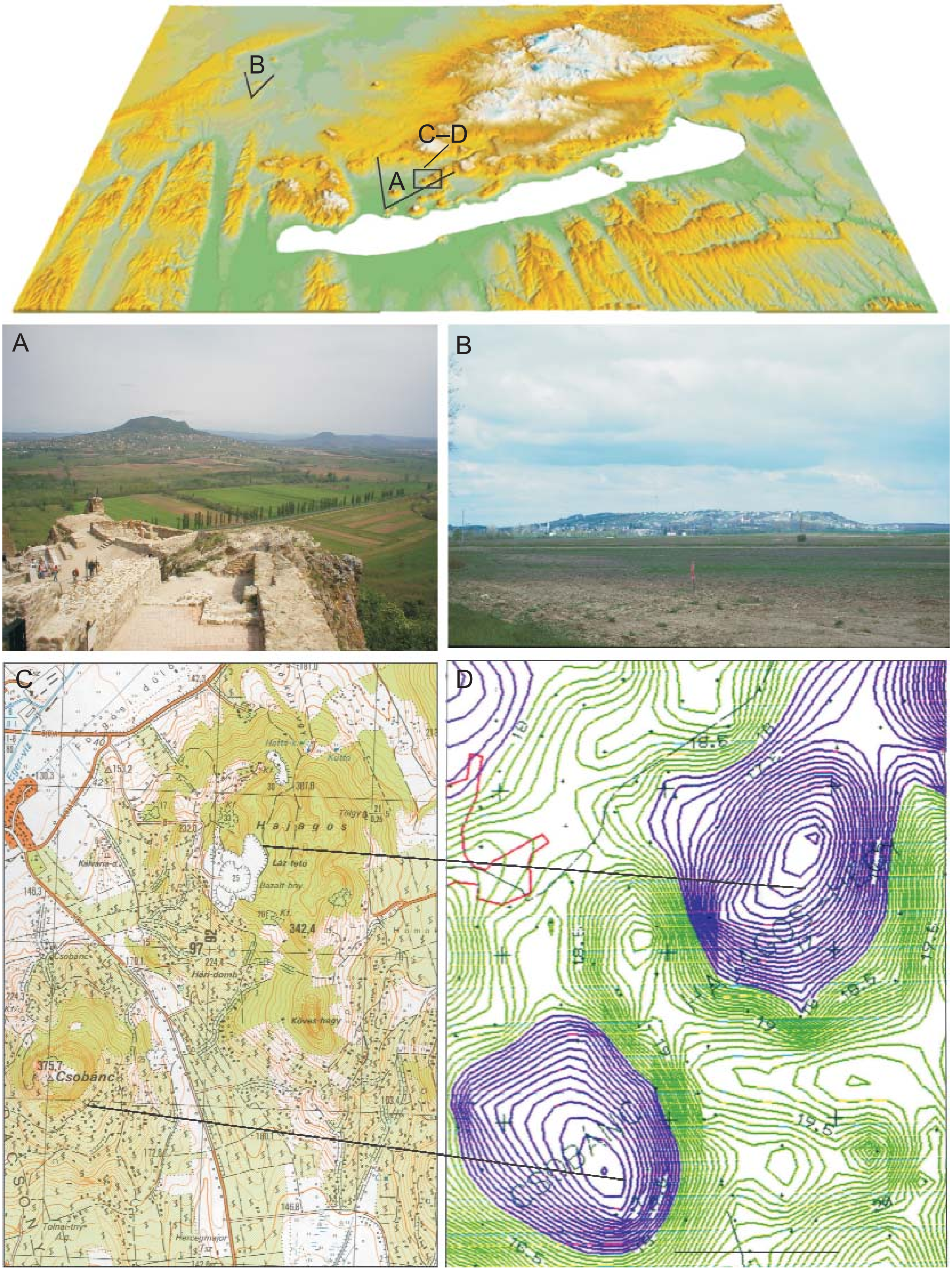
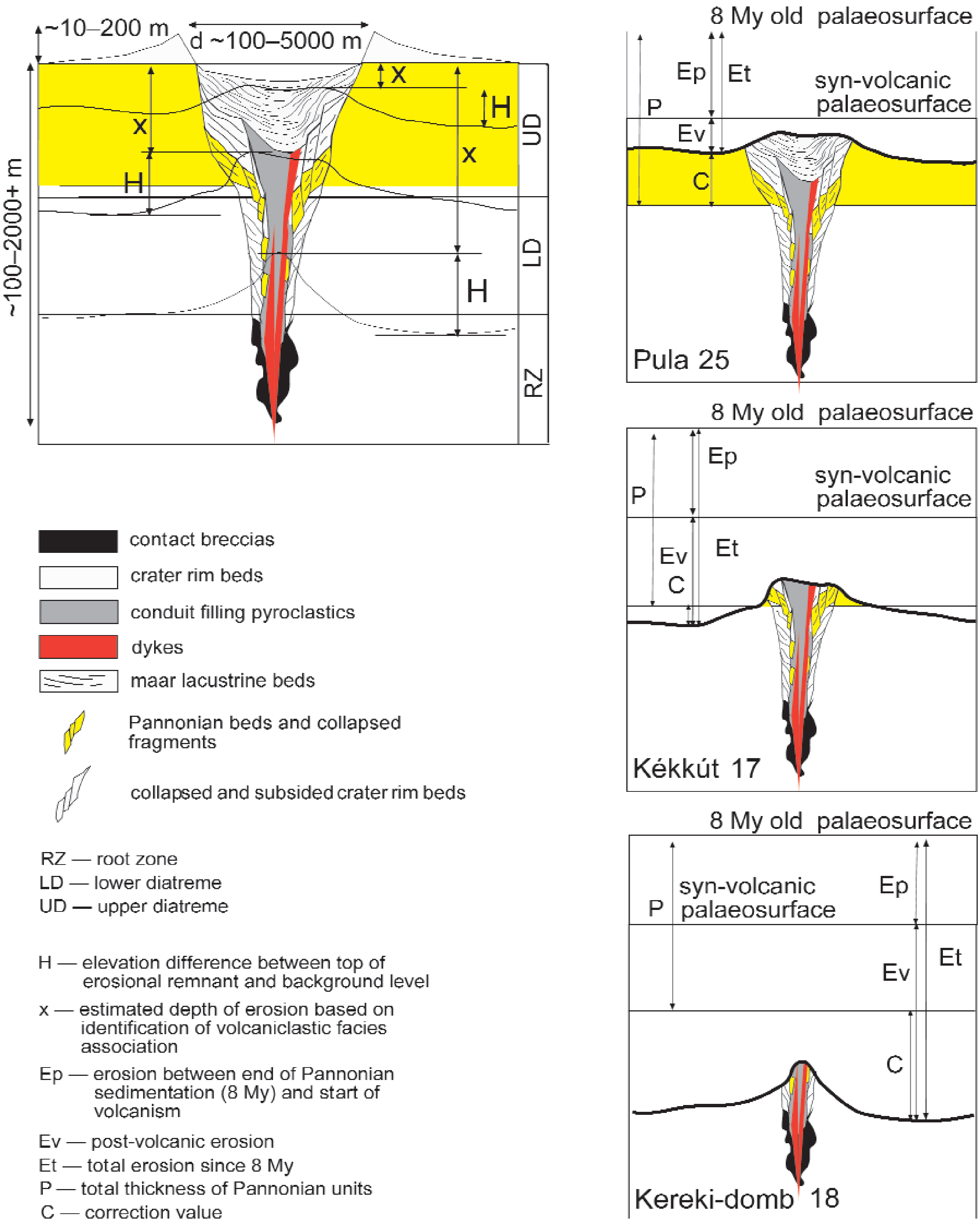
Digital terrain model (A) of the western Pannonian Basin with erosional remnants of small-volume intraplate volcanoes (stars). A simplified geological map of the Bakony – Balaton Highland Volcanic Field (B) gives an overview of the relationship between distribution of volcanoes and type of exposed units. The lines on this map correspond to the cross sections shown on Plate. 2.4. Numbers refer to the identified volcanic erosion remnants and the names are shown on Table 2.1



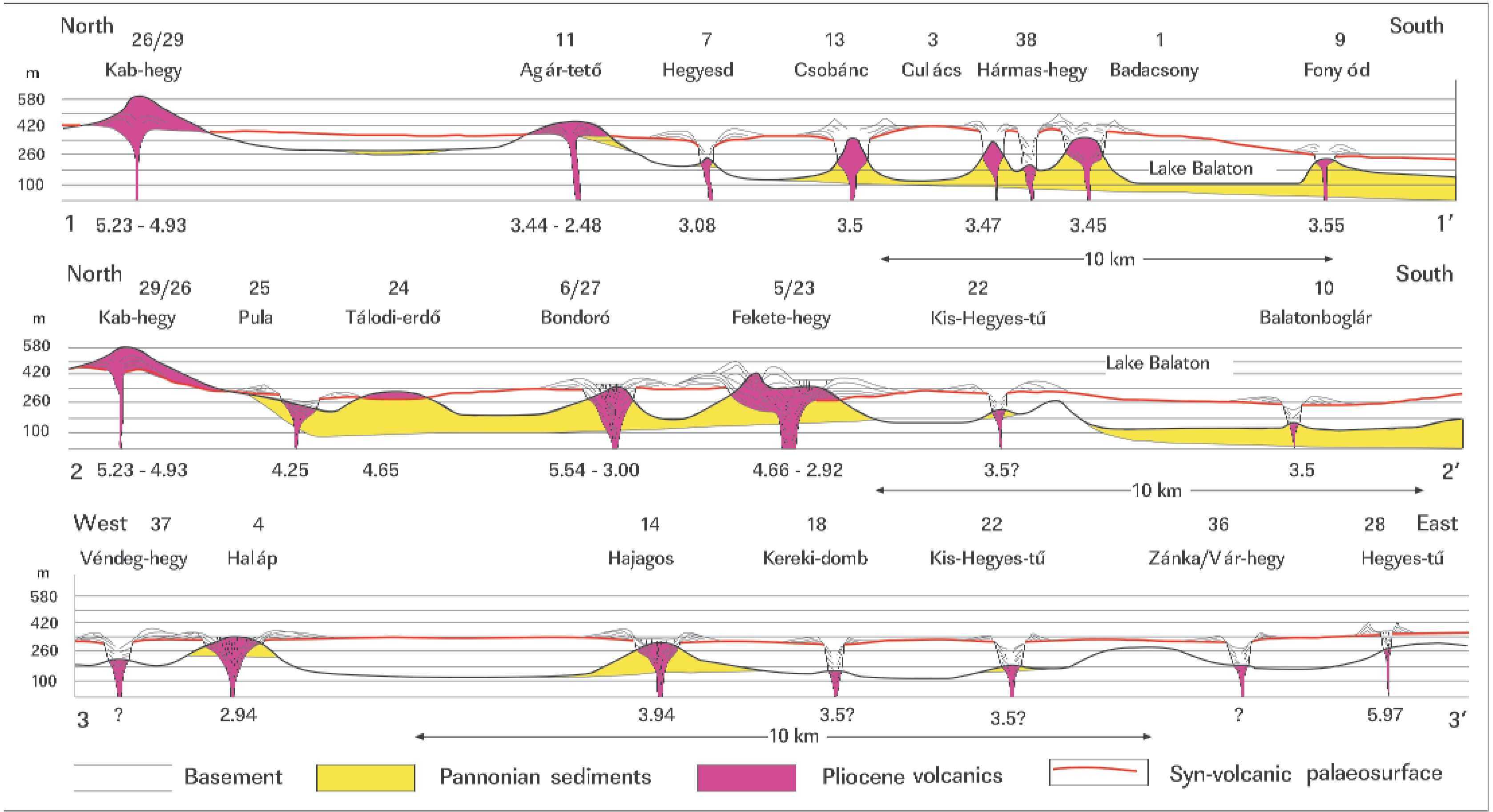
General landscape features of eroded monogenetic volcanoes, A) Szent György-hegy, BBHVF, B) Kis-Somlyó (LHPVF). The general negative Bouguer anomalies where lava capped buttes are located suggests a significant mass deficit below these erosional remnants. On figure D an example is shown from the Hajagos in comaprison to the topography of the hill itself (C). The values on the gravity map are in milligalls and recovered from unpublished data of Kovácsvölgyi (Eötvös Lorand Geophysical Institute, Budapest)



Estimation of the erosion level on the basis of identified volcanic lithofacies from the BBHVF (Németh et al., 2003b). Model of a monogenetic phreatomagmatic volcano and its erosion through time. Note the 3 different stages of erosion can be established by the identification of different volcanic lithofacies from different examples from the field. Abbreviations correspond to Table 2.1



Reconstructed syn-volcanic palaeosurfaces through the BBHVF showing a very uniform, flat landscape (NÉMETH et al. 2003b). Cross-section lines are shown on Plate 2.1, B



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Chapter 3

Second International Maar Conference — Hungary–Slovakia–Germany

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 massé" fall back of collapsing phreatomagnetic eruption collumn, accompined by occasional the vent

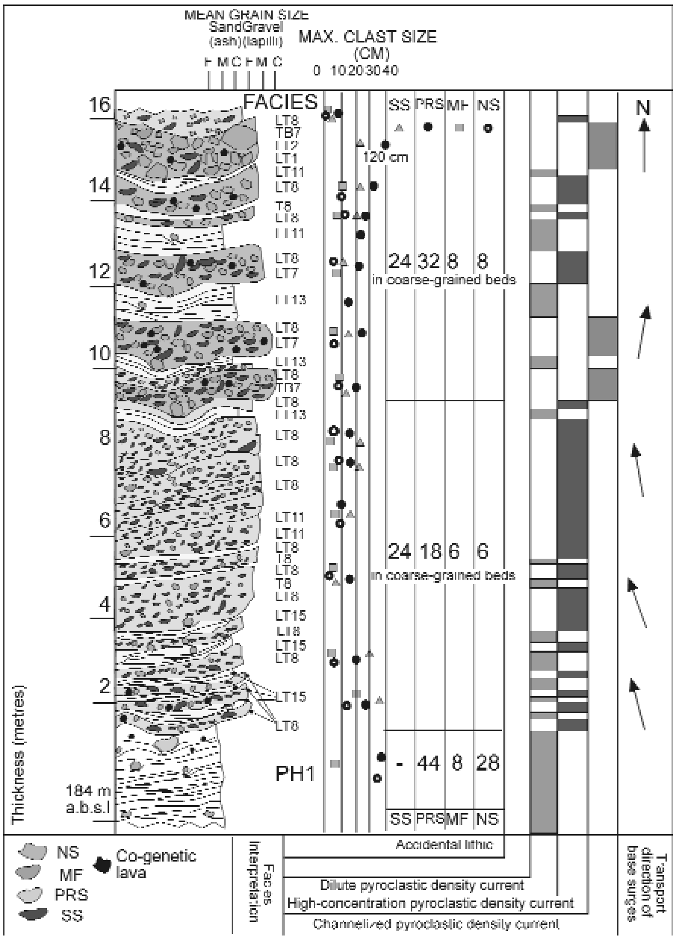

Figure 3.1. Simplified stratigraphic log of Barátlakások (Tihany; NÉMETH et al. 2001) SS = Silurian schist, PRS = Permian red sandstone, MF = Mesozoic formations, NS = Neogene siliciclastic sediments

(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.

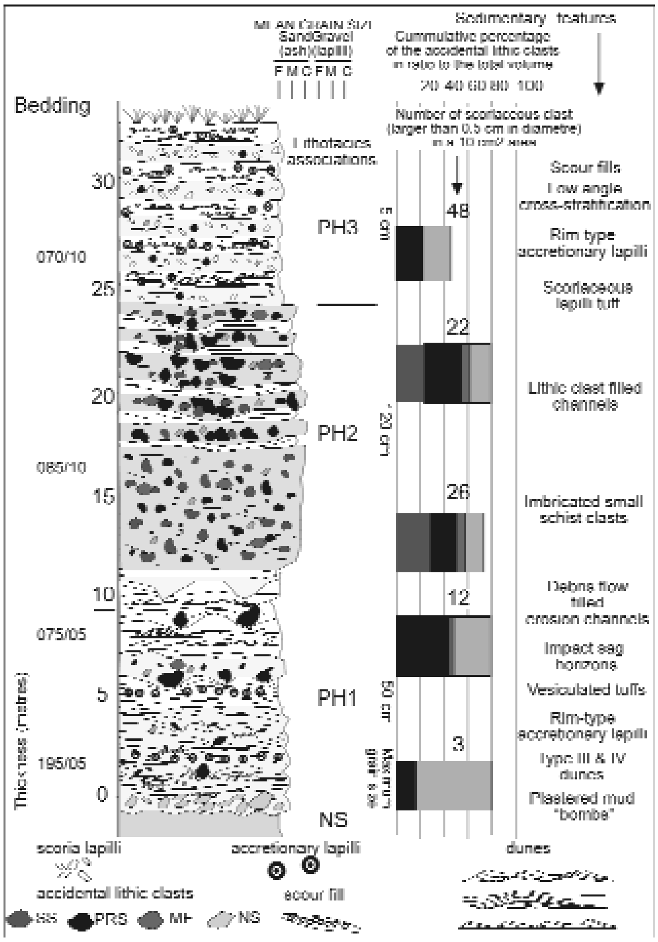

Figure 3.2. Composite stratigraphic log of the eastern outcrops (Barátlakások) at Tihany (Németh et al. 2001). Legend see on Figure 3.1.

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).

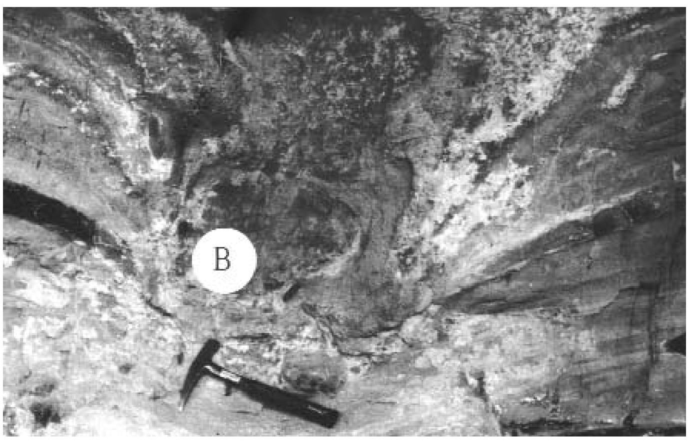

Figure 3.3. Impact sag (B) caused by a ballistically emplaced block from the Permian units, photo is taken from the Barátlakások outcrop

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.

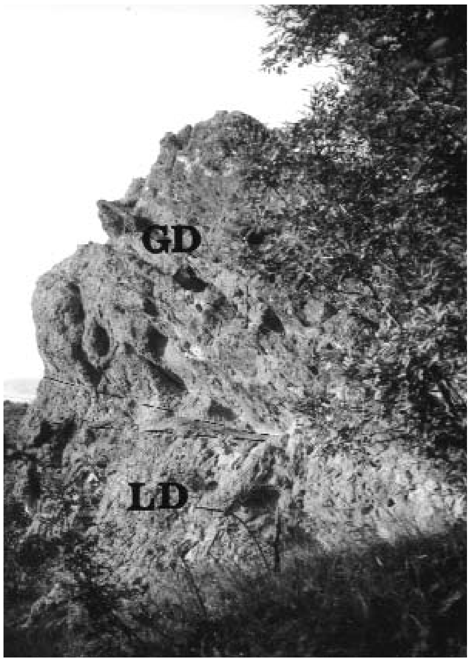

Figure 3.5. Massive, unsorted tuff breccia from PHLD of the basal zone Kiserdő-tető (LD)
The tuff breccia is rich in clasts of every known pre-volcanic rock unit. GD = is interpreted to be a Gilbert-type delta. (NÉMETH 2001)

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

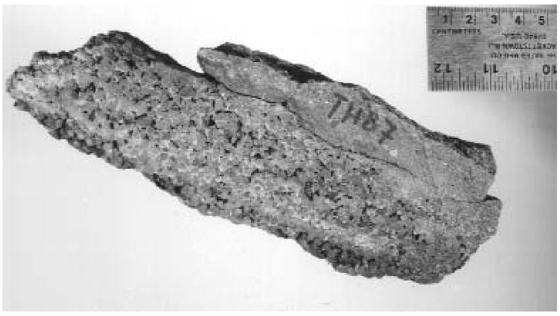

Figure 3.8. Close up of an inversely graded lapilli tuff, which is part of the Gilbert-type delta front from the Csúcs-hegy (western part of the Tihany Peninsula)

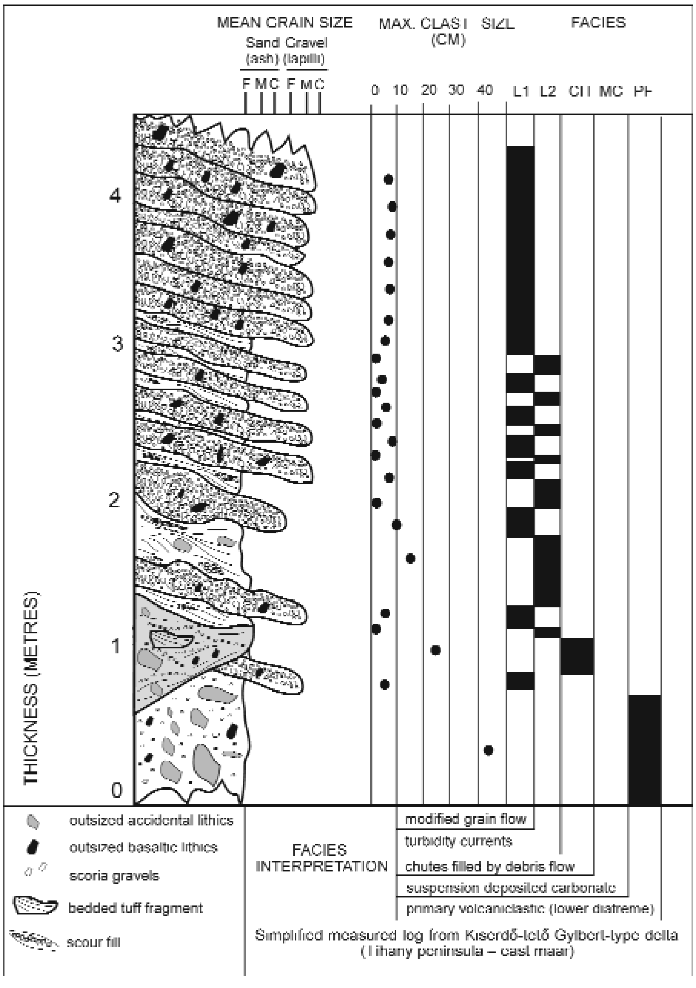

Figure 3.7. Simplified log from the Kiserdő-tető succession exposing the PHLD facies capped by volcaniclastic succession interpreted to be the result of a Gilbert-type delta in a former maar basin (NÉMETH 2001)

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.

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).

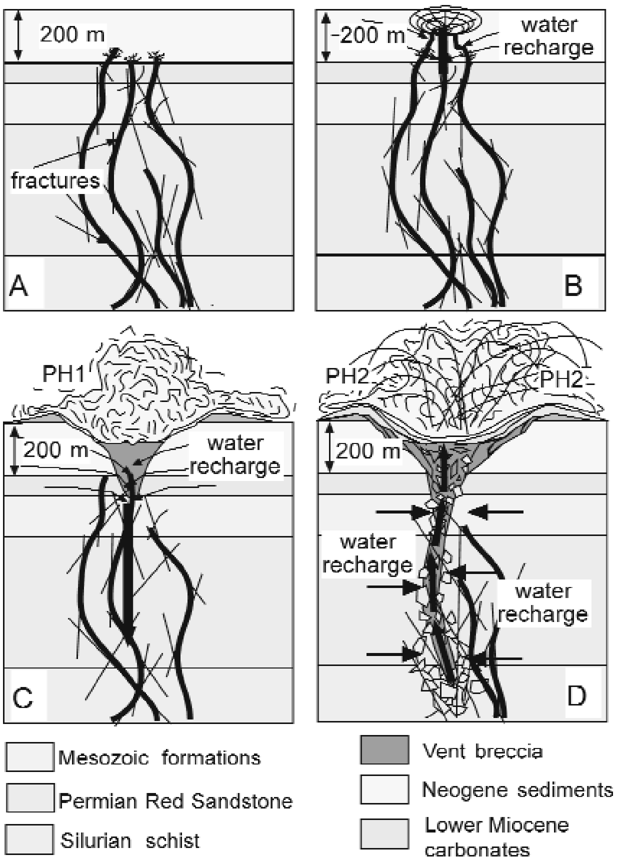

Figure 3.10. Eruption mechanism of the TMVC (NÉMETH et al. 2001)
A, B, C and D represent evolutional sequence of the eruption

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 fluviolacustrine 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

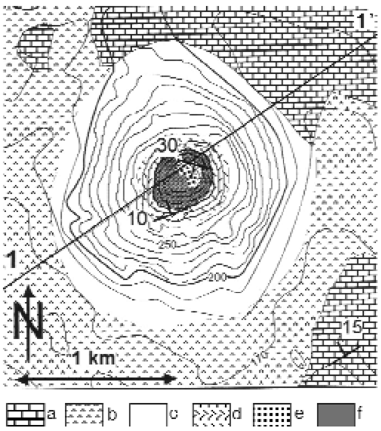

Figure 3.12. Simplified geological map of the region around Haláp maar/tuff ring volcano
a = Upper Triassic carbonates, b = Middle to Upper Miocene limestone, c = neogene siliciclastic units, d = bedded lapilli tuff (crater rim units) e = massive to bedded lapilli tuff (conduit filling facies), f = solidified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

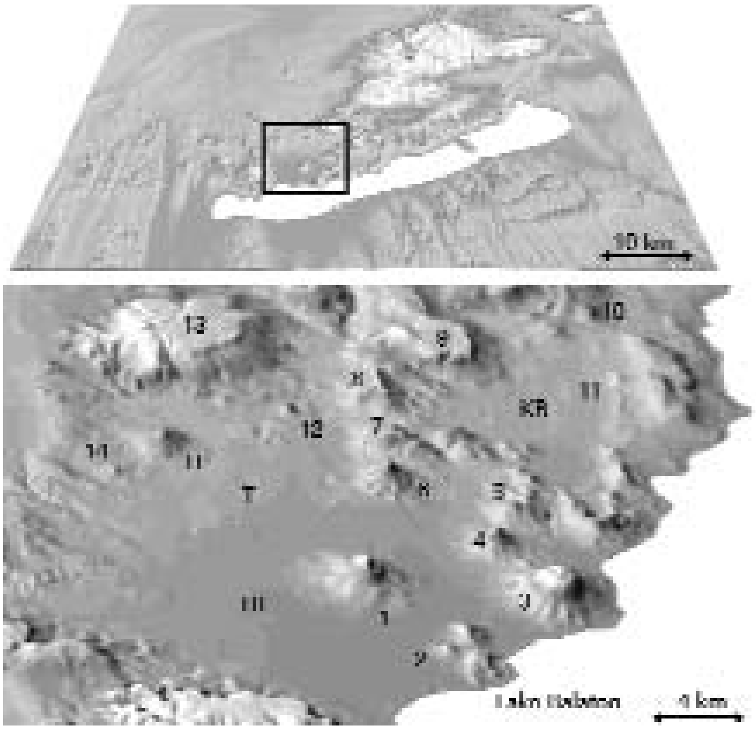

Figure 3.13. Digital Terrain Model of the Tapolca Basin looking from south-west to north-east
H – Haláp, 1 – Szent György-hegy, 2 – Szigliget group, 3 – Badacsony, 4 – Gulács, 5 – Tóti-hegy, 6 – Csobánc, 7 – Hajagos, 8 – Sátorma, 9 – Fekete-hegy, 10 – Hegyes-tű, 11 – Kis-hegyes-tű, 12 – Hegyesd, 13 –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

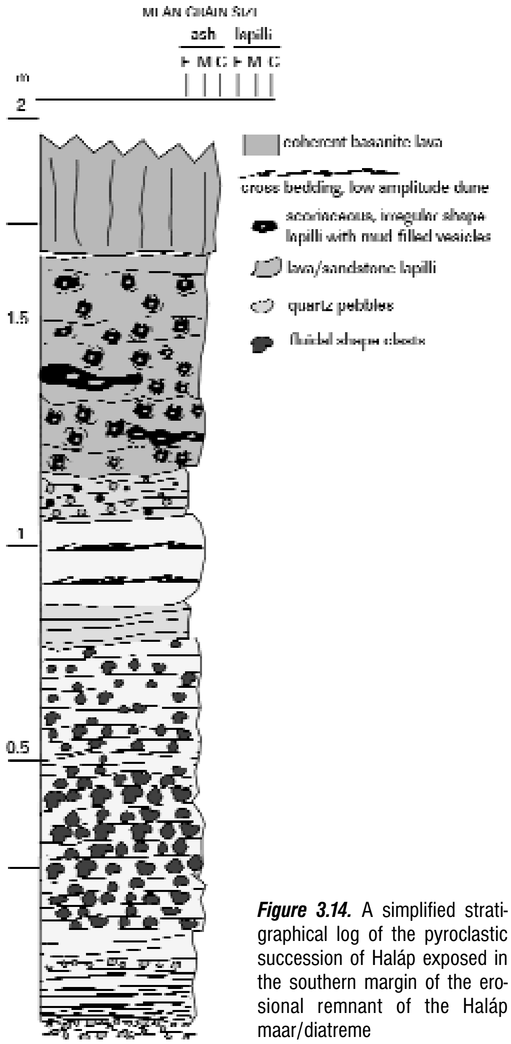

Figure 3.14. A simplified stratigraphical log of the pyroclastic succession of Haláp exposed in the southern margin of the erosional remnant of the Haláp maar/diatreme

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).

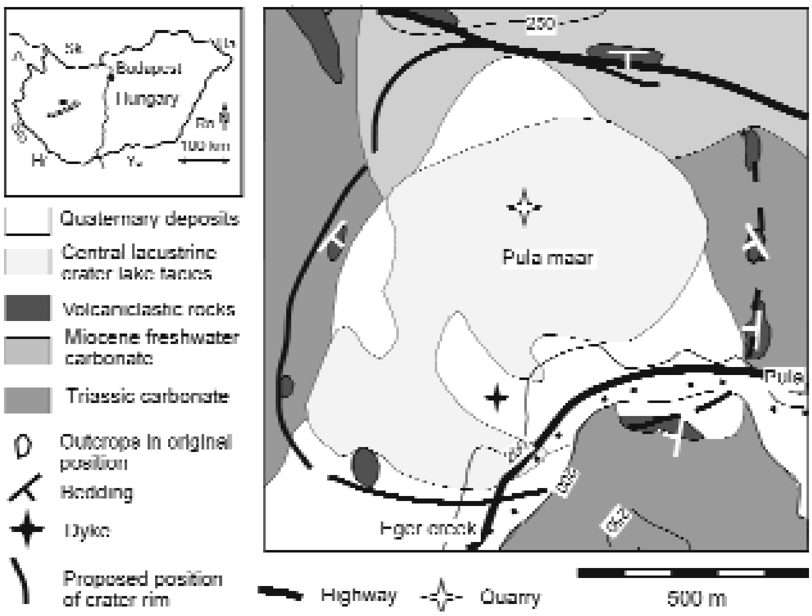

Figure 3.15. Simplified geological map of the Pula maar region

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).

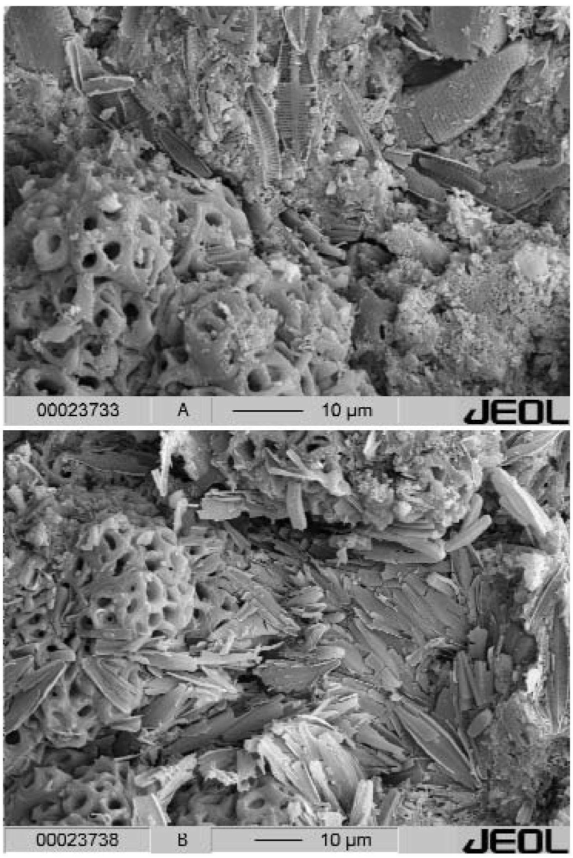

Figure 3.16.
A) Alginite from the Pula maar lake, B) Botryococcus colonies and diatom frustules on SEM images [photo courtesy of Kurt Goth (2004)]

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 –

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)].

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

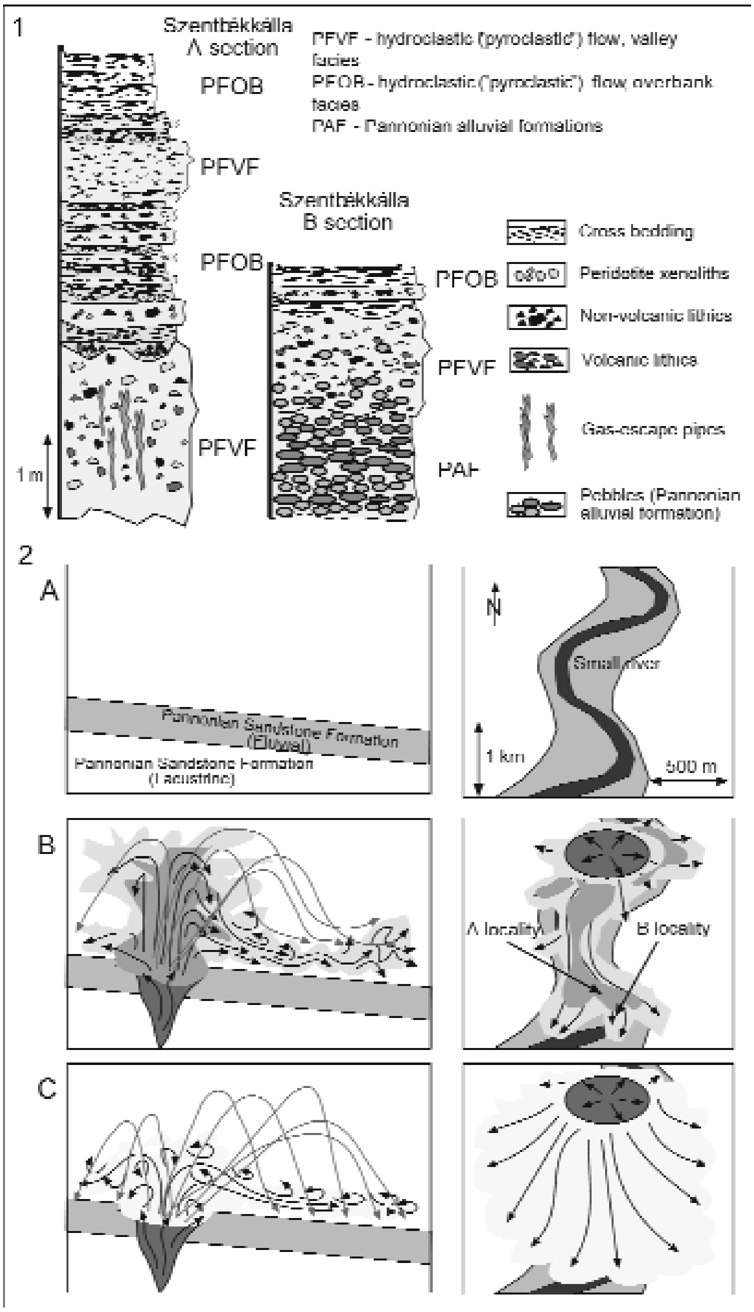

Figure 3.17. Simplified stratigraphic columns (1) from the Szentbékkálla section showing vertical relationship between massive and dune-bedded pyroclastic facies that largely correspond with valley filling and overbank facies of a phreatomagmatic pyroclastic flow (NÉMETH and MARTIN 1999). Localities 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)

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 syn-volcanic 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).

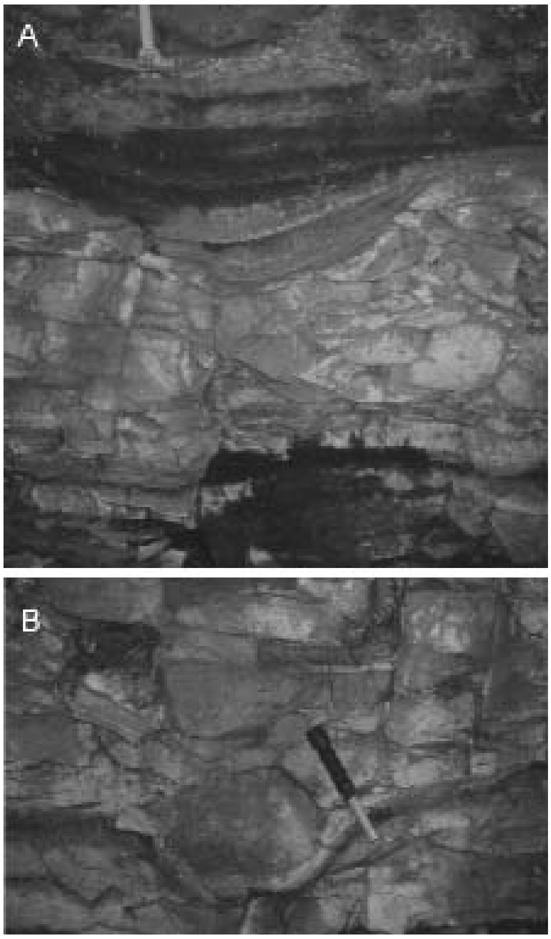

Figure 3.18. Alternating tuff and lapilli tuff (A) of phreatomagmatic origin from the Fekete-hegy southern flank (Vaskapu-árok)

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).

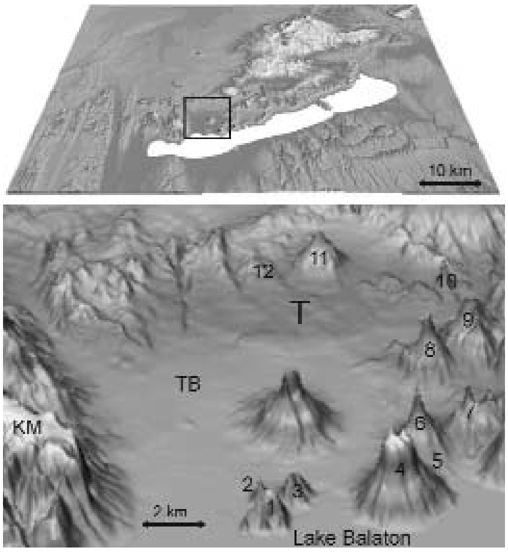

Figure 3.19. DTM model of the Tabolca Basin, showing eroded remnants of small-volume intraplate volcanoes forming small hills up to 350 metres above the basin floor

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

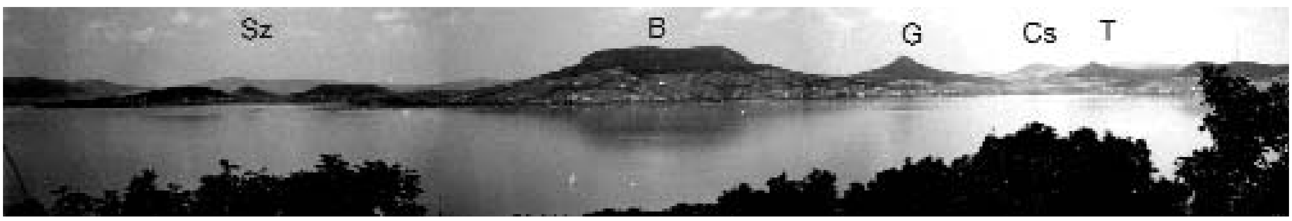

Figure 3.20. View toward Szigliget (Sz), Badacsony (B), Gulács (G), Csobánc (Cs) and Tóti-hegy group (T) from the southern shoreline of the Lake Balaton

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-

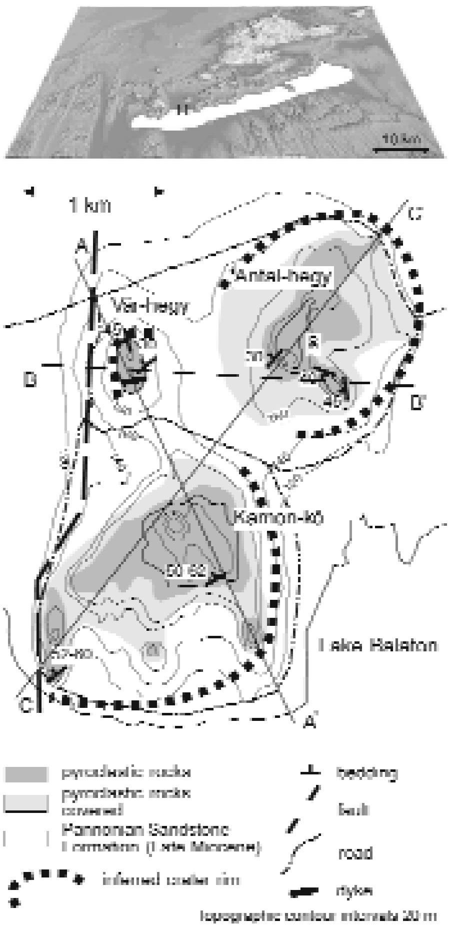

Figure 3.21. Simplified geological map of the Szigliget Peninsula

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

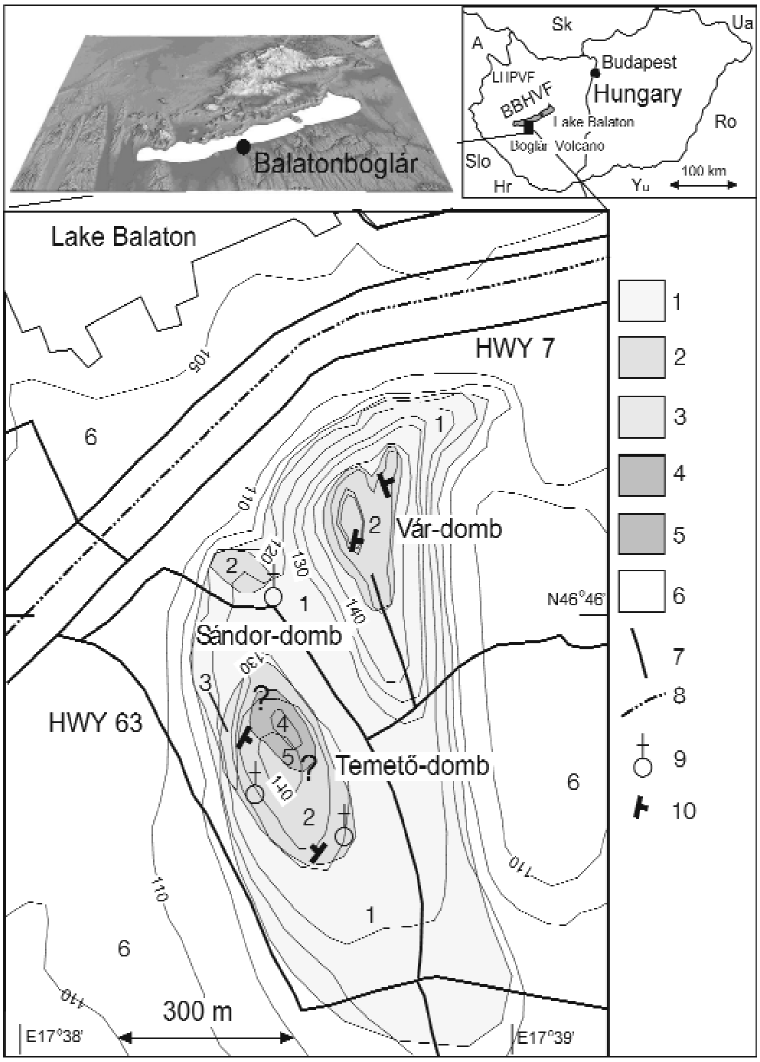

Figure 3.22. Simplified geological map of the Boglár diatreme region (after NÉMETH et al. 1999b)
1 – Neogene rock units, 2 – Lapilli tuff, 3 – Lapilli tuff with fossil tree trunks, 4 – Maar lake sediments, 5 – Peperite, 6 – Quaternary deposits, 7 – road, 8 – Railway, 9 – Church, 10 – Bedding

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.

