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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-

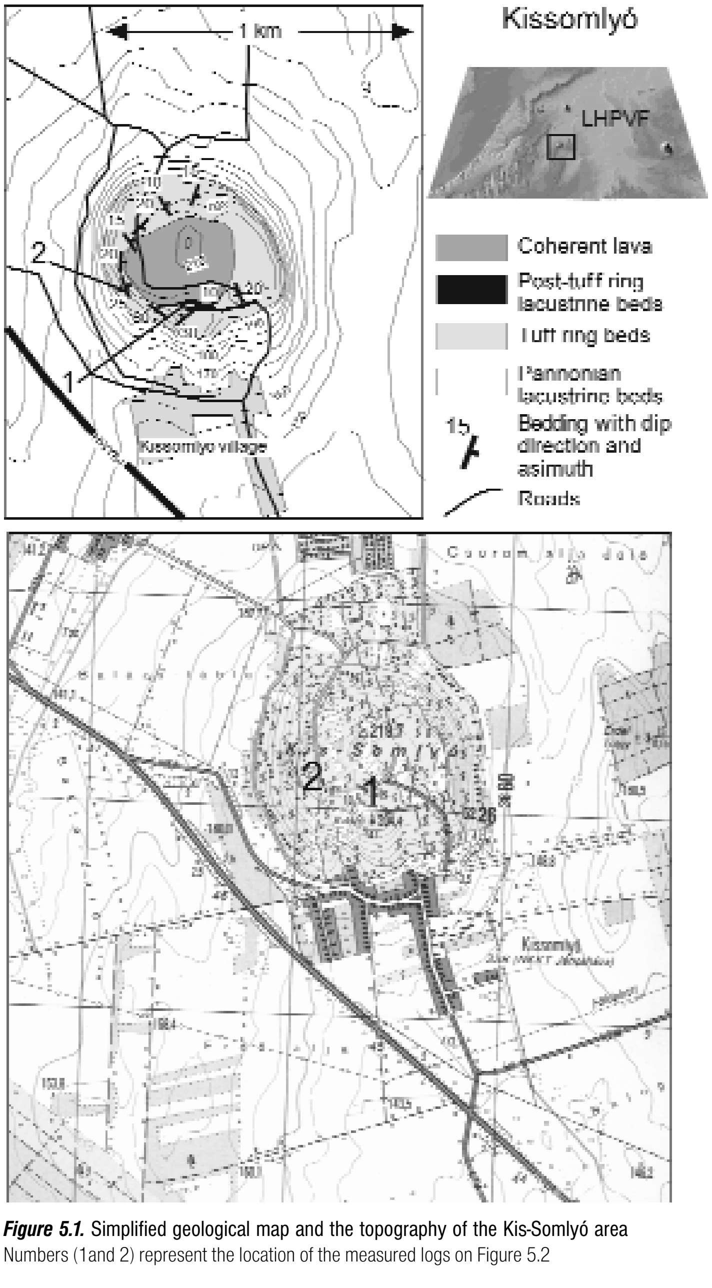

Figure 5.1. Simplified geological map and the topography of the Kis-Somlyó area
Numbers (1and 2) represent the location of the measured logs on Figure 5.2

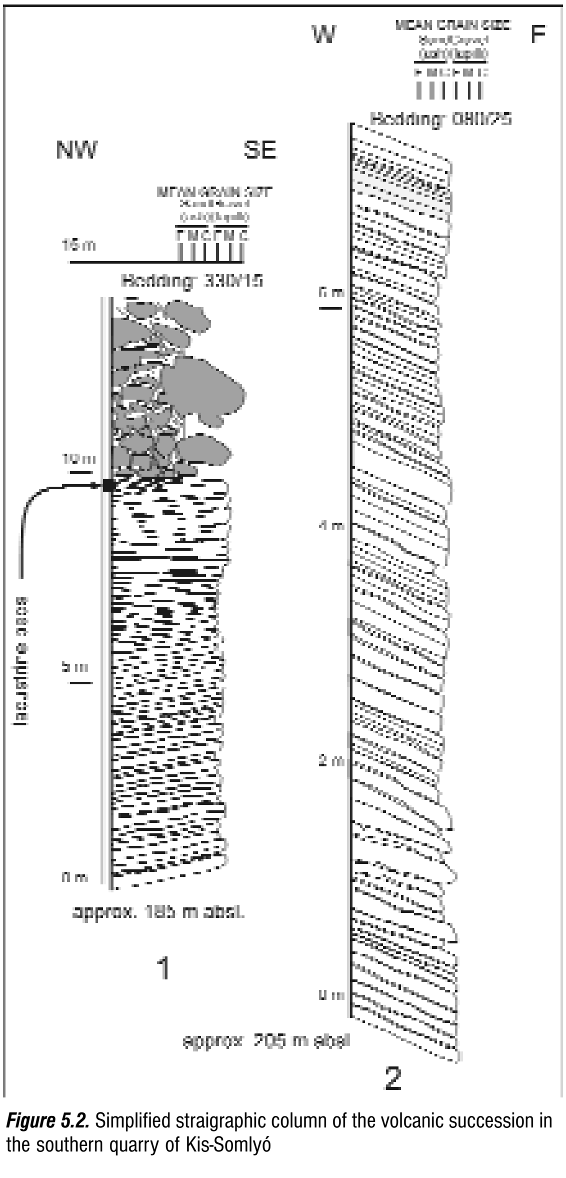

Figure 5.2. Simplified straigraphic column of the volcanic succession in the southern quarry of Kis-Somlyó

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.

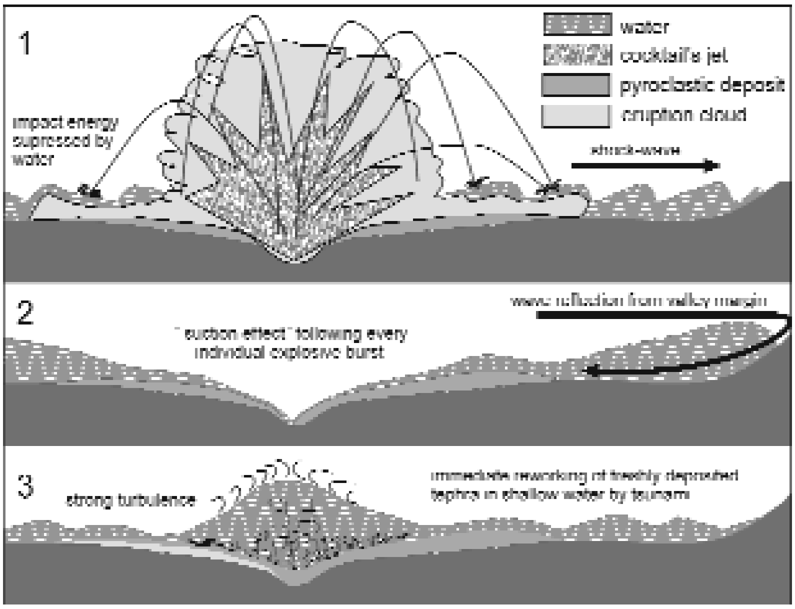

Figure 5.3. Model for the generation of different pyroclastic lithofacies at Kis-Somlyó
1. phreatomagmatic explosive eruption in shallow water produced a pyroclastic mound, 2. eruption advanced and produced shock waves that travelled through the shallow water, and 3. caused immediate reworking of the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

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.

