World Geomorphological Landscapes

Dénes Lóczy *Editor*

# Landscapes and Landforms of Hungary

Springer

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*Authors dedicate this volume to the 60th birthday of the Editor, Prof. Dénes Lóczy*

# Series Editor Preface

Landforms and landscapes vary enormously across the Earth, from high mountains to endless plains. At a smaller scale, nature often surprises us by creating shapes which look improbable. Many physical landscapes are so immensely beautiful that they receive the highest possible recognition—they hold the status of World Heritage properties. Apart from often being immensely scenic, landscapes tell stories which not uncommonly can be traced back in time to tens of million years and include unique events. In addition, many landscapes owe their appearance and harmony not solely to natural forces. For centuries, and even millennia, they have been shaped by humans who have modified hillslopes, river courses and coastlines, and erected structures which often blend with the natural landforms to form inseparable entities.

These landscapes are studied by geomorphology—'the science of scenery'—a part of Earth Sciences that focuses on landforms, their assemblages, surface and subsurface processes that moulded them in the past and that change them today. To show the importance of geomorphology in understanding the landscape, and to present the beauty and diversity of the geomorphological sceneries across the world, we have launched a book series, *World Geomorphological Landscapes*. It aims to be a scientific library of monographs that present and explain physical landscapes, focusing on both representative and uniquely spectacular examples. Each book will contain details on geomorphology of a particular country or a geographically coherent region. This volume presents the geomorphology of Hungary—a country that may seem small but presents a remarkable diversity of landscapes, from vast plains to spectacular karst plateaus and ruined volcanic landforms. Nearly 30 case studies introduce the finest examples of geomorphology in Hungary, providing guidance to geoscientists as to where to go to enjoy the very best scenery.

*The World Geomorphological Landscapes* series is produced under the scientific patronage of the International Association of Geomorphologists (IAG)—a society that brings together geomorphologists from all around the world. The IAG was established in 1989 and is an independent scientific association affiliated to the International Geographical Union (IGU) and the International Union of Geological Sciences (IUGS). Among its main aims are to promote geomorphology and to foster dissemination of geomorphological knowledge. I believe that this lavishly illustrated series, which sticks to the scientific rigour, is the most appropriate means to fulfil these aims and to serve the geoscientific community. To this end, my immense thanks go to Prof. Dénes Lóczy—a long-standing supporter of the IAG activities and its past Secretary—for adding to his agenda the hard task of editing this volume and successfully coordinating the large team of authors. I hope he is as pleased with the final outcome as I am. I also acknowledge the excellent work of all individual authors who accepted to share their expert knowledge of the country with the global geomorphological community.

Piotr Migoń

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# 1 Introduction

Dénes Lóczy

The novelty of this publication is that this is the first which intends to be a modern introduction to the most interesting geomorphological sites of Hungary—at least in the English language. The idea of such a book has been long raised by several Hungarian geomorphologists (including the present editor), but none of the manuscripts has ever reached the stage of completion.

It cannot be claimed, however, that the book has no precedents at all. In the early decades of the 20th century, an eminent geomorphologist and our most prolific geographical writer, Cholnoky (1929), published numerous books, both academic and popular, with valuable geomorphological descriptions in his unimitable style, including a "Geography of Hungary". In order to enhance public knowledge on the earth sciences, a scientific guidebook entitled "Geological Excursions around Budapest" (Schafarzik et al. 1929) was published in the same year. Although much of the information contained is still valid today, no further update after the 3rd edition, issued exactly half a century ago, has come out.

To meet an increasing demand for geological guides presenting other parts of Hungary too, a popular series of geological guides (Budai et al. 2002) was launched by National Park directorates in cooperation with the Geological Institute of Hungary. The well-illustrated summary of spectacular sights in one of Hungary's most scenic landscapes, based on decades of detailed mapping in the Balaton Highland by the team of authors, was favourably received in the circle of geomorphologists, but the first volume printed in a relatively small number of copies has not been followed by further volumes. A new series of geological guides written by geologists of the Eötvös Loránd University has just begun to be published by Hantken Publishers (Pálfy and Pazonyi 2007; Palotai 2010). However, inventories of the key geological sections have never been supplemented with the descriptions of related landforms. The closest attempt was made by Juhász (1987) in his popular summary of earth history on the territory of Hungary and in a geological atlas for tourists (Budai and Gyalog 2010).

Although the publisher of agricultural books has been showing respectable steadfastness in regularly issuing monographs on the National Parks of Hungary since the 1970s, an interested reader can find very little and mostly outdated geomorphological information there. Likewise, earth sciences monographs (for instance, Karátson 1997; Mészáros and Schweitzer 2002) only rarely give some space for a brief description of typical geomorphological sites.

The most painful gaping hollow on the bookshelves of Hungarian geomorphologists, however, is the lack of a new regional geomorphology of Hungary, which could be the source for an English translation.

Under the above presented circumstances, it is a fortunate coincidence that the *Landscapes and Landforms* series was just recently launched by Springer publishers (modelled on the book *Geomorphological Landscapes of the World*—Migoń 2010) and the Hungarian efforts to the end of producing a comprehensive introduction of geomorphological sites in Hungary could be successfully channelled in this direction. This does not mean that the book is striving for completeness: the limitation of space only allows for the presentation of a *selection of sites*. From a class of landforms the most spectacular example is included and presented in some detail (Fig. 1.1), while others have to be 'satisfied' with only brief mentions.

Some chapters focus on classical examples of landforms, which regularly appear in the standard geomorphological/geographical textbooks, old and new (for instance, the dunes of blown-sand areas or oxbows on lowland floodplains). They are so typical for the country that simply cannot be left out. Others are exactly at the other end of the scale and, as a consequence, may be even more interesting: they are relatively unknown even to Hungarian geographers, such as the karst features under and around the basalt lava cap of Kab Mountain. These are mostly presented here by the teams of researchers who first described them.

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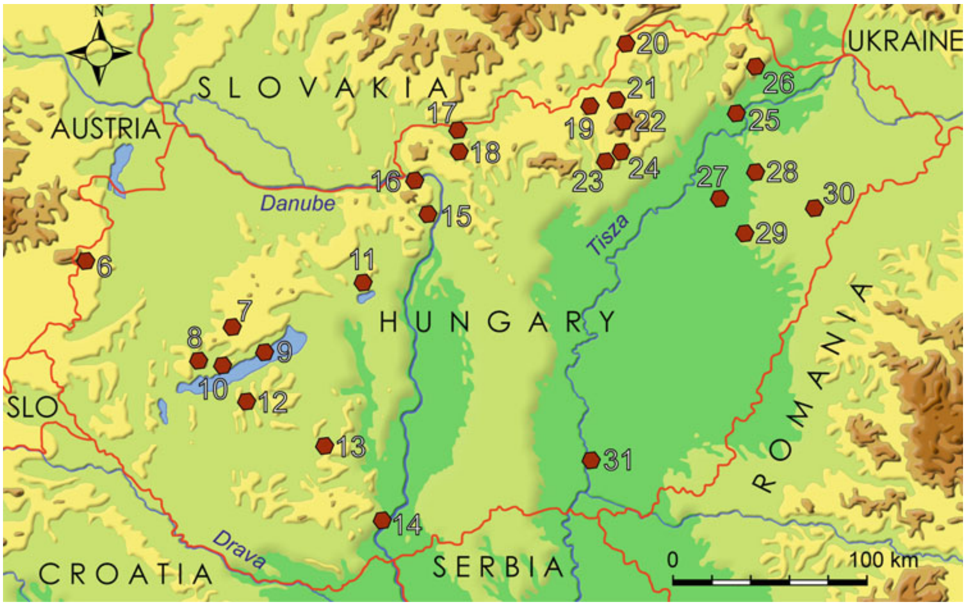


**Fig. 1.1** Locations of the geomorphological sites presented in the volume. Chapter numbers: *6* Kőszeg Mountains; *7* Kab Mountain in the Bakony Mountains; *8* Tapolca Basin; *9* Tihany Peninsula; *10* Kál Basin; *11* Velence Mountains; *12* Somogybabod gully in the Somogy Hills; *13* Kapos Valley; *14* Dunaszekcső loess bluff; *15* Caves in the Buda Mountains; *16* Danube Bend; *17* Medves region; *18* Kazár in the Nógrád Basin; *19* Vajdavár Hills; *20* Baradla–Domica cave system; *21* Uppony Hills; *22* Bükk Mountains; *23* Egerszalók; *24* Beehive rocks in the Bükk Foothills; *25* Tokaj Hill; *26* Megyer Hill in the Tokaj Mountains; *27* Hortobágy puszta; *28* Nagyhegyes Crater Lake; *29* Lyukas Mound; *30* Southern Nyírség; *31* Mártély oxbow

The question arises: how ‘final’ are the conclusions on the origin of the individual landforms? We can never be absolutely sure about the validity of our interpretations, but while the development of some landforms seems not to present a problem any more, the origin of others is still unresolved. The stages and chronology of the evolution of the picturesque Visegrád Gorge (Danube Bend) has been and remains to be among the great puzzles geologists and geomorphologists keep trying to solve time after time. We do not know who carved the niches, certainly anthropic features, into the “beehive rocks” or built the numerous tumuli. At any rate, the volume reflects the state-of-the-art explanations of geomorphological features of variable size and age. The age of landform is probably an issue still open to debate in many cases since novel geophysical dating methods may change the views of researchers at this point even overnight.

The lack of space has prevented most of the authors to include cultural geography in more detail. In some chapters, however, in addition to the preservation of geoheritage, some insight is also provided into the social, cultural and economic significance of landforms.

The sites are arranged geographically. Starting in the west, the reader is guided across Transdanubia to Budapest and then on to northern Hungary and the Great Hungarian Plain. Although very few of the landforms presented in this volume are unique to Hungary, the editor hopes that the curiosity of the reader will be satisfied and his/her interest in the topical issues of the geomorphology of Hungary maintained.

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# Part I

# Physical Environment

# Geological and Tectonic Background

# 2

János Haas

**Abstract**

The Pannonian region (the Pannonian or Carpathian Basin and the surrounding mountain ranges) is part of the Mediterranean Mountain System, which was formed during the last plate-tectonic cycle since the latest Paleozoic times. In Europe it is an about 300–800 km wide belt (Neo-Europe) accreted to the previously consolidated parts of Europe (Hercynian/Variscan Europe or Meso-Europe) as a result of the Alpine orogeny caused by convergence of the European (Eurasian) and African Plates. The present-day geological structure of the region is mostly determined by the evolution of the Tethys and Atlantic Ocean systems, i.e. the dismembering of the European and African continental plate margins during the early evolutionary stages and their tectonic deformation and uplifting as consequences of plate and microplate collisions. Plate-tectonic processes led to the formation of the large Pannonian Basin in the Late Cenozoic times. Hungary lies in the central part of the Pannonian Basin that is actually a system of several basins separated by isolated ranges of Palaeozoic and Mesozoic, sedimentary, magmatic and metamorphic formations and Cenozoic sedimentary and igneous rocks.

**Keywords**

Plate tectonics • Pannonian Basin • Tisza Megaunit • ALCAPA Megaunit

## 2.1 Geological Evolution of the Pannonian Region

The pre-Neogene structure of the Pannonian (Carpathian) Basin exhibits a complex mosaic-like collage structure (Fig. 2.1). The basement is divided into two large structural units by the Mid-Hungarian Lineament trending east-north-east to west-southwest (Csontos et al. 1992; Fodor et al. 1999; Haas and Kovács 2001; Schmid et al. 2008; Haas et al. 2010). These large units, namely the Tisza and Alpine-Carpathian-Pannon (ALCAPA) Megaunits, show markedly different geological features and evolution history (Fig. 2.2).

The Tisza Megaunit consists of blocks accreted during the Late Paleozoic Variscan (Hercynian) orogenic phases, when it formed a part of the European Variscan Belt. It separated from this belt in the Middle Jurassic and since the late Early Cretaceous it has moved as a separate entity, i.e. a microcontinent. The Variscan crystalline complexes are covered by Upper Paleozoic continental siliciclastic series, and continental, and shallow-marine Triassic formations (Bleahu et al. 1994; Haas 2012). The Jurassic is typified by facies diversification with coal-bearing and then neritic and deep-marine siliciclastic formations in the Lower and lower Middle Jurassic and deep-sea cherty limestones in the upper Middle and Upper Jurassic in the Mecsek Zone and condensed swell facies in the Villány-Bihor Zone. The Lower Cretaceous is characterised by basic volcanites and conglomerates and sandstones derived from the coeval volcanic rocks in Mecsek Zone whereas shallow-marine limestones prevail in the Villány-Bihor Zone (Haas and Péró 2004). The Paleozoic–Mesozoic series of the Tisza

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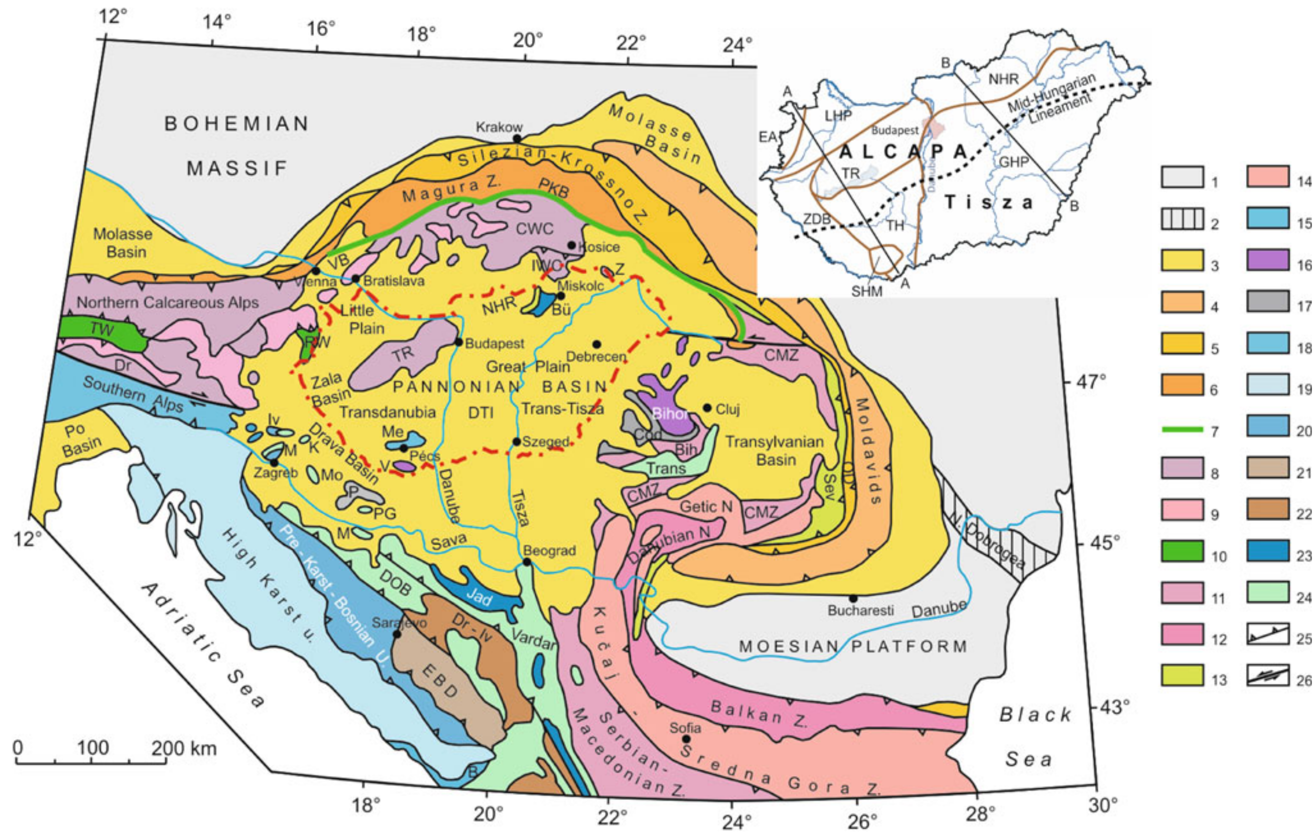


**Fig. 2.1** Major structural units of the Carpathic–Balkan–Dinaric region (after the maps by Schmid et al. 2008; Zagorchev 1994, Dimitrijević 1997). *1* Precambrian–Paleozoic platforms; *2* North Dobrogea Unit; *3* molasse basins; *4–7* Carpathian Flysch Zone: *4* Moldavides; *5* Silesian–Krossno Zone, Outer Dacides (*OD*); *6* Magura Zone; *7* Pieniny Klippen Belt (*PKB*); *8* Upper Austroalpine Unit, Transdanubian Range Unit, Fatric, Hronic and Silicic Units; *9* Lower Austroalpine Unit, Tatric, Veporic and Gemeric Units; *10* Penninic Unit; *11* Crystalline-Mesozoic Zone (*CMZ*), Serbian-Macedonian-Rodope Zone, Biharia Unit (*Bih*); *12* Danubian Nappes, Balkan Zone; *13* Severin Nappe (*Sev*); *14* Getic Nappes, Kučaj-Sredna Gora Zone; *15* Mecsek Zone (*Me*); *16* Villány (*V*)–Bihor Zone; *17* Papuk (*P*)–Codru (*Cod*) Zone; *18* Southern Alpine Units; *19* High karst Unit; *20* Pre-Karst–Bosnian Unit; *21* East Bosnian–Durmitor Unit (*EBD*); *22* Drina–Ivanjuca Unit (*Dr-Iv*); *23* Jadar Unit (*Jad*), Bükk Unit (*Bü*); *24* Vardar Zone, Transylvanian Nappes (*Trans*), Dinaridic Ophiolite Belt (*DOB*); *25* overthrust; *26* strike-slip fault. *Tw* Tauern window; *Rw* Rachnitz window; *Dr* Drau Range; *K* Kalnik; *Iv* Ivanscica; *Mo* Moslavačka Gora; *PG* Požekša Gora; *Z* Zemplén Mountains. *Inset map* Megaunits of Hungary with main lineaments and sites of profiles in Fig. 2.3. *AA* and *BB* sites of profiles. *EA* East-Alpine; *LHP* Little Hungarian Plain; *TR* Transdanubian range; *ZDB* Zala and Drava Basin; *TH* Transdanubian Hills; *NHR* North Hungarian range; *GHP* Great Hungarian Plain; *SHM* South-Hungarian Mountains

Megaunit are exposed in the Papuk Mountains in Croatia, in the Mecsek and Villány Mountains in South-Hungary and in the Apuseni Mountains in Romania. They were also recognised in a great number of wells in the basement of the Pannonian Basin.

Parts of the ALCAPA Megaunit constitute the basement of the northwestern segment of the Pannonian Basin as a continuation of the Austroalpine units exposed in the Eastern Alps (Fig. 2.2). The Transdanubian Range Unit is considered as an Upper Austroalpine-type nappe (Fodor et al. 2003; Tari and Horváth 2010). It is built up of a Lower Palaeozoic low-grade metamorphic complex, Permian fluvial sandstones and Triassic shallow marine sedimentary formations, mostly shallow-marine carbonates. The mostly deep-marine Jurassic–Lower Cretaceous formations are overlain by continental to deep-marine sediments formed by tectonically controlled transgression-regression cycles from the Late Cretaceous to the Paleogene (Haas 1991, 2012).

The Mid-Hungarian Zone is a narrow belt at the southern margin of ALCAPA containing strongly sheared, displaced elements of the South Alpine and Dinaridic origin. The Bükk Unit, exposed in the Bükk Mountains, Northeast-Hungary, is considered as part of the Mid-Hungarian Zone. It is composed of low-grade metamorphosed Upper Paleozoic to Jurassic sedimentary and volcanic rocks, which are overthrust by mélange with fragments of the Neotethys accretionary complex (Haas and Kovács 2001).

Large-scale strike-slip movements and coeval opposed rotation of the megaunits led to the juxtaposition of the basement units during the Early Tertiary (Csontos et al. 1992; Fodor et al. 1999; Csontos and Vörös 2004). These motions were controlled by indentation of the Adria

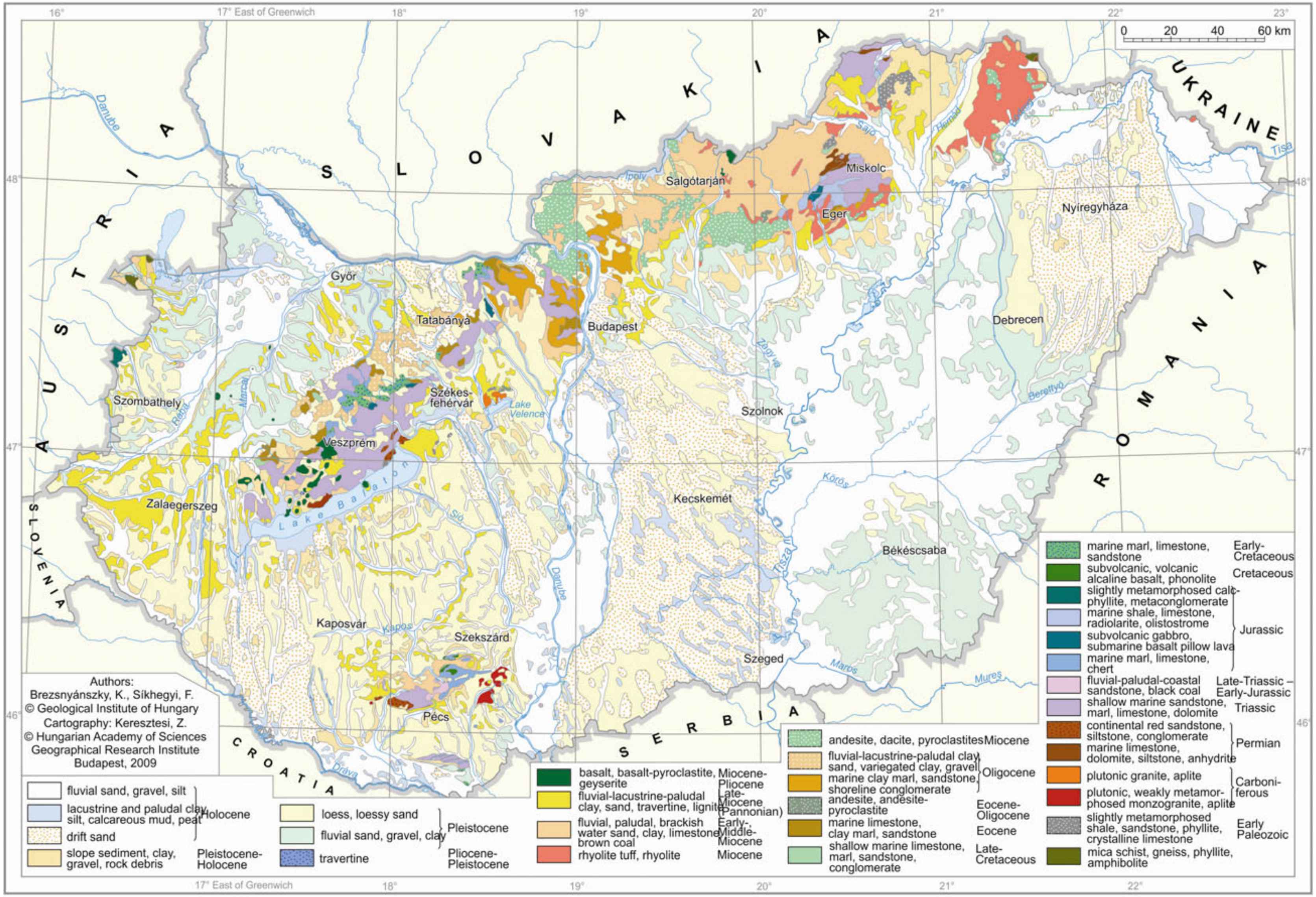


**Fig. 2.2** Geological map of Hungary (Brezsnyánszky and Síkhegyi 2009)

Microplate and rollback of the subducting slabs along the European Plate margin which led to the formation of a young basin system through crustal thinning beneath the area (Horváth et al. 2006). The phase that formed the Pannonian Basin was initiated by attenuation of the crust, leading to intense volcanism and significant but uneven subsidence during the Miocene. An andesite-dacite stratovolcanic chain, sub-parallel to the Carpathian arc, was formed 17–12 Ma ago (Harangi et al. 2007). It was followed by intense subsidence and sedimentary upfilling of the basins between 11.5 and 5 Ma. In the meantime, due to uplift of the Carpathian arc the previous connection with the Black Sea had ceased to exist and a large lake (Lake Pannon) took shape (Magyar et al. 1999). Parallel to the intense subsidence, basalt volcanism started in some parts of the Pannonian Basin ca 8 Ma ago (Németh and Martin 1999; Harangi 2001). Sediments derived from the rising Alps and Carpathians gradually filled up the lake, step-by-step through advancing deltas (Juhász et al. 2007; Magyar et al. 1999). By the Pliocene, a fluvial-lacustrine system with large swamps and wetlands replaced the lake. An intense uplift of Transdanubia, the western part of the Danube-Tisza Interfluve, and of the present-day mountains began 5 Ma ago, whereas the subsidence of the deep basins continued, giving rise to the deposition of thick fluvial sediments during the Pleistocene.

The main stages of the evolution of the geological structure of the Pannonian region can thus be summarized as follows:

1. Pre-Alpine, mostly Variscan, evolution that determined the geological structure of the plate margins at the beginning of the Alpine plate-tectonic cycle. Large fragments of the Variscan Belt became dismembered from the margins and incorporated into the Alpine orogenic system.
2. The early stage of the Alpine plate-tectonic cycle is characterised by the opening of oceanic basins: the western Neotethys Ocean from east to west from the Middle Triassic to Early Jurassic and the Penninic branch of the Atlantic Ocean from west to east from the Middle Jurassic to Early Cretaceous.
3. The stage of the mountain building processes that was the consequence of closure of the Neotethys basins from the Middle Jurassic to the Late Cretaceous–earliest Tertiary and of the Penninic basins ('Alpine Tethys') from the early Late Cretaceous to the Early Miocene.
4. Development of molasse basins in the foreland of the Alpine nappe stacks and in backarc setting related to the subduction of the European Plate in the Late Cenozoic.

## 2.2 Regional Geological Units

### 2.2.1 East Alpine Ranges in Western Hungary

Metamorphosed Palaeozoic and Mesozoic complexes representing the continuation of the East Alpine ranges are exposed in the northwestern corners of Hungary, in the Sopron and the Kőszeg Mountains, along the Austrian border. The Sopron Mountains consist of mica-schist and gneiss formations which can be correlated with the Raabalpen "Grobgneiss" Complex of the Lower Austroalpine (East Alpine) Nappe System (Fig. 2.3a). The history of metamorphism for these rocks commenced probably during the Caledonian orogeny, and continued during the Variscan phases. They were also affected by shearing in connection with the Alpine nappe movements (Lelkes-Felvári et al. 1984).

The rocks of the Kőszeg Mountains (belonging to the Rechnitz Window) and to Vashegy Hill (forming an independent window) appear from beneath the Austroalpine nappes in two blocks. There is a metasediment complex of Jurassic age, made up of quartz-phyllite and calcareous phyllite with coarse-grained meta-conglomerate bodies, and an ophiolite complex with serpentinised ultramafic–metagabbro–greenschist and blueschist rock associations of Early Cretaceous age. Both complexes were subject to very low to low-grade Alpine metamorphism (Koller 1985).

After a long period of continental erosion, the deposition of fluvial clastics and coal-bearing lacustrine clayey, silty, sandy sequences initiated in small basins of the Sopron Mountains during the Early Miocene. The Middle Miocene (Badenian) transgression led to the deposition of conglomerates followed argillaceous sedimentation under relatively deep marine conditions. The upfilling of the basins was reflected in the deposition of shallow marine biogenic limestones. The late Middle Miocene (Sarmatian) sedimentation is characterised by coarse-grained delta facies and the deposition of gravelly and sandy sediment continued during the Late Miocene lacustrine stage (Pannonian).

### 2.2.2 Little Hungarian Plain

The Little Hungarian Plain Basin (Danube Basin in Slovakia) is a large sub-basin of the Pannonian Basin system (Fig. 2.3a). The Lower and Upper Austroalpine nappes and Paleozoic and Mesozoic formations of the Transdanubian Range unit form the basement of the basin on the territory of Hungary. Parallel with intense subsidence, deposition of terrestrial clay, breccia and conglomerate began in the central part of the basin during

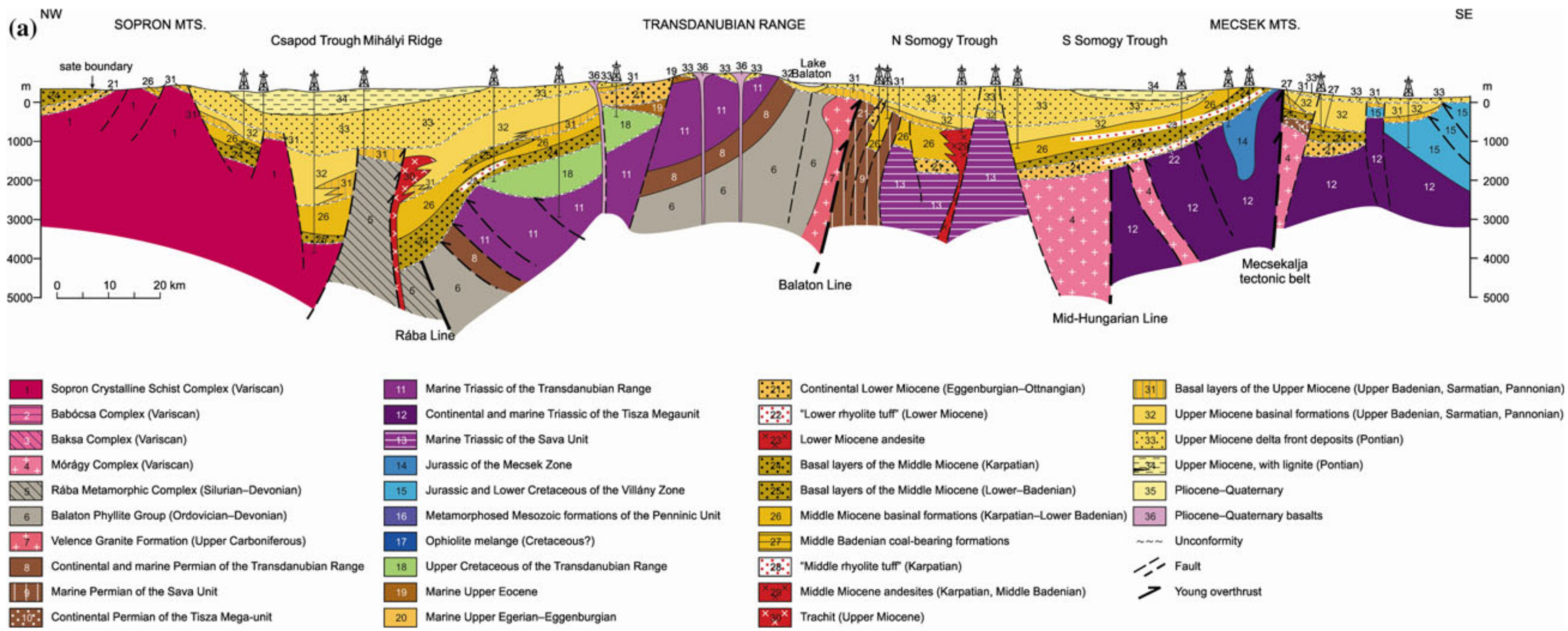


**Fig. 2.3** Profiles across Transdanubia (**a**) and eastern Hungary (**b**) (after Haas 2012)

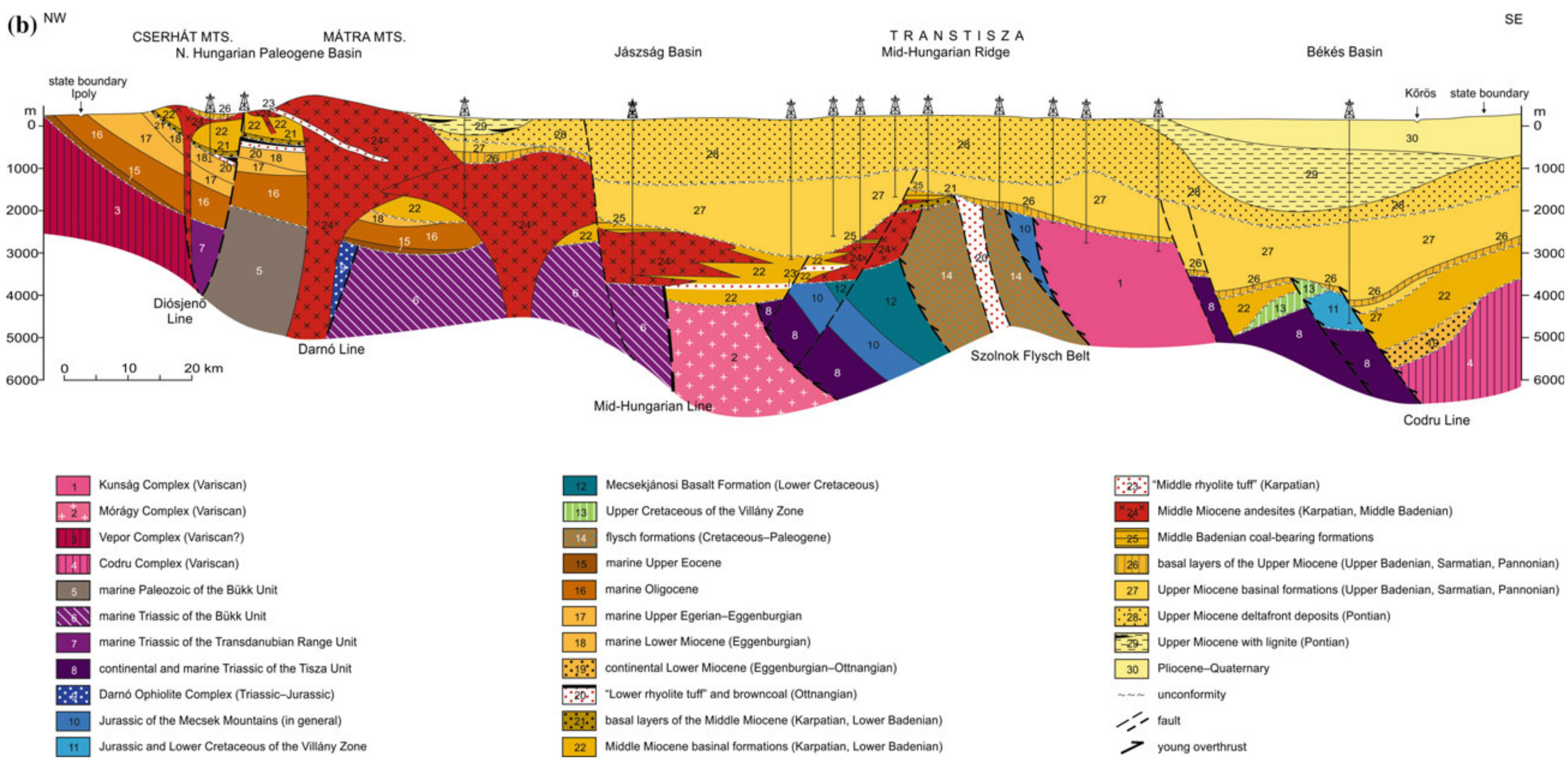


**Fig. 2.3** (continued)

the Middle Miocene (Karpatian, Early Badenian). It was followed by deposition of marine sandstone and limestone along the basin margins and clayey sediments in the deeper inner parts during the Badenian, while clayey marl, sandstone and tuffitic sandstone in the late Middle Miocene (Sarmatian) (Hámor 2007). Intense alkaline volcanic activity took place in the centre of the basin in the late Middle to early Late Miocene (Late Badenian to Early Pannonian) interval (Harangi 2001). At the beginning of the Late Miocene, i.e. in the Early Pannonian (11–12 Ma) lacustrine conditions already prevailed (Magyar 2010). Clayey sediments were deposited in the deeper sub-basins, whereas along the basin margins (Sopron Mountains, Transdanubian Range) and around elevated blocks (e.g. Mihályi High) conglomerates were formed from local sources. Then the southeastward prograding large Pannonian-delta reached the area between 9.7 and 9.0 Ma ago (Magyar 2010) and led to the deposition of predominantly sandy sediments in a thickness of several 100 to more than 1,000 m. It was followed by fluvial–lacustrine sedimentation of sand, sandstone, siltstone and variegated clay locally with lignite intercalations. The thickness of the Late Miocene to Pliocene fluvial succession may exceed 1 km. In the Pliocene, ca 5 Ma ago, explosive magmatic activity emerged at some sites of the Little Plain (Kemenesalja) that resulted in the formation of maar complexes and related tuff rings (Németh and Martin 1999 see Chap. 7). Intense subsidence of the area continued in the Quaternary; the thickness of the fluvial Quaternary formations reaches 450 m in the Győr Basin (Gábris and Nádor 2007).

### 2.2.3 Transdanubian Range

The Transdanubian Range, extending for 250 km in a NE–SW direction, consists of hills and mountains with a great variety of geological components (Fig. 2.3a). Lower Palaeozoic phyllite and carbonates are known north of Lake Balaton (Balaton Highland), while Early Permian granite makes up a great part of the Velence Hills located northeast of the Balaton. In a narrow belt north of Lake Balaton Lower Paleozoic rocks are overlain by Permian terrestrial red conglomerate and sandstone and a Lower Triassic succession of alternating marl, siltstone, limestone and dolomite formations. Other parts of the Transdanubian Range (Keszthely, Bakony, Vértes, Gerecse, Pilis and Buda Mountains) are built up mainly of Triassic dolomites and limestones of 2–3 km in thickness (Fig. 2.4a—Haas 2012). The Jurassic and Lower Cretaceous formations occur in the central zone of a large syncline, which was created by compressional tectonic movements in the middle Cretaceous. The Jurassic sequences are characterised by red pelagic limestones and in the Middle to Upper Jurassic segment also by radiolarites. The Lower Cretaceous is mostly represented by cherty deep marine limestones in the Bakony and marls, sandstones and conglomerates also of deep sea facies in the Gerecse Mountains. The orogenic processes in the middle Cretaceous were also manifested in uplifting, subaerial exposure of a significant portion of the Transdanubian Range and deposition of karstic bauxites in the Bakony, followed by a transgression and accumulation of marine limestones and marls of several 100 m in thickness. The next deformation phase during the Late Cretaceous caused even more intense denudation, karstification and deposition of bauxites of commercial value. The older Mesozoic formations are unconformably covered by continental and shallow marine Upper Cretaceous formations in the western Transdanubian Range (Bakony). General uplifting took place after the Cretaceous and created terrestrial conditions all over the Transdanubian Range area, intense karstification and bauxite deposition (Haas 2012). After a long continental period marine transgression resumed only in the Middle Eocene and deposited coal seams, and mostly clayey formations in the basins and carbonates in the shallow marine margins. Subsequent to the Early Oligocene uplift and erosion, fluvial sedimentation prevailed over the predominant part of the area during the Late Oligocene, whereas in the northeastern Transdanubian Range (Buda Mountains), which belonged to the North Hungarian Paleogene Basin, mostly clayey sediments were deposited from the Late Eocene to the Oligocene in a relatively deep marine basin (Nagymarosy 1990). In the Miocene the overwhelming area of the Transdanubian Range was exposed but several sub-basins formed with the deposition of shallow marine gravelly–sandy sediments and limestones. During the early Pannonian times the range constituted a large peninsula surrounded by shallow lacustrine environments with gravelly-sandy sediment deposition from local sources near the shore and clayey sedimentation in the deeper offshore zones. Later on, as a result of rising lake level, a significant portion became inundated (Magyar 2010). The basaltic volcanic activity in the South Bakony–Balaton Highland area began in the latest Miocene (7.9 Ma—Balogh and Németh 2005) and continued till the Late Pliocene.

### 2.2.4 Zala and Drava Basins

The Northern and Southern Zala and the Drava Basins are located in Southwest-Hungary and extend over the territory of Slovenia and Croatia. The basins began to take shape during the Lower to Middle Miocene and became sub-basins of Lake Pannon during the Late Miocene. Reflecting their different pre-Miocene evolution, the basements of the basins are significantly different. The basement of the Northern Zala Basin belongs to the Transdanubian Range structural unit. Accordingly, it is composed of Paleo-Mesozoic and Paleogene formations akin to those in the western part of the Transdanubian Range. The Miocene sequence begins with


**Fig. 2.4** Widespread rock types of geomorphological significance in Hungary. **a** Triassic (Dachstein) limestone slope at Kesztölc in the Pilis Mountains, Transdanubian Range (Photo by János Haas); **b** Triassic dolomite cliff in Veszprém, Bakony Mountains, Transdanubian Range (Photo by János Haas); **c** Main Conglomerate of Jakab Hill, Mecsek Mountains (Photo by János Haas); **d** Loess bluff of the Drava River at Heresznye, Inner Somogy Hills (Photo by Dénes Lóczy)

Karpatian to Badenian conglomerates or sandstones overlain by silt and clay with marine fossils; the Sarmatian is represented by marl and sandstone. The thickness of the Middle Miocene succession may exceed 1,000 m (Hámor 2007). Clayey marl of pelagic lake facies was formed in the early times of the Late Miocene. This sub-basin was approached by the south-eastward prograding delta 8.9–8.6 Ma ago (Magyar 2010), leading to the deposition of a 1 km thick sand-dominated succession. It was followed by the deposition of a fluvial–lacustrine series, ca 1.5 km in thickness.

The basement of the Southern Zala Basin is assigned to the Mid-Hungarian Zone. In a belt south to the Balaton Lineament displaced fragments of the South Karavanken and Julian–Savinja Units consisting of Permian and Triassic formations form the basement. Further southward slightly metamorphosed Triassic and Jurassic rocks and ophiolite mélange of the Kalnik Unit were encountered below the Miocene formations (Haas 2012). In the depocenter above terrestrial conglomerate about 3 km thick Middle Miocene (Badenian) marine clayey marl was formed. This sub-basin may have been reached by the large Pannonian delta somewhat later, in the 8.6–8.0 Ma interval. Otherwise, the Pannonian succession is similar to that in the Northern Zala Basin.