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.

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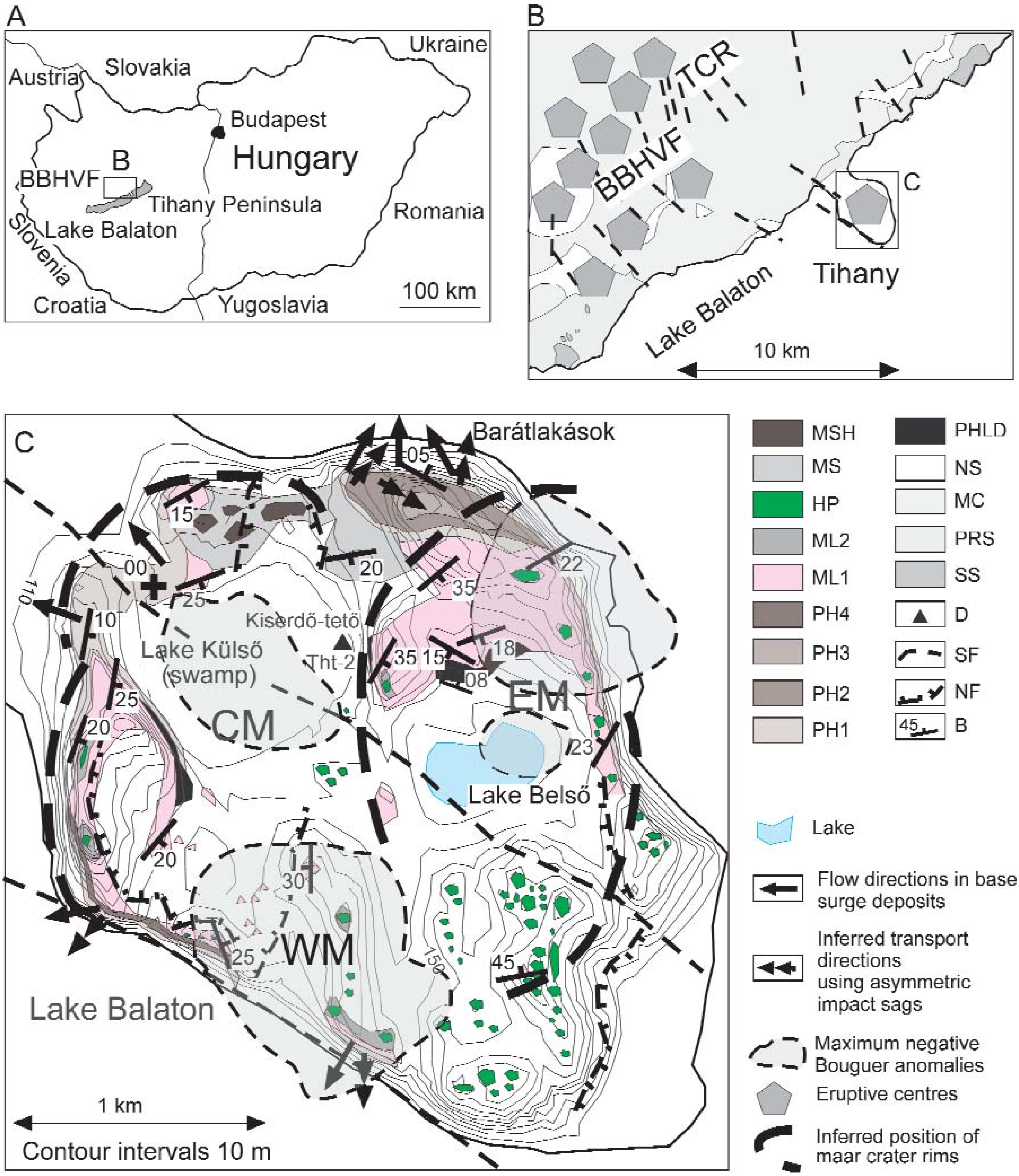
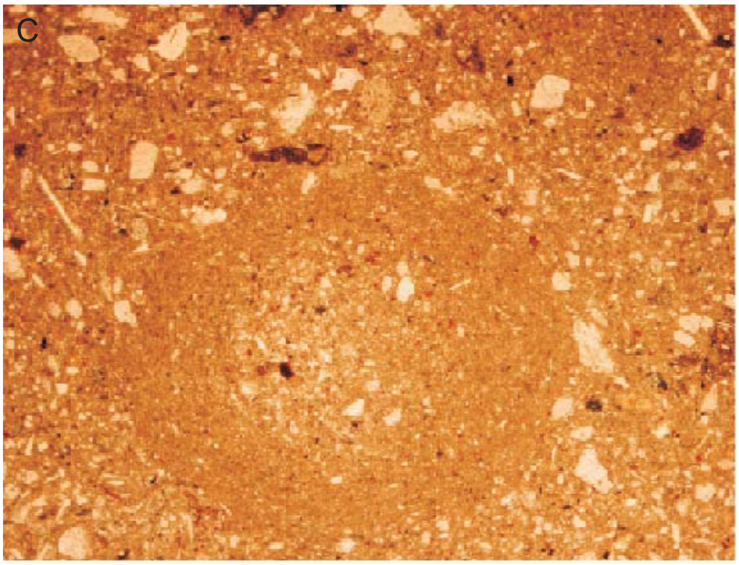
Simplified geological map of the Tihany Peninsula (after NÉMETH et al. 1999, 2001)



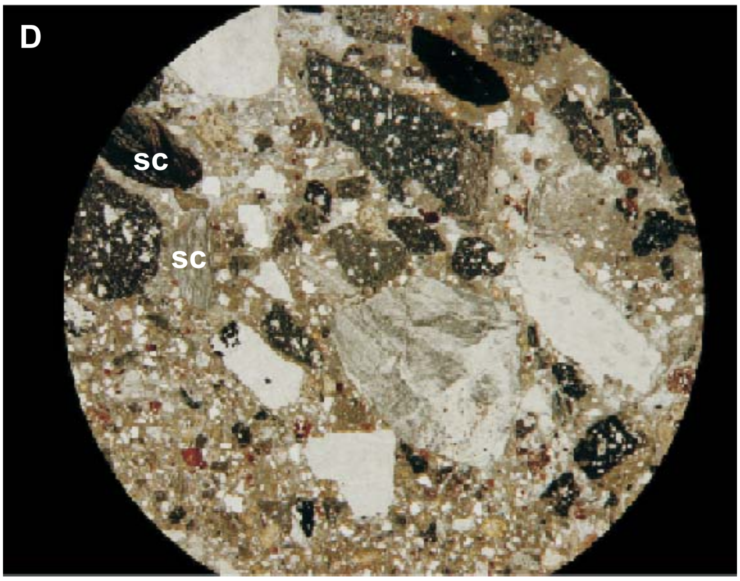
B = bedding
NF = normal fault
SF = strike slip fault
D = drill core
PHLD = phreatomagmatic lower diatreme
NS = Neogene siliciclastic sediments
MC = Mesozoic carbonates
PRS = Permian red sandstone
SS = Silurian schist
MSH = magmatic Strombolian/Hawaiian units
MS = magmatic Strombolian units
HP = hot spring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



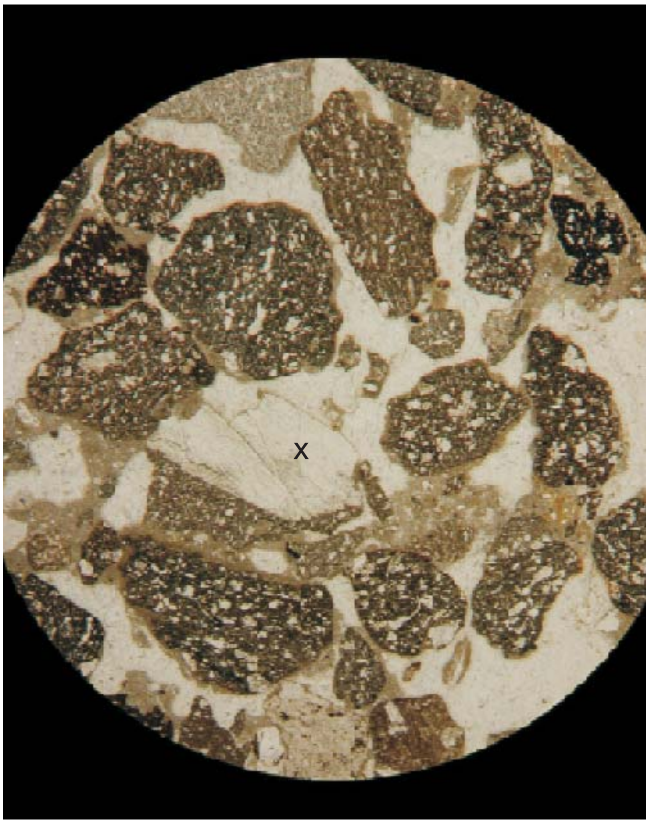
Photomicrographs of a fine-grained tuff predominantly composed of fragments derived from the Neogene siliciclastic units. Photo is 1cm across (paralell polarised light)



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

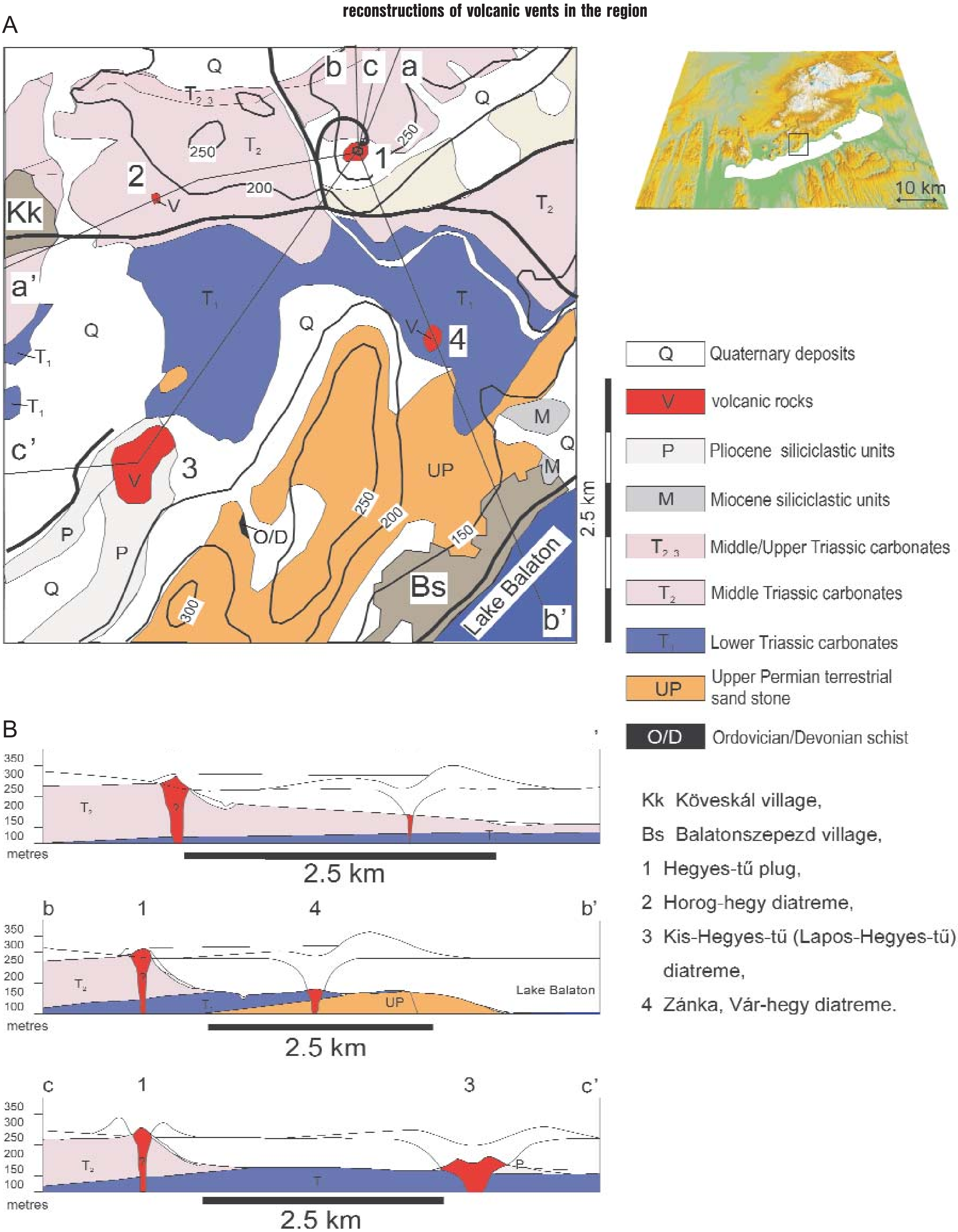


Photomicrographs of a lapilli tuff inferred to be part of the Gilbert-type delta front. Note the large amount of rounded lapilli and broken pyrogenic and xenocryst (x), cemented by calcite. The photo is 2 cm across

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)

Plate 4 / Chapter 3 *Second International Maar Conference — Hungary–Slovakia–Germany*

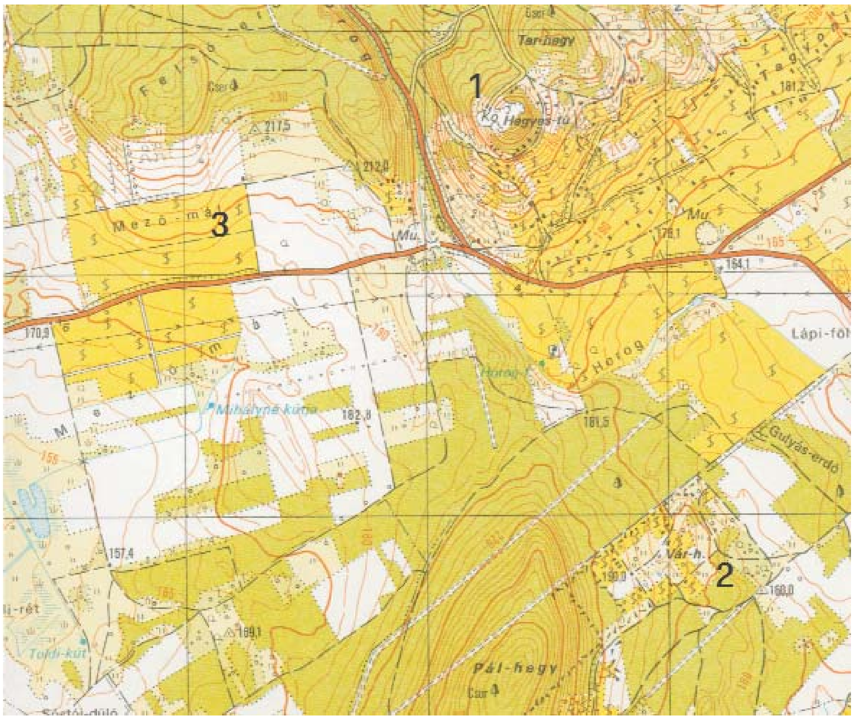
A) Simplified geological map of the eastern part of the Kál Basin. B) Cross sections through the eastern part of the Kál Basin show possible reconstructions of volcanic vents in the region



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

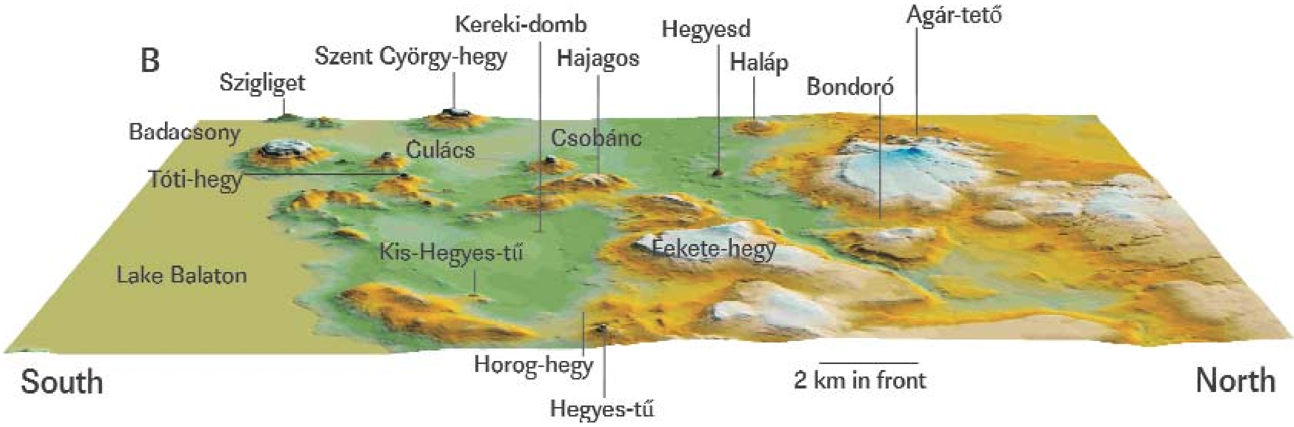
1 – Hegyes-tű, 2 – Zánka, Vár-hegy, 3 – Horog-hegy. Rectangular grid spacing on the map is 1 km. Map is a detail from the 1 to 25,000 scale topography map of Hungary

A

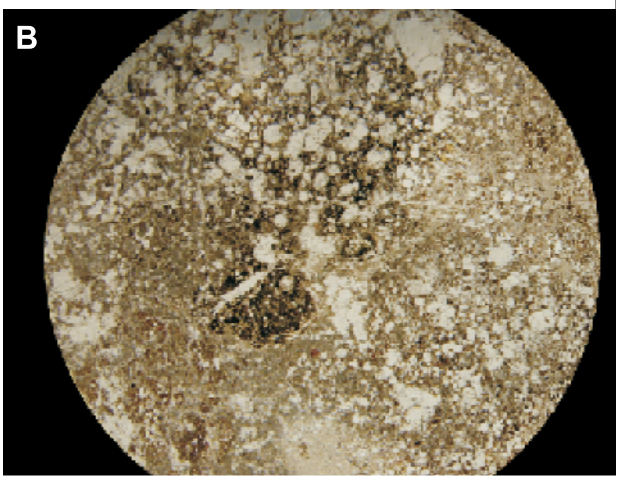


Panoramic view toward the Kál Basin on DTM from the Hegyes-tű. In the centre of the basin are small diatreme remnants (labeled) identified as small morphological irregularities

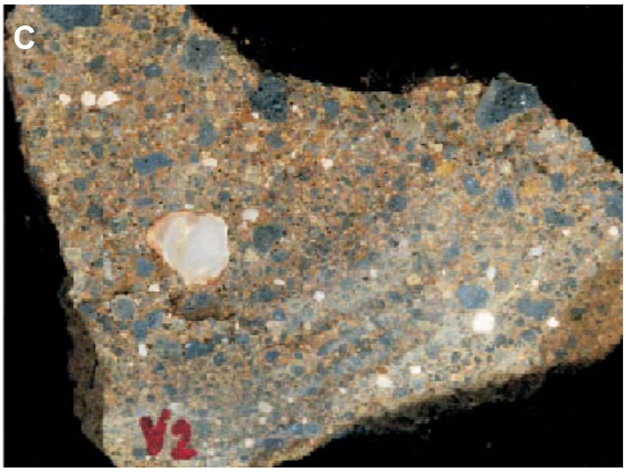
B



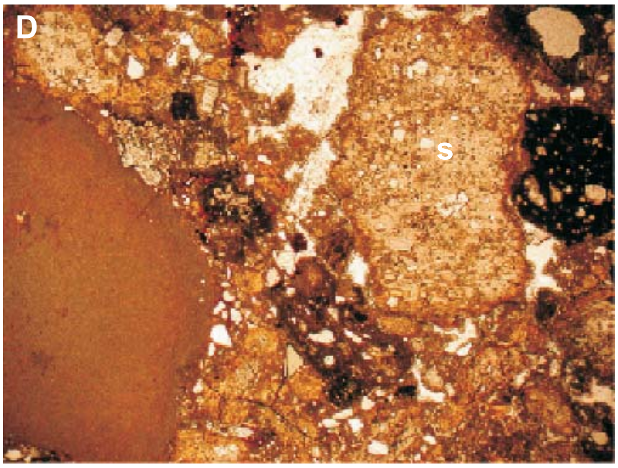
Highly vesicular basanitic lava clasts in a strongly palagonitized, mud-rich matrix adjacent to the coherent lava body of Hegyes-tű. The clastic zone is somehow surrounded by coherent lava indicating its well-localized structure, perhaps a bubble, that formed inside of the still liquid basanitic lava



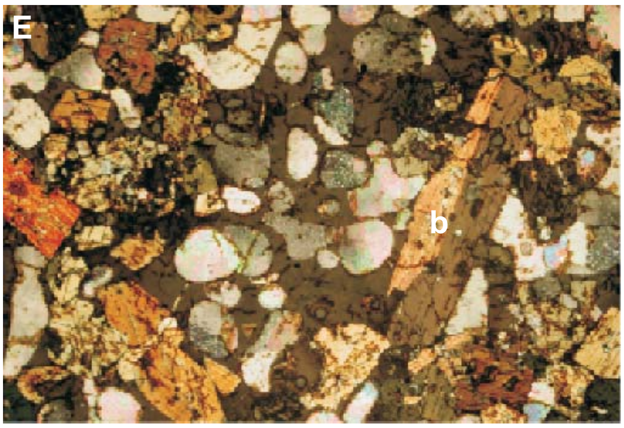
Photomicrograph of the pyroclastic breccia near the Hegyes-tű basanite plug. Note the irregular shape of tachylitic lapilli (black). The picture is ~2 cm across.



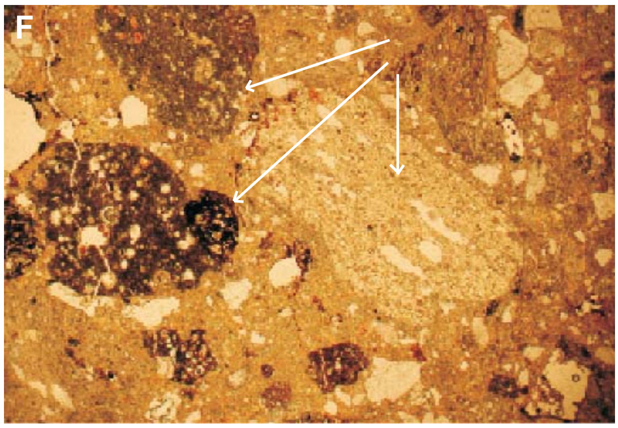
Handspecimen of a lapilli tuff of the zánkai Vár-hegy diatreme. Note the limestone lapilli (white) derived from Mesozoic rock units. The shorter side of the photo is about 15 cm



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

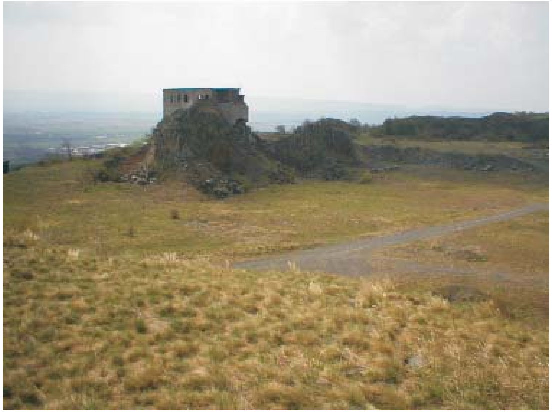


Photomicrograph of a tephriphonolitic glass hosted amphibole aggregate as a clast in a pyroclastic rock recovered from in situ debris from the Horog-hegy, near Hegyes-tű. Short side of the photo is ~2 mm



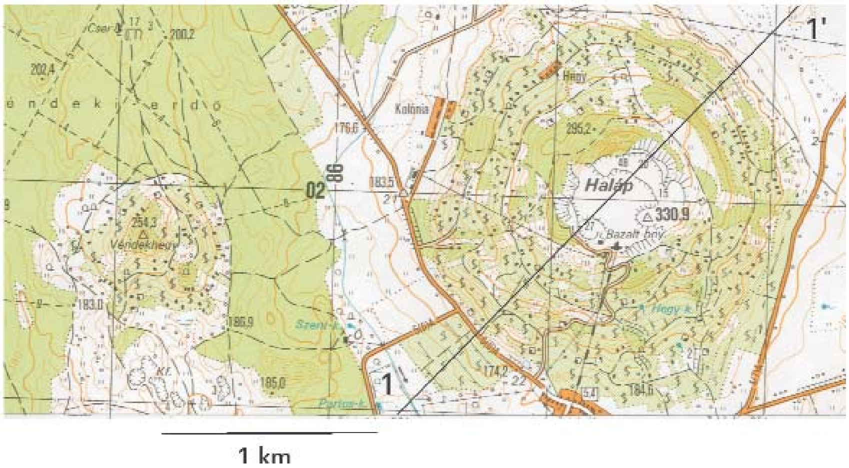
Photomicrograph of volcanic glass shards (arrows) from the Kis-Hegyes-tű. Shorter side of the picture is ~4 mm

A



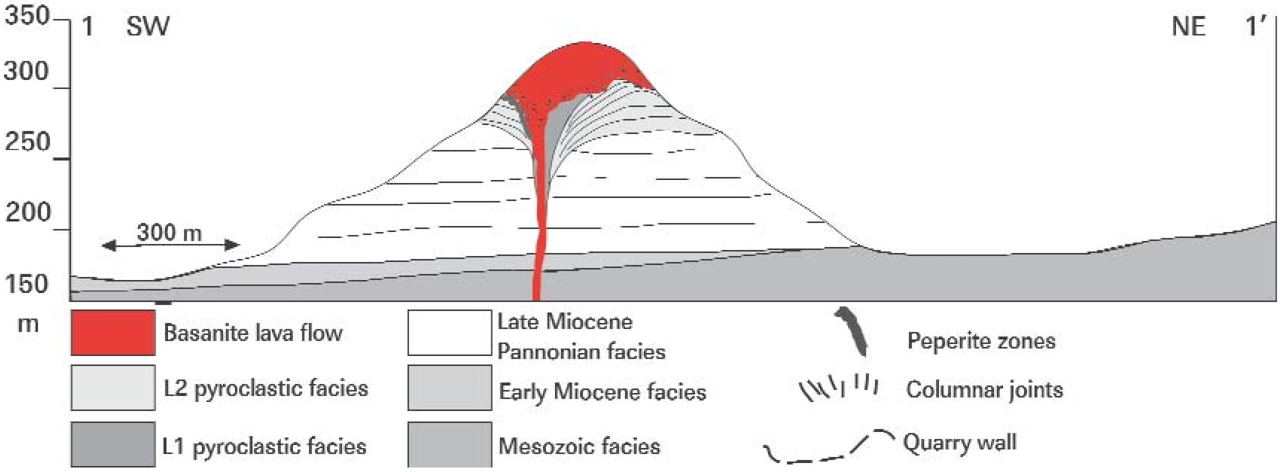
The "castle-like" architecture of Haláp [the log on Figure 3.14 represents a section just below the abandoned building to the left of the photo]

B

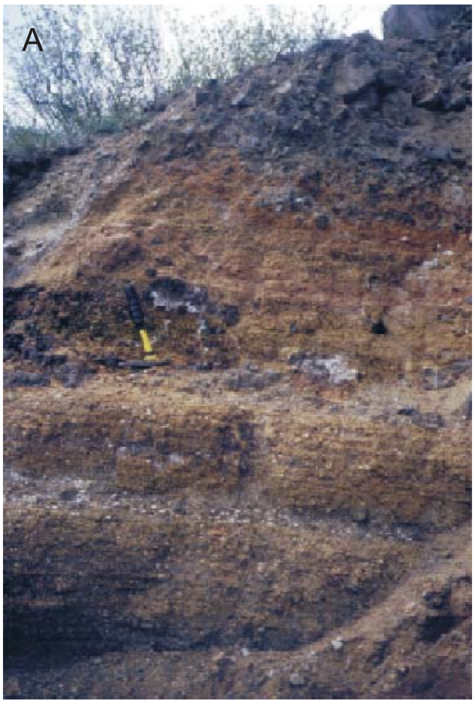


Location of the Haláp, and Véndek-hegy volcanic remnant [a detail from the 1 to 25,000 scale topographic maps of Hungary]. Active quarrying removed the former coherent basanitic lava, leaving behind collar of pyroclastic rocks. 1–1' line represents the line along the cross section is made on "C"

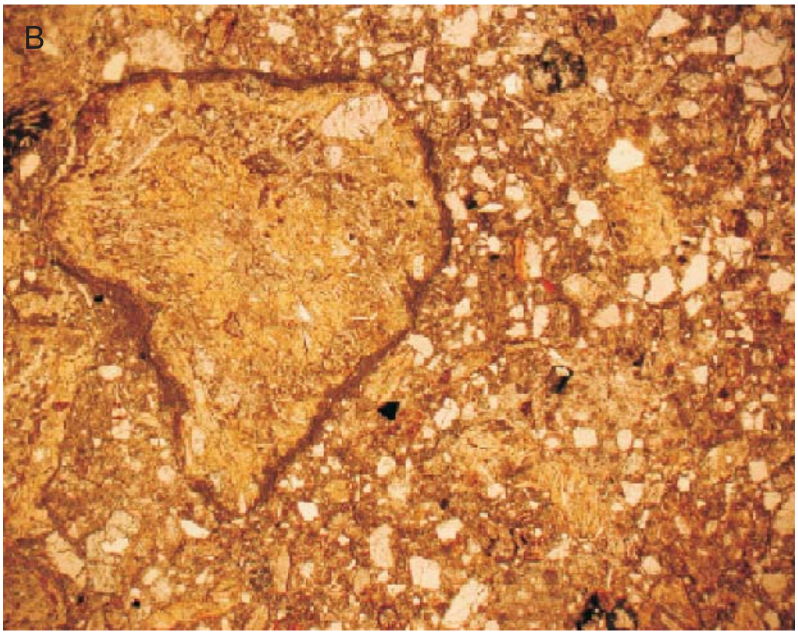
C



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



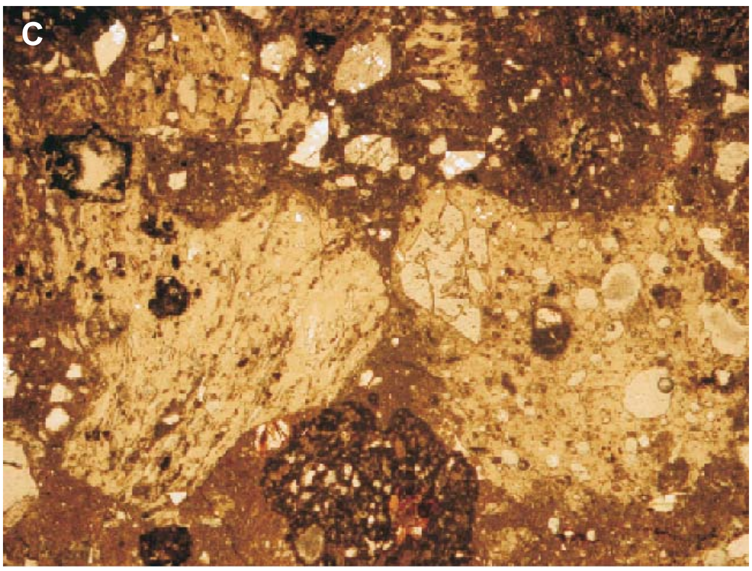
Photomicrograph of strongly palagonitised sideromelane glass shard from a lapilli tuff of the central part of the Haláp maar/tuff ring remnant

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

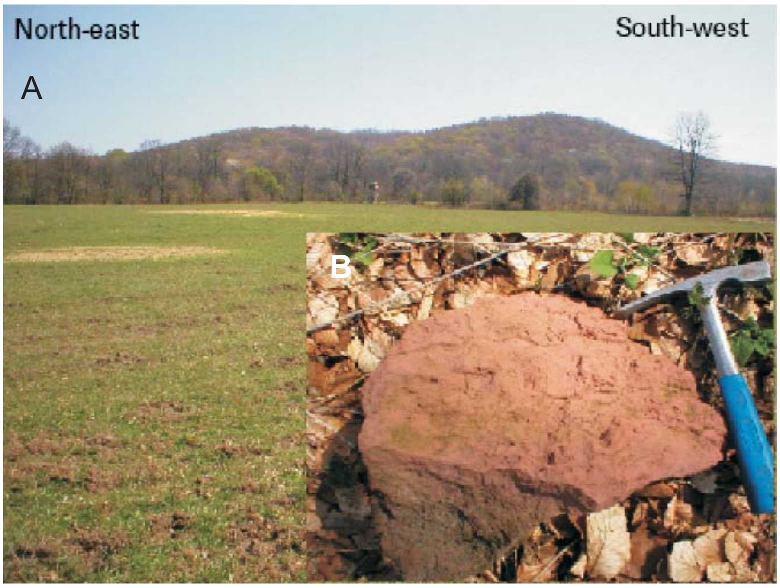
Mud filled vesicles of a scoriaceous lava spatter in a phreatomagmatic lapilli tuff of Haláp near the contact of pyroclastic succession and coherent lava

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

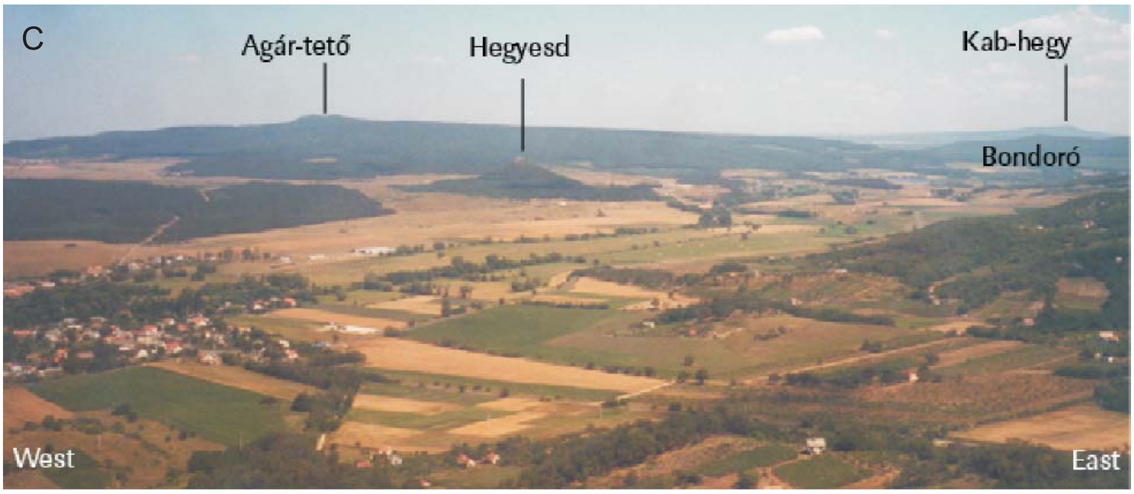


Photomicrograph of a volcanic glass shards from the Véndek-hegy pyroclastic succession. The short side of the photo is ~4 mm

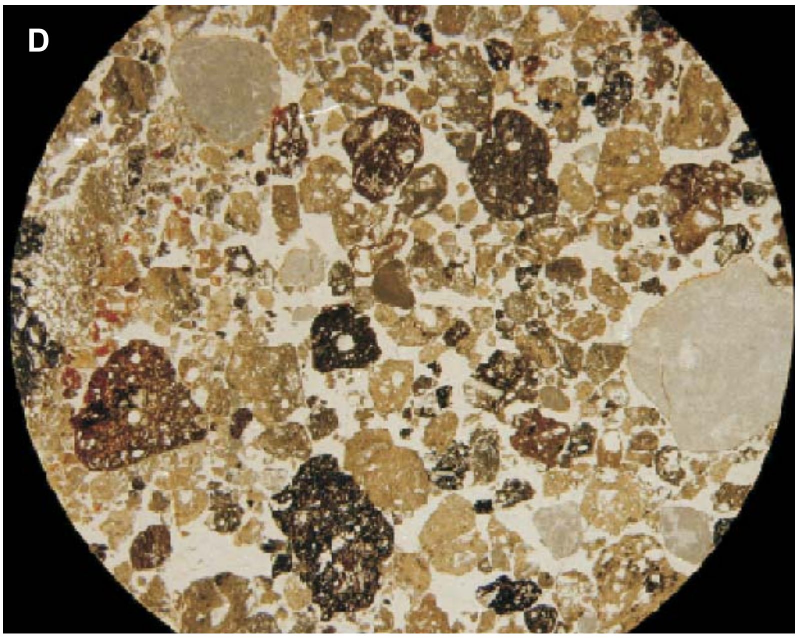


Erosional remnant of scoria cone at the top of the Agár-tető lava shield still retaining its original volcanic morphology, suggestive for young age

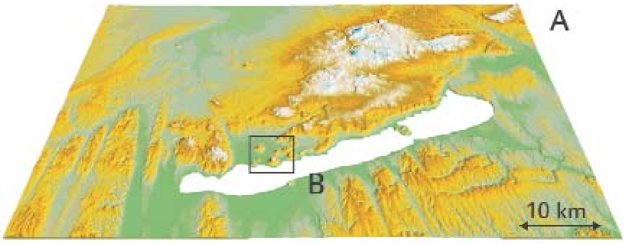
On the flank of the scoria cone remnant of Agár-tető a large number of fragments from red spindle bombs are present



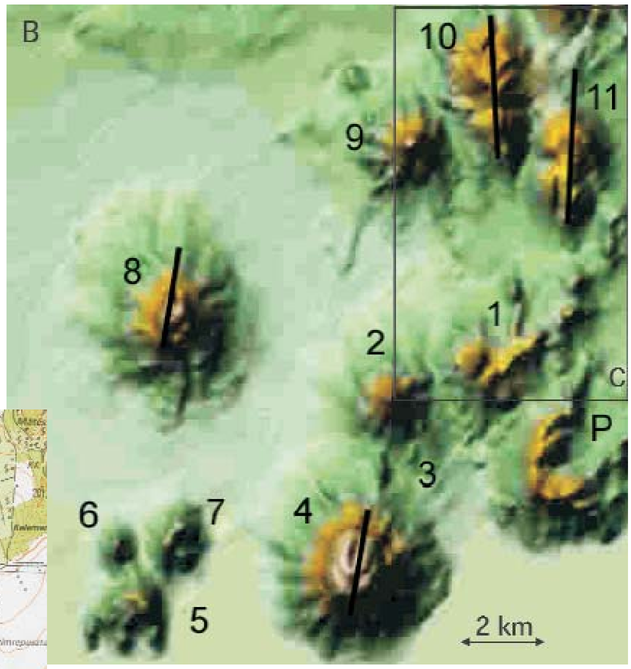
The scoria cone cap is clearly visible (from south to north) on the Agár-tető volcanic complex



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

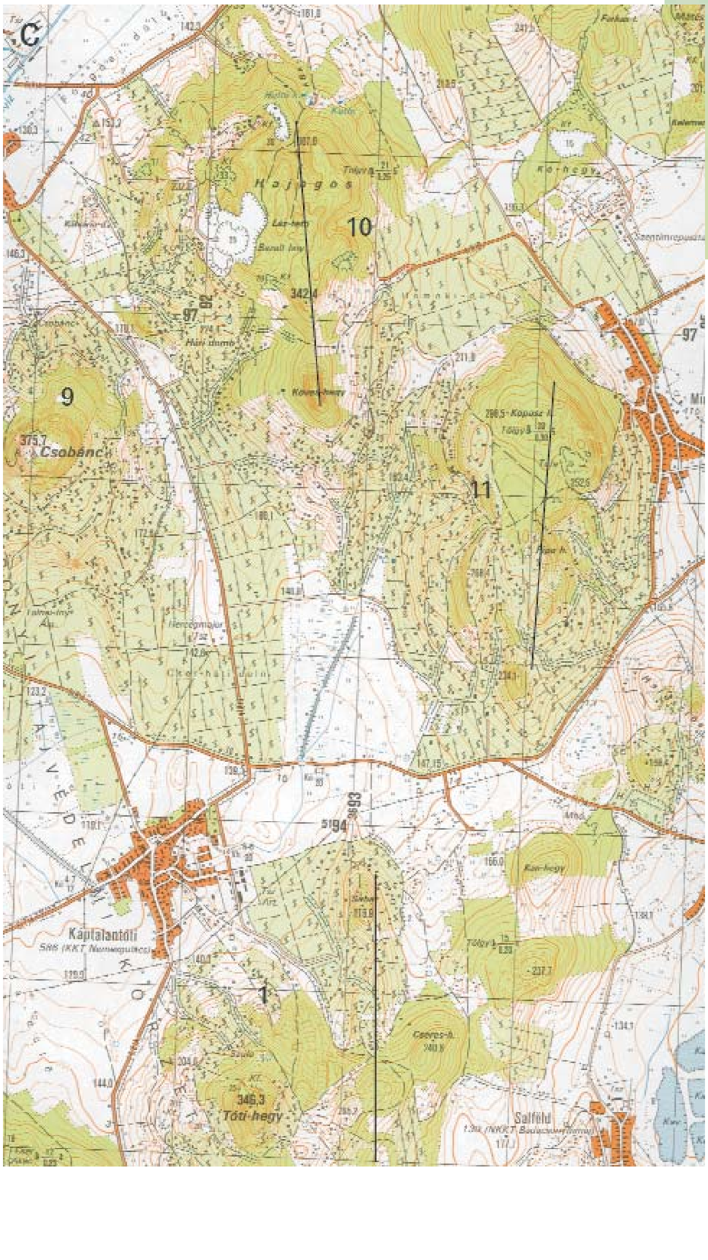


An overview DTM map shows the location of the Tapolca Basin.