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.

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

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).

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-

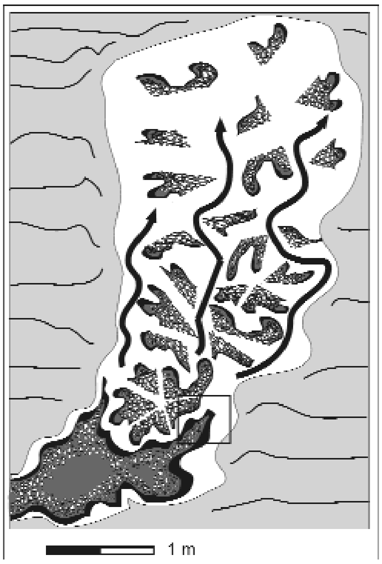

Figure 5.8. Formation of a fines enriched halo along intrusions and their relationship with different peperite textures as a result of phreatomagmatic explosive disruption of the dyke tip due to breakage of the steam envelope. Lines show original bedding. In rectangule see the still intact dyke/sill 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).

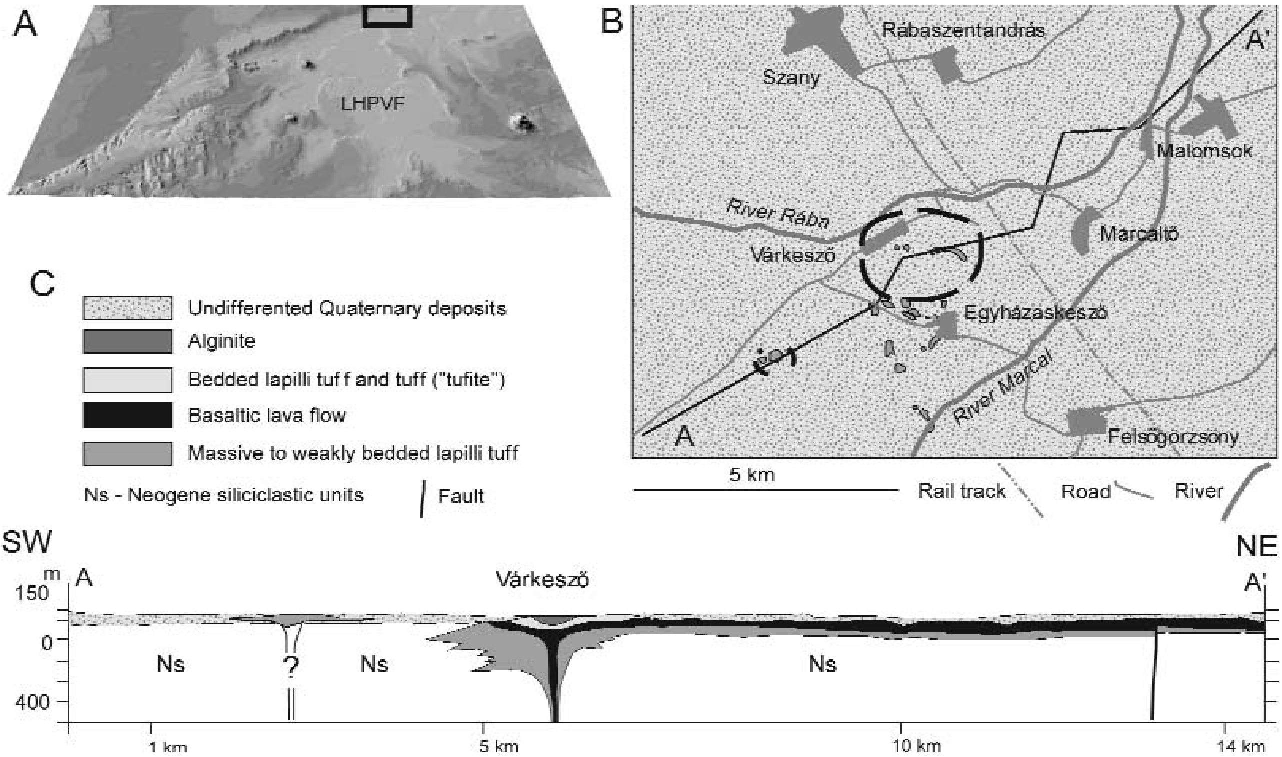

Figure 5.9. Location maps (A, B) of double tuff ring between Várkesző and Egyházaskesző villages. [Geological data are after Bence et al. 1978] Cross-section (C) is predominantly based on shallow drill cores. Dashed line circles on B locate the tuff rings

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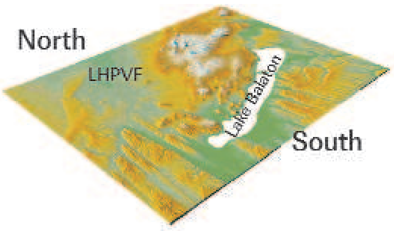
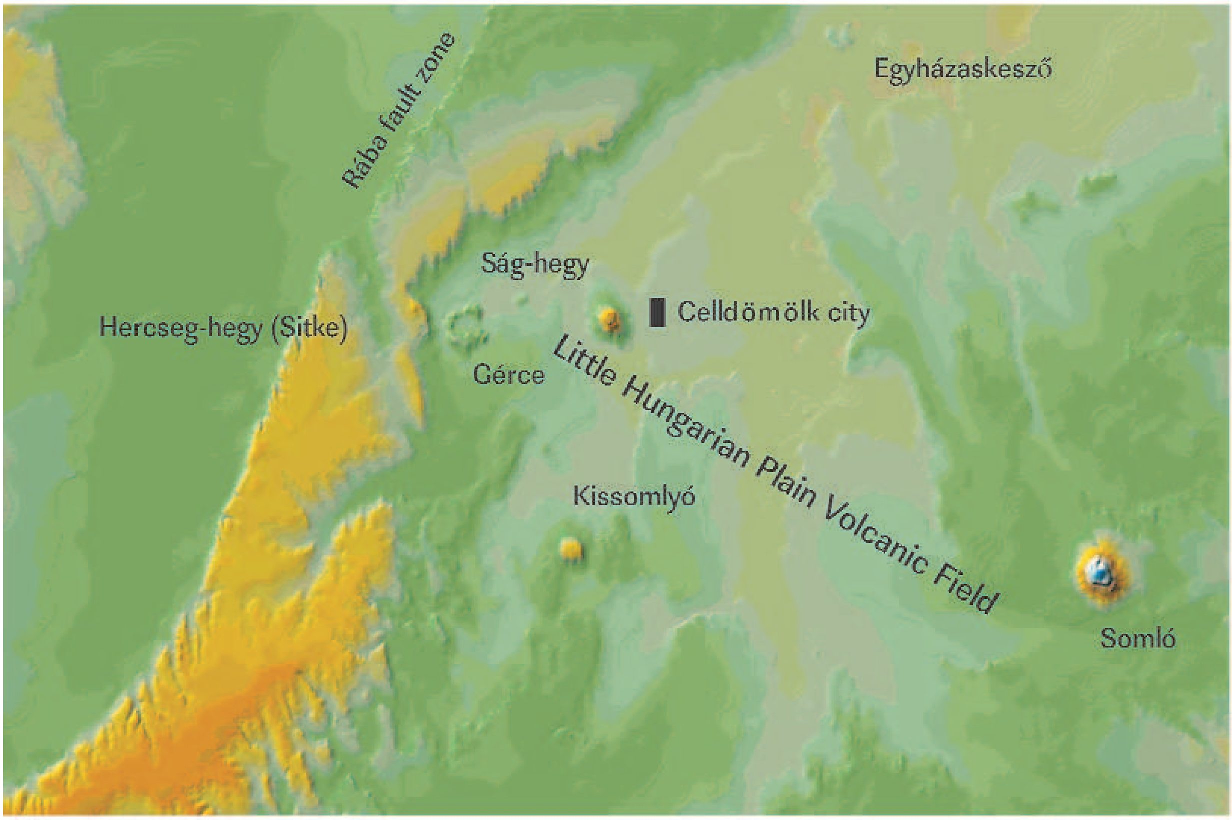
Digital terrain models of the Little Hungarian Plain Volcanic Field showing the main exposed Neogene erosional remnants of intraplate alkaline basaltic volcanoes