In the Drava Basin a basement of Paleozoic medium-grade metamorphic rocks and Mesozoic carbonates was detected. The Neogene started with Lower Miocene non-marine conglomerates, sandstones and clayey marls in a thickness of 2 km. The Middle Miocene is represented mostly by marine marls and clays in the inner part of the basin and shallow marine carbonates along the basin margins. The Pannonian delta may have occupied this area ca

6.8 Ma ago. Above the 1 km thick delta-related lacustrine sequence 2.5 km thick fluvial-lacustrine series was formed. In the axial belt of the basin the thickness of the Quaternary is more than 250 m.

### 2.2.5 Transdanubian Hills

This area is located in southern Transdanubia between the Transdanubian Range and the Mecsek Mountains and bordered by the Dráva Basin to the southwest and the Great Hungarian Plain to the east. Late Paleozoic to Mesozoic predominantly carbonate formations of the Mid-Hungarian Zone and Paleozoic metamorphic complexes and Mesozoic formations of the Tisza Megaunit form the basement of the Neogene sedimentary sequences (Fig. 2.3a—Haas 2012). In the Early to early Middle Miocene a fluvial conglomerate and sandstone succession was accumulated in a remarkable thickness (0.8–1.2 km) in a tectonically controlled continental basin. Continuing subsidence led to transgression and establishment of open marine conditions during the Badenian, followed by shallow marine sedimentation in the Sarmatian. The Pannonian lacustrine-pelagic conditions were changed in this area in the 8.6–7 Ma interval due to the effect of the prograding delta system that led to deposition of ca 1 km thick turbiditic sandy sediments and a subsequent fluvial-lacustrine succession. However, in contrast to the Dráva Basin in this area the subsidence was followed by Pliocene and Quaternary uplift in the inversion phase of basin evolution (Horváth et al. 1988).

### 2.2.6 South Hungarian Mountains

Carboniferous granite is exposed in the southeastern foreland of the Mecsek Mountains (Mórágy Hills). 3–4 km thick Permian and Early Triassic continental red-beds and 600–700 m thick Middle Triassic carbonate sequences constitute the anticline of the Western Mecsek Mountains (Fig. 2.3a), Jurassic sedimentary sequences and Lower Cretaceous magmatic and sedimentary formations constitute the syncline of the Eastern Mecsek. The thickness of the Jurassic is more than 3 km in the south and decreases to 500 m northward. The succession begins with a coal-bearing formation that is covered by shallow to deep-marine marls and sandstones of remarkable thickness. A thin series of deep-marine limestones represents the Middle and Upper Jurassic. Intense volcanism dominated the Early Cretaceous evolution producing a ca 1 km thick complex of alkaline basalt and marine sandstone and conglomerate consisting mostly of reworked volcanic rocks.

Located south of the Mecsek Range, the Villány Hills have an imbricate structure consisting mainly of Mesozoic carbonates; Triassic shallow marine dolomites and limestones, a condensed and discontinuous marine Jurassic succession and a thick Lower Cretaceous shallow-marine limestone formation.

In the early Cenozoic times the area of southern Transdanubia was an emerged land subjected to erosion. In the Early Miocene a large continental basin developed in the northern foreland of the Mecsek where thick fluvial formations were formed and similar sequences occur in sub-basins within the Mecsek Mountains. In the northern part of the Mecsek, above the Mesozoic basement or the Miocene terrestrial and rhyolite tuff succession, andesitic subvolcanic rocks occur. The marine sedimentation began only in the Badenian, when shallow-marine limestones deposited directly upon the bedrocks. However, large parts of the Mecsek–Villány area were probably still in emerged position. The Pannonian sequence commences usually with basal conglomerates that are overlain by calcareous marl and clayey marl of open lake facies and followed by sand-dominated delta successions.

### 2.2.7 North Hungarian Range

The North Hungarian Range is very complex geologically. In a geological sense the western Cserhát Hills (the Naszály and other Mesozoic blocks in its environs) belong to the Transdanubian Range unit. In contrast, the Visegrád Mountains, on the western bank of the Danube, is a part of the North Hungarian Miocene volcanic range.

The oldest formations occur in the northeastern part of the region, in the Szendrő and the Uppony Mountains where slightly metamorphosed Palaeozoic shallow and deep marine sedimentary rocks—phyllites and carbonates—outcrop. The Bükk Mountains is built mostly of slightly metamorphosed Upper Palaeozoic to Jurassic series. The northern part of the mountain consists predominantly of Carboniferous to Permian shales and carbonates. The Bükk Plateau is constituted mostly of Middle to Upper Triassic shallow marine limestones. Jurassic shales and conglomerates of deep-marine basin and slope facies and large basalt and gabbro bodies prevail in the western portion of the mountains. These complexes are locally covered by a marine Palaeogene sequence. The Rudabánya and Aggtelek Hills are built up of nappes of the Inner West Carpathian unit. The Rudabánya Hills contain non-metamorphosed and slightly metamorphosed Triassic and Jurassic shales and carbonates. The Aggtelek Hills consist of Triassic rocks, mostly shallow-marine limestones (Haas 2012).

The western North Hungarian Range consists of Paleogene and Neogene sedimentary formations and Neogene volcanic rocks. Metamorphic complexes of the Central Carpathian Vepor unit and Mesozoic sequences of the Transdanubian Range and the Bükk units form the basement of the Cenozoic formations. In the Paleogene Basin the Cenozoic transgression led to the formation of shallow-marine limestone in the Late Eocene, followed by the deposition of deep-marine marls in the Early Oligocene. During the latest Eocene–Early Oligocene the upbuilding of an andesitic stratovolcano in the Eastern Mátra Mountains started. Deposition of deep-marine marls and siltstones continued in the centre of the basin whereas shallow marine sandstones were formed along the margins during the Late Oligocene–Early Miocene. Subsequent uplift resulted in the establishment of terrestrial conditions and intense erosion that was followed by the deposition of rhyolite tuff over large areas in the late Early Miocene (Ottnangian). Transgression in the early Middle Miocene (Badenian) led to the deposition of clayey marine sediments in the deeper basins and sandy sediments and biogenic limestones in the shallow marginal zones (Hámor 2007), accompanied by intense volcanism. The bulk of the andesitic volcanic rocks that make up the Visegrád, Börzsöny, Cserhát and Mátra Mountains were formed in the Badenian, 15–16 Ma ago (Fig. 2.3b—Harangi 2001). The thickness of the lava, volcanic breccia and tuff succession of the stratovolcanic complexes may reach 1–2 km. Pliocene to Quaternary (5.6–1.8 Ma) basalts occur the Karancs–Medves area (Nógrád–Gemer Volcanic Field) and in the northern Cserhát Hills.

The eastern section of the North Hungarian Range includes the Cserehát Hills, which consist mostly of Neogene sedimentary rocks, and the Tokaj Mountains that are built up of Neogene and Quaternary deposits and volcanic complexes. Here volcanism began in the Late Badenian (13 Ma) and a 1–3 km thick stratovolcanic complex of rhyolite, dacite, andesite and their pyroclastics accumulated in the course of several eruptions until the earliest Pannonian (10.5 Ma) (Harangi 2001).

### 2.2.8 Great Hungarian Plain

Formed by upfilling of a large Neogene basin of articulated basement topography, it is an extensive plain area extending far over the territory of Hungary. The basement of the Great Plain is heterogeneous (Fig. 2.3b). The southern portion is assigned to the Tisza Megaunit, whereas the northern part belongs to the ALCAPA Megaunit and they are separated by the Mid-Hungarian Zone. Mostly Variscan medium-grade metamorphic complexes consisting of gneisses and mica schists form the basement of the Tisza Megaunit. These complexes are locally covered by Mesozoic successions. Highly deformed, Cretaceous to Paleogene imbricate flysch sequences occur in the northernmost belt of the Tisza Megaunit. North of the Mid-Hungarian Lineament, predominantly Mesozoic carbonates were encountered under the Cenozoic sequences.

Controlled by the Middle Miocene extensional tectonics, very deep sub-basins and intrabasinal highs developed. In the former, thick (3–7 km) and nearly complete Middle Miocene to Pliocene successions accumulated (e.g. Jászság, Nyírség, Derecske, Makó, Békés Sub-basins) whereas in the latter the Middle Miocene formations are usually missing and the thickness of the Upper Miocene (Pannonian) to Pliocene deposits is less than 2 km (Fig. 2.3b—Nagymarosy 1981). After transgression in the Badenian islands and shallow to deep-sea environments were established. This general paleogeographical setting prolonged to the Late Miocene, parallel to changes in the salinity of water. As a result of intense fluvial transport two large delta systems developed in the Pannonian which led to gradual upfilling of the basin from to NW and NE to SE during the late Pannonian (8–5 Ma—Juhász 1991; Juhász et al. 2007; Magyar 2010) and coeval extension of the fluvial-lacustrine sedimentary environments. The thickness of the Quaternary sediments is usually more than 50 m, but in the still subsiding parts of the basin it may exceed 500 m (Gábris and Nádor 2007).

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# 3 Climate and Drainage

Dénes Lóczy

**Abstract**
Located in mid-latitude eastern Central Europe, Hungary has a moderately humid continental climate and, due to its situation in the Carpathian Basin, the basin effect is also observable. Regarding average conditions, the climate is rather humid than arid, water availability, however, is a decisive component of the climatic system and in most parts of the country drought is a recurring phenomenon. Since year-to-year or season-to-season variability is more remarkable than regional variations, long-term average values are not really informative as far as the climatic properties of individual regions are concerned. Warming and drying trends are predicted for the 21st century. The drainage system, developed since the Pliocene, is adjusted to two hydrographical axes: the Danube and the Tisza Rivers within the drainage area of the Danube. The water regimes of these rivers and their major tributaries depend upon the runoff conditions of their drainage basins: for the Danube first of all the Eastern Alps and for the Tisza River the Northeastern and Eastern Carpathians. The largest and best studied lake of Central Europe is the shallow Balaton. The single extensive reservoir was impounded on the Tisza River at Kisköre, mainly for irrigation purposes. As far as the groundwater resource is concerned, thermal, medicinal and mineral water reserves are particularly appreciated.

**Keywords**
Climatic regions • Basin character • Temperature • Precipitation • Winds • Runoff • Rivers • Lakes • Groundwater • Climate change

## 3.1 Introduction

It is difficult to overemphasize the significance of the climatic factor, i.e. past and present climatic conditions and events, in the physical environment in general and particularly in governing geomorphic processes. Equally important are rivers and groundwater conditions in landform evolution. Many of the geomorphological sites presented in this book owe their existence to a particular interplay between topography, climate and drainage in some period of geological history or even at present. This interconnectedness justifies that the climate and drainage of Hungary are treated jointly in this chapter.

## 3.2 Brief Climate History

To outline the climatic background to the geomorphic evolution of the Carpathian Basin (or the lands on its predecessor lithospheric plates), we have to reconstruct the conditions under which the surfaces now exposed were once formed. To this end, we have to reach back to the late Mesozoic and provide a brief overview of climate history since then—even if serious deficits exist in the data necessary for a more complete reconstruction.

Sedimentological evidence points to an extensive but generally shallow marine inundation in the Jurassic with equable and humid climate on the sporadic islands

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(Császár et al. 2008). Under the tropical warm and humid Cretaceous climate lateritic weathering provided favourable conditions for bauxite formation. The ore-bearing deposits were preserved in the depressions of the karst surface.

Recently, a short spell with rapid warming was identified in the sedimentological records of the Paleocene to Eocene transition (Paleocene/Eocene Thermal Maximum—Gingerich 2006). Subtropical to tropical ("hothouse") conditions survived well into the Eocene (Császár et al. 2008), when transgressions and regressions alternated in the area of the ALCAPA lithospheric plate. At the Eocene/Oligocene boundary (38 Ma ago) abrupt cooling is assumed. In addition to sedimentological evidence, it is also testified by a drastic drop in the level of world ocean. In the Oligocene a warm subtropical climate (mean annual temperature (MAT): 20 °C; annual precipitation (AP): ca 1,500 mm; both with marked seasonal variations) is assumed from the sporadic palynological data (e.g. the analysis of the Eger profile—Nagy 2005). In the Lower Miocene (Eggenburgian or Acquitanian) high temperatures (MAT: around 18 °C) and abundant rainfall (AP: 1,200–1,500 mm) with even distribution prevailed on the islands of the Carpathian archipelago (see also Chap. 4). Probably the dry season was longer than the wet period, but drought was moderate, possibly due to the proximity of the sea (Hably 1979). In the Ottnangian (Burdigalian) swamp forest conditions were typical with slightly lower temperatures (MAT: 16–17 °C), a marked cooler and drier season and probably somewhat lower (AP: 1,000–1,500 mm) precipitation (Nagy 2005). Simultaneously regional variations began to increase. In the largely marine Karpatian (Langhian) age subtropical climate dominated (MAT: 15–16 °C). Badenian (Serravallean) orogenic movements and volcanism (Chap. 4) further diversified the climate of the Carpathian Basin and in higher altitudinal zones temperate vegetation was also present (Hámor 2001). Tropical floral elements tend to disappear during the Sarmatian (Tortonian), while the surviving subtropical elements indicate connections towards the Near and Far East. The Carpathian Basin is in a transitional zone from the subtropical to the warm temperate zone with hot and dry summers and rainy winters (MAT: 14 °C, AP: 700–800 mm) close to the present-day values, also regarding its uneven distribution (Nagy 2005). Pannonian (Messinian) climate was expressedly warm temperate (MAT: 13 °C) of Mediterranean character, locally with particularly favourable conditions for vegetation growth. The existence of Lake Pannon attenuated climatic oscillations. In the mountain frame even conifer forests appeared.

During the complete upfilling of Lake Pannon, parallel to the Messinian Salinity Crisis, dry desert climate prevailed (AP: 150–250 mm) with wind action (Bérbaltavárian subage, 7–6 Ma ago). At the beginning of the Pliocene, the opening of the Gibraltar Strait restored water fill in the Mediterranean Sea (Császár et al. 2008; Schweitzer 2013). In the Ruscinian-Csarnótan subage (Zanclean-Piacenzan, 4.4–3 Ma ago) warm temperate (MAT: 12 °C) conditions with a summer dry season prevailed in the Carpathian Basin.

The Pleistocene began with warm and dry climate favouring pedimentation (Villányian, Calabrian, 3–1.8 Ma), followed by gradual cooling with loess and river terrace formation (Biharian, 1.8–1.2 Ma) (Schweitzer 2013). In the late Pleistocene cold and dry periods of periglacial climate alternated with warmer and wetter interglacials. The cyclicity of Quaternary climate is also detectable from an analysis of terrestrial deposits (e.g. loess-paleosol sequences or fluvial sequences from the Great Plain). Until ca 1 Ma ago the average length of the Milankovitch cycles was 40 ka and after that date 100 ka (Császár et al. 2008). At the peak of the last glacial (20–18 ka BP) MAT was −2 to 0 °C and AP 200–400 mm.

According to a recent Holocene chronology (Gábris et al. 2002), this age is subdivided into Bölling, Older Dryas, Alleröd, Younger Dryas and postglacial stages. In the last cold spell, the Younger Dryas (12.5–11.2 ka BP) July mean temperatures did not rise above 13 °C in the Carpathian Basin. Climatic amelioration (5–7 °C rise in MAT within millenia or even centuries) was a clear trend in the still cool Preboreal (11.6–10.2 ka BP; in July 18 °C) and in the Boreal stage (10.2–8.3 ka BP), when winters became mild. Among the stages the Atlantic (8.3–5.8 ka BP) stands out with its equable climate, probably warmer than today. Under the Boreal and Atlantic climates the hydrological cycle accelerated and this favoured fluvial erosion, which reached its maximum (Gábris et al. 2002). The wet and mild Subboreal stage (AP: 900 mm) was replaced by the moderately wet Subatlantic (AP: 750 mm) 2.5 ka ago, when human impact became decisive in the environmental history of the Carpathian Basin.

## 3.3 Climatic Influences and Regions

At present, Hungary lies in the temperate belt, in the zone of westerly winds, but in a relatively great distance from oceans. Its climate results from spatially and temporarily highly imbalanced Atlantic, Mediterranean and continental climatic influences, producing, on the whole, a moderate continental climate (Péczely 2009). The influence of Atlantic air masses is primarily manifested in milder winters, cooler summers and a more or less uniform distribution of precipitation (overwhelmingly rainfall) throughout the year. The most common area of origin for Mediterranean cyclones, the Ligurian Sea, lies even closer to the western borders of the country than the Atlantic Ocean: at a mere 600 km distance. The cyclones arriving from the Mediterranean bring rains in autumn and, with increasing frequency, Mediterranean air masses make winter weather milder—and sometimes more turbulent. The opposite can be said of the continental influence: the formation of the so-called Voeikov axis (a lasting coupling between the Azoran and Siberian high-pressure centres across the Carpathian Basin) is responsible for particularly severe and prolonged winters (like in 1928/1929, 1939/1940, 1941/1942, 1953/1954, 1963/1964, 1984/1985, 1986/1987, 2002/2003, 2012/2013). The difference in the mean temperature of the coldest month (January) between years of Atlantic and Siberian type may amount to 15 °C. In years of continental type anticyclonal conditions are also typical in the summer half-year, when air masses arrive from the Middle East (Persian Gulf) and bring hot and dry weather.

Comparing the climate of the Carpathian Basin with that of the neighbouring regions, the basin effect is clearly detectable (Pajtók-Tari 2012). The highest drought hazard is observed in the centre of the basin: the length of precipitation-free period may reach 50 days (particularly in the summer). In winter a cold "air pillow" may settle in the basin for longer periods and contributes to the accumulation of pollutants.

In the classification system proposed by Wladimir Köppen and Rudolf Geiger, most of Hungary belongs to the Cfb class (warm and uniformly wet temperate with long and relatively cool summers), while the southernmost part is Cfa (the same with hot summers) and the northeastern part is Dfb (uniformly wet boreal with long summers). According to Glenn Thomas Trewartha's modified classification, Hungary's climate is mostly Dcw (wet subtype of continental with a longer warm period) or DCb (emphasizing warm summers). According to Charles Warren Thornthwaite, by thermal efficiency the country belongs to the $B_3$ mesothermal type and by the moisture index to the $C_1$ (dry subhumid) areas. Instead of deciding on one or the other scheme, a more recent and simplified climatic regionalization (Bartholy and Weidinger 2002—Fig. 3.1) is presented here. The climatic regions, the boundaries of which are largely adjusted to topography, are the Great Plain (subdivided into three subregions), the Little Plain, the mountains and hills of Transdanubia (also subdivided into three subregions) and the North-Hungarian Mountains.

It has to be noted, however, that meso- or microscale variations in climate may surpass those observed at macrolevel. In spite of the limited elevation range in the topography of the country (maximum relief: 942 m), altitudinal differences in climate are considerable, being most pronounced in the North-Hungarian Mountains. For instance, the Bükk karst plateau (600–900 m) receives high precipitation (800–900 mm, usually also in the form of snow) and has cool summers (OMSz 2003). Extreme local climate is formed in valleys open to the north: a "cold pole" came about in the North Hungarian Mountains, at Zabar (see the location of the settlements mentioned in Fig. 3.1), with repeated daily negative temperature records, while somewhat to the west, in the Nógrád Basin drier and warmer climate prevails. In some regions (like on Mátra and Bükk foothills) abrupt change from mountain to lowland climate is observable. In the generation of local variations human impact is often decisive. For instance, the best developed and studied urban heat islands in Hungary are found in Budapest and Szeged (Unger and Pongrácz 2008).

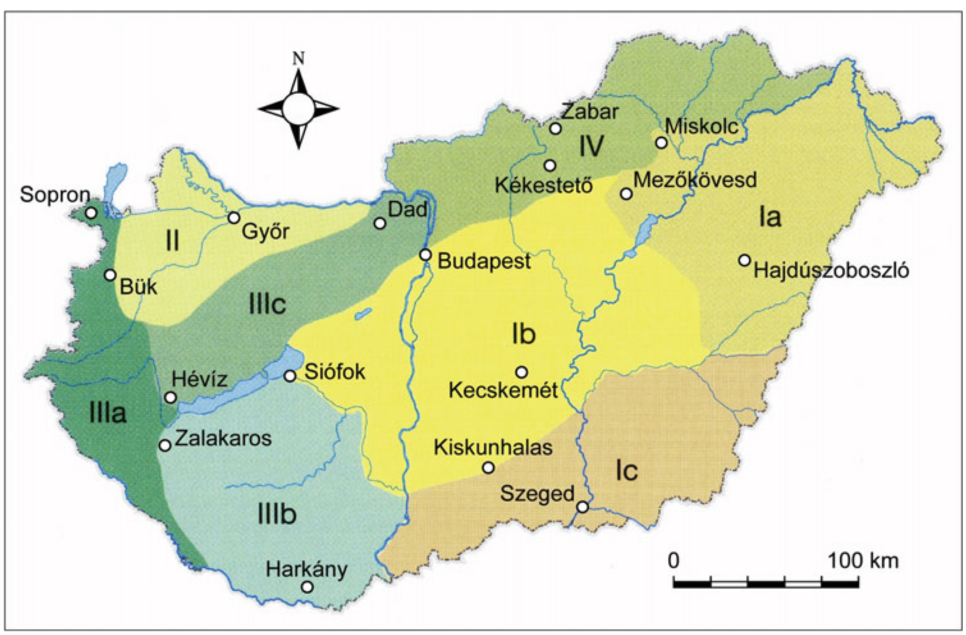


**Fig. 3.1** Climatic districts of Hungary (after J. Bartholy) with the locations of settlements mentioned in the text. *Ia* Northeastern Great Plain (severe winters); *Ib* Central Great Plain (dry summers); *Ic* Southern Great Plain (hot summers); *II* Little Plain; *IIIa* Southwestern Transdanubia (humid); *IIIb* Southern Transdanubia (mild and humid); *IIIc* Transdanubian Mountains (cool summers); *IV* North-Hungarian Mountains

## 3.4 Climatic Elements

### 3.4.1 Solar Radiation

Abundant incoming solar radiation is an outstanding property of Hungarian climate. The annual duration of sunshine is more than 2,000 h (more than 4,500 MJ $m^2$ global radiation) over the entire Great Plains (maximum: 2,496 sunshine hours at Kecskemét recorded in 1950) and this allowed for the plantation of orchards and vineyards. Accordingly, cloudiness is also lowest in the centre of the basin. It shows a high range between regions (annual average: 53–66 %) and a considerable annual variation (highest in the western borderland and in December: up to 80 %).

Local influences further improve heat balance. In the Kiskunság sand region, for instance, the reflection of radiation from bare surfaces enriches fruits (mostly apricot and peach) in sugar and vegetables (red seasoning paprika) in aromatic stuffs. The shallow lake water of Balaton absorbs much heat and this creates a characteristic microclimate within some kilometres' distance in the shore zone. In summer cooler water and air temperatures reduce air moisture content and cloud formation. Clear sky is more common here than elsewhere in Transdanubia and more incoming radiation has again a beneficial influence on viniculture.

### 3.4.2 Temperature

Under the mesothermal climate of Hungary typical MAT values range between 10 and 11 °C for most of the territory. Much colder areas, however, are also found, the coldest being some marked frost pockets in valleys opened to the north at the northern border. The absolute minimum temperature at Miskolc-Görömbölytapolca was −35.0 °C (on 16 February 1940) and −31.9 °C at Zabar, where record daily minima have been measured recently—particularly in the transitional seasons. Although quite low MATs were recorded in the northwestern corner (for the 1901–2000 period: 9.4 °C at Sopron), its altitudinal position makes Kékestető (the highest point, 1,014 m) the coldest place of Hungary (5.6 °C for 1971–2000 and 4.2 °C in the coldest year of 1980)—but it is partly counterbalanced by the ca 1.5-fold higher number of sunshine hours in the winter.

The warmest area is the Tisza lowland along the southern border (long-term MAT in Szeged: 11.1 °C) and the absolute maximum temperature (41.9 °C) was also observed in the Great Plain, in the Kiskunság sand region (Kiskunhalas), on 19 July 2007. According to predictions, the frequency of extreme events will increase in the future (Table 3.1).

### 3.4.3 Precipitation

In Hungary the highest AP (1,554 mm) was observed in the most recent record during the wet year of 2010 in the Bükk Mountains. It is more than twice as large as the climate normal of the country-wide average AP (around 650 mm) and 7.5 times more than the annual minimum of 203 mm, measured in 2000 at the Szeged meteorological station. The highest monthly amount (444 mm) was detected in Dobogókő in June 1958. On a finer temporal scale, maximum daily precipitation (260 mm) was observed during a prolonged cloudburst in Dad (Eastern Gerecse Mountains) on 9 June 1953. Similar events, however, are likely to occur anywhere in the mountain and hill regions in the future.

There are usually two peak periods of precipitation: May and June due to Atlantic air masses and September–October, when rains are brought by Mediterranean cyclones. July and August are months of drought, which mostly affects the Great Hungarian Plain, where the return period for disastrous drought is 20–50 years (Pálfai 2002). Pálfai's Drought Index (PAI) is calculated from the ratio of mean daily temperatures from April to August to precipitation total from October to August, modified by correction factors (of precipitation, groundwater table and temperature) (Fig. 3.2).

The duration and depth of snow cover are among the least predictable climatic elements, but it can be claimed that on the average there are more than 80 days with snow cover in the mountains above 700 m elevation and in western

**Table 3.1** Expected extreme temperatures in Hungary in the 21st century based on PRECIS model simulation, A1B scenario (after Department of Meteorology, Eötvös Loránd University in: NAK 2013)

| Extreme temperature indices | Average value (days per year) | Expected value (days per year) | |
|---|---|---|---|
| | 1961–1990 | 2021–2050 | 2071–2100 |
| Number of frost days ($T_{min}$ < 0 °C) | 93 | 58 | 39 |
| Number of summer days ($T_{max}$ > 25 °C) | 67 | 105 | 135 |
| Number of hot days ($T_{max}$ > 30 °C) | 14 | 48 | 79 |
| Number of very hot days ($T_{max}$ > 35 °C) | 0.3 | 12 | 34 |
| Number of heat alarm days ($T_{mean}$ > 25 °C) | 4 | 34 | 63 |

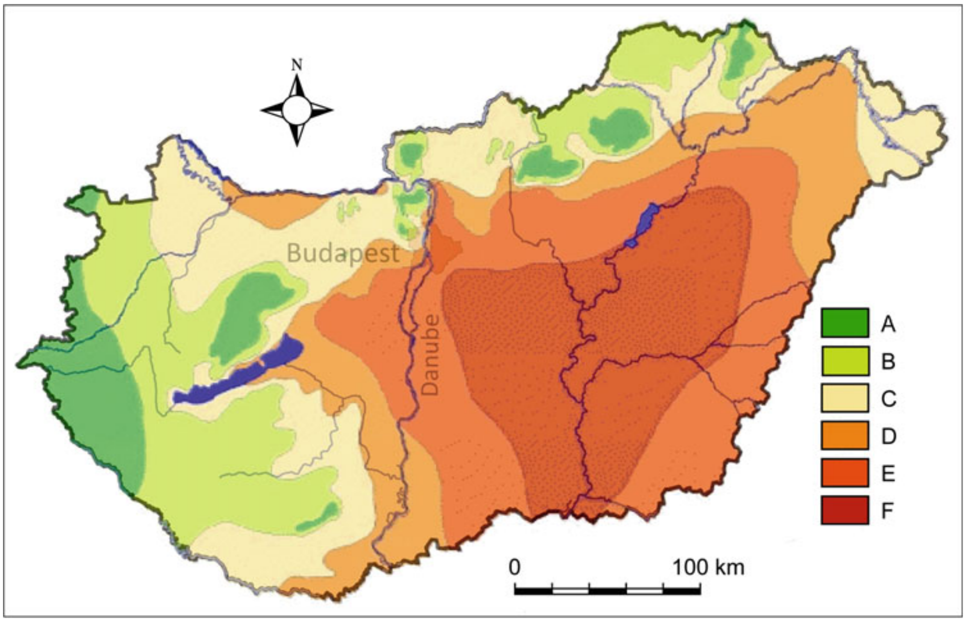


**Fig. 3.2** Drought hazard map based on Pálfai's drought index (*PAI*) (Pálfai 2002). *A* no hazard (PAI <5 °C/100 mm); *B* low hazard (PAI = 5–6 °C/100 mm); *C* moderate hazard (PAI = 6–7 °C/100 mm); *D* medium hazard (PAI = 7–8 °C/100 mm); *E* serious hazard (PAI = 8–9 °C/100 mm); *F* extreme hazard (PAI >9 °C/100 mm)

exposure (maximum average snow depth: 66 cm on Kéketető), while in the lowlands snow hardly lasts for more than some weeks in most years.

### 3.4.4 Winds

Westerly winds are prevailing in the Carpathian Basin. The windiest parts are found in two opposing corners of Hungary: in the northwestern Little Plain (Moson Plain) wind speeds are constantly high (average: 3.8 m $s^{-1}$), while in the southeast, on the Körös–Maros Interfluve maximum wind gusts during rainstorms are common (absolute record: 44.5 m $s^{-1}$ at the Szarvas station, close to the Triple Körös River). Wind channels develop in river valleys in the typically mountainous northern part. In the grabens between mountain groups highly variable distributions of wind directions are observed and locally higher than 40 m $s^{-1}$ wind speeds are also recorded.

The Lake Balaton region is in a special position concerning wind conditions. Lakeshore breeze of reversing direction is common. Similarly, the so-called Vázsony wind, a katabatic wind blowing from the Bakony Mountains, is also typical and can bring sudden thunderstorms. Stormy days (with wind speed above 10 m $s^{-1}$) are common in the Lake Balaton region (more than 75 days per year), only surpassed in Sopron (more than 100 days).

As far as seasonal distribution is considered, the strongest winds blow in spring and early autumn. The most destructive wind event in Hungary was the Biatorbágy tornado on 8 June 1955, when wind speeds reached 100 m $s^{-1}$ and demanded a death toll of 11 people (Bartholy and Weidinger 2002).

## 3.5 River Network Evolution in Brief

The conditions became favourable for the development of the river network after the infilling of Lake Pannon. The Late Miocene deltaic gravels, found in the Transdanubian and North-Hungarian Mountains and previously associated with the Paleo-Danube, are now explained as deposits of Carpathian rivers in the coastal zone of Lake Pannon, cemented by travertine at base level (Schweitzer 2013). In the Late Miocene, in the arid Bérbaltavárian stage (contemporaneous with the Messinian), only intermittent watercourses existed. The formation of the river network dates back to the warm and wet Ruscinian-Csarnótan times and was interrupted under the dry climate of the Villányian (Schweitzer 2013). The previously assumed course of the Danube towards the south, the still existing Slavonian Lake (Fig. 3.3a), cannot be proved by sediment chronology. The highest Danube terraces in the Visegrád Gorge (at 220–240 m elevation) are of Late Pliocene age (Chap. 26). At first, the Paleo-Danube entered the Carpathian Basin through the Ebenfurth-Sopron Gate, then shifted into the Bruck Gate and finally, in the middle Pleistocene, into the Devín (Dévény) Gate between the Little Carpathians and Hainburg Hill. The Danube built the huge alluvial fan of the Danube-Tisza Interfluve and shifted on its surface gradually westwards (Fig. 3.3b). By the Late Pleistocene it formed an anabranching system in the Danubian Plain and reached its present-day north to south course in the Last Glacial (Fig. 3.3c).

In the eastern half of Hungary the Tisza gradually developed into a main hydrographical axis. Although in the Pleistocene the entire Great Plain was a subsidence area, the rate of subsidence was regionally variable. Until the late

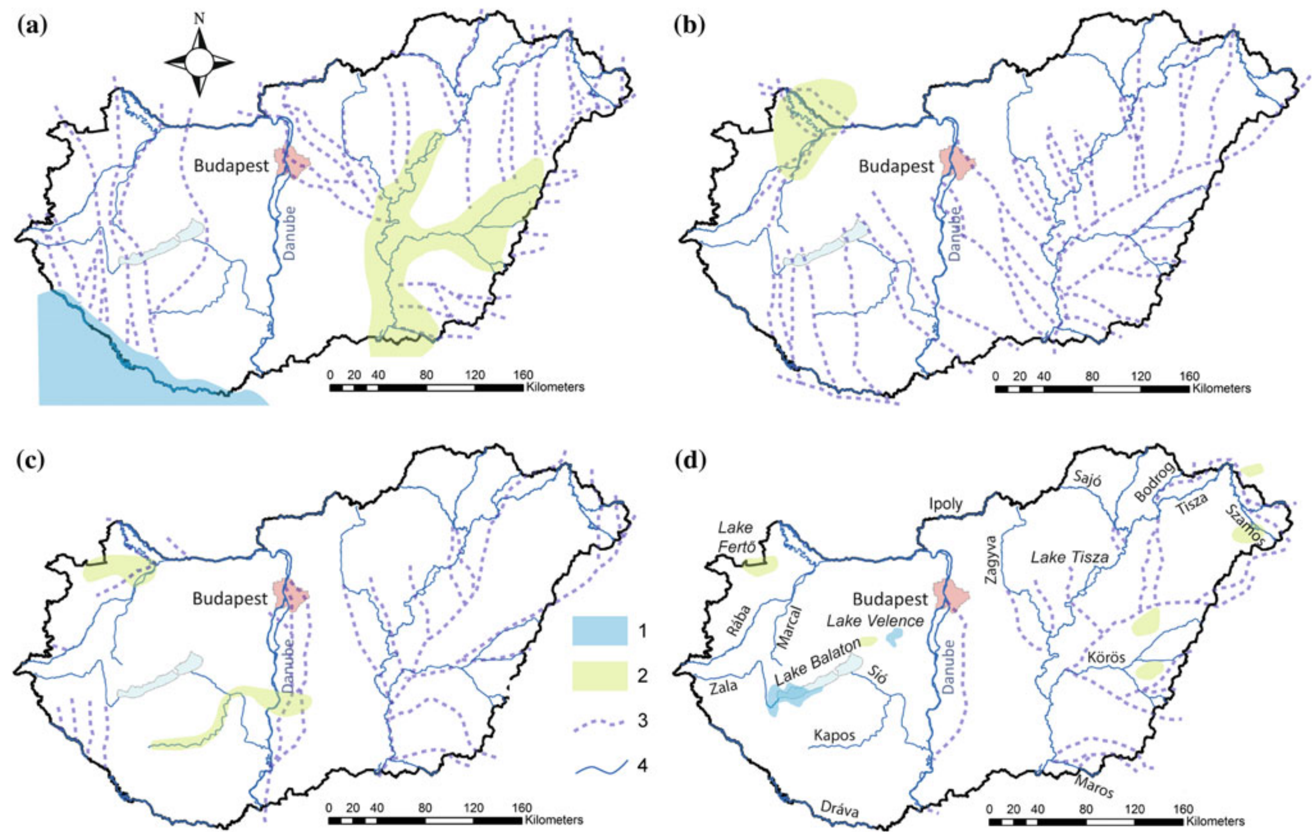


**Fig. 3.3** Evolution of river network in Hungary since the Late Pliocene (modified after Borsy 1990). **a** Late Pliocene; **b** early Pleistocene; **c** late Pleistocene; **d** middle Holocene. *1* Water surface; *2* subsidence areas with uncertain drainage; *3* paleochannels; *4* present-day river network

Pleistocene (Fig. 3.3a, b) the Tisza cut its valley along the present Hungarian-Romanian border, but accelerated subsidence in the north forced it to shift and take a course around the Nyírség alluvial fan, built by its right-bank tributaries (Fig. 3.3c—Borsy 1990). Since the end of the Pleistocene large expanses of the Great Plain became subsidence areas and formed hydrographical centres, like along the lower reaches of the Körös Rivers (Fig. 3.3d). In the Pleistocene interglacials meandering pattern was typical, while in the glacials and the Older and Younger Dryas more sediment load supported a braiding tendency (Mezősi 2011). The now observable meander traces are all of Holocene age. The present courses of rivers are mostly due to human intervention (channelization—Szabó 2011).

## 3.6 Drainage Today

### 3.6.1 Runoff Conditions

Since Hungary is mostly a lowland, the values of specific runoff are rather low (less than 3 L $s^{-1}$ $km^{-2}$). Even higher runoff in hilly (4–5 L $s^{-1}$ $km^{-2}$) and mountain areas (up to 8 L $s^{-1}$ $km^{-2}$) is unable to ensure the discharge of the rivers of the basin—their flow depends on the precipitation conditions in the Alps and the Carpathians (Somlyódi 2011).

### 3.6.2 Rivers and Drainage Patterns

Out of the 2,860 km total length of the Danube, 417 km falls to Hungary. The Danube is the only major European river that follows a mostly west to east course. The entire territory of the country belongs to its drainage basin of over 817,000 $km^2$ area. In the Eastern Alps more than 1,000 $km^2$ of the catchment is glaciated and glacier meltwater is an important source of discharge for the 15 major right-bank tributaries. There are 20 major left-bank tributaries to the Danube, mostly fed by rainfall (WWF 2010). The annual water discharge of the Danube at Budapest is 2,340 $m^3$ $s^{-1}$, and for mean water discharge at mouth the Danube ranks the 22th among the rivers of the world (Czaya 1981).

The channel conditions along the Danube are rather extreme (Lóczy 2007). At Budapest the river is only 3 m deep at some point, while at Liberty Bridge it is more than 10 m. Mean current velocity is 0.5 m $s^{-1}$, while it amounts to a maximum of 2.5 m $s^{-1}$ during floods. The Danube freezes in cold winters. The average duration of drift ice is 30–40 days, usually between mid-December and early February. The first ice on water is recorded on the average on 15 November and the last on 15 April.

The Danube enters the Carpathian Basin in the Devín Gate (Porta Hungarica) and the ancient Danube built a double alluvial fan in the Little Plain (Pécsi 1959), where two meandering and anabranching channels form two large islands: the Szigetköz island (375 $km^2$) and the Žitný ostrov (in Hungarian: Csallóköz, 1,600 $km^2$). The Old Danube has a braided channel. River competence decreases rapidly in the present-day gravel river bed. At the upper tip of the Szigetköz interfluve maximum grain size is 15–20 mm, at the lower one only 5 mm. The environment along this section has been fundamentally altered by the construction of the Gabčíkovo Barrage. The major right-bank tributary, the Rába (Raab, 80 $m^3$ $s^{-1}$) only transports suspended load. In the terraced valley east of Győr the Danube flows in a braided channel with numerous shoals built of the Váh and Hron deposits.

A dramatic change takes place when the Danube arrives at the Visegrád Gorge (see Chap. 26), where the channel is 300 m wide and at least 5 m deep. Seven terraces are identified in the Danube Bend of marked horseshoe shape. Downstream Pleistocene anastomosis is still attested by extensive islands, the largest being the Szentendre (gravels) and and the Csepel Islands (coarse sands). Valley asymmetry is striking: on the right bank the high undercut loess bluffs of the Mezőföld occur in contrast to the flat alluvial left bank.

The Drava (695 km; drainage area: 43,238 $km^2$; mean discharge at Barcs: 595 $m^3$ $s^{-1}$) form the Croatian-Hungarian border along a 135-km long section. Its channel on the elongated alluvial fan is partly meandering, partly anastomosing.

The Tisza is the longest tributary of the Danube (966 km) and drains an area of 157,186 $km^2$. In Hungary it has more tributaries than the Danube, including the Bodrog and the Slaná/Sajó (both fed by water from the Carpathian Mountains in Slovakia and Ukraine) and the Şomeş/Szamos, Crişul/Körös River System and Mureş/Maros from Transylvania. Mean discharge at Szeged is 865 $m^3$ $s^{-1}$. Before river regulations in the second half of the 19th century the Tisza of minimum (0.0002–0.0003) gradient formed huge meanders across the Great Plain. In 2000 two tailings dam bursts caused severe pollution along the river.

Motivated by climatic and topographic endowments floods are common in Hungary, in two periods: early spring and early summer (Somlyódi 2011). Sudden snowmelt along the Upper Danube may cause ice-jam floods in early spring (like in Buda in 1838). Early summer 'green' floods (already green twigs and branches carried along) are associated with the precipitation maximum on the Upper Danube catchment, often coinciding in time with the melting of Alpine glaciers. The lowest water stages are observed in autumn and winter. On the Tisza the floods of the tributaries are often superimposed (as it happened in 1970) and cause 8–10 m water level rise within 2 or 3 days and enduring high water levels (up to 120 days along the lower sections) unprecedented on other European rivers.

### 3.6.3 Lakes

Water surfaces only make up 0.75 % of the area of Hungary. Lake Balaton is the largest freshwater lake in Central Europe (area: 598 $km^2$) (Virág 1997). Its four partial basins took shape in the Pleistocene and filled up during the climatic amelioration following the last glaciation. Its water level and extent have fluctuated remarkably in the Holocene. Its depth is 3.25 m on the average and reaches a maximum of 11.5 m at its narrowest point (1.5 km). Its water level is stabilized by its outflow, the Sió Canal, whose discharge is regulated by the Siófok sluice. Because of its shallowness the lake warms up rapidly in summer (to a maximum of 30 °C). On the other hand, in severe winters the lake freezes (for a maximum period of 110 days, as in the winter of 1962–1963). Recently, lake water quality has been substantially improved through the restoration of the Little Balaton wetland, which retains much of the nutrient load of the only major inflow, the Zala River. Some finest sceneries are found on the northern shore of Lake Balaton (see Chaps. 11 and 12).

Lake Fertő (Neusiedler See, area: 352 $km^2$, three quarters of it belongs to Austria) is even shallower: on the average 1.5 m deep. It dried out on several occasions over its history, for the last time during the 1866–1869 drought period. The lake is filling up rapidly. Its southern, Hungarian portion shows very little open water surface and predominantly constitutes of reed-beds. The concentration of dissolved salts is threefold higher than in Lake Balaton.