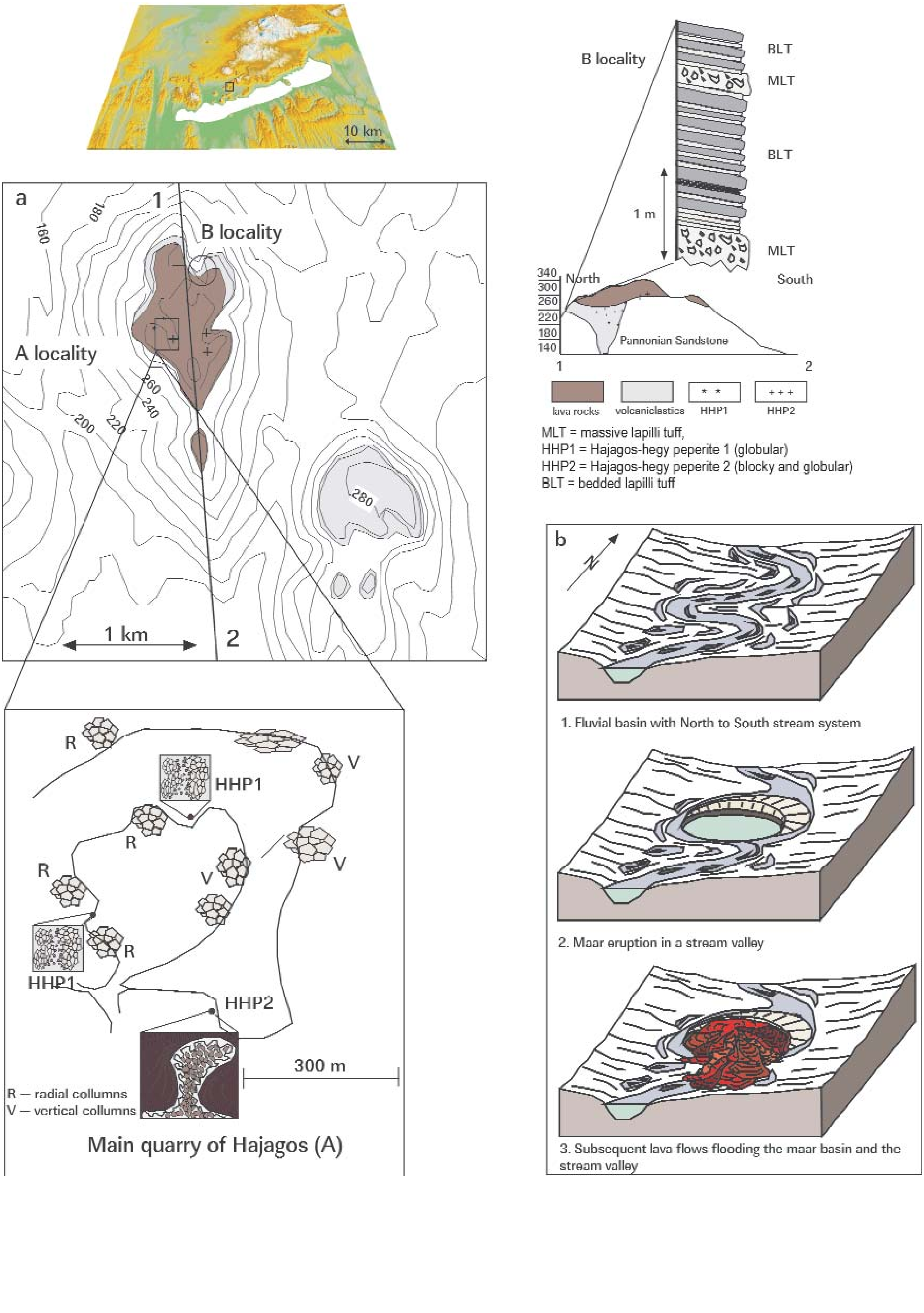
DTM of the eastern margin of the Tapolca Basin, showing the location of the presented volcanic erosion remnants. Lines over the vent remnants indicate the elongation direction of vent remnants. Rectangular field corespondes to topographical map on "C"

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



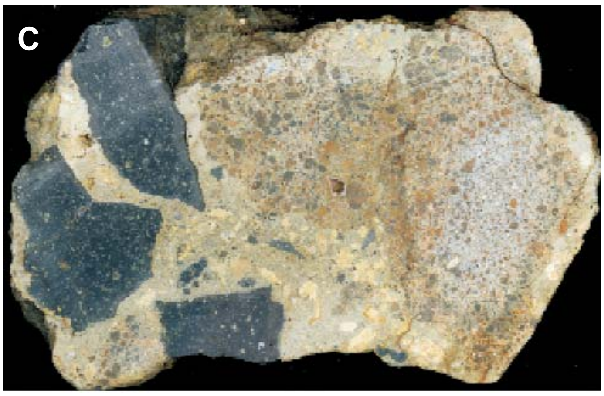
A detailed topographic map shows the fine morphology of the pyroclastic hills of the eastern Tapolca Basin. Numbers represent the same localities introduced on the "B" figure. The map source is a 1 to 25,000 scale topographic map series of Hungary. The rectangular grid on the map has 1 km spacing

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

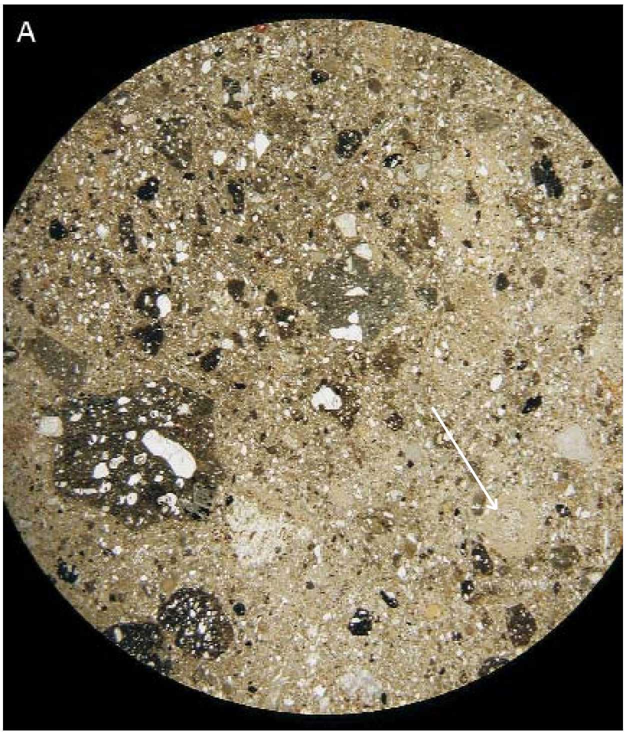
Fluidal peperite (globular) in lapilli tuff host at Hajagos. Note the horizontal fingerlike protrusions of basanitic lava that truncating the host pyroclastic deposit

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

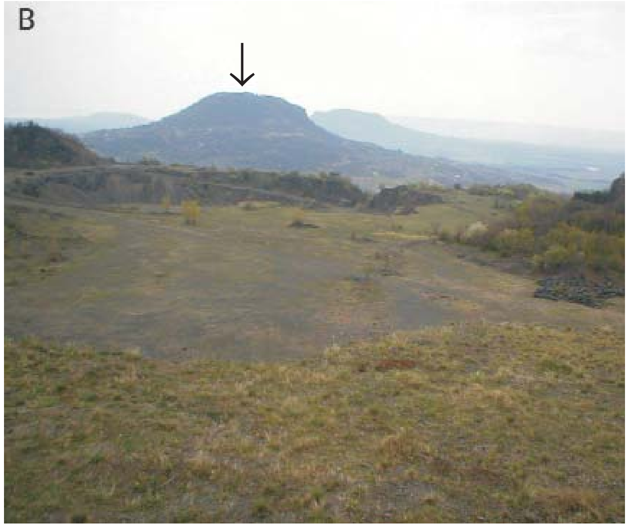
Peperitic zone along the foot of a lava flow at Hajagos. The lava flow movement is inferred to be from right (north) to left (south)

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

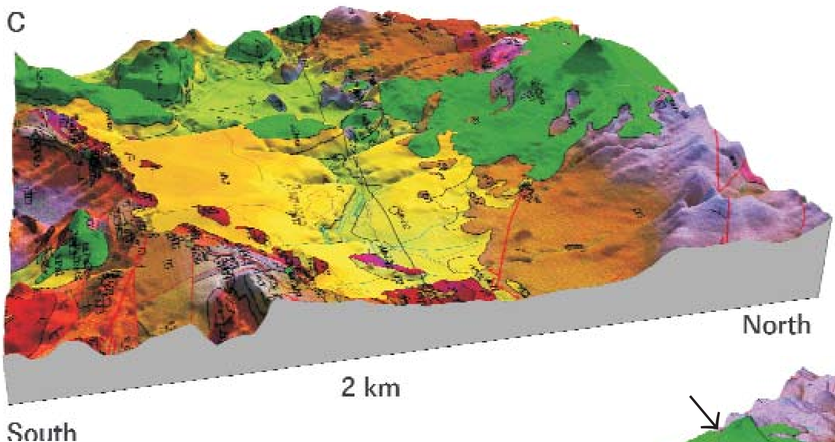
Vesicular lava flow foot texture is known from the lower flow unit of the lava flows from the southern margin of the Hajagos and from the Köves-hegy, just a few hundreds metres toward south of Hajagos



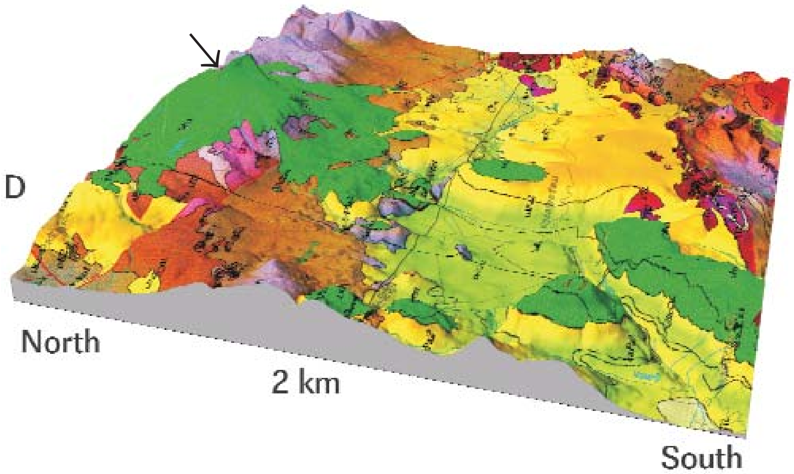
Photomicrograph of a phreatomagmatic tuff of the eastern sector of the tuff ring remnant of Bondoró. Note the palagonitized volcanic glass shards and accretionari lapilli (arrow)



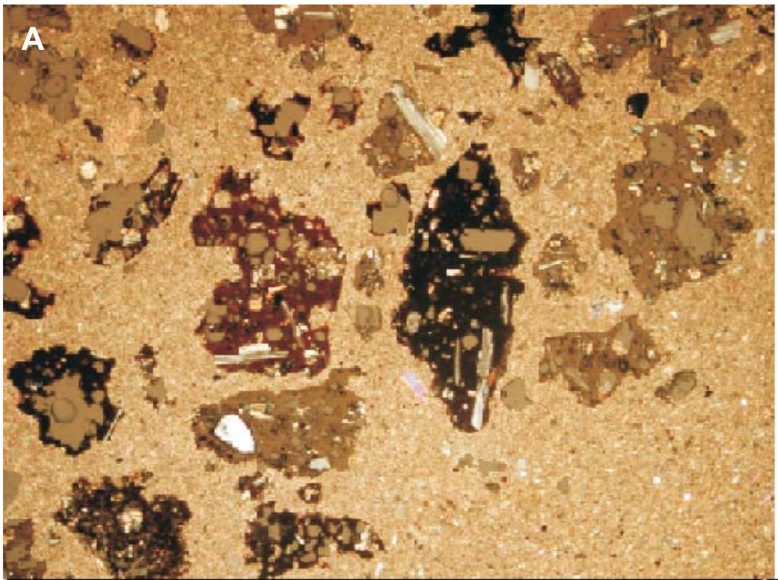
Panoramic view to the Csobánc (arrow) from Hajagos. The hill is capped by lava spatter that has been intruded by basanitic feeder dykes, today preserved as columnar jointed basanite. The lower section of the erosional remanant is formed by a phreatomagmatic lapilli tuff succession, which is only poorly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



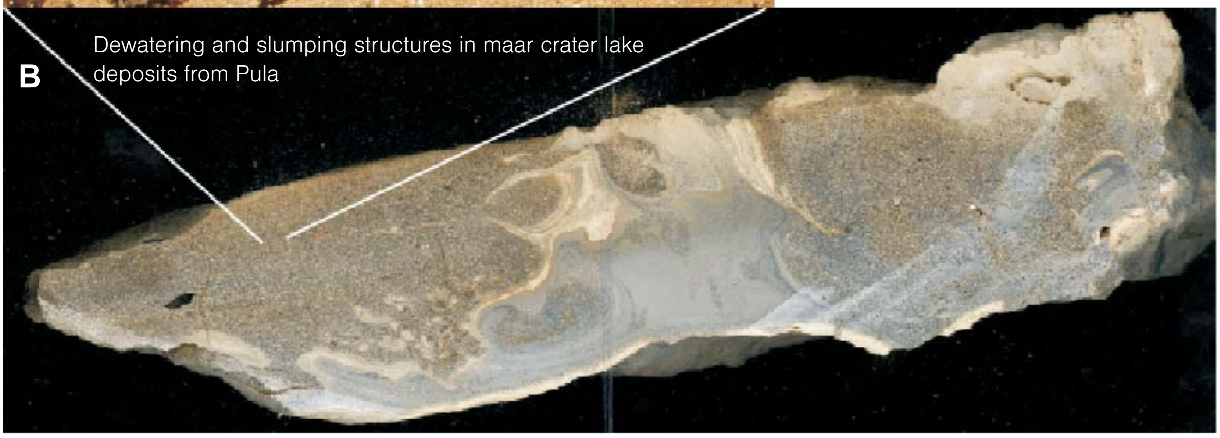
DTM model for the Pula region looking toward west



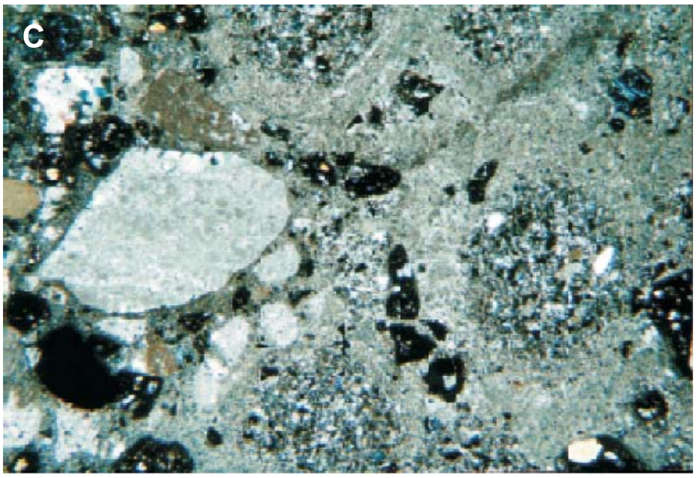
DTM model of the Pula region fitted with the 1:50,000 scale geological map of the BBHVF (Budai et al., 1999). Pula is a small depression in the centre of the 3D model between the background of the Kab-hegy shield volcano (arrow) and the foreground Tálodi-erdő lava field



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

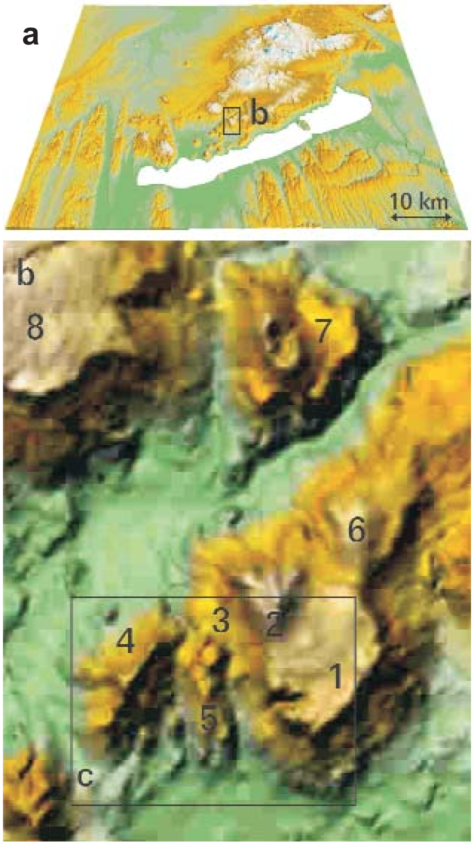


Dewatering and slumping structures in maar crater lake deposits from Pula



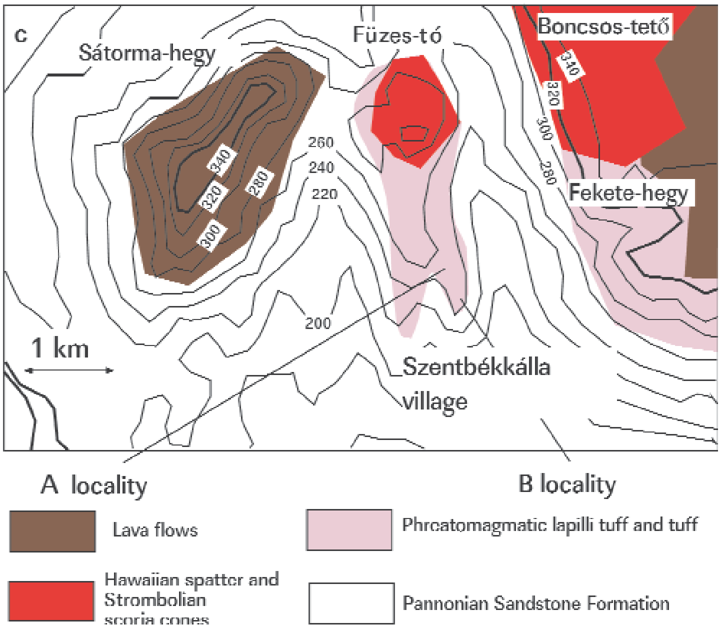
Photomicrograph of rim-type accretionary lapilli from a phreatomagmatic crater rim sequence of Pula

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").

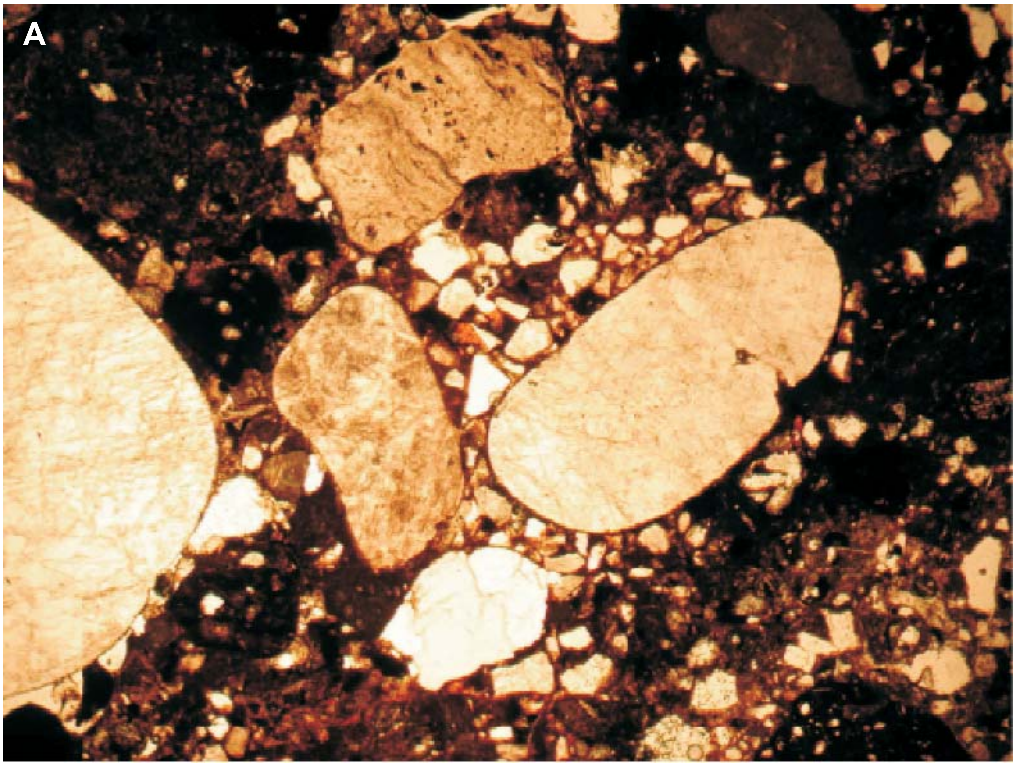
1 — Fekete-hegy, 2 — Boncsos-tető, 3 — Füzes-tó (Kopácsi-hegy), 4 — Sátorma-hegy, 5 — Szentbékkálla mafic pyroclastic flow, 6 — Kapolcs diatreme, 7 — Bondoró, 8 — Agártető lava field



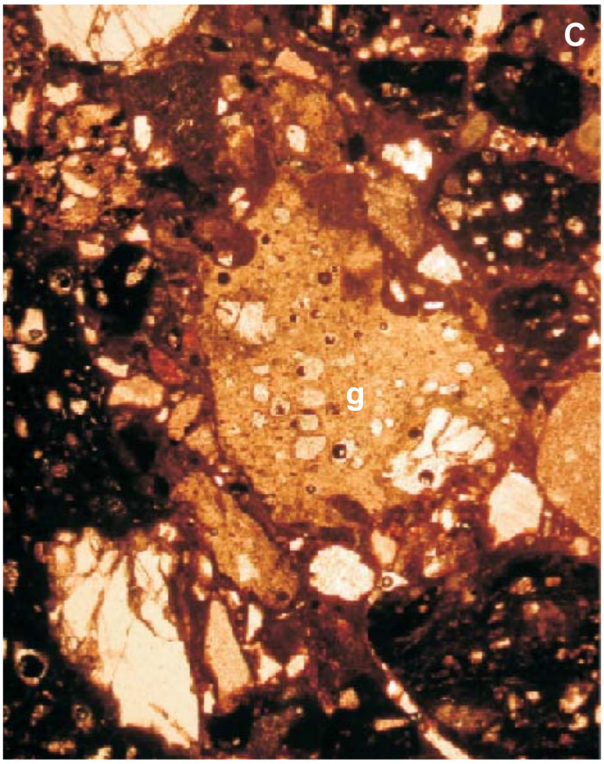
A simplified geological map shows the relationship between different volcanic facies nearby Szentbékkálla village. Localities A and B refer to two section that have been identified and described

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)

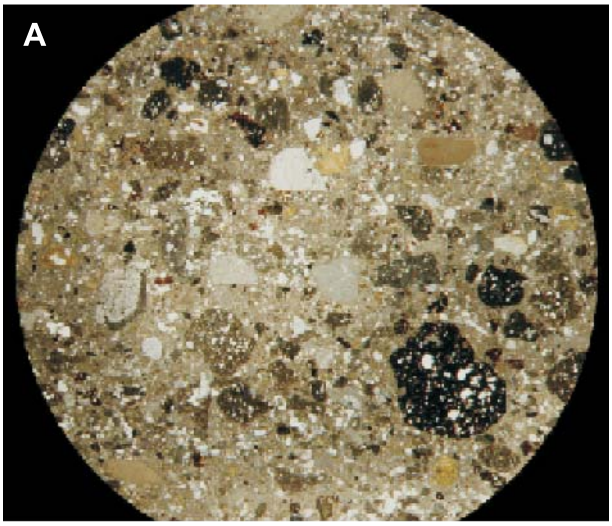


Photomicrograph of picked up pebble from the basal zone of the pyroclastic succession near Szentbékkálla, which is in contact to the underlying gravel beds indicating, that the gravel was still unconsolidated and easy to be picked up en route from horizontally moving pyroclastic density current

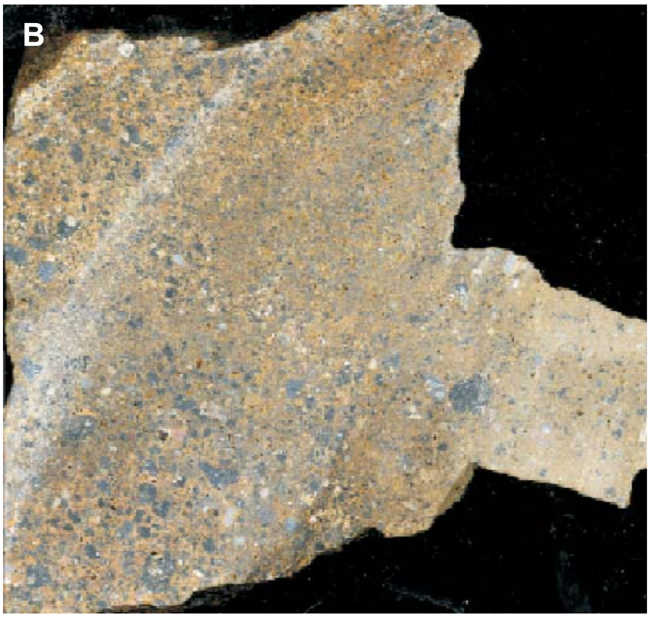


Photomicrograph of volcanic glass shards (g) in a fine siliclastic fragment-charged matrix of the massive lapilli tuff from the Szentbékkálla loacality. Short side of the photo is about 1 mm

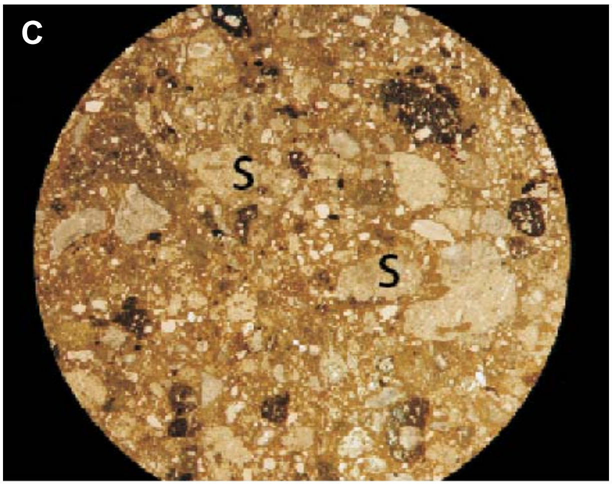
Gas-segregation pipes (arrows) in the basal zone of the massive lapilli tuff of Szentbékkálla filled with angular to subrounded predominantly accidental lithic fragments



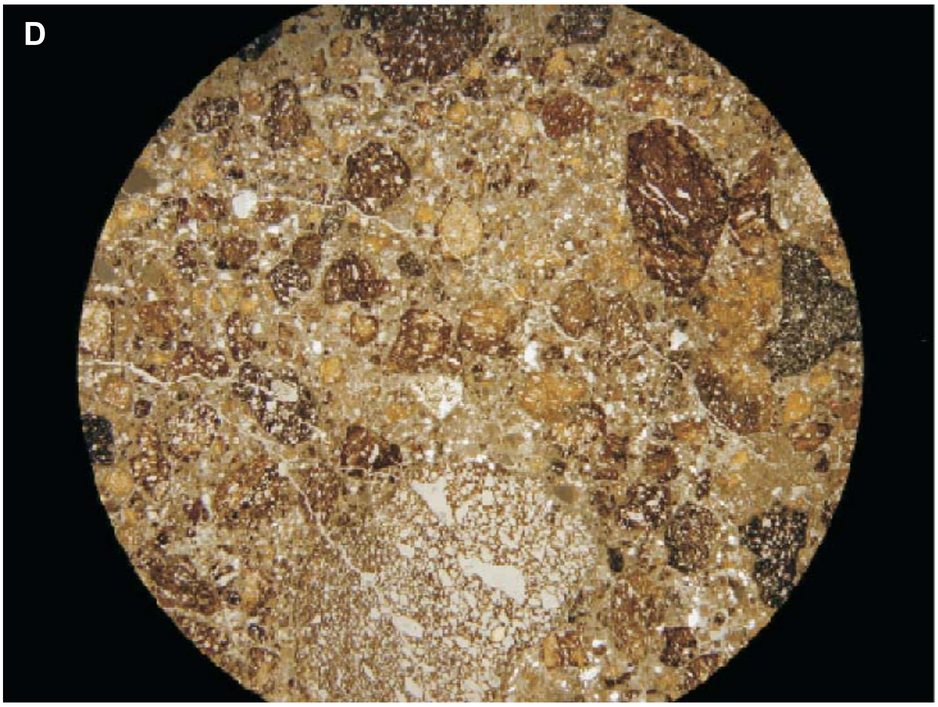
Photomicrograph of a phreatomagmatic lapilli tuff rich in quartzo-feldspatic sediment-derived mineral phases indicating the water saturated nature of the sediments that played an important role during the eruption at Fekete-hegy. The photo is ~2 cm across



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

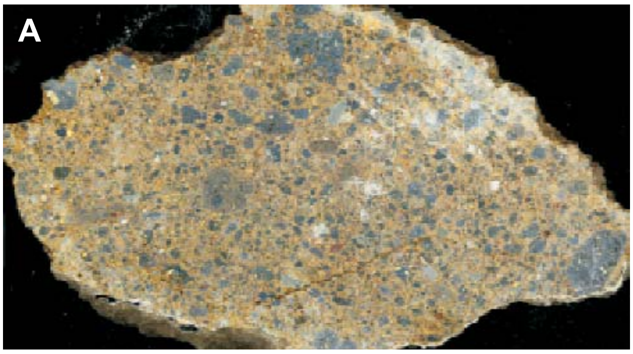


Photomicrograph of a lapilli tuff from Kereki-hegy, that is rich in fragments derived from siliciclastic sediment and blocky glass shards (s). The photomicrograph is ~1 cm across

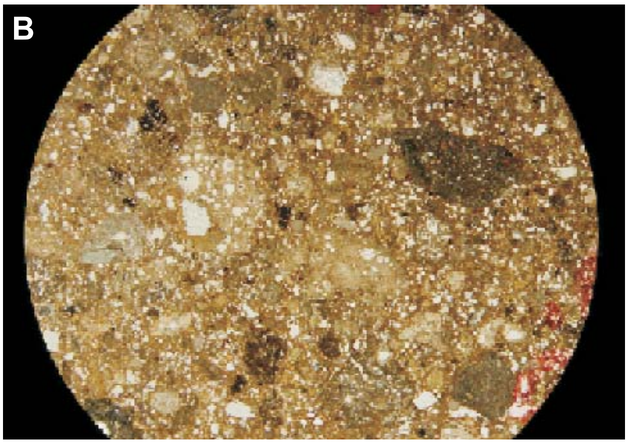


Photomicrograph of a juvenile lapilli with trachitic texture from the Harasztos-hegy pyroclastic succession. The photo is ~2 cm across

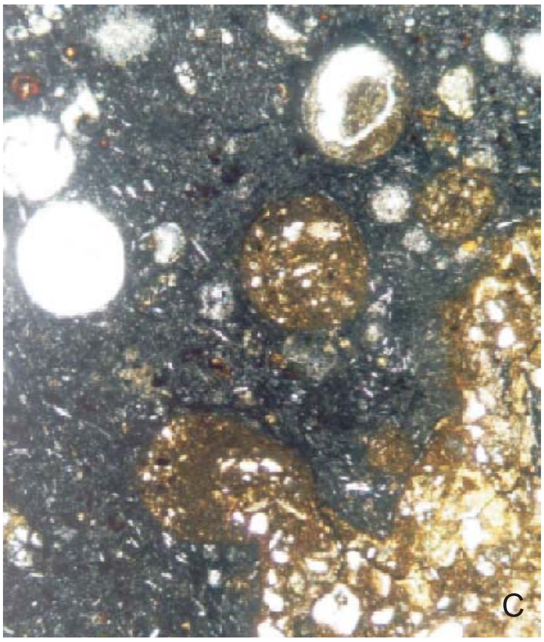
Plate 20 / Chapter 3 Second International Maar Conference — Hungary–Slovakia–Germany



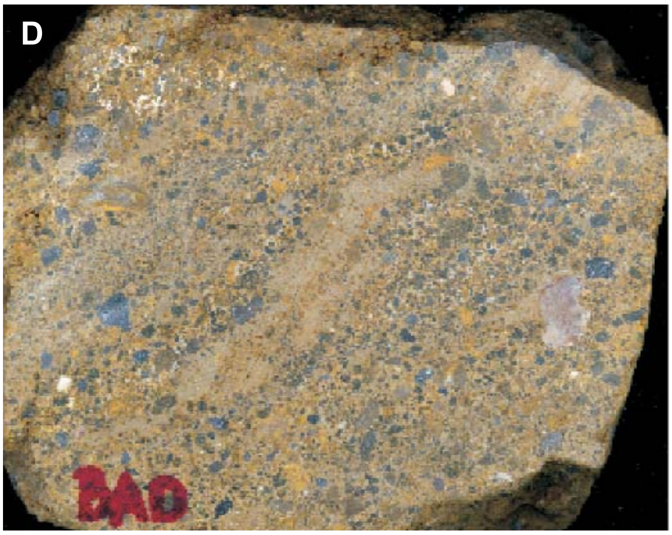
Handspecimen of a lapilli tuff from Hármas-hegy, characteristically yellow, to light brown in colour and rich in volcanic glass lapilli in a quartzofeldspathic matrix. The shorter side of the picture is about 10 cm



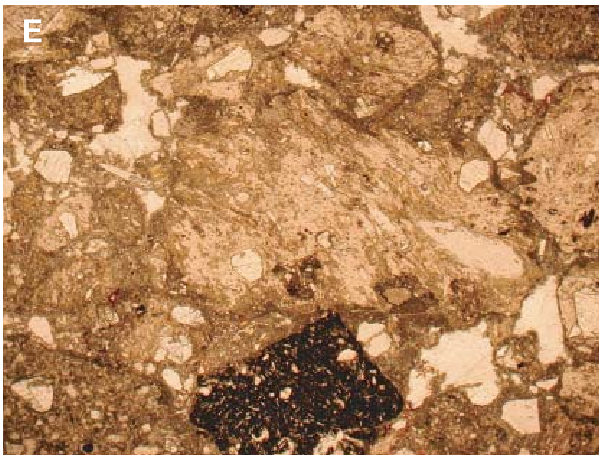
Photomicrograph [plan parallel polarized light] of a lapilli tuff recovered from the Hármas-hegy rich in volcanic glass shards with variable microvesicularity. The photo is ~ 2 cm across



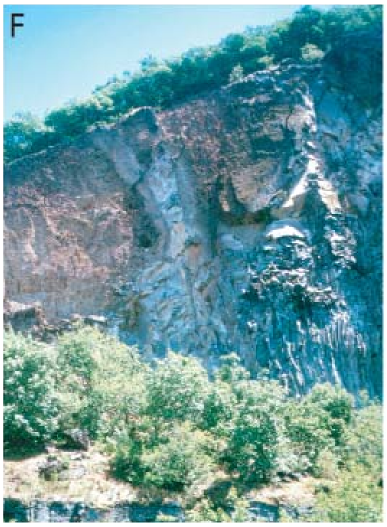
Photomicrograph (parallel polarized light) of a lapilli tuff from the Hármas-hegy near Badacsony that shows a tachylite glass shard (black vesicular clast in the left hand side), that entrapped siliciclastic and/or volcaniclastic mud/ash indicating a premixing and possible recycling of pyroclast through 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



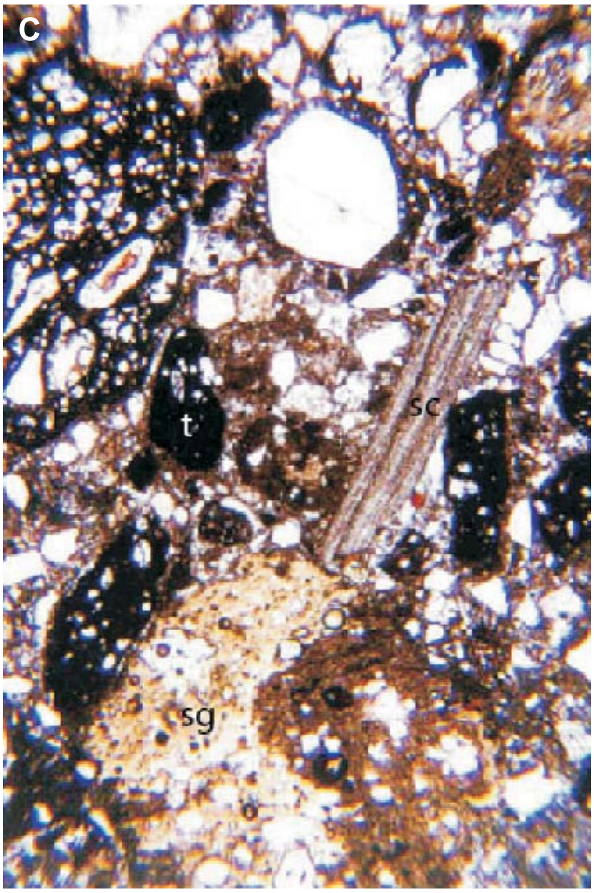
Photomicrograph of a lapilli tuff from Badacsony, containing elongated, blocky, moderately vesicular volcanic glass shards. Broken pyrogenic and xenocrysts are characteristic constituents of the pyroclastic rocks at Badacsony. The short side of the photo is 2 mm



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

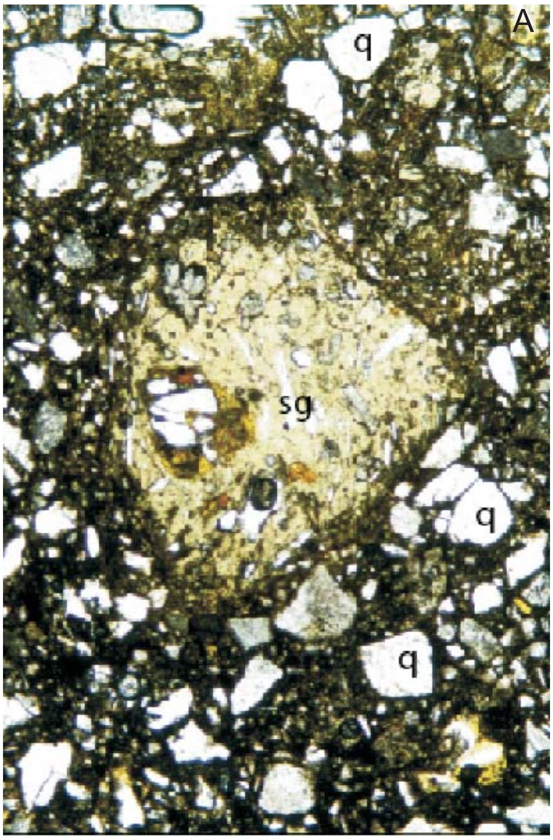
Steep bedded pyroclastic succession capping the Vár-hegy in the north-western margin of the Szigliget area, p = pyroclastic rock units, d = basanite dyke, n = Neogene subhorizontal siliciclastic rock units

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

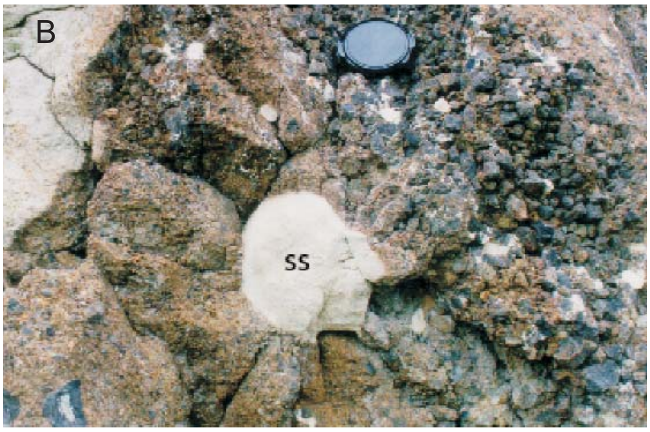


Photomicrograph of a lapilli tuff from the southern hill of Szigliget (Kamon-kő), that contains schist (elongated angular clast in the centre of the picture – sc) and blocky volcanic glass shard (in the left side of the picture – sg) and tachylite (t) cemented by calcite. The short side of the photo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



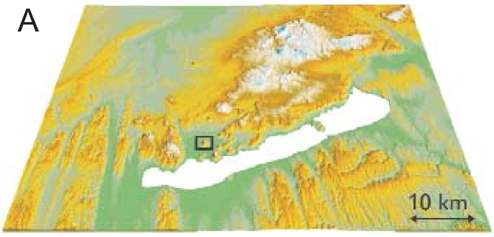
Large (dm-size), often rounded Neogene sedimentary clasts (creamy coloured clast in the centre of the photo — ss) of Unit 3 in the topmost part of the pyroclastic succession of the Vár-hegy

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

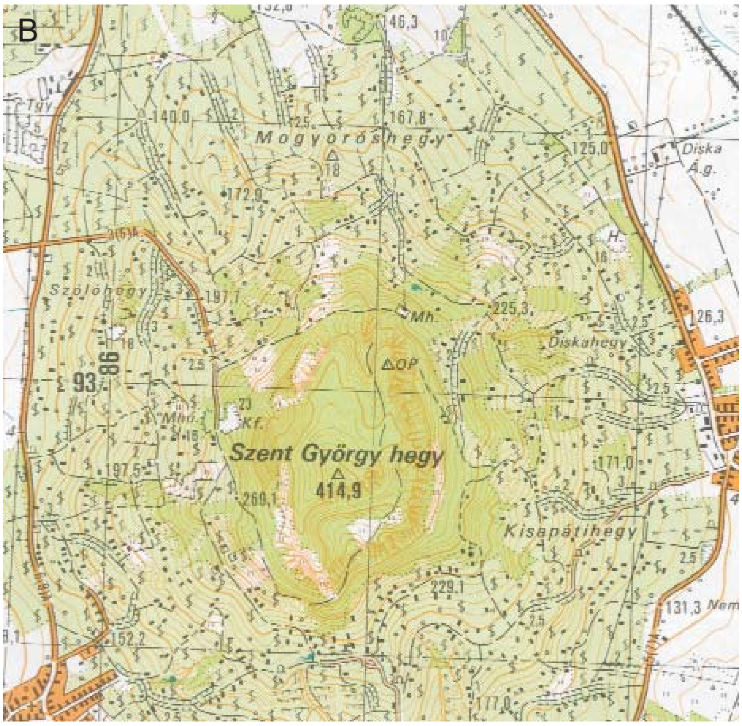
Abraded lapilli tuff fragment as a clast (outlined) in a lensoid steep bed from Vár-hegy, indicates some degree of reworking and remobilization of tephra upon formation of this succession

A



Location of the Szent György-hegy on DTM

B

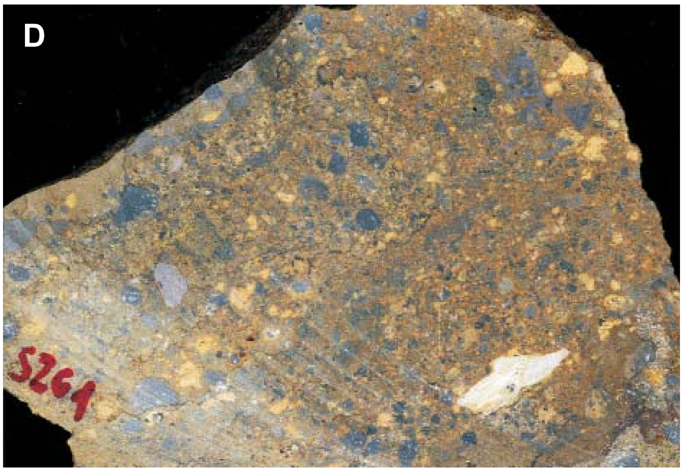


Detailed topographic map of the Szent György-hegy erosional remnant. Note the north to south elongated erosional remnant. Map data is from the 1 to 25,000 scale topographic map series of Hungary. Rectangular grid has 1 km spacing

C

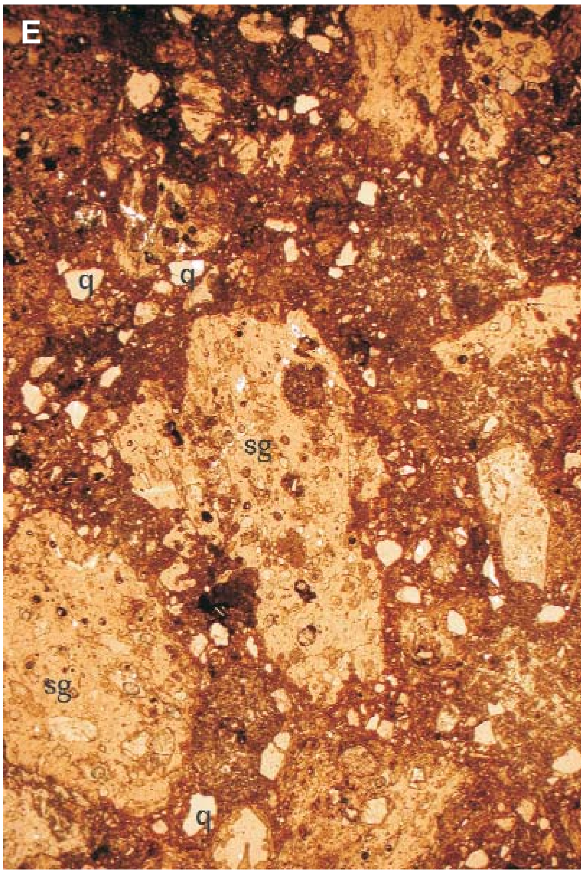
View to Szent György-hegy from south. Note the capping irregular shape structure of the hill above the grape yards. This zone is a spatter rich scoriaceous succession that was invaded by rosette-like columnar jointed basanite