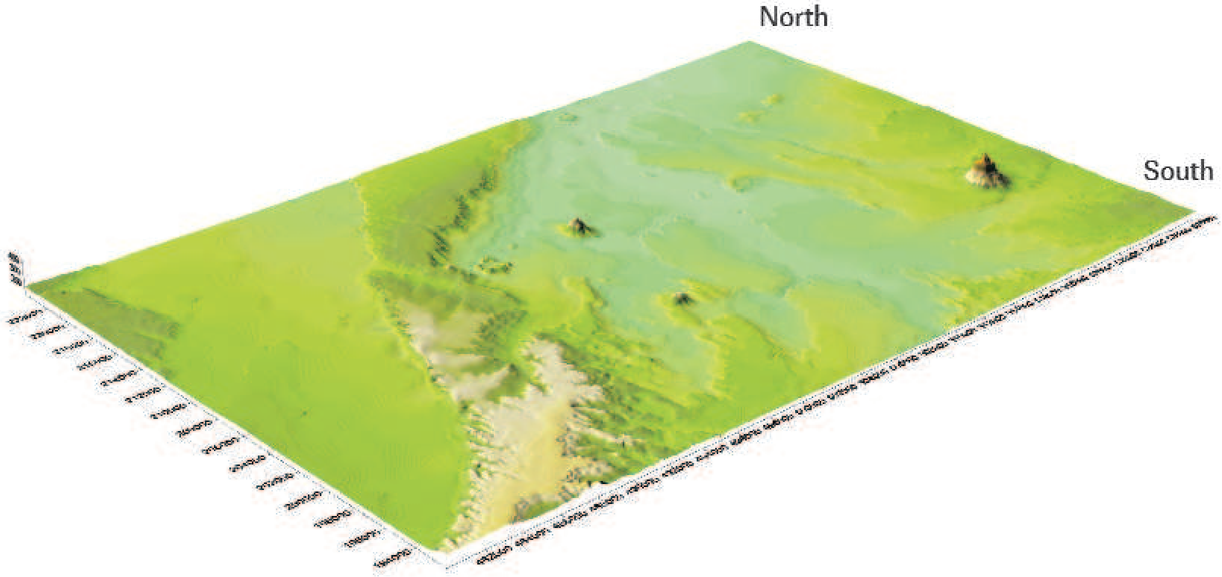
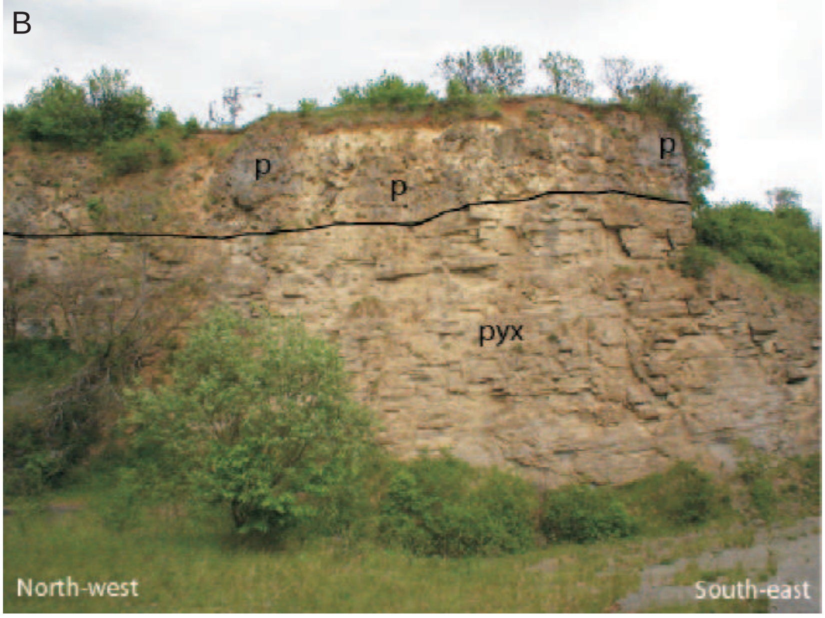
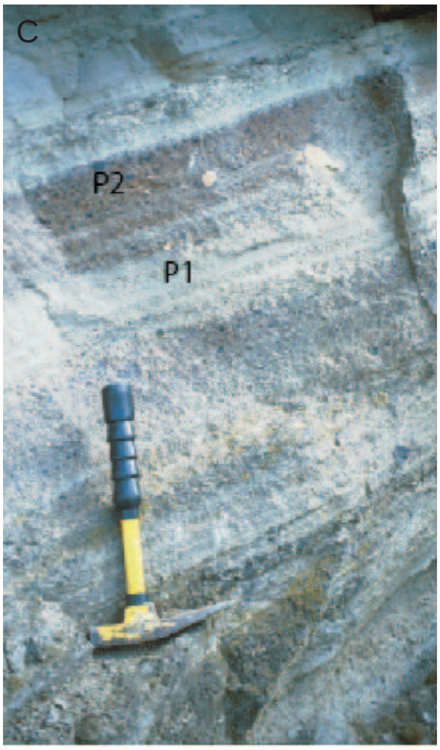


Plate 2 / Chapter 5

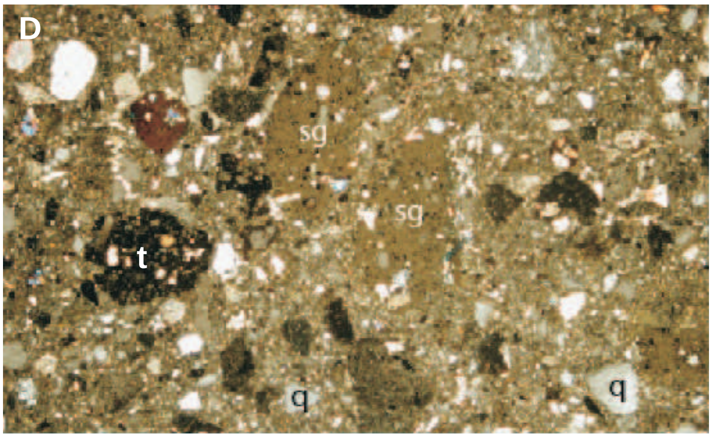
Panoramic view of the pyroclastic mound (line points to the location of the main quarry) of Kis-Somlyó from south