The third major natural standing water, Lake Velence (area: 26 $km^2$), is 1.2 m deep (at maximum 2 m). The open water surface is surrounded by 12 $km^2$ of reed-beds. Lake Hévíz is the second largest thermal water lake on Earth (4.75 ha) and the largest with aquatic life.

The largest reservoir in Hungary, Lake Tisza, resulted from the impoundment of the Tisza River in 1974. Its extension depends on the water level of the Tisza (ca 100 $km^2$). Its positive flood retention effect has been confirmed during recent floods.

### 3.6.4 Groundwater

Hungary is rich in subsurface water reserves, particularly in thermal and medicinal waters (Somlyódi 2011). In most lowland and hilly areas, unconfined groundwater table is

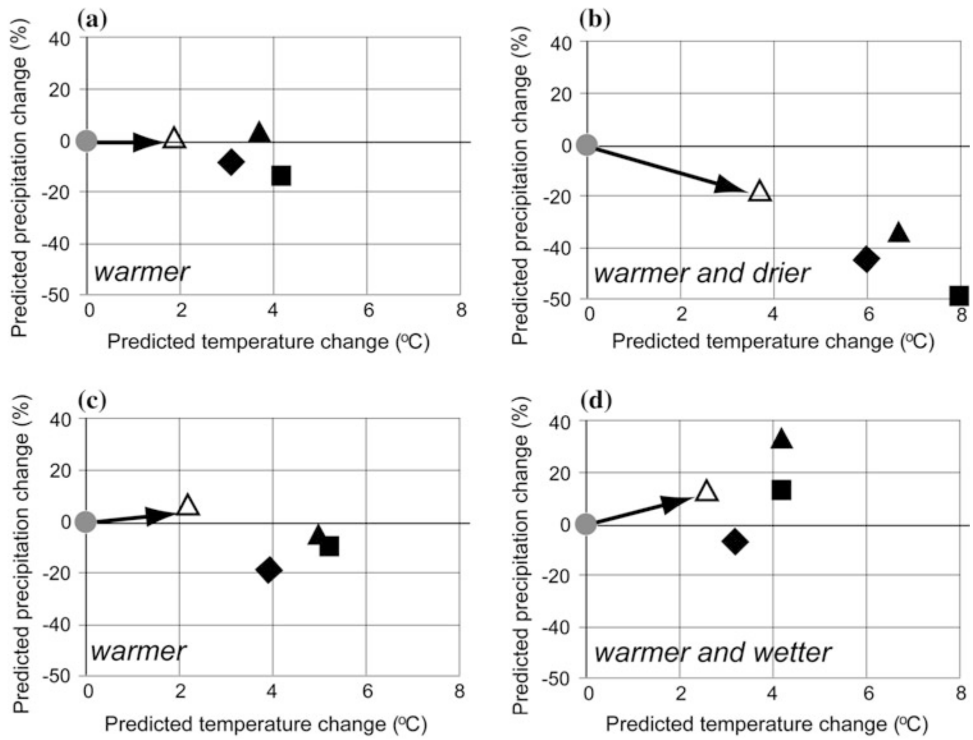


**Fig. 3.4** Predictable climate change for Hungary based on simulations using the PRECIS regional climate model (after Bartholy et al. 2013). Reference period: 1961–1990 (*grey circle* at point 0). **a** spring; **b** summer; **c** autumn; **d** winter. *White symbols* change between 2021 and 2050; *black symbols* change between 2071 and 2100. *Squares* change according to IPCC scenario A2; *triangles* A1B; *rhombus* B2

found at 3–6 m below the surface, but its depth shows a major seasonal range and a dropping trend in some regions of sand deposits (for instance, on the Danube–Tisza Interfluve). The infiltration of nutrient-rich water caused a serious contamination of groundwater reserves. There are more than 50,000 artesian wells in Hungary and this figure points to the importance of the confined groundwater resource in drinking water supply. On thermal water wells a highly developed network of thermal spas is based (Bük, Hévíz, Zalakaros, Harkány, Mezőkövesd, Hajdúszoboszló). Their water also has medicinal effects. In the Transdanubian Mountains, built of limestones and dolomites, a contiguous karst water system is found. To make bauxite mining possible, a huge artificial depression had been created, but with the termination of mining activities the karst water table is steadily rising and some karst springs are reactivating (Babák et al. 2013). One of the geomorphological sites presented in this volume, the travertine hill of Egerszalók (see Chap. 25), directly owes its existence to the precipitation of carbonates from groundwater.

## 3.7 Conclusions

Changeable weather, the increasing emergence of new temperature and precipitation records, is now widely regarded as evidence for global climate change (Mika 2012). A comparison of three different scenarios (A2, A1B and B2) described in the assessment reports of the Intergovernmental Panel on Climate Change (IPCC) points to a remarkable warming trend (1–2.5 °C rise in mean temperatures) for the 21st century (Bartholy et al. 2013; NAK 2013), strongest in summer and winter and less intensive in spring and autumn (Fig. 3.4). Models predict that extremely cold weather will occur less frequently in the 21st century (Pongrácz et al. 2011). A significant drying trend will be typical, particularly in summer (−20 % precipitation compared to the reference period 1961–1990), while in winter precipitation (increasingly in the form of rainfall) is expected to increase considerably (+15 %) by the end of the century (Fig. 3.4). This means that the driest months of the year will probably be July and August (instead of January and February) with 10–30 % less precipitation (NAK 2013), while the wettest months will likely be April and perhaps also May and June. The number of consecutive days without precipitation will decrease in winter (by ca 10–15 %) and grow in summer (by 15–25 %)—particularly in the eastern half of the country (Bartholy and Pongrácz 2010; NAK 2013). Climate change simulations by geographers using various indices indicate that the landscape units of Hungary with the highest drought hazard will be the Körös-Maros Interfluve and the Gödöllő Hills (Blanka et al. 2013) during the last decades of the 21st century.

Such climate changes will influence hydrological and geomorphic processes too. A foreboding sign of new conditions is that flood frequency rose from once in 10–12 years to once in 7–9 years in recent decades. Modified climate may influence floodplain evolution, landslide activities, wind erosion, karst and alkali flat processes and other geomorphological environments (for examples see the relevant chapters of this volume).

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# 4 Long-Term Geomorphological Evolution

Gábor Csillag and Krisztina Sebe

**Abstract**

Hungary occupies the inner parts of the largest basin in the Alpine orogenic belt, the Pannonian (Carpathian) Basin. Elements of its present-day topography have taken shape from the Late Paleozoic to modern times. A uniform basement of the basin was amalgamated from terranes of different origin as late as in the Middle Miocene. Accordingly, older landforms had been created on either the European or the African plate, while since the Miocene geomorphic evolution has happened within a common framework. The oldest relict landforms are fragments of multiple generations of tropical planation surfaces (e.g. tower karsts, shallow tropical karsts, tors), which had formed until the Middle Miocene. From the late Neogene other types of landforms have been preserved as well. The Middle to Late Miocene andesitic-rhyolitic volcanism built mountain ranges consisting mainly of stratovolcanoes. From the Pliocene mostly glacis but also pediments formed in the piedmont zones. The Late Miocene–Early Pleistocene basalt volcanoes include a few shield volcanoes and numerous maars. The spatial distribution of recent geomorphic processes is controlled by a compressional stress regime due to basin inversion acting from the end of the Neogene. Mountains and hills are uplifting and being eroded, dissected by incising watercourses. Uplift and climatic oscillations initiated the formation of flights of terraces along rivers. Mass movements are common on slopes. Aeolian processes had increased importance under the periglacial climate of the Pleistocene glacials: an extensive, thick loess cover was deposited, while wind erosion carved mega-yardangs, wind corridors and deflation hollows. Deflated material accumulated in large dune fields active up to the Holocene. Permafrost features are common in the northern and western parts of the basin. The majority of Hungary's surface is Pleistocene–Holocene alluvial plains.

**Keywords**

Geomorphology • Hungary • Pannonian basin • Planation surfaces • Volcanic landforms • Periglacial features • Terraces • Alluvial plains

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## 4.1 Introduction

The landscape of the Pannonian (Carpathian) Basin formed during geologically long times, and the present-day topography retains the geomorphic signature of hundreds of millions of years (Fig. 4.1). Relict landforms were shaped partly parallel with periods of sediment accumulation but mostly during major denudation intervals. Older geomorphic features originally formed on different microplates, thus in geographically distant regions: the basement of the Pannonian Basin is constituted of tectonic units originating both from the European and the African plates (Chap. 2), which arrived at their present-day positions as late as the Middle Miocene (Fodor et al. 1999). Due to their different tectonic histories, these plate fragments experienced dissimilar pre-Miocene geomorphic evolution, though under comparable climatic conditions at a given time.

Mountains and hills in the Pannonian Basin show erosional landscapes; even volcanic forms are degrading at present. The landforms of these regions have preserved 'memories' of important events in the earth history of the area. This chapter describes only those events in the geological evolution of Hungary which left visible signs, i.e. surface remnants in the present-day topography. For a long time, until the Paleogene, tropical planation was the dominant geomorphic process, therefore the relict landforms of this age are planation surface remnants. Though also known from the earlier geological record, other landforms, such as terraces, volcanic or aeolian features have only been preserved in the landscape from later times. Paleosurfaces make up only fragments of the modern topography, in many cases they crop out in areas of limited extent and their major parts are buried under younger sediments (Pécsi 1998). After their formation they have usually been heavily altered, tectonically deformed, and their superposition or reactivation at different dates is also not uncommon.

## 4.2 Post-variscan Denudation

Tropical or subtropical planation was a major geomorphic process shaping Miocene or older, consolidated rocks, chiefly in the Transdanubian and North Hungarian Ranges and in the Mecsek Mountains. The oldest landform created by this process and presently recognizable on the surface is found in a semi-exhumed position near Kővágóörs

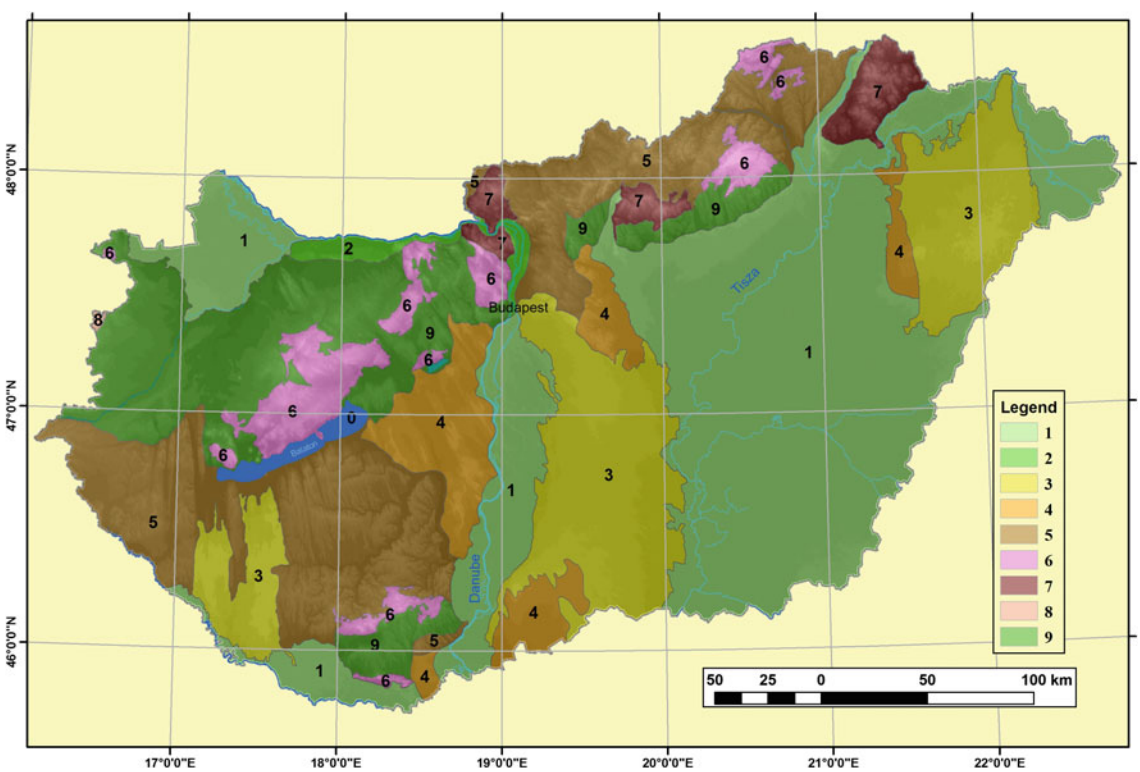


**Fig. 4.1** Geomorphological overview of Hungary. Legend: *1* Late Pleistocene–Holocene surfaces of alluvial plains; *2* Pliocene (?)–Holocene terraces; *3* Late Pleistocene–Holocene blown sand fields; *4* Chiefly Late Pleistocene–Holocene landscape formed on loess; *5* Dissected hills on Cenozoic molasse sediments and Miocene acidic volcanics; *6* Relief on Paleozoic–Eocene rocks with planation surface remnants; *7* Heavily eroded Miocene volcanic ranges; *8* Late Miocene–Holocene erosional landscape of the Penninic Rechnitz Window; *9* Pediments and erosional plains on unconsolidated Cenozoic sediments

(Balaton Uplands or Highland), represented by silcrete and ferricrete fragments, together with debris from the base of the overlying sandstone (Csillag 2004). The planation creating these crusts happened during the post-Variscan denudation interval. It produced a 10–40 m thick variegated, oxidized weathering zone overlying Paleozoic slate and dacite in the Balaton Highland, which includes a thick, red, kaolin-rich weathering mantle on the dacite (Fülöp 1990). The crusts were buried under red fluvial sandstones and conglomerates of Upper Permian age.

## 4.3 Mesozoic–Eocene: A Story of Intense Tropical Planation

From the Triassic to the Early Cretaceous, the dominantly marine sedimentation was interrupted only by minor denudation events leaving no visible signs in the present-day topography. A remarkable change ensued with the closure of the Tethys in the Cretaceous. By that time, both major tectonic units of the Pannonian Basin had detached from the continents as separate plate fragments. Caught between the colliding African and European plates in the Cretaceous, from then on these plate fragments experienced multiple denudation periods under tropical climate. The denudation events can be best distinguished in the Transdanubian Range, though they affected other structural units as well.

In the Transdanubian Range three major, long-lasting denudation periods can be recognised between the Early Cretaceous and Middle Eocene. The resulting regional unconformities, which are tropical planation surfaces in the geomorphological sense, partly overlap in the Southern Bakony Mountains (around the Halimba bauxite field), but their spatial extent is different; this has been explained by the tectonically controlled migration of the elevated area (specifically of a forebulge) (Mindszenty et al. 2001). The superposition of the effects of the three denudation periods created the flat-topped gross morphology of the Transdanubian Range.

The first tropical planation took place during the Early Cretaceous. The gap between the Albian bauxites and the underlying Lower Jurassic and Upper Triassic limestones refers to a terrestrial period with intense karstification and weathering (Kaiser 1997). The next event is indicated by the Upper Cretaceous (Santonian) bauxites, partly in situ, partly redeposited, which overlie an extensive planation surface. This surface was relatively dissected, with paleokarst landforms 90–100 m high. It formed dominantly on Upper Triassic dolomites, rocks older than the basement of the Albian bauxites, thus indicating a prolonged denudation period (Juhász 1988). The major gap in the sedimentation represented by this paleosurface is a consequence of the Alpine orogenic movements. Somewhat later thick sediments of a Late Cretaceous transgression covered the entire Transdanubian Range, but in the early Paleogene another important denudation period followed (until the Middle Eocene), which finally produced the uniform, karstic tropical planation surface of the Transdanubian Range, visible in the Bakony and Vértes Mountains in the vicinity of and underneath the Eocene bauxite occurrences. The Eocene etchplain of the Velence Hills (Kaiser and Csillag 2004) was most probably also a part of this extensive planation surface. A small patch of spectacular steep landforms excavated from under a manganese ore cover at Úrkút (Bakony Mountains) is traditionally known as a cockpit karst belonging to this surface (Bárdossy and Kordos 1989); however, this view has recently been challenged and a submarine origin has been ascribed to these forms (Polgári et al. 2012). In the Albian and at the beginning of the Middle Eocene the humid climate was interrupted by semiarid intervals with pedimentation (Kaiser 1997). Coarse alluvial fans accumulated below or above the bauxites (Kercsmár et al. 2008) or occasionally on the eroded surface of karstic terrains filled with bauxitic clays (Kósa et al. 2003).

In the Bükk Mountains Upper Eocene terrestrial sediments overlie a dissected topography (Less 2005). The relief is partly a consequence of the tropical karstification of carbonate rocks (Hevesi 1978), partly of differential weathering of Triassic carbonates and less resistant Jurassic sediments. According to Hevesi (2002), an open, allogenic karst terrain formed in the Bükk Mountains during the Eocene (Chap. 22). Although it became temporarily buried in the Miocene, karstification has affected almost the same surface (the present-day Bükk Plateau) from the Eocene up until now, thus reshaping the original tropical planation surface.

Little is known about the Mesozoic landscape evolution of the Aggtelek-Rudabánya Hills. According to Gyuricza and Sásdi (2009), subaerial denudation started in the Late Cretaceous, when a low-relief tropical karst landscape formed; however, this was erased by subsequent continuous erosion lasting until the Early Miocene.

In the Villány Hills, a minor bauxite occurrence of Berriasian–Valanginian age and the underlying karstic features indicate subaerial conditions in the Early Cretaceous. However, no landforms produced by this event have been preserved on the surface.

## 4.4 Late Eocene–Oligocene: Erosion and the Paleogene Basin

Through a multi-step transgression in the Eocene, the so-called Paleogene Basin formed and advanced. By the end of the Eocene it inundated the Transdanubian Range and the low-lying regions of the Bükk Unit, thus terminating the most important planation period in the area. Tectonically, the

Paleogene was characterised by large-scale horizontal movements. The plate fragments of the basement of the future Pannonian Basin migrated eastward, towards their later position in the embayment of the Magura Basin. During the Oligocene, strike-slip or transpressional deformations (Fodor 2008) caused local uplift and partial or complete erosion of the Eocene sediments in parts of the Alcapa Megaunit. Karstification of Mesozoic and Eocene carbonates ensued; the karstic cavities and depressions were filled by Oligocene fauna-bearing sediments (Jámbor et al. 1971) and redeposited bauxites in the Transdanubian Range. In the Bükk Mountains the Mesozoic basement was exposed and eroded in the Oligocene, as shown by coarse gravel in the foreland originating from the mountains (Less 2005; Pelikán 2005).

Most of the Tisza Megaunit experienced prolonged denudation from the Cretaceous through the Paleogene. No paleosurfaces of Cretaceous or Paleogene age have been proven to exist in the Mecsek and Villány area, and the lack of Paleogene sediments (except for a very local subsurface occurrence) shows the dominance of erosion in the region.

## 4.5 Miocene: Birth of the Pannonian Basin, Volcanism and the Miocene Climatic Optimum

The first half of the Miocene was crucial for the evolution of the Pannonian Basin. In the Early Miocene the hitherto separate tectonic megaunits were already located close to each other, and their final amalgamation in the Middle Miocene created the modern tectonic framework for landscape evolution. From then on, geomorphic processes acted on a uniform continental basin, in the same climatic system.

In the Early and Middle Miocene (Burdigalian to Langhian), during the so-called Miocene climatic optimum, climatic condition favoured tropical-subtropical weathering. As already proposed by Bulla (1958), this was the last period of tropical planation in the Pannonian Basin, when most of the older paleosurfaces underwent erosion and new surfaces formed.

Tropical weathering in the western Transdanubian Range is attested by dolines with diameters of 15–30 m and depths of up to 100 m. They are filled with Eocene–Oligocene reworked clay, which underwent kaolinitic weathering under warm and humid climate (Budai et al. 1999). On the Veszprém Plateau, in the Southern Bakony, more extensive dolines occur (several 100 m in diameter), filled with red clay, bauxitic clay and dolomite debris (Fig. 4.2). Along the SE margin of the unit, the Eocene surface of the Velence Hills was reshaped by Miocene tropical deep weathering. The spectacular tors (Chap. 11) exhumed by the late Neogene–Quaternary denudation represent the deepest part of the weathering zone, close to the weathering front.

In the highest zone of the Bükk Mountains denudation and karstification continued in the Early and Middle Miocene. Karstification is indicated by the Middle Miocene freshwater limestones (Sásdi 1999; Pelikán 2005). Due to still existing lithological contrasts, the karstic planation surface had a considerable relief, as indicated by the variable thickness (up to 200 m) of the terrestrial sediments partly covering it. Along the margins (e.g. at Nagyvisnyó), remnants of the late Early Miocene shorelines appear in the present-day topography as wave-cut rock surfaces covered by a veneer of littoral gravel or perforated by boring bivalves. According to Pelikán (2005), even the highest parts of the Bükk may have suffered a temporary marine inundation in the early Middle Miocene but became re-exhumed shortly thereafter, still in the Middle Miocene.

In the Aggtelek Hills terrestrial landscape evolution continued until the Early Miocene (Gyuricza and Sásdi 2009), when the karstified surface was covered by pyroclastics. Their weathering products—red clays with minor amounts of bauxite—accumulated in dolines (Gyuricza and Sásdi 2009). The doline lake Vörös-tó and the rock formation Medve-sziklák (“Bear Rocks”) are remnants of this Miocene surface.

In the Tisza Megaunit, the dominance of erosion lasting since the Cretaceous ceased in the Early Miocene. The final paleosurface of the denudation period, a karstic limestone plateau dotted with dolines and ponors in the western Mecsek is now a marked trait of the landscape. It has been preserved under a Lower Miocene fluvial gravel mantle, but was exhumed again after the Miocene, with only scattered remnants of the gravel left mainly in dolines.

Initiated by the subduction of the European Plate under the basement of the Pannonian Basin, intense andesitic-dacitic-rhyolitic volcanism built up the Inner Carpathian Volcanic Chain in the Middle to Late Miocene, creating the most prominent volcanic landscapes. Under the favourable, warm and humid climate the volcanic edifices underwent heavy erosion right from their emergence. Hundreds of metres of material have been eroded, and by now only the strongly remoulded remnants of the primary volcanic landforms have been preserved. One of the largest of them is the horseshoe-shaped caldera of the Visegrád Mountains, which controlled the shape of the Danube Bend (Chap. 16—Karátson et al. 2006; Karátson et al. 2007). Similarly large calderas as well as lava domes and stratovolcanoes rise in the Börzsöny (Karátson et al. 2000) and in the Tokaj (Zemplén) Mountains (Zelenka et al. 2012). According to Karátson (2007) and in contrast with previous views, no calderas formed in the Mátra, the largest volcanic range in Hungary. Instead, the range hosts several crater remnants, which

**Fig. 4.2** Miocene karstic planation surface on Upper Triassic dolomite (Veszprém Plateau) (photo by Gábor Csillag)

dominantly produced lavas and by now have widened and have experienced hundreds of metres of erosion. The largest of them are the craters of Galyatető, Kékes and Nagy-Szár-hegy (Karátson 2007). Subduction-related volcanism also produced multiple horizons of extensive and locally quite thick rhyolitic and dacitic pyroclastics. Where these rocks outcrop, small-scale badlands (Chap. 18) or "beehive rocks" could develop on them (Chap. 24).

The geological and geomorphological evolution of the westernmost areas of Hungary was governed by intense tectonic processes related to extension in the Eastern Alps, including the Hungarian parts of the Austroalpine nappes. As far as landscape evolution is concerned, the Kőszeg Mountains (Chap. 6) are unique in Hungary: in contrast to other elevated areas built up by basement rocks, where planation processes are generally known from the Mesozoic, Kőszeg Mts. were deeply buried under the Lower Austro-alpine Nappe until the Miocene. As shown by fission track data, their tectonic unroofing happened as late as the Early to Middle Miocene (Dunkl et al. 1998). They emerged as a tectonic window (Rechnitz Window) from below the overlying, downsliding nappe in the Serravallian, but became covered again by clastic basin-filling sediments, which did not start to erode until the inversion of the Pannonian Basin, here latest Miocene to Pliocene. This was the time when the landscape of the Kőszeg Mountains could begin to develop.

The Sopron Mountains form an emergent part of the Lower Austroalpine Nappe. Their geomorphic appearance as an elevated dryland is first shown by littoral and fluvial sediments in the Early Miocene, though some authors mention pre-Miocene planation surfaces as well (Kárpáti and Ádám 1975). Apatite fission track ages ranging between 40 and 60 Ma (Dunkl and Frisch 2002) allow that the Sopron Mountains might have been exposed already in the Paleogene.

## 4.6 Late Miocene–Early Pleistocene: Lake Pannon, Basin Inversion, Volcanism and Pedimentation

The Late Miocene of the Pannonian Basin is the story of the formation and upfilling of Lake Pannon. At its largest extent this long-lived lake occupied most of the basin. It was filled mainly from the NW and from the NE by rivers carrying sediment from the uplifting and eroding mountain frame. Through a gradual southward progradation of the northern lakeshore and the final upfilling of the lake by the Early Pliocene, a basinwide alluvial plain formed, with only minor, insular elevated areas. Relict landforms from the lacustrine period are syn-sedimentary fault scarps with littoral gravel and slope debris, chiefly along the margins of the Keszthely and Vértes Mountains (Csillag et al. 2004; Budai and Fodor 2008), and locally also wave-cut platforms have been preserved (e.g. next to Sümeg in the Bakony—J-Edelényi 1985). Further geomorphic significance of Lake Pannon lies in the fact that the lake sediments covered the old planation surfaces, sheltering them from erosion. These surfaces (e.g. in the Balaton Highland) either remained buried or have been exhumed by the later basin inversion in a practically unaltered state.

The Late Miocene–Early Pleistocene alkali basalt volcanism (Martin and Németh 2004; Balogh et al. 1986; Wijbrans et al. 2007; Prakfalvi et al. 2007) produced perhaps the most spectacular landscapes of the Pannonian Basin. It occurred in two regions: in Transdanubia (Little Hungarian Plain, Bakony, Balaton Highland) and in North Hungary (Nógrád-Gemer basalt region). Volcanic activity was dominated by maars, which were often filled by lava flows and lava lakes, and frequently capped by scoria cones (e.g. Fekete-hegy and Boncsos-tető in the Balaton Highland). The famous buttes of the Tapolca Basin (Chap. 8) are remnants of these lava lakes and were created by geomorphic inversion, by erosion of the surrounding loose sediments. A few of the maars have been eroded to their roots, their diatremes are now exposed as small hills (e.g. Vár-hegy of Zánka). Shield volcanoes are less numerous [Kab-hegy (Chap. 9), Agár-tető, Medves Plateau (Chap. 17)] and can exceed 10 km$^2$.

After the Miocene basin formation, a compressional stress field developed in the Pannonian Basin by the Pliocene due to the continued conversion of Africa and Europe (Fodor et al. 1999). This resulted in the inversion of the basin, which was manifested in differential vertical motions and dissection of the hitherto relatively level topography. Uplift and erosion started in the present-day mountain ranges and hills, while in subbasins—in the northern part of the Little Hungarian Plain, in the Drava Basin and in the Great Plain—subsidence and fluvial deposition continued.

Kretzoi and Pécsi (1979) and Schweitzer and Szöőr (1997) consider the Pliocene as the main period of pedimentation (including glacis formation) in the uplifting mountain forelands, followed by the accumulation of thick red clays. Remnants of these surfaces are located in the southern foreland of the Mátra and Bükk Mountains (Pécsi 1985); their former presence in southern Transdanubia (near Beremend, Villány Hills) is indicated by karstic cavity fills of Pliocene age in the underlying limestone, which originate from material transported on a glacis surface (Koloszár 2004). However, in the Transdanubian Range, the recognisable pediments lie in topographically lower positions—and are consequently younger—than the Pliocene–Early Pleistocene paleosurfaces, the age of which is known due to superposed lava flows and maars. In the Kál Basin (Chap. 10) the oldest post-Miocene erosional surface has been dated to be 1.5 Ma old (Ruszkiczay-Rüdiger et al. 2011). In the foreland of the Vértes Hills eight glacis surfaces have been distinguished (Csillag and Fodor 2008); even the fifth of them is barely 100 thousand years old (Thamó-Bozsó et al. 2010). A definitely old erosional paleosurface, but more extensive than the mentioned mountain foreland glacis, is the denudation plain of the southern Little Hungarian Plain. As shown by dated maar volcanoes in the area (e.g. Ság, Somló), it has been evolving and downwearing since the latest Miocene, both through fluvial and aeolian processes.

## 4.7 Middle Pleistocene–Holocene: Periglacial and Modern Landscape Evolution

The major part of Hungary's land surface has been modelled by geomorphic processes acting in the younger Quaternary. The alternating periglacial climate of the glacial intervals and temperate continental climate during the interglacials brought varied processes shaping the landscape, many of them still active.

The most important sediment and also a landscape-forming factor of the Pleistocene glacials is loess. Distributed over mountains, hills and plains, its thickness is highest in southeastern Transdanubia (up to nearly 100 m—Koloszár and Marsi 2010). Loess deposition started in the Early Pleistocene—the oldest loess strata are more than 1 Ma old (Koloszár and Lantos 2001)—and lasted until the very end of the Pleistocene (Frechen et al. 1997). The loess-paleosol sequence contains pedogenic red clays in its lower and forest and steppe soils in the upper units, as well as two young tephra horizons (Horváth 2001). Major loess regions occur in Transdanubia (Baranya and Outer Somogy Hills, Mezőföld) and on the Great Plain (Hajdúság). Typical loess-related landforms include rapidly evolving narrow, deep gorges (20–25 m deep in the Tolna Hills), both naturally formed and anthropogenic (hollow ways or sunken lanes), and various loess karst features: dolines, sinks, piping features. Along the walls and at junctions of gorges several metres high loess columns form. Loess plateaus are frequently characterised by shallow dry valleys (dells) and deflation hollows. Loess areas are highly prone to landsliding (Chap. 14).

Besides depositing loess, wind also played a major geomorphic role as an erosive agent, especially in the driest and windiest periods of the Pleistocene. Strong air flows entering the Pannonian Basin from the NW, between the Alps and the Carpathians, carved a fan-shaped system of long, linear valleys (wind corridors) and mega-yardangs (ridges in between) in the weakly consolidated sediments in the uplifting western part of the basin (Sebe et al. 2011; Fig. 4.3). On the leeward side of the Transdanubian Range, a series of shallow deflation hollows formed. These include the initial subbasins which coalesced in the Holocene to form Lake Balaton, the largest lake of Hungary (Cserny and Nagy-Bodor 2000), and the basin of Velence Lake. The marginal areas of the Transdanubian Range have been uncovered from below 200 to 300 m thick Lake Pannon sediments, with the wind taking an important part in the process.

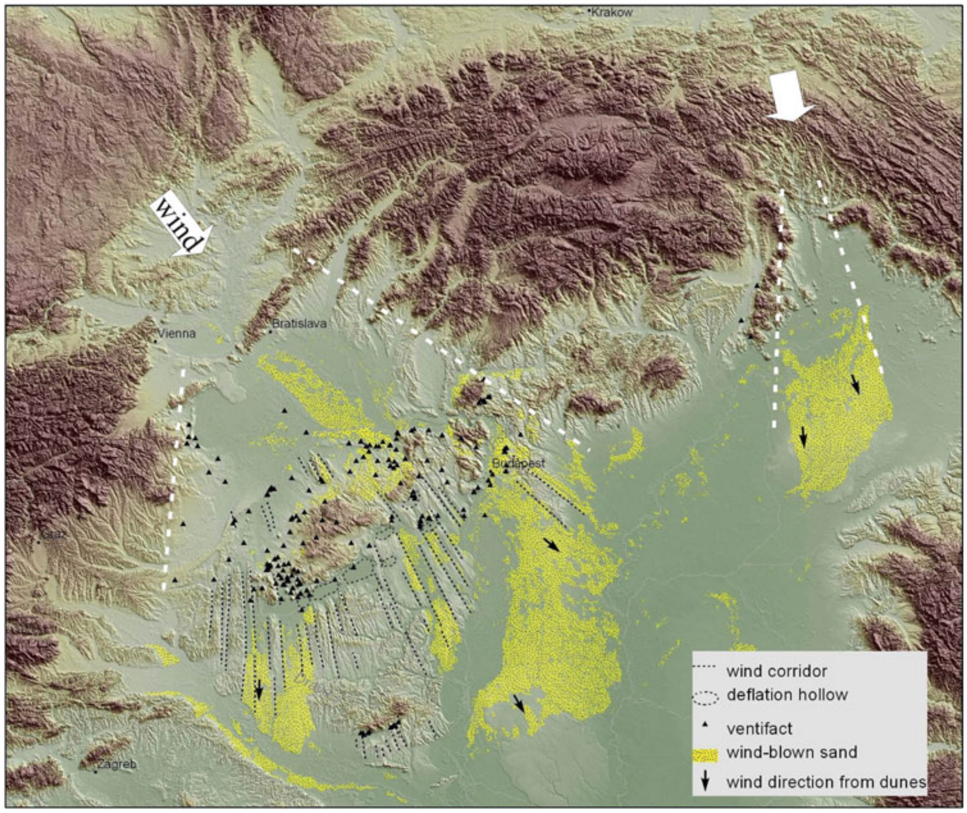


**Fig. 4.3** Periglacial aeolian feature system of the Pannonian Basin (after Sebe et al. 2011)

Loose material removed from the uplifting source areas was deposited downwind as aeolian sand, and alluvial plains have also been surficially reworked by the wind. The major blown sand regions (Belső-Somogy, Kiskunság, Nyírség) display various meso-scale landforms depending mainly on vegetation cover, wind velocity and sand supply during their formation. The dominant geomorphic features are typical of partially vegetated sands and include parabolic dunes of sometimes considerable sizes (Chap. 30), blowouts and blowout dunes, but sand sheets (coversands) are also widespread (Borsy 1977). Sand movement recurred several times during the Pleistocene and also in the Holocene (Ujházy et al. 2003; Kiss et al. 2012).

During the Pleistocene glaciations, a variety of periglacial features formed. Permafrost-related phenomena like ice and sand wedges and cryoturbations are common primarily in unconsolidated sediments in the north and northwest (Pécsi 1997; Fábián et al. 2014). Volcanic ranges are rich in periglacial forms like block fields, block streams (Fig. 4.4) and cryoplanation terraces (Pinczés 1977; Székely 1977).

The recent landscape evolution of uplifting areas is dominated by surface processes typical of the relatively wet, temperate continental climate prevailing today and also in the interglacial and interstadial periods. Uplifting mountain ranges and hills are being dissected by incising drainage networks. Terraces are characteristic along watercourses. On poorly consolidated molasse sediments of several hilly areas, e.g. in north Hungary or in southern Transdanubia, valleys can be surprisingly closely-spaced and deep despite of the low average elevation of the terrain. On clayey and silty rocks landslides are very common (Szabó 1996); less frequently they also occur on hard rocks and must have been widespread in the humid intervals throughout the Quaternary. Slope movements have occured in historical times as well. For instance, Lake Arló near Ózd in northern Hungary was dammed by a human-induced landslide triggered by undermining.

The exhumation of subvolcanic bodies could result in relatively high relief even in areas of low average elevation. In the western part of the North Hungarian Range, in the Cserhát Hills, andesite dykes intruded into soft sediments in the Middle Miocene. Selective erosion of these few tens of metres wide but 10–15 km long dykes created long and narrow ridges.

Limestone surfaces have been experiencing intense karstification, rejuvenating or remodelling paleokarsts. Karren fields, ponors and dolines are being formed, simultaneously with hot- and cold-water cave systems (Chap. 22). Caves have formed at numerous elevations in uplifting mountain ranges, with the oldest ones lying in the highest positions (Szanyi et al. 2012). The best-developed large caves as well as the most pronounced karren phenomena are located in the

**Fig. 4.4** Periglacial block stream on a north slope in the Mátra Mountains, near Mátraháza (photo by Krisztina Sebe)

Aggtelek karst area (Chap. 20). In repeated periods since the Late Miocene karst waters issuing from carbonate rocks have deposited freshwater limestones (travertines) along the mountain fronts, mainly around the northeastern end of the Transdanubian Range (Gerecse and Buda Mountains). Here many travertine bodies capped and preserved underlying sediments of the Danube terraces and now form flat-topped hills at various elevations depending on their ages (Scheuer and Schweitzer 1988; Sierralta et al. 2010).

Structural movements have also contributed to landscape evolution by producing tectonic landforms among others. Probably the most spectacular of them is the fault scarp of the Telegdi-Roth Line in the southern Bakony Mountains (Fig. 4.5), along which nearly 5 km of right-lateral displacement happened in the Neogene (Tari 1991). Active tectonics strongly controls surface processes, the distribution of eroding and upfilling areas as well as the drainage network.

Under the recent compressional stress field of the Pannonian Basin, the main accumulation areas are subbasins nearly continuously subsiding since the Miocene: the northern part of the Little Hungarian Plain, the Great Plain and the Drava Basin.

On entering the Little Hungarian Plain, the Danube, its tributaries from the NW Carpathians and the river Rába built a several hundred metres thick alluvial fan in the northern part of the basin in the Quaternary. Here the Danube shows an anabranching drainage pattern, with migrating bars and islands. On its way to the Great Plain, the Danube has cut an antecedent gorge between the uplifting Transdanubian and North Hungarian Ranges (Chap. 16). The formation of a flight of eight, Pliocene to Holocene terraces in the gorge has been governed both by uplift and climatic factors (Pécsi 1959; Ruszkiczay-Rüdiger et al. 2005; Gábris and Nádor 2007).

The modern landscape of the Great Plain has been created by three major rivers since the Late Pleistocene: the Danube, the (paleo) Tisza and the (paleo) Bodrog (Gábris and Nádor 2007). Alluvial fan systems have been built on the plain by the Danube from the northwestern and by the Tisza and its tributaries (Sajó-Hernád, Bodrog, Körös rivers, Maros) from the northeastern and eastern directions. The thickness of the Quaternary succession exceeds 500 m. By the end of the Pleistocene, the Danube had gradually abandoned its alluvial fan in the present-day Kiskunság region, thus losing its role in shaping the plains. The river Tisza and its tributaries have created a huge, perfectly flat floodplain in the Late Pleistocene and Holocene; the Hortobágy region (Chap. 27) is part

**Fig. 4.5** Long, straight, steep rock slope, the fault scarp of the Telegdi-Roth line in the South Bakony Mountains (near Várpalota) (photo by Zsuzsanna Haraszti)

of this area. Even in the Great Plain, the impact of neotectonics on fluvial action can be detected along with that of the climate (Gábris and Nádor 2007). The characteristic landforms of the modern topography are meanders, oxbow lakes (Chap. 31) and point bar systems.

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# 5 Geomorphological Regions

Dénes Lóczy

**Abstract**

Since the present territory of Hungary is not a natural physico-geographical unit, its subdivision into geomorphological regions (or districts) is a challenge for experts. Adjusted to the major structural units of the Pannonian (Carpathian) Basin six geomorphological macroregions could be identified—also in the context of the entire basin. In addition to structure (lithology and tectonics), the basis for distinguishing between geomorphological units was the overall character of their topography (elevation, relative relief) and prevailing geomorphic processes. The recognition that most of them constitute 'fuzzy sets' with transitional strips rather than rigid boundaries between them is another obstacle to the map representation of geomorphological regions. While describing geomorphological units, brief descriptions of the topography are supplied (for the underlying geology see Chap. 2) and the most important geomorphological sites in the given region are mentioned.

**Keywords**

Geological structure • Topography • Geomorphic processes • Subdivisions • Hungary

## 5.1 Introduction

In Hungarian geography, the boundaries of landscape units, when drawn manually, are adjusted to the spatial distribution of the components of the physical environment, i.e. the so-called "landscape-forming factors": tectonic units, surface rocks, topographic elements, surface drainage pattern, groundwater reserves, climate, soil and vegetation types (Pécsi and Somogyi 1969). The broader context for relief classification was established by the preparation of the geomorphological map of the Carpatho-Balkan region in international collaboration (Pécsi 1977). Since the 1950s, for political reasons, however, the physico-geographical units of the Carpathian Basin (Bulla and Mendöl 1947) could not be represented on maps any more, and this resulted in the awkward situation that regions terminated at the national border.

It was usual in the terminology of Hungarian geography that when the distribution of a single factor serves as a basis of subdivision, districts are identified, and when the physical environment is considered in its complexity, regions (or landscape units) are defined. In this concept regions can be mapped through the superposition of maps of geological, topographical (geomorphological), hydrographical, climatic, soil and vegetation districts.

The traditional geomorphological districts delimited for the territory of Hungary (Bulla 1964; Pécsi and Somogyi 1969; Pécsi 1970—Fig. 5.1) at the highest level of hierarchy are

- the Great Hungarian or Middle Danubian Plain (in Hungarian: Alföld or Nagyalföld);
- the Little Hungarian or Little Danubian Plain (Kisalföld);
- the Alpine foreland (Alpokalja) or Western Transdanubia;
- the Transdanubian Hills;
- the Transdanubian Mountains;
- the North-Hungarian Mountains with its intramontane basins.