D

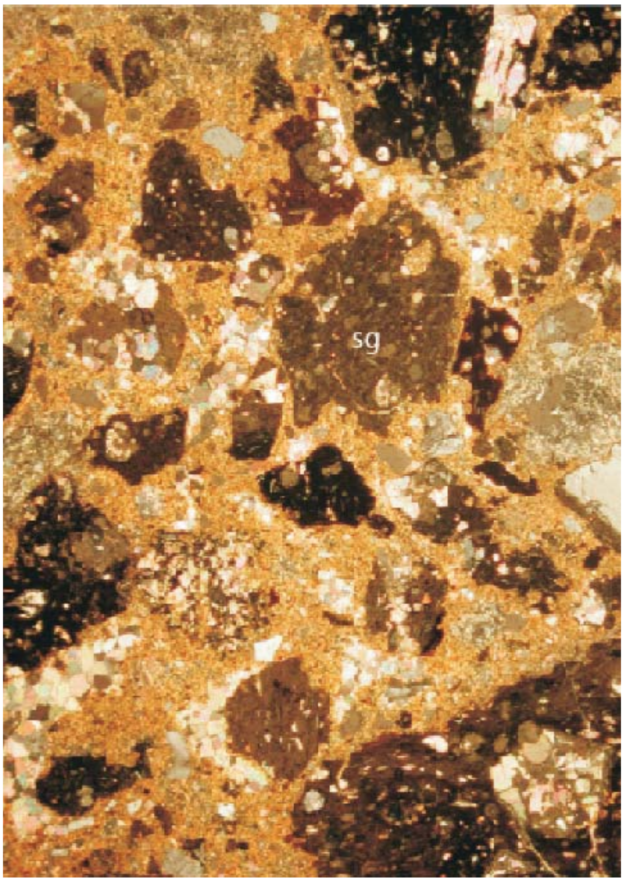


Handspecimen of a phreatomagmatic lapilli tuff from Szent György-hegy. The black lapilli are chilled pyroclasts. The short side of the photo is 10 cm

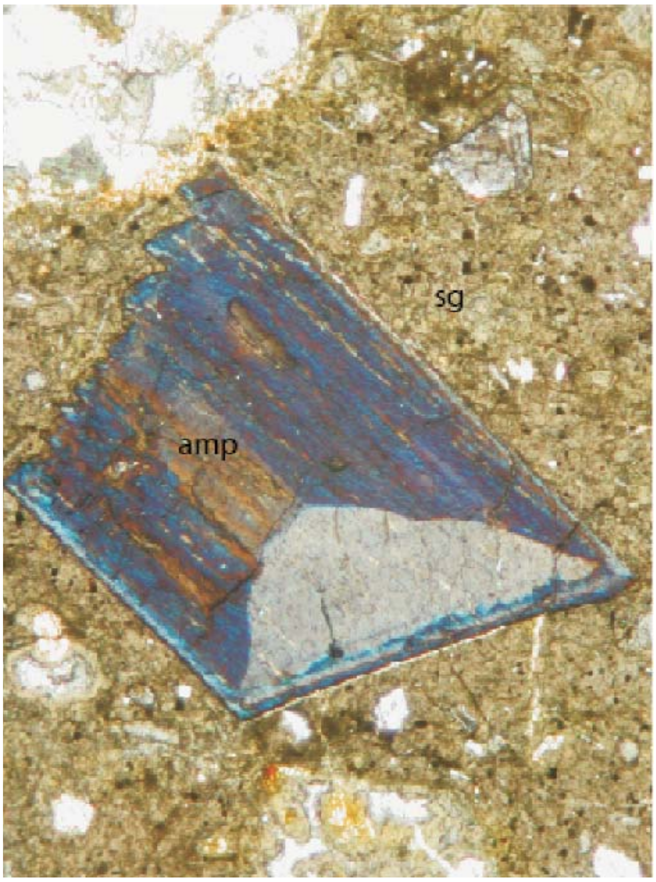
E



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



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)

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Chapter 4

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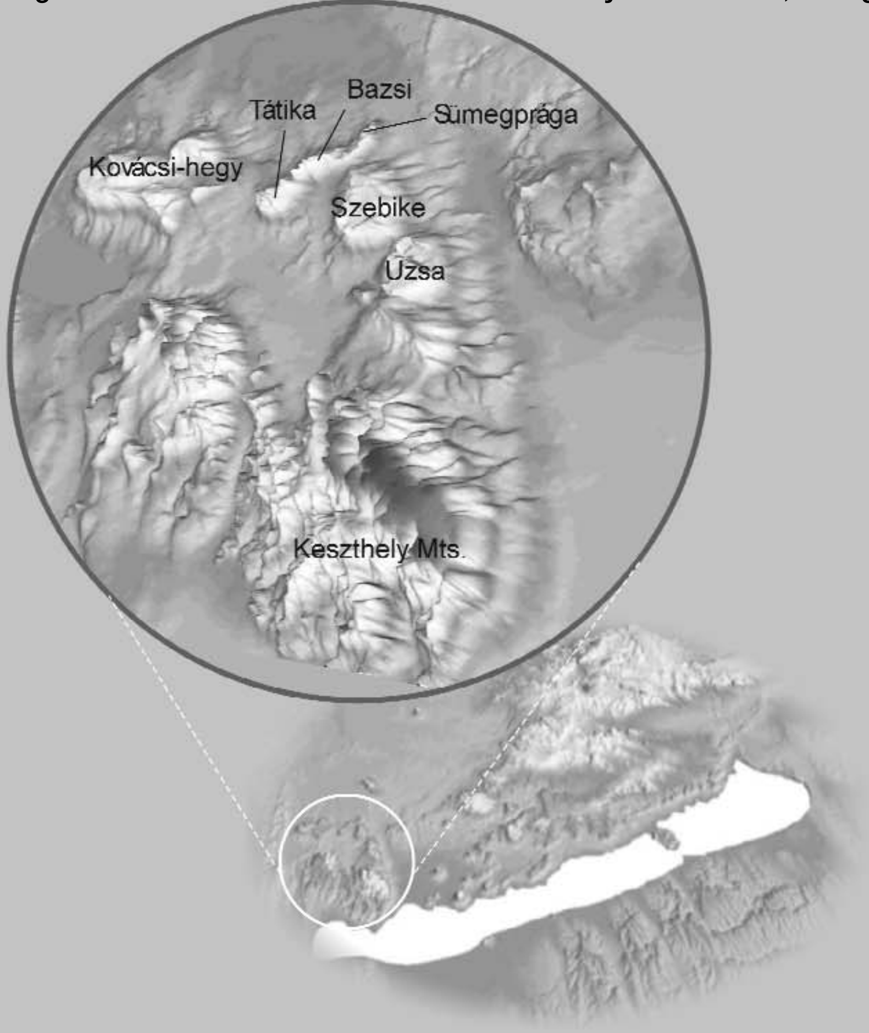
Shallow sub-surface intrusive processes associated with phreatomagmatic volcanism north of the Keszthely Mountains, Hungary



Contents

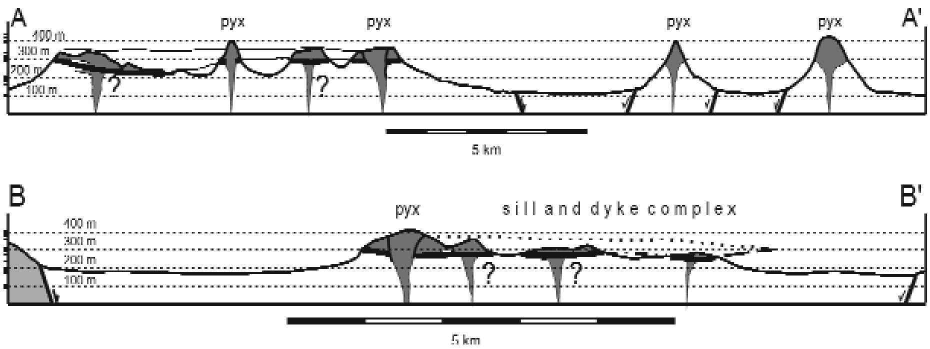
Abstract

Neogene alkaline basaltic rocks in the western Pannonian Basin are eroded remnants of former maars, tuff rings, tuff cones, scoria cones and lava fields. The erosion level of these volcanoes is deep enough locally to expose diatreme zones associated with the phreatomagmatic volcanoes. West of the Bakony – Balaton Highland Volcanic Field the erosion level is deeper yet, exposing shallow subsurface dyke and sill swarms related to former intra-plate volcanoes. The basanitic sills are irregular in shape and their lateral extent is highly variable. Individual sills reach a thickness of a few tens of metres and they commonly form dome-like structures with rosette-like radial columnar joint patterns. The largest sill system identified in this region is traceable over kilometres, and forms a characteristic ridge running north-east to south-west. Elevation differences in the position of the basanitic sills within an otherwise undisturbed "layer cake-like" siliciclastic succession indicate emplacement of the basanite magma at multiple levels over kilometre-scale distances. The margins of sills in the system are irregular at a dm-to-mm-scale. Undulating contacts of the sills together with gentle thermal alteration in the host sediment over cm-to-dm distances, indicate the soft, but not necessarily wet state of the host deposits at the time sills were intruded. Parts of the sill complex show a complicated relationship with the host sediment in form of peperitic zones and irregularly shaped, disrupted, peperite textures. This is interpreted to reflect inhomogenenities in water content and rheology of the siliciclastic deposits during intrusion. The current summit of this ridge preserves a small diatreme that seems to cut through an otherwise disk-like sill indicating to some degree of relationship between sill emplacement and phreatomagmatic explosive eruptions. A complex pyroclastic-to-lava succession is exposed in a large, still active quarry in the eastern part of an area inferred to represent a maar basin that was filled by post-maar lava flows, volcanic debris avalanche deposits, and by scoria cones.

Keywords: maar, diatreme, monogenetic, erosion, sill, dyke, dome, basalt, peperite

Introduction

Pliocene volcanic rocks crop out north of the Keszthely Mts and form an elevated ridge reaching more or less the same elevation (>300 m) as the southern Triassic limestone and dolomite blocks (Plate 4.1 and Figure 4.1). The volcanic rocks are somehow separated from the main volcanic zone of the Balaton Highland and form a distinctive cluster (Plate 4.1). The volcanic rocks in this area form a north-west to south-east trending zone, but the ridge alignments are different from this direction. The erosional remnants in the central part of the area form a very characteristic north-east to south-west ridge like feature (Plate 4.1). Volcanic rocks form mesa-like hills west (Kovácsi-hegy) and south-east (Szebike) of this ridge (Plate 4.1). The immediate pre-volcanic rocks are Neogene siliciclastic siltstones, sandstones, silt and sand. The southern margin of the area is fault bounded against Triassic limestone with a dolomite ridge that reaches an elevation of over 400 metres. Lava capped mesas (e.g. Tátika – 413 m, Uzsa lava plateau – 340 m) are more or less at the same elevation. The volcanic rocks are situated in a graben-like zone bounded by faults, a north-westward continuation of the Tapolca Basin (Plate 4.1 and Figure 4.1).



Figures 4.1. Simplified cross sections through the volcanic hills north of the Keszthely Mts. Thick line shows the possible lower margin of the coherent lava inferred to be a sill complex
Dashed line projects the possible upper contact of the sill complexes (pyx = pyroclastic rocks). Cross section lines are shown on Plate 4.1

Pyroclastic rocks crop out in large volumes only at the Uzsa locality (Plate 4.1). Small outcrops of pyroclastic rocks have been identified high up on Tátika (at about 380 m). The other localities show only coherent basanite. The basanite ridge of the Tátika–Sümegprága (Sarvaly) is a chain of irregularly shaped, commonly rosette-like columnar jointed basanites that are in intrusive contact with the host Neogene siliciclastic rock units. In contrast, the Kovácsi-hegy is a tabular basanite flow with multiple flow units forming vertically oriented columnar joints. A basanite ridge formed by dissected bud-like basanite intrusions forms a north-west to south-east ridge just south of Uzsa (Kő-orra).

The ages of the lava flows are among the youngest of the Pliocene intraplate volcanic rocks in the western Pannonian Basin and range between 3.4 to 2.7 My (BALOGH and PÉCSKAY 2001, BALOGH et al. 1986). This age range is similar to that of volcanic rocks of the Tapolca Basin (BALOGH et al. 1986, BORSY et al. 1986). This similarity also highlights the genetic relationship between these volcanic rocks and those of the Tapolca Basin.

Uzsa

Uzsa is the most voluminous Pliocene volcanic remnant north of the Keszthely Mts. It is diverse in type of preserved pyroclastic and effusive/intrusive rocks and their origin not yet fully understood. One of the largest still-active basalt quarries is located here. The intensive quarrying, which began in the fifties, has removed a significant part of the coherent basaltic lava, and gives access to the inner architecture of the lava capped mesa (Plate 4.2).

The lava flow units inferred to locate on horizontally bedded Neogene siliciclastic units. The contact zone, which is not exposed, is estimated to be at an elevation of about 300 m. The lava flows in the quarry wall expose multiple flow units. The basanitic rock is in general vertically jointed (Plate 4.3, A, B) with a few dm-wide columns which may reach heights up to 20 m. The basanite unit base to the pre-volcanic sedimentary rock units is not exposed anywhere in the quarry.

Between lava flow units large irregular zones of highly vesicular lava domains in a whitish, sandy, silty matrix are common (Plate 4.3, C). A few dm-sized pillow-like, vesicular lava bodies in the foot wall form a well-packed structure with preferred orientation of the individual pillow lobes. The transition to the coherent lava body is continuous, and a drop-like texture and gradual increase of vesicularity of the lava flow unit toward the pillowed zone can be identified (Plate 4.3, C). Such pillowed lava foot zones are present in different levels of the quarry and indicate that lava effusion may have been discontinuous. The silt and sand forming part of the matrix of these pillowed zones suggests that following a lava effusion stage, a short period of sedimentation took place on the surface of the older lava flows. The lava flows immediately above the pillowed lava foot zone are platy jointed (Plate 4.3, C). This joint pattern changes gradually to more-vertical jointing patterns up-section. In several parts of the quarry an onion-like joint pattern is prominent near the known and/or inferred margin of the lava flow (Plate 4.4, A, B), which suggests that the lava may be close to the margin towards the sediment.

Pyroclastic successions

The largest volume of pyroclastic rocks is preserved at Uzsa just north of the Keszthely Mts. The pyroclastic rocks are grouped in two distinct units:

1. In the central and western part of the quarry massive, weakly bedded, matrix-supported, grey to brown, accidental lithic-rich lapilli tuff and tuff breccia (Plate 4.4, C) crop out, and form a few tens of metres thick succession truncated by coherent basanite lava flows.

2. Red, weakly bedded, scoriaceous lapilli tuff and tuff breccia (Plate 4.5) crop out on the top of the major succession of coherent lava flows in the southern side of the quarry.

The first unit is rich in sideromelane glass shards that are moderately to weakly vesicular and blocky in shape (Plate 4.6), which suggests phreatomagmatic fragmentation (BÜTTNER et al. 1999, 2002, DELLINO and LIOTINO 2002). The glass shards are tephritic in composition and often show well-developed gel palagonitic rims (Plate 4.6). The pyroclastic rocks are characteristically rich in large, angular, non-vesicular blocky volcanic lithic fragments up to a metre in size (Plate 4.6, C). These blocks are porphyritic in texture and commonly fractured.

The weakly bedded, ill-sorted lapilli tuff unit dips approximately 15 degrees more steeply outward than the pyroclastic rocks of the same unit in the centre of the Uzsa volcanic complex, which may indicate the presence of a tuff ring (or pyroclastic mound) that was part of a former phreatomagmatic volcano. However, an alternative possibility is that basalt intrusion resulted in tilting of originally more shallow-dipping beds. Moderate amounts of non-volcanic accidental lithic fragments have been identified as predominantly abraded fragments of siltstone and sandstone. These clasts are characteristically derived from the immediate pre-volcanic Neogene units. The basal massive part of the pyroclastic rocks of unit 1, however, contains white, smooth surfaced, carbonate-rich silt fragments interpreted to have been derived from lacustrine units which are however unknown in the local strata (Plate 4.7, C).

In the upper level of the Uzsa quarry, reddish, moderately bedded, scoriaeous lapilli tuff and tuff breccia units (unit 2) form at least a few tens of metres thick succession capping the hills. The pyroclastic rocks of this unit contain a large number of fluidally shaped, dense to highly vesicular lava fragments (Plate 4.8, A, B). The size and the packing of vesicular lava clasts change gradually from south to north. Ambiguously, the succession also bears textural characteristics of a lava delta foot zone (SCHMINCKE et al. 1997, SKILLING 2002). There is a gradual decrease in dip from 20–25 degrees in the pyroclastic beds toward a shallow-dipping platy jointed, undulating coherent basanite flow.

In the upper section of the quarry, the contact zone between coherent lava and phreatomagmatic lapilli tuff units (unit 1) is exposed (Plate 4.7, A). The contact zone is sharp but irregular (Plate 4.7). It is obvious that the lava flow, now quarried away to leave behind a castle-like structure, post-dates the deposition of the pyroclastic units, and it is inferred to have rather had a shallow intrusive contact (e.g. lava lake emplacement) with the pyroclastic succession similarly to that at Ság-hegy (MARTIN and NÉMETH 2002, 2004).

Interpretation

The predominantly chaotic textural characteristics and the thickness of the tuff breccia and lapilli tuff beds of unit 1 suggest that the fragments of this unit were transported and deposited from volcanic debris flows associated with phreatomagmatic volcanoes similar to those of the Iblean Mts, Sicily (SCHMINCKE et al. 1997). However, ongoing research highlights that not all "massive" beds associated with phreatomagmatic volcanoes result from secondary remobilization of tephra (e.g.

Ferrar, Antarctica – HANSON and ELLIOT 1996) but that some may represent primary products of a phreatomagmatic eruption (e.g. Ferrar, Antarctica – WHITE and MCCLINTOCK 2001). To distinguish these sediments of different origin are not easy as it is known from the Ferrar province (ELLIOT and HANSON 2001), and it is a subject of ongoing research in the western Pannonian Basin, as well. The deposits may represent crater and/or vent filling units that accumulated in the crater and/or conduit of a phreatomagmatic volcano, e.g. a maar. The presence of oriented vesicles in some of the otherwise dense volcanic blocks of the pyroclastic rocks of unit 1 suggests that the vesicle elongation reflect stretching or shear of the magma either in the vent (e.g. dyke – BÜCHEL and LORENZ 1993) or afterwards, e.g. that they were parts of lava foot breccias (FISHER and SCHMINCKE 1984). The common presence of these dense volcanic lithic fragments in the massive to weakly-bedded lapilli tuff and tuff breccias, especially in the basal part of the unit 1 indicates that

1. coherent lava bodies must have existed prior to the phreatomagmatic eruptions, similarly to other fields such as the Crater Elegante (Pinacate, Sonora, Mexico – GUTMANN 1976), and/or

2. they have been derived rather from dykes than from previous lavas (HOUGHTON and SMITH 1993, VESPERMANN and SCHMINCKE 2000).

Fluid-form vesicular lava fragments in unit 2 are inferred to be small lava bombs and lapilli that rolled down-slope on the flanks of a former volcanic edifice (HEAD and WILSON 1989, THORDARSON and SELF 1993, ELLIOT and HANSON 2001). The gradual change of packing from south to north suggests a rapid change in volcanic facies architecture reflecting a nearby source. The bedding and the clast distribution pattern within this unit suggest that it originated by piling up of volcanic material in conjunction with ongoing remobilization on the flank of a volcano (HOUGHTON and SCHMINCKE 1989, VESPERMANN and SCHMINCKE 2000), rather than by successive deposition from spatter-rich pyroclastic density currents (VALENTINE et al. 2000). The gradual transition in bedding dips could alternatively suggest that the deposit represents a lava delta (SKILLING 2002), though no pillow lava nor pillow-fragment breccias that would add support for this interpretation have been identified. The abundance of fine-grained, lithic-derived clasts in the matrix of the succession is also suggestive of a scoria cone origin.

In summary, the Uzsa quarry (Láz-hegy) provides an excellent view into the inside of a lava capped mesa. Such mesas are very common in Western Hungary, but the rocks forming them are often poorly exposed, with only coherent lava flows traceable at the surface. The great thickness of the massive to weakly-bedded pyroclastic rocks at the basal zone of the Uzsa volcanic succession are inferred to have phreatomagmatic origin and indicate that magma/water interaction played an important role in the formation of the Uzsa volcanic complex. The relatively low proportion of non-volcanic accidental lithic fragments in these pyroclastic rocks indicates that the magma/water interaction that triggered explosive fragmentation occurred near the surface and/or there was not very much water available to fuel magma/water interaction. However, the large proportion of volcanic lithic fragments may be interpreted as fragments from a thick pile of coherent flows that formed the immediate pre-volcanic succession, and through which the explosive eruptions quarried. The topmost scoriaceous pyroclastic succession could be interpreted as either part of a scoria cone or as a lava delta. The hill south of the southern limit of the Uzsa quarry is interpreted to be a remnant of a scoria cone. Thus both lava deltas and scoria cones (which may have been the source of the lava flows) may have coexisted in Uzsa.

Sümegprága

Sümegprága is the northernmost part of a ~5 km long ridge exposing volcanic rocks and trending from north-east to south-west (Plate 4.9, A–E), with the ridge at Sümegprága reaching an elevation of about 260 m. Coherent lava flows appear at elevations around 220 metres and above (Plate 4.9, A–E). At this site basanitic lava has been quarried in the past, exposing the three dimensional architecture of the coherent lava unit. The lava flows form a small, flat hill, that has been opened up during the quarrying (Plate 4.9, A–E). The basanite is in intrusive contact with the host Neogene siliciclastic units. There are tabular and rosette-like coherent basanite that are in sharp but irregular contact with the host sediment (Plate 4.10, A, B). The contact zones of the coherent basanite are thermally affected (Plate 4.10, C), with sand and silt slightly hydrothermally altered at the contact zone of up to a metre near the intrusive bodies (Plate 4.10, C). Small protrusions from the master sills commonly form irregularly shaped dm-to-metre thick dykes with chilled margins (Plate 4.10, D). Peperitic margins are rare, and only exposed in small, dm-scale zones where slightly baked sand/silt is in contact with coherent flows. The textural characteristics of the host sediment are the same as for other Neogene siliciclastic units forming the immediate pre-volcanic successions elsewhere in the region. Small outcrops around Sümegprága show intrusive contacts between coherent lava bodies and host sediments, indicating that the ridge is consist of a sill and dyke complex.

Bazsi to Tátika

In the continuation of the ridge from Sümegprága to the south-west a sill/dyke complex is exposed in another, smaller, quarry (Plate 4.11, A–C). The sills are very irregular in shape and have chilled margins (Plate 4.12, A). Along the margin of the coherent lava body the host sediment is thermally affected in a dm-to-metre wide zone. In places, irregular margins and globular peperites can be identified (Plate 4.12, B). In these zones, the lava showed fluidal behavior and blobs which are mixed with the host sediment. Because there is no obvious evidence of high temperature alteration of the silt/sand along the coherent magmatic bodies (e.g. hornfels) we can only state that at least some hydrothermal effect on the host sediment (Plate 4.12, B) took place along the intrusive bodies.

In the upper quarry small finger-like lava intrusions into the host sediment can be observed, and terminate into an irregularly shaped zone of volcaniclastic rock (Plate 4.12, C). The clasts in this unit are angular, non-to-moderately vesicular, black and basanitic in composition, and are hosted by a fine, homogenised silt/sand (Plate 4.12, C). The basanitic clasts are generally finely crystalline or tachylitic in texture, but have cm-thick palagonite rims (Plate 4.12, C).

The 3D relationships of the coherent basanite with the host rock indicate that at Bazsi a sill and dyke complex is exposed. The location of this quarry is close to the Sümegprága quarry, and the elongation of the basanite body along the length of the ridge indicate that Bazsi is part of the same major sill and dyke system that runs from north-east to south-west. Similar dyke and sill complexes in poorly exposed settings have been mapped out to the south-west. Other major ridges, south-west of Bazsi, each represent bud-like basanite bodies that have rosette-like columnar jointing.

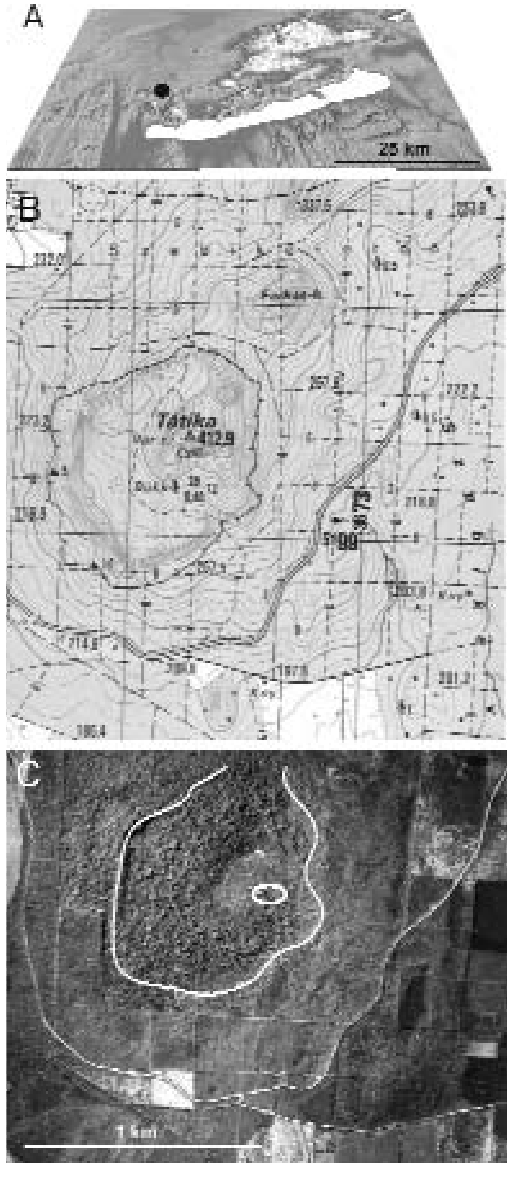
The most south-western hill of the ridge terminates in a circular shaped plateau-like lava region forming the plateau of the hill Tátika (Figure 4.2). The quarry of Bazsi exposes basanitic sills up to the level of 300 metres, and is covered by the same sand as it is exposed in the quarry itself. The plateau at Tátika reaches 350 metres (Figure 4.2). This plateau is cut through by few basanite buds that are characterized by rosette-like columnar jointing. In the basal zone pyroclastic rocks are exposed.

Pyroclastic rocks, collected from poor outcrops away from the quarry, are rich in fine sand, silt, quartz grains, and mud chunks (mm-to-cm size), derived from rock types characteristic of the immediate pre-volcanic Neogene rock units (Plate 4.12, D). The pyroclastic rocks of Tátika contain sideromelane glass shards that are blocky in shape and moderately vesicular, typical products of phreatomagmatic fragmentation (Heiken and Wohletz 1986, 1991, Wohletz 1986). The glass shards are tephritic in composition, similar to glass shards from Neogene phreatomagmatic lapilli tuff and tuff units elsewhere in the western Pannonian Basin (Martin et al. 2003).

The general spatial relationship between coherent lava units of Bazsi and Tátika, and the Neogene siliciclastic succession indicate that the basanite is in intrusive contact with the sedimentary rocks, and is part of an elongated sill and dyke system in this region. The only location where pyroclastic rocks are exposed is Tátika, indicating that explosive eruption took place in that site. The textural characteristics of the recovered pyroclastic rocks indicate that the explosive eruptions of Tátika were driven by magma/water interaction. The pyroclastic succession seems to be cut by basanitic sills and dykes, which suggests sill/dyke intrusion after the phreatomagmatic eruption of Tátika. The erosion levels at these locations are deep. The minimum level of the syn-volcanic palaeosurface is at the top of the Tátika, assuming that

1. there has been no uplift nor tectonic dissection of the volcanic rock-capped ridges between Tátika and Sümegprága, and

2. Tátika pyroclastic rocks represent the top, rather than lower parts, of a diatreme. Even if the present hill top of Tátika really is the top of a diatreme, its position therefore should represent a level below the syn-volcanic palaeosurface using general considerations for the 3D architecture of a diatreme (Lorenz 1987, 2000a, b, White 1991a, b). It is hence a conservative estimate that the immediate pre-volcanic Neogene siliciclastic units were all present at the time of eruption in this region, and that the sill complex hence intruded at depths of at least 50-100 metres below the palaeosurface. It is very likely, however, that this depth is an underestimate, and that the Neogene sedimentary cover has suffered some 200–300 metres of erosion in this region, as has been calculated for the Bakony – Balaton Highland Volcanic Field (Németh and Martin 1999).

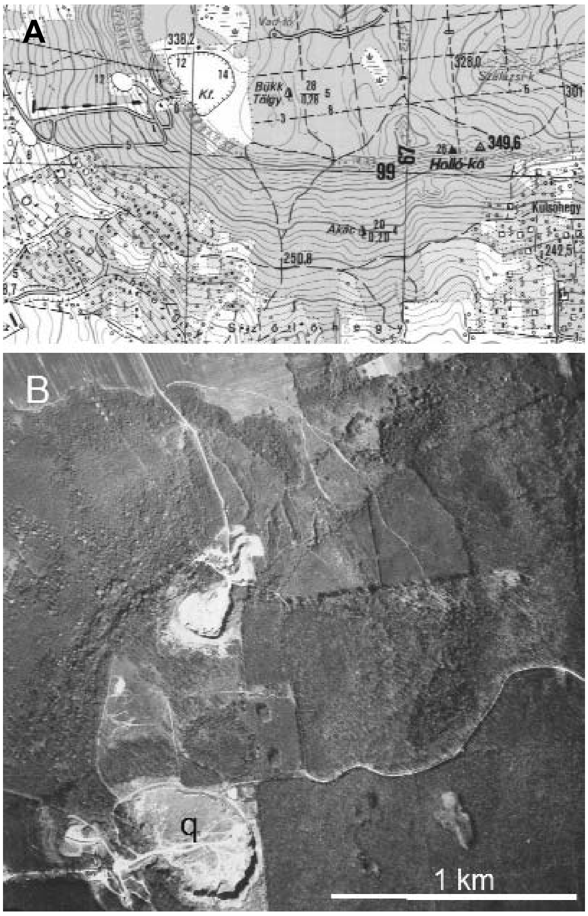


Figures 4.2. Location map of Tátika (A), which marks the termination of the volcanic rock ridge between Sümegprága and Tátika

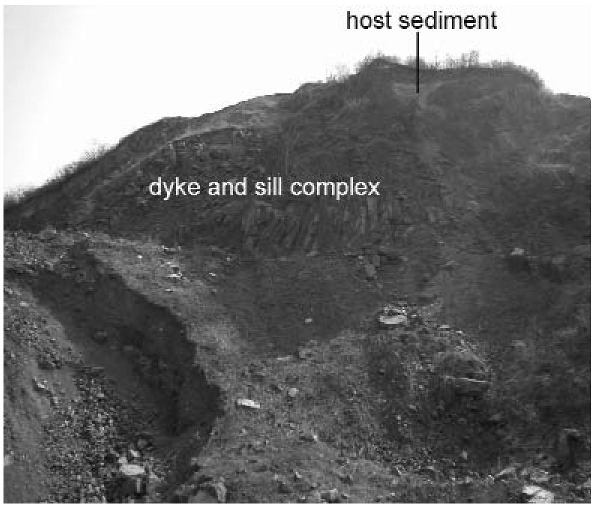
Note the plateau forms a circular zone at Tátika (B, C) that corresponds well with the extent of coherent basanite units (white line). On this plateau a cliff is well visible on aerial photo (C). This cliff preserves a massive, lapilli tuff succession (white circle), however, in poorly preserved condition

Kovácsi-hegy

The Kovácsi-hegy comprises the westernmost Neogene alkaline basaltoid volcanic hills north of the Keszthely Mts. (Figure 4.3). The Kovácsi-hegy forms a circular plateau about 300 m in elevation marking the extent of the lava flows over the Neogene sands and silts. East of this lava plateau, the morphology is more rugged, and consists of north-east to south-west, and north-west to south-east, trending ridges, similar to the ridge between Sümegprága and Tátika (Figure 4.3). These ridges consist of coherent basanite buds having intrusive contacts with Neogene siliciclastic host units similar to those exposed in east (Figure 4.4). The contact between the pre-volcanic rocks and coherent basanite is not exposed at the Kovácsi-hegy. In a large quarry system along the western margin of the Kovácsi-hegy a basanite unit at least 50 m thick and comprises at least two cooling units. The basanite body is columnar jointed with perpendicular, very regular columns that are 20 to 40 cm in diameter. In the area there are no pyroclastic rocks exposed.



Figures 4.3.
Location map of the Kovácsi-hegy (A). On the airphoto it is clearly visible, that the Kovácsi-hegy forms a circular plateau around 300 m elevation (B). East of this plateau, the morphology is more rugged, and consists of north-east to south-west as well as north-west to south-east trending ridges, similar to the ridge between Sümegprága and Tátika. These ridges consist of basanite buds with intrusive contact to a host Neogene siliciclastic units similar to those exposed in east, q = quarry



Figures 4.4. Intrusive basanitic lava bud – similar to the lava buds at Bazsi – in a host Neogene sand and silt from a small quarry east of the Kovácsi-hegy

Conclusion

The volcanic rocks mapped north of the Keszthely Mts have long been a subject of geological research in the region (LÓCZY 1913, 1920, JUGOVICS 1948). Each researcher noted the general lack of pyroclastic rocks in these locations in comparison to other Neogene alkaline basaltic regions in the western Pannonian Basin. Current research confirms that there is a general lack of pyroclastic rocks in this region, and this may be of some significance in analysis of the volcanic evolution of the Neogene alkaline basaltic systems in the western Pannonian Basin. The volumetrically largest known pyroclastic succession is at Uzsa, which had a significant phreatomagmatic explosive eruptive history, and may have formed a maar and/or tuff ring. The Uzsa phreatomagmatic succession subsequently was covered by effusive and eruption products generated by magmatic fragmentation. The topmost volcanic succession at Uzsa has been interpreted as the erosional remnant of scoria cones or possibly of lava deltas. The circular distribution of the lava fields at Uzsa suggests some form of control on the extent of the lava, such as a tuff ring and/or maar crater wall. Inferring accumulation of the volcanic material in a confined maar basin with its base below surrounding ground level, the present highest level of scoriaceous pyroclastic rocks in a mound-like hill south of the Uzsa quarry therefore is inferred to represent the syn-volcanic palaeosurface. Clarification of this interpretation is a subject of future research.

The ridge between Sümegprága and Tátika is clearly a shallow subsurface sill and dyke complex. This system, at least in one place, has been cut through by a diatreme (near Tátika). This implies that the emplacement of sills and dykes pre-dates formation of the phreatomagmatic volcanoes.

At the Kovácsi-hegy, a sill and dyke system similar to the Sümegprága and Tátika system has been identified. The Kovácsi-hegy rocks are inferred to have formed as a basanitic sill on the basis of its general recent elevation and its 3D relationship with the neighbouring coherent flow units, which are clearly intrusive in origin.

The intrusive origin of the majority of the basanitic rocks north of the Keszthely Mts suggests that significant erosion took place since their emplacement around 3 My. A conservative estimate would reconstruct the level of the palaeosurface at the level of the present top of the Tátika, where pyroclastic rocks crop out. However, if the Tátika is the remnant of an exhumed diatreme, it would imply that its present top section represents a level well below the syn-volcanic palaeosurface. A realistic estimate would add a minimum of 50 metres above the peak of the present Tátika to establish the reference elevation for the syn-volcanic palaeosurface. With this calculation, the minimum missing Neogene sedimentary cover would be in the range of 250–350 metres in comparison to the presently preserved average of ~150 metres of strata. This would imply that the sill and dyke complex in the region developed around 100 to 250 metres below the syn-volcanic palaeosurface. The present day high altitude of the intrusive rocks at this region in comparison to other effusive coherent lava rock locations in the BBHVF in a more or less similar elevation suggests some sort of differential base level changes through the Neogene which process needs further study (MARTIN and NÉMETH 2003).

The recognition that a majority of the coherent igneous rocks exposed north of the Keszthely Mts. originated as intrusive sill and dyke complexes also highlights the complexity of magma emplacement and feeding systems for small-volume intracontinental alkaline basaltic volcanic systems. There are growing examples worldwide that demonstrate that maar/diatreme volcanism is often associated with complex effusive (extensive lava flows) and intrusive (dykes, sills and laccoliths) processes such i.e. as it has been reported from the Saar-Nahe Basin, Germany (LORENZ and HANEKE 2004). Both recent and preceding seismic studies have identified several mound-like, high velocity zones within Neogene strata a few tens to a hundred metres below the surface in the Lake Balaton basin (CSERNY and CORRADA 1989, SACCHI et al. 1999, SACCHI and HORVÁTH 2002). These structures are best interpreted to represent similar structures as the Sümegprága–Tátika sill and dyke systems that never made it to the syn-volcanic surface, and which are not yet exhumed.