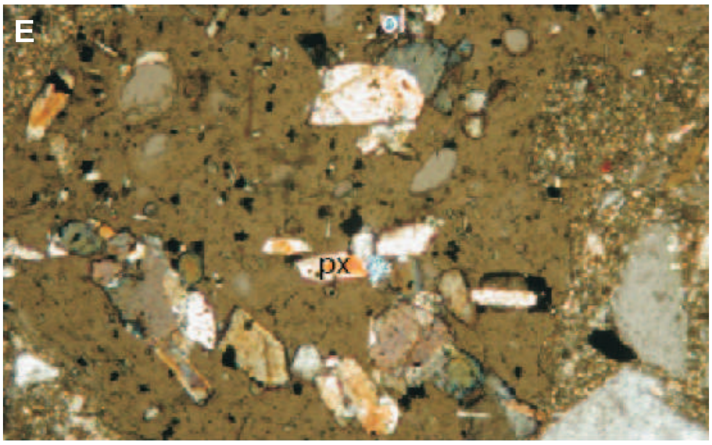
Well-exposed succession of pyroclastic rocks (pyx) overlain by siliclastic rocks that have been invaded by basanitic lava often forming pillows (p) and/or lava tubes cross cutting the siliciclastic sediment. Thick line represents the sharp but irregular contact between the pyroclastic succession of the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

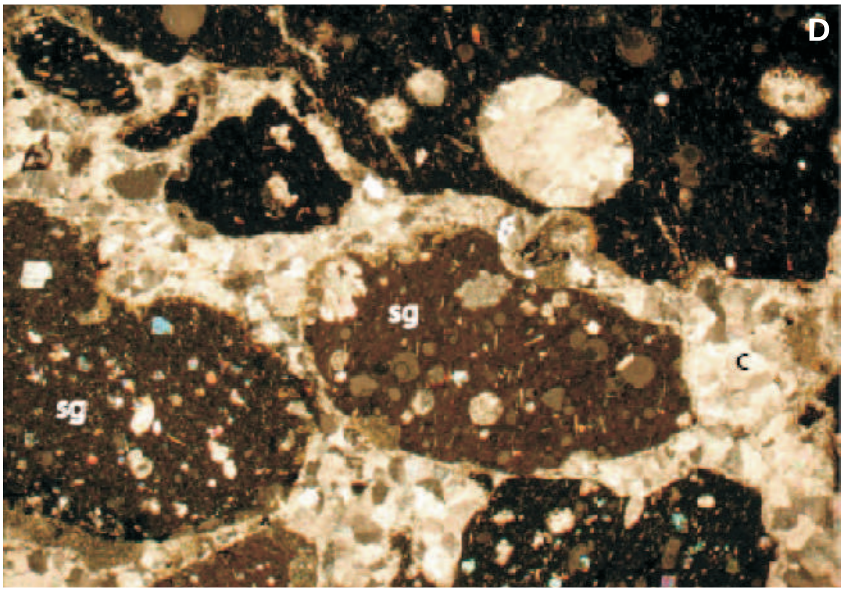


Photomicrograph of a glass shard from Kis-Somlyó lapilli tuff and tuff beds. They are commonly non-vesicular, and have a moderate amount of microlite and/or microphenocryst such as pyroxene (px) and/or minor plagioclase or olivine (ol) (the short side of the photo is about 1 mm)

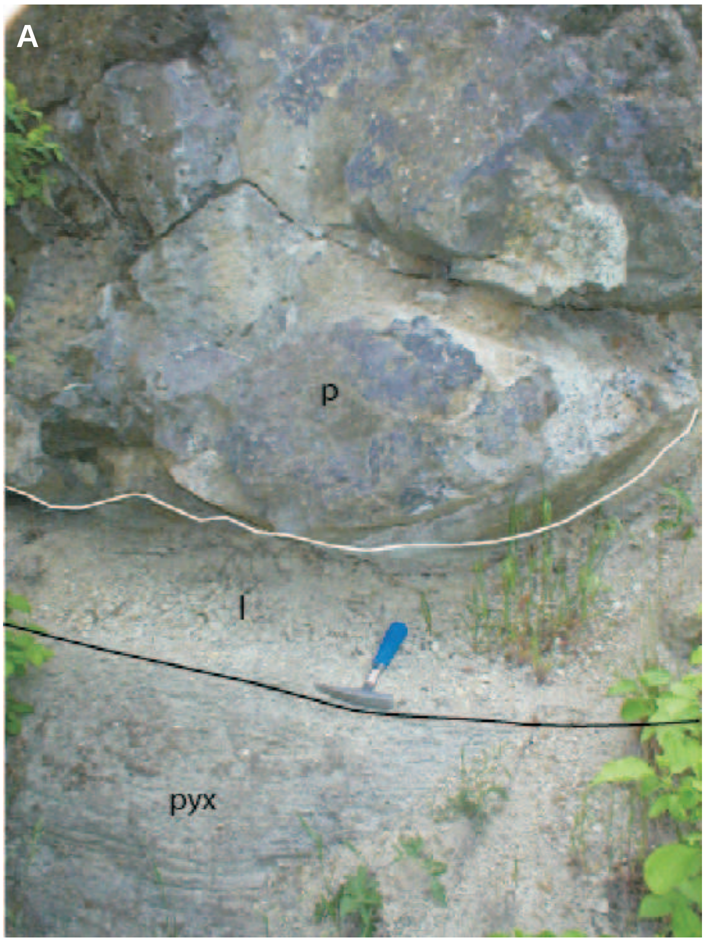
Impact sag (line) on an exposed bedding plain of the Kis-Somlyó pyroclastic succession

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

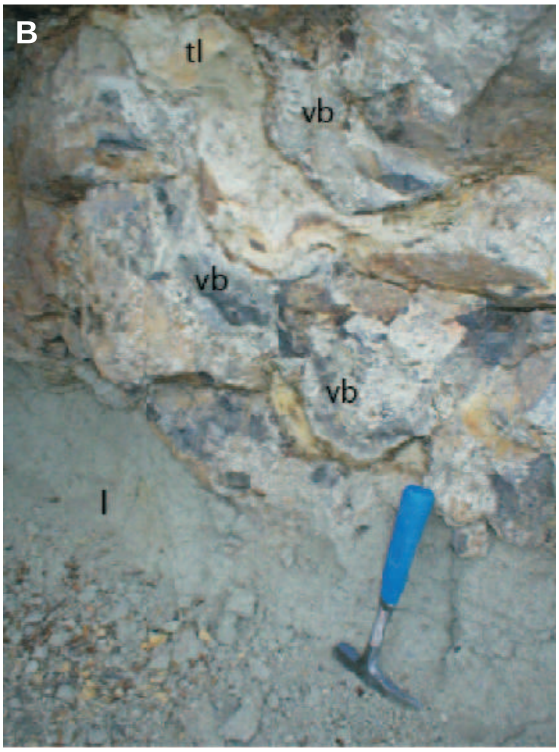
Thickly bedded, accidental lithic and xenocryst rich, fine grained tuff (f) in lensoidal bedding from the tuff ring sequence of Kis-Somlyó



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)