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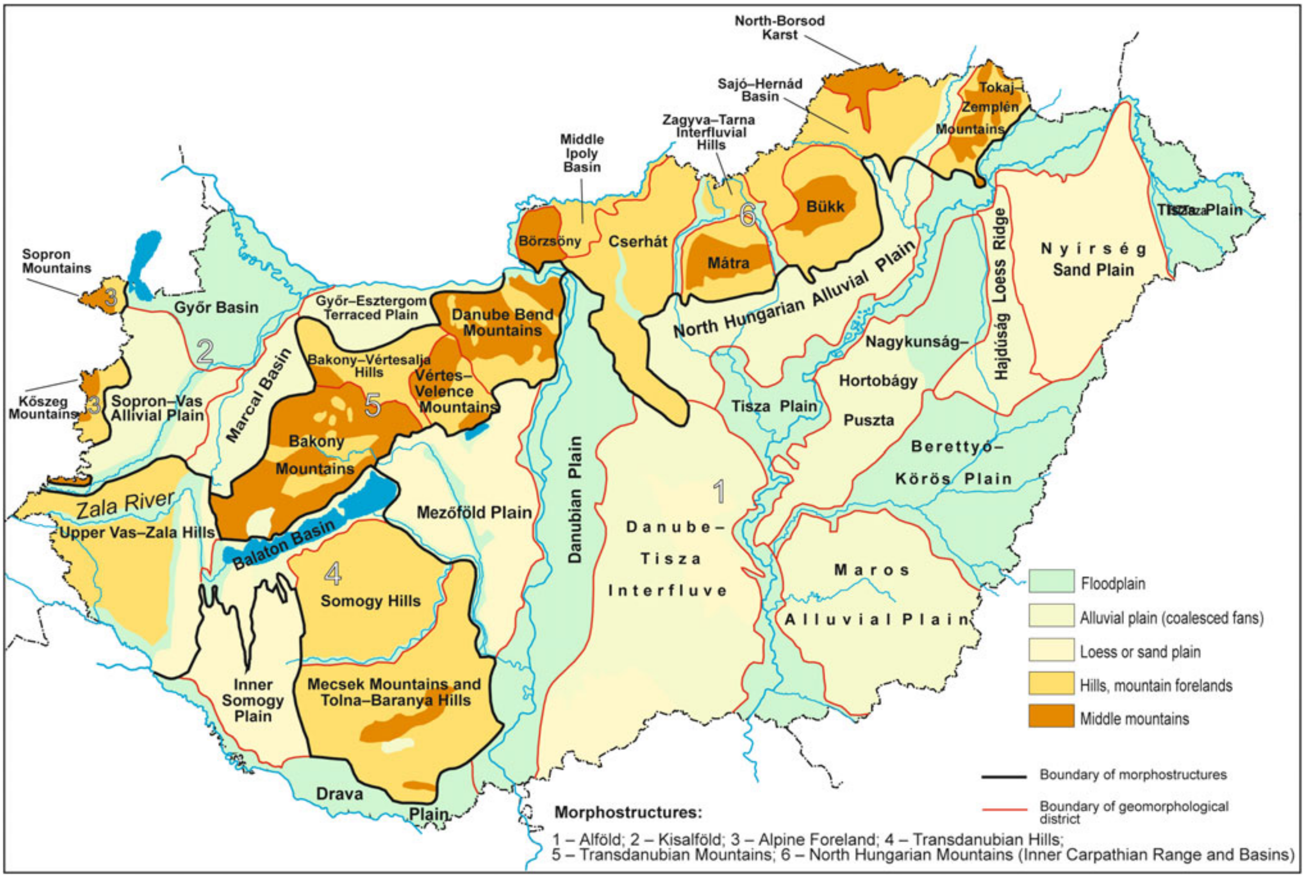


**Fig. 5.1** Major geomorphological districts of Hungary (after Pécsi and Somogyi 1969; Kocsis and Schweitzer 2009)

With the exception of the Transdanubian Mountains, all the geomorphological units continue beyond the country's borders and their boundaries closely follow those of the highest-level landscape units (Hajdú-Moharos and Hevesi 2002). Hajdú-Moharos and Hevesi (2002) even find the North-Hungarian Mountains a misnomer since the indication of its northern position had made no sense before the 1920 Peace Treaty, when the Carpathians formed the national border in the north. However, no alternative denomination acceptable for use in academic language have been offered for this mountain range.

The boundaries of macroregions have been modified with time. Acknowledging that both banks of rivers are usually of the same physical character, boundaries were moved away from rivers in several cases. The lowlands west of the Danube (i.e. in Transdanubia) are now grouped with the Great Plain if they are adjacent to it (Mezősi 2011).

## 5.2 The Great Hungarian Plain

Lying in the central part of the Carpathian Basin, the topography of Hungary is dominated by lowlands. The Great Hungarian Plain itself occupies almost two-thirds of the country's territory (62,300 km$^2$) and, thus, is considered *primus inter pares* among the landscapes (Bulla and Mendöl 1947). It extends in all directions over the territory of neighbouring countries (under various names including the Slovak Lowland, the East Croatian Plain, the Vojvodina Plain, the West-Romanian Plain, the Transcarpathian Lowland of the Ukraine and the Eastern Slovakian Plain) with the total area of more than 100,000 km$^2$ and average elevation around 150 m (lowest: 78 m; highest: 183 m). The Great Plain results from long-term basin subsidence and the accumulation of an up to 7 km thick marine and lacustrine sedimentary sequence (Chaps. 2 and 4) with 400–700 m of Pleistocene fluvial deposits. Its boundary roughly follows the contour of 200 m above sea level and does not everywhere correspond to tectonic lines (often blurred by marginal volcanic hills or alluvial fans). The boundary mostly coincides with the outer limit of the largest extension of the Upper Miocene Lake Pannon.

Along its margins there is a usually gradual transition to foothill areas. The present-day surface is a product of Pleistocene and Holocene fluvial and fluvio-aeolian accumulation that occurred under varying regimes of tectonic subsidence (the highest in the southernmost zone, the Körös and the Zagyva river hydrographical centres), and is subdivided accordingly to three types: low and high levels of alluvial plains (most characteristically in the 'valley' of the Tisza River), loess-mantled alluvial fans (the largest being that of the Maros/Mureş River) and alluvial fans with blown-sand dunes (first of all in on the Danube-Tisza Interfluve and in the Nyírség). The floodplains and loess-mantled fans protected from flooding are presently shaped by human action (mainly arable farming), the blown-sand areas by seasonal wind action, and the active floodplains by fluvial processes during floods.

In spite of the relative monotony of lowland topography, several geomorphological sites have been selected for presentation in this volume. Each represents a typical landscape unit: the cut-off oxbows of the Tisza the pristine channel environments once accompanying major rivers (see Chap. 31); the alkali puszta of Hortobágy the now drained river floodplains (Chap. 27), while the dunes of the Nyírség the blown sand surfaces of alluvial fans (see Chap. 30). Two additional sites are included to illustrate the significance of human impact on the landscape: a crater of natural gas explosion (see Chap. 28) and the tumuli, which are landmarks visible from a great distance, built by one or several prehistoric people(s) (see Chap. 27).

The engineering geomorphological issues of the loess bluffs along the Danube are presented on the example of the Dunaszekcső bluff (Chap. 14).

## 5.3 The Little Hungarian Plain

In the northwestern corner of Hungary (Transdanubia), in the foreland of the Eastern Alps and Northwestern Carpathians, the Little Plain (5,300 km$^2$) encompasses Danubian landscapes in the north: the two-tier Pleistocene alluvial fan and the right-bank terrace system; the lacustrine-paludal Fertő(Neusiedler See)-Hanság depression in the west and surfaces shaped by the tributaries of the Danube in the south. The Little Plain also belonged to the area of the Miocene Pannonian Sea and its subsidence and sedimentation history also resembles the evolution of the Great Hungarian Plain. The thickness of the gravelly-sandy-silty accumulations amounts to several thousands of metres in the hydrographical centre (the Győr Basin), while the southern portion of the plain is dominated by eroded surfaces (the Marcal Basin) and smaller and younger alluvial fans (Rábaköz). In the east multiple terrace levels fringe the Danube floodplain. Today the impacts of land drainage, ensuing cultivation, gravel quarrying and river deposition determine the character of the landscape.

The Little Plain displays a variety of landforms of interest, such as meandering channels, river terrace sequences, terrace "islands". In Chap. 8 some mention is made of the isolated, 5.5–3 Ma old volcanic hills on the margin, Ság and Somló Hills (see also Budai and Gyalog 2010).

## 5.4 The Alpine Foreland or Western Transdanubia

Only a fraction of the Alpine foreland (750 km$^2$), the easternmost spurs of the ranges of the Northern Limestone Alps, belongs to Hungary. They are part of the Lower Austroalpine Nappe (the Sopron Mountains), the Upper Austroalpine Nappe (under the sediment fill of the Rába River Valley or Sopron–Vas Plain) and the Pennine Nappe (the Kőszeg Mountains and Vas Hill), which are bordered in the southeast by a major lineament, the Rába line. Consequently, mostly metamorphic and sedimentary rocks of very different properties, including the resistance to erosion, are found here. On the rock surfaces weathering and slow hillslope processes are predominant, while streams incise into the gravel fan surfaces which surround the mountains.

The landforms originating from differential weathering are clearly demonstrated on the Hat Rock in the Kőszeg Mountains (see Chap. 6). Additional sites in the Sopron area which deserve a visit are the Brennberg Valley cut into Paleozoic mica schist (exposed in the Vöröshíd quarry) and the flat-topped hill row of the Balf Block (where Badenian, 15.5–12 Ma old, Leitha Limestone is worked in the Bishop's Quarry with a wealth of fossils) (see Budai and Gyalog 2010).

## 5.5 The Transdanubian Hills

The Transdanubian Hills (10,200 km$^2$) is an equally heterogeneous geomorphological unit of Hungary. This heterogeneity does not only refer to the Paleozoic-Mesozoic basement, but also to the rock formations exposed in the isolated mountains surrounded by hills. In addition to locally minutely dissected, asymmetric loess-mantled or sand hills of 200–300 m elevation with broad river valleys (for instance, the Kapos Valley, see Chap. 14), the region includes lowlands (for instance, around Lake Balaton) and folded-faulted mountains, such as the Mecsek Mountains, a denuded and exhumed anticline of Permo-Triassic clastic sediments and Triassic carbonates in the west and a syncline of Jurassic sedimentary and Miocene volcanic rocks in the east; and the Villány Hills of imbricated strata of Mesozoic carbonates. The Somogy Hills are interpreted as a foothill region of the Transdanubian Range formed before the subsidence of the Lake Balaton basin in the last phase of the Pleistocene and are composed of loess-mantled alluvial fans of streams flowing towards the tectonic depressions of southern Transdanubia. The hills are predominantly under crop cultivation. Sheet wash, intense rill formation, locally piping, gully erosion (as described in Chap. 13 on the example of the Somogybabod gully) and soil creep are the main geomorphic processes.

In a broader sense the Transdanubian Hills also include the asymmetric hill ridge of the Kemeneshát, a remnant of a gravel-dominated alluvial fan, and the Zala Hills, which are mostly sandy accumulations of ancient Alpine rivers (probably even including the Ancient Danube), minutely dissected by Quaternary watercourses of various size. The origin of the N to S aligned "meridional" valleys of the Eastern Zala Hills has been the subject of a long-standing debate in Hungarian geomorphology: there have been proponents of both fluvial and eolian generating processes. Less uncertainty surrounds the double capture of the Zala River (the first at the village of Türje and the second at Zalavár—see map in Fig. 5.1).

The most interesting geomorphological sites are found beyond the northern shore of Lake Balaton, in the Balaton Highlands—the blockfields of the Kál Basin (described in Chap. 10); the Hegyes-tű columnar basalt exposed in a quarry; the Tihany thermal spring cones (see Chap. 9) and the Balatonkenese bluff of Pannonian marine sediments; in the Mecsek Mountains—the Abaliget dripstone cave; the conglomerate "puppet cliffs" (Babás szerkövek) of Jakab Hill; the rimstones in the Melegmány Valley; the opencast Karolina black coal mine under reclamation; and in the Villány Hills—the karren field of the "Hill Ploughed by the Devil" (Ördögszántás on Szársomlyó) (see for details Budai and Gyalog 2010; Dezső et al. 2004).

## 5.6 The Transdanubian Mountains

The Transdanubian Mountains (6,400 km$^2$) are geologically an elongated synclinal structure and geomorphologically a series of slightly folded-imbricated and block-faulted horsts of southwest to northeast general strike (Pécsi 1970), separating two subbasins of the Carpathian Basin. The mountains were planated in the Mesozoic; dismembered by and uplifted along a dense network of faults during the Tertiary into numerous plateaus at 300–700 m elevation. The planated surfaces of the range were repeatedly buried and exhumed. The individual members, the Bakony region, the Vértes, and the Danube Bend Mountains (Gerecse, Pilis and Buda Mountains), are separated by tectonic basins and grabens filled with Miocene sediments (the Mór, Tata and Pilisvörösvár grabens), with present-day seismic activity. Sparse drainage is typical with gorge-like valleys. Spring caves are elevated to various altitudinal positions. Most of the forested, plateau-like Mesozoic and Tertiary limestone mountains typically show covered karst phenomena, while dolomite surfaces show characteristic weathering phenomena and mass movements. The Plio-Pleistocene tectonic movements, reaching the extent of 200–300 m as attested by travertine horizons, were accompanied by basalt volcanism.

The variety of landforms in the Transdanubian range is represented by two sites of basalt volcanism: the Kab Mountain in the Bakony Mountains (see Chap. 7) and the buttes of the Tapolca Basin (see Chap. 8). In addition, the landforms on the granite of the Velence Mountains (Chap. 11) are of interest to geomorphologists—along with a number of other sites.

## 5.7 The North-Hungarian Mountains

Beyond the Pilisvörösvár graben, the continuation of the Transdanubian Mountains of 8,000 km$^2$ area is called the North-Hungarian Mountains. It is a range of highly variable geological structure (member of the Inner Carpathian Range) and elevation (the only mountains reaching above 1,000 m, highest point: Kékes, 1,014 m). The range includes both Mio-Pliocene volcanic mountains (complex andesite paleovolcanoes in the Visegrád, Börzsöny and Mátra Mountains; andesite dyke systems in the Cserhát Mountains; rhyolite and andesite laccoliths, necks, dykes and lava sheets in the Tokaj Mountains) and Mesozoic-Paleogene limestone mountains with a wealth of karstic features (the Bükk Mountains—see Chap. 21). After the Neogene-Pleistocene uplift the exhumed karstic plateaus became more or less conserved surfaces, while the volcanic landforms have been mostly affected by intensive denudation. An almost continuous broad belt of pediments, glacis and alluvial fans have formed along the southern margin of the range. The higher surfaces are forested, while foothills are used as arable land, with southern slopes as gardens and vineyards. The present-day geomorphic processes also vary on a wide range: hillslope processes, mass movements, karstification and all types of erosion by surface runoff are typical. The members of the range are separated by broad fluvial valleys deepened along faultlines, draining to the south, towards the Tisza River.

Of all the geomorphological districts of Hungary, the North-Hungarian Mountains offer the largest number of sites selected for inclusion into this book: the extensive dripstone cave system of Baradla-Domica in the Aggtelek Karst (see Chap. 20); loess features on Tokaj Hill itself (Chap. 25), and the sandstone gorges and hoodoos in the Vajdavár Hills (Chap. 19). The spoil heaps in the Medves mining region (Chap. 17), the rhyolite badlands at Kazár (Chap. 18); the tufa mound of Egerszalók (Chap. 23), and the millstone quarry of Megyer Hill in the Tokaj Mountains (Chap. 26) are directly or indirectly of human origin.

The great diversity of the North-Hungarian mountain range is best exemplified by the landforms of the Uppony Mountains (Chap. 21). The origin of the Danube Bend gorge

(Chap. 16) between the Visegrád and Börzsöny Mountains receives a special emphasis since it is still an unsolved problem in Hungarian geomorphology.

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# Part II

# Landscapes and Landforms

# Hat Rock: Solution Features in Metamorphic Mountains

6

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**Abstract**

In the chapter the denudation of metamorphic rocks with carbonate content is presented from the Kőszeg Mountains and its impact on geomorphic evolution is described. The calcareous phyllites and greenschists contain calcium carbonate in considerable amounts and have different dissolution properties. At the same time, dissolution is also influenced by other factors such as the degree of frost shattering. Their joint action creates solution notches, holes, hat shape forms and terraces. The most spectacular rock formation generated from greenschists in the Kőszeg Mountains is the Kalapos-kő (Hat Rock). The selective dissolution of the rock results in hat-shaped forms, which, on losing support, tend to collapse. Parallel slope retreat can be an enduring phenomenon, eventually generating terrace-like features which are characteristic morphological elements of valley slopes of calcareous phyllites and greenschists.

**Keywords**

Greenschist • Calcareous phyllite • Semicarbonate dissolution • Solution notch • Hat • Terrace • Kőszeg Mountains

## 6.1 Introduction

Siliceous rocks are often considered insoluble. If, however, they have a considerable carbonate content, they can be affected by solution processes, which create morphological features, formerly attributed exclusively to soluble rocks. The dissolution of calcareous rocks and its morphological footprint have been described for various types of rocks. Dissolution processes in marls, carbonate conglomerates (Avias and Dubertret 1975), calcareous sandstone (de Vaumas 1970; Sanlaville 1973; Waltham 1984; Goeppert et al. 2011); rhythmic sandstone and clay sequences (flysch) (Slabe 1992; Kéri 2011) and loess (Boros 1982) were likewise demonstrated in the literature (Jakucs 1971).

The rocks of the Kőszeg Mountains are affected by the process called partial or semicarbonate dissolution (Veress et al. 1998; Veress 2004). In this case only a part of the rock mass, the carbonate content, enters the solution. As a consequence, the rocks are disintegrated into grains of different sizes.

## 6.2 Geology and Geomorphology of the Kőszeg Mountains

The Kőszeg Mountains are the easternmost unit of the Eastern Alps in Hungary, bordered in the north by the Gyöngyös and in the south by the Pinka Rivers, in the east by the Kőszeg-szerdahely Hills and the Gyöngyös River plain (Fig. 6.1). The main mass of the mountains is built of crystalline rocks, structurally assigned to the Penninicum Unit (Schmidt 1956). The mountains represent a tectonic window (Kőszeg-Rohonc/Reichnitz Window), similar to the Tauern-Engandin Window,

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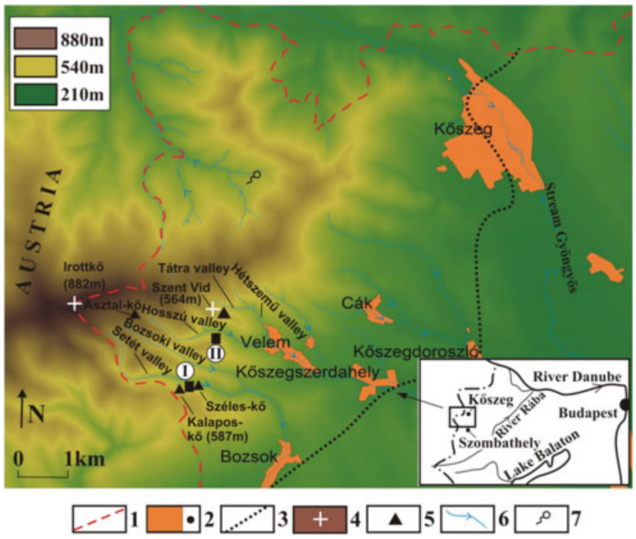


**Fig. 6.1** Location of measurement sites in the southern Kőszeg Mountains. *1* National border; *2* settlement; *3* boundary of the Kőszeg Mountains; *4* peak; *5* rock formation; *6* stream; *7* spring

which exposes rock series from below the Austro-Alpine Nappe (Wieseneder 1932). The major part of the window lies in Austria, only the Kőszeg Mountains and Vas Hill (to the south) are situated in Hungary.

The main component rock, phyllite, was generated from littoral pelitic, calcareous and detrital (sands and gravels) deposits in the southern shelf region, not far from the coastline. The lowest topographic position on surface is taken by quartzite phyllite, overlain by normal phyllites, a variety of which is calcareous phyllite, in turn overlain by greenschists. The greenschists derive from the metamorphosis of the oceanic crust (Lelkes-Felvári 1998), dated at 28–31 Ma (Balogh et al. 1983). They occur in five major and eight minor patches (Kisházi and Ivancsics 1984). The carbonate content is high in both calcareous phyllite and greenschists and may even reach 30 % (mostly in greenschists). This high content bears on calcareous and tuffaceous deposition in some streams. There is a spring in the area where $Ca^{2+}$ content reaches 90 mg $L^{-1}$ (Veress et al. 2005). The elevated calcareous content accounts for the fact that certain opened cavities show carbonate precipitations (stalactites and/or stalagmites of some centimetres length).

Carbonate in the greenschists occurs in two forms: either as interbeddings between the strata or along the foliation planes as thin crusts and veneers, or in disperse distribution. The latter has continuous distribution in calcareous phyllite whereas in greenschists it occurs locally. The local occurrences can have various positions: horizontal, vertical, developing as a matter of fact in any spatial direction.

The Kőszeg Mountains forms a ridge of north to south direction. The tributaries of the Gyöngyös River created a system of valleys of west to east alignment. The interfluves between these valleys form secondary ridges with steep walls and locally even canyon-like morphology (Kárpát and Ádám 1975). The southern dip of schistosity planes is perpendicular to the general directions of the valleys. Consequently, while the northern sides of the valleys are formed along the planes of schistosity, in the southern sides the heads of schistose beds are exposed.

The southern valley sides, built of greenschists and calcareous phyllite, show benches alternating with cliffs. On the interfluvial ridges spectacular rock formations are found, including the Kalapos-kő (Hat Rock, the name indicating its shape, widening upwards) on the high ridge stretching southwards from the Sötét Valley.

Caverns of various sizes are characteristic for both the valley sides and the ridge tops. Earlier they were explained by "pseudokarstic" processes (Földvári et al. 1948).

## 6.3 Features of Calcareous Rocks

On the surface of these rocks (particularly of those on valley sides) semicarbonate elements were formed due to the differential dissolution, manifested either as pits or protrusions. The former group of greenschist features includes solution notches (kamenitzas), niches, ducts (vertically elongated, circular or elliptical channels, sometimes of cave size); while the latter are pillars, hats and terraces (Veress et al. 1998). On calcareous phyllite solution notches, niches, narrow passages (caves), mouldings, pillars and terraces are observed, but hat-shaped features are missing. The likely reason is that this type of rock is not sufficiently hard to resist erosion and to maintain a protruding form. Carbonate distribution is more uniform in the calcareous phyllite than in the greenschists and, as a consequence, denudation is also more uniform in the former.

In addition to the Hat Rock, similar landforms in the greenschist domain are the Széles-kő (Wide Rock) and Asztal-kő (Table Rock). Morphological features of calcareous phyllite of characteristic shape are found in the Hosszú Valley and on terraces accompanying the Sötét, Bozsoki, Hosszú, Tátra and Hétszemű Valleys.

Solution notches (Fig. 6.2) are horizontal pits 1–2 cm high and even several metres' long on cliff faces. Their depths range from 1–2 cm to 1–2 m. Similar morphological elements are found in carbonate rocks; for instance, mostly in tropical environments, on the boundary between the parent rock and the soil (deposits), where the permanent presence of water and $CO_2$ augments solution (Slabe and Liu 2009). The process was named 'corrosion under wash' (Balázs 1962), and the generated features are grooves (Paton 1963), solution notches (Jennings 1973) and swamp notches (Ollier 1984). Due to the dissolution on the boundary between the soil and the underlying deposits, characteristic

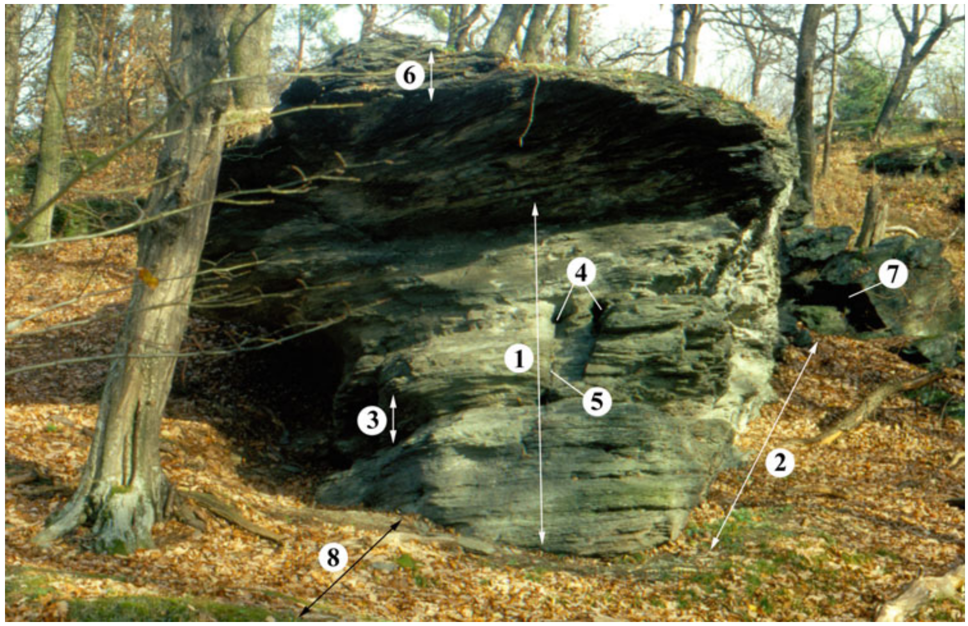


**Fig. 6.2** Complex notch (at the western edge of Hat Rock). *1* Uncovered portion of a big notch; *2* filled-up notch; *3* medium-size notch; *4* passage; *5* fissures; *6* top of the hat; *7* ruined cave; *8* present-day terrace development

for the piedmont of inselbergs, the resultant features may have lengths over 100 m, heights of 6 m and depths of 20 m. According to Sweeting (1973) this dissolution is responsible for the development and maintenance of abrupt slopes of inselbergs.

The length of the niches is double their height and actually they are nothing more than short solution notches. The pillars are the bastion-shaped rock formations, protruding from the surface between niches. The mouldings are step-like features of some tens of centimetres in size. The "hats" (Fig. 6.3) rise above their surroundings by several metres and their upper parts are characteristically wider. If notches develop on one side, an asymmetrical hat results. If they occur on both sides, the hat becomes symmetrical. The terraces are planar features on the valley sides, sloping towards the valley axis and tens of meters wide. Their length can reach even 100 m.

## 6.4 The Origin of Different Features

In the greenschists the distribution of carbonate content is not uniform. It shows considerable variation, especially in vertical profile (Figs. 6.4 and 6.5). It is evident that hollows

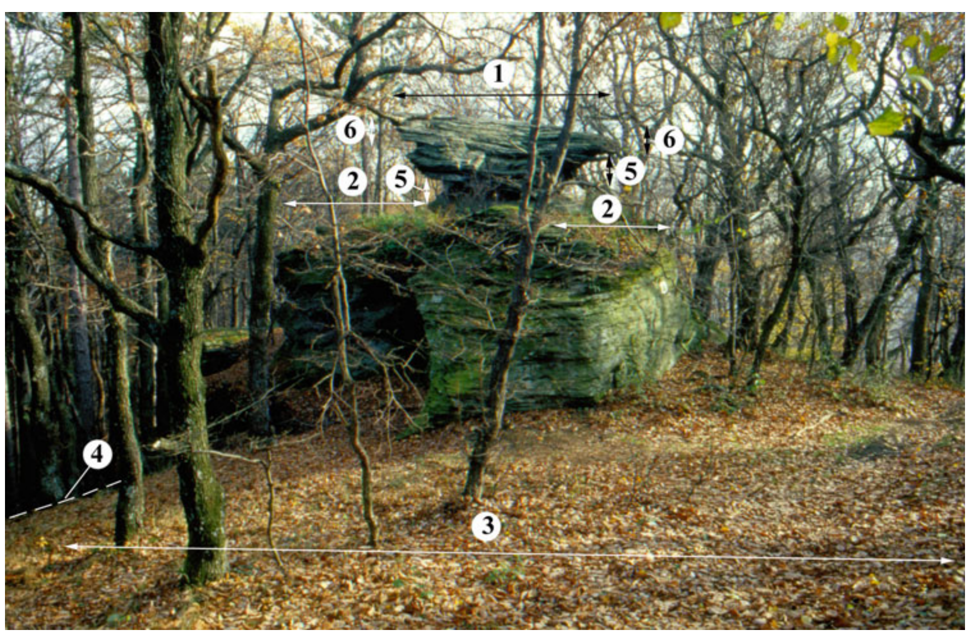


**Fig. 6.3** Symmetrical, double-sided hat and terraces of various ages on Hat Rock. *1* Remnant of the original surface; *2* remnant of the upper terrace; *3* remnant of the lower terrace; *4* presently forming terrace edge; *5* notch; *6* top of the double-sided hat

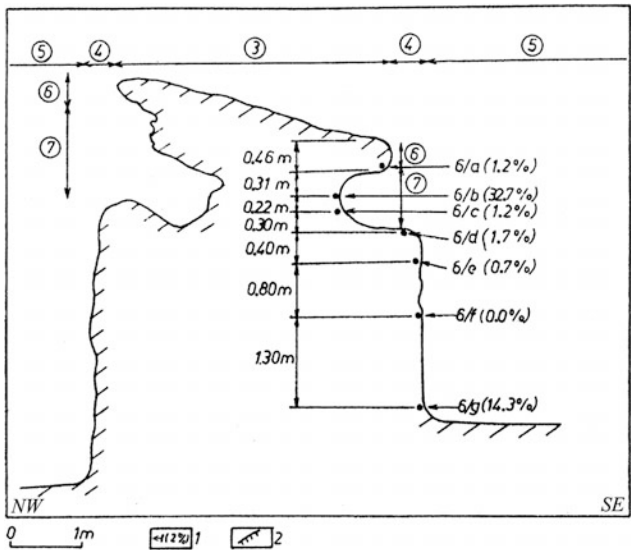


**Fig. 6.4** Location and carbonate content (Veress et al. 1998) of the samples taken from the double-sided hat (Fig. 6.3). *1* Sampling location and number (in brackets the $CaCO_3$ content in percentage); *2* Greenschist; *3* original surface; *4* upper level; *5* lower level; *6* hat; *7* notch

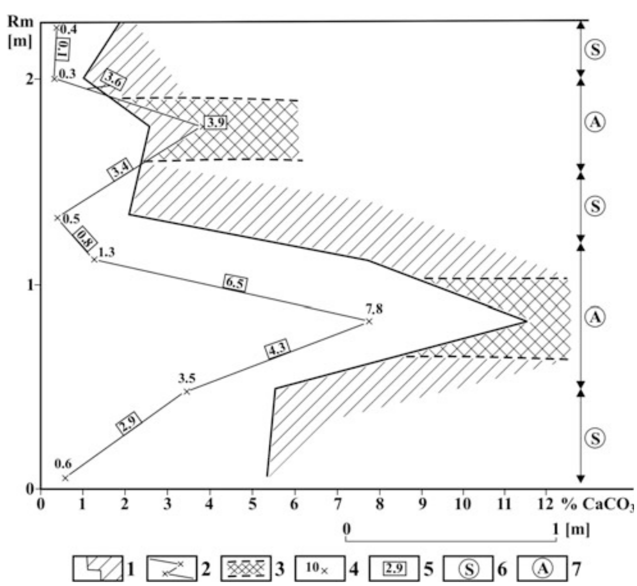


**Fig. 6.5** Zones of different morphological evolution according to the carbonate content for a big solution notch (Veress et al. 1998). *1* Rock (*side view*); *2* carbonate content along the wall (pointwise sampled); *3* high carbonate content compared to environs; *4* sampling sites; *5* difference of carbonate content between neighbouring samples (in %); *6* rock without dissolution traces due to the low carbonate content; *7* rock rich in morphological features due to high carbonate content; *Rm* Relative height

**Fig. 6.6** Ice from the niches in greenschists, proving the presence of infiltration and water outflow

appear on the walls due to elevated carbonate content because in such portions of the rock the presence of carbonates promotes more intensive denudation through differential dissolution, which affects both the surface and the internal mass of the rock. On those surfaces upon which schistosity creates favourable conditions (by properly inclined planes), water can infiltrate into the rock. The dissolution affects the entire surface of the rock between the infiltration points. In this case only minor features can be created. On the surfaces where the inclination of the foliation is towards the valley, water is flowing out from the rock (Figs. 6.6 and 6.7). At these points dissolution capacity is increased because water flow is concentrated. The resulting features are fewer in number, but larger in size. Direct dissolution is more active on rock surfaces with gentle inclination. Overhanging cliffs, on the other hand, are fully protected from raindrop impact.

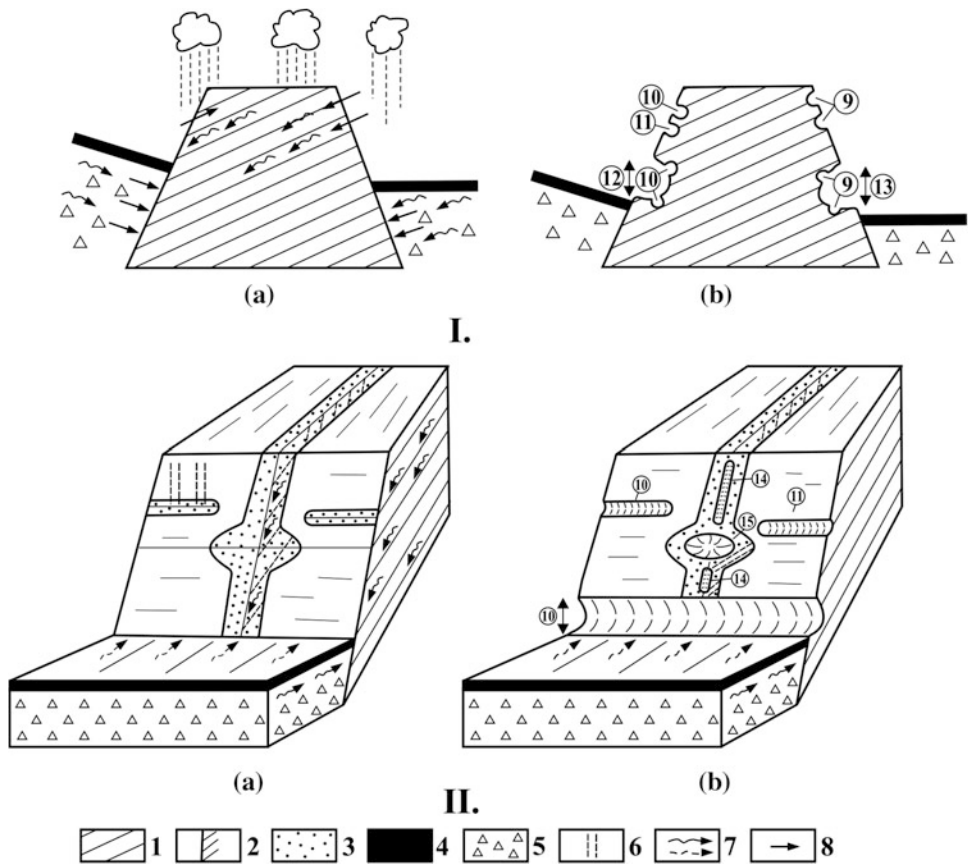


**Fig. 6.7** Morphological elements originated by the partial dissolution in greenschists. *I* Side view; *II* in 3D; **a** original situation; **b** fully developed notches; *1* Schistosity surface; *2* fault plane; *3* carbonate pocket; *4* soil; *5* debris and weathered material; *6* precipitation; *7* water infiltration; *8* dissolution; *9* small to medium-size solution notch on the infiltration point; *10* small to medium-size solution notch due to direct dissolution; *11* small to medium-size solution notch with water outflow from rock; *12* large solution notch generated under the soil due to water outflow; *13* large solution notch generated under the soil due to infiltration; *14* vertically elongated hole; *15* circular hole/cave with water outflow

Water flow through the rocks is allowed by foliation and faulting (Fig. 6.7). If the water flow follows foliation planes, and the carbonate pocket is more or less horizontal, then small (several centimetres' high) or medium-size (a few tens of centimetres' high) solution notches and niches are generated. Fissures are generated along fault planes and if the carbonate bed is vertical. If the carbonate has a uniform distribution and the water flows along the foliation and the faults, circular or elliptical niches are generated.

The rock can suffer indirect dissolution if the precipitation is retained by the soils and the weathered zones, which continuously recharge water to the rock surface and maintain an intensive dissolution due to long-lasting water-rock contact and increased $CO_2$ content. As a result, medium-size notches are formed on the escarpment and may develop into larger features. It happens if the surface next to the escarpment is lowered. Thus, notch formation takes place in a deeper position and the notch height increases.

On barren slopes partial dissolution and frost action mutually strengthen each other, particularly if foliation planes conduct water into the rock. The infiltrated water can freeze along the foliation planes, which open and thus allow more water to enter into the rock, enhancing further dissolution and frost action.

## 6.5 Slope and Form Evolution on Greenschist

Spectacular manifestations of the above processes can be traced on the Hat Rock in the form of notches and outcrop asymmetry (Fig. 6.8). The hat is disintegrating due to interactions between notch evolution and the resulting weakened support. The development of solution notches and the resulting slope retreat are hindered by the thinning carbonate layer and the transport of soil and weathered

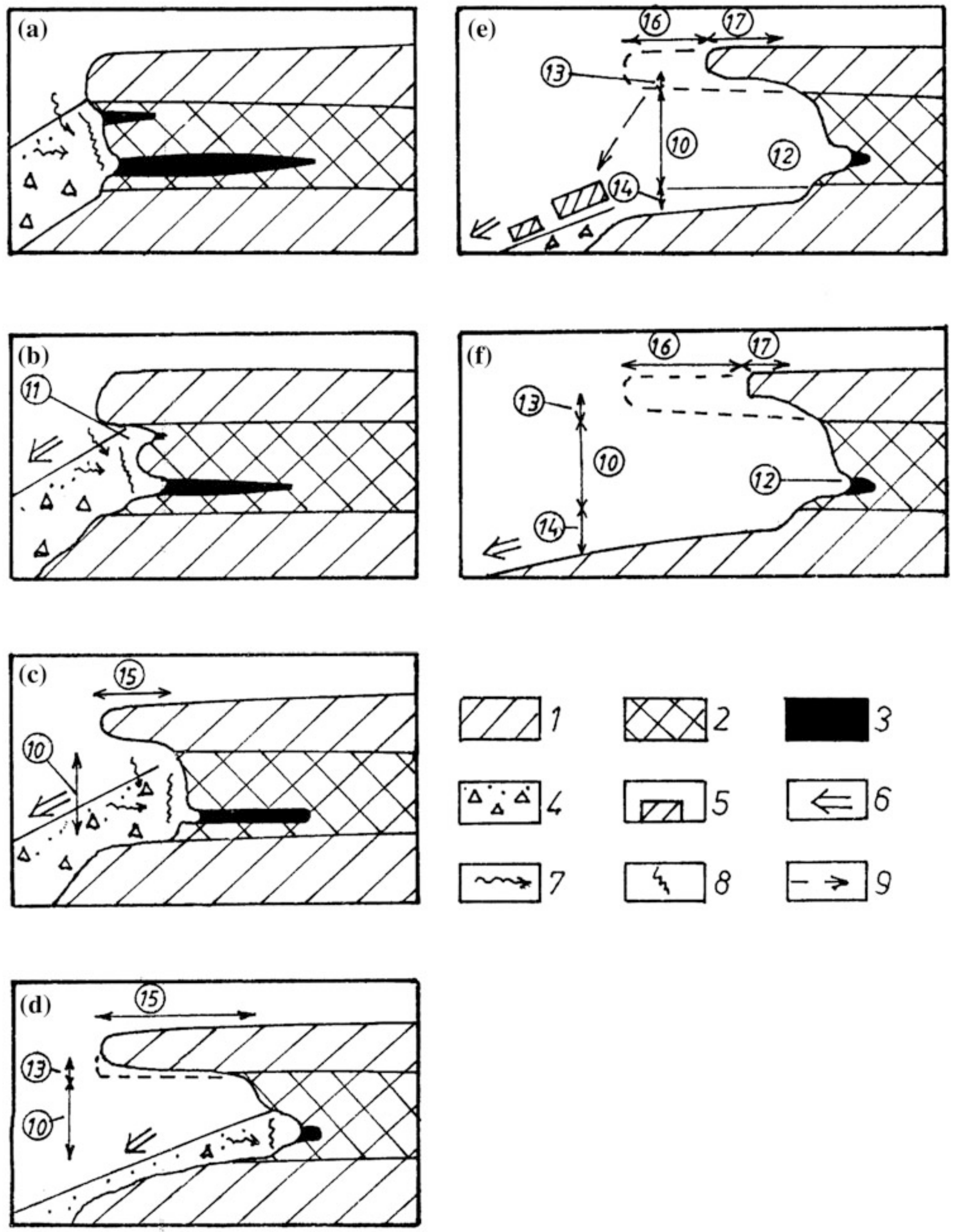


**Fig. 6.8** Origin and denudation of the hat. *1* Greenschist; *2* carbonate pocket; *3* limestone; *4* soil and frost weathering debris; *5* disintegrated blocks; *6* colluvia; *7* infiltration; *8* dissolution; *9* trajectory of weathered material; *10* segment of large solution notch; *11* weathered medium-size notch; *12* conserved medium-size notch; *13* segment of notch generated by disintegration and frost weathering; *14* segment of notch generated by frost weathering; *15* hat; *16* destroyed segment of hat; *17* remnant of hat; **a**–**b** generation of combined (large) notches; **c**–**d** hat generation; **e**–**f** denudation of hat and large notch

deposits from the notch. The deposits can be replenished with the material eroded from the hats, and backwearing continues as long as material supply exceeds transport. The rate of destruction (and thus the supply of material) depends on the consistency of the rock and the rate of development of solution notches. If backwearing lasts long, the lower sheets of the notches will become terraces (Fig. 6.9).

On the Hat Rock various traces of partial dissolution and backwearing are easily identifiable: double-sided and one-sided hat shapes, terraces and all remnants of denudation are present. The terraces are found on the more elevated zones of the Hat Rock area, on both sides (Fig. 6.3). The upper terrace (no. 2) was eroded into a remnant due to the widening of terrace no. 3 lying below then terrace no. 3 also turned into a remnant through slope retreat (Fig. 6.2).