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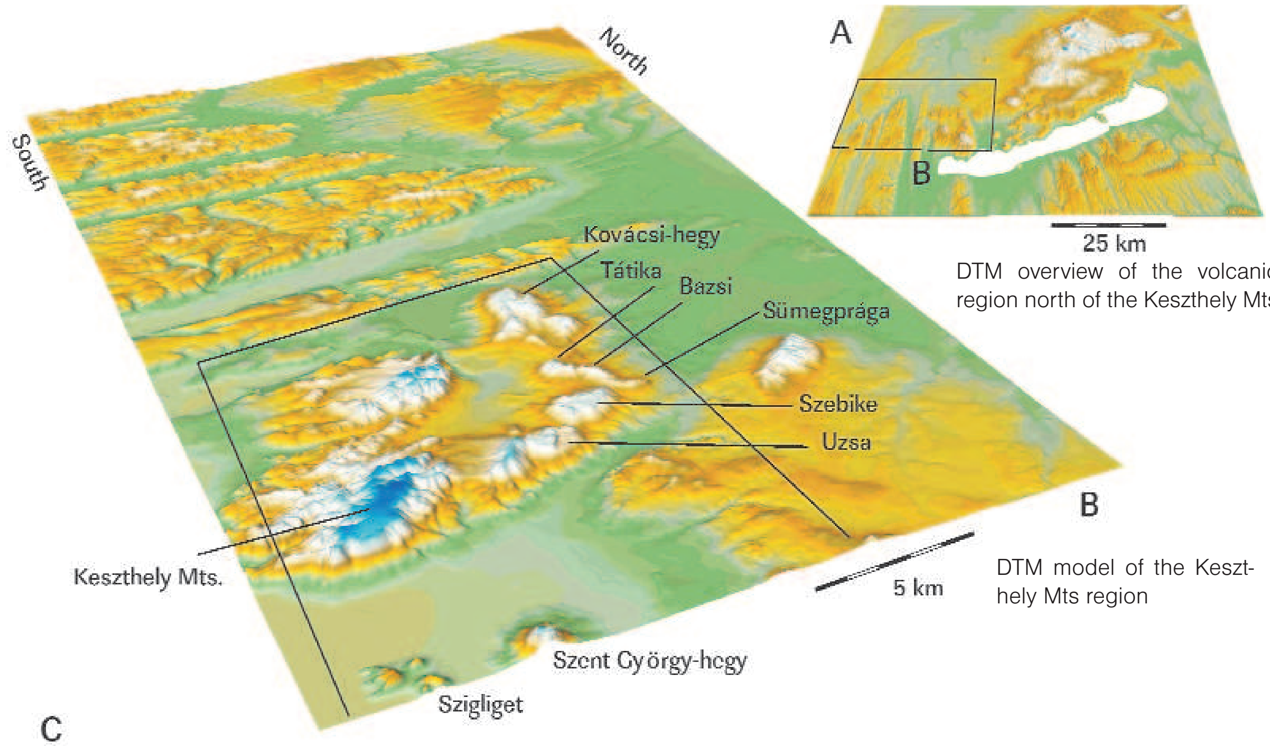
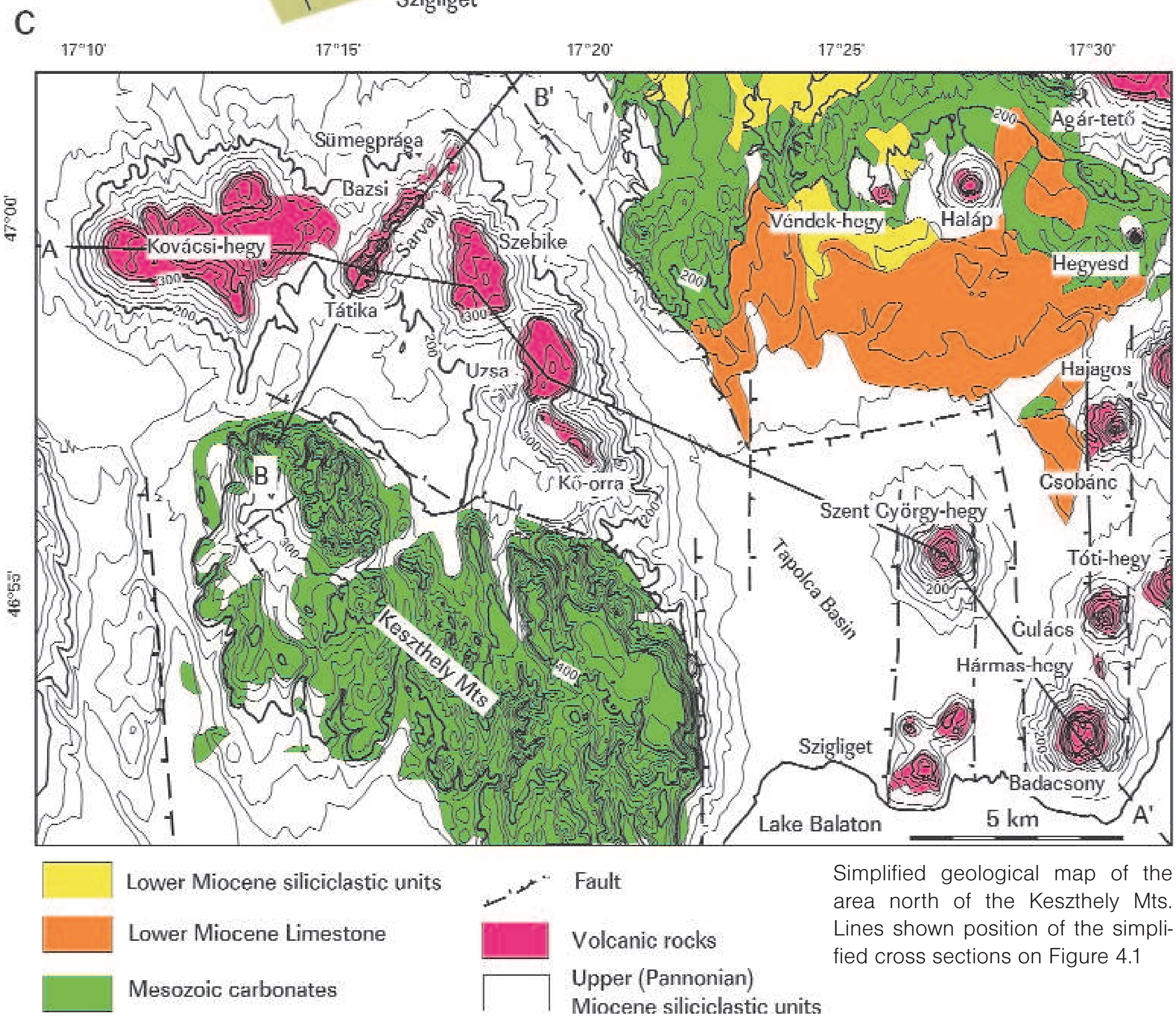
Chapter 4 / Plate 1



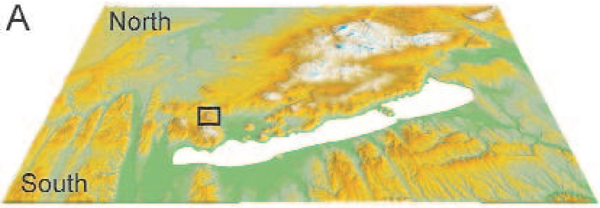
DTM overview of the volcanic region north of the Keszthely Mts

DTM model of the Keszthely Mts region

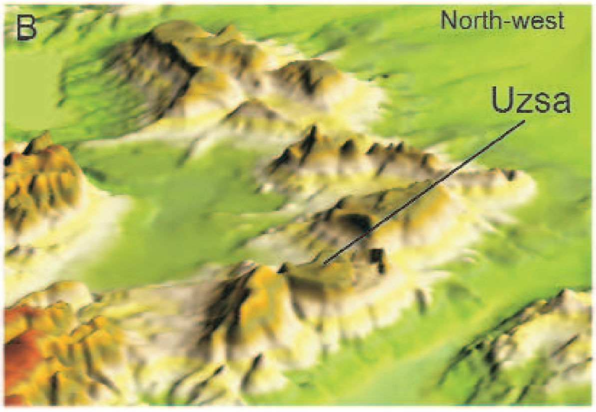


Simplified geological map of the area north of the Keszthely Mts. Lines shown position of the simplified cross sections on Figure 4.1

Plate 2 / Chapter 4 Second International Maar Conference — Hungary–Slovakia–Germany

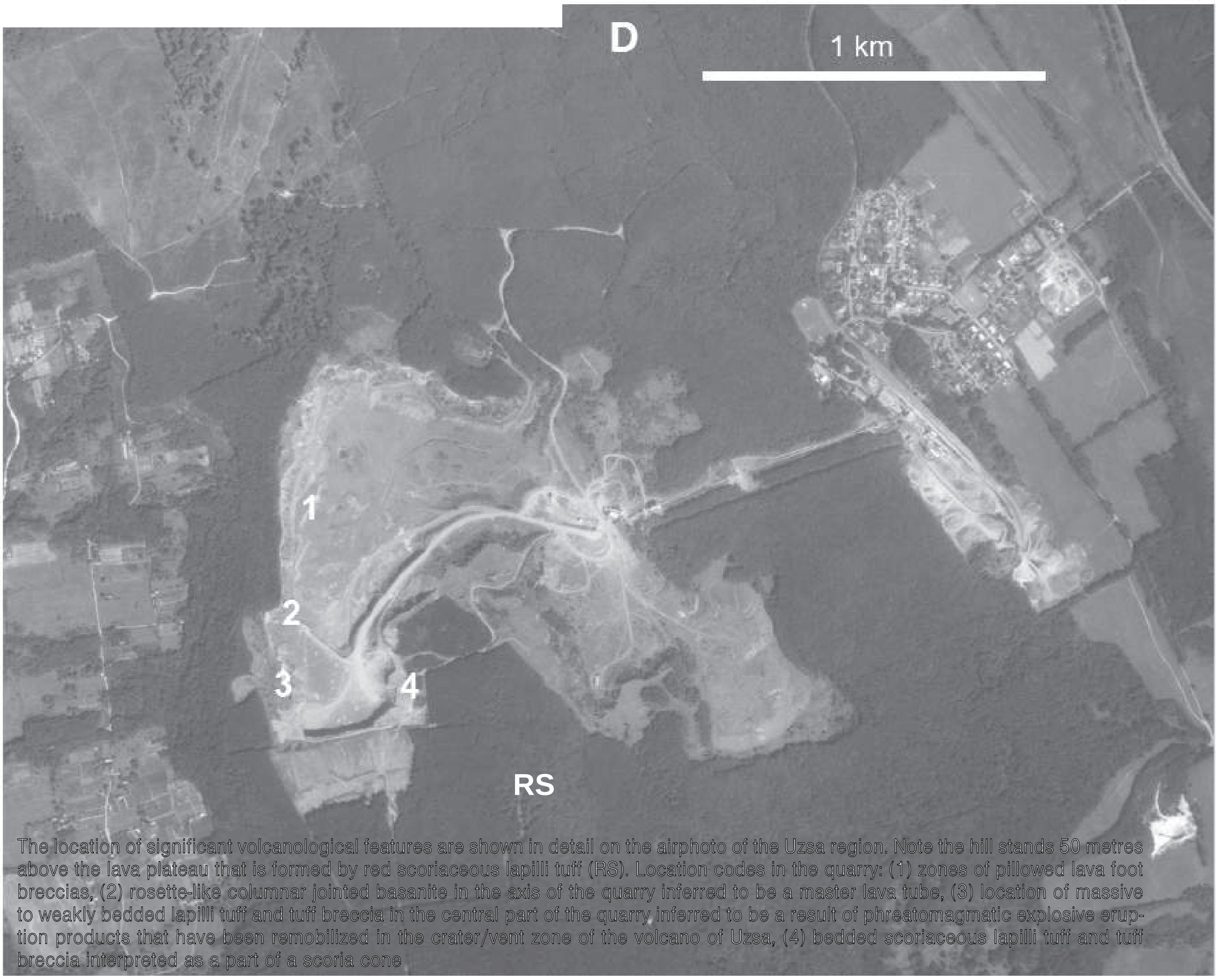


Location of the Uzsa basanite capped mesa in the western Pannonian region



The Uzsa region shown on a DTM looking to north-west

A detail of the 1 to 25,000 topography map of the Uzsa region shows the 3D relationships of the topography in relationship with the location of the large basalt quarry in Uzsa



The location of significant volcanological features are shown in detail on the airphoto of the Uzsa region. Note the hill stands 50 metres above the lava plateau that is formed by red scoriaceous lapilli tuff (RS). Location codes in the quarry: (1) zones of pillowed lava foot breccias, (2) rosette-like columnar jointed basanite in the axis of the quarry inferred to be a master lava tube, (3) location of massive to weakly bedded lapilli tuff and tuff breccia in the central part of the quarry inferred to be a result of phreatomagmatic explosive eruption products that have been remobilized in the crater/vent zone of the volcano of Uzsa, (4) bedded scoriaceous lapilli tuff and tuff breccia interpreted as a part of a scoria cone

A

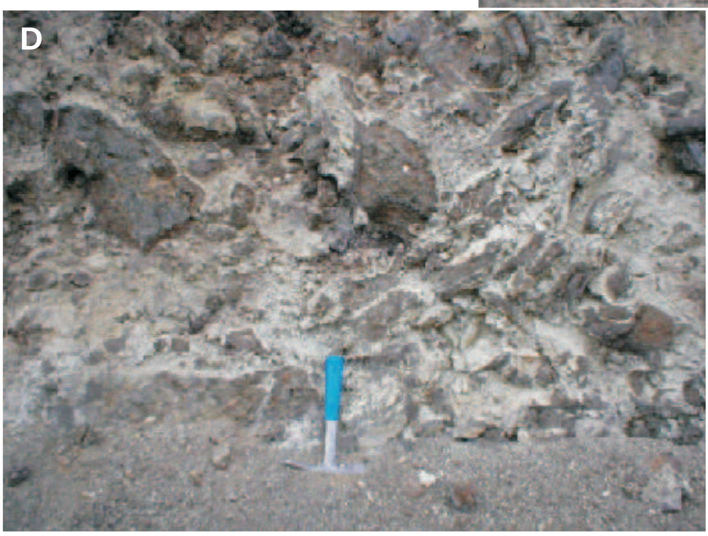
Columnar jointed coherent basanite in the western quarry wall of Uzsa

Basanite columns at Uzsa are about 20–25 cm wide. A perpendicular view to the columns shows that columns are predominantly pentagonal

B

C

Vesicular basal zone of a coherent basanite at Uzsa



D

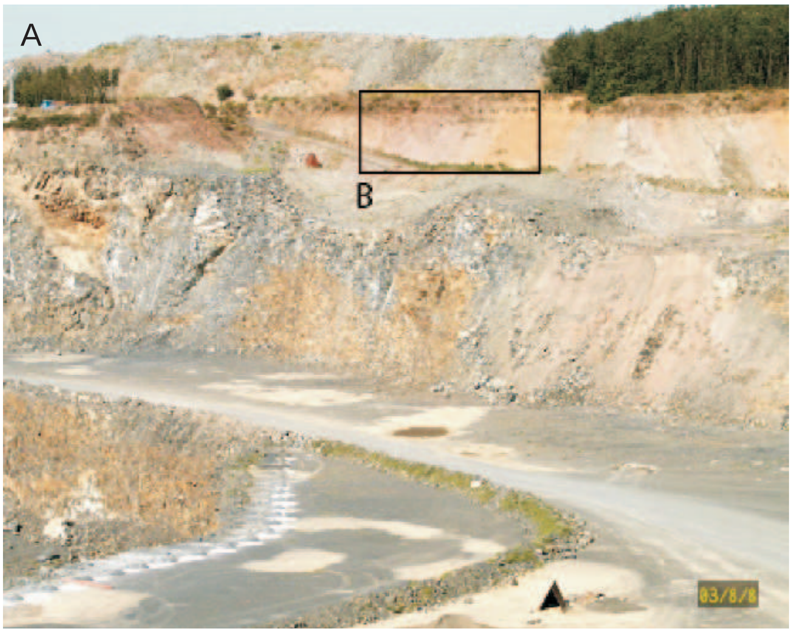
Well packed pillow-like structure of the lava foot of a lava flow at Uzsa

Onion-like jointing pattern near the margin of the coherent magmatic body at Uzsa

A coherent basanite tube forming a half circle (white line), rosette-like joint pattern in the western side of the quarry at Uzsa. Note that the rosette-like jointing pattern gradually defect to more platy jointing pattern near the margin of the flow

Massive to weakly bedded lapilli tuff and tuff breccia in handspecimen in the western part of the Uzsa quarry

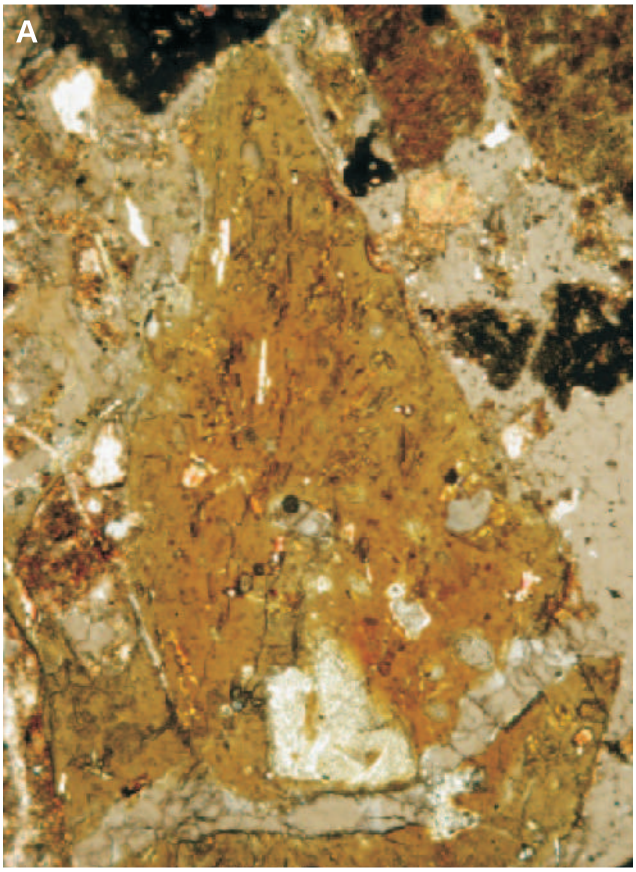
Red scoriaceous, weakly bedded lapilli tuff and tuff breccia in the southern margin of the Uzsa quarry



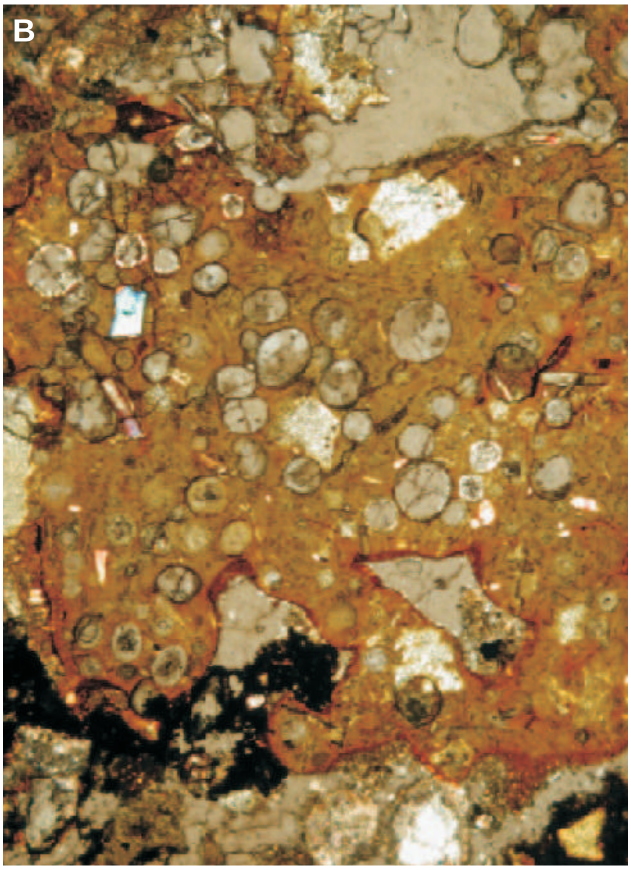
Textural features of the scoriaceous lapilli tuff succession show characteristics for remobilisation of tephra on a volcanic flank

The massive structure of the pyroclastic breccia of Uzsa is more prominent in a close up view

Massive fine grained lapilli tuff unit of Uzsa

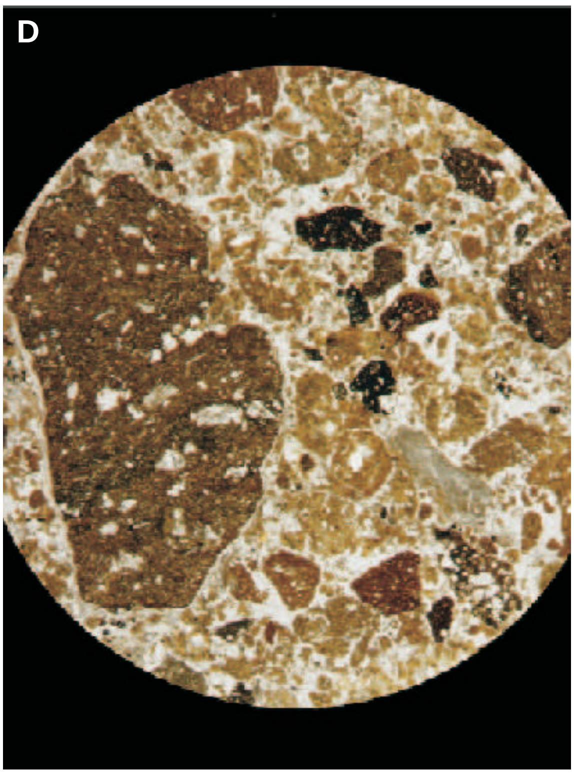


Photomicrograph of blocky, moderately vesicular tephritic glass shard from the Uzsa pyroclastic succession. The short side of the photo is 2 mm. [plan parallel polarized light]



Photomicrograph of a glass shard of the lapilli tuff unit of Uzsa. They are commonly palagonite rimmed (dark zone). The short side of the photo is 2 mm. [plan parallel polarized light]

Blocky shaped, dense volcanic lithic fragment in a fine-grained matrix of a tuff breccia of Uzsa



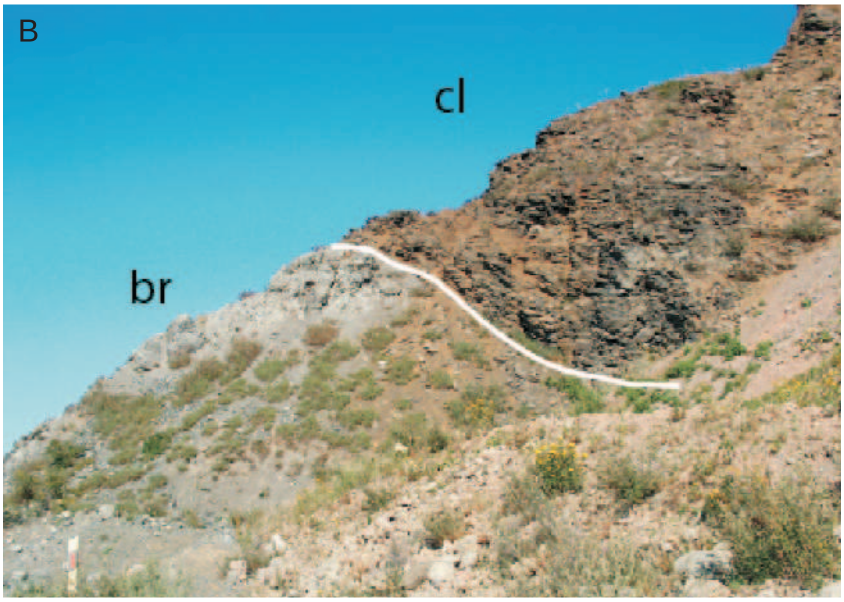
The matrix of the tuff breccia of Uzsa is rich in volcanic glass shards indicating its phreatomagmatic origin as it is shown in the photomicrograph. [plan parallel polarized light, the view is about 2 cm across]

Irregular contact (thick white line) between weakly bedded (straight white line represents bedding), ill-sorted lapilli tuff and basanitic intrusion in the upper level of the Uzsa quarry

In a close up view of the lapilli tuff at Uzsa the bedding is more prominent and cauliflower bombs (white circle) can be recognized indicating magma–water interaction during fragmentation

White, semi-rounded, carbonate-rich silt (white circles) as accidental lithic fragment in the basal tuff breccia of Uzsa

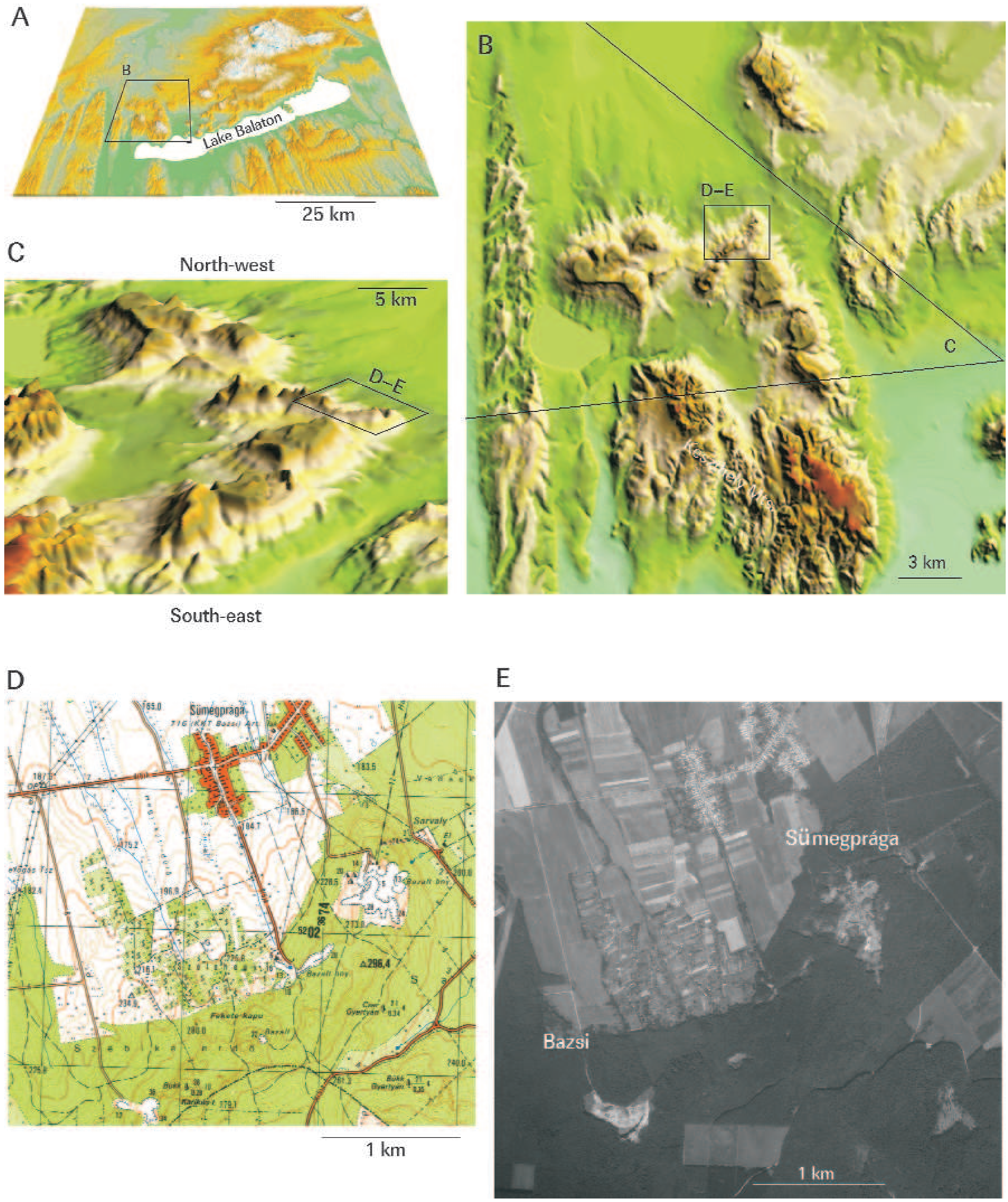
Fluidally shaped moderately vesicular lava spatter in a scoriaceous lapilli matrix from the south-eastern margin of the Uzsa quarry

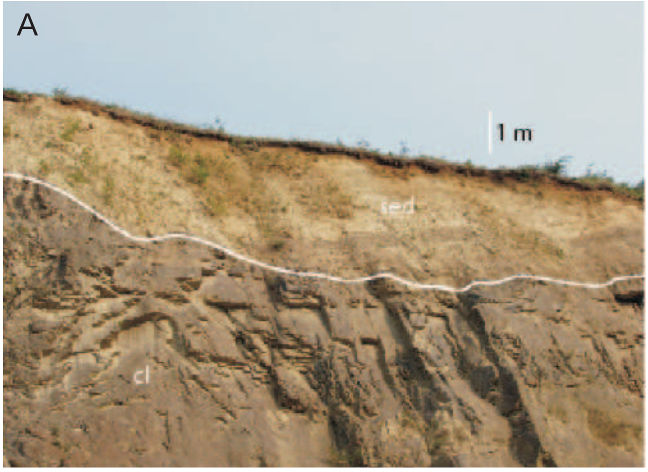


Coherent basanite (cl) and pyroclastic breccia (br) irregular contact in the top-most volcanic units of Uzsa indicating some close relationship between lava flow and clastic rocks in this part of the quarry

Coarse grained pyroclastic breccia at Uzsa

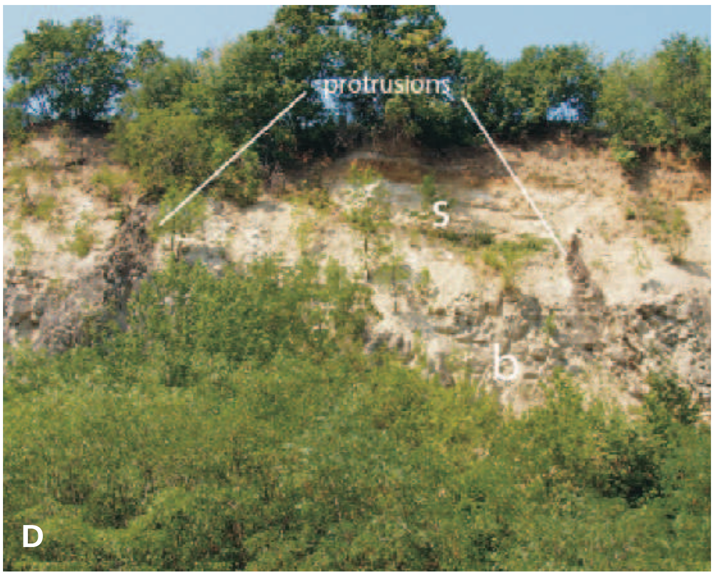
A well developed ridge north of the Keszthely Mts runs from north-east to south-west where predominantly basanitic coherent lava bodies crop out (A, B, C). The northernmost location is a former quarry, Sümegprága (D and E), where basanite shows radial jointed pattern and have intrusive contact with the host siliciclastic sediments (silt, sand).





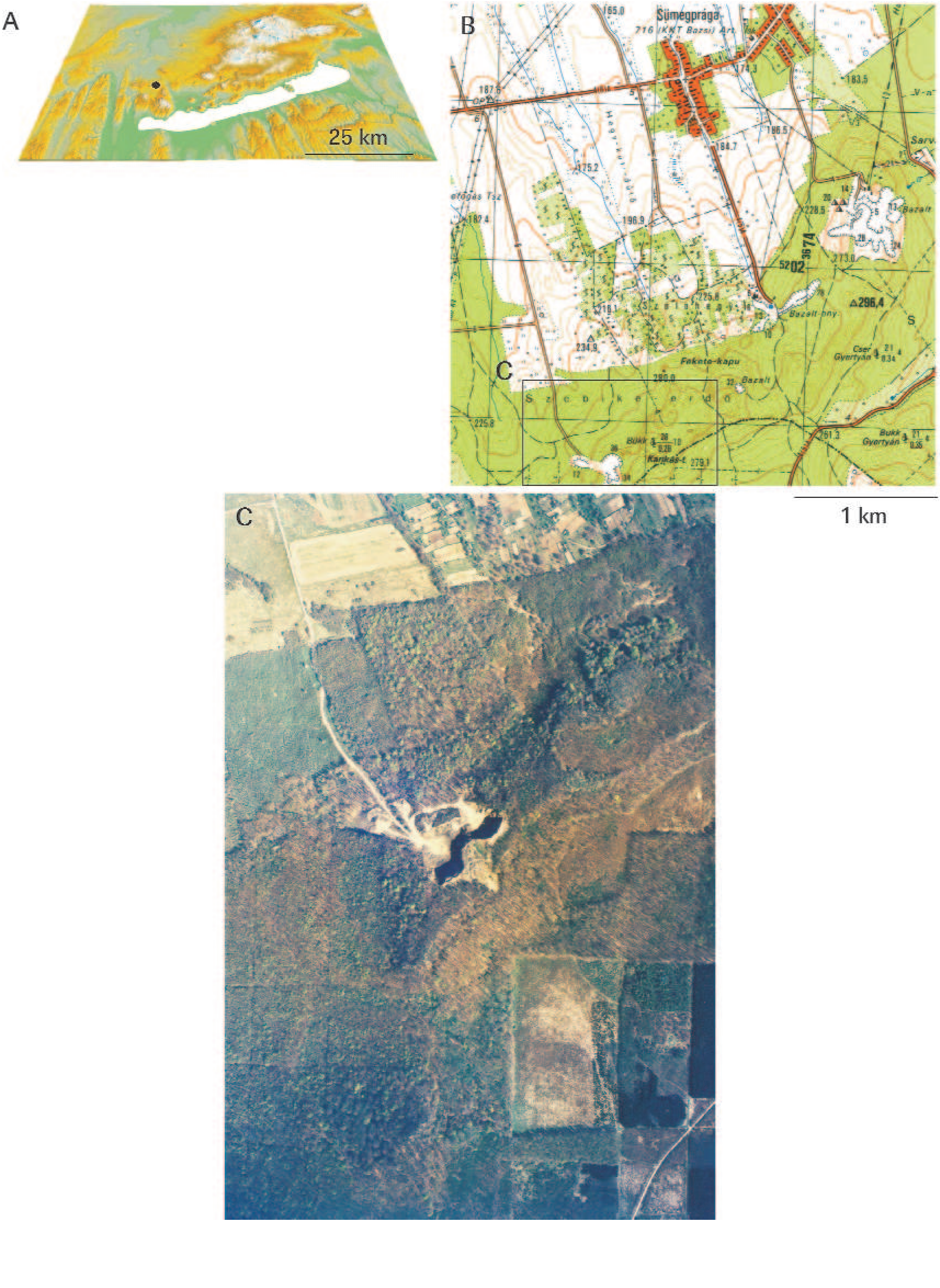
Two types of intrusive coherent magmatic bodies from Sümegprága. A) a tabular unit (cl) with irregular contact (white line) to the host sediment (sed) and B) a rosette-like columnar jointed magmatic unit (cl) with irregular contact to the host sediment (white line)

Baked silt/sand (s) at the contact of basanitic body (cl) and host sediment (photo courtesy of Manuella Kramer)



Small, vertical protrusion of basanite (b) that intruded from a master sill into the host sediment (s) at Sümegprága (photo courtesy of Claudia Henke)

Location of the Bazsi quarry (A, B) with an outline of the currently active quarry on an airphoto (C)



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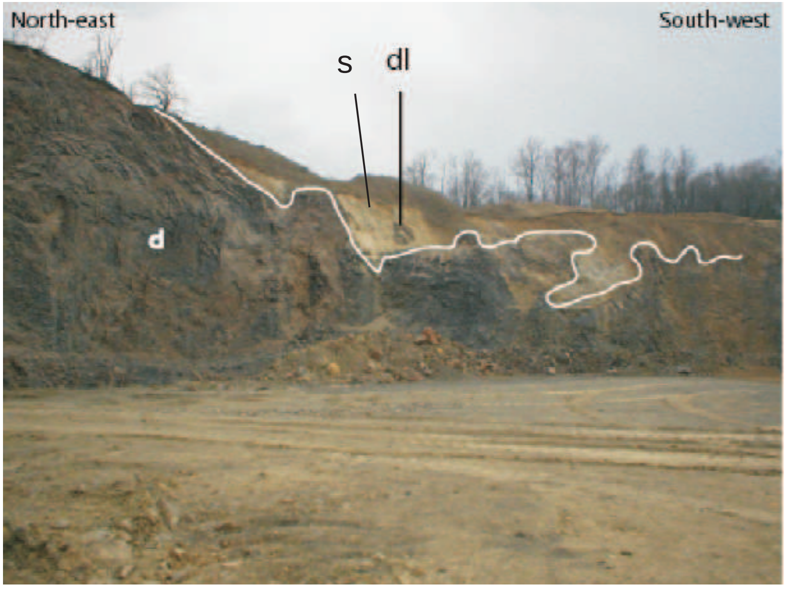
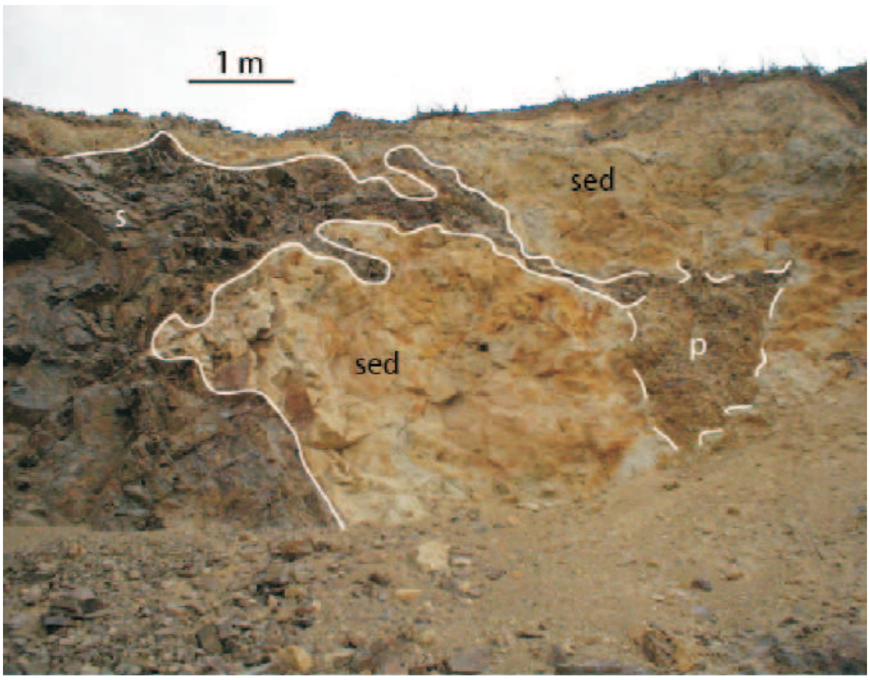
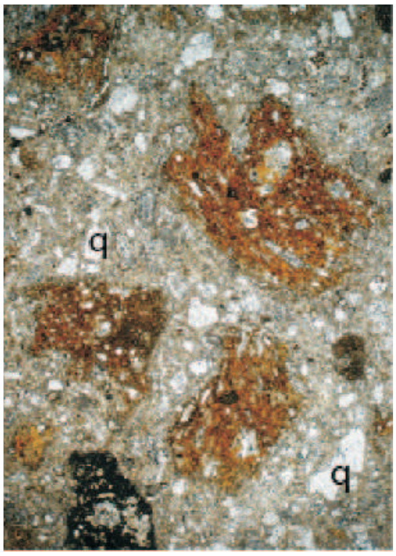


Overview of the master sill (d) of Bazsi. Note the very irregular shape of the sill (white outline) and the detached coherent lobe (dl) "floating" in the host (s)

Peperitic margin of a sill (s) at Bazsi



Small sill (white line) that protruded from a master sill (s) into a host sediment (sed) at the upper level of Bazsi. The sill then inflated and terminated into a volcaniclastic unit (white dashed line) that is seemingly disperse into the host sand/silt matrix, and forming peperite (p)



Photomicrograph [plan parallel polarized light, the short side of the photo is about 4 mm) of a lapilli tuff derived from the Tátika. Note the angular quartz fragments (q) derived from the immediate pre-volcanic Neogene rock units. The pyroclastic rocks contain abundant sideromelane glass (s, which are blocky and moderately vesicular and thus indicative for phreatomagmatic fragmentation

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Mio/pliocene phreatomagmatic volcanism in the Little Hungarian Plain Volcanic Fied (Hungary) and at the western margin of the Pannonian Basin (Austria, Slovenia)

Contents

Abstract

The Little Hungarian Plain Volcanic Field (LHPVF) is a Mio/Pliocene alkaline basaltic volcanic field that is located in a Mio/Pliocene sub-basin of the Pannonian Basin called the Danube Basin. Volcanic rocks crop out as erosion remnants of former mound-like phreatomagmatic volcanoes often topped by scoria cones and/or lava flows. They are similar to the volcanic erosional remnants in the Bakony – Balaton Highland Volcanic Field. In the westernmost margin (Austria and Slovenia) of the Pannonian Basin, small strongly eroded phreatomagmatic volcanoes evolved in fluvial environments. They are preserved on alluvial fans, and/or their debris was reworked into Quaternary alluvial deposits. The volcanic erosional remnants of the LHPVF are generally mound-like pyroclastic rocks that comprise volcanic glass shards and fine siliciclastic clasts suggesting near surface phreatomagmatic explosive fragmentation of the rising melts through a thick, water-rich Neogene siliciclastic succession. The presence of sideromelane glass shards, glassy juvenile lithic fragments as well as the variable amount (but always present) siliciclastic detritus indicates near surface phreatomagmatic explosive eruptions, which may have even occurred in shallow standing water body on an alluvial plain near to the palaeo -ground water table.

The level of exposure of the erosional remnants is relatively shallow in comparison to the Balaton Highland examples and often exhibits coherent lava flows and/or crater lake lacustrine units accumulated in tuff ring and/or shallow maar craters. The generally flat dip of the phreatomagmatic lapilli tuffs, especially from the LHPVF is interpreted to be a result of broad tuff rings. There are also intact volcanic edifices that resemble skeleton structures of tuff rings with wide craters that have been exhumed recently.

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

Introduction

The Mio/Pliocene erosional remnants of the alkaline basaltic, small-volume, intraplate volcanoes in the Little Hungarian Plain Volcanic Field (LHPVF– Plate 5.1) are located near to major tectonic lines, such as the Rába detachment fault and the perpendicular strike-slip faults (TARI et al. 1992, TARI 1994, SCHAREK et al. 1995, SCHAREK 1996). The erosion remnants of this volcanic field consist of moderately eroded, lensoidally shaped (in map view) mounds of pyroclastic rocks, often covered by variably thick lava caps. In spite of the large number of surface exposures of volcanic rocks in the region, covered (or semi-exposed) volcanic structures are also present and are identified on the basis of geophysical (TÓTH 1994) and/or drill core data (SCHLÉDER and HARANGI 2000).

The underlying basement of the LHPVF consists of Palaeozoic to Mesozoic metamorphic rocks (gneiss, schist etc.) (BALLA 1993, NEMESI et al. 1994) covered by thick (up to 6000 m) Miocene siliciclastic sediments (PHILLIPS et al. 1992, TARI et al. 1992, KOVAČ et al. 1993, TARI 1994, HORVÁTH and TARI 1999, MAGYAR et al. 1999, SACCHI et al. 1999, SZAFIÁN et al. 1999). The Little Hungarian Plain is a Neogene to Quaternary sedimentary basin filled by thick Miocene to Quaternary, predominantly siliciclastic sediments. The basement rocks of this basin consist of crystalline units belonging to the Upper and Lower Austroalpine terrain (HORVÁTH 1993). The basin formation was facilitated by the uplift of the Penninic metamorphic core complexes and the development of an extensional basin system bounded by low angle normal faults (TARI et al. 1992, HORVÁTH 1993). Several seismic profiles, magnetotelluric studies as well as geochemical evidence suggest the existence of a supracrustal fault and an astenospheric dome along the axis of the Little Hungarian Plain (HORVÁTH 1993). Both may have significance in the development of the volcanic field as suggested in similar basins (WALKER 1989, CONNOR et al. 2000). In the Neogene, during an extensional tectonic regime (TARI 1993), just shortly before volcanism started, a large lake occupied the Pannonian Basin, the Pannonian Lake (KÁZMÉR 1990, HORVÁTH 1993, MAGYAR et al. 1999, SACCHI et al. 1999). The Lower Miocene units in the LHPVF are deep-water siliciclastic deposits (HORVÁTH 1993, NEMESI 1994). In Upper Miocene from the Pannonian Basin is characterised by prograding deltas that developed from NW to SE, which led to the diminishing of the sub-basins first in the LHPVF area (VAKARCS et al. 1994, MAGYAR et al. 1999). Shallow lacustrine sandstones, mudstones, and marls of the brackish Pannonian Lake are widespread in the western part of the Pannonian Basin and form the immediate pre-volcanic rocks around all of the volcanic remnants of the LHPVF (TREGELE 1953, VARRÓK 1953). All of these sediments are good porous media aquifers today (ROTÁR-SZALKAI 1998). A southward prograding delta system gradually filled the Pannonian Lake, and in the Pliocene time led to the development of an alluvial plain (JÁMBOR 1989, JUHÁSZ et al. 1997, 1999, MÜLLER 1998). Large, shallow, standing water bodies (10-m-scale) may have developed in the region especially during wet seasons. Consequently, volcanism occurred in subaerial settings, along fluvial valleys likely filled with swamps, small streams or shallow lakes all providing substantial surface (as well as near-surface) water to fuel phreatomagmatic volcanism. Water-saturated sediments (mud) played an important role in magma-water interaction (NÉMETH and MARTIN 1999). Pre-volcanic sedimentary rocks at each location consist predominantly of gravel, sandstone, siltstone and mudstone, with marly inter-beds deposited in a shallow sub-lacustrine to fluvio-lacustrine environment (VARRÓK 1953, KAISER et al. 1998). These pre-volcanic lake deposits commonly are very fine-grained and distinct in their creamy colour (JÁMBOR 1989, MAGYAR et al. 1999). Individual siliciclastic beds immediately underlying the volcanic rocks are structureless to weakly bedded and/or cross-stratified and form cm-to-dm-scale beds. The contact between pre-volcanic and pyroclastic beds is in most cases not exposed, but it is inferred to be in angular unconformity. Drill core data on the database held at the Geological Institute of Hungary show that the contacts between volcanic rocks and immediate pre-volcanic siliciclastic successions are erosional rather than a purely depositional.

Recent studies, based on comparative drill core analyses, seismic sections and palaeontological studies show that extensive lacustrine sedimentation of the Pannonian Lake with open surface water masses (tens of metres deep) likely ceased in the Little Hungarian Plain ~9 My ago (VAKARCS et al. 1994, MAGYAR et al. 1999, SACCHI et al. 1999, SACCHI and HORVÁTH 2002).

Kis-Somlyó tuff ring

Introduction

Volcanic rocks of Kis-Somlyó are part of a Pliocene erosion remnant of an alkaline basaltic tuff ring located in the southern edge of the Little Hungarian Plain Volcanic Field (LHPVF – Plate 5.1 and Figure 5.1). Late Miocene shallow subaqueous, fluvio-lacustrine sand(stone) and mud(stone) units underlie sub-horizontally bedded lapilli tuff and tuff beds with an erosional contact (Figures 5.1 and 5.2). The pyroclastic units build up a ~20 m thick sequence, forming a semi-circular mound structure (JUGOVICS 1915, VARRÓK 1953, JUGOVICS 1968, MARTIN and NÉMETH 2002, 2004a) with gentle (<15 degrees) inward dipping beds (Figure 5.1 and Plate 5.2, A). Sedimentary features and field relationships indicate that the pyroclastic units were formed in a terrestrial setting, in a shallow lake and/or swamp. Phreatomagmatic explosions occurred at shallow depth or close to the water surface, producing a large amount of disrupted juvenile ash and lapilli, transported and deposited predominantly by pyro-

Figure 5.1. Simplified geological map and the topography of the Kis-Somlyó area Numbers (1and 2) represent the location of the measured logs on Figure 5.2

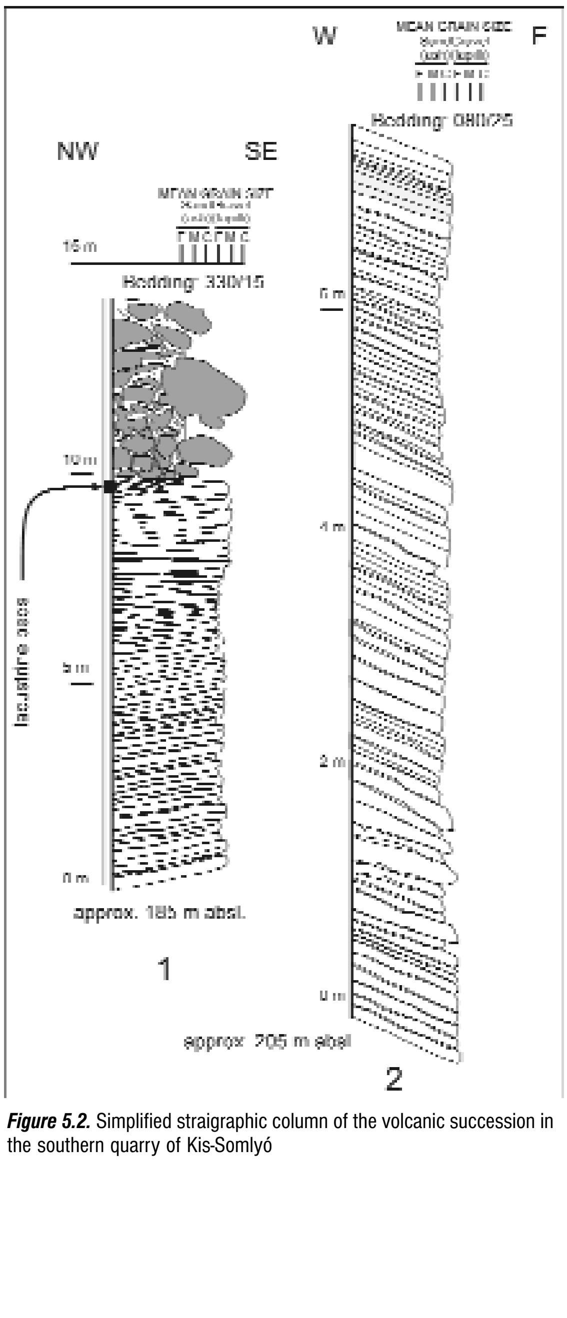

Figure 5.2. Simplified straigraphic column of the volcanic succession in the southern quarry of Kis-Somlyó

clastic density currents and subordinate fallout. The pyroclastic units are overlain by cross- and parallel laminated siltstone and mudstone deposited in a lake that developed over the tuff ring. The textural and structural differences between the underlying and overlying lacustrine units suggest that they did not belong to the same lacustrine sedimentary environment. It is inferred that a lake developed shortly after the formation of the tuff ring. The preserved thickness of the post-tuff ring lacustrine units is approximately 5 m resulting in a water depth of the subsequent lake on a scale of metres. The post-tuff ring lacustrine sequence is invaded by basanite lava. The lava shows a peperitic margin partially destroying the original texture of the lacustrine beds due to fluidisation and heat effects. The time gap between the tuff ring formation and the emplacement of the lava flow is estimated to be a few thousand years, calculated from the thickness of the laminae of the post-tuff ring lacustrine sediments.