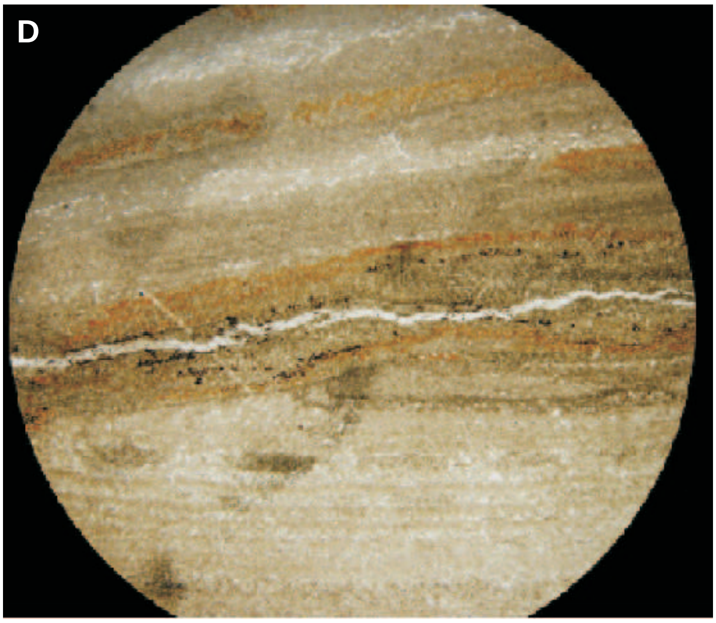


Siliciclastic lacustrine unit (l) that overly the pyroclastic (pyx) succession of Kis-Somlyó. The lacustrine unit is invaded by coherent pillowed basanite (p)



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"

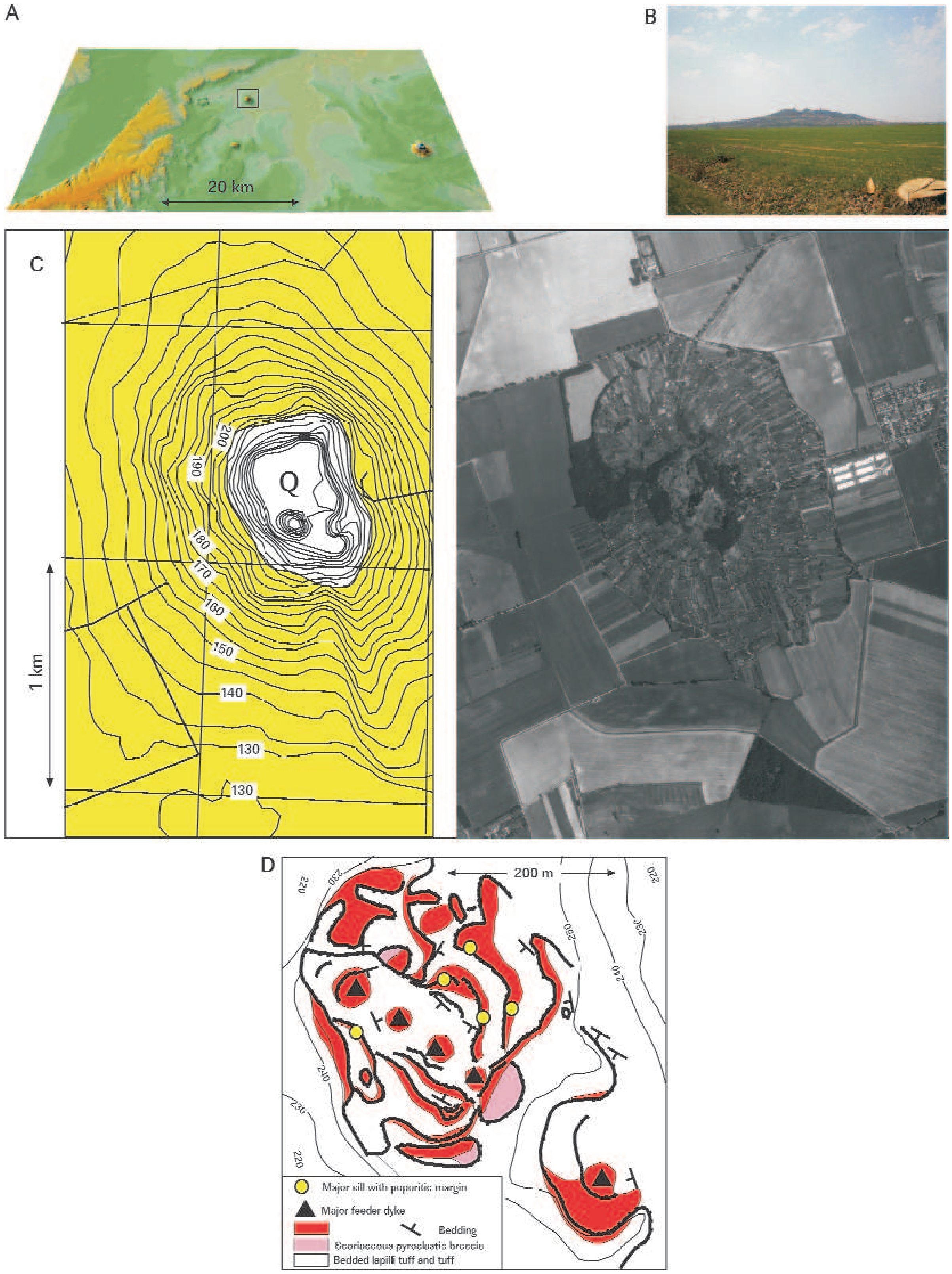
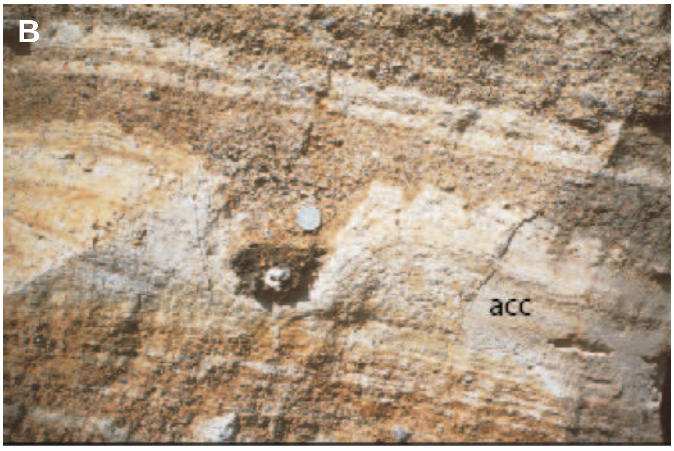
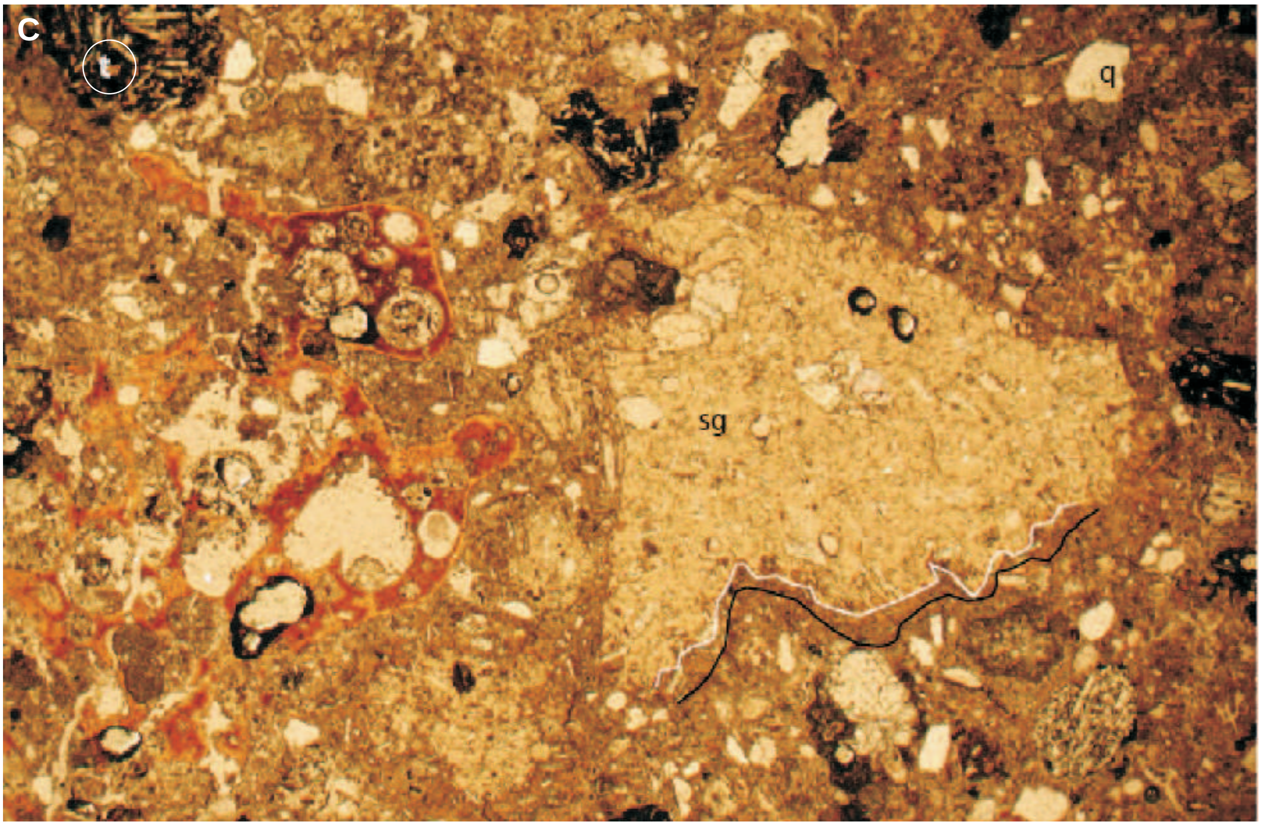