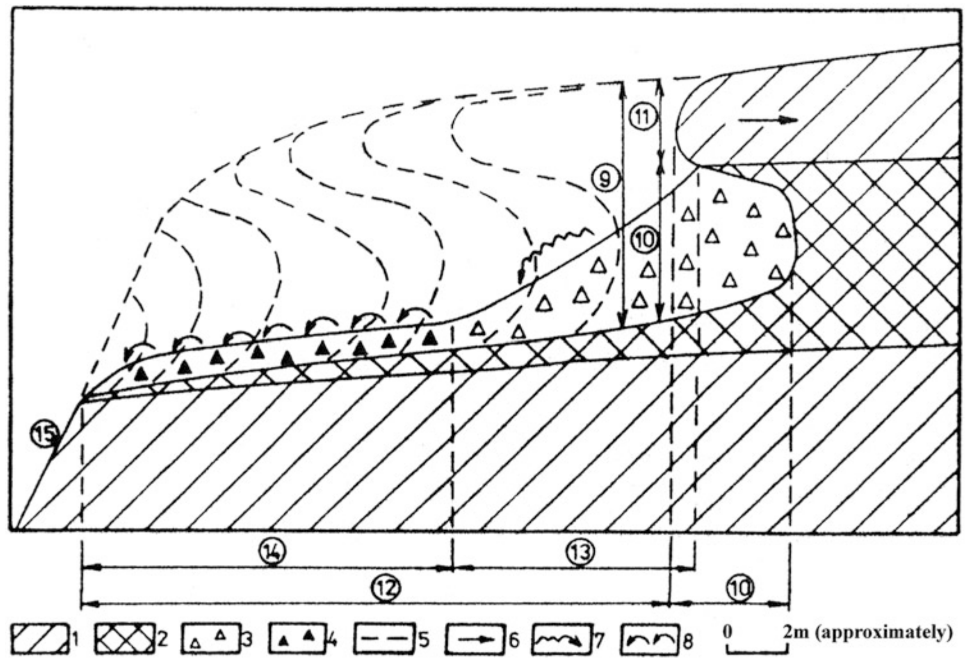


**Fig. 6.9** Origin of a terrace. *1* Greenschist; *2* calcareous pocket; *3* deposits of partial dissolution and frost shattering; *4* further transformation and disintegration of deposits; *5* bedrock at different dates; *6* direction of backwearing; *7* transport of debris by mass movement and sheet wash; *8* remnants of debris transported by sheet wash; *9* backwearing rockwall; *10* solution notch; *11* hat; *12* terraces; *13* inner part of terrace; *14* outer part of terrace; *15* valley side

## 6.6 Conclusions

In the Kőszeg Mountains the rocks with carbonate content undergo differential dissolution and disintegrate. The resulting denudation features, of which the Hat Rock is the most picturesque, are considered unique in Hungary.

The morphology of such features is determined by the structure of the rock, the extent of its carbonate content, and the regime of water flow along schistosity or fault planes. The dimensions of landforms are again controlled by the carbonate content, the existence or lack of a soil mantle and weathered deposits, the coverage and longevity (stability) of the surface affected. Under such conditions notches, niches, hats and other rock formations are generated. Notch development involves the backwearing of the walls, a self-supporting process. The deposits are mostly transported, but also supplied with material from further disintegration of the hat. The bases of notches form terraces on the slopes of the major valleys. On calcareous phyllite, however, backwearing does not generate any morphological elements with hat shape.

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# 7 Kab Mountain: Karst Under a Basalt Cap

Márton Veress and Zoltán Unger

**Abstract**
A limestone-dolomite horst of the Southern Bakony Mountains, Kab Mountain is capped by basalt lava and mantled by loess. Karstic phenomena occur on the rim of the basalt terrain and in the karstic windows. On the rim of the basalt one can find ponors in blind valleys. Karstic windows, an outstanding feature of Kab Mountain, were formed on sites where no or only thin basalt covered the limestone or where the limestone outcrops from below the basalt layer in the valleys. Where the limestone is exposed ponors occur, whereas subsidence dolines are found in places where the limestone is mantled by loess. If the limestone is covered by a thin basalt layer, caprock dolines generated by surface collapse may transform into ponors with time. In karstic windows extensive systems with specific elements could develop: epigenetic valleys, ponors, caprock dolines and subsidence dolines.

**Keywords**
Basalt • Loess • Karstic window • Ponor • Blind valley • Inflow cave • Subsidence dolines • Caprock dolines

## 7.1 Introduction

According to the relative position of rocks prone or not prone to karstification, allogenic, autogenic and mixed karsts are identified (Jakucs 1977; Ford and Williams 2007). The autogenic karst is exclusively fed by precipitation, whereas for the allogenic karst additional water supply comes from the surrounding non-karstic areas sloping in the direction of the karstic terrain. In mixed karsts the elements of autogenic and allogenic karsts occur side by side. In the case of covered karsts, the overlying deposits are non-karstic. If the superficial deposits are impermeable, it is a cryptokarst, if permeable, it is a concealed karst. All these types are present on Kab Mountain.

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The interface of rocks prone and not prone to karstification may run along a rock (lithological) boundary. On an allogenic karst with cryptokarst contact this rock boundary is a true junction (boundary of the limestone outcrop), but for a concealed karst it is a concealed junction (Veress 2009). In the case of a true junction, runoff from the non-karstic surfaces reaches the karstic rocks through ponors. A concealed junction is formed at thin and permeable caps, through which infiltrating water induces karstification of the bedrock, manifested morphologically in dolines of superficial deposit. The origin of such dolines is explained by the process that fragments of the caprock fall into the shaft of the bedrock.

The various types of karst are characterized by well-defined morphological features. On autogenic karsts the morphology is dominated by solution dolines, uvalas, and solution caves, while in the case of allogenic karsts mostly ponors and erosion caves develop on the margins. On cryptokarsts caprock dolines are found, ponors on valley floors and epigenetic valleys are formed. With valley incision the rock boundary retreats, this allows further ponors to form, while the older ponors are turned into dolines (Jakucs 1977).

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Caprock dolines result if the non-karstic consolidated caprock collapses into karst hollows.

The dolines of concealed karsts are dropout dolines and suffosion dolines. Solution chimneys and shafts accompany concealed karst evolution (Veress 2009). If solution is relatively rapid or the superficial deposits are cohesive rocks, dropout dolines result, and if it is a less rapid process, suffosion dolines will be the outcome (Drumm et al. 1990; Tharp 1999; Williams 2003). A remarkable morphological element of the covered karst is the depression of superficial deposits, which forms where the locally eroded superficial deposit is resedimented within the karst through subsidence dolines and ponors (Veress 2009).

## 7.2 Geology and Geomorphology

The above processes can be well demonstrated in the geomorphic evolution of Kab Mountain, a 600-m high horst of the Southern Bakony (Transdanubian Mountains), capped by basalt rock (Fig. 7.1). The area of the basalt cap is ca 5 km$^2$. The mountain is built up of Triassic dolomite and Mesozoic and Eocene limestones, while Jurassic and Cretaceous limestones only outcrop in a limited area along the northern edge. Thin Pannonian basalt lava flows covered a slight dip on the surface of sedimentary rocks. The radiometric age of the volcanic activity is determined at 7–5 Ma (Balogh et al. 1982). The generation of numerous individual lava flows from distinct centres was preceded by pyroclastic activity (Lóczy 1916; Jámbor 1980). In the periods of quietude between volcanic phases, red clays deposited on the basaltic surfaces under a wet and warm climate (Jámbor 1980). The volcanic activity is considered to be part of the Balaton Highland volcanism, during which, due to crustal thinning and fragmentation, melted material from the mantle reached the surface in several places of the Carpathian Basin (Stegena et al. 1975).

The lava flows covered a surface dissected by karstic features and epigenetic valleys. Paleokarstic features include solution dolines, uvalas separated by mounds. The karstic

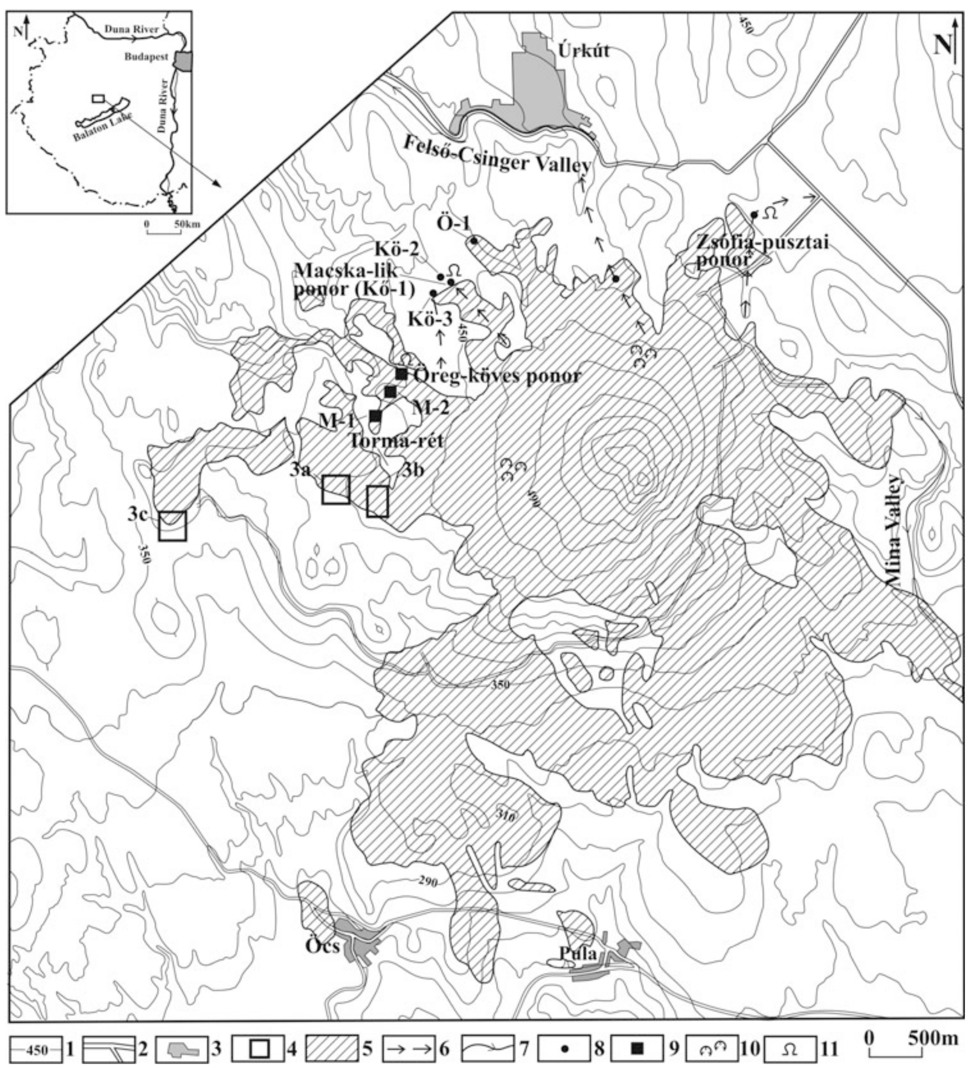


**Fig. 7.1** Map of Kab Mountain. *1* Contour line; *2* road; *3* settlement; *4* the area shown in Fig. 7.3; *5* basalt cap; *6* intermittent watercourse; *7* watercourse; *8* karst feature; *9* complex karst feature, depression of superficial deposit (subsidence dolines, caprock dolines, ponor); *10* dolines in debris; *11* inflow caves

mounds were either totally covered by the basalts of varying thickness or only surrounded by the lava flow. The epigenetic valleys incised into the limestone were partially or totally filled with basalt. After the basalt solidified, regressing valleys cut into the basalt surface, starting from its edge. Valleys due to karstification (ponor formation) also occurred locally within the basalt cap. In such places the watercourses incised into the basalt cap and created epigenetic valleys in the limestone. During the glacial periods loess deposited both on uncovered karsts and on the basalt cap. The basalt cap has been fragmented into blocks of various size to our days.

## 7.3 Karst Processes and Landforms

Present-day karst processes on Kab Mountain affect the edge of the basalt cap (at the boundary of the limestone outcrop), the cryptokarst in karst windows (i.e. where the limestone is exposed within the basalt terrain, or where the basalt is thinning out), the concealed karst and basalt debris (Fig. 7.2). The following landforms can be identified: ponors (at the limestone boundary); caprock dolines (on basalt); dolines in basalt debris; subsidence dolines (in loess); depressions of superficial deposit (a complex karstic system); shafts or inflow caves (under ponors). The karstic processes are controlled by the development of chimneys and shafts in the bedrock.

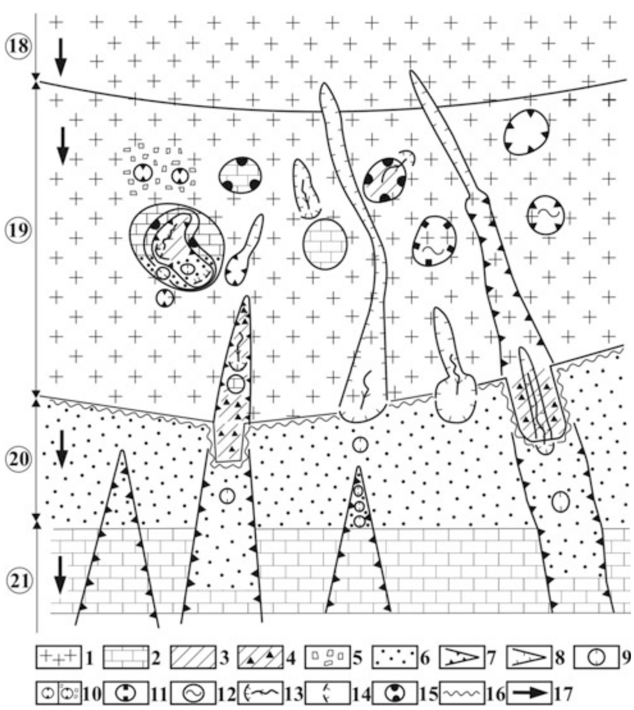


**Fig. 7.2** Sketch of karstic zones on Kab Mountain (without scale). *1* Impermeable rock (basalt); *2* limestone; *3* weathering residue; *4* redeposited fluvial sediment; *5* in situ basalt debris; *6* permeable superficial deposit (loess); *7* epigenetic valley; *8* creek; *9* subsidence doline; *10* caprock doline; *11* former depression filled with basalt; *12* pond, waterlogged area; *13* ponor; *14* passive ponor; *15* depression of superficial deposit; *16* boundary of exposed limestone; *17* surface slope; *18* buried karst; *19* buried karst dissected by karst windows; *20* concealed karst; *21* uncovered karst

The ponors mostly occur in blind valleys, on the outcropping limestone mounds, along the edge of the basalt cap, on the floors of valleys which cut through the basalt cap and at the limestone outcrops within the basalt terrain. In the latter cases solitary ponors may have been formed, or depressions functioning like ponors (for instance, no Bk-4, Fig. 7.3a). Due to a prolonged evolution, however, complex karstic systems (M-2, Fig. 7.4) are likely to have been generated.

The caprock dolines on Kab Mountain are not typical representatives of this landform type. While typical caprock dolines have steep (vertical) walls, with limestone outcrop and debris pile on the bottom (Waltham et al. 2005), the caprocks on basalt terrain show more gentle slopes, without limestone outcrops or debris accumulations. They are often horizontally (consisting of subdepressions) as well as vertically complex dolines (several dolines embedded in main dolines). The reason for the low inclination of the doline slopes is the development of shafts in the bedrock, which continue into the basalt, without the formation of major caves. The embedded internal dolines are subsidence dolines if the basalt (and also the main dolines) is covered by red clay or loess deposits. Among the caprock dolines some are not linked to any watercourse or morphological element (as, for instance, the depression marked Ö-1), while others have such links (as the depression Bk-6, Fig. 7.3a).

The subsidence dolines of the concealed karst (in loess deposits above limestone shafts) are of two types differentiated by the following conditions:

1. the basalt cap has no impact upon the dolines generated in the loess (e.g. Kö-2 dolines—Fig. 7.1 and Fig. 7.3b);
2. or the subsidence dolines generated near the edge of the basalt cap are fed by runoff from the basalt terrain. In the latter case the ponor-like character of the doline is more highlighted (subsidence dolines transforming to ponors). The dolines may continue in inflow caves (as the Bujólik cave—Fig. 7.3c).

Ponors are formed at the margin of the basalt terrain, at the boundary of the limestone outcrop, and fed by permanent or temporary runoff from the basalt cap (Fig. 7.2). The valleys are transformed into blind valleys. Such a ponor is the Macska-lik (Fig. 7.5) or the nearby ponor marked Kö-3. It may happen that the ponor is not associated with the rim of the basalt cap, but occurs on the bottom of the epigenetic valley without any basalt coverage if the bottom of the epigenetic valley is covered by the debris transported from the valley generated in the basalt terrain. Ponors may also

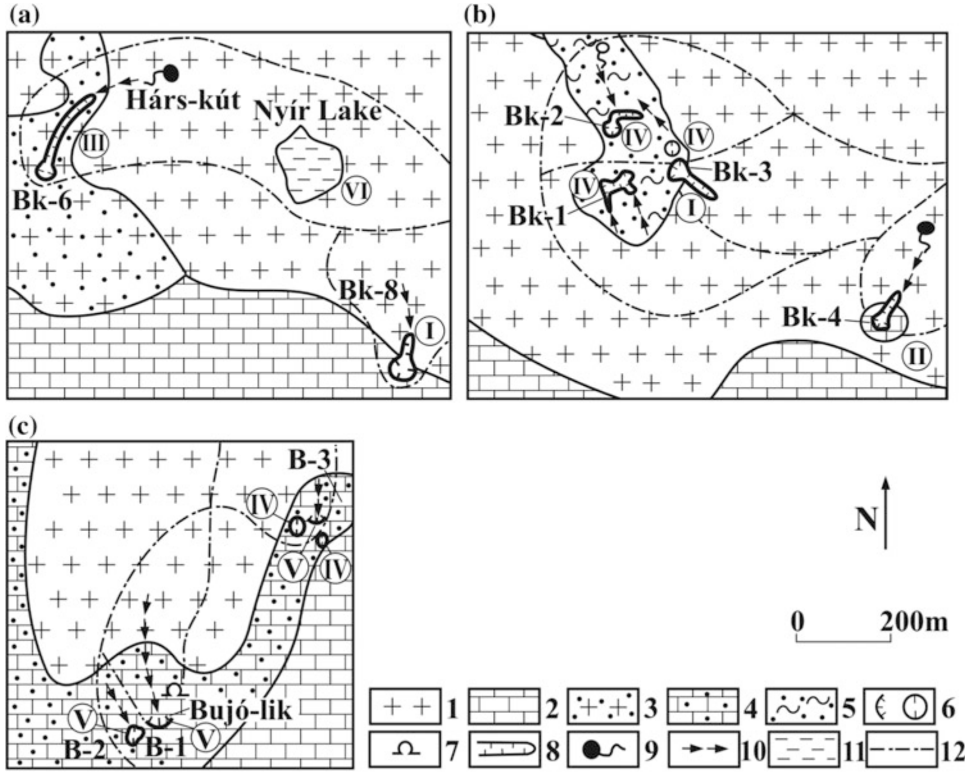


**Fig. 7.3** Karst features on Kab Mountain and environs. *1* Basalt; *2* limestone; *3* loess on basalt; *4* loess on limestone; *5* loess on red clay; *6* karstic depression; *7* inflow cave; *8* gully, ravine; *9* spring; *10* (intermittent and permanent) watercourses; *11* pond; *12* watershed. *I* ponor on outcrop boundary; *II* karst depression turning into ponor on limestone surface; *III* caprock doline turning into ponor; *IV* subsidence doline; *V* subsidence doline turning into ponor (fed by runoff from the basalt terrain); *VI* pond, waterlogged depression

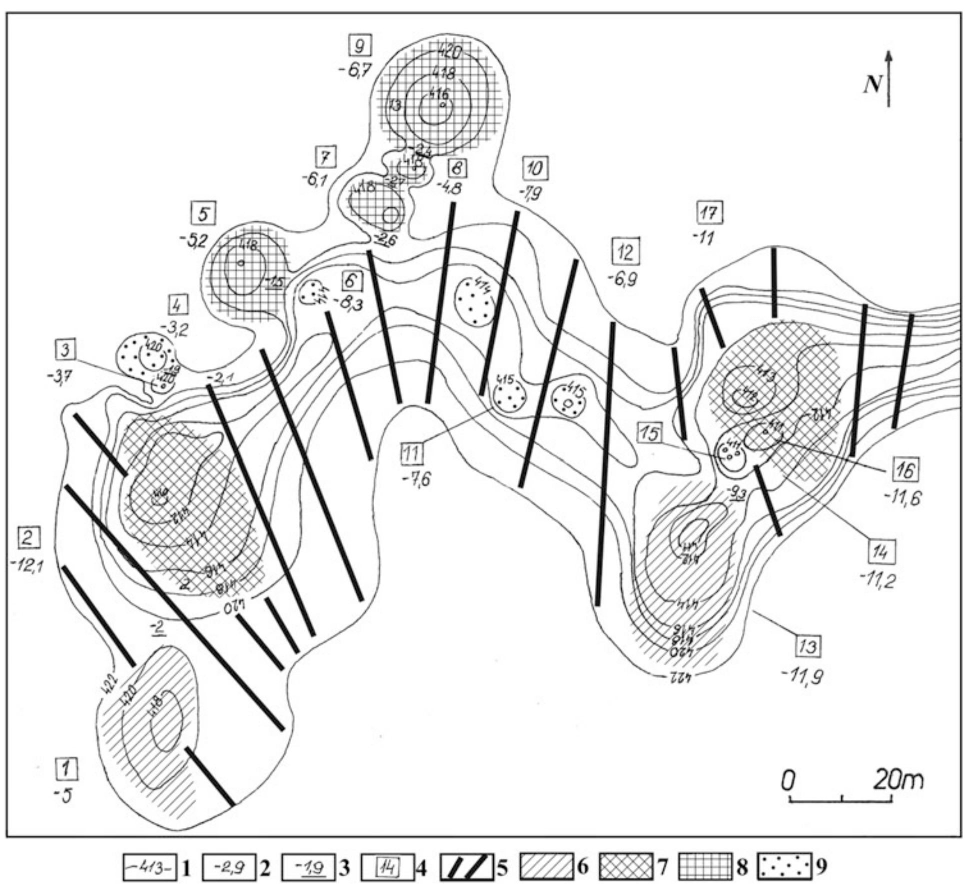


**Fig. 7.4** Contour map of the M-2 system. *1* Contour lines; *2* the deepest point of a karstic depression (started from 422 m elevation); *3* the depth of sill between subdepressions (from 422 m level); *4* code of depression; *5* epigenetic valley; *6* passive ponor (doline); *7* intermittently active ponor; *8* caprock dolines; *9* subsidence doline

**Fig. 7.5** The Macska-lik ponor

emerge on the basalt terrain where limestone is exposed on the valley bottom. This usually happens on the elevated segments of the limestone surfaces, where the basalt coverage is thinner.

Karstification in karst windows can happen in four different ways:

1. If the basalt partly covers the limestone mounds, a basalt-limestone boundary is formed and this circumstance promotes ponor generation, and, through back-step denudation, gully formation (Fig. 7.6a). Presently, an initial stage of the phenomenon can be studied at the ponor marked Bk-4 (Fig. 7.3b). Gullies can develop into true valleys if the surface is steep enough and there is no other watercourse in the vicinity. The resulting valley will cut into the thin basalt layer (Fig. 7.6b). On the bottom of the valley a new outcrop boundary is formed, retreating and generating new ponors (Fig. 7.6c, d), while the older ponors are filled up (fossilized). Such an epigenetic valley developed into a karstic system is the depression marked as M-2 (Fig. 7.4). If all these are covered by loess, on the flanks and bottoms subsidence dolines are generated (Figs. 7.6e and 7.7), while on and near the edges caprock dolines also occur (Fig. 7.6e). The doline marked M-1 functions as a ponor with water inflow, as a valley developed leading to it. The caprock dolines might have been generated near the M-2 system, which had hosted a former ponor whose shaft continued in the limestone below the basalt. Over this shaft, the limestone and the basalt have been crashed. The development of caprock dolines near the M-2 system might have been promoted by the fact that precipitation infiltrated towards the shaft, which consequently expanded and additional shafts also developed (Fig. 7.8). The complex karstic system (depression of superficial deposit) has a various morphological footprint: some of those preserve the character of the former valleys, but they are separated into counterslope segments with ponors, gullies, creeks, remnants of mass movements, filled ponors, subsidence dolines, caprock dolines (Fig. 7.6). The ponors continue in inflow caves as dissoluble shafts (Fig. 7.9).
2. If the limestone elevations were surrounded by basalt rock, no karstic processes will happen under the caprock, but ponors are created on the limestone crests. Due to the substantial runoff from the basalt, the surface of the limestone is intensely lowering. Solution dolines of a very short lifetime are formed. As shafts emerge, the morphological elements are being transformed into ponors. The limestone mounds lower below the basalt surface and, accordingly, are filled with or buried under debris from its surroundings. The surface is becoming more impermeable, and, as a consequence, after some time rainwater may flow off from the limestone terrain. On the edge of the covered surface, near the basalt, ponors are generated.
3. If the basalt entirely covers the limestone mounds, caprock dolines are generated. It was already mentioned that caprock doline formation is controlled by the existence of chimneys and shafts in the bedrock. Water infiltrating

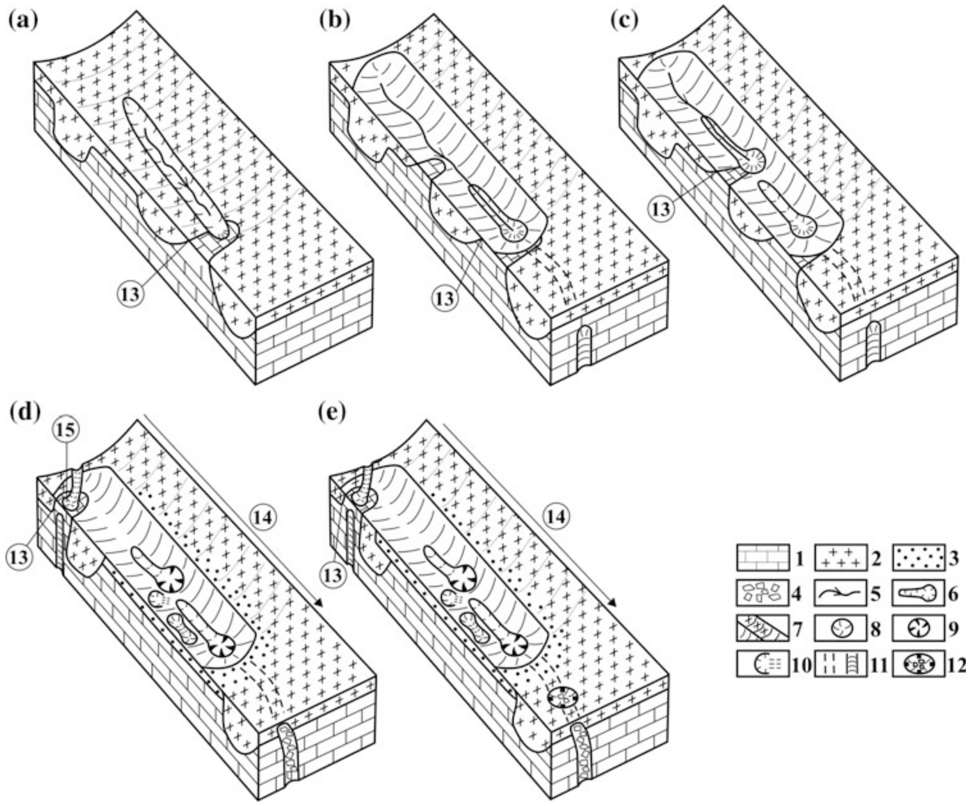


**Fig. 7.6** The development of a complex karstic system (depression of superficial deposit) presented by the M-1 caprock dolines and M-2 system. *1* Limestone; *2* basalt; *3* loess; *4* debris; *5* watercourse; *6* ponor and its valley; *7* depression on basalt and side of epigenetic valley; *8* subsidence doline; *9* passive ponor (doline); *10* asymmetric suffosion doline; *11* water burrow of the ponor/chimney; *12* caprock doline; *13* limestone outcrop; *14* depression of superficial deposit; *15* emerging rock boundary; **a** ponor formed at uncovered karstic bedrock; **b** epigenetic valley segment on limestone; **c** rest of the epigenetic valley on the limestone with new ponors; **d** a third ponor generated on the new rock boundary, older ponors filled and transformed into dolines; **e** loess deposition, subsidence dolines develop on the floor, in the side and at the edge of the valley, while a caprock doline develops on basalt


**Fig. 7.7** Subsidence dolines on the margin of the M-2 system. *1* Subsidence dolines; *2* valley segment

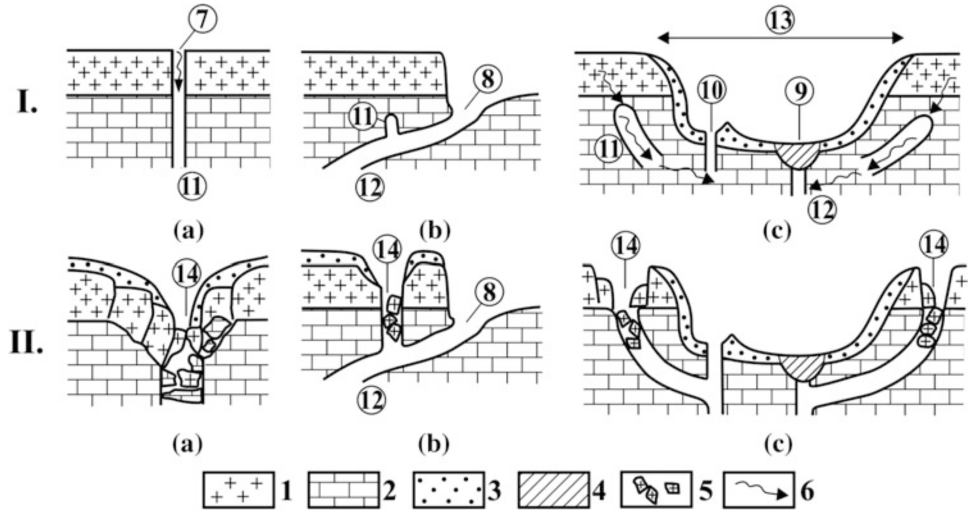


**Fig. 7.8** Modes of shaft and caprock doline formation. **a** water infiltration through basalt fissures; **b** at inflow cave; **c** by infiltration from basalt terrain towards depression; *1* basalt; *2* limestone; *3* loess; *4* infilling of ponor; *5* water infiltration; *6* fracture; *7* ponor; *8* infilled ponor; *9* subsidence doline; *10* shaft; *11* water burrow of inflow cave; *12* former valley; *13* caprock doline; *14* debris

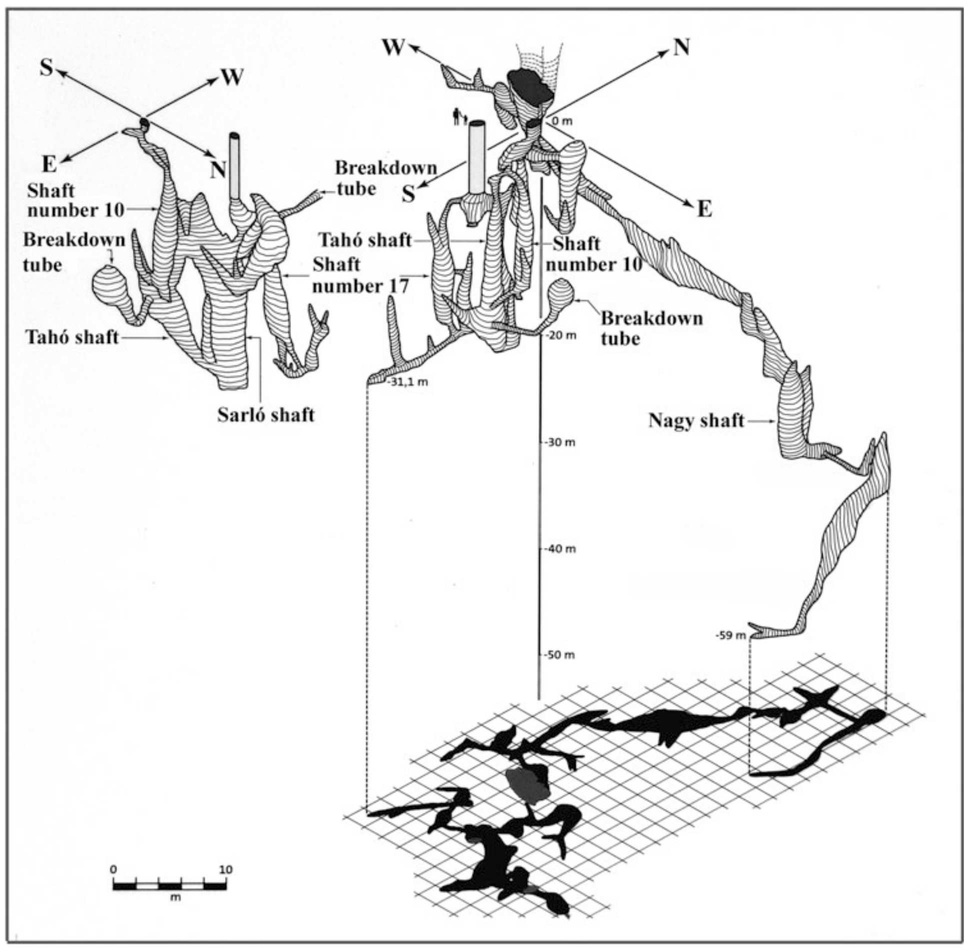


**Fig. 7.9** Isometric map of Öreg-köves ponor cave

through the fractures of the basalt dissolves the limestone bedrock (Győrffy 1957). Similarly, runoff from the clay which is situated between the beds also dissolves the slopes of the limestone mounds (Németh 2005). The caprock dolines on Kab Mountain are at different phases of development and, as a consequence, are of various sizes: from 1 to 2 m diameter and tens of centimetres' depth to dolines 20–30 m across and 5–10 m deep. They are fed by water inflow, which ensures rapid development and transformation into ponors, while their chimneys and shafts turn into inflow caves. In such karstic windows complex karstic systems can also evolve.

4. If the valleys expose the limestone, the karstic processes affect the limestone mounds. Ponors are formed on the bottom of the valleys, which are not blind valleys, but continue over the ponor.

Groups of depressions several metres across and deep may occur in places where the basalt is thin (from 10–20 cm to 1–2 m) and fragmented into blocks. Their origin can be explained by the dissolution of the bedrock and the cover debris suffering local subsidence—although human origin cannot be excluded either.

Where loess occurs, subsidence dolines are formed beyond the edge of the basalt cap, mostly on the bottom of valleys, and also in karst windows where loess is deposited on limestone (Figs. 7.4 and 7.7). Subsidence dolines (such as Bk-1 and Bk-3) are generated in the loess which overlies red clay (Fig. 7.3c). Fed by runoff from the basalt cap, subsidence dolines may also develop into ponors.

Locally shallow, wet, swampy ponds are found on the basalt surface (Fig. 7.3a). In addition to those generated as caprock elements (see above), they could form through the collapse of paleoholes/caves (Jugovics 1954) or by the subsidence of the basalt lava (Vörös 1966). It is also a plausible explanation that the paleodolines had been just partially filled by the lava flow.

As evidenced by speleological explorations, several inflow caves of hundreds of metres' length are related to ponors (Kocsis 2008) and dissolution shafts or shaft systems (Figs. 7.9). They follow fracture systems, show vertical segments linked by oblique passages, which are developed along bedding planes. Erosion only slightly influenced the original morphological elements (blind chimneys, bridge arches, windows) or did not influence them at all. The dissolution was facilitated by the fact that in these places considerable amount of water feeds the karst from the basalt terrain through ponors.

## 7.4 Conclusions

A wide range of recent karstic processes are active on the basalt terrain of Kab Mountain. The karstic processes either operate within the basalt terrain (in karstic windows), at the edge of the basalt cap or on the loess mantle (sometimes overlapping with the area of karstic windows).

Karstic windows are unknown in international literature and seem to be unique to Hungary. Karstic windows are found either at limestone outcrops or within thin basalt caps. Ponors are formed along the edges of the basalt cap and in karstic windows with limestone exposed. In karstic windows where the basalt is thin, caprock dolines are formed which may gradually turn into ponors. Due to the erosion of the basalt cap, the limestone bedrock is exposed, and no caprock dolines, but ponors are generated immediately.

Karstic windows are affected by dissolution (shaft formation in the bedrock) and fluvial erosion (valley generation on caprock). The karstic processes are complex: ponors, caprock and subsidence dolines, chimneys, shafts and inflow caves are generated, which are completed by valley formation and the appearance of epigenetic valleys. This is the reason why complex and vast karst systems (depressions of superficial deposits) are formed on the basalt cap of Kab Mountain.

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# 8 Buttes in the Tapolca Basin

Péter Gadányi

**Abstract**

The group of residual hills of the Tapolca Basin, extinct alkaline basalt volcanoes mostly of truncated cone shape, are picturesque landforms of the Transdanubian landscape. Pliocene volcanism began with phreatomagmatic explosions, generating tuff rings, continued by Strombolian activity and ended in Hawaiian effusions. Since the resistant crater fills are eroding at a much slower rate than the material of the tuff rings and the sands deposited in Lake Pannon during the Upper Miocene, high relief has resulted. The parallel retreat of the basalt fills created cryoplanational cliffs with platforms and screes. The basalt-capped mesas have been dissected and the resulting truncated or pointed cone shapes of the residual hills makes them similar to the famous buttes of the American West. Burial under soil cover intensified the weathering of basalt columns of more or less regular shape and modified their cross-sections. After exhumation the stability of the columns reduced and they became liable to toppling and disintegration. To the northwest another volcanic field of eruption centres about the same age borders the Little Hungarian Plain. Other features indirectly related to magmatic activity, including the Tapolca Lake Cave, are also geomorphosites popular among visitors.

**Keywords**

Basalt volcanism • Selective erosion • Tuff rings • Crater fill • Cryoplanation • Lake cave

## 8.1 Introduction

In the Transdanubian landscapes of subdued relief (also called “Pannonian” after the Roman province), a number of spectacular basalt volcanoes of 3–5 Ma age rise above the flat surface of the Tapolca and Marcal Basins and the Kemenesalja region (Fig. 8.1). Morphologically, the hills are composed of a broad “skirt” of gentle slopes at the base, steep basalt cliffs and flat, slightly convex, undulating or even pointed tops (pinnacles), where ruins of once strong border fortresses remind us of eventful times in Hungarian history. On the southern slopes with nutrient-rich soils formed on weathered volcanic debris vineyards produce exquisite wines, which were favoured by the Romans too.

The volcanic origin of the hills has been known since the first half of the 19th century. Scientists were also aware that the loose Upper Miocene sediments of Lake Pannon have been eroded from the environs of the volcanoes, while the more resistant basalt lava flows protected the underlying, easily erodible, Pannonian sands (Böckh 1874; Hofmann 1878; Vitális 1911; Lóczy 1916; Mauritz 1948). Thus, the residual hills, which would be called basalt-capped buttes, were they located in the western United States, bear the name of “witness hills” in Hungarian geographical literature, emphasizing that they preserve the level of the terrain present at the time of volcanism.

The good-quality basalt of the hills was intensively quarried between the early 20th century and the late 1950s and used as basalt cubes for road pavement or crushed basalt

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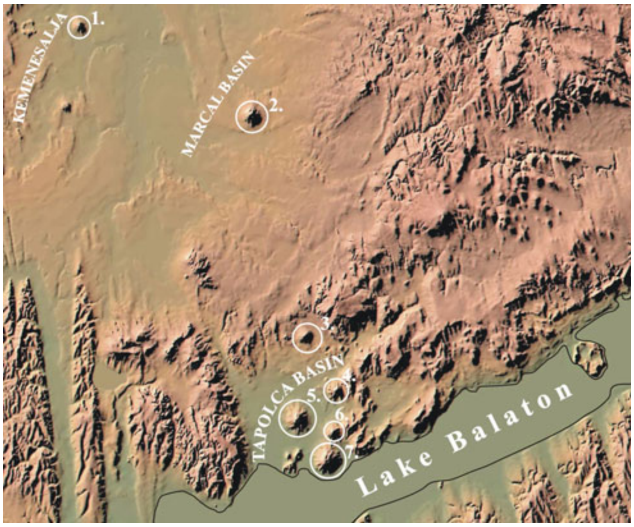


**Fig. 8.1** Location of basalt volcanoes in central Transdanubia. *1* Ság; *2* Somló; *3* Haláp; *4* Csobánc; *5* Szent György-hegy; *6* Gulács; *7* Badacsony

for railway embankments and other purposes (Kónya 2007). The quarries, however, also created exposures with insight into the geological buildup of the hills (Jugovics 1953, 1954, 1971). Comprehensive modern studies of volcanism in Transdanubia began in the 1950s (Martin and Németh 2004; Harangi 2011).

Unfortunately, the destruction accompanying mining did not spare the buttes and the top section of Haláp (Kónya and Bodnár 2008) and Ság Hills, once rising as majestically as all the others, have been almost completely removed.

## 8.2 Geological Setting

The Tapolca Basin volcanoes are part of the Bakony-Balaton Highland Volcanic Field (Németh and Martin 1999; Martin and Németh 2004; Németh 2012) of ca 3 $km^3$ material ejected from 150 to 200 centres, including the residual hills of Badacsony (438 m), Szent György-hegy (St. George Hill, 414 m), Gulács (393 m), Csobánc (376 m) and Haláp (330 m) (Figs. 8.1 and 8.2). The hills seem to be arranged randomly at 3–5 km distances, but Martin and Németh (2004) claim that they are aligned along NE–SW faultlines.

The neighbouring Little Hungarian Plain Volcanic Field (Németh 2012) includes Somló Hill (432 m) in the Marcal Basin and the buttes of Kemenesalja: Ság (278 m) and Kis-Somlyó (219 m) (Fig. 8.1).