Locally, at least 5 m thick siliciclastic deposits overlie the pyroclastic units of Kis-Somlyó volcano, and were invaded by subsequent lava extrusions and intrusions forming peperite (MARTIN and NÉMETH 2002). There are two K/Ar ages available on whole rocks (4.97±0.31 My and 4.04±0.17 My: (BALOGH et al. 1986). However, excess Ar has been inferred to have caused an error in the older age. An age of 4.11±0.29 My measured on the most magnetic fraction has been considered to be the most likely age by K/Ar method (BALOGH et al. 1986). A new $^{39}Ar/^{40}Ar$ age determination gave an isochron age of 4.61±0.04 My. This is older than the K/Ar age, however, and the reason for this is under debate (WIJBRANS et al. 2004).

The existence of siliciclastic sediments above pyroclastic units which can mimic the sedimentary texture of the pre-volcanic sediments derived from the Pannonian Lake, highlights the difficulty to distinguish pre- syn- and post-volcanic sedimentary cycles in an intracontinental setting, as well as the palaeogeographical importance of this locality in regard to reconstructing a complex fluvio-lacustrine system in this region during the Pliocene.

Erosion remnant morphology

Kis-Somlyó (220 m) is a small flat volcanic remnant consisting mainly of pyroclastic rocks and sporadic lava flows. The pyroclastic rocks form a semi-circular mound, preserved below an up to 5 m thick lava cap in its centre (Figure 5.1). The contact between pre-volcanic siliciclastic and pyroclastic rocks is not exposed but it is inferred to be at an elevation between 195-200 m (Figure 5.1). The mound shaped erosional remnant sits ~150 m above the basin floor of the Little Hungarian Plain (Plate 5.1). The pre-volcanic siliciclastic units are exposed in small sand pits about 160 m in elevation, having distinct sub-horizontal bedding. Similar sub-horizontal bedding of the pre-volcanic siliciclastic units have also been reported from shallow water research drillings and other geophysical research reports (DUDÁS et al. 1994, HOBOT and DUDÁS 1994). The pyroclastic rocks have gentle hill-ward dipping orientation, inferred to be the primary dip angle. A few metres-sized blocks with steep internal bedding are inferred to have been tilted subsequently by younger landslides.

The preserved pyroclastic sequence is at least 20 m thick. Beds of the pyroclastic rocks dip radially (<15°) towards the centre of the hill, form a collar of exposures in the southern side of the remnant and are hundred metres in length and few metres in thickness. The exposed sections exhibit a relatively uniform coarse/fine alternation of poorly sorted, bedded lapilli tuffs and tuffs (Plate 5.2, B). Bedding is sub-horizontal in the marginal zones of the pyroclastic mound (Figure 5.1). Steep dip directions toward the centre of the preserved pyroclastic units have been reported (JUGOVICS 1915, VARRÓK 1953) and are confirmed during recent mapping on the SW and W side of the hill, where large (10 m-scale) tilted blocks are located close to floor (~165 m) of the Little Hungarian Plain (JUGOVICS 1915).

The description and interpretation of the pyroclastic units and post-volcanic lacustrine units are summarised in MARTIN and NÉMETH (2004a).

Pyroclastic units

Description: There are two major types of lithofacies (P1 and P2) that have been distinguished in the exposed pyroclastic succession on the basis of bedding, grain size, componentry and the ratio between juvenile and accidental clasts (MARTIN and NÉMETH 2004a). There are fine grained accidental lithic or accidental lithic-derived mineral phase rich, thinly bedded, often cross bedded lapilli tuff and tuff beds (P1) randomly intercalated with coarser grained, rounded juvenile lapilli-bearing, calcite cemented thickly bedded, massive to weakly stratified lapilli tuffs (P2 – Plate 5.2, C). These two lithofacies types form two end-members but there are also transitional types.

Single pyroclastic beds (P1) are normally or reversely graded and occasionally show low angle cross stratification, antidunes or undulating beds (especially the fine grained tuffs and lapilli tuffs). Beds are typically a few cm thick, but 15–20 cm thick massive, fine grained tuff beds are also present in the lower part of the succession.Tuffs and fine lapilli tuffs often contain sporadic accretionary lapilli, as well as armoured lapilli. In general, beds are laterally continuous on 10 metres-scale. Tuffs and lapilli tuffs are formed mainly by semi-rounded to blocky sideromelane glass shards (moderate to highly vesicular), glassy volcanic lithics, tachylite, microcrystalline to aphanitic basaltic and pre-volcanic lithic clasts as well as armoured lapilli (Plate 5.2, D). Larger accidental lithic clasts are predominantly rounded to sub-rounded, dense, often radially fractured sandstone fragments (up to 70 cm in diameter) and/or flat, fluidal shaped plastically deformed mud (up to 20 cm in length). The most common pre-volcanic lithic clasts are various siliciclastic fragments variable in size (Plate 5.2, D, E). Large, up to 1 m in size, hard, often semi-rounded medium to coarse grained sandstone, as well as fine grained mudstone are the most common non-volcanic lithics. These hard lithic fragments often cause impact sags on the underlying bed (Plate 5.3, A), however, there is no systematic correlation between the size of lithics and the depth of impact sags. Deep bedding sags are bed specific and appear commonly below coarse lapilli tuff beds. In contrast, there are beds with no or very shallow bedding sags in fine-grained beds, although blocks commonly reach diameters up to 50 cm (Plate 5.3, B). The transport direction determined from impact sags shows a radially outward direction from the centre of the erosional remnant. Sorting of the pyroclastic beds regardless to their average grain size is poor to moderate. Deep-seated crystalline or other exotic accidental lithic clasts are rare. Cauliflower bombs (up to 15 cm in diameter) are characteristic in the whole section, commonly preserving olivine megacrysts in their interior.

Coarse grained lapilli tuff beds appear to exhibit more pronounced inverse or inverse-to-normal grading (P2), however, because many contacts are diffuse the grading is often unclear. Most of these beds have a characteristic separation of a lower lapilli-rich and an upper fines-enriched layer (Plate 5.3, C), which is more enhanced by the colour difference, which is yellowish tan in the fine grained and greyish in the lower lapilli tuff beds. This gives a prominent appearance of bed couplets in certain outcrops. There is cross lamination in fine-enriched, muscovite-rich (<5 cm thick) pyroclastic beds with diffuse contacts with coarser grained lapilli tuff beds. These beds consist of juvenile glass shard rich lapilli tuffs with a low amount of matrix but are often strongly cemented by micritic, as well as sparritic calcite (Plate 5.3, D). The lapilli are semi-rounded to well-rounded, and have abraded outer rims. The vesicularity and microlite content of the glass shards vary extremely (Plate 5.3, D).

Interpretation: The presence of sideromelane glass shards, cauliflower lapilli and bombs in P1 and the presence of characteristic (often bed-specific) impact sags, as well as the large volume of accidental lithic and accidental lithic-derived mineral phases suggest a phreatomagmatic and primary origin (HEIKEN 1971, 1972, 1974, WOHLETZ 1983, HEIKEN and WOHLETZ 1991, WHITE 1991a, b, DELLINO et al. 2001, ZIMANOWSKI et al. 2003). Low angle cross bedding and antidune structures are both indicative of traction sedimentation of dilute pyroclastic density currents such as base surge (FISHER and WATERS 1970, FISHER and SCHMINCKE 1994). The pyroclasts have been transported and deposited from base surges around the eruptive centre forming a tuff ring. The moderate to high vesicularity of the volcanic glass shards suggests magma/water interaction during vesiculation of magma. The fact that the number of many blocks of impact sags associated with large clasts are small indicate that the energy of the impacts were suppressed by e.g.

1. high density pyroclastic density current activity, and/or

2. the presence of deeper water or sediment-laden water (slurry) in the depositional environment reducing the impact energy of larger bombs and/or blocks. In contrast the deep sags, caused by small clasts, may have developed in shallower water.

The large amount of accidental lithics and/or mineral phases derived from such pre-volcanic rock units (Neogene) suggest near surface phreatomagmatic fragmentation of uprising melt (LORENZ 1986, 1987, 2003b). The low amount of large accidental lithic bombs indicates that the disrupted pre-volcanic material was unconsolidated with low density, which facilitated an easier breakage during eruption. The presence of siliciclastic lithic and/or siliciclastic-derived mineral phases all suggest that the volcanic eruption occurred in a soft rock environment (LORENZ 2003a, b), thus the pre-volcanic Neogene (Pannonian) sediments must have been still unconsolidated at the time of volcanism at Kis-Somlyó. In addition, effective fragmentation of the magma upon magma/water contact produced fine grained (ash, fine lapilli) clasts instead of large bombs.

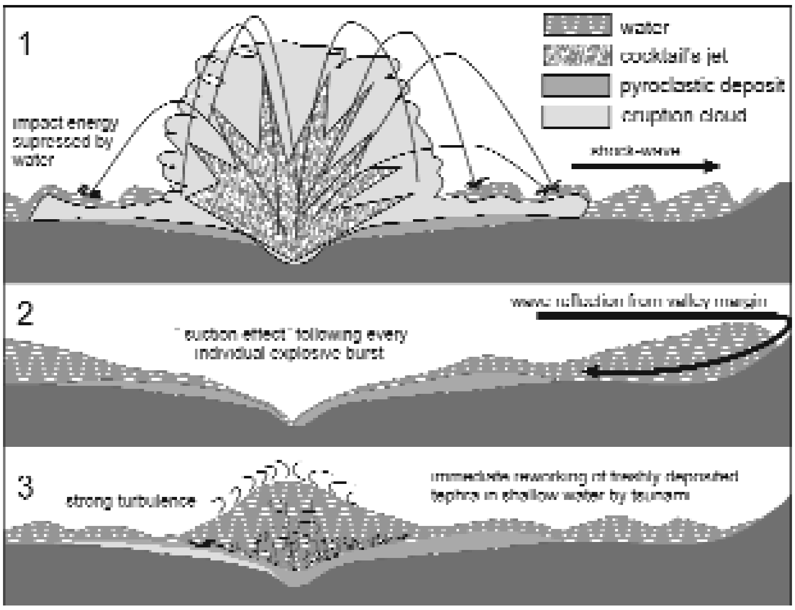

Figure 5.3. Model for the generation of different pyroclastic lithofacies at Kis-Somlyó
1. phreatomagmatic explosive eruption in shallow water produced a pyroclastic mound, 2. eruption advanced and produced shock waves that travelled through the shallow water, and 3. caused immediate reworking of the previously (seconds, minutes) deposited pyroclasts (after MARTIN and NÉMETH 2004a)

In contrast, P2 lapilli tuff beds that are randomly distributed among these primary pyroclastic beds (P1) are inferred to be in situ syn-eruptive reworked tephra, on the basis of the common presence of inverse grading and structure-less texture. The beds comprise abundant abraded lapilli of predominantly juvenile origin, in tabular, undulating or lensoid beds that are interbedded with primary pyroclastic density current deposits (BELOUSOV and BELOUSOVA 2001). The random distribution of primary and reworked tephra beds suggests a depositional environment where particles were free to get remobilized right after initial settling by any type of current movement.

The most favourable environment to generate such deposits is an eruption that occurred in a wet environment, or even in free water (very shallow water). Each individual explosive burst displaced lake water and water saturated mud, transporting pyroclasts radially outward via base surges and building a mound like pyroclastic apron (Figure 5.3). The soft host rock environment allowed the formation of a wide bowl-shaped depression in and around which the pyroclastic deposits have been accumulated. The presence of two major types of lithofacies in the pyroclastic sequence suggests an immediate reworking of freshly deposited tephra forming alternating reworked and primary beds in a non-uniform distribution. The lack of intercalated beds from suspension settling in the pyroclastic units indicates that

1. the volcano erupted only in a very shallow lake and/or
2. there was continuous wave activity during eruption in case of deeper water depth (few tens of metres).

Post-volcanic lacustrine sequence invaded by lava flows

A well-bedded ~5 m-thick, laminated and cross-laminated fine grained yellow, grey siltstone unit overlies the pyroclastic units (Plate 5.4, A, B). The siliciclastic unit is only exposed in the south-western side of the erosional remnant, below the capping lava flow unit and has been interpreted as a lacustrine unit that is truncated by lava flows (JUGOVICS 1915). The areal extent of the siliciclastic unit capping the pyroclastic succession is unclear, however the presence of sandy patches within columnar joints, as well as baked siltstone xenoliths in the lava suggest, that this sedimentary unit likely covered the pyroclastic sequence extensively at Kis-Somlyó. There are low-angle (5–10°) cross-laminated packages in the preserved siliciclastic units. Above these, the sedimentary structures are truncated by invading magma (Plate 5.4, C). This post-volcanic siliciclastic unit differs from the pre-volcanic quartzo-feldspathic units in being richer in oriented muscovite, clay minerals, and in being micro-laminated with 0.2–0.4 mm laminae (Plate 5.4, D). Dark and light coloured laminae form a distinct rhythmic structure (Plate 5.4, D). There is no conclusively identifiable organic material, fossils or pollen. However, scoriaceous, strongly altered volcanic clasts form lenses of basaltoid lapilli and have been identified in the lowermost 30 cm zone of the post-pyroclastic units. The topmost scoriaceous lapilli layer of the pyroclastic sequence is infiltrated by fine mud. Such transition zones extend up to 5 cm in thickness. The contact between the pyroclastic units and the overlying siliciclastic unit is sharp. Contact with the overlying lava is discordant and irregular with brecciated zones of coherent lava and fluidal, highly vesicular detached lava fragments. Subsequent lava flow emplaced onto the post-tuff ring siliciclastic units produced mega-pillows, pillow lobe breccias and peperitic margins along lava and/or dyke margins (MARTIN and NÉMETH 2002). The lava flow forms a semi-circular distribution in map view, with one major rosette-like columnar jointed part, currently forming the highest topographic relief on the volcanic remnant. The thickness of the exposed coherent lava flow changes largely from a few metres to tens of metres toward the centre of the erosional remnant based on sporadic outcrops. Globular peperite occurs beneath the lava flow and formed by pillow-shaped lobes up to 50 cm that penetrated into the wet sediment. Lava mixed with the unconsolidated sediments partly incorporating them. Fluidly shaped clasts, but also some blocky ones, in a wide size-range (cm to metre scale) are dispersed along the margin of the lava flow (MARTIN and NÉMETH 2002). Locally there are also disconnected pillows with near-spherical bulbous shapes, which are detached from the main lava flow supported by the fluidised host sediment (Plate 5.4, C). Up-section the originally laminated lacustrine sediment became homogenised due to intense fluidisation by the intruding magmatic bodies. The new high precision Ar, Ar incremental step heating measurements gave a 4.63±0.03 My age for the lava flows at Kis-Somlyó (WIJBRANS et al. 2004). The closest volcanic remnants (~10 km) where large volume of lava flow has been preserved, it is in a complex lava lake and flow setting is Ság-hegy (Plate 5.1). The new $^{39}Ar/^{40}Ar$ ages suggest that the age of Ság-hegy is significantly older than Kis-Somlyó being 5.48±0.03 My old (WIJBRANS et al. 2004).

The localised, semi-circular distribution of the lava at Kis-Somlyó indicates that its movement was controlled by a semi-circular barrier (e.g. crater rim) and therefore it has been inferred to be accumulated in a volcanic crater of a broad tuff ring. Conversely the post-tuff ring siliciclastic deposits which were invaded by the lava represent crater lake deposits. The origin of the lava from other sources than the Kis-Somlyó volcano is not supported due to the fact, that the nearby volcano that produced a larger amount of lava, is located approximately 10 km distance from Kis-Somlyó. In addition, there are no preserved lava remnants between these two localities. The rapid changes of the estimated thickness of the lava at Kis-Somlyó also point to its local origin.

The presence of laminae and micro-laminae with aligned, platy muscovite flakes in the beds of the post-pyroclastic lacustrine unit indicates suspension settling. Deposition is inferred to have occurred in a water mass with no, or just limited water movement. Water movement is recorded in some non-uniform cross bedding parts in the lower section of the unit which is interpreted to be a result of wave action. More persistent unidirectional cross bedding would be expected if inflowing water had generated the cross bedding. The large amount of muscovite is inferred to have been derived from nearby Neogene sand (Pannonian) ridges. The lack of clear evidence of an in-flowing water course into the post-tephra ring lacustrine system suggests that muscovite may have been transported by wind action and deposited by suspension. The presence of interbedded, coarser grained inverse-to-normal graded laminae or thin beds suggests remobilization by small-scale grain flows, or traction carpets due to turbidity currents. The dominance of the aligned muscovite-rich laminae in this unit supports the interpretation of the wind-blown input of extra-basinal detritus into the sedimentary basin rather than deposition from a sedimentary gravity flow. The presence of volcanic grains in the lowermost sequence of the post-pyroclastic units indicates that the pyroclastic remnant was still unconsolidated by the time a lake developed. Lack of volcanic detritus in the post-volcanic lacustrine beds in higher stratigraphic position indicates that the source area changed. The fact that there is no characteristic facies change in the pyroclastic mound (e.g. primary P1 beds overlain by reworked P2 gravity mass flow deposits rich in basaltic lapilli) suggests that no significant destructive events took place prior to the development of the lacustrine siliciclastic units overlying the pyroclastic beds and/or the original landform was too flat to allow significant mass redistribution into the subsequently developed crater lake. This allows reconstruction of

1. a relatively low-lying original tephra ring into which no, or insignificant detritus was transported or
2. significant erosion of the volcanic succession prior to the lacustrine sedimentation or
3. the preserved part of the volcanic remnant is a distal part of a volcano.

The first interpretation is more likely and therefore it is inferred that either

a) the original tephra ring morphology did not allow enough source material to build up the post-pyroclastic lacustrine deposits in comparison to other sources,

b) the tephra ring was already eroded away and its tephra deposited elsewhere, e.g. outside of the tephra ring by the time the post-volcanic lacustrine sedimentation took place, or

c) a combination of both.

Lack of organic material, fossils or pollen in the post-tuff ring deposits indicate unsuitable conditions for life in the lake, which could have been caused by

1. originally bare landscape and environment,
2. destructive processes of the tuff ring forming event, erasing life in the vicinity of the crater prior to development of the post-pyroclastic lacustrine environment,
3. unsuitable living conditions in the post-pyroclastic lake due to poisonous degassing, alkaline rich water input etc. and/or
4. too short time scale to reinhabit the region after volcanic eruption.

Cross bedding is inferred to have been initiated by

1. hot spring activity,
2. continuous inflow and underflow of stream water into the lake and/or
3. wind action.

Post-volcanic crater lakes with the similar age than Kis-Somlyó volcano have been described in the vicinity of Kis-Somlyó with more than 50 metres of suspension sediments with very similar characteristics to the post-pyroclastic sediments from Kis-Somlyó (Hably and Kvaček 1998). However, these volcanic crater lake deposits are rich in fossils and often record evidence of mesophytic forests in the Pliocene around crater lakes (Hably and Kvaček 1998). palaeobotanical evidence supports the reconstruction of a dry and hot climate in the area of LHPVF (Hably and Kvaček 1998). The surroundings of the craters must have been humid but the climate, in general, was presumably quite dry (Hably and Kvaček 1998). On the basis of this finding, it is likely that large open surface sand ridges may have existed giving substantial source material of wind-blown dust, which was able to deposit in volcanic depressions such as Kis-Somlyó. The lacustrine sedimentation in the crater lake of Kis-Somlyó, based on the presence of ~5 m thick lacustrine unit, took place over a period of a few thousand years. Nevertheless wave action was insignificant in the crater lake which suggests

1. the lake was fairly shielded by valley shoulders,
2. the tuff ring formed a barrier to prevent a significant disturbance in lake water,
3. fine dust was rather a permanent suspension constituents in the air than a direct wind blown blast,
4. the size of the crater lake was relatively small (few kilometres across) and with shallow depth (few metres deep), which would inhibit the generation of large waves.

Ság-hegy tuff ring

Introduction

Ság-hegy is located on a main NW–SE-trending fault zone and forms a complex phreatomagmatic volcano that have been a subject of various geological mapping in the past few decades (Jugovics 1937, Mauritz and Harwood 1937, Kulcsár and Gucyzné Somogyi 1962, Jugovics 1971, Török 1993, Harangi et al. 1994, Harangi and Harangi 1995).

Ság-hegy is a complex volcano located in the central part of the LHPVF (Plate 5.1) consisting of several phreatomagmatic pyroclastic sequences preserved under a thick (~50 m) coherent lava body, which in part has been quarried away (Plate 5.5). Due to the intensive quarrying, the inner part of the coherent lava body has been removed, leaving behind a castle-like architecture of pyroclastic rocks. The outcrop walls thus represent the irregular morphology of a coherent lava body, emplaced in the NW–SE-trending ellipsoidal shaped crater/conduit zone of a phreatomagmatic volcano. Pyroclastic beds in the quarry wall, truncated by oblique dykes to horizontal sills, are inferred to have been fed from a central magma zone. Thin (<10 cm) strongly chilled, black, angularly jointed, aphanitic basaltic lava, mantling the preserved pyroclastic sequence, forms corrugation zones as a consequence of sudden chilling upon contact with the cold and wet phreatomagmatic tephra in the inner wall of the crater of the former tephra ring. These corrugation zones are inferred to be textural feature characteristics for precursor of extensive mixing of lava and host tephra leading to peperite formation along the outer rim of the emplaced lava lake. A whole spectrum of peperite formed along the lava lake margin where fluid oscillation, due to fluidisation of the wet tephra, disrupted the steam envelope around the lava body allowing basaltic magma to invade and mix with the phreatomagmatic tephra. Unconformities in the tephra ring enhanced sill formation fed from the central lava body due to decreased stress, which allowed an easier emplacement.

Out of 7 K/Ar whole rock ages on 2 distinct age groups are recognized: 5.87–5.14 My and 3.46–3.02 My. On additional magnetic fractions 2 isochron ages were calculated and gave ages of 6.27±0.58 My and 3.43±0.61 My (BALOGH et al. 1985, 1986). The great variety of ages obtained by K/Ar method highlights the problem of excess Ar, and some sampling difficulty. Newly obtained $^{39}Ar/^{40}Ar$ geochronology gave an isochron age of 5.42±0.06 My for the Ság-hegy (WIJBRANS et al. 2004).

Phreatomagmatic pyroclastic units of Ság-hegy

The basal pyroclastic series of Ság-hegy comprises weakly to well-bedded, unsorted and poorly graded to normal graded, alternating tuff and lapilli tuff beds (Plate 5.6, A). Soft sediment deformation (Figure 5.4), cross bedding, undulating bedding, accretionary lapilli beds (Plate 5.6, B), and deep, plastically deformed impact sags (Figure 5.5) are common in the upper section of the pyroclastic succession. Juvenile clasts of fine ash to fine lapilli size are predominantly angular sideromelane glass shards of tephrite composition that show no to high vesicularities (Plate 5.6, C). Vesicular sideromelane shards tend to be stretched and slightly fluidal, both showing intense palagonitisation. A variable amount of tachylite shards are present but are less common than sideromelane. Juvenile lithics are rare, and predominantly microgabbroid textured mafic rock fragments up to coarse lapilli size (Plate 5.7, A). Accidental lithic clasts (<5 cm in diameter) are predominantly derived from the Late Miocene fluvio-lacustrine units immediately underlying the volcanic sequence. They often appear clot-like, plastically deformed, fragments or as single crystals (Plate 5.7, B). Large (cm-to-dm-scale) zones of fines enriched, mica-rich, irregular shaped clots are common, especially in medium bedded lapilli tuffs in the lower pyroclastic strata. Such beds have been described earlier as lake beds inter-bedded with the pyroclastic succession and providing evidence for the interpretation that the volcanism and the immediate pre-volcanic lacustrine sedimentation are coeval (JUGOVICS 1915, KULCSÁR and GUCYZNÉ SOMOGYI 1962). Recent studies demonstrate that the beds contain volcanic glass shards and their bed-forms are more characteristic of base surges generated during phreatomagmatic explosive eruptions (MARTIN and NÉMETH 2004b). In micro-scale, similar mica-enriched clots are also common in most of the fine matrix supported lapilli tuffs, as well as in the fine tuffs. Large intact sand, silt and mudstone clasts are prominent and often form bed-flattened, strongly elongated irregularly shaped clasts in the lapilli tuff beds (Plate 5.7, C). In association with certain beds, these siliciclastic clasts show intense heat alteration such as hematite enrichment, and mud crack-like radial joints. The deepest exposed stratigraphic level comprises thickly bedded, structureless or weakly stratified, accidental lithic-rich lapilli tuffs and/or tuff breccias (Figure 5.6). A gradual improvement in bedding is obvious up-section. High up (~50 m above the deepest exposure level of the pyroclastic units) in the accidental lithic-rich pyroclastic sequence, well and thinly-bedded, unsorted, accretionary lapilli- and/or

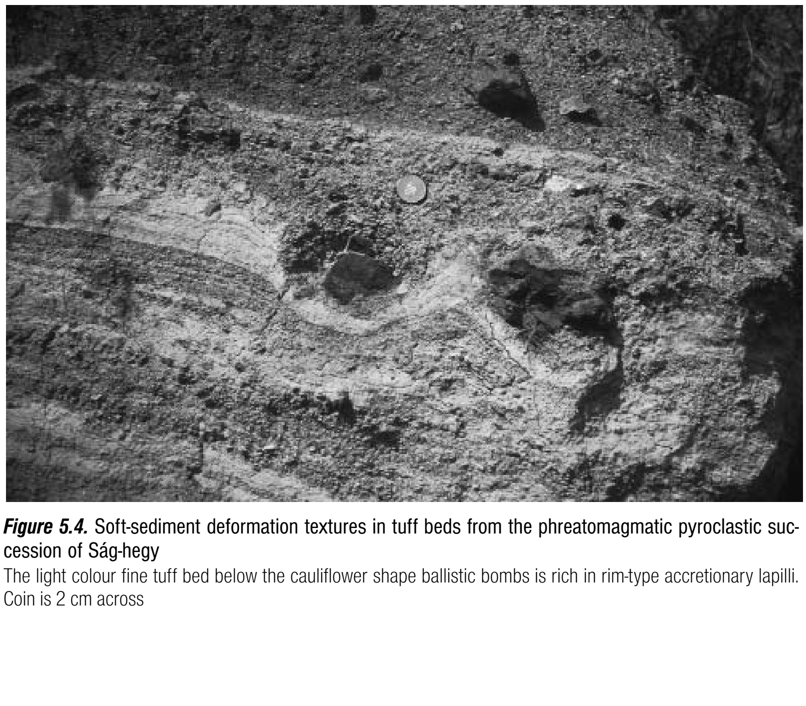

Figure 5.4. Soft-sediment deformation textures in tuff beds from the phreatomagmatic pyroclastic succession of Ság-hegy
The light colour fine tuff bed below the cauliflower shape ballistic bombs is rich in rim-type accretionary lapilli. Coin is 2 cm across

armoured lapilli-rich, and Neogene sediment-derived mineral phase-rich tuff and lapilli tuff are more prominent. Beds in the upper section are also richer in bomb sags, scour fills, vesiculated tuff layers, soft sediment slumping, dish structures and irregular lower bed contacts.

Interpretation

The features above suggest that the pyroclastic units at Ság-hegy resulted from phreatomagmatic eruptions generated due to interaction of rising basaltic magma and water-saturated unconsolidated sediments (HEIKEN 1971, LORENZ 1974, 1986, WHITE 1989, 1990, 1991a). The pyroclastic units are inferred to have been deposited by alternating base surges and fall-out, which gradually built an initial tephra ring around the erupting vent(s). The large amount (often over 50 vol.% by visual estimate) of accidental lithic clasts and especially the mineral phases characteristic of the Neogene fluvio-lacustrine units indicate that the magma water interaction was predominantly driven by ground water stored in the porous media aquifer, likely in the near surface. The eruptions must have taken place in soft sediment (LORENZ 2003b), otherwise intact country rocks would have been a more common constituent in the tephra, as has been reported from Turkey (GEVREK and KAZANÇI 2000) or Mexico (ARANDA-GOMEZ and LUHR 1996). Only there are rare accidental lithic fragments, which were derived from other than Neogene sedimentary units. This fact and the general paucity of intact Neogene accidental lithic fragments indicate that the phreatomagmatic explosions were driven by surface or near-surface water sources. The explosion took place in the uppermost, still unconsolidated and wet, water-saturated Neogene sediments, which is a characteristic eruption style of tuff rings that may produce extreme wide craters surrounded with flat rims according to the water availability and the eruption rate and duration (HEIKEN 1971, MCCLINTOCK and WHITE 2000, WHITE and MCCLINTOCK 2001). The lower massive part of the accidental fragment-rich unit suggests deposition from high concentration, laminar gravity-driven mass flows such as volcanic debris flows (SMITH and LOWE 1991). The angular, ragged and irregular shape of the juvenile pyroclasts, as well as the presence of the chilled glassy pyroclasts such as volcanic glass and/or glassy juvenile fragments with low to moderate vesicularity indicate a primary, eruption-fed origin for these deposits. The sub-horizontal bedding characteristics and the abundance of coarse lapilli and block size juvenile fragments are suggestive of deposition from pyroclastic density currents in a near vent setting. The abrupt textural change of the pyroclastic units in the upper and lower part may indicate changes in the eruptive environment from shallow subaqueous to subaerial (SOHN and CHOUGH 1992, WHITE 1996, 2000, 2001, SCHMINCKE et al. 1997, WHITE and HOUGHTON 2000).

Figure 5.5. Deep impact sag, caused by a cauliflower bomb in the upper pyroclastic succession of Ság-hegy Note the fault that cuts the pyroclastic succession

Figure 5.6. Thickly bedded pyroclastic breccia, coarse lapilli tuff in the basal pyroclastic succession of the Ság-hegy

Contact zone between solidified lava lake and tuff ring units

Due to the intensive quarrying at Ság-hegy in the past decades, the inner part of coherent lava body of the former phreatomagmatic volcanic complex has been completely removed, leaving behind a castle-like architecture of pyroclastic rocks. The outcrop walls thus represent more or less the irregular morphology of a coherent lava body emplaced in the crater of a phreatomagmatic volcano. The quarry walls of pyroclastic bedsare truncated by oblique to horizontal coherent lava layers (Figure 5.7 and Plate 5.7, D), which are inferred to have been connected with a central magma zone (presumably to the already quarried lava) through narrow (cm-to-dm-scale) lava necks. There are large areas (tens of m^2), where thin (<10 cm) strongly chilled, black, angularly jointed aphanitic basaltic lava mantles the preserved pyroclastic sequence (Plate 5.8, A), that are interpreted to be the irregular contact zone between the inner crater wall of the tuff ring and the emplaced lava lake. The contact zone of the coherent lava body with the pyroclastic rocks

Figure 5.7. The sills (s) of Ság-hegy are preferentially intruded along the unconformity of the tuff ring sequence

is undulatory and resemble a 'shark skin' or ´elephant hide' texture (Plate 5.8, A). These chilled contact zones in places seem to cover entire quarry walls, giving an impression that they are embedded coherent lava flows in the pyroclastic sequence. There are mm-to-cm-scale undulations with a wavelength in a cm-to-dm-scale, forming a skin-like lava crust instead of sub-horizontal lava units. These zones, commonly form closely spaced ridge like features in a cm-to-dm-scale. However, in places they rather resemble turtle shell structure. In close view these corrugation zones form closely spaced and very irregularly shaped basaltic protrusions. This can be examined in hand-specimen-scale on oriented samples containing both the lava crust and the neighbouring host lapilli tuff. Similar protrusions have been reported from the Peninsula Tuff Cone, California (LAVINÉ and AALTO 2002). At Ság-hegy, these corrugated zones often feed centimetre-to-metre-scale, straight or slightly twisted protrusions (Plate 5.8, B). These small sills are commonly sub-horizontal, but are not necessarily intruded along bedding planes. There are also sills that are gradually brecciated and form globular peperitic zones in a cm-to-dm-scale along their margins. Peperitic zones between the intruding small sills and the host pyroclastic sediments are randomly distributed and have no characteristic distribution pattern in the sense of stratigraphic elevation of a particular sill. Microphenocrysts of plagioclase and pyroxene are mostly aligned parallel to the sill margin, and are less abundant along the margin. Similar patterns exist in the main lava lake body. Due to the quarrying, however, only the chilled margin is preserved commonly which is mostly vesicle and microphenocryst free.

Feeder dykes intruded and thus crosscut the phreatomagmatic deposits at Ság-hegy, often forming peperite along their margins. Peperitic margins along sills and oblique dykes are more prominent in the lower section of the pyroclastic units suggesting its water-saturated and unconsolidated conditions (Plate 5.7, D). In contrast, in the upper section the intrusions often caused reverse faulting of the pyroclastic units (Plate 5.8, C) indicative of its drier, more rigid state during intrusion. However, there are also peperitic margins along the contact of the lava lake and tuff ring at the topmost pyroclastic sequence, which is inferred to represent the subaerial tuff ring units. This peperitic zone is thinner and sharper than peperitic sill and/or dyke margins in lower stratigraphic positions, and is interpreted to be the result of exhalation of magmatic gases and vapour that both together may have promoted the formation of these peperites. In the reverse fault planes the intrusion contacts toward the pyroclastic units are generally sharp, however, cm-scale undulations, boudinage-like structures are characteristic. These intrusions form narrow globular peperitic margins with narrow (cm-to-dm-scale) dispersed peperite zones. They are surrounded by finely dispersed, slightly oriented, homogeneous, and fluidised haloes, which are often channelled far away (m-scale) from the tip of the intrusion along the reverse fault plane(Plate 5.9, A).

All the coherent lava bodies at Ság-hegy, especially those which were involved in the peperite forming process, have a cm-to-m-scale strongly fluidised zone forming a whitish, fine-enriched sediment halo around them (Plate 5.9, B). This halo is generally easy to distinguish from the undisturbed lapilli tuff and tuff by its lighter colour and lack of sedimentary structures. Commonly, this halo is narrow (cm-scale) in places where globular peperite has developed, whereas the halo reaches meter size in areas where blocky peperite is apparent, indicating that these zones may have functioned as pathways to disperse lava clasts deep into the host tephra via clastic dykes (MARTIN and NÉMETH 2004b). These zones may represent areas where the steam envelope has been disrupted and direct contact between cool water-rich sediment and the hot melt led to suppressed phreatomagmatic disruption of the sill along the sill margins (Figure 5.8).

Intrusion along unconformities in the tephra ring sequence may have enhanced sill formation fed from the central, large-volume lava body that gradu-

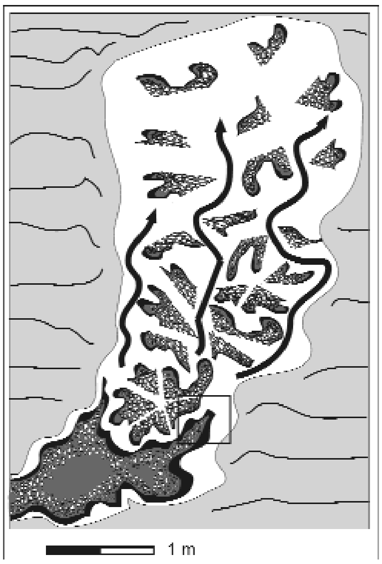

Figure 5.8. Formation of a fines enriched halo along intrusions and their relationship with different peperite textures as a result of phreatomagmatic explosive disruption of the dyke tip due to breakage of the steam envelope. Lines show original bedding. In rectangule see the still intact dyke/sill tip and the already fragmented clast. Arrows show the fluid movement through the fluidised host tephra

ally filled the crater due to decreased stress, which allowed an easier emplacement (MARTIN and NÉMETH 2004b). The lava-lake fed sills have jagged and brecciated margins and intrusion occurred preferentially along unconformities of any type in the tuff ring sequence (MARTIN and NÉMETH 2004b).

Conclusion

At Ság-hegy, there is a clear transition from irregular margins of sills or dykes (globular peperite) to disrupted, angular shaped (but originally globular fluidal) clasts or blocks (blocky peperite – MARTIN and NÉMETH 2004b). The mixed appearance of globular and blocky peperite at the same location indicates a change in fragmentation and mixing mechanism of host and intruding magma body during magma-wet sediment interaction (BUSBY-SPERA and WHITE 1987, KANO 1989, GOTO and MCPHIE 1996, HANSON and HARGROVE 1999, DOYLE 2000, DADD and VAN WAGONER 2002, HOOTEN and ORT 2002, MARTIN and WHITE 2002, MCCLINTOCK and WHITE 2002, SKILLING et al. 2002, WOHLETZ 2002, MARTIN and NÉMETH 2004b). Intrusion along unconformities in the tuff ring sequence may have enhanced sill formation due to decreased stress, which allowed an easier emplacement. The initial magma fragmentation and mixing with sediment is interpreted to have been the result of tearing apart of magma and shaping of the magma-sediment interface into globular, pillow-shaped bodies by contact-surface interaction. During a second stage blocky peperite along the sill, as well as along the lava lake margin was formed by phreatomagmatic events during breakdown of insulating vapour films at the sediment-magma interface The presence of peperitic zones as well as the whitish, strongly fluidised halo along the entire lava lake and along all the sills derived from the lava lake indicates that pore water was easily remobilised from the host tephra due to the heat of the lava. The relatively homogeneous distribution of the peperitic margins along the intrusive bodies regardless of their stratigraphic position indicates that the weight of the lava lake was not large enough to suppress pore fluid oscillation in the basal and marginal zone of the lake. Conditions for peperite formation at Ság-hegy are inferred not to have been favourable for phreatomagmatic explosive disruption because

1. the water content of the tephra was insufficient to fuel highly efficient phreatomagmatic disruption, and
2. the magma discharge rate was relatively high.

The latter caused large enough magmatic pressure on the inner crater wall to suppress larger-scale explosive disruption. In the upper stratigraphic level, the common brittle-fragmented pyroclastic units indicate drier conditions during lava lake-fed sill emplacement. Because phreatomagmatic tephra dries out relatively quickly (days to weeks), and because there is no indication of disruption in the lava lake emplacement, it is a plausible interpretation that the lava lake emplacement took place in one major event, and in shorter time than the time required to partially dry out the tephra (MARTIN and NÉMETH 2004b). The tephra, however, after deposition would have remained wet longer in the deeper stratigraphic level at Ság-hegy, because the base of the volcano apparently developed in a shallow standing water body (MARTIN and NÉMETH 2004b).

Similarly, there are irregular contact zones of lava lakes emplaced in the crater of Plio/Pleistocene phreatomagmatic volcanoes in southern Slovakia (KONEČNÝ et al. 1995, 1999, KONEČNÝ and LEXA 2000) and along intrusive bodies emplaced into maar crater filling lacustrine units from phreatomagmatic volcanoes of the Eger rift (Germany – SUHR and GOTH 1996).

Egyházaskesző tuff ring

Introduction

About 15 km north-east of Ság-hegy small dissected mafic pyroclastic rocks crop out, forming a well-defined region near the River Marcal, along the major tectonic line of the LHPVF (Rába Fault Zone – Plate 5.1). The exposed pyroclastic rocks form flat-topped hills that are a few tens of metres above the base level of the Little Hungarian Plain. They are inferred to have been partially covered by Quaternary terrestrial sediments (gravel and sand beds). In spite of the poor exposures in this region, they have been known for a long time (JUGOVICS 1915). They were the subject of geophysical investigations (TÓTH 1994) that revealed significant reservoirs of alginite and basaltic bentonite associated with dish-like structures inferred to be craters of tuff rings (SOLTI 1987, TÓTH 1994). Between Várkesző and Egyházaskesző villages (Figure 5.9) in a small area of about 0.3 km^2, in a 75 m deep volcanic depression a basal alginite (30 m) is covered by a 42 m thick basaltic bentonite (SOLTI 1987). Just south-west of Egyházaskesző village below a thin (less than a metre) soil, a 36.6 m thick basaltic bentonite and 4 m thick alginite succession have been identified that form a crater filling sedimentary unit (SOLTI 1987). Between these two volcanic depressions pyroclastic rocks are known in small outcrops that have been used for community purposes. Among many pyroclastic rock pits, one is still in use today in the village of Egyházaskesző (Figure 5.9).