Plate 6 / Chapter 5

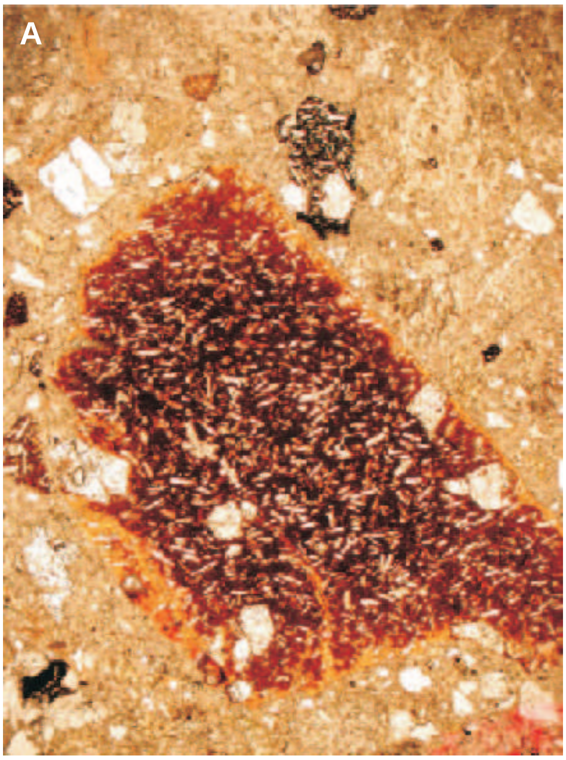
Basal phreatomagmatic pyroclastic succession at Ság-hegy



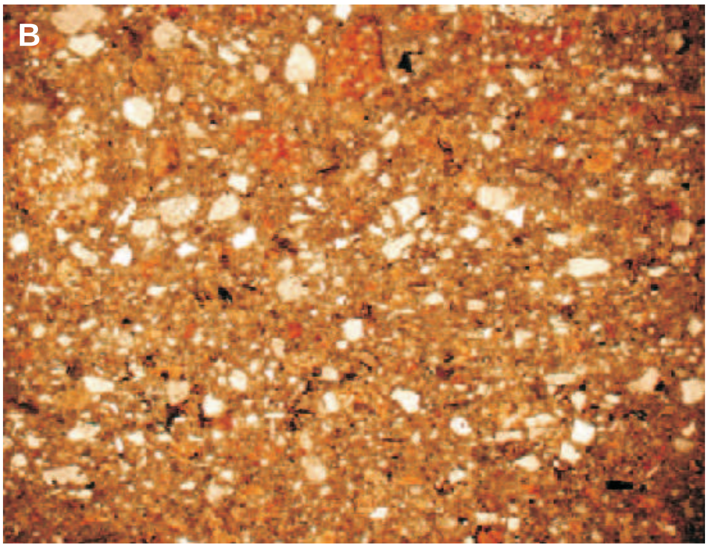
Accretionary lapilli bed (acc) in the upper pyroclastic succession in the north-western side of the erosion remnant of Ság-hegy. Note the deep impact sag caused by a cauliflower bomb (c)



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



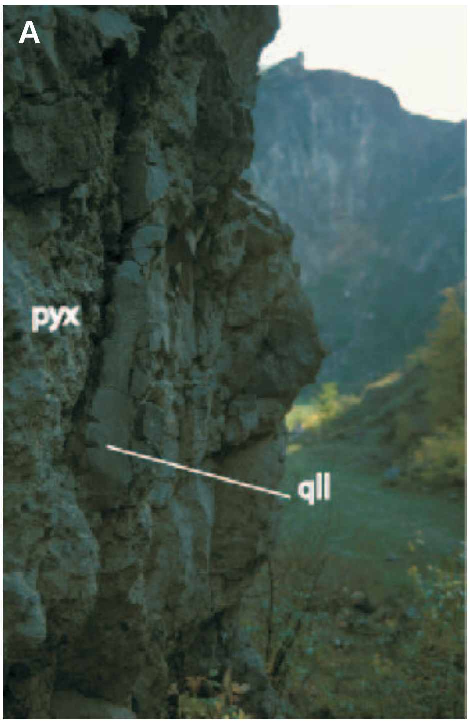
Photomicrograph (plan parallel polarized light) of microgabbroid volcanic lithic clast of the Ság-hegy pyroclastic beds. The short side of the photo is 2 mm



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

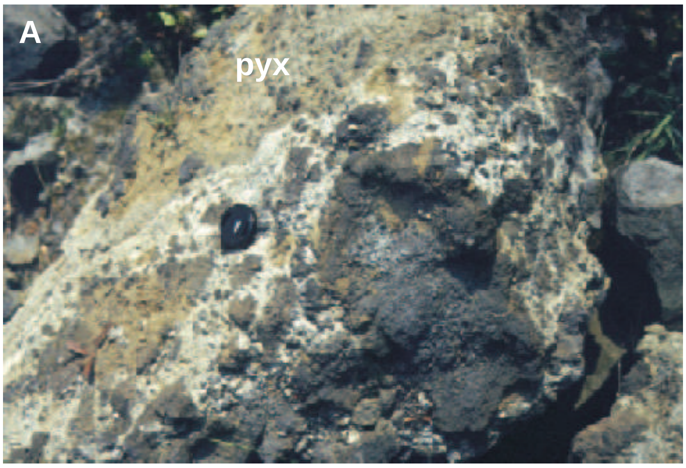


Preserved chilled margin (white line) contact of the former lava lake (qll) to the pyroclastic succession (pyx). The lava lake is inferred to have been occupied the crater/vent zone of the Ság-hegy tuff ring