As background information to volcanism in the Carpathian (Pannonian) Basin, it is necessary to point out that the lithosphere below it is thinner than average, only 80 km, while the thickness of the crust is 25–27 km. The explanation lies in the pull exerted by the lithospheric plate along the Carpathian arc on the basement of the Pannonian Basin 18–14 Ma ago and this pull resulted in the fracturing and thinning of the plate (Harangi 2011). The basement, however, is of variable thickness and the hills are located on the boundaries between plate fragments of different thickness (Harangi 2011).

## 8.3 Evolution of Volcanic Landforms

The model elaborated for the evolution of basaltic volcanism in Transdanubia (Martin and Németh 2004; Harangi 2011; Németh 2012) assumes magma flow from below the thicker

**Fig. 8.2** Aerial view of the Tapolca Basin from the direction of Lake Balaton looking north, showing Badacsony (in the *foreground*), Szent György-hegy (to the *left*), Gulács and Csobánc (to the *right*) and Somló Hill (on the *horizon*) (Photo by János Scheffer)

(ca 100 km) plate of western Transdanubia towards the base of the thinner (ca 60 km) plate of the Great Hungarian Plain. This meant an almost vertical upward movement which generated a loss of lithospheric pressure and favourable conditions to magma accumulation, reinforced by water content (phreatomagmatic influence) and lowered melting point of rocks. The large amounts of magma created this way followed the fractures on the boundaries between the thicker and thinner plates and flowed to the surface. As a consequence, a multitude of minor centres began to erupt, most intensively between 5 and 3 Ma ago. The buttes of Ság, Kis-Somlyó, Somló, Badacsony, Szent György-hegy, Gulács, Csobánc and Haláp all derive from this time interval. They are similar not only in age, but also in geological setting, morphology and denudation history.

The study of the interior structure of Ság Hill, a ruined volcano, exposed by quarrying revealed useful information on the origin of all the other volcanoes too. The edifice of Ság Hill volcano evolved to the effect of the following influences (Martin and Németh 2004; Harangi 2011—Fig. 8.3). After the regression of Lake Pannon (ca 5 Ma ago) the magma reached the surface along extensional cracks. When the lava solidified, the terrain was waterlogged with lakes, swamps and rivers on several hundreds of metres of sand and gravel deposits, saturated with water. As a consequence of magma–water interactions explosive eruptions built tuff-rings around the vents. In the tuff-rings wet sand was mixed with basalt pyroclasts of heavily fractured, splintery and vitreous fabric. Subsequently, when most of the heated water evaporated and the ascending magma contacted with ever lower amounts of water, the explosions tended to reduce in intensity and the Strombolian eruptions built cinder cones within tuff-rings. With the declining influence of water and further degassing of magma, Hawaiian activity with lava flows followed. The lava flows of inflated pahoehoe type filled up the craters of the Strombolian cones. In this phase of volcanism the low-viscosity pahoehoe lava was impounded by the inner walls of the tuff-ring in the form of lava lakes (Figs. 8.3 and 8.4). It was the effusive compact basalt from this stage that was recently quarried. The cones built of the products of the Strombolian eruptions were left partially intact.

In the final phase of volcanism lava flows were overlain by fallen pyroclasts, partly including remains of spatter cones formed in the vicinity of lava fountains (Fig. 8.4). Sporadic explosions also produced cinder cones again, the most spectacular examples being found on Badacsony and Somló Hills.

The compact alkaline basalt lavas, Strombolian cinder layers and tuff beds in the rings do not show evidence of weathering, indicating that the different types of activity followed each other without interruption. The monogenetic character of the volcanoes is often emphasized (Martin and Németh 2004; Németh 2012). As an exception Kis-Somlyó can be cited, where the development of the tuff ring was followed by a longer pause of several hundred or even thousand years. As a consequence, the crater became inundated and lacustrine sands deposited in it (Harangi 2011). When volcanic activity resumed, pillow lava accumulated in the crater lake (Harangi 2011).

## 8.4 Denudation History

As soon as volcanic activity ceased, exogenic forces became active on the volcanic edifices. Compared to the fill of vents and craters of compact basalt lava, tuff rings and the neighbouring unconsolidated sandy series of Lake Pannon were less resistant to erosion and, thus, were affected by intensive denudation. The removal of the Pannonian sediments was further enhanced by regional uplift, which was the most intense in the Pleistocene (Futó 2003). Streams transported the mobilized material towards the Kaposvár–Kalocsa subsidence zone in Southeast-Transdanubia (Futó 2003). In the Trandanubian Mountains, including the Bakony immediately

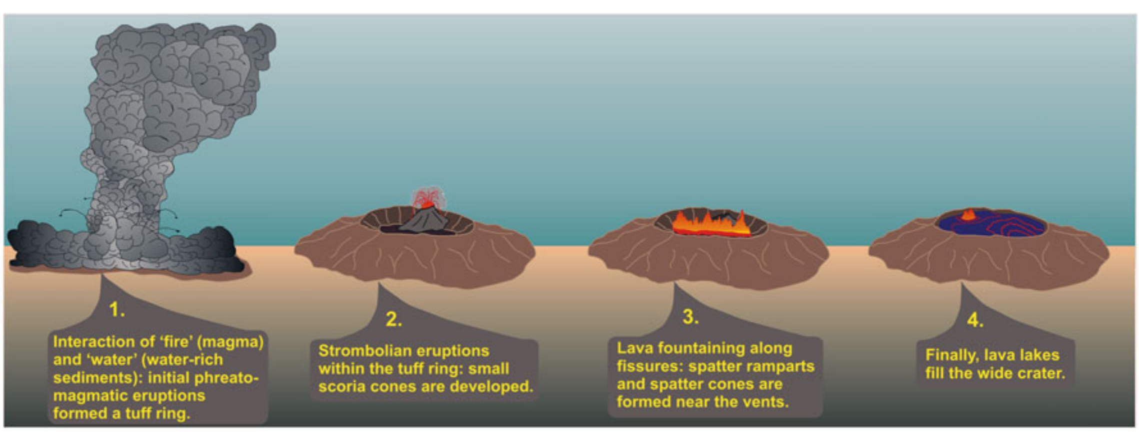


**Fig. 8.3** Evolution of residual hills in Transdanubia on the example of Ság Hill (after Harangi 2011)

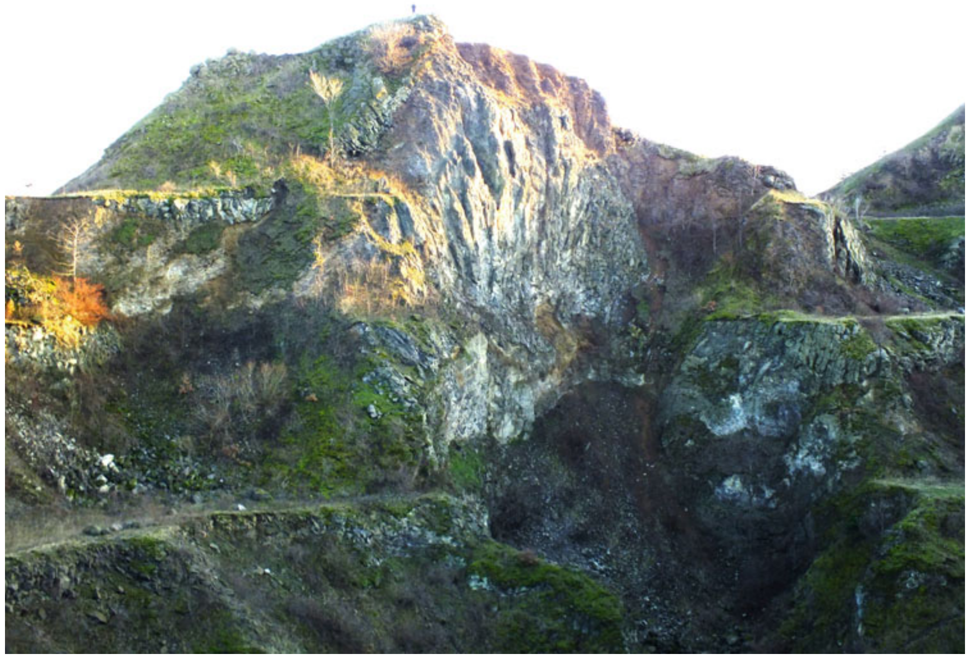

**Fig. 8.4** One of the vents of Ság Hill with radially fissured lava from the last stage of volcanism, reddish scoria and spatter cone (photo by Péter Gadányi)

to the NE of the Tapolca Basin, the terrain between craters was lowered, but the former craters filled with basalt lava preserved their topographic positions and emerged as basalt-capped mesas or buttes. During the Pleistocene glacials the residual hills and their environs were eroded through the action of periglacial processes. In the relatively dry glacials wind erosion was prevalent (Erdélyi 1954). Deflation further deepened the areas between basaltic crater fills. With the destruction of tuff-rings, the margins of the basalt crater fills, compact but fractured by solidification joints, were exhumed and emerged as vertical cliffs, primarily moulded by frost shattering. Therefore, the slopes of residual hills retreated at high rates and cryoplanation cliffs with platforms resulted. The rate of retreat was substantially increased by the deflation of unconsolidated Pannonian sand underlying the compact basalt. Landslides also contributed to slope evolution. At the feet of retreating basalt cliffs periglacial rockflows and blockfields accumulated.

It is estimated that the overall terrain lowering in the environs of volcanic hills amounted to 50–350 m (Martin and Németh 2004), while the butte tops and crater fills denuded at a much slower pace. This is how the buttes became witnesses of topographic conditions during the Pliocene volcanic activity (Fig. 8.5).

## 8.5 Landform Diversity

In lateral view, the residual hills of Transdanubia present almost the same shape from all sides. The broad and gentle base slopes are built of sandy deposits of Lake Pannon. Subvertical walls of basalt rise above them, while the summits are slightly undulating surfaces with remnants of cinder or spatter cones. In planform the residual hills are circular (Fig. 8.1), because the lava once accumulated in the circular craters of tuff-rings. The circular groundplan has not been substantially modified by retreat of cryoplanation cliffs either since the retreat was directed from the crater margin towards the centre of the hill. It is assumed that the present diameter of the basaltic crater fills is somewhat smaller than the original diameter of the enclosing crater. In the case of smaller craters parallel retreat has almost reached the centre of the hill, consumed its flat top, and transformed the butte into a pointed pinnacle with concave slopes as exemplified by Gulács Hill (Figs. 8.1 and 8.2). The Gulács basalt is rather friable and, thus, its cliffs experience a more rapid retreat (Erdélyi 1954). The more subdued relief of this hill is also explained by lithological properties.

Ság Hill (278 m) is a landmark in the Kemenesalja section of the Little Hungarian Plain. The remnants of its tuff ring, created in the initial, phreatomagmatic, stage, were exposed by quarrying. The alternation of layers of fallen pyroclasts and cross-bedded surges are clearly visible (Fig. 8.4). Several Stromboli-type eruption centres from the next stage of volcanic activity can also be studied on quarry walls, cutting across the vents and encircling welded cinder. In drier environment Hawaiian eruptions followed and the spatter cones of lava fountains and minor patches of inflated pahoehoe lava appear on the wall of the former vent. It is interesting to observe that in a contact zone with the once wet tuff ring the lower lava flow shows gas bubbles generated from water vapour.

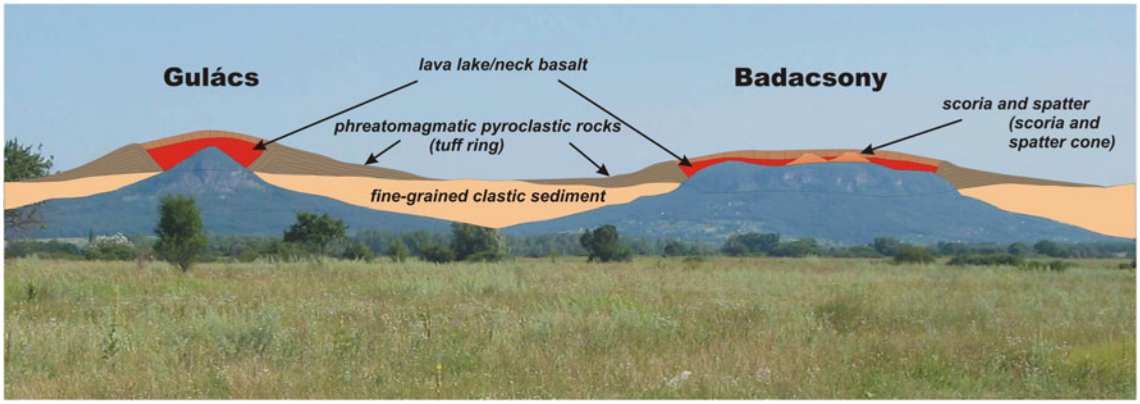


**Fig. 8.5** Reconstruction of Gulács and Badacsony Hills, the situation 3.5–4 Ma ago (after Harangi 2011)

The basalt lava of Somló Hill (432 m), on the boundary between the Bakony Mountains and the Marcal Basin, is thick-bedded and easily disintegrates into fragments 1–2 cm across. Luckily from the viewpoint of nature conservation, the basalt of Somló was not suitable for industrial extraction and the hill retained its regular butte shape. An individual trait of Somló Hill is the pointed cinder cone on its flat top.

The largest of all residual hills of Hungary, Badacsony (438 m) rises on the southern margin of the Tapolca Basin somewhat protruding into Lake Balaton as a peninsula (Fig. 8.1). On its northern and northeastern flanks two scars of quarries with steep cliffs of compact thick-bedded basalt lava are visible. Extensive blockfields from repeated topples of rock fragments detached from the cliff occur around the hill. One of the most conspicuous cliffs of Badacsony are the thick, massive columns called "rows of bastions" on the southwestern flank. The rather irregular columns of thick-bedded basalt on the northeastern side, the marked Kőkapu ("stone gate"), are remnants of a cliff produced by parallel retreat of basaltic crater fill. On the flat top of Badacsony, ruins of reddish-brown cinder cones have been preserved. The trench-like depression of NE–SW strike on the top can probably be explained by subsidence in the underlying unconsolidated Pannonian deposits (Erdélyi 1954). The evidence for it is provided by the basalt of the Badacsonytomaj quarry, heavily fractured in irregular directions below the surface depression.

## 8.6 Weathering Features on Columnar Basalt

The most beautiful residual hill of the Tapolca Basin is Szent György-hegy (414 m). On the northeastern flank a renowned basalt colonnade, the Nagyorgona ("great organ"—Fig. 8.6) comprises basalt columns 20–25 m high and 1–1.5 m across. Their tops have been fashioned to a rounded shape through mechanical weathering and are called locally "stone sacks" (Erdélyi 1954). During the solidification of basalt lava in the crater within the former tuff ring contraction led to columnar basalt formation (described in detail in international literature by Budkewitsch and Robin 1994). The process can even take place in inflated pahoehoe lava of several metres of thickness and subvolcanic rock masses. During cooling from top to bottom, transverse fissures perpendicular to vertical contraction cracks also developed at 5–10 cm average intervals, but locally even at 2–3 cm from each other. More rapid cooling may have resulted in a denser joint network. Due to the particular rock structure the basalt of the Nagyorgona disintegrated into cylindrical columns, divided into discs similar to oversize coins (Fig. 8.7). Transverse jointing makes the columns liable to break, topple and fall into pieces. During disintegration fragments 1–5 cm across are detached from the walls as it is clearly visible on the discs of toppled columns. If the network of transverse joints is less dense, sizeable blocks are weathered out and columns with angular cross-sections or thick discs are produced, as in the blockfield upslope of Nagyorgona. The fragments of the blockfield are of much greater dimensions than in the blockfield of Nagyorgona (Fig. 8.8).

The formation of rounded basalt columns mostly takes place under soil cover, most efficiently 10–20 cm underground. The soil cover ensures continuous water discharge, enhancing weathering, mainly freeze-thaw action in early spring. Water infiltrates into cracks and the sides of the columns begin to retreat by mechanical weathering. Frost-shattered debris accumulates in the soil and is transported to greater depths by rainwater and snow meltwater. Runoff is the most efficient geomorphic agent between columns, where gullies function as water conduits after showers. The smaller shattered fragments are transported by runoff, while most of the larger blocks remain in place or are dislodged by episodically recurring topples to increasingly lower positions. The columns exhumed from below the soil which

**Fig. 8.6** Columnar basalt of Nagyorgona with rounded tops on Szent György-hegy (photo by Péter Gadányi)

**Fig. 8.7** Fragment of column separated by transverse joints, "giant coin" of the Nagyorgona colonnade. The columns are weathered into cylindrical shape (photo by Péter Gadányi)

stand separately cannot be wetted so deeply as under soil cover since the rainwater runs down rapidly from their surfaces and sections of their walls are not wetted at all. Neither is frost weathering active on them. The exhumed columns are gradually separated and shifted away from one another and acquire a rounded shape and oval cross-section, because they become elongated in the direction of the exhumation process. Dislodged fragments get stuck between columns locally. Due to the daily and seasonal temperature regime as well as to displacements of other origin, the blocks arrive to lower and lower topographical positions and exert a wedging influence on the columns.

**Fig. 8.8** Blockfield upslope of Nagyorgona (photo by Péter Gadányi)

With diminishing soil cover, more and more columns are exhumed and become thinner until, in lack of support, they eventually topple. In the case of some columns this thinning only affects the uppermost third or quarter, which collapses often towards the colonnade, hitting against the neighbouring column. Locally, the solidification cracks branch out like tree roots and some columns stand on "several legs". The cracks between such "legs" are occasionally broadened through frost action to form caves.

Rock fragments of the screes around columns are further disintegrated and mixed, primarily due to mass movements generated in the loose Pannonian sands under the load of accumulating basalt debris. Susceptibility to sliding is enhanced by the emergence of slip-planes along clay interbeddings in the sands. When reaching a threshold depth, certain sections of screes may become so unstable that gravity induces episodic movements.

## 8.7 Another Geomorphosite: Tapolca Lake Cave

The Tapolca Basin also offers additional sights to an interested tourist. Its geological/geomorphological heritage includes the Lake Cave of Tapolca with 3.3 km total length of passages. Cold karst water arrives into the 14-Ma-old Sarmatian limestone and marl, in which the cave formed, from an Upper Triassic dolomite aquifer. It is mixed with thermal water of almost 40 °C temperature, attesting to postvolcanic activity. Cave formation was due to the mixture corrosion of the resulting 19 °C warm water.

The Tapolca Lake Cave was discovered in 1902. It was the second cave to be protected legally in Hungary and the first equipped with electric lighting as a show cave. Water levels in the lake, however, dramatically sank during the operation of bauxite mines in the Bakony Mountains and most sections became dry. The lake could only be restored after the mines closed down in the 1990s and the cave was reopened to visitors in 2012. The outflow of the cave is the Mill Lake in the centre of the town Tapolca.

## 8.8 Conclusions

The Tapolca Basin is counted among the most picturesque landscapes and most valuable group of geosites in Hungary with relatively high relief, columnar basalt formations and superb vistas on Lake Balaton and marked buttes on the horizon.

The last of the basalt quarries, opened after 1903, was closed down in 1964 and the landscape scars were at least partly eliminated by the end of the 20th century. With the purpose of conservation of natural monuments around Lake Balaton, the Balaton-felvidék National Park of 56,997 hectares area was established in 1997.

Three nature trails present the sights of the Basin to visitors. The 4-km-long geological and botanical trail of Badacsony Hill informs about the origin of the "basalt organs", historical and architectural monuments (castles, manors) as well as typical landscape patterns. The most beautiful basalt columns and the best panoramic view can be enjoyed on the Szent György-hegy Trail. Next to the lake shore, around the much lower butte of Szigliget, the Kamon Rock Nature Trail presents the history of volcanism, the origin of Lake Balaton, vernacular architecture, viticulture, reed economy and animal life.

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# Thermal Spring Cones of the Tihany Peninsula

9

Márton Veress, Péter Gadányi, and Gábor Tóth


**Abstract**

Tihany is a spectacular volcanic peninsula of Lake Balaton, where more than a hundred cones deposited by thermal waters rise. The calcareous and siliceous building materials overlie basaltic tuff layers of maar-type volcanism. Their formation is related to the existence of an underlying magma chamber, which heated up the surrounding karst water. Hot karst waters spouted to the surface, where the dissolved silica and carbonates precipitated and deposited. During the stage of mofetta development hot spring water accumulated in ponds where organic limestone deposited. The further heating of karst water modified the composition of spring water and, in parallel, the composition of the cones. The mofetta stage was succeeded by the fumarola stage, when siliceous minerals precipitated from spring water and added further substances to the spring cone edifices. These minerals supplanted the calcites and filled cavities and cracks.



**Keywords**

Hot springs • Spring cones • Post-volcanic activity • Karst water • Tihany peninsula


## 9.1 Introduction

The Tihany peninsula of 12 km$^2$ area protrudes into Lake Balaton from the northern shore. The basins of the permanent Inner Lake, the intermittent Outer Lake and more than hundred discrete mounds make the landscape of the peninsula variable. These high grounds are spring deposit cones of calcareous and siliceous sinter which formed upon Pliocene (7.8–6.6 Ma old) basaltic tuff layers. The lake basins are volcanic craters of maar type (Németh 1995). The volcanic activity was phreatomagmatic (Németh et al. 2001; Martin and Németh 2004) with three eruption centres, in the northern, central and southern part of the peninsula. The bedrock under the basaltic tuff is slate of Silurian age (Lovas Slate Formation), Permian red sandstone (Balatonfelvidék Sandstone Formation), Triassic limestones and dolomites, and Late Miocene sands, clays, conglomerates and limestones.

The cones accumulated due to precipitation at and around hot springs in the Pleistocene. The uprush of hot spring waters made ideal conditions for building of conical edifices: water flow is directed upwards and, as a consequence, the precipitates from outflows are superposed one above the other. The amount of dissolved materials is higher in hot and pressurized waters and precipitates under reduced hydrostatic and lithostatic pressure as the cooling water reaches the surface. According to water temperature, the following stages of post-volcanic activity could be distinguished: mud volcano (hotter than 300 °C), solfatara (200–300 °C), fumarole (100–200 °C), geyser (over 100 °C), mofetta (50–100 °C) and thermal water stage (below 50 °C).


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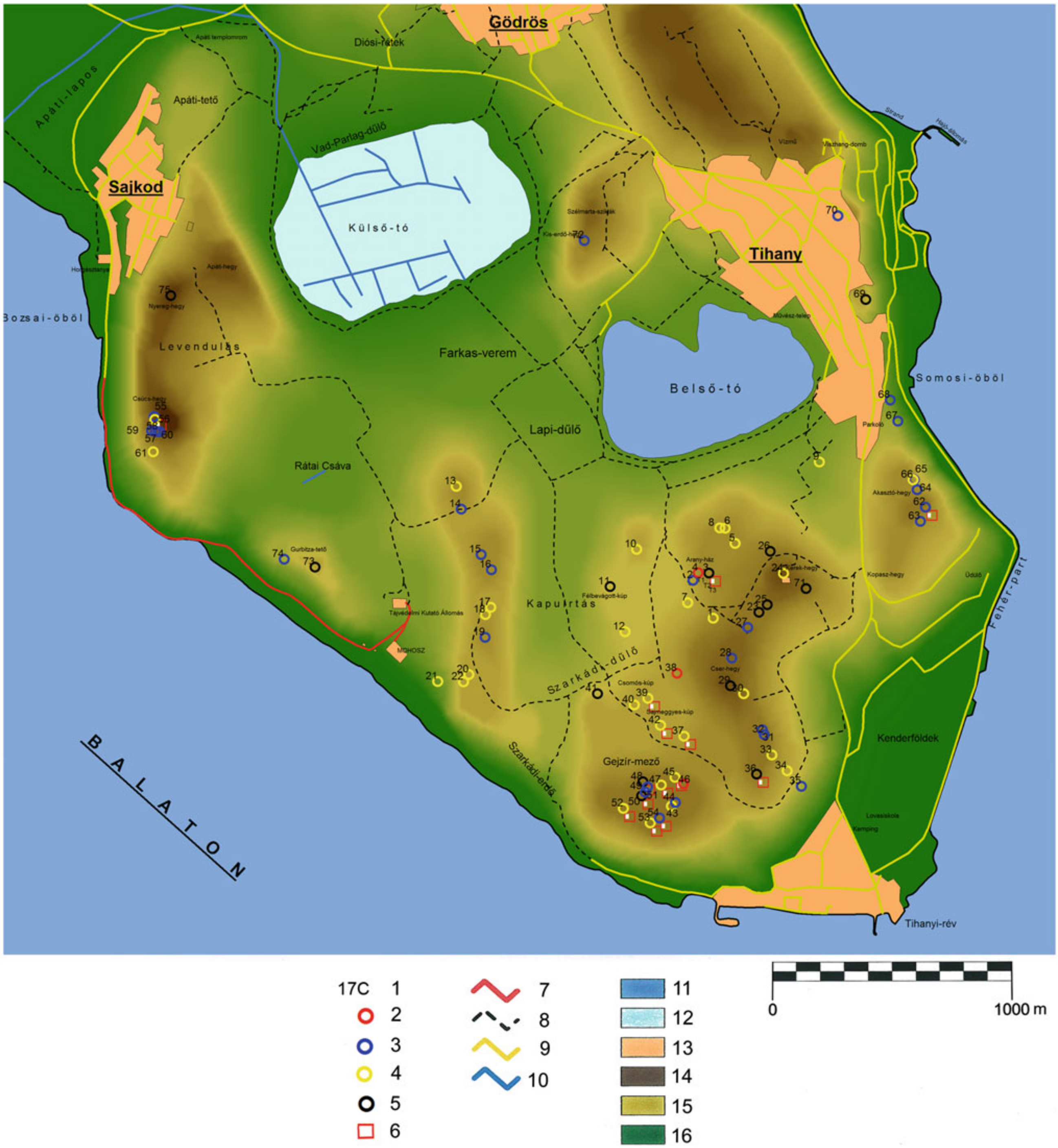


**Fig. 9.1** Map of the thermal spring cones of the Tihany-peninsula (Veress 2003). *1* Thermal spring cone with identification mark; *2* cone with uncertain diameter; *3* cone less than 25 m across; *4* cone 25–75 m across; *5* cone more than 75 m across; *6* quarried cone; *7* road closed to vehicle traffic; *8* dirt road; *9* gravel road; *10* watercourse, stream channel; *11* lake; *12* swamp with open water surface and reed; *13* built-up area; *14* elevation 230 m above sea level; *15* elevation 160 m; *16* elevation 100 m

## 9.2 Location and Morphology

The cones are located in clusters, significantly different in terms of density and size. In the peninsula there are 10 cones per $km^2$ on average, but there are places such as the "Geyser Field" where 24 cones occur in an area of 0.25 $km^2$. Their size is significant too. The diameter of 32 % of the mapped cones (75 pieces) at their longer axis is larger than 75 m (Fig. 9.1). Clusters of cones form rows along north to south direction in general. The exceptions are the irregularly distributed cones of the "Geyser Field", where the most famous thermal spring cone, Aranyház ("golden house"), is found (Fig. 9.2). The clusters are elongated in ground plan and their N–S alignment coincides with the orientation of cracks and fissures along which hot waters and steam once spouted to the surface. The existence of these fault lines is proven by aerial photographs (Németh 1995).

The major part of the cones is located along the eastern and western margins of the peninsula and in the interior, south of Inner Lake. The building material of the thermal spring cones is resistant against erosive forces. This could be the reason of the lack of valleys in the peninsula, because the marginal cones prevented regressive valley evolution from the shores of Lake Balaton, although relative relief above the lake level exceeds 50 m.

The cones can be classified as primary and secondary. Primary cones are independent from each other, and some of them have altered significantly during further evolution by additional growth through further precipitation. Secondary cones are connected to each other and form ridges several hundred metres long. The tops of the cones are dissected by smaller mounds called adventive cones. The adventive vents and cones (Figs. 9.3 and 9.4) were built during later stages of hot water eruptions, whose sites gradually shifted with time. On the flat-topped cones there are pools 1–2 m across and 10–20 cm deep (Fig. 9.4). The margins of the pools are rimmed by rings of silica and calcareous sinter and their residual forms. The rings formed by precipitation from water overflowing from the pools. On the bare flanks of the travertine cones, karren (pits, thimble karren) developed due to the solution effect of rainwater.

Exposed naturally or by quarrying, spherical or shaft-like cavities are also characteristic features of spring cones. The spherical cavities usually widen downwards. The surfaces of the more spacious spherical cavities (around 1 m in diameter) are compound, their walls are dissected by niches of 10–20 cm in diameter on average and miniature hollows of great density, several centimetres or even millimetres across. The cavities evolved by post-genetic solution at lowered pH or through enclosure by precipitation.

The shafts (Fig. 9.3) are either closed or open at their upper end, and often form branches of larger cavities, pointing towards the surface. Most of them are vertical, but there are also oblique and almost horizontal cracks. According to the shape of their cross-sections, they are complete, cut in half or destructed. The surfaces of the shaft walls are smooth or dissected by small open cavities of 2–5 cm across. The origins of shafts are variable. They either develop through the erosive effects of waters which erupted to the surface (open shafts of stope-up solutional origin) or through enclosure by the accumulation of mineral precipitations around them. The above-mentioned features are indicative of the environment of shaft formation. Thus, those of vertical position prove an upward waterflow in the late phase of spring cone development.

**Fig. 9.2** Aerial photograph of Aranyház with Inner Lake in the background

**Fig. 9.3** Opened shaft of Aranyház

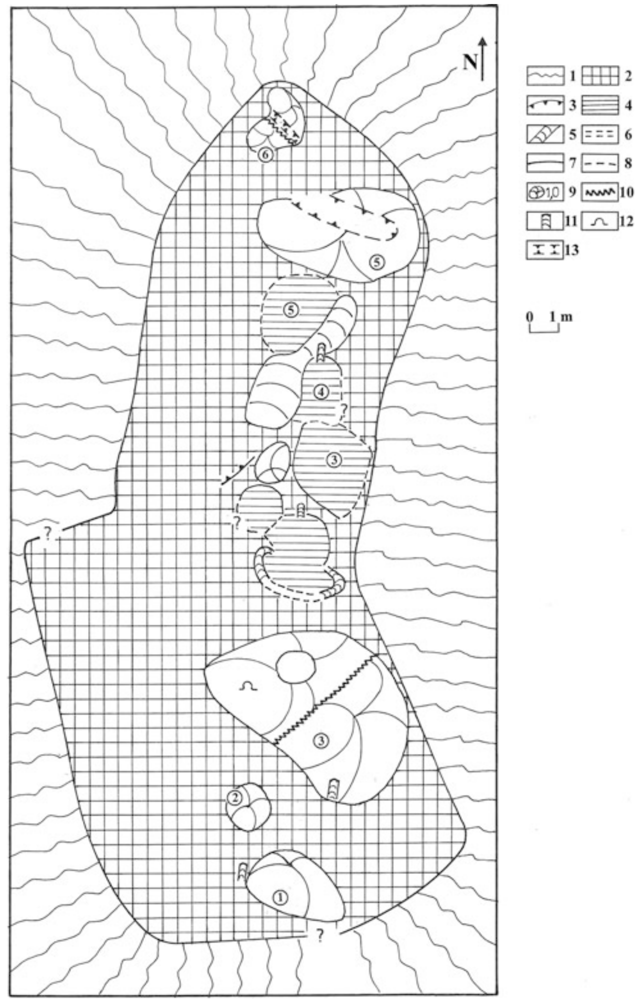


**Fig. 9.4** Geomorphological map of the upper Aranyház cone (Veress 2003). *1* Cone flank; *2* gentle slope of cone top; *3* stairs; *4* floor of spring pool; *5* sinter ring; *6* damaged sinter ring; *7* margin of spring-pool floor; *8* assumed margin of spring-pool floor with adventive cone (elevation in metres); *10* coalescence site; *11* shaft; *12* cavity of solutional origin; *13* half shaft

The depressions on the top of the cones indicate former pools and geysers or fumarole-type operation of adventive cones.

On the flanks of spring cones there are numerous traces of mass movements, which could have happened on the bottom of the pools around the cones. Subaquatic mass movements seem to be relatively old, and only identified in the interior structures of building rocks. The younger mass movements, developed on already existing cones, are roof collapses of cavities, episodic rockfalls from steeper and unstable wall segments or rock slides due to tilting affecting significant portions of the cones. Irregular scars of old quarries on the flanks of a number of cones add anthropogenic features to their morphology.

## 9.3 Substance and Conditions of Cone Formation

In the opinion of Lóczy (1916), thermal spring cones of Tihany evolved by geyser activity. According to the studies by Hoffer (1943), the structures of the lower parts of the cones are layered and laminated and rich in $CaCO_3$, while the top portions are porous, unstratified, non-laminated, with relatively more quartz and hydroquartzite than $CaCO_3$. Varrók (1957) claims that the lamination of the cones is the result of alternation of calcite and silica (with siliceous algae) and their diagenesis started with the precipitation of polluted calcite. Subsequently, calcite was dissolved by erupting hot waters and replaced by spherocalcites. Then siliceous substances began to deposit in two manners: they either supplanted the calcites or filled the pore spaces of cones.

Recently Kovács-Pálffy et al. (2007) analysed the mineralogical composition of three cones, taking 48 samples (Fig. 9.5). The carbonate minerals of the samples were micrite, microspar, sparite, magnesium-calcite, dolomicrite,

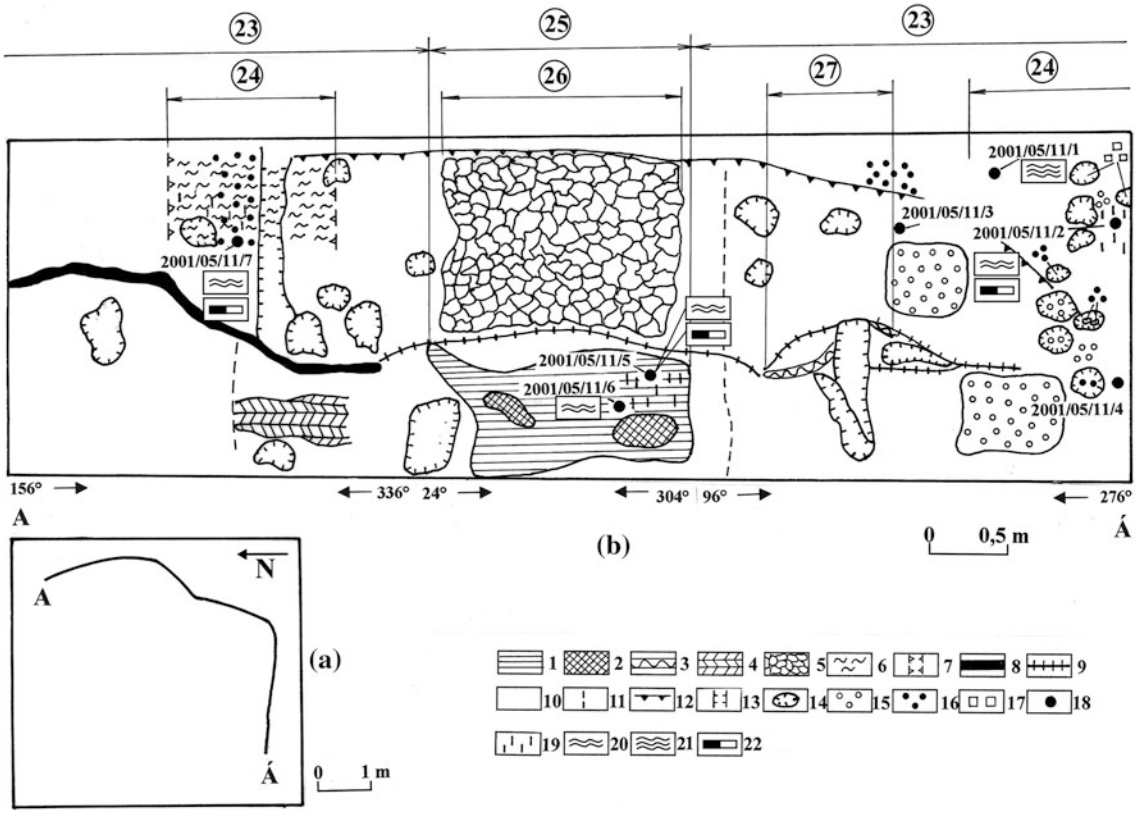


**Fig. 9.5** Cross section of opened cavities of the split cone. **a** Overview map; **b** detailed map (Veress 2003). *1* Laminated beds; *3* partly laminated beds; *4* layered beds; *5* beds of distinct block; *6* massive shaft or pool fills; *7* fill boundary; *8* dark, thick, homogeneous intercalation; *9* discordance; *10* massive beds; *11* fracture; *12* edge of opened cavity; *13* shaft, *14* solutional cavity of medium size; *15* small cavity enclosures; *16* small solutional cavity; *17* botryoids; *18* sampling sites; *19* algal mats; *20* rocks with algae; *21* rocks dominated by algae; *22* recrystallized parts; *23* cone flank; *24* former pool or shaft; *25* former pool; *26* possible sign of fumarole activity; *27* cone coalescence

dolosparite, magnesite and aragonite. Clay minerals included montmorillonite, illite, illite-montmorillonite and kaolinite. The siliceous minerals which supplanted carbonate rocks or secondary minerals were quartz, chalcedony, tridimite and opal. These minerals, with the exception of the tridimite, were secondary minerals precipitated in cavities, pores and cracks. Secondary ferric minerals (from siderite) were hematite, goethite and limonite gel. In addition sulfides, sulfates, gypsum, extraclasts derived from falling dust (plagioclase feldspar), bioclasts (skeletal components of gastropods) and residuum of plant-fossils were also identified.

In the geological description of the evolution of thermal spring cones we follow the nomenclature suggested and used by Folk (1959), Todd (1966) and Ford and Pendley (1996). The building materials of cones are primary and secondary minerals, which generated in two main evolution stages. In the first stage the calcareous sinter (mudstone) accumulated, while in the second silica sinter deposited simultaneously with hydrothermal mineralization.

In the first stage of evolution calcareous sinter (or mudstone) formed through the recrystallization of calcareous and dolomitic mud of algal origin. This mudstone consists of carbonate grainlets and micrites. A laminite of the alternating of micrite and microsparite laminae is typical, but where no laminae exist, the mudstone is massive in appearance. The mudstone derived from blue-green algae (Cyanophytae). Gas bubble enclosures generated by their autolysis are clearly recognizable in thin sections. The termination of the first stage is marked by desiccation cracks on the surface of the deposited mudstone.

In the second stage silica minerals supplanted calcareous mineral grains in the pores and on the laminae surfaces. The first generation of silica minerals consists of chalcedony, cristobalite, then substituted by the second generation of spherical quartz. After silicification fissure fillings formed in the previously deposited material. Open cracks evolved by desiccation or by atectonic movements of the cones material. Three generations of cracks are distinguished. On the walls of the first generation cracks, evolved after silicification, coatings of chalcedony, quartz, gypsum and aragonite occur. The fillings of the second crack generation consist of chalcedony, dolomite and holocrystalline calcite. This swarm crosses the cracks filled by quartz. The precipitation of colloidal limonite and opal, belonging to the third generation of fills, continues to our days.

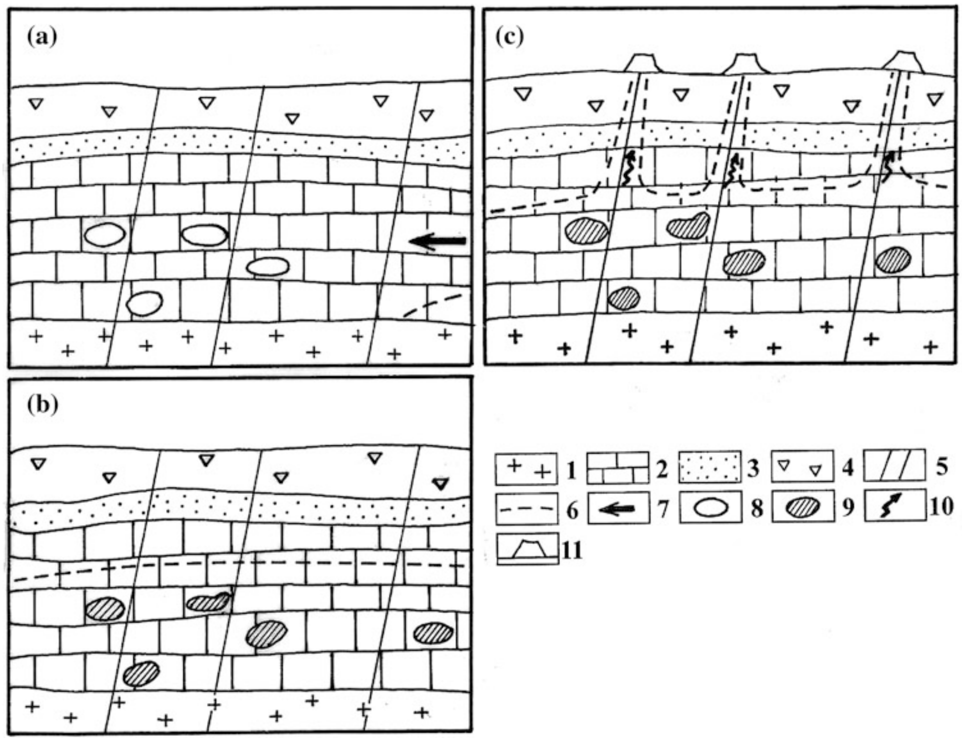


**Fig. 9.6** Formation of thermal spring cones in Tihany Peninsula (Veress 2003). For stages see the text. *1* Permian red sandstone; *2* triassic carbonate rocks; *3* Pannonian sand; *4* volcanites; *5* faults; *6* karst water table; *7* karst water flow; *8* karstic cavity; *9* karst water; *10* hot karst water rising towards the surface; *11* spring cone

The evolution of carbonate minerals started at 64 °C temperature, but carbonates of biogenic origin can also occur at temperatures 10–20 °C lower (Vasconcelos and Mckenzie 1997). According to Pentecost (1990), the optimal reproduction temperature for the carbonate forming blue-green algae is between 45 and 60 °C, but they are also capable of living in waters of 80 °C. The first stage of spring cone evolution took place at such temperatures. The amorphous silica gel could dissolve at higher (95–110 °C) temperatures and 9.5 pH (Moray et al. 1962). These conditions were usual during the second evolution stage when the temperature of spouting waters increased due to the heating effect of the underlying active magma chamber. The precipitation of magnesium silica minerals occurred at temperatures in the range of 30–100 °C. As the waters cooled, the siliceous components transformed to clay minerals.