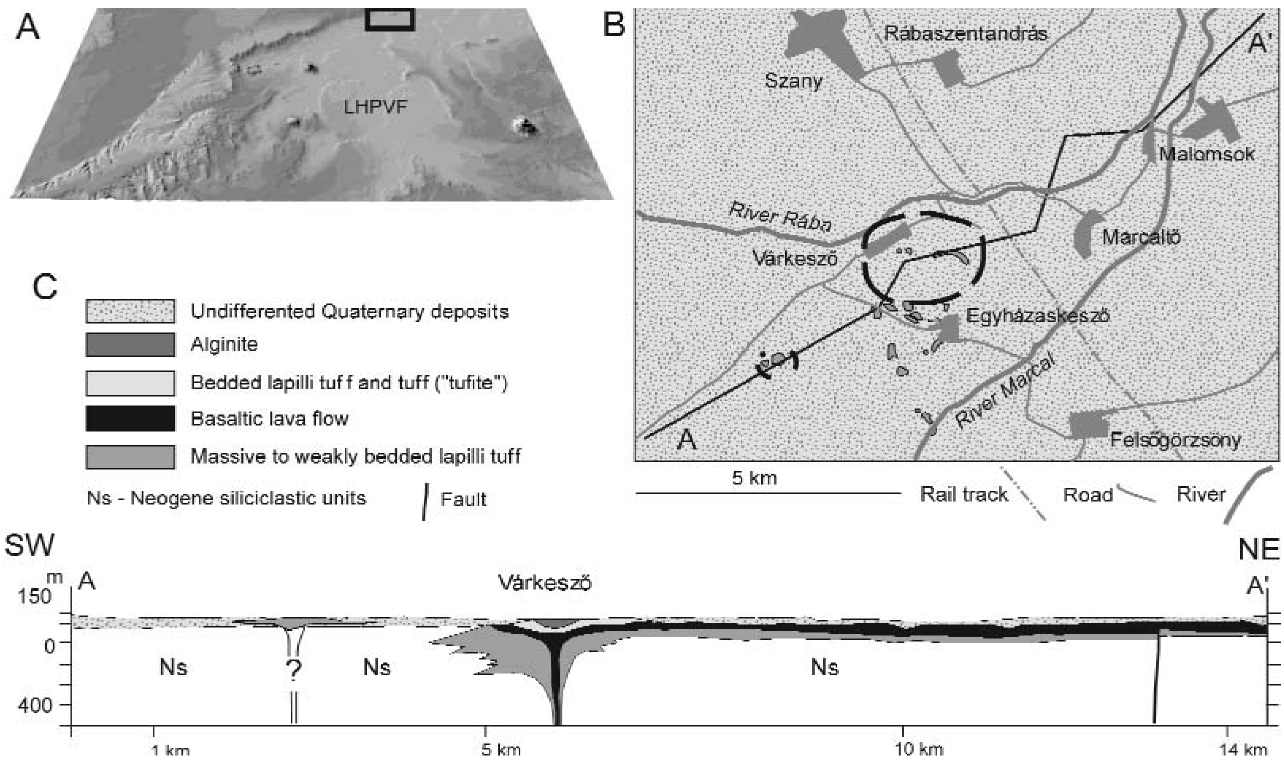

Figure 5.9. Location maps (A, B) of double tuff ring between Várkesző and Egyházaskesző villages. [Geological data are after BENCE et al. 1978] Cross-section (C) is predominantly based on shallow drill cores. Dashed line circles on B locate the tuff rings

Pyroclastic succession

In the active quarry of Egyházaskesző the pyroclastic succession (Figure 5.9, C) dips toward the north at a low angle (5–15°). The dip of the bedding of the pyroclastic rocks in all of the outcrops is low, however, it varies in direction (JUGOVICS 1915). The steepest dipping angles have been recorded in the northern limit of Egyházaskesző village, nearby the inferred crater rim (JUGOVICS 1915, SOLTI 1986, 1987). Coherent lava fragments of large vesicular to dense basaltoid lapilli and blocks (JUGOVICS 1915) are either in the lapilli tuff or they are weathered out and accumulated as in situ debris. The lapilli tuff and tuff succession is monotonous; neither significant unconformity nor textural changes of the rocks in terms of composition, grain size, or colour has been recognised. The lapilli tuff is rich in a fine muddy matrix that contains large amounts of muscovite flakes, quartz grains and other clay minerals inferred to have been derived from the immediate pre-volcanic siliciclastic units (Plate 5.9, D). The juvenile fragments are volcanic glass shards (Plate 5.10, A) that are characteristic products from phreatomagmatic explosive eruptions. The glass shards are moderately to strongly palagonitized (Plate 5.10, A). The lapilli tuffs often contain small pebbles as accidental lithic clasts picked up during eruption. In the active quarry of Egyházaskesző, large bedding planes are exposed (Plate 5.10, B). The exposed bedding planes of the lapilli tuff form low amplitude long wave length undulating beds (Plate 5.10, B), similar to those that have been interpreted from geophysical sections in subsurface tuff ring beds (CAGNOLI and ULRYCH 2001a, b). On the surface of these exposed bedding planes, asymmetric impact craters up to 40 cm can be identified that are similar to impact craters on young maar/tuff rings such us Ubehebe California (Plate 5.10, C, D). The impact craters are shallow (cm-to-dm-scale) and angular volcanic lithic fragments and/or cauliflower bombs are still preserved within. On the basis of the asymmetry of these impact craters a north to south transportation can be concluded (Figure 5.9). In the active quarry of Egyházaskesző fine lapilli tuff and tuff beds are commonly rich in rim-type accretionary lapilli (up to cm size) and/or clot-like features formed by mud similar to those reported from phreatomagmatic volcanic remnants of the Hopi Buttes, Arizona (WHITE 1991b).

In the outcrops of an abandoned quarry south of Egyházaskesző village a pyroclastic succession similar to the one described above is preserved (Figure 5.9). In this locality the pyroclastic rocks form a gentle (<10°) south-westward dipping succession a few tens of metres thick. There are neither large lithic fragments and impact sags nor accretionary lapilli beds in this locality. The pyroclastic succession is covered by a few dm thick Quaternary gravel bed.

Interpretation

The textural characteristics of the pyroclastic outcrops around Egyházaskesző suggest that near surface magma/water interaction caused phreatomagmatic explosive eruptions, which formed broad and flat tuff rings. The bedding characteristics (Plate 5.10, B) suggest deposition from base surges (FISHER and WATERS 1970, WATERS and FISHER 1971, CHOUGH and SOHN 1990, BULL and CAS 2000, COLE et al. 2001). The geometrical parameters of the original landforms of these tuff rings are inferred to be similar to those reported from Oregon from a former lake basin with a 1:20 crater-rim-height/volcano width ratio (HEIKEN 1971). The large proportion of mud and silt, as well as mineral phases derived from these rock units indicate a shallow level of magma/water interaction and the formation of tephra rings. After formation of the tephra rings, their craters were flooded by water, giving way to crater lake deposition and the formation of alginite, as well as basaltic bentonite. In the Quaternary, the tuff rings were covered by young alluvial plain deposits, which were stripped away by recent erosional processes. It is inferred that this is a result of vertical movements and associated changes of a depositional versus erosional regime, controlled by the base level changes of the region.

Gérce–Sitke tuff ring

Introduction

Gérce–Sitke is a double hill north-west of the Ság-hegy tuff ring (Plate 5.1) and consists of a skeleton like tuff ring structure (Figure 5.10). The volcanic origin and structure of eroded tuff rings in this area was recognised by JUGOVICS (1915).

The eastern group of small hills forming a castle-like group east of the village of Gérce (Plate 5.1, A) stands just ~50 metres above the basement floor of the LHPVF (Plate 5.1). Pyroclastic rocks crop out in each of these hills with a bedding dip direction toward a small (1 km wide) depression between the hills.

Figure 5.10. Location map of the Gérce-Sitke tuff ring remnants (A) and their surroundings (B)
On the aerial photograph (C – Hungarian Military Photo Collection) the castle like architecture of the erosional remnants of the Gérce–Sitke tuff rings is clearly recognisable. White circle represents the location of the alginite quarry, the black rectangle shows a section where the tuff ring rim pyroclastic rocks are well-exposed

A ring of small hills that are built up by gentle dipping lapilli tuff beds are located between the villages of Gérce and Sitke (Plate 5.1 and Figure 5.10) called Hercseg-hegy. The top of these hills reach the same ~180 m elevation as the hill tops of the Gérce pyroclastic rocks. At Hercseg-hegy, four distinct hills can be identified each standing slightly above the pyroclastic succession and composed of rosette-like columnar jointed alkaline basaltic lava buds.

Pyroclastic succession

The pyroclastic rocks of these hills are well-bedded, unsorted, and commonly inverse graded, 15–25 cm thick lapilli tuff beds with mm-to-cm thick tuff couplets. The coarser grained lapilli tuff beds are commonly calcite cemented and rich in abraded xeno- and/or pyrogenic crystals such as olivine or clinopyroxene. Crystal distribution of the lapilli tuff beds are equal regardless of the stratigraphic position of the pyroclastic rocks, however, their grain size may vary randomly. The lapilli tuff is rich in blocky to moderately elongate, moderately vesicular sideromelane glass shards that suggest magma–water interaction during phreatomagmatic explosive eruptions of their source volcano. Mud, as well as mineral phases derived from fine-grained siliciclastic deposits are common in the matrix of these pyroclastic rocks, however, matrix poor, volcanic glass shard rich calcite cemented rocks are also common at this site. The origin of the calcite cement, and the interpretation of the clast-supported nature of a few of these lapilli tuff beds are the subject of current research.

The pyroclastic rocks of the hills between Gérce and Sitke, dip towards the hill centre in a radial arrangement. The lapilli tuff of the hills between Gérce and Sitke are richer in matrix than those exposed east of Sitke. The pyroclastic rocks of Hercseg-hegy are mud, silt and muscovite rich. Sideromelane lapilli are commonly strongly altered to palagonite and show reddish staining. They are moderately vesicular, and commonly form fluidal, but irregular shaped clasts. Peridotite lherzolite xenoliths up to 35 cm in diameter are common. The pyroclastic rocks often contain vesicular, ribbon like lava bombs, with mud filled vesicles. The tuff ring crater toward the west was gradually filled by alginite, which is quarried today. The alginite beds here reach a few tens of metres, indicating a volcanic depression, where lacustrine sedimentation took place.

Interpretation

It is inferred, that the hills east of Gérce, are exposed tuff ring skeletons that formed during shallow surface magma/water interaction and formed tephra rings surrounding shallow maar craters.

The origin of the Hercseg-hegy is probably similar to the pyroclastic rocks east of Gérce. The difference in character is inferred to be the depth of magma/water interaction, and the possible choking of the vent at Hercseg-hegy by mud.

Somló butte

Introduction

Somló is a large butte (~1 km across) capped by basaltic lava in the south-eastern margin of the LHPVF (Plate 5.1). The hill is as high as the largest buttes of the BBHVF such as the Badacsony, or Szent György-hegy (Plate 5.11, B). On the flank of the Somló the immediate pre-volcanic Neogene siliciclastic rock units crop out up to 250 m in elevation. In the southern flank of the butte, dark grey basaltoid lava flows are inferred to cover the pre-volcanic siliciclastic rocks. There is no field evidence to support the existence of pyroclastic rocks below the coherent lava body in this side of the butte. The coherent lava body is platy jointed in the basal zone of the flow, but thickly columnar jointed in the upper part. The total thickness of the coherent lava body is at least 70 m in the southern part of the butte. The coherent lava body forms a smooth surface plateau around an elevation of 350 metres. This plateau is capped by a gentle sloping hill reaching over 400 m today. This capping hill consists of a reddish, scoriaceous halo that can be mapped in the field by the colour of the soil. In the northern side of the butte, around 250 m in elevation, just below the lava pyroclastic rocks crop out that are inferred to be around 50 m in thickness. There is no outcrop of the pyroclastic succession here, because of the thick Quaternary rock fall unit flanking to the coherent lava body.

Pyroclastic succession

In small outcrops and in situ debris, grey to yellowish lapilli tuff and grey tuff have been recovered. In the small dm-to-m scale poor outcrops, as well as handspecimen size samples, poor bedding, and grading have been recognised. The pyroclastic rocks are rich in muscovite flakes, quartz, small pebbles and small irregularly shaped mud- and siltstone clasts. In the studied samples, no deep seated xenoliths have been identified yet. The lapilli tuff and tuff samples are fine, muddy and matrix supported (Plate 5.11, C). The volcanic glass shards are moderately vesicular and blocky (Plate 5.11, D). The glass shards are often reddish, and an advanced stage of palagonitisation is common at the rim and along fractures of the shards (Plate 5.11, D). In the topmost section of the Somló pyroclastic rocks consists of black scoria rich lapilli tuff seemingly overlies the lava plateau. These scoriaceous lapilli tuff beds are also rich in siliciclastic clasts, which are inferred to have been derived from the Neogene successions. The exact 3D relationship between the scoriaceous pyroclastic beds and the lava plateau is not yet clear. According to the geographical distribution of the scoriaceous beds, it is possible that their formation post-dates the lava flow emplacement, and thus that they are similar to those capping scoria cones described from Badacsony, Szent György-hegy or Agár-tető. However, the lava flows may have been derived from a scoria cone in the centre of the Somló may have breached the wall of the cone and flowed around the cone.

Interpretation

For the eruption mechanism of the Somló a shallow level magma–water interaction is suggested on the basis of the basal fine grained, accidental lithic-rich (or mineral phases derived from those rocks) tephra, moderately vesicular sideromelane glass shards, and the advanced palagonitization of glass shards. The unsorted and matrix supported texture of the basal pyroclastic rocks indicates that these rocks are the results of diagenised deposits transported and deposited predominantly by base surges. The lack of ballistic bombs as judged from the poor outcrops and the finely dispersed quartzofeldspatic matrix of the pyroclastic rocks indicate that the phreatomagmatic explosion occurred in a soft rock environment.

Erosional remnants of phreatomagmatic volcanoes from Austria and Slovenia

Introduction

Mio/Pliocene alkaline basaltic rocks are also known from deeply eroded outcrops in Austria and Slovenia, near the Hungarian state border. The Austrian examples are often referred to as the Styrian Basin Volcanic Field. This volcanic field extends with few very poor outcrops into Slovenia, close to the triple border of Austria, Hungary and Slovenia. In the northernmost areas of the Styrian Basin Volcanic Field there are basanite lava flows that erupted on an erosional surface of a metamorphic core complex, which forms the basement just 200 km west at the LHPVF. The best known locality of this region is the Pauliberg (Pál-hegy – JUGOVICS 1915, 1939). No pyroclastic rocks are known from this locality and no detailed study of the origin and nature of the basanitic lava is available yet. A group of hills around the town of Güssing (Németújvár) standing about 100 m above the local base level form a similar morphology to that, which is common in the BBHVF. The castle hill of Güssing is entirely built up by pyroclastic rocks and its similarity to diatremes of the western Hungary was recognised a long time ago (JUGOVICS 1915). The pyroclastic rock unit is underlain by the similar Neogene siliciclastic fluvio-lacustrine rock succession known elsewhere in the western Pannonian Basin. The contact between the pyroclastic and pre-volcanic sand, silt and gravel beds is at an elevation of about 250 m, just ~50 m above the base level of the Quaternary deposit filled valley in the surrounding areas.

Pyroclastic succession

The pyroclastic hill at Güssing reaches 310 m in elevation, and is formed by gently to moderately (5–25°) dipping pyroclastic beds (Plate 5.12, A) that dip always inward to the centre of the hill (however irregularities can be measured in small outcrops where the castle stands – JUGOVICS 1915). Dip values are steeper in the marginal zones of the pyroclastic succession than in the centre (top) of the hill (Plate 5.12, B). The pyroclastic rocks of Güssing show more matrix supported characteristics in the basal areas, and a gradual increase in glassy pyroclasts and xenocrysts and pyrogenic minerals up-section. In spite of these variations, the pyroclastic rocks are very similar with regard to components and bedding characteristics across the entire erosional remnant. The matrix of the lapilli tuff beds is rich in mud and silt, and lapillus sized irregular shaped mud chunks (Plate 5.12, C). In the muddy, silt-rich matrix of the rock, a great variety of volcanic glass shards has been recognised (Plate 5.12, C).

Near Güssing, small hills have been reported to host pyroclastic rocks such as the Bindergberg (Kálvária-hegy) near Tobaj (JUGOVICS 1915). From this location amphibole crystals can be collected from the soil topping the hill, but no outcrop is known that would allow study the texture of the possible source pyroclastic rocks. The hill itself is an insignificant mound standing a few tens of metres above the valley of the Strem creek, just ~6 km toward north-west of Güssing. Further north-west, around Limbach (Hárspatak) a small group of hills form an irregular surface of mounds of pyroclastic rocks. The pyroclastic rocks are only exposed in very poor outcrops. These are strongly palagonitized, however, moderately vesicular sideromelane glass shards and a large volume of muddy, silt-rich matrix can be recognised. Some poorly preserved tree trunks have been reported from the pyroclastic rocks earlier, however, their botanical identification has not been possible yet (JUGOVICS 1915).

A very similar occurrence of lapilli tuff has been reported from Slovenia (KRALJ 2000a). Pliocene volcanic rocks have been reported near the village of Grad (Slovenia), accumulated over an alluvial fan (KRALJ 2000a). On the basis of poor outcrops in the region a preliminary model on the formation of the Pliocene volcanic rocks at Grad is given by KRALJ (2000a). An early development of a scoria cone and associated lava flow(s) is inferred, that were destroyed by subsequent phreatomagmatic explosive eruption(s) (KRALJ 2000a). The preserved pyroclastic rocks in the region are diverse in texture, exhibit matrix rich and fines-depleted lapilli tuffs, and are often rich in scoria (KRALJ 2000a). Accretionary lapilli have been reported from fine grained lapilli tuffs indicating phreatomagmatic eruptions (KRALJ 2000a, b).

Interpretation

The presence of the sideromelane glass shards with variable vesicularity and shape parameters indicate a near-surface phreatomagmatic fragmentation of the uprising basaltoid magma leading to the formation of a tephra ring. The circularly inward dipping pyroclastic beds with common reworked textural characteristics such as inverse grading indicate deposition from grain flows. Abraded clasts, bed couplets of thick coarse and thin fines enriched beds, as well as the common presence of free, broken pyrogenic minerals and/or xenocrysts have been interpreted as a result of early remobilisation of tephra, probably during syn-eruption time. The textural characteristics of the pyroclastic succession of Güssing is remarkably similar to those that have been described from Szigliget (e.g. Várhegy – NÉMETH et al. 2000) and or observed from the South Slovakian Pliocene alkaline basaltic fields, e.g. from Filakovo (Fülek – JUGOVICS 1948, KONEČNÝ, et al. 1995, 1999, KONEČNÝ and LEXA 2000). The origin for such pyroclastic rocks is suggested to be the direct result of syn-eruptive remobilization on a flank (inner and/or outer) of a growing pyroclastic cone (NÉMETH et al. 2003). However, post-eruptive reworking of tephra in a volcanic crater and/or conduit zone may also be a reasonable interpretation of such volcanic texture, but further investigation required to improve these models.

The texture of the pyroclastic rocks of the Bindergberg succession suggests that this locality is also an erosional remnant of a former phreatomagmatic volcano that very likely erupted in a fluvial valley filled with wet siliciclastic sediments. Due to post-volcanic erosion, the former tuff rings were quickly destroyed, leaving behind a mound like pyroclastic succession. The pyroclastic mounds were probably repeatedly covered and exhumed since the volcanism ended leaving behind a small veneer of pyroclastic rocks.

Large lava flows in alluvial debris flow deposits associated with the erosional flank of the volcanic edifices near Grad have been interpreted as clasts eroded from former peperite units (KRALJ 2000a). However, this conclusion cannot be supported unless a clear demonstration of a direct relationship with the magmatic body and the host sediment is demonstrated (GOTO and MCPHIE 1996, WHITE et al. 2000, SKILLING et al. 2002). Overall, the examples from near Grad (Slovenia) and near Limbach (Austria) suggest that the original volcanic edifices were destroyed by fluvial processes, and their erosional debris was redistributed into debris flows of the alluvial fans. Such a process is widespread (RIGGS et al. 1997) and has even been observed in recently erupted phreatomagmatic volcanoes in such settings such as in Rininahue, Chile (MUELLER and VEYL 1956). The preservation potential of phreatomagmatic volcanic fields in terrestrial settings is generally poor (UFNAR et al. 1995), and the volcanic deposits of such volcanoes are expected to be preserved only under special circumstances such as quick burial processes.

The study of the Pliocene alkaline basaltic intraplate volcanoes from the western margin of the Pannonian Basin is far from complete, and is the subject of much ongoing research and is also a promising area to identify phreatomagmatic volcanoes (POSCHL 1991).

Conclusion

In shallow water, small-volume basaltic explosive volcanic eruptions form cones, rings, or mounds consisting of bedded pyroclastic deposits that are formed by fall out, density currents and/or down-slope remobilization of tephra to the water level or above (FISHER and SCHMINCKE 1984). The formation of volcanic fields such as those in the LHPVF, are often related to phreatomagmatism generated by shallow or deep groundwater sources, where often seasonal climatic changes, as well as the availability of surface and groundwater play an important role in the evolution of volcanic landforms (NÉMETH et al. 2001). This suggests that in low lands, a great variety of volcanic landforms can develop depending on the status of the hydrological environment during eruptions. Attempts to characterise the depositional palaeoenvironment of a volcanic field, especially in well-drained low-lying areas are not common, but volcanic fields often seem to occur in such settings (HAMILTON and MYERS 1962, HEIKEN 1971, GODCHAUX et al. 1992, ORT et al. 1998). In basin-like settings, water surplus is likely to occur in periods of higher rainfall, so seasonality may influence the water availability of these regions. Studies of volcano remnants and their volcaniclastic sedimentary records, facies relationships between pre-, syn- and post-volcanic rock formations, as well as the reconstruction of the eruptive environment, may give vital information of the dynamics of the palaeoenvironment where intracontinental volcanoes have erupted. Examples of eroded tuff rings from the LHPVF are inferred to have erupted in a low-lying area that has been covered by young alluvial sediments during periods of basin subsidence and wet climatic periods, and have recently been exhumed. The exhumation of these former tuff rings provides relict exposures of pyroclastic rocks that nevertheless record the detailed evolution of phreatomagmatic centres within a fluvio-lacustrine sedimentary basin, whose nature varied in both space and time since the Early Pliocene.

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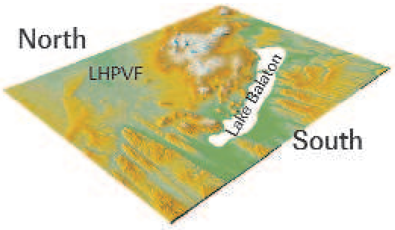
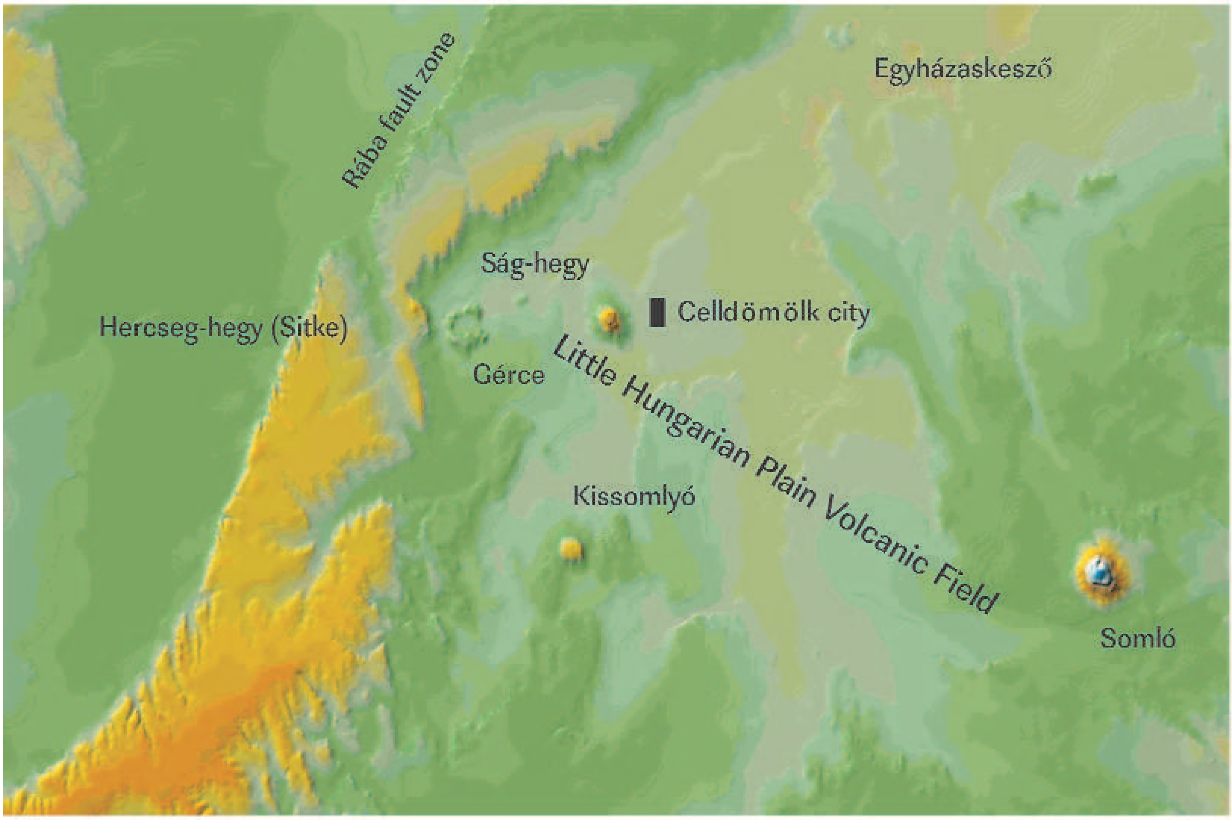
Digital terrain models of the Little Hungarian Plain Volcanic Field showing the main exposed Neogene erosional remnants of intraplate alkaline basaltic volcanoes





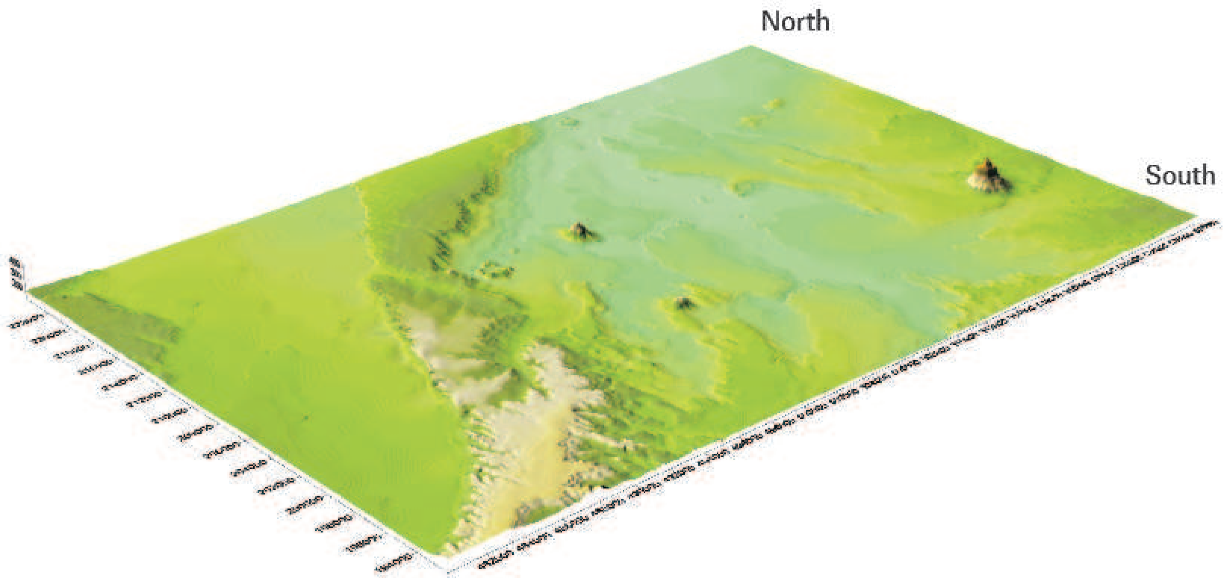
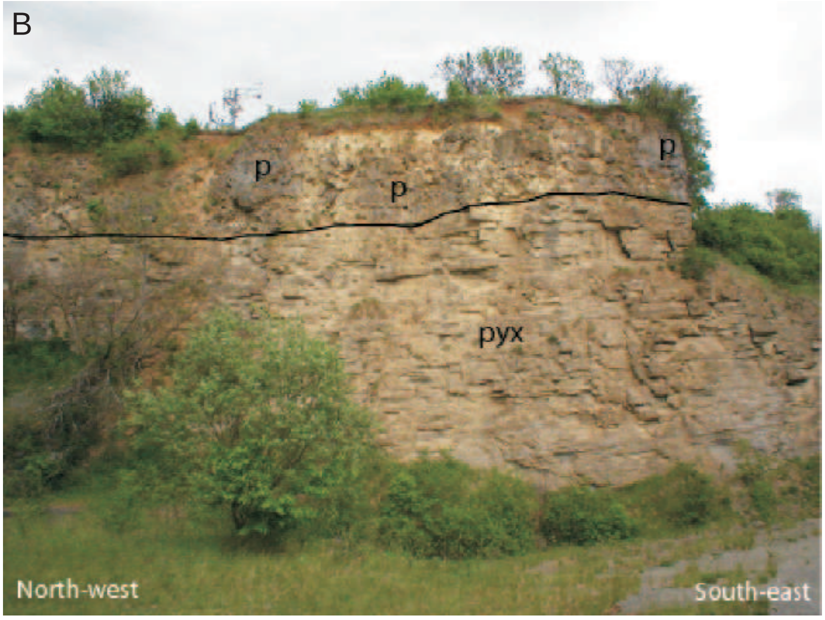
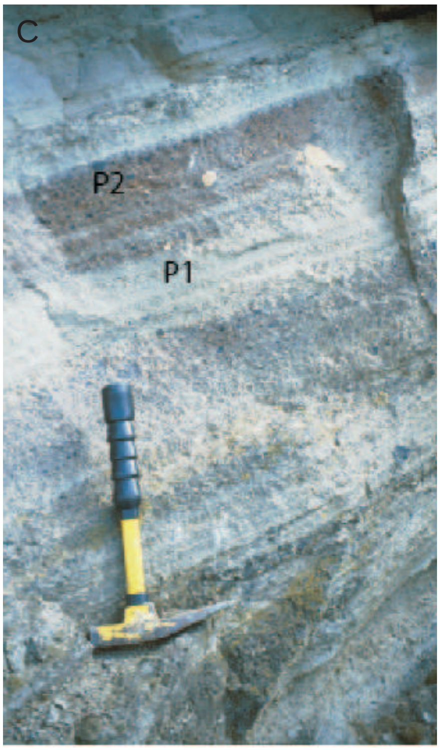


Plate 2 / Chapter 5 Second International Maar Conference — Hungary–Slovakia–Germany

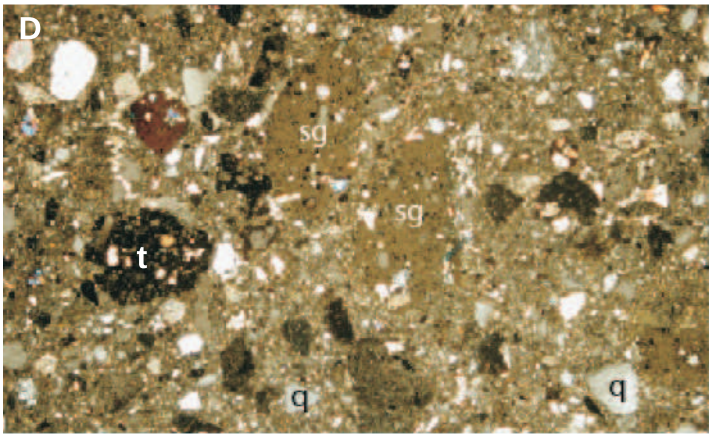
Panoramic view of the pyroclastic mound (line points to the location of the main quarry) of Kis-Somlyó from south



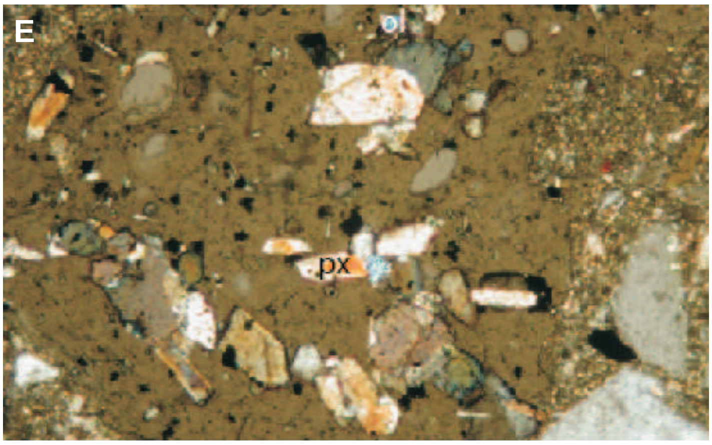
Well-exposed succession of pyroclastic rocks (pyx) overlain by siliclastic rocks that have been invaded by basanitic lava often forming pillows (p) and/or lava tubes cross cutting the siliciclastic sediment. Thick line represents the sharp but irregular contact between the pyroclastic succession of the tuff ring and the post-tuff ring rock units



Two type of pyroclastic lithofacies in the pyroclastic succession of Kis-Somlyó. P1 is a fine grained lapilli tuff, tuff, P2 is a coarse grained, volcanic glass shard rich lapilli tuff, also rich in abraded pyrogenic and xenocrystals that are often calcite cemented



Photomicrograph (parallel polarized light) of blocky, moderately vesicular sideromelane glass shards (sg) from a tuff layer (the short side of the photo is about 4 mm) of Kis-Somlyó volcano. Tachyilite (t) is less dominant in the tuff. The matrix of the tuff is rich in quartz (q)

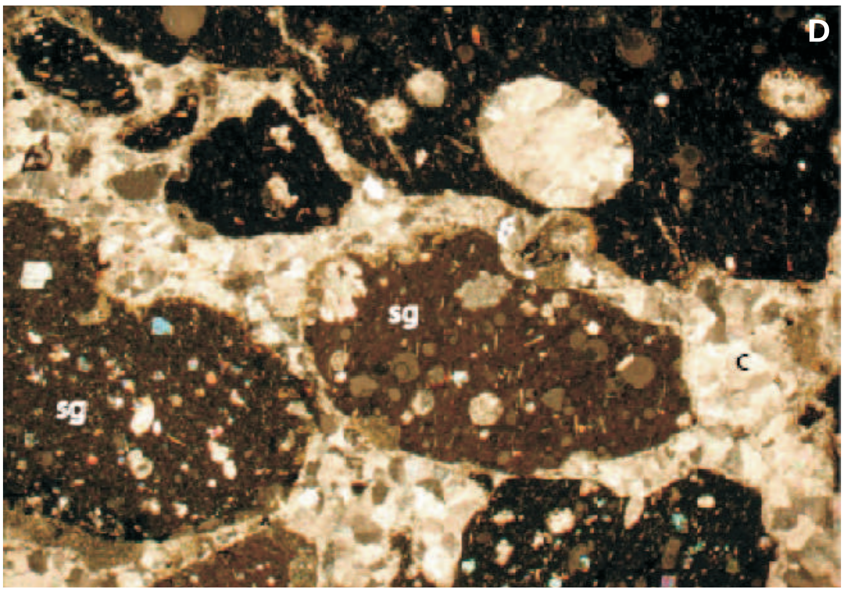


Photomicrograph of a glass shard from Kis-Somlyó lapilli tuff and tuff beds. They are commonly non-vesicular, and have a moderate amount of microlite and/or microphenocryst such as pyroxene (px) and/or minor plagioclase or olivine (ol) (the short side of the photo is about 1 mm)

Impact sag (line) on an exposed bedding plain of the Kis-Somlyó pyroclastic succession

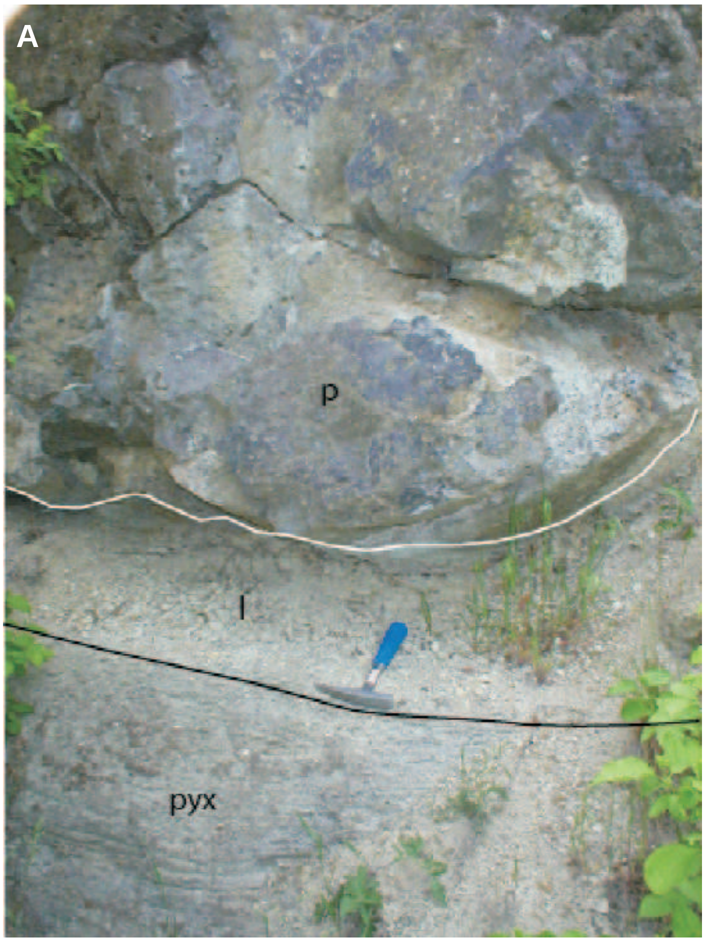
Large clast (s) from a Neogene pre-volcanic siliciclastic succession that has not caused significant impact sag on the bedding surface at Kis-Somlyó

Thickly bedded, accidental lithic and xenocryst rich, fine grained tuff (f) in lensoidal bedding from the tuff ring sequence of Kis-Somlyó

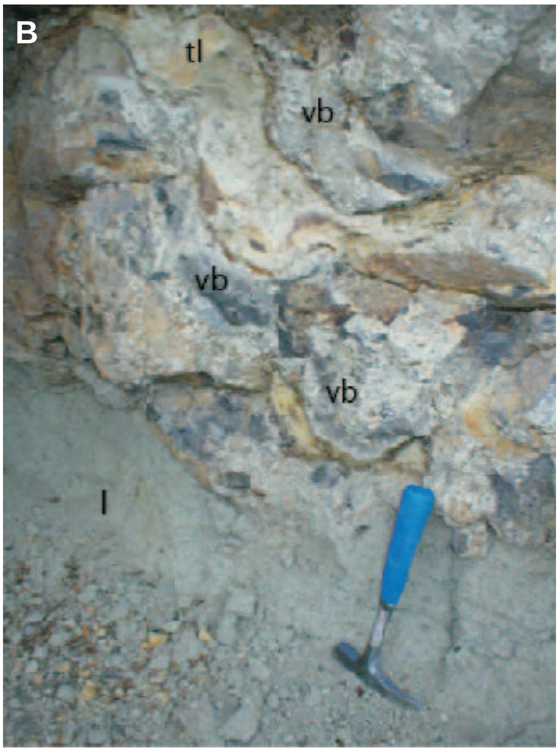


Photomicrograph (cross polarized light) of a calcite cemented (c), trachytic textured sideromelane glass shard (sg) from coarse grained lapilli tuff beds at Kis-Somlyó. The short side of the photo is about 2 mm)

Plate 4 / Chapter 5 Second International Maar Conference — Hungary–Slovakia–Germany

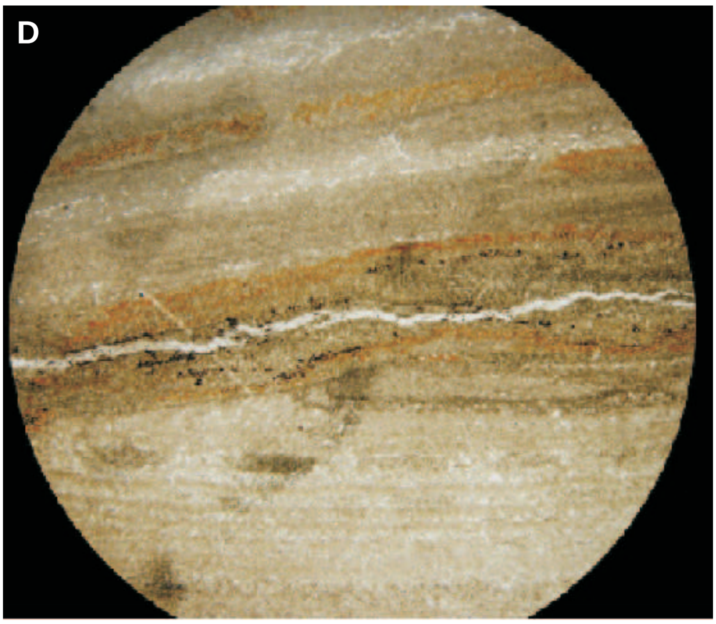


Siliciclastic lacustrine unit (l) that overly the pyroclastic (pyx) succession of Kis-Somlyó. The lacustrine unit is invaded by coherent pillowed basanite (p)



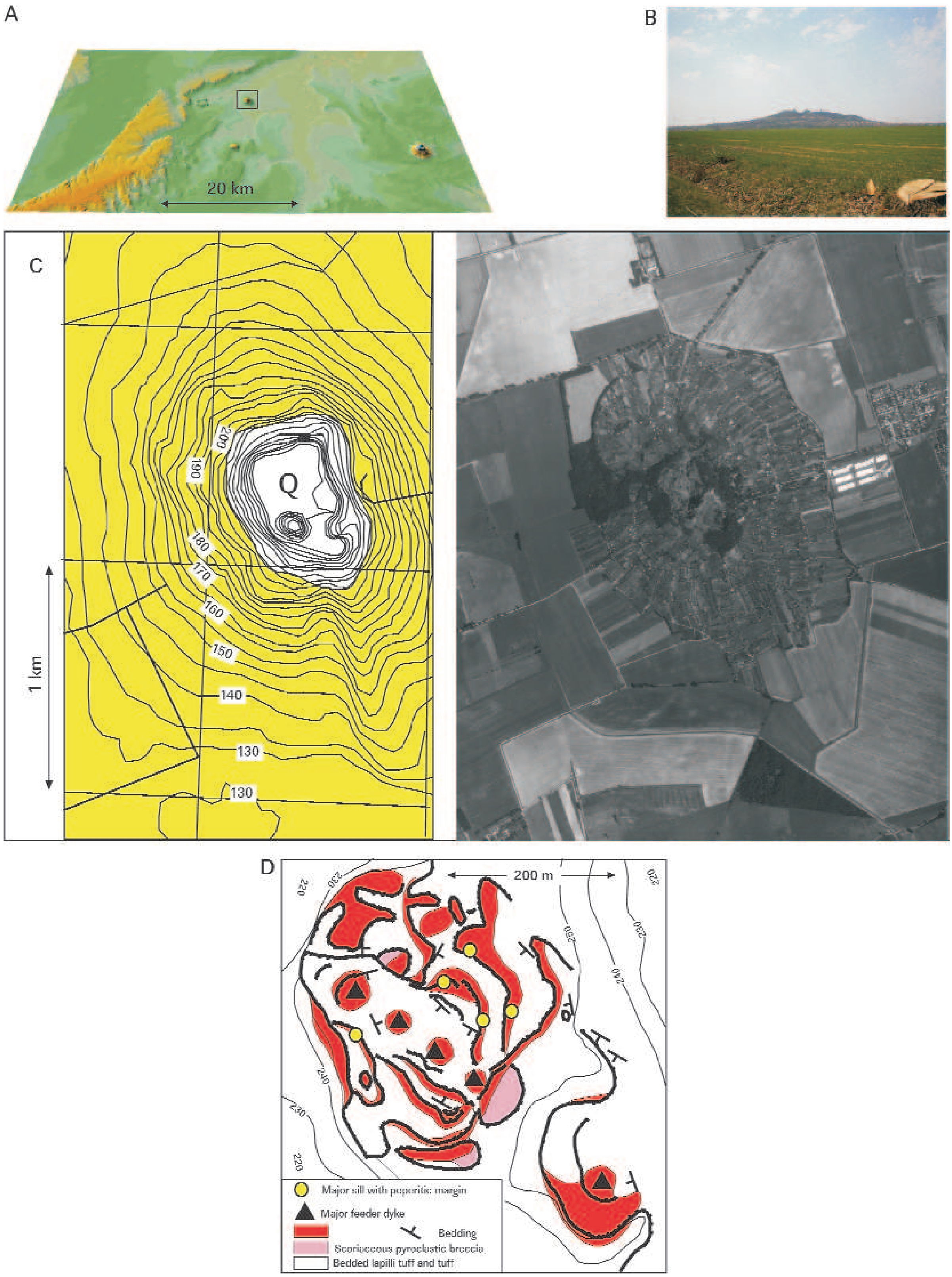
The lacustrine unit (l) at Kis-Somlyó has been truncated by basanitic lava that are commonly highly vesicular in the contact zone (vb) and forming globular peperite. The siliciclastic sediments are thermally altered (tl) where they have been captured in a basanite melt