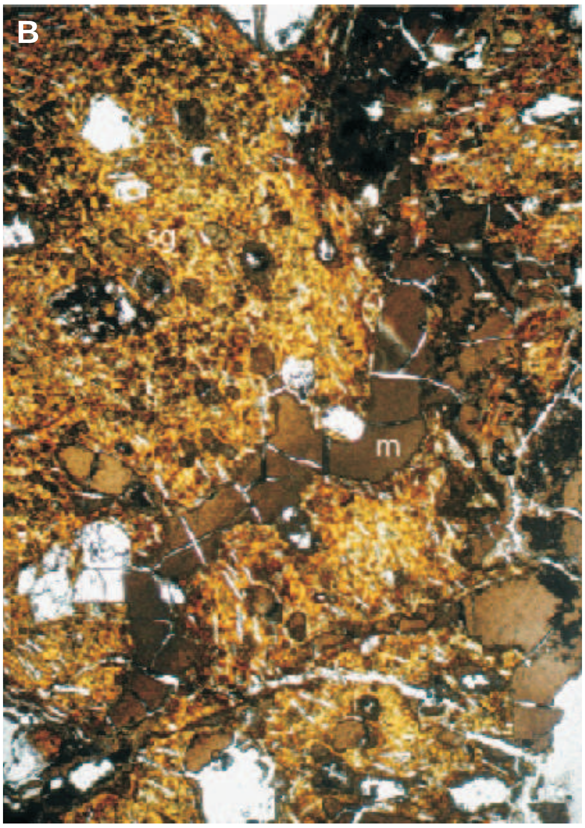
Small basanitic (dm-thick) lava protrusion (black line) that intruded into the tuff ring sequence interpreted to have been fed from the central coherent lava body filled the crater zone of the Ság-hegy tuff ring

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

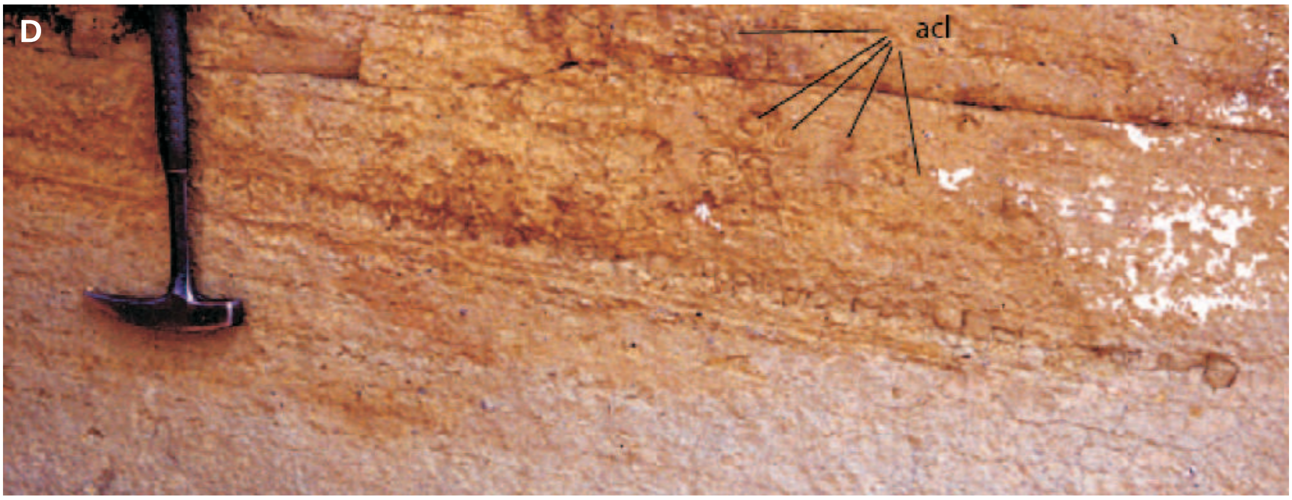


Fluidization (white region around dark basanite clasts, black lines) texture preserved along an intruded basanite body in the upper section of the tuff ring (pyx) sequence

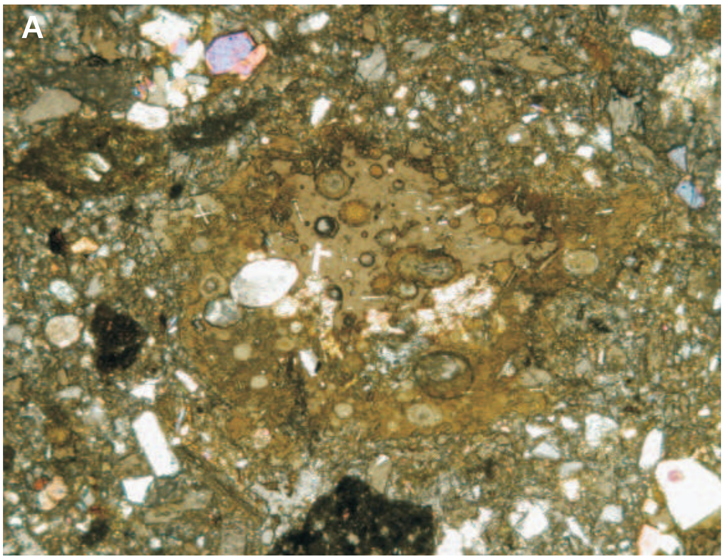


Photomicrograph (parallel polarized light) of a globular peperite from the margin of large coherent lava body at Ság-hegy. Note the smooth surface mud (m), inferred to have been boiled during the interaction of basanite melt (sg) and the host mud. The short side of the photo is about 2 mm

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)

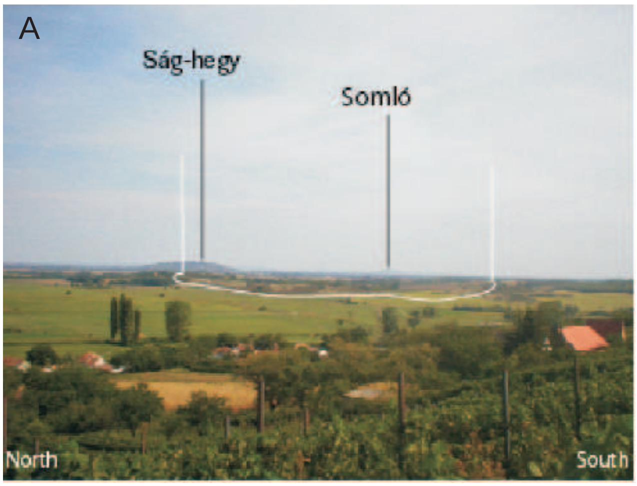


Photomicrograph of moderately vesicular, blocky sideromelane glass shard from the Egyházaskesző lapilli tuff indicative for phreatomagmatic explosive eruption [plan paralell light polarised]. Note the darker coloured rim, a result of advanced palagonitization front progressing inward. Shorter side of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