## 9.4 Thermal Spring Activity and Cone Formation

According to the new investigations by Veress (2003) and Kovács-Pálffy et al. (2007), the formation history of the thermal spring cones can be reconstructed in the following way (Fig. 9.6). During the phreatomagmatic volcanic activity the karst water reserves stored in limestone and dolomite were consumed (Fig. 9.6a), but later the emptied reservoirs were gradually recharged with karst water originating from the Bakony Mountains (Fig. 9.6b). Heated by the still hot magma, the karst water rushed towards the surface (Fig. 9.6c). The spouting hot waters transported small rock particles, grainlets and dissolved carbonate, which partly reached the surface and partly entered and deposited in the joints of the basalt.

As a consequence of the continual heating effect of the magma chamber, the temperature of the uprushing waters was increasing. At first warm springs appeared on the surface, and later their temperatures exceeded 50 °C, so the conditions of cone formation became similar to a mofetta-type post-volcanic activity. At this evolution stage hot-water ponds emerged around the springs, rimmed by calcareous materials precipitated from the overflow waters of pools. On the bottom of the pools skeletal components of algae accumulated (Fig. 9.7a, b). Further heating occurred due to the effect of the proximity of magma and hot waters penetrated into the building materials of the cones. Siliceous minerals from Pannonian sand were dissolved and deposited in the cones (Fig. 9.7c). Higher water temperature is indicated by chalcedony (formed at 100–150 °C) and $\beta_1$-tridimite (at 117–163 °C), detected in a sample from the Aranyház thermal spring cone. Hot water spring activity was accompanied by steam eruptions, whose previous existence is attested by adventive cones (Figs. 9.3, 9.4 and 9.7d), which evolved by fumarole activity. In the course of time, with reducing heating by the magma, the temperature of waters penetrating into the cones dropped and, finally, the rise of hot waters ceased (Fig. 9.7e).

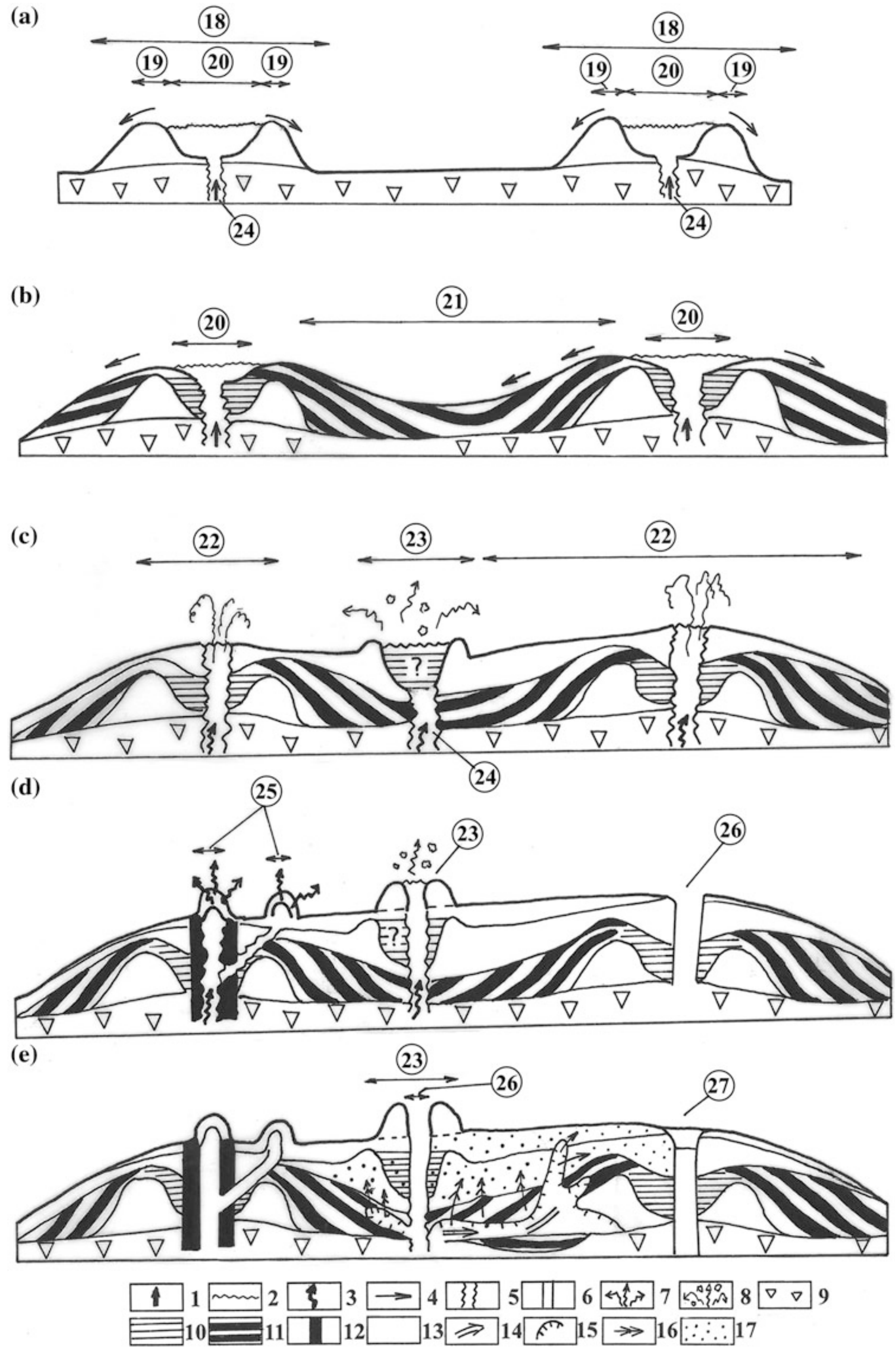


**Fig. 9.7** Model of formation of secondary thermal spring cones (Veress 2003). **a** Sinter rings formed during mofetta stage; **b** coalesced spring cones; **c** secondary cone due to geyser and fumarole activity; **d** adventive cones formed on secondary cones; **e** dissolution at lower cone, precipitation on upper cone. *1* Erupting waters with temperatures below 100 °C; *2* level of hot water; *3* erupting waters above 100 °C; *4* surface watercourse; *5* active shaft; *6* inactive shaft; *7* geyser activity; *8* fumarole activity; *9* basaltic tuff; *10* laminar deposits in pools; *11* laminar deposits outside the pools; *12* shaft fills; *13* non-laminar calcite and silica deposits; *14* direction of waters with hydrocarbonate and alkaline solution; *15* cavity formed by hydrocarbonate and alkaline solution; *16* circulation of oversaturated waters; *17* calcite modifications (e.g. orthopatite II), opal replacing calcites and precipitated in small cavities of different origin; *18* thermal spring cone; *19* sinter ring; *20* pool with pond; *21* common parts of coalesced cones; *22* geyser cone; *23* cone of fumarole activity; *24* opened shaft formed by stope-up solution; *25* cone from precipitations around spring; *26* open shaft; *27* covered stope-up shaft

Parallel with hot water eruptions, the pools were partly or entirely filled with precipitated and deposited materials, the cones developed further, and several adjacent cones coalesced (Figs. 9.5 and 9.7e).

## 9.5 Conclusions

The thermal spring cones of Tihany Peninsula result from post-volcanic activity. They are striking and spectacular formations, not only for their high density and impressive dimensions, but also for their intriguing morphology. The uniqueness of these cones is lies in their evolutionary history in which post-volcanic processes did not follow the usual sequence governed by decreasing temperatures, but in a reverse order. Karst waters played a crucial part in thermal spring cone formation. The dissolved calcareous materials precipitated and deposited in and on the cone structures, making their composition and morphology more variable.

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# Kál Basin: Pseudokarren on Sandstone

# 10

Márton Veress and Gábor Tóth

**Abstract**

The Kál Basin, located in the central part of Transdanubia, on the northern, mountainous shore of Lake Balaton (Balaton Highland), is one of the most spectacular landscapes of Hungary, a focal area of geological and geomorphological research for more than a century. Its sandstone was formed from the sand beaches of Lake Pannon during the Pliocene and the blockfields (felsenmeers) of the basin are also intriguing geomorphological sites. In addition to the geomorphic diversity, they are also valuable because they survived earth history almost unmodified. After the regression of the sea, on the surfaces of stone blocks a wealth of pseudokarren features began to emerge, adding further diversity to the landscape. The most common features on the horizontal block surfaces are kamenitzas, rinnen, grikes and pits, while pockets and solution notches occur on the walls of the blocks.

**Keywords**

Blockfield • Pseudokarren • Sandstone formation • Weathering • Lake Pannon • Balaton uplands

## 10.1 Introduction

The Kál Basin—together with the Pécsely Basin—is the largest basin of the Balaton Uplands or Highland. Its geological structure is varied: in addition to Permian red sandstone, conglomerated sandstone, limestone and dolomite, several basalt cones (Hegyestű, Kis-Hegyestű, Lapos-Hegyestű, Kereki Hill) rise above the basin floor. The cones result from basalt volcanism subsequent to the Miocene marine transgression (ca 4–5 Ma BP). Springs with waters of high gas content emerging to the surface at several localities within the basin remind us of volcanic activity. The blockfields near Szentbékkálla, Salföld and Kővágóörs, where millstones were cut and carved for hundreds of years, are geological values famous throughout Europe. The most beautiful is the blockfield of Szentbékkálla (Fig. 10.1), which has survived almost intact until the present day and exhibit numerous types of pseudokarren features.

## 10.2 Geological and Geomorphological Settings

The Kállai formations of the Balaton Uplands occur in several varieties, described by Budai and Csillag (1999) and Budai et al. (2002). They include the so-called “quartz sand” of the Kál Basin, deposited along the margins of the Keszthely Mountains, where the sand grains of the sandstone are partly cemented by amorphous silica (Kállai Gravel Formation). Emszt (1911) was the first to investigate the composition of the rock which constitutes one of the blockfields (at Szentbékkálla). He found that this rock contains $SiO_2$ (97.77 %), $Al_2O_3$ (0.91 %), $Na_2O$ (0.36 %), $Fe_2O_3$ (0.14 %), $H_2O$ (0.41 %) and $K_2O$ in traces.

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**Fig. 10.1** Panoramic view of the Kál Basin with blockfields (photo by Gábor Tóth)

According to Lóczy (1913) and Cholnoky (1929), the ridges on which the blockfields developed, were beaches of Lake Pannon ca 10–5 Ma ago. The ridges were subject to weathering and wind erosion in the Pliocene under semiarid climate. According to Bulla (1954), the sand was cemented into sandstone to the effect of hot water percolation (hydrothermal activity). Hydrothermal processes were probably associated with postvolcanic activity subsequent to the basalt volcanism at the end of the Pliocene. According to Győrffy (1957), the silica precipitated from magmatic hot water cemented the sand. He proved the effect of high temperature by the recrystallisation of angular sand grains, also supported by mineralogical investigations (Kalmár 2000). In the literature there are numerous other processes proposed to account for the cementation. According to Balogh (1992), the processes of precipitation could have happened through the mixing of salty and freshwater and did not require high temperature. Thiry and Bertrand-Ayrault (1988) explain the cementation of the Fontainebleau Sandstone in France, which is similar to the rock building the blockfields of the Kál Basin, by the dissolution of silica by infiltrating waters and precipitation at base level.

According to Bulla (1954), the sandstone strata were separated into blocks through freeze-thaw action in the Pleistocene. Győrffy (1957) claims that the blocks were dislodged and tilted from their original positions because the surrounding loose deposits were carried away by aeolian action and pluvial erosion. We assume that the cementation of sand hardened the rock only locally, which is supported by the fact that blocks of variable size are found in the quartz sand quarry of Salföld (particularly at the upper level), buried under loess. These blocks were partly crushed to smaller fragments at later time.

The dimensions of sandstone blocks in the blockfields are highly variable. Cemented gravel beds, with grains smaller than 6 mm in diameter, are found in the sandstone in bedded or lenticular intercalations. The conglomerates are classified into fine-grained (gravels of 2–10 mm diameter) and medium-grained (over 10 mm) conglomerates. Conglomerated sandstones are dominated by sand, while finer to medium-grained conglomerates mostly contain gravels of 2–10 mm diameter but those over 10 mm also occur. Fine-grained conglomerates are widespread in the Szentbékkálla blockfield, while medium-grained conglomerates only occur in lenticular intercalations of 5–10 cm or 40–50 cm typical thickness between the sandstone and fine-grained conglomerate strata. Where the sandstone is continuously exposed on the surface, the south-looking scarps show gradients between 5° and 20°. The inclination of the tilted block surfaces range between 0° and 90° and are not formed along bedding planes, but on the scarps of strata.

## 10.3 Pseudokarren

Karren features on granite, gneiss and sandstone were first described by Cvijič (1924). Bögli (1960) used the term pseudokarren for features on granite, gypsum, sandstone and basalt. Anelli (1973) distinguishes para- and pseudokarsts. Parakarst denotes solution forms on limestone, while pseudokarst does not result from solution. Several authors studied pseudokarren on different rocks, including granite (Klaer 1956; Rasmusson 1959), halite (Andreichuk and Eraso 1996; Macaluso and Sauro 1996), gypsum (Macaluso and Sauro 1996 and Calaforra 1996) and sandstones (Bulla 1954; Robinson and Williams 1992) and described various forms from different rocks. These include rillenkarren, rinnenkarren, kamenitzas from granite (Rasmusson 1959; Hedges 1969; Dzulynski and Kotarba 1979; Migoń and Dach 1995), ripples on sandstone (Dyga et al. 1976), kamenitzas, rillenkarren, rinnenkarren and different types of sandstone polygons (Robinson and Williams 1992); rinnen and pits on basalt (Bartrum and Mason 1948), rinnen, grikes, labirynth karst on quartzite (White et al. 1966; Marker 1976; White 1988; Brook and Feeney 1996) and several microforms (for example, microrills and micro-scale meanderkarren) on halite and gypsum (Szablyár 1981; Macaluso and Sauro 1996). Large-scale pseudokarst forms (e.g. avens) are identified on sandstone (Szczerban and Urbani 1974; Dyga et al. 1976). Siegel et al. (1968) found dripstones with silica content. This fact indicates that silica is carried into the solvent. There are basins probably of pseudokarren origin and several metres across, generated on sandstone. Working on granite, Goudie and Migoń (1997) designate them as "pits" (if they are minor) and "sumps" (if they are large) and claim that their development can be jointly explained with biogenetic impact, aeolian erosion and dissolution. Colveé (1973) presented solutional features from a cave in the Roraima area, southeastern Venezuela, formed in sandstone.

### 10.3.1 Factors Affecting Solution on Sandstone

Denudation of the amorphous siliceous sandstone takes place in the following steps:

- As the rock disintegrates into grains, amorphous silica dissolves (Jakucs 1977; Veress and Szabó 2000), as known from dolomite or greenschist outcrops too (Veress et al. 1998). If the dissolved material is carried away by a stream or if the water infiltrates into the rock, denudation only affects the surface.
- The size of crushed grains is further diminished by solution. Chalcraft and Pye (1984) present slow dissolution of the crystalline quartz studied under the electron microscope. The remaining grains are fragmented by freeze-thaw action or thermal (insolation) weathering and transported away by the above mentioned processes and by wind.

  The rate of solution depends on the following factors:
- The quartz material of the sandstone may be amorphous or crystalline. The rate of dissolution for amorphous silica may exceed that of crystalline quartz a hundred times (Siffert 1962).
- Under tropical climate, crystalline quartz becomes amorphous quartz (White 1988) and, as a consequence, dissolution intensifies.
- According to Siffert (1962), Kennedy (1950), White et al. (1956), Moray et al. (1962, 1964) and Siever (1969) the solubility of amorphous silica depends on temperature, pH and the duration of solution. Amorphous silica is a Si–O–Si bond polymer compound, which disintegrates through the splitting of $OH^-$ ions during the solution process. The resulting smaller grains are scattered in the water. At first, in the initial stage of solution, when pH is low, significant expansion probably happens, and heavier silicic acid is generated with silica depositing faster. Then, if solution is prolonged at pH 8.5, silicate ions emerge. Different plants may increase the pH of the water.
- The amount of the dissolved material depends on water availability and the duration of solution. The volume of water is determined by the type of rainfall and its intensity, evaporation, soil, vegetation and block size. Solution may last for several weeks, depending on the shape of preexisting pseudokarren features and on evaporation, influenced by temperature, amount of rainfall, and again the shape and size of pseudokarren (Veress et al. 1999). Evaporation is low if a pseudokarren form is deep and has an overhanging wall. The duration of solution depends on the slope of the block surface, which controls the rate of runoff, with slower flow increasing the efficiency of solution, and on rock transmissibility.

### 10.3.2 Main Types of Pseudokarren

Kamenitzas are the most common features on blockfields and classified as simple or complex (Fig. 10.2). Seen from above, they are of circular or elongated shape, developed along cracks or on the sides of blocks, often fed by rinnen and delimited by wavy margins. Kamenitzas occur on vertical or overhanging, more soluble block walls (Fig. 10.3). If the kamenitza is deepened into sandstone, its bottom is flat. Circular kamenitzas have varied cross-sections, shallow features with gentle slopes and flat bottoms being most common. Kamenitzas may merge into uvala-like kamenitzas. Double kamenitzas with internal round ridges are common. Complex kamenitzas are single complex or manifold complex, divided into partial kamenitzas. Manifold kamenitzas enclose ridges and karren "inselbergs".

In the Kál Basin kamenitzas of diverse, often complex, morphology are found on blocks (Fig. 10.4). Shallow kamenitzas are elongated in north to south direction. Since the solution on the block surface takes place at remarkably fast rates, neighbouring kamenitzas may combine, with lower ridges remaining between them. The majority of the deepest and largest kamenitzas are found around the main crack across the block (Veress and Kocsis 1996).

The pits (pipes, karst wells) on the sandstone of the Salföld blockfield are either simple pits or uvala (i.e. merged) pits. In the case of complex pits the floor is divided into smaller pits. Straight or meandering rinnen (troughs), 10 cm–2 m long, 1–15 cm wide and at most 10 cm deep, form on the surface of lodged blocks and feed kamenitzas. Decantation rinnen begin at kamenitzas and usually end at the margin of the block ("marginal rinnen"). Rinnen with various cross-section emerge at cracks and represent transition from rinnen to rills on limestone. Polygonal features on sandstone are straight or arcuate, sometimes crossing each

**Fig. 10.2** Kamenitza developed next to rinnen and on the margin of a block (Szentbékkálla blockfield) (photo by Gábor Tóth)

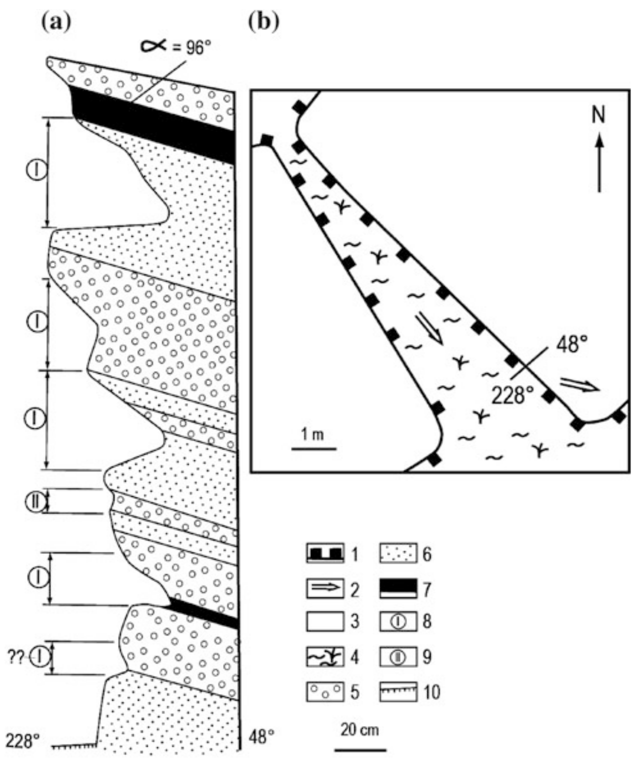


**Fig. 10.3** Typical section of a fissure wall (Szentbékkálla blockfield). **a** Lateral view. *5* Medium-grained conglomerate; *6* fine-grained conglomerate; *7* sandstone; *8* tafoni; *9* solutional notches; *10* soil; *α* direction of the dip of strata; **b** plan view. *1* Corridor; *2* slope direction; *3* type section (with *direction*); *4* soil and vegetation

other, 1–2 cm wide and deep. A pattern is created when they connect to each other and depends on the length and shapes of the polygons linked. The surfaces of some major blocks with steep slopes are subdivided by step features. If they are gentler, their width is 40–50 cm, if steeper, it is 10–15 cm. Kamenitzas frequently exist on the thread of steps. Grikes of some tens of centimetres are solitary and rarely create grike joints. They evolve from cracks through solution.

“Thimble karren” of 1–2 cm diameter and depth are often elongated, but the direction of their longer axis differs from the slope direction of the bearing surface. They primarily occur in considerable density (5–6 karren in 1 m$^2$) on the walls of kamenitzas on the blocks at Salföld. Karren bridges are formed if the oblique pits or pits changing direction cut through the block. Miniature karren “inselbergs” are residual forms (Fig. 10.5), which originate where the sandstone is dissolved on the surface or where neighbouring kamenitzas merge. They appear on flat or slightly dissected surfaces or, as mentioned above, inside larger pseudokarren. Marginal karren “inselbergs” usually appear on the blocks at Salföld.

The following major features, in fact large pseudokarren, occur on block walls: solutional notches with length exceeding their height and depth are medium to large sized features, described from the sides of limestone hills (Willford and Wall 1965; Jennings 1985), where runoff from the soil at the limestone boundary reaches the limestone and dissolves it, and pockets of some tens of centimetres in horizontal direction, similar to solutional notches (Fig. 10.6).

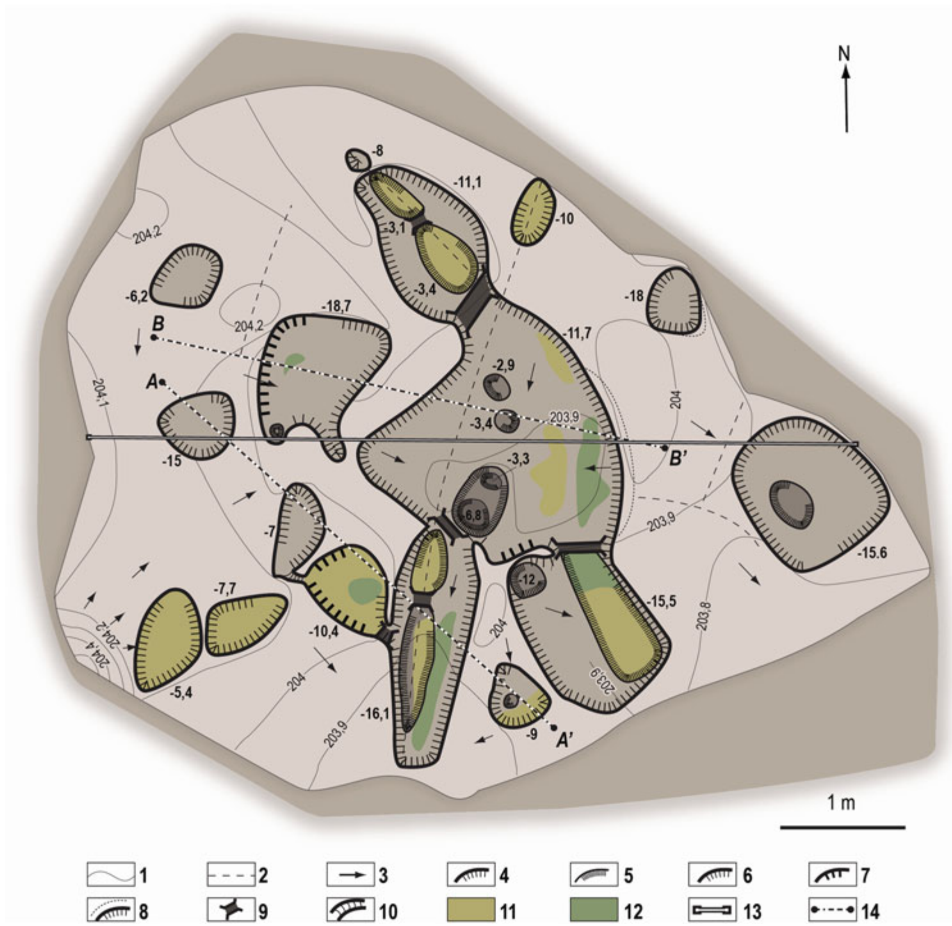


**Fig. 10.4** Morphological map of a block (Szentbékkálla blockfield). *1* Side of the boulder with depth (m); *2* contour line; *3* angle of the direction of the sloping block surface; *4* Kamenitzas with depth (tens of cm); *5* inner kamenitza; *6* gently sloping side of kamenitza; *7* vertical side of kamenitza; *8* overhanging wall of kamenitza; *9* ridge between kamenitzas; *10* rinnen; *11* lower plants (moss, lichen); *12* grass-like vegetation; *13* cross-section (with specific solution); *14* cross-section (longest axis of gravels is measured at 5 cm intervals)

**Fig. 10.5** Miniature inselberg in the Kál blockfield (photo by Gábor Tóth)

**Fig. 10.6** Pockets on the side of a boulder (Szentbékkálla) (photo by Gábor Tóth)

## 10.4 Origin of Microforms

Kamenitzas and pipes can only form by dissolution because they do not have outflows. The following evidence point to dissolution:

- amorphous silica is present in the rock;
- gravels and sand grains in pseudokarren indicate disintegration through the dissolution of the cementing material.

The number of pseudokarren increases with slope and kamenitzas also mainly occur on gentle block surfaces. Kamenitzas are elongated where block surfaces are cracked or steep (Fig. 10.4). Transitional forms between kamenitzas and troughs are also observed. Troughs on the margin of blocks are the outflows of kamenitzas (decantation runnels). The number of the troughs, however, decreases with increasing slope.

A relationship is found between the size of pseudokarren of the Salföld blockfield and the area of block surfaces. The larger the block, the larger are the pseudokarren. The reason behind this is that the pseudokarren on larger blocks receive more water from their surroundings than those on smaller boulders. The depth of kamenitzas depends to a lesser degree on block size, but the size of pipes is a function of block size. We can explain this relationship as follows. In the case of kamenitzas surface runoff prevails, but in the case of pipes more water can percolate into the rock along cracks and helps to remove the dissolved material. Therefore, pipes can deepen more intensively on larger than on smaller boulders.

The development of kamenitzas of various size and shape is explained by variations in the transmissibility of rocks. Kamenitzas on sandstone tend to widen rather than deepen because water percolation is moderate into this type of rock. Therefore, runoff water flows out of the pseudokarren, promoting the horizontal development of kamenitzas. Part of the water percolates into the conglomerate carrying dissolved material into the rock and resulting in kamenitzas being deepest along cracks (Fig. 10.4). The dissolved material may precipitate again between the grains, within cracks, along the bedding planes or it can be carried away by water which flows off the block. We noticed siliceous crusts of a few centimetres' thickness on bedding planes of a block in the Szentbékkálla blockfield, probably precipitated from runoff water. Dissolution is most efficient on fine-grained conglomerate, where runoff or percolation may be at optimum, hence drainage is neither too fast nor too slow. We think the duration of dissolution is quite long and the amount of dissolved material is higher than during fast percolation and short-duration dissolution. Percolation induces deepening of pseudokarren. The quantity of the dissolved material increases with depth of the feature and decreases with the rate of runoff. Intensive dissolution produces larger pseudokarren. Kamenitzas in medium-grained conglomerate deepen rather than widen, suggesting that most of the runoff percolates into the rock.

If we compare the grain size on surfaces which are lowered by dissolution at different rates, we can prove that the intensity of dissolution depends on grain size. Because of the optimum runoff rate, dissolution is more intensive where smaller gravels occur.

If we compare specific dissolution in both blockfields pseudokarren formation is represented by kamenitzas. Furthermore, rinnen and grikes occur in the Szentbékkálla blockfield, while pits are more typical for the Salföld blockfield. The rate of specific dissolution is higher in the Szentbékkálla blockfield (54.48 cm $m^{-1}$ on average) than at Salföld (24.30 cm $m^{-1}$). The likely explanation is that the blocks of the Szentbékkálla blockfield are larger and had been earlier exhumed than those at Salföld.

The shape of pseudokarren depends on the quality and position of the different strata. Flat-bottomed kamenitzas emerge if the kamenitza reaches a fine-grained bed during deepening. Later the excavation of the form stops and it expands laterally. Extended kamenitzas, uvala kamenitzas and rinnen appear along cracks if slope is steep on fine-grained sandstone surface. In the later case the shape of pseudokarren is determined by the behaviour of rivulets or by the rate of percolation. Finally, manifold complex kamenitzas result if fine-grained sandstone strata underlie the medium-grained conglomerate zone. The manifold complex kamenitza assembles if beds of different grain size alternate. Inner kamenitzas appear in the sandstone on the bottom of major kamenitzas and reach the fine-grained conglomerate bed as they are growing. Some sandstone patches are preserved on the floor as pseudokarren 'inselbergs'.

Sandstone polygonal forms develop independently from the rock joints. Probably desiccation cracks appeared on the rock surface preceeding diagenesis, then filled with amorphous silica. The dissolution of the rock should have happened along the desiccation cracks. The steps formed on blocks where the difference between the slope direction of the block surface and the dip of beds is 180°. The risers of the steps are built up of fine-grained conglomerate scarps, which retreat faster than the sandstone beds because of dissolution.

Tafoni evolve on sandstone banks (Segerstrom and Henriquez 1964; Jakucs and Csuták 2000; Veress et al. 2002). Since their present evolution is not affected by rainwater runoff, they must have been generated by percolation into the rock. They could also develop under circumstances different from those of today if the surface surrounding the blocks had been more elevated than at present. Previously rainwater of the older surface flowed off on the side of the blocks. As the surface was lowered, more and more tafoni could have emerged. Pseudokarren "inselbergs" result where the surface is less soluble, e.g. in lenticularly intercalated sandstone. At the top of the blocks, in lack of solution, the original surface is preserved and marginal "inselbergs" emerge.

## 10.5 Conclusions

Sandstones and different conglomerates of the Kál Basin were cemented by amorphous silica deriving from hydrothermal activity. The blocks were formed by local cementation and later fractured by freeze-thaw action.

The most typical forms on blockfields are kamenitzas, troughs, grikes and pipes. In the Kál Basin blockfields, karren "inselbergs", polygons, solution notches and pockets seldom occur. There are pseudokarren, which develop through percolation (for example, kamenitzas and pipes), and originate if surface slope of a boulder is gentle or they are far from margins of a block. A kamenitza becomes more elongated on sloping surfaces. Joints promote the development of grikes, pipes and elongated kamenitzas. Steep block surfaces also give rise to troughs, which receive abundant water from the block top of gentle slope (marginal trough) or from kamenitzas (decantation runnel).

Pseudokarren are elongated because surface slope and rock joints conduct runoff and percolation. Kamenitzas develop if the rock is well-bedded and amorphous silica content is low. Poorly developed bedding and high joint density favour grike and pipe formation. The shape of the kamenitzas and pipes primarly depends on grain size: in coarse grained rock pseudokarren are deeper, while with decreasing grain size, their depth is reduced. First of all, solution notches are cut by non-local dissolution in the walls of blocks along bedding planes, while tafoni and pockets are due to local dissolution. The density of pseudokarren features also depends on the gradient of a block surface: on steeper surfaces, lower density of pseudokarren is observed.

The size of pseudokarren and the intensity of dissolution depend on the size and slope of the block, its amorphous silica content and the grain size of the rock—pseudokarren on larger blocks collect more water and, therefore, more material is dissolved. Transmissibility also depends on grain size and jointing. Percolation provides ideal conditions for pseudokarren formation on fine-grained conglomerate since water with the dissolved material moves not only on the rock surface, but also within the rock. The mass of the dissolved material is higher, and because of slow infiltration, solution will operate for a longer time. Similarly, the duration of the solution is increasing with gentler surface slope.

The shape of pseudokarren depends on the development and configuration of beds of different grain size. Low inclination walls on sandstone and steep walls on conglomerate are typical. In the first case pseudokarren widen, in the second case they mainly deepen. Elongated forms are established along cracks or if the block surface is tilted. Asymmetrical forms of scarps appear on the block: steeper on fine-grained conglomerate and more gentle on sandstone. We explain the different dissolution of the wall with variations in permeability.

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# Granite Weathering in the Velence Hills

11

Gábor Mezősi

**Abstract**

The landforms of granite weathering in the Velence Hills are unique in Central Europe. Through intensive weathering since the end of the Permian the exposed granite mass was transformed to saprolite. Weathering combined with sheet wash worked on the Eocene andesite too and resulted in heavy denudation. Since the Late Tertiary tectonic movements also contributed to the fragmentation of the granite mass, which was accompanied by the decay of the surface saprolite. The Velence Hills are best known for granite ridges and tors (balanced rocks). Tectonics had a fundamental contribution to the development of landforms. The water entering from both below and above to the joint network had a significant weathering effect on the one hand through its hydrothermal effect and on the other hand (from the other direction) through hydrolysis (both affected the feldspars and talcs of the granite). The landscape on surfaces decomposed hydrothermally or through hydrolysis are clearly distinguishable from the land form related to the intact granite surfaces.

**Keywords**

Variscan granite • Saprolite • Spheroidal weathering • Tors • Ridges • Tectonics • Hydrothermal affect • Velence hills

## 11.1 Introduction

Primarily due to its geological properties, the Velence Hills show particular morphological characteristics within the Carpathian Basin. According to the Inventory of Microregions in Hungary (Dövényi 2010), the microregion covers an area of 80 km$^2$, while geologists mention that over an area of 140 km$^2$ bedrock is in near-surface position (Benkó 2008). The highest point is Meleg Hill (352 m a.s.l.). West of it, the base point for the geodetic survey of Hungary is located (at Nadap, elevation above the Baltic Sea level: 173.16 m).

The geomorphologically most valuable and most spectacular part of the area, the Pákozd Perched Rocks Nature Reserve (44 hectares), has been protected since 1951. Today it is under the administration of the Directorate of the Danube-Ipoly National Park established in 1997. The origin and special properties of this landscape are presented in this chapter.

## 11.2 Geographical Setting

The low Velence Hills are located to the north of Lake Velence in northern Transdanubia (Fig. 11.1). The hills present a typical Transdanubian landscape. The climate is moderately cool and moderately dry, with mean annual temperature around 10 °C and average annual precipitation 550–600 mm. The surface is drained by small streams running towards the lake or by the Császár-víz Stream in the

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**Fig. 11.1** The Velence Hills with the lake to the south at Pákozd (photo by Tamás Fábián)

northwest. The soils are predominantly brown forest soils (Cambisols) (Barczi et al. 2003).

Despite the subdued relief, the landscape has a great aesthetic value due to the presence of spectacular granite weathering forms. In addition, in the southern foreland of the hills Lake Velence is located (area: 236 km$^2$, depth: 1.5 m—Fig. 11.1), with a significant recreational value due to the proximity of Budapest. Although granite features are the main attractions, the gentle hillside slopes covered by forests and vineyards as well as reedbeds covering half of the lake and the associated tussocks together with their unique bird population (nature reserve) make the region a tourist attraction.

## 11.3 Geological Setting

A typical allochtonous segment of the Mid-Transdanubian Unit (see Chap. 2), the Velence Hills, are divided into two sections:

- The western part is ca 10 × 3 km in size, with hilly relief and NW-SE strike. It is built of a granite intrusion complex of Upper Carboniferous, Variscan (280–300 Ma—Buda 1981) age, transformed by planation processes. The highest peaks (including Meleg Hill) are composed of metamorphosed granite and quartzite.
- The central and eastern parts are built of Oligocene to Eocene Alpine calc-alkaline intrusive quartzite and metamorphosed andesite. Volcanism and associated hydrothermal activity occurred in the Eocene (Kubovics 1958), similarly to the North Hungarian Mountains. Such activities are most likely caused by tectonism or, more precisely, mark the onset of subduction in the Magura Ocean, in the northern part of the Tethys Ocean. Thus, from the middle Eocene on an andesite-dacite volcanic ridge developed in the area of the ALCAPA plate (see Chap. 2).

The host rocks of magmatic formations are early Paleozoic (Ordovician–Silurian) schists and phyllites (Benkó 2008), which are the oldest formations of the hills. The schists under the Velence Mountains and their neighbourhood underwent metamorphism, thrusting and folding in the Carboniferous period (Dudko 1988). The granite itself, however, was not affected by Variscan metamorphism. The intruding granite upwarped the old sedimentary rocks as a dome and cooled and solidified very slowly in the form of a laccolith and several smaller intrusion bodies (dykes and minor stocks) (Vendl 1914).

On the basis of paleomagnetic investigations, it is claimed that the Velence Hills were a part of the African Plate and the Paleozoic granite intrusions in the core of the mountains solidified into rock in the southern hemisphere, along the Tropic of Capricorn (Mezősi 2011; Haas 2012), and merged with the European plate during the late Paleogene–early Neogene.

The area was characterized by a lack of sedimentation and prolonged denudation from the Permian period until the end of the Mesozoic. Eocene volcanic formations only extend over a limited area. In the later Tertiary andesite lava reached the surface through fractures and subsequent hydrothermal activity produced ores of negligible economic significance (Jantsky 1957; Tarsoly 2013).

The eastern border of the mountains is the Vál fault. To the west the granite gradually subsides below the Neogene sedimentary cover. The mountains are surrounded by Late Miocene and Holocene marine-lacustrine sediments and fluvial deposits.

## 11.4 The Variscan Granite

The granite of the Velence Hills is not of the classical granite type characteristic of other areas of the Hercynian/Variscan orogeny. The Variscan intrusion is a well-differentiated biotitic granite crystallized at ca 600–700 °C temperature and at ca 2 kbar pressure (Buda 1981). The biotite is accompanied with quartz, orthoclase and plagioclase (Jantsky 1957; Buda 1981) as well as various accessory minerals such as apatite, zircon, rutile, magnetite, ilmenite, allanite, xenotime and titanite (Vendl 1914; Buda 1981). The intrusion may be postorogenic or rift related (Uher and Broska 1994) as after differentiation the residual granite melt intruded into the opening extensional fractures.

The main body of granite intrusion is crossed by several joint systems (Gohkahle 1964; Horváth et al. 2004). NE-SW and NW-SE directions are predominant and the N-S and E-W oriented joints are subordinate. However, the statistics of the orientation of joints slightly varies in different parts of the mountains (Horváth et al. 2004).