Pillow lava and lava tube (lt) hosted in the capping siliciclastic succession of Kis-Somlyó

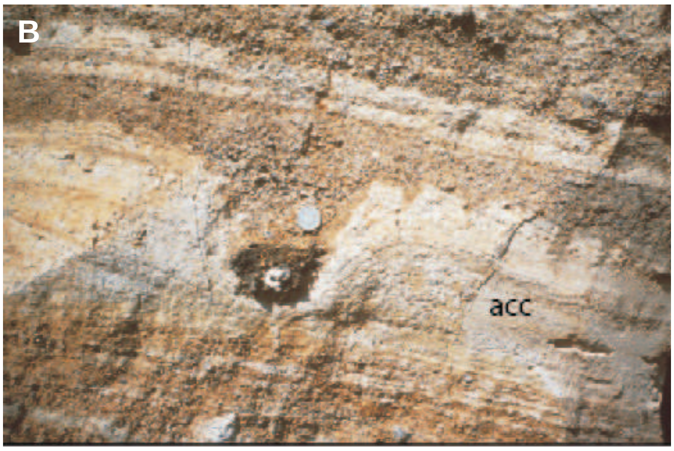


Photomicrograph (parallel polarized light) from the siliciclastic unit capping the pyroclastic succession of Kis-Somlyó. The shorted side of the view is 2 cm

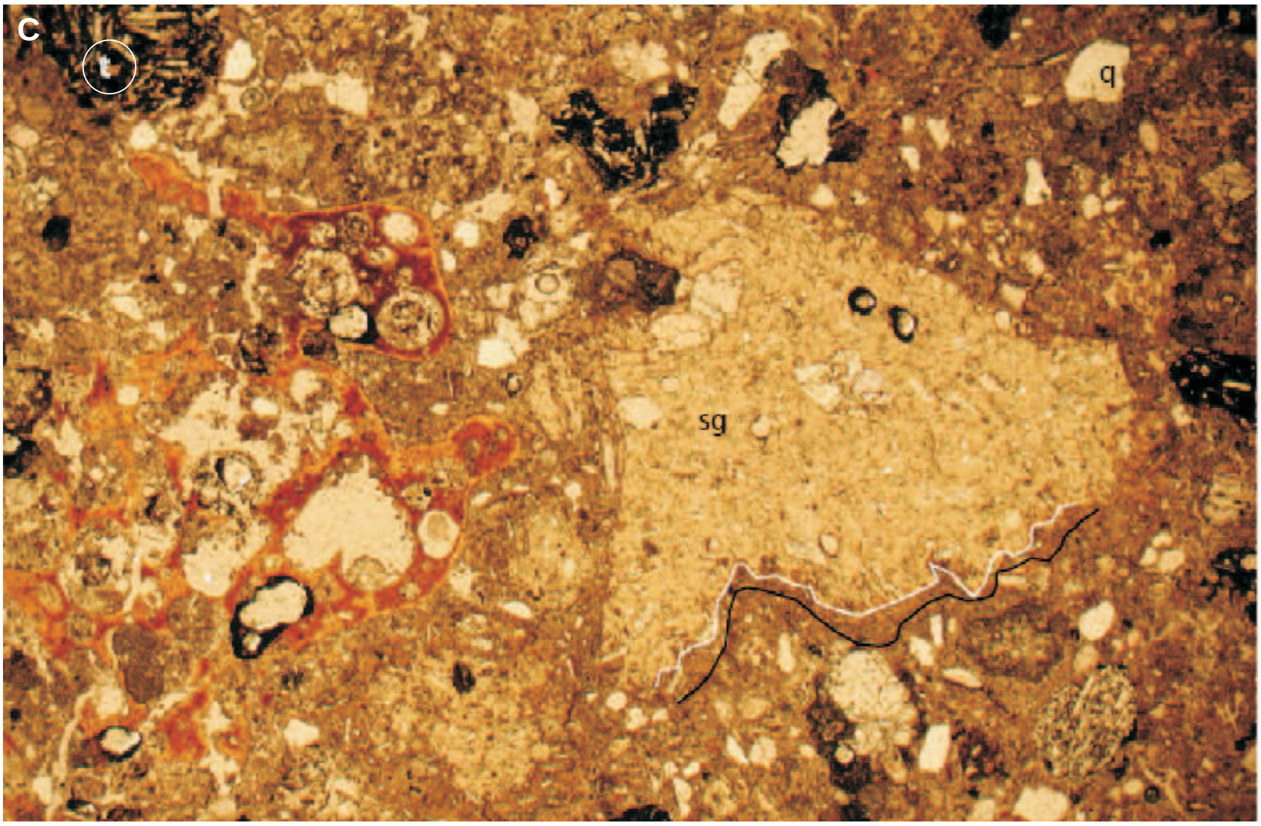
Location map of Ság-hegy (A) and a view from south to the Ság-hegy volcano (B). A simplified geological map (white = volvanites, yellow = Neogene sand, “Q” = the location of the quarry in comparison to the topography of the hill [airphoto from the Hungarian Military courtesy] shows the erosional remnant of the Ság-hegy (C). The outline of the quarry and the location of key features is shown on “C” and "D"



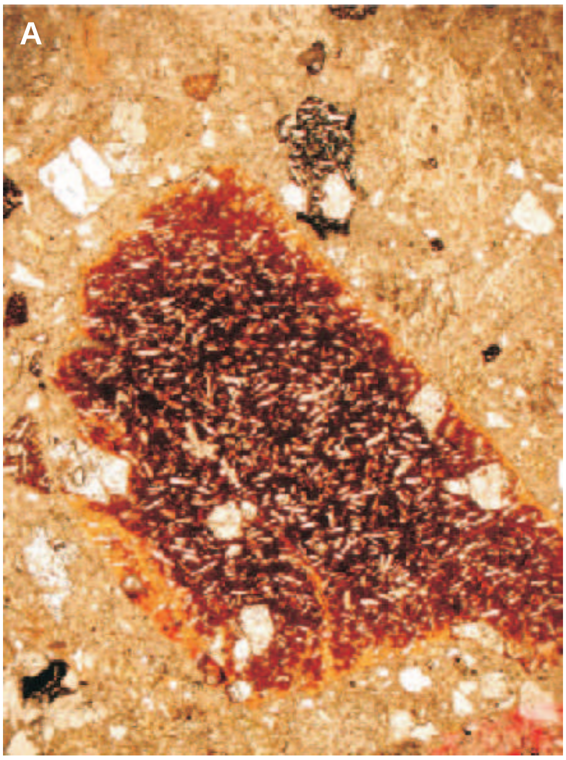
Basal phreatomagmatic pyroclastic succession at Ság-hegy



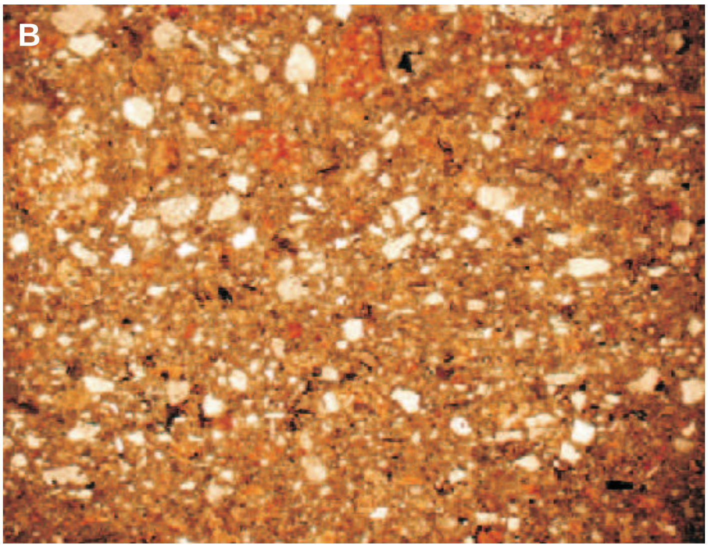
Accretionary lapilli bed (acc) in the upper pyroclastic succession in the north-western side of the erosion remnant of Ság-hegy. Note the deep impact sag caused by a cauliflower bomb (c)



Photomicrograph (plan parallel polarized light) of moderately vesicular tephritic glass shard (sg) from the Ság-hegy lapilli tuff succession. Note the irregular palagonite rim (marked between a white and black line in the lower limit of the glass shard) around the glass shard. The short side of the photo is 2 mm. The lapilli tuff is rich in quartz (q). Tachylite is less common (t). Note the highly vesicular brown glass that is completely turned to pe gel palagonite in the left of the picture



Photomicrograph (plan parallel polarized light) of micro-gabbroid volcanic lithic clast of the Ság-hegy pyroclastic beds. The short side of the photo is 2 mm

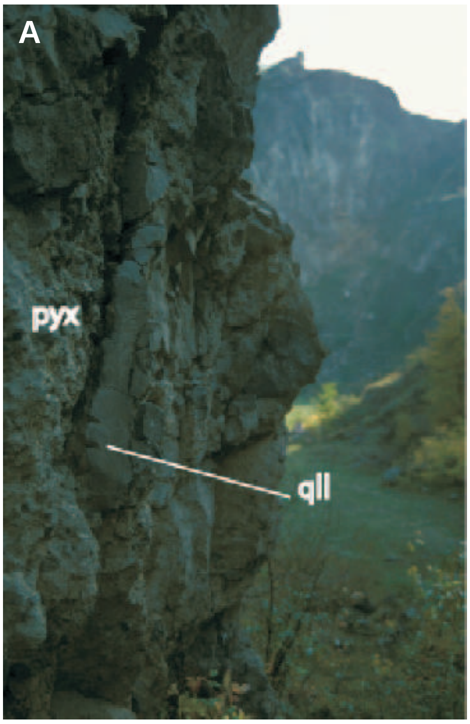


Photomicrograph (plan polarized light) of a muscovite and quartz rich tuff of the Ság-hegy pyroclastic beds. The short side of the photo is 2 mm)

Bed-flattened mud chunk (mc) in an otherwise fine grained lapilli tuff layer of Ság-hegy

A thick sill (s) that intruded into the former tephra ring (pyx). The sill has peperitic contact (p) to the host pyroclastic unit that is truncated by fluidised zones (black line and "f"). The entire succession is subsequently cut by an oblique dyke (d). Numbers represent relative timing of events

Plate 8 | Chapter 5 Second International Maar Conference — Hungary–Slovakia–Germany

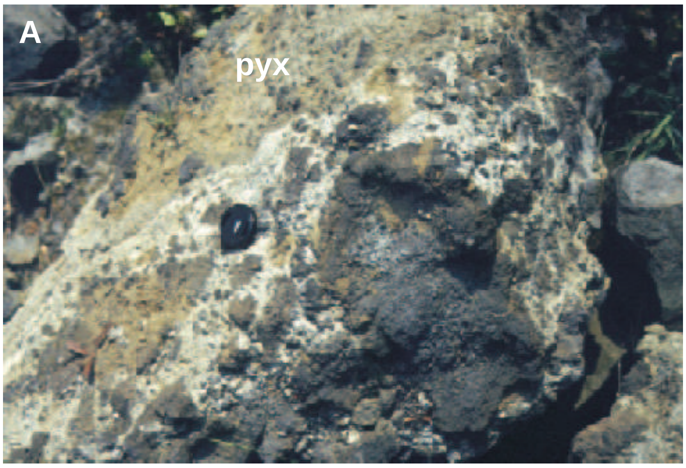


Preserved chilled margin (white line) contact of the former lava lake (qll) to the pyroclastic succession (pyx). The lava lake is inferred to have been occupied the crater/vent zone of the Ság-hegy tuff ring

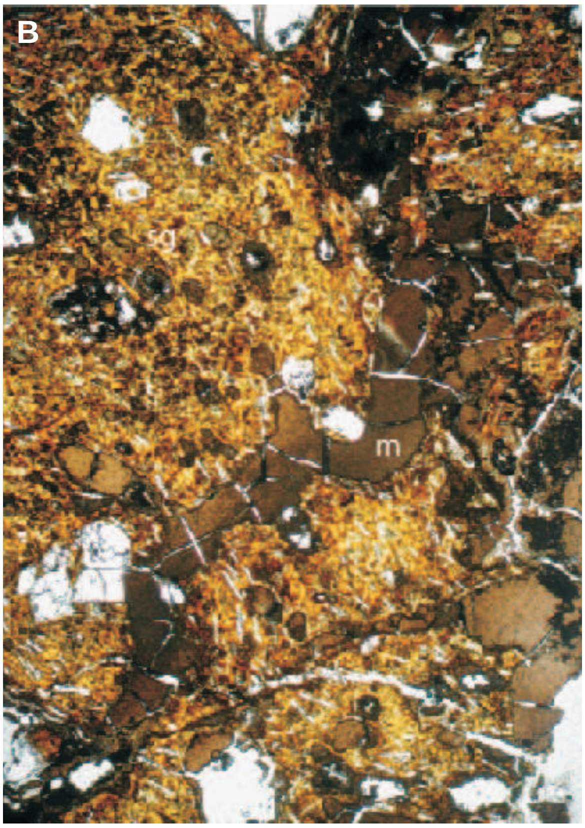
Small basanitic (dm-thick) lava protrusion (black line) that intruded into the tuff ring sequence interpreted to have been fed from the central coherent lava body filled the crater zone of the Ság-hegy tuff ring

Reverse faulting [lava movement inferred to be from left to right caused by the growing lava lake in the upper section of the Ság-hegy tuff ring rim

Bulldozing [lava movement inferred to be from right to left of the former tephra ring caused by the growing lava lake in the upper section of the Ság-hegy tuff ring rim

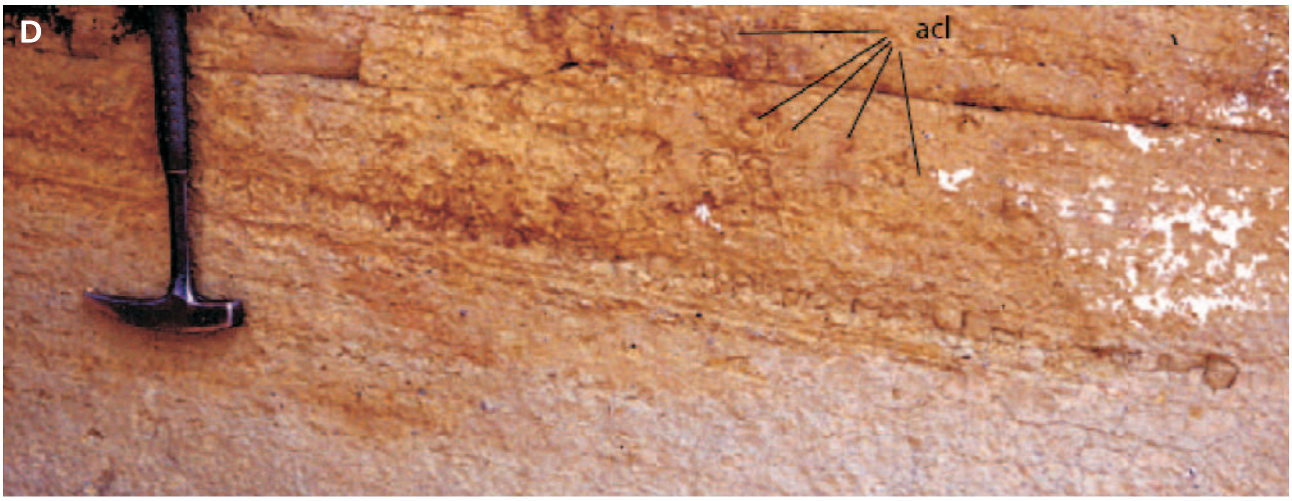


Fluidization (white region around dark basanite clasts, black lines) texture preserved along an intruded basanite body in the upper section of the tuff ring (pyx) sequence

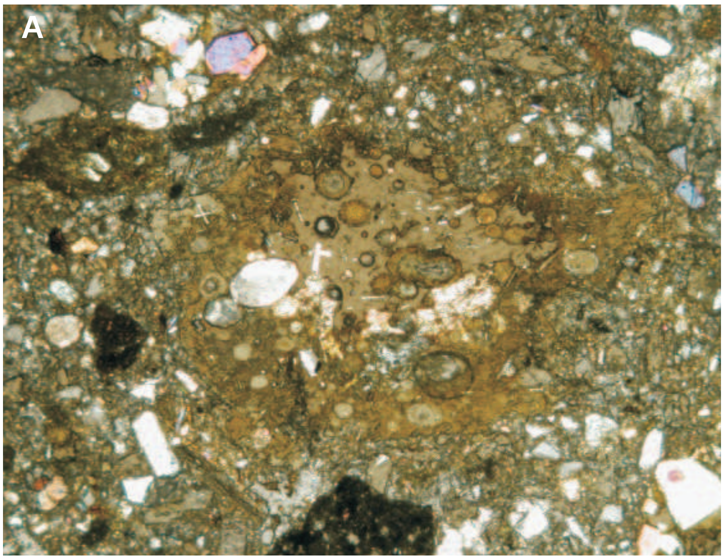


Photomicrograph (parallel polarized light) of a globular peperite from the margin of large coherent lava body at Ság-hegy. Note the smooth surface mud (m), inferred to have been boiled during the interaction of basanite melt (sg) and the host mud. The short side of the photo is about 2 mm

Lapilli tuff and tuff succession in an active quarry in the northern limit of the village Egyházaskesző exposing low angle dipping toward the inferred crater located north of this quarry beside few hundred metres as well as low amplitude dunes (white lines)



Muddy matrix of a lapilli tuff from Egyházaskesző, rich in muscovite flakes. Tuff commonly contains rim-type accretionary lapilli (black lines – acl)



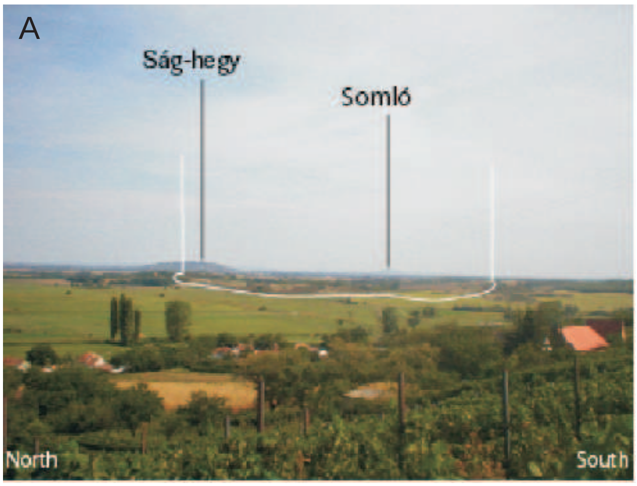
Photomicrograph of moderately vesicular, blocky sideromelane glass shard from the Egyházaskesző lapilli tuff indicative for phreatomagmatic explosive eruption [plan paralell light polarised]. Note the darker coloured rim, a result of advanced palagonitization front progressing inward. Shorter side of picture is about 4 mm

B

Exposed bedding plane of a lapilli tuff of Egyházaskesző that form low amplitude long wave length undulating beds, characteristic for base surge deposition

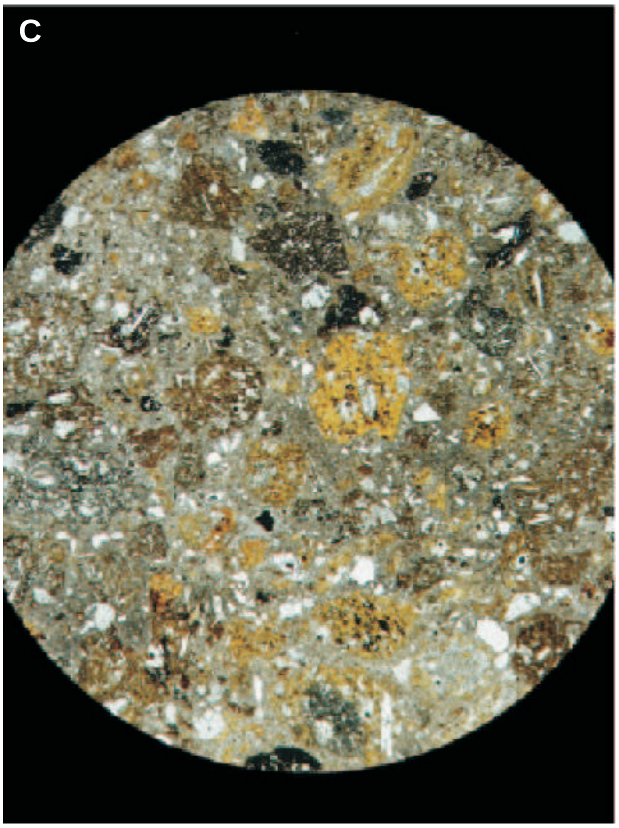
Shallow impact sag on the bedding surface of a base surge bed at Egyházaskesző

Shallow impact sags from the Ubehebe Crater, California. Compare Plate 5.10, C, D with Plate 5.3, A

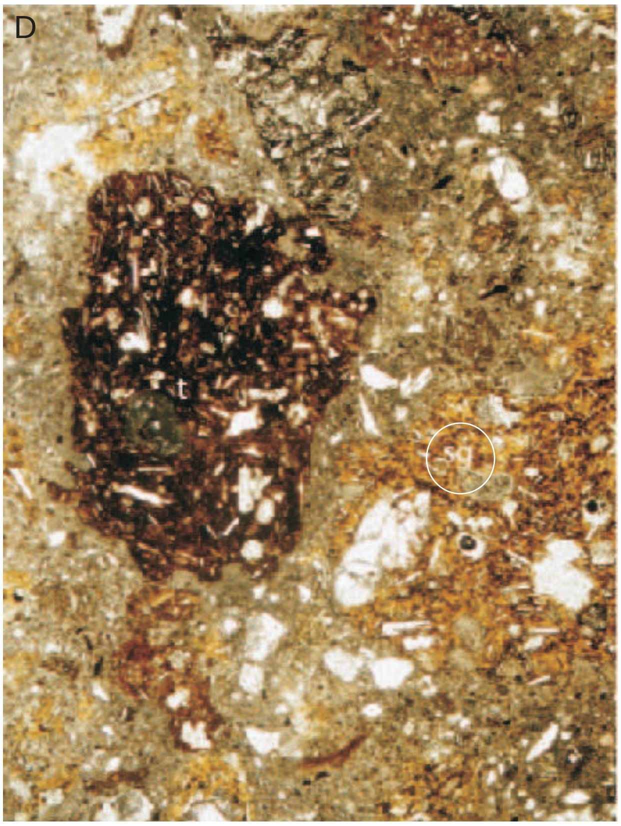


Panoramic view toward the pyroclastic hills east of the village of Gérce (rim marked by white curve line and straight lines point to the proposed rim) forming similar castle-like architecture to the Fort Rock tuff ring in Oregon (HEIKEN 1971)

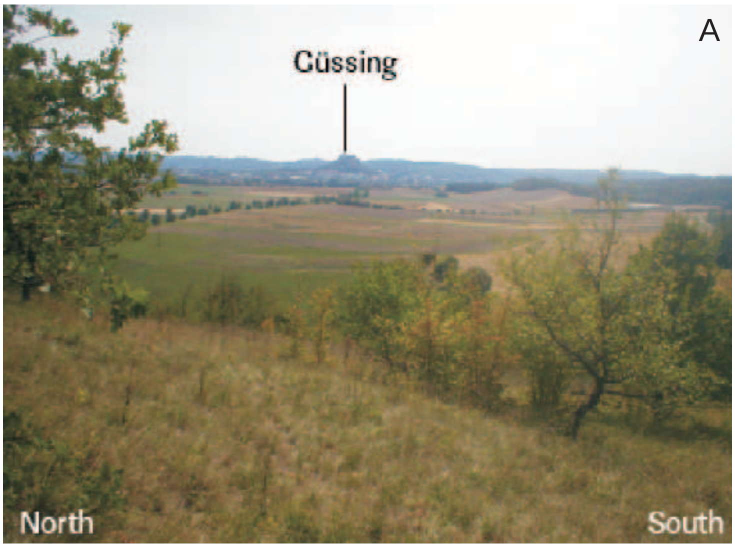
Panoramic view toward the Somló butte from south. Note the small hill on the top of the butte which consist of red, scoriaceous pyroclastic mound



Photomicrograph (plan parallel polarised light) of a mud/silt rich lapilli tuff of the northern basal pyroclastic succession of Somló. The view is about 1 cm across

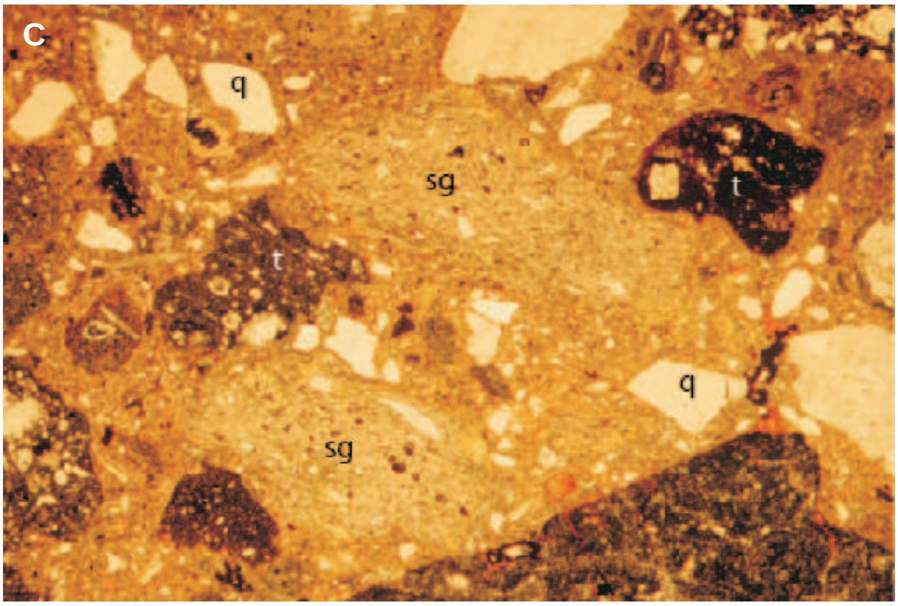


Photomicrograph (plan parallel polarised light) of a sideromelane glass (light colour "sg") and tachylite (dark colour "t") shard from the basal lapilli tuff and tuff succession of Somló. The short side of the photo is about 2 mm



Overview of the Güssing diatreme

Inward dipping pyroclastic beds in the upper section of the pyroclastic succession of Güssing, exposing syn-eruptive remobilisation of tephra



Photomicrograph (plan parallel polarized light) of a lapilli tuff from Güssing, rich in fluidally shaped, moderately vesicular sideromelane glass shards (sg) with moderate palagonitisation and angular tachylie glass (t). The lapilli tuff also rich in quartz (q). The short side of the photo is about 2 mm

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Concluding remarks

Contents

Summary

On the basis of volcanological research of the past 10 years it can be concluded that the Neogene alkaline basaltic rocks in the western Pannonian Basin (Plate 6.1) represent eroded remnants of former maars, tuff rings, tuff cones, scoria cones, dykes, sills and lava fields. The erosional remnants can be clearly identified on satellite images as semicircular buttes (Plate 6.1). The number of studied sites almost includes every location where pyroclastic rocks have been reported over the past hundred years of research history of the western Pannonian region (LÓCZY 1913, 1920, MAURITZ and HARWOOD 1936, 1937a, b, c, d, 1939, HOFFER 1943a, b, BUDAI and CSILLAG 1998, 1999, BUDAI et al. 1999). Each of the studied sites show clear evidence of some degree of magma–water interaction during the eruption of the alkaline basaltic magma. The evidence includes

1. the common presence of volcanic glass shards (HEIKEN and WOHLETZ 1986, ZIMANOWSKI 1998, BÜTTNER et al. 1999, 2002),

2. large volume of predominantly non-volcanic lithics that are often angular in shape and samples of the entire known pre-volcanic rock formations (LORENZ 1986, WHITE 1991, ORT et al. 1998),

3. common presence of peperite (WHITE et al. 2000, SKILLING et al. 2002, WOHLETZ 2002, ZIMANOWSKI and BÜTTNER 2002) along intrusive bodies and/or in the foot zone of lava flows (SKILLING 2002) and

4. typical bedforms such as pyroclastic breccia, sandwave, massive and planar beds (WOHLETZ and SHERIDAN 1979, WOHLETZ and HEIKEN 1992) that are characteristic for base surge transportation. Such textural characteristics of the studied pyroclastic rocks and their contact with intrusive and/or effusive bodies are common in the western Pannonian region, and indicate that phreatomagmatism was a key issue during the eruption of these volcanoes. The eruption style (e.g. predominance of phreatomagmatism) and the present state of erosion (e.g. the level of exposures) make the western Pannonian region comparable to other known eroded phreatomagmatic volcanic fields, such as the Hopi Butte in the western US (WHITE 1990), the western Snake River (Idaho) volcanic field (GODCHAUX, et al. 1992, WOOD and CLEMENS 2004), Waipiata in southern New Zealand (NÉMETH and WHITE 2003) or the Saar-Nahe Basin in Germany (LORENZ and HANEKE 2004).

By total volume of eruptive products, however, the western Pannonian volcanic fields together do not reach the total erupted volume estimated for the Hopi Buttes, western Snake River fields nor the Waipiata fields. A preliminary estimation of the total erupted volume has been made for the Bakony – Balaton Highland Volcanic Field (BBHVF) and gave around 4 km^3 (NÉMETH et al. 2000). The relatively smaller size and the smaller number of known vents of the Little Hungarian Plain Volcanic Field (LHPVF) and the Styrian Basin Volcanic Field (SBVF) probably would not add significantly more to the total erupted volume to reach more than 10 km^3 of eruptive products over 6 million years of eruption history in the western Pannonian region. This volume estimate is preliminary, and perhaps two major facts may alter this number such as (1) the estimation of erosion and (2) the existence of covered volcanic rocks. There is evidence from seismic sectioning, gravimetry, magnetic surveys, and drill cores that there are buried Neogene alkaline basaltic rocks (effusive, intrusive and pyroclastic) in the western Pannonian region, especially in the LHPVF (NEMESI et al. 1994, TÓTH 1994). Similarly, there is evidence from various geophysical studies that along the axis of the Lake Balaton, buried volcanic rocks are likely to exist (CSERNY and CORRADA 1989, SACCHI and HORVÁTH 2002). The exact location of these remnants and their interpretation could be a future research subject, and may contribute significantly to refine our knowledge about the total magma involvement in the Neogene alkaline basaltic volcanism in the western Pannonian region.

The duration of the volcanism in the western Pannonian region estimated to be approximately 6 My (BALOGH et al. 1982, BORSY et al. 1986, BALOGH and PÉCSKAY 2001, BALOGH and NÉMETH 2004), which generally indicates a long lasting volcanic activity in comparison to other similar volcanic fields (CONNOR and CONWAY 2000). This time length also suggests a relatively low recurrence rate of the volcanism in this region in comparison to other similar volcanic fields (CONDIT and CONNOR 1996, CONWAY et al. 1998). The available data base is not sufficient to state more about age grouping of volcanic events in the region and therefore this also needs to be addressed in future research subjects.

The general stratigraphical relationship of various pyroclastic successions indicates an initial phreatomagmatic phase which commonly turned into a purely magmatic phase. This stratigraphic relationship also indicates that the water sources have been exhausted during the initial state of the magma uprise, and subsequently magmatic gas-driven fragmentation followed. This relationship is in good agreement with the model of LORENZ (1986) and therefore, the western Pannonian volcanic fields are similar to other maar volcanic fields world-wide. However, there is evidence that deep groundwater reservoirs commonly have been involved in the magma–water interaction especially in those areas, where the cover of immediate pre-volcanic siliciclastic rocks was thin (tens of metres). In these areas, the initial interaction between magma and the water from the porous media aquifer is inferred to have been significantly influenced by water stored in fracture controlled aquifers (e.g. karst water). In extreme case when no such water was involved in the magma–water interaction just a limited magma-water interaction may have occurred and built a thin tephra ring sequence that was topped by scoria cones. In contrast, where water from fracture controlled (e.g. karst water) aquifers was significantly involved the magma–water interaction may have lasted long enough to build deep maars. It is sug-

gested, that due to the seasonality of the fracture controlled (e.g. karst water) aquifers, some sort of control could be expected in a volcanic field, which developed over such pre-volcanic buildup (NÉMETH et al. 2001). The studied pyroclastic successions in the western Pannonian region all exhibit textural features indicating that the immediate pre-volcanic Neogene silicilastic sediments must have been water saturated and unconsolidated soft rock environment. These sediments functioned as an impure coolant (WHITE 1996) during the magma–water interaction. In areas where the thickness of this immediate pre-volcanic sediments exceeded a few hundreds of metres, lateral excavation of craters led to the formation of broad, shallow volcanic depressions, like e.g. in the LHPVF. In contrast, where the fracture controlled aquifer was near to the syn-volcanic palaeosurface, steep walled maars are inferred to have been formed. In either case, the resulting pyroclastic rocks became rich in finely dispersed siliciclastic rock fragments and/or mineral phases derived from such sediments (e.g. quartz, muscovite). This textural appearance of the pyroclastic rocks from the western Pannonian region commonly led to the interpretation that these rocks are the result of contemporaneous volcanism and sedimentation (KULCSÁR and GUCYZNÉ SOMOGYI 1962, JÁMBOR 1980). Indeed, to distinguish accidental lithic fragment rich pyroclastic rocks from normal volcanogenic sediments from any environment (e.g. subaqueous or terrestrial) is extremely difficult (RIGGS and BUSBY-SPERA 1990, RIGGS et al. 1997, BULL and CAS 2000). In spite of these difficulties most of the lapilli tuff and tuff units are interpreted to be the result of deposition form base surges and or phreatomagmatic fall out on the basis of their bedding characteristics and/or the general 3D relationships mapped in these areas. Similarly, detailed revision of drill core data with interbedded volcaniclastic and Neogene siliciclastic units suggested that such apparent interfingering is rather a result of near crater rim subsidence of tuff ring beds (e.g. Pula) that may cause apparent bed repetition (JÁMBOR and SOLTI 1975, 1976). There have also been interpretations, which concluded that the volcanic rocks were quickly covered by the immediate pre-volcanic silt and sand. These observations gave an impression, that the Late Miocene siliciclastic sedimentation was contemporaneous with the volcanism (JÁMBOR 1980), and therefore the volcanoes were interpreted to have been evolved in a shallow marine environment (JÁMBOR and SOLTI 1976). There is evidence that such siliciclastic sediments that become deposited above the volcanic successions often resulted from sedimentation in a volcanic depression, e.g. Fekete-hegy (MARTIN et al. 2002, 2003) or Kis-Somlyó (MARTIN and NÉMETH 2004a). However, there is also growing evidence, that coherent alkaline basaltic bodies interbedded with siliciclastic rock units, are inferred to be predominantly intrusive bodies, e.g. volcanic region north of the Keszthely Mts. In the LHPVF, where flat volcanic edifices, such as broad tuff rings, have developed during the magma/water interaction, is evidence, that such craters have been flooded by water and subsequently siliciclastic sediments became deposited in their crater. At this stage it is not yet understood whether this events are related to

1. general basin subsidence,
2. temporal increase of water input into the sedimentary basin (e.g. climatic forcing) or
3. they are just local events with no significant regional implication.

It is also evident that maar basins became flooded by water either quickly, or rather slowly. These two end members resulted in significantly different volcanic facies association from

1. complete scoria cones (feeding intra-crater lava flows that in case of subsequent flooding could have been covered by siliciclastic maar crater deposits) and/or
2. Surtseyan volcanoes that may evolved in maar basins that have been flooded quickly.

The complicated stratigraphic relationship of various volcanic rock units and their relationship to the pre- and post-volcanic non-volcanic succession suggests that the studied volcanoes in the western Pannonian region are complex and cannot be really classified as pure monogenetic volcanoes. At this stage it is also not clear, if various volcanic facies in the same location truly represent volcanic events that took place more or less in the same time. There is evidence in a few cases, that a complex volcano has been constructed in a short period of time (MARTIN and NÉMETH 2004b). However, their eruptive time may already exceed the common time that is necessary to freeze a feeder dyke and, therefore, a volcano, regardless of its small volume in some instances could rather be viewed as a closely spaced group of individual vents that may even have been fed from completely different sources.

One of the major outcomes of the research in the past 10 years is the recognition the complexity of the small-volume intraplate volcanoes in the western Pannonian Basin. There is evidence showing the recurrence of volcanic activity in the same locations. In a relatively small area, such as e.g. the Szigliget volcanoes, each preserved hill in an area of 2 km^2 is interpreted as an individual diatreme pipe. One of these closely spaced vents, recognised recently, may be even more complex and represent a nested diatreme in an area smaller than 500 metres across.

In summary it can be stated that phreatomagmatism was the main eruptive mechanism that created the original volcanic landforms in these volcanic fields. The palaeoenvironment of the volcanic fields are best modelled as relatively flat lying areas, where fluvial incision was insignificant during onset of volcanism. In the relatively broad, flat land, elongated valleys likely have been temporarily flooded by surface water, and/or occupied by flat, shallow lakes all enhancing the development of phreatomagmatic volcanoes during magma uprise. The fine distinction between volcanoes that formed purely subaerially (maars and/or tuff rings) from volcanoes that may have been at least in their initial eruptive phase subaqueous is still a subject of future long term research plans in the region. Overall, to study the volcanism in the western Pannonian Basin may contribute to our understanding of a pre- syn and post-volcanic sedimentary environment and its relationship with an ongoing intraplate predominantly alkaline basaltic volcanism. In this respect, the evolution of the western Pannonian Basin and its Mio/Pliocene intraplate volcanism is remarkably similar to the Neogene to present volcanic fields of the Basin and Range Province, where large lacustrine systems have been present before, during and after the volcanism, such as e.g. Lake Idaho (SMITH et al. 1989). Such lacustrine systems actively were effected by the volcanism causing base level fluctuations and associated sudden, often dramatic drainage as well as flooding (ORE et al. 1996, SADLER and LINK 1996, TALBOT and ALLEN 1996). In such lacustrine systems complex subaqueous to emergent volcanoes (e.g. Pahvant Butte, Utah – WHITE 1996, 2001) developed in time of high water stand and purely subaerial phreatomagmatic ones in low water stands (e.g. Western Snake River Plain — WOMER et al. 1980, GODCHAUX et al. 1992) often associated with complex volcano-sedimentary (e.g. Challis volcanic field, Idaho — PALMER and SHAWKEY 1997) deposition. In the western Pannonian Basin active shallow marine to lacustrine sedimentation (Lake Pannon) took place at least up to 8 My ago (MAGYAR et al. 1999). Subsequent Pliocene lacustrine sedimentation in the region is generally assumed, but was never constrained well on the basis of large scale sedimentary basin analysis. However, preliminary results highlight the difficulty to identify different lacustrine events on the basis of texturally similar siliciclastic successions (SACCHI and HORVÁTH 2002) a common problem in terrestrial sedimentary facies analysis (e.g. YOUNGSON et al. 1998). The western Pannonian Basin is an excellent area to study volcanic rocks resulted during Neogene intraplate volcanism. The region is also an area where deeper levels (crater or diatreme) of small volume intraplate volcanoes are exposed partially due to quarrying. In this regard, the volcanic regions in the area of the western Pannonian Basin are suitable for full development of volcanic national monuments, volcanological exhibition sites and perhaps to be part of a larger geopark networks in Central Europe as it has been suggested in several places already (e.g. NÉMETH 1996, CSILLAG 2004).

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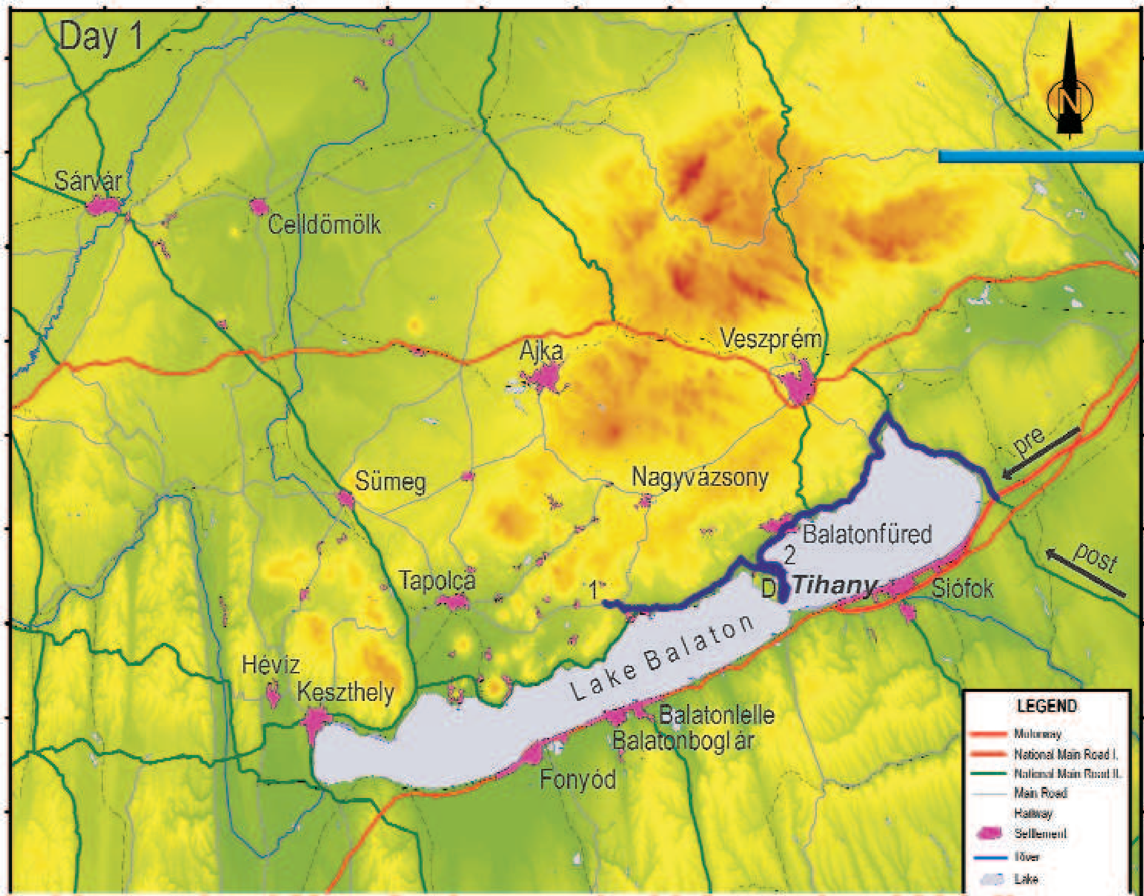
MrSID satellite image (NASA) of the Bakony – Balaton Highland Volcanic Field
Selected vent remnants are marked, 1 — Szigliget, 2 — Badacsony, 3 — Gulács, 4 — Szent György-hegy, 5 — Agár-tető, 6 — Bondoró, 7 — Fekete-hegy, 8 — Tihany, 9 — Kab-hegy, 10 — Tátika, 11 — Haláp, 12 — Fonyód, 13 — Boglár, 14 — Somló (part of the Little Hugnarian Plain Volcanic Field). [Green is forest, pink is agricultural land.] View is 70×50 km

MrSID satellite image (NASA) of the Little Hungarian Plain Volcanic Field.
Vent remnants are marked, 1 — Ság-hegy, 2 — Kis-Somlyó, 3 — Somló. 4 — Gérce, 5 — Hercseg-hegy (Sitke), 6 — Várkesző. [Green is forest, pink is agricultural land.] View is 60×40 km



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