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

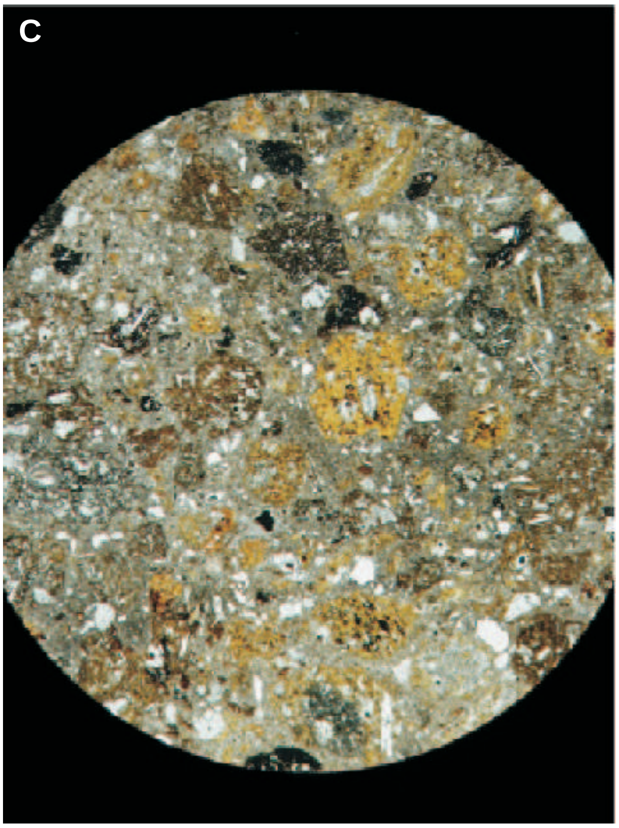
Shallow impact sag on the bedding surface of a base surge bed at Egyházaskesző

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

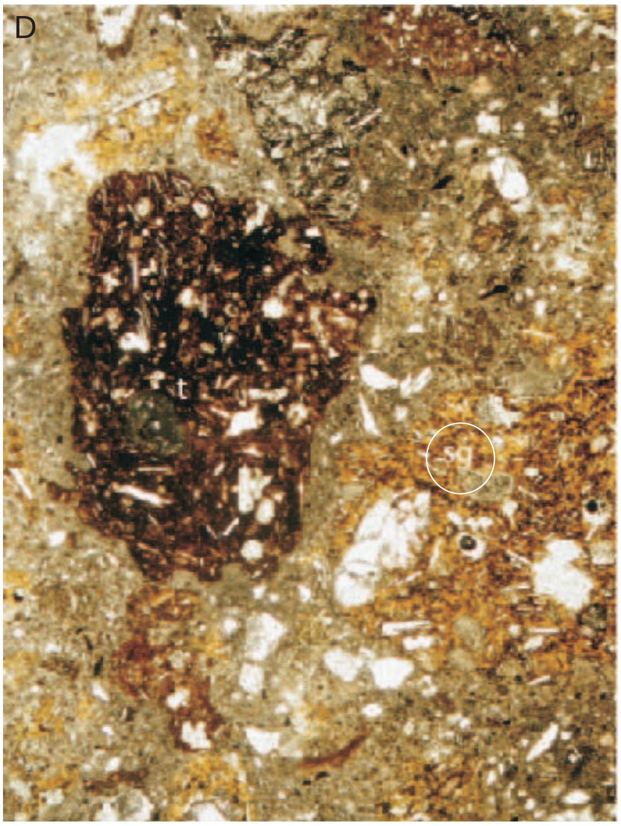


Panoramic view toward the pyroclastic hills east of the village of Gérce (rim marked by white curve line and straight lines point to the proposed rim) forming similar castle-like architecture to the Fort Rock tuff ring in Oregon (HEIKEN 1971)

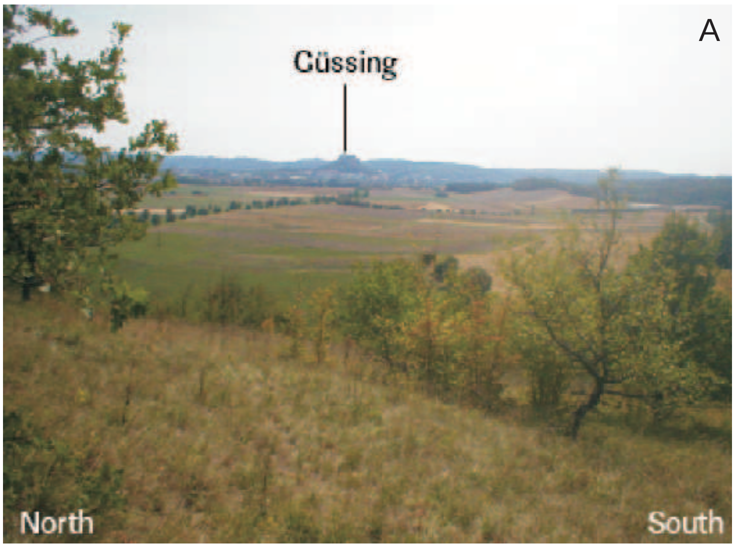
Panoramic view toward the Somló butte from south. Note the small hill on the top of the butte which consist of red, scoriaceous pyroclastic mound



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

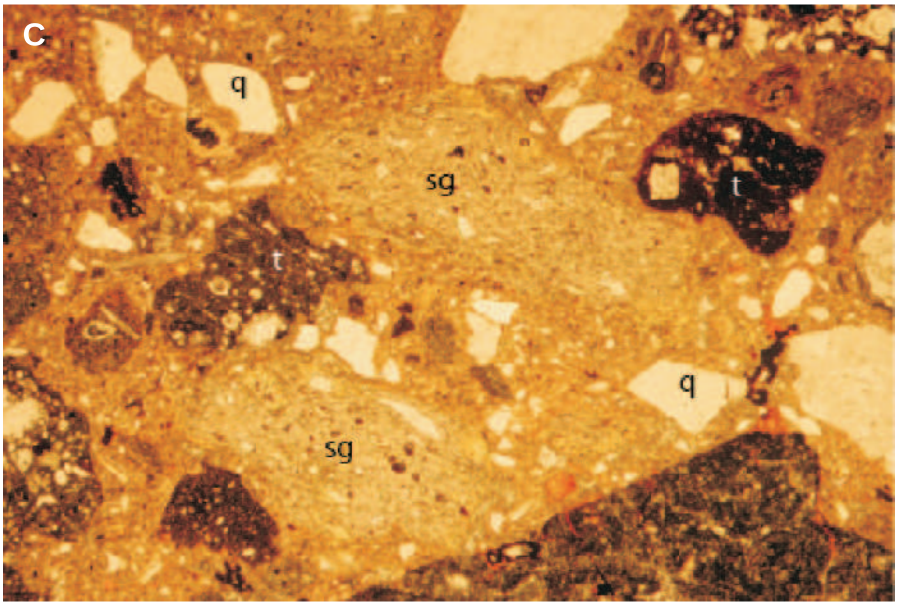


Photomicrograph (plan parallel polarised light) of a sideromelane glass (light colour "sg") and tachylite (dark colour "t") shard from the basal lapilli tuff and tuff succession of Somló. The short side of the photo is about 2 mm



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