### 11.4.1 Geomorphic Evolution and Typical Landforms

In the present-day topography of the low Velence Hills landforms are subdued, among them the slightly fragmented planated surfaces of 250–300 m elevation (Figs. 11.2 and 11.3) and fossil pediments of low inclination are predominant. Since the end of the Permian (the last period of granite intrusions) planation under tropical and subtropical climate has continued without interruption. Intensive weathering also affected the schist, which was transformed to saprolite. Weathering combined with sheet wash worked on the Eocene andesite too and resulted in high denudation rates (Ádám 1993). Since the Late Tertiary tectonic movements also contributed to the fragmentation of the granite mass, which was

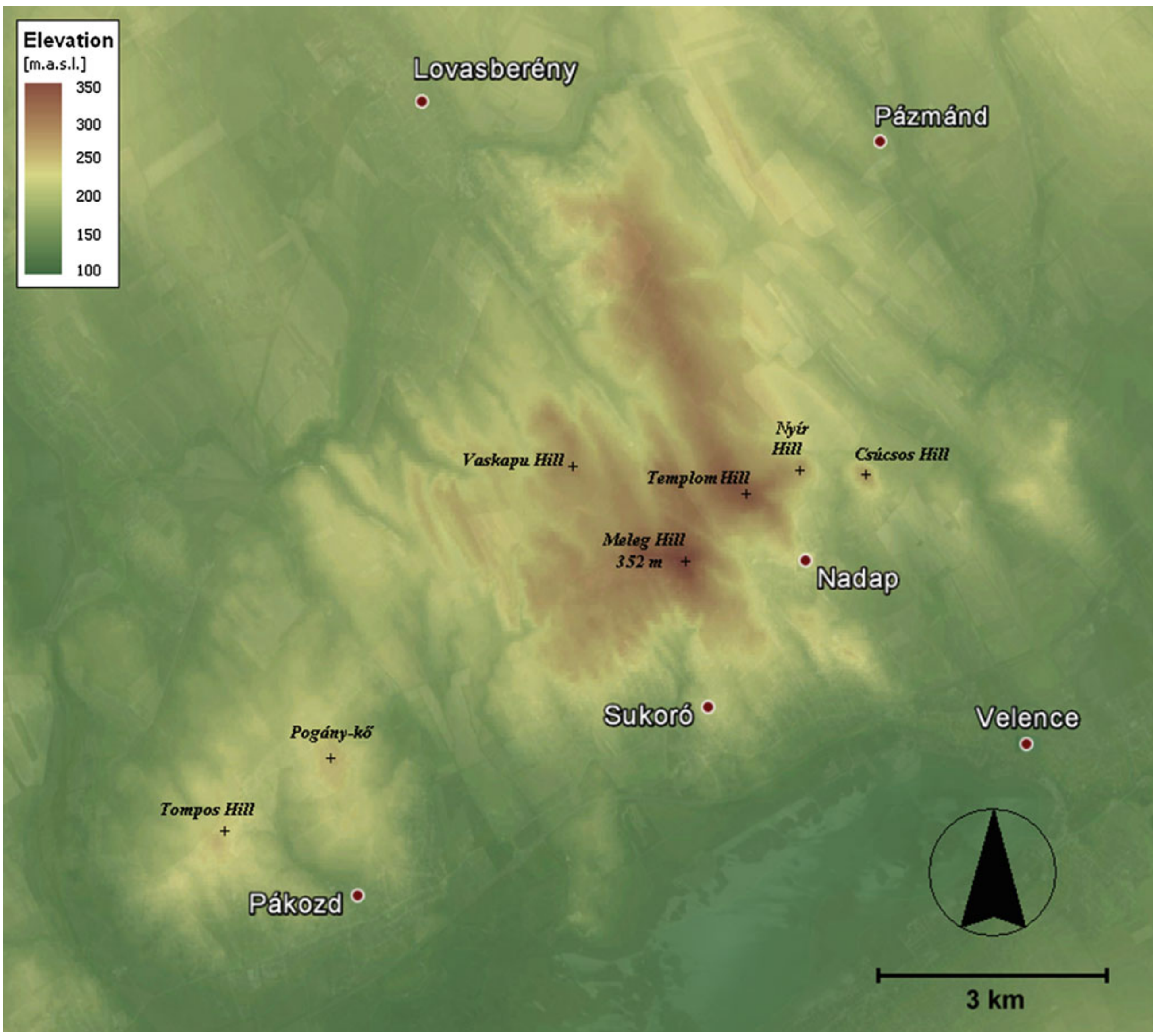


**Fig. 11.2** Digital elevation model of the Velence Hills

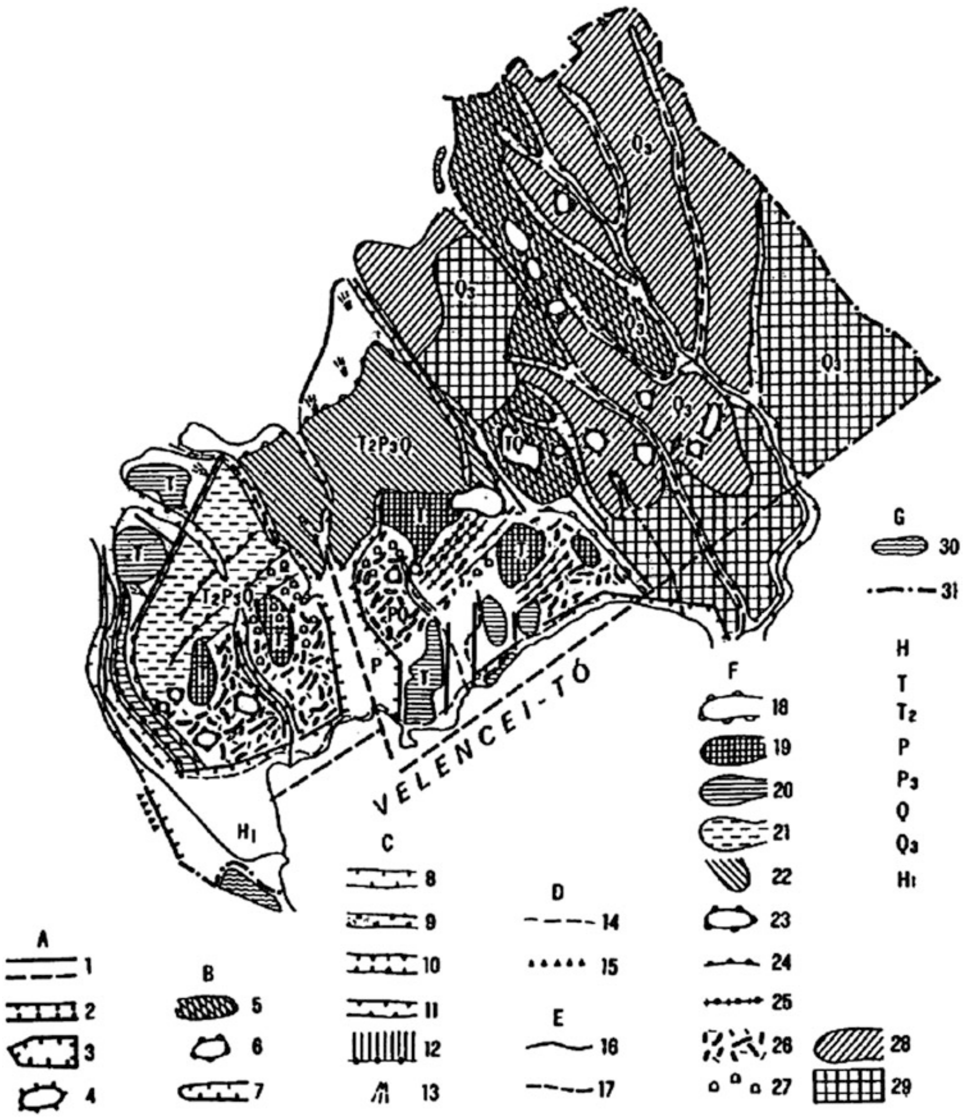


**Fig. 11.3** Geomorphological map of the Velence Hills (after Ádám 1993). **A** Exogenic landforms: *1* fracture; *2* fault; *3* tectonic basin; *4* secondary volcanic cone. **B** Derasional landforms: *5* derasional ridge; *6* residual mound (monadnock); *7* derasional valley. **C** Erosional landforms: *8* erosional valley undifferentiated; *9* flat-floored erosional valley; *10* high-gradient erosional valley; *11* ravine; *12* late Pleistocene terraces (nos IIa and IIb); *13* colluvium. **D** Lacustrine landforms: *14* old shoreline; *15* fossil abrasional platform. **E** Drainage: *16* permanent stream; *17* intermittent stream. **F** Landforms of complex origin: *18* planation surface; *19* uplifted planation remnants; *20* subsided and later exhumed planation remnants; *21* exhumed fossil pediment; *22* covered fossil pediment; *23* granite monadnock; *24* eroded dyke; *25* eroded hogback; *26* peneplain remnants; *27* tors, balanced rocks; *28* erosional-derasional ridge; *29* loess-mantled plain. **G** Anthropogenic landforms: *30* fish-pond; *31* Watershed. Age of landforms: *T* Tertiary undifferentiated; $T_2$ Late Tertiary; *P* Pliocene undifferentiated; $P_3$ Late Pliocene; *Q* Quaternary undifferentiated; $Q_3$ Late Pleistocene; *H* Early Holocene; $H_2$ Late Holocene

accompanied by the removal of the surface saprolite. The impact of these processes was enhanced by post-magmatic processes, during which hydrothermal decomposition of granite and dyke rocks took place (Kubovics 1958).

Lake Velence to the south as well as the Zámoly Basin to the north originated by a very recent (Late Pleistocene, 15–10 ka old) subsidence (Fig. 11.2), which accompanied the latest stage of uplift of the Velence Hills.

From a geomorphological aspect, the hills are best known for granite landforms due to selective weathering and stripping, mainly granite ridges, tors (in the sense of Linton 1955 and Gerrard 1988), and balanced (or perched) rocks (Fig. 11.4). The aforementioned tectonics, involving heavy jointing, played a fundamental part in the development of landforms. As a result of contractions and tectonic forces, fissures and cracks emerged in the cooling granite mass.

**Fig. 11.4** Typical balanced rock (Pandour Rock) at Pákozd (photo by Attila Barczi)

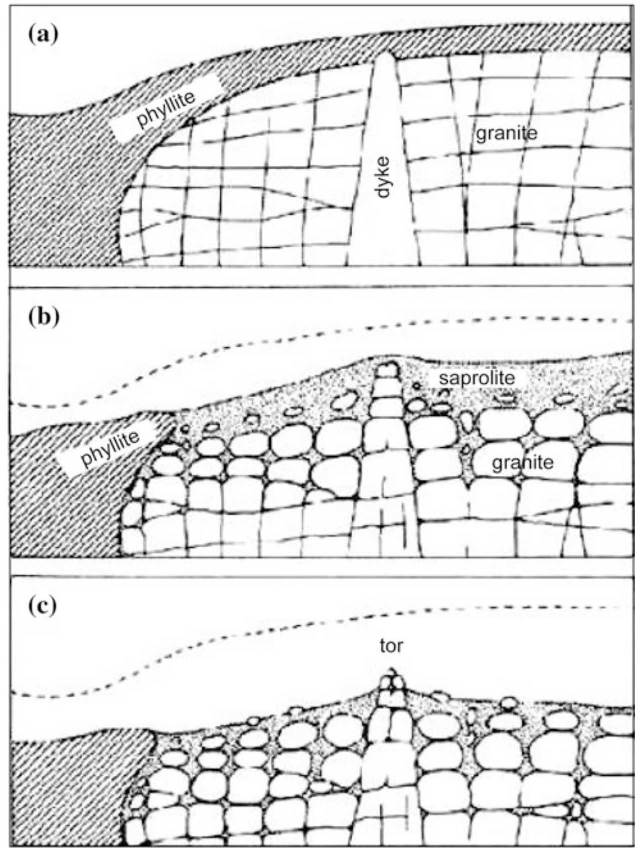


**Fig. 11.5** Selective (spheroidal) weathering of the Velence granite. **a** Solidification joints in granite laccolith. **b** Weathering into blocks. **c** Blocks exposed through the removal of saprolite (after Eszterhás 2009)

Water entering the joint network from both below and above had a significant weathering effect with its hydrothermal effect on the one hand and (from the other direction) with hydrolysis on the other hand (both affecting the feldspars and talcs of the granite). The landscape on surfaces decomposed hydrothermally or through hydrolysis are clearly distinguishable from the landforms on intact granite (Fig. 11.4). The resulting regolith, rich in clay minerals and colloidal acids, has been washed away by running water. As a result of differential weathering, saprolite of irregular depth and characteristics was formed near the topographic surface.

In the course of selective weathering, granite blocks began to separate the subsurface due to joint widening, according to Linton's (1955) tor formation model. Due to the small extension of the area the chemical composition of water involved in the process had a limited impact on the landforms produced, while the joint pattern has been a much more significant factor.

The conditions resulting from slow evolution during hundreds of millions of years fundamentally changed when the mountains emerged to the surface. Uplift started in the late Tertiary but mainly happened in the Pleistocene. Then the granite blocks, which previously had evolved underground, became exposed from below the debris mantle. Exhumation and further modelling of granite ridges is still taking place (Fig. 11.5). Thus, granite ridges and blocks exposed as tors or balanced (or perched) rocks appear on the surface (Fig. 11.6). The position of such features indicate a parallel drainage network, which emerged after the erosion of the weathering mantle.

Today the specific erosional landforms appear most attractively as balanced or pedestal rocks a few metres high, locally called "woolsacks" (Figs. 11.6 and 11.7). They occur at 180–240 m elevation in the eastern (north of the village Pákozd: Pagan, Lion, Cube and Pandour Rocks) and central area (around Sukoró: Woolsack, Hollow Rock) of the mountains. Their rounded and ellipsoidal shapes were developed by spheroidal weathering. This process is governed by major differences in temperature and pressure, smaller at the surface compared to the conditions at the depth where the rock solidified. During weathering thin sheets of a few millimetres' thickness are produced by fine jointing on the granite surface and detached over time (exfoliation). As a consequence of the repeated exfoliation process, sharp angles are smoothed and disappear and the shape of rocks is

**Fig. 11.6** Balanced rock at Pákozd (photo by Tamás Fábián)

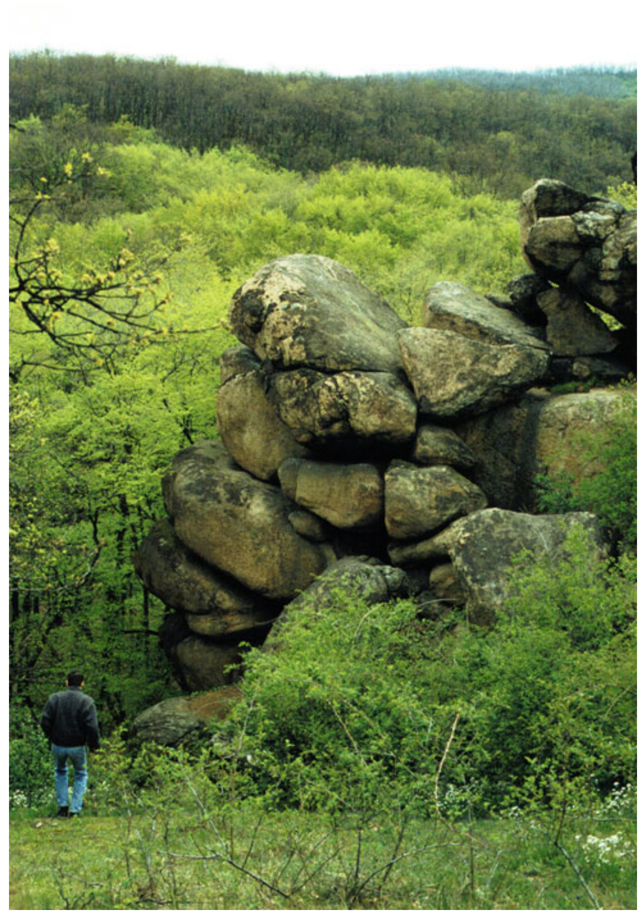

**Fig. 11.7** Tor at Pákozd (photo by Tamás Fábián)

becoming increasingly rounded. The exposed corestones surrounded by weathered material are called woolsacks. In the Pákozd group of tors ten woolsack caves have been identified (Eszterhás 2009). As opposed to caves formed through the displacement of blocks and dilatation of cracks (pseudocaves), woolsack caves are true caves, resulting from the removal of saprolite between blocks (Tarsoly 2010, 2013). They are of variable dimensions, a three-dimensional passage system of tens of metres of length, usually with several entrances and storeys.

## 11.5 Conclusions

Some of the granite ridges and tors are picturesque and sometimes even bizarre components of the landscape. The balanced rocks often appear as if they were stones thrown upon each other by human hands, being in unstable equilibrium threatening with immediate collapse. Actually, in the course of selective weathering, with the expansion of the cracks the granite blocks began to separate underground. In recent times, more and more huge granite blocks, which previously evolved under the surface, are becoming exposed. The exhumation and further modelling of blocks of the granite ridges through repeated exfoliation is still taking place.

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# The Somogybabod Gully: Hidden Erosion (Piping) in the Somogy Hills

12

Gergely Jakab and Zoltán Szalai

**Abstract**

Gully formation is a serious problem in crop production, landscape management and freshwater eutrophication. A recent spectacular example, the Somogybabod gully came into being in the early summer of 2010 after an extremely wet spring. Usually a very rapid process, especially on loess, gully development in this case was supported by improper land use, catchment management and previous piping. Piping and tunnel formation take place invisibly underground. When tunnel roofs collapse, rather deep gullies are exposed within few minutes—as it happened at Somogybabod. Although several active permanent and ephemeral gullies are found in the surroundings, only the Somogybabod gully attracted the attention of scientists because it was initiated under forest cover and, threatening houses, was also highlighted by the media. The aroused public interest could provide an opportunity to face and solve the problem. Instead of runoff mitigation, however, only the filling up of the gully was considered—no long-term solution without the control of surface hydrology and field-scale complex amelioration.

**Keywords**

Gully erosion • Piping • Runoff • Loess features • Outer Somogy hills

## 12.1 Introduction

Soil erosion is normally too slow to produce spectacular landforms. In the summer of 2010, however, after an extremely wet spring and early summer, huge ravines were created in the loess covered hill regions of Hungary. The media reported about “craters” or “the earth opened up”. In reality, these phenomena resulted from gully erosion of unusual scale. The biggest and most impressive gully was formed next to the village of Somogybabod, in the western part of the Outer Somogy Hills (Fig. 12.1).

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The ravine rapidly became a widely known symbol of improper land use and the resulting soil erosion related problems. However, it is not a unique phenomenon: there are several gullies of the same size within 1 km distance from each other. The older gullies are located in a forest and formed over several decades (Fig. 12.2). It is usual in Hungary that, as a consequence of improper activities in forestry, huge gullies develop under forest vegetation and they do not attract public interest. On the other hand, if a gully extends over an agricultural field and even endangers buildings, it could be viewed as a sensation. Moreover, the abrupt excavation of the Somogybabod gully also seemed to be exceptional.

For geomorphologists, the short-lived popularity of this spectacular landform provided a great opportunity and can be exploited to supply scientific information, relevant to many other regions of Hungary, to the educated public on geomorphic processes and hazards. In this chapter the background knowledge to piping in loess in Hungary is

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**Fig. 12.1** The Somogybabod gully (photo by Gergely Jakab)

summarized and then exemplified by the Somogybabod gully. This is the most recent landform among all presented in this volume.

## 12.2 Mechanisms of Gully Formation

One of the most important influences of climate change in Hungary is the increasing frequency of extreme precipitation events (Bartholy and Pongrácz 2013). The rainstorms generate huge amounts of surface runoff, which is a very effective agent of geomorphic evolution. Accordingly, the landforms created by heavy runoff are less stable and can change considerably over weeks. While walking in the hilly countryside after a heavy rainstorm, numerous signs of soil erosion can be observed: rills and gullies are easily recognizable on the surface. But there is another, almost invisible erosion process beneath the surface called piping (suffosion) (Bryan and Yair 1982; Jones 2004). Concentrated surface runoff enters underground through a hole and carves pipes and tunnels of variable dimensions.

When water collects on the surface, an increased volume of runoff involves higher and higher velocities and energy. When the kinetic energy exceeds certain thresholds (i.e. the critical shear stress value of the surface), it can detach and deliver soil particles. As a consequence, at first rills emerge on the ground surface, which may develop into gullies. The velocity of flow necessary for gully formation depends on several conditions, such as surface cover, topography and others, but the role of land use and the parent material is of primary importance (Gábris et al. 2003). On arable lands it is not unusual to find bare soil surfaces without any protection against the impact of falling raindrops. On bare surfaces raindrops easily destroy soil structure, break down the aggregates and create a sealed crust of nearly zero porosity. Due to this compacted surface layer, infiltration capacity dramatically diminishes and the volume of surface runoff is multiplied. Consequently, the proportion of the catchment used as arable land is a significant factor in the emergence of hazard, largely contributing to runoff generation. Approximately two thirds of the territory of Hungary is arable land and a considerable part is situated in hilly regions, where surface runoff and soil erosion are serious problems (Kertész 2009). On the other hand, the compact solid material is less susceptible to erosion than a loose or even collapsible deposit.

**Fig. 12.2** Panorama of an unnamed permanent gully 1 km from the Somogybabod gully. Note the collapses on the walls (photo by Gergely Jakab)

## 12.3 Piping in Gully Formation

Since large areas of Hungary are covered by unconsolidated sediments, linear (rill and gully) erosion plays an important part in recent geomorphic evolution. In terraced agricultural landscapes human impact can be highly influential in their generation (Csorba 2010). According to Kertész and Centeri (2006), neither gully erosion nor piping are of major significance in Hungary and there is no reference to piping either in the Hungarian chapter of the monograph on recent landform evolution in the Carpathian-Balkan-Dinaric region (Kertész 2012).

Faulkner (2006) classified piping phenomena in Europe into three main groups, such as piping in peat, dispersive materials and collapsible materials. However, large areas of Hungary are situated on Na-rich saline soils, where dispersivity is one of the main factors of soil formation, but flat topography prevents piping from acquiring any significance.

In some cases a considerable amount of concentrated surface runoff consistently infiltrates into the soil on the same spot. The enduring and intensive leaching mobilizes the cementing components of the soil or sediment. In Hungary this mobilization is mostly due to dissolution of $CaCO_3$ or through suspension of the finest clay particles in the presence of a high sodium content (LaFleur 1998). Losing its fundamental components, the structure of the material collapses and the individual particles are transported by subsurface flow (throughflow), resulting in a complex network of underground pipes or tunnels. Generally these forms develop from pre-existing macropores, such as cracks, root channels or animal burrows (Faulkner 2006). In general, the pipes end at a steep bank and the transported sediment is deposited in fans at the outlets, or, if there is no outlet, the sediment invisibly disappears in the groundwater.

In Hungary, both piping and gully phenomena are believed to be concentrated in the loess-mantled hill regions of the country. The most important property of loess is its homogeneous particle-size distribution: most of the grains fall into the 0.02–0.05 mm diameter class. There is, however, a very effective adhesive component, calcium carbonate content, which is essential for creation of a structure with higher than 50 % porosity and, consequently, a fairly high water infiltration capacity. This structure is quite strong and allows the formation of high loess bluffs (most spectacularly along the Danube—see Chap. 14). The aggregating $CaCO_3$ is a relatively soluble mineral and can be removed even by infiltrating rainwater, depending on its acidity (Kerényi 1994). The dissolved $CaCO_3$ is transported downwards in the soil profile, resulting in lower structural stability in the upper horizons and secondary calcite enrichment in the lower part of the profile. Recrystallization generally appears in certain layers as hardpans or spheroidal concretions (in Hungary often called "loess dolls"), which are important tools for the interpretation of environmental influences on the evolution of loess profiles (Barta 2011).

The erosion of the more or less homogenous loess profiles is often influenced by the presence of paleosol horizons of higher clay contents than the loess itself and, consequently, lower porosity and water conductivity. As the leaching water reaches the nearly impermeable horizon, it is impounded and forced to find an alternative flow direction, generally on the surface of the paleosol horizon. Since paleosols are much less prone to erosion, at paleosol interbeddings the rates of gully erosion or enlargement of suffosion tunnels are dramatically diminished (Fig. 12.3). In such cases, incision is replaced by widening as the principal process of gully evolution (Kertész 2004).

Continuous water flow along the same path gradually dissolves $CaCO_3$, reduces the stability of soil structure, and the loess collapses. Therefore, pipes often emerge on the interfaces between layers with different porosity or infiltration capacity. Such alternating layers in the loess profiles are paleosols or hardpans in soils, resulting from natural (e.g. lessivage in Luvisols) or human-induced processes (e.g. improper tillage). Repeated tillage operations, always affecting the soil at the same depth, can produce a sealed layer at the bottom of the tilled horizon, called plough pan. Considerable subsurface flow is generated on the top surface of the plough pan.

**Fig. 12.3** Ephemeral gully on paleosol, where the loess cover was removed (Somogybabod). Note the flow resistant 'steps' formed by ploughing (photo by Zoltán Szalai)

Another frequent setting of pipe formation is connected with active permanent gullies. Wet loess has much lower stability than dry loess, because of the partly dissolved $CaCO_3$ content. At the end of a rainy period the uppermost section of loess profiles could be wetted to considerable depths. Such wet banks of a permanent deep gully often cave in and create a temporary dam in the gully floor. The water impounded behind such a dam infiltrates into the loess and creates pipes (Fig. 12.4) and tunnels running relatively deep underground.

The features mentioned are widespread in the loess covered areas of Hungary. They are common but present particular hazards only in the famous wine growing areas, such as Tokaj (see Chap. 25), or as an accompanying phenomenon to huge landslides (see Chap. 14) along river bluffs. Pipes under arable land, pastures or woodland cannot be considered serious hazards. Piping features are very dynamic phenomena, but little is known about the temporal evolution of loess tunnels. They are usually thought to be ephemeral landforms with lifetimes shorter than a year, although Verachtert et al. (2010) present data on loess pipes with 5–10 years' lifetime.

**Fig. 12.4** Secondary pipe developed due to side-wall slumps in a permanent gully at Somogybabod. Runoff in the gully infiltrates behind the dam. Note the collapsed roof at the bottom right corner (photo by Gergely Jakab)

Although it is undoubtedly true that the most impressive and rapidly evolving features are cut into loess, for recent geomorphic evolution other unconsolidated rocks are also relevant. Although rhyolite tuffs are less common in Hungary, appearing in landscapes with high relief energy, they can be affected by intensive piping (Horváth et al. 2010), gullying or even badland formation (see Chap. 18). Gullies also appear on solid rock surfaces. The gullies and ravines on hard rocks, however, are mostly overgrown by trees and, therefore, are virtually stable elements of the contemporary landscape.

## 12.4 The Somogybabod Gully

Piping in loess could extremely accelerate gully development—as happened in the case of the Somogybabod gully in early summer of 2010 after a wet winter and spring. This huge gully of 600 m length, 10 m deep and almost equally wide at the most impressive part, is cut through a road and passed by a house within 50 m distance.

The composite Somogybabod gully presents three separate types of gully morphology. The uppermost section crosses an arable field on a ridge. Here the depth does not exceed 0.5 m and, consequently, it is regarded an ephemeral gully. The lowermost section, on a valley floor of moderate slope just before the fan, has become well-known for its significant depth and considerable volume. Between these two sections there used to be a dirt road and the gully is the steepest here. This part has varied morphology. Locally it is just a few cm deep and divided into several rills, but large headcuts and vertical walls are also observed.

The steepest upper part of the catena is forested to regulate soil erosion, but both the valley bottom and the ridge are intensively cultivated (Fig. 12.5). Huge amounts of runoff water generated on the crusted surfaces of arable fields on the ridge reach the steepest part of the slope and create hollow roads, gullies and ravines even under the forest. An additional reason for the extremely high runoff volume is the near-surface occurrence of plough pan and paleosol (Fig. 12.4)—both much less permeable than the loess or the recently tilled soil.

The middle section of the gully was formed along a dirt road which runs exactly downslope (Fig. 12.5). This road is already the third in the same place since the mid-20th century. Because of the steep slope and the lack of ditches runoff cuts rills into the road surface. To ensure safe traffic on these roads they have to be flattened time after time. Hollow road formation could be very fast due to artificial cutting. Once a road is deep enough or so much dissected that it cannot be flattened easily, it is abandoned and replaced by a new track parallel to the former.


**Fig. 12.5** Location and parts of the Somogybabod gully (Base map: Google Earth)

The spectacular, deep and wide lower section was formed at the bottom of the steepest slope. Most visitors believe that the gully ends at this lower headcut.

Local inhabitants reported that the lower part of the gully had been already present at a smaller scale several years before but was filled up with scraps and covered by loess. The porosity of the scraps was much higher than its surroundings which attracted surface and subsurface runoff into this filled section. This high volume of water presumably triggered piping which contributed to the size of the gully substantially.

Only a small amount of deposition originates from soils. The sediment load entering the gully will be delivered to the fan at the very end. Along the higher (ephemeral gully) section most of the soil loss is fertile topsoil, while along the lower sections the gully cuts into the parent material and detaches loess particles. The main source of load is not gully incision but mass movements affecting the wet vertical sidewalls. Consequently, the fan is mostly built of loess and very little humus.

The Somogybabod gully terminated at a dirt road, which served as a dam and prevented the fan from developing its usual morphology. The road conducted the sediment load to the surrounding arable field, where the crops were covered by reworked loess (Fig. 12.5). It is typical for the region that the soil parent material is exposed at the surface, but sometimes not as a Regosol profile since under the redeposited loess a whole in situ soil profile is found.

Fortunately, the sediment load of the Somogybabod gully was deposited on the valley bottom and did not reach the recipient Tetves stream, where the neighbouring gullies often deliver materials (mainly loess), causing agricultural and environmental problems. According to the inhabitants, rapid runoff is an imminent threat to their buildings. Therefore, they try to stop and regulate runoff water at the fans of the neighbouring gullies through makeshift earthen and wooden dams (built of tree branches) and waterways. These facilities are only able to withstand minor events, while heavy runoff destroys them time after time.

## 12.5 Piping Hazard and Its Mitigation

The danger of piping lies in the fact that once a pipe is formed, it can increase its dimensions without any recognizable sign on the surface. When the thickness of the sediment or soil above the tunnel reaches a critical value, it collapses, creating a huge gully often within moments (Fig. 12.6). From that time on the new gully transports an increased amount of water because it traps additional surface runoff. The average length of pipes in the Central European loess areas does not exceed 10 m. Even if detailed surveys only reveal relatively short tunnels (Bíl and Kubecek 2012), the amount of the entrained and transported sediment could be at least of the same order of magnitude as the soil loss due to gully erosion (Verachtert et al. 2011).

In Hungary black locust *(Robinia pseudoacacia)* trees are widely used to inhibit further erosion on steep and already eroded loess slopes. This type of tree is quite modest in its habitat requirements, but grows very fast and its roots can protect the tunnel roofs against collapse (Lukic et al. 2009). Accordingly, tunnels under forests could be more stable than beneath ground used as pasture or arable land.

In general, soils protect the underlying loess against the destructive effect of rain and meltwater. But because of the high slope angle and the intensive cultivation on the most endangered sites, the soil had been removed by decades of intensive tillage. These areas are mainly used as orchards

**Fig. 12.6** Collapsed tunnel in heavily eroded loess on the border of an intensively tilled arable field. Note the grassed ephemeral gully in maize in the background (photo by Tamás Huszár)

and vineyards now. Therefore, they are very prone to rill and gully erosion, even if land levelling or terracing were employed (Fábián et al. 2006).

Piping often takes place on terraces or other manmade surfaces built on loess (Újvári et al. 2009). During construction the natural loess structure is disturbed causing significant spatial differences in infiltration capacity. If the terrace tread has no sufficient slope to conduct water in the original slope direction, water can be stored on its surface (in arid landscapes counterslope terrace treads of increasing water conservation efficiency were constructed.) On loess the impounded rainwater often leads to the formation of sinkholes, pipes and tunnels (Csorba 2010). The biggest pipe networks or even loess caves tend to form in the vicinity of natural bluffs along rivers (Lukić et al. 2009), at manmade supporting walls or terrace risers.

The typical cross-section of a pipe can vary both in shape and size, even within a short distance (Holden and Burt 2002). Under temperate climatic conditions the average cross-section area is around 0.3 $m^2$ and of a circular shape (Verachtert et al. 2010). Moving from humid temperate to more continental climates, their diameter increases and under semiarid conditions it can reach several metres (Bryan and Yair 1982; Bryan and Jones 1997). Although Hungary is in the temperate zone, Kerényi (1994) reports pipes with 1 m diameter from Tokaj, but there is evidence of even larger pipes from the western (wetter) part of the country (Fig. 12.7). The triangular pipe cross-section is unusual and could be the result of an intensive or long-lasting development, when repeated flow in the tunnel had adequate time to deepen the floor, meanwhile the transported volume of water increased.

The most important mode of pipe destruction is roof collapse, which turns the pipe into a gully. The other possibility of destruction is sealing, when pipe sections are filled up with sediments.

Once a pipe developed into a gully, it cannot be expected to fill up naturally. Over geological time scales, it may be buried, but it would take several thousands of years. Accordingly, if the gully endangers artificial facilities or inhibits effective land management, it should be filled up. This is quite an expensive procedure. It is estimated that the Somogybabod gully needs €28,300 to be eliminated, but gully levelling alone cannot solve the problem. Improper

**Fig. 12.7** Tunnel with triangular cross-section, ca 1 m across, in the Outer Somogy Hills, probably formed on the interface between soil and loess, with soil recently collapsed from the roof in the middle. Note the traces of dissolution and suspended colloid flow on the wall (photo by Gergely Jakab)

tillage on slopes generates a huge amount of runoff, which contributes to gully formation even under forest. The best practice would be the inhibition of runoff through precision/conservation tillage or afforestation. A more rational but less effective alternative solution is hydrological planning and drainage system construction. If surface flow is concentrated and conducted to the valley bottom under controlled circumstances, no rill or gully erosion occurs. The construction of such a system is rather expensive—not to mention the necessary continuous maintenance. Preferably, the construction and operation of drainage network should be implemented jointly by land owners and the municipality with clear rights and task sharing.

## 12.6 Conclusions

Piping is often associated with gullying and promotes gully development. On the other hand, gullies are rather persistent landforms, which can only be filled up naturally on geological time-scales. The Somogybabod gully had impressive volume and shape but it was just one among several others in the vicinity. It has become famous because in this case soil erosion appeared within the village and affected everyday life. Although the total length of the gully exceeds 600 m, just the lower impressive section has emerged as a tourist attraction. Both piping and gully erosion at the Somogybabod gully detach and deliver loess, while the humus and nutrient-rich topsoil are much less endangered, consequently the fan is also built up of loess. Almost the entire sediment load has been deposited on an arable field at the end of the gully, creating a large loess-covered patch on the valley floor. It looks like Regosol, however, the whole in situ soil profile is buried under the loess cover. Accordingly, gully erosion reduces the productivity of agriculture not only by removing fertile soil and dissecting the field, but through burying the plants and topsoil by loess. Gully levelling alone cannot solve the problem at Somogybabod. A long-term solution is hydrological planning and drainage system construction. If surface runoff is concentrated and conducted to the valley bottom under controlled circumstances, no rill or gully erosion would occur.

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# Kapos Valley: The Most Asymmetrical in Hungary

13

Dénes Lóczy, Ervin Pirkhoffer, and Péter Gyenizse

**Abstract**

Neotectonic movements often result in marked large-scale asymmetry of river valleys. The characteristically rectangular ("chequerboard") drainage pattern of Southern Trandanubia has provoked discussions among Hungarian geologists and geomorphologists for more than a century. In the centre of this area is the Kapos River valley, the asymmetry of which is obvious to any visitor. Although this asymmetry has been undoubtedly controlled by neotectonic movements to a large extent, it is probably also enhanced by geomorphic processes. The degree and distribution of asymmetry is quantified using different morphometric indices calculated from a Digital Elevation Model. An overview of the tectonic models proposed to identify the reasons behind asymmetry in the study area is also presented. It is pointed out that the asymmetry is associated with the alignment of two tectonic lines crossing the catchment and also manifested in the planform of the valley floor (i.e. in the shape of the morphological floodplain).

**Keywords**

Floodplain • Neotectonics • Valley asymmetry • Valley confinement • Asymmetry indices • Undercutting • Transdanubian Hills

## 13.1 Introduction

Travelling from Budapest to Pécs by train along the middle and lower section of the Kapos River Valley, a passenger observes a marked slope asymmetry: the hillslopes on the right-bank side are consistently much steeper than those on the left bank (Fig. 13.1). Here the river flows on the boundary between two loess-mantled hill landscapes.

The Kapos is a medium-size river by Hungarian standard (length: 112.7 km, catchment area calculated from DEM: 3,259 km$^2$) (Lóczy 2013). It is a tributary of the Sió Canal, the outflow of Lake Balaton to the Danube, which allows the regulation the water level of this shallow lake, strongly influenced by the actual weather conditions. The topographical/morphological floodplain (the flat area bordered by terrace margins and hillslopes where fluvial landforms can be detected, without the floodplains of tributaries) extends over 104.2 km$^2$ and makes up merely 3.3 % of the total catchment area. The source of the Kapos is at ca 180 m above sea level, and it is a junction of two headwaters of 2 km length each, now impounded to form a series of fishponds. The river is fed by 27 right-bank tributaries and 28 left-bank tributaries of relatively permanent flow. Among the tributaries there is only a single (left-bank) fourth-order

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**Fig. 13.1** The steep right-bank slope of the asymmetrical Kapos valley with inundated floodplain at Döbrököz (photo by Dénes Lóczy)

**Table 13.1** Topographic data for the catchments of the Kapos River and its major tributaries (derived from 50-m resolution DEM)

| Catchment | Kapos | Koppány (left-bank) | Orci stream (left-bank) | Surján stream (right-bank) | Baranya canal (right-bank) |
|---|---|---|---|---|---|
| Lowest point (m) | **100.5** | 103.3 | 121.3 | 122.8 | 114.3 |
| Highest point (m) | 583.2 | 311.1 | 274.2 | 280.8 | **583.2** |
| Relative relief (m) | **482.7** | 207.8 | 152.9 | 157.9 | 468.9 |
| Mean elevation (m) | 176.4 | 176.8 | 167.7 | 191.1 | **209.9** |
| Steepest slope (°) | **27.1** | 15.4 | 10.5 | 16.3 | 24.9 |
| Mean slope (°) | 3.5 | 3.5 | 2.7 | **6.0** | 5.6 |
| Drainage density (km km$^{-2}$) | 0.75 | 0.69 | 0.60 | 0.74 | **1.09** |

**Bold** values show extreme high or low values

tributary (the Koppány River), one third-order stream on the left bank and two on the right (Table 13.1).

## 13.2 The Kapos Catchment

The Kapos drains large areas of the Transdanubian Hills, one of the major geomorphological regions of Hungary (see Chap. 5 in this volume), including the Outer Somogy Hills (maximum elevation: 300 m) on the left bank and the hilly regions of the Northern Zselic (358 m), Baranya (321 m) and Tolna Hills (285 m) on the right bank. Only a smaller portion of the catchment (ca 5.9 km$^2$) is constituted by the northern slopes of the Mecsek Mountains (648 m) (Fig. 13.2), which are mostly built up of Mesozoic-Tertiary calcareous sediments, locally with some old volcanics.

Valley density is highest in the Mecsek segment of the Kapos catchment, but its values are also high in the Zselic Hills and along the Koppány River (8–10 valleys km$^{-2}$) (Table 13.1). Interfluvial ridges, rising to 300 m in the north, are usually not wider than 150–200 m. The central part of the catchment is a flat loess-mantled plateau of ca 150 m elevation, dissected by broad and bowl-shaped dry valleys (called “derasional” valleys in Hungary). Even the statistics of the topographic variations (elevation and slope angles) show the contrast between sub-catchments (Table 13.1).

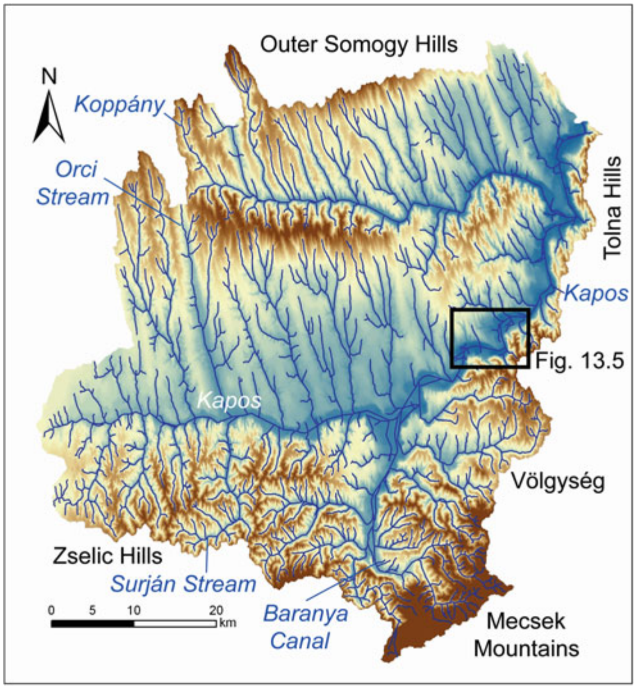


**Fig. 13.2** Topography of the Kapos River catchment with segment covered in Fig. 13.5 (by Ervin Pirkhoffer and Dénes Lóczy)

## 13.3 Valley Asymmetry

Asymmetrical valleys are not at all uncommon landforms—in fact, river valleys with perfectly symmetrical walls are rather the exceptions than the rule. As recognized in international research at early date (Kennedy 2004), asymmetrical valleys can result from various processes. Moving from large-scale or valley-wide to smaller scale or localized phenomena the following reasons of asymmetry may be recognized:

(a) Uneven uplift during neotectonic movements leads to asymmetry (Bull and McFadden 1977; Burbank and Anderson 2001).
(b) Glacier action may truncate one side of the valley.
(c) Periglacial processes active at different rates on slopes of various exposure (e.g. Demek 1968; French 2007).
(d) Landslides change valley cross-sections dramatically but usually only locally.
(e) Bank undercutting by meandering rivers also causes local asymmetry—although the effectiveness of the Coriolis force, also mentioned by Kennedy (2004), is open to debate.
(f) At a smaller scale even differential rock weathering (in combination with aspect/microclimate and the induced slope processes) may generate asymmetry (e.g. Ollier and Thomasson 1957).

In the mountains and hills of Hungary small-scale asymmetric topography, i.e. gentle southern and southwestern as opposed to steeper northern and northeastern hillslopes, were first described in the 1960s and primarily explained by frost processes (such as solifluction, cryoplanation) driven by microclimate (both temperature and moisture) variations during the Pleistocene (Pécsi 1966).

## 13.4 Basin and Valley Asymmetry Indices

In the international literature morphometric indices are used to provide quantitative evidence of the asymmetric character of catchments and valleys. This approach is particularly useful in reconnaissance surveys of tectonically heavily affected, remote mountainous regions, such as the Himalaya (Dar et al. 2013). This is not the only way of application, however, as recent tectonic deformations were also detected by morphometric methods in the relatively well-studied Carpathian (Pannonian) Basin (Pintér 2005).

In this chapter the Digital Elevation Model of the Hungarian Institute for Remote Sensing and Photogrammetry (FÖMI) (horizontal resolution: 10 m) was applied to extract the required index values with the help of ArcGIS 9.2 Spatial Analyst extension.

### 13.4.1 Asymmetry Factor (AF)

The Asymmetry Factor is often interpreted as a an indication of (tectonic) tilting of a drainage basin as a whole with respect to the main channel (Cox 1994; Raj 2012—Fig. 13.3a):

$$AF = 100(A_r/A_t),$$

where $A_r$ is right-bank drainage basin area and $A_t$ is total drainage area.

This index does not only reflect overall tilting, but also the uplift/subsidence of individual blocks resulting in valley asymmetry (Pintér 2005). AF values above 50 % the show that the main river channel is shifted to the left, whereas values below 50 % indicate channel shift to the right side.

### 13.4.2 Transverse Topography Asymmetry (T)

Transverse Topography Asymmetry (T) also indicates the tilting of river basins and the wandering of river channels away from the symmetry axis of the catchment (Cox et al. 2001—Fig. 13.3b). The shifting of rivers is analyzed for consecutive 2-km segments for third- and fourth-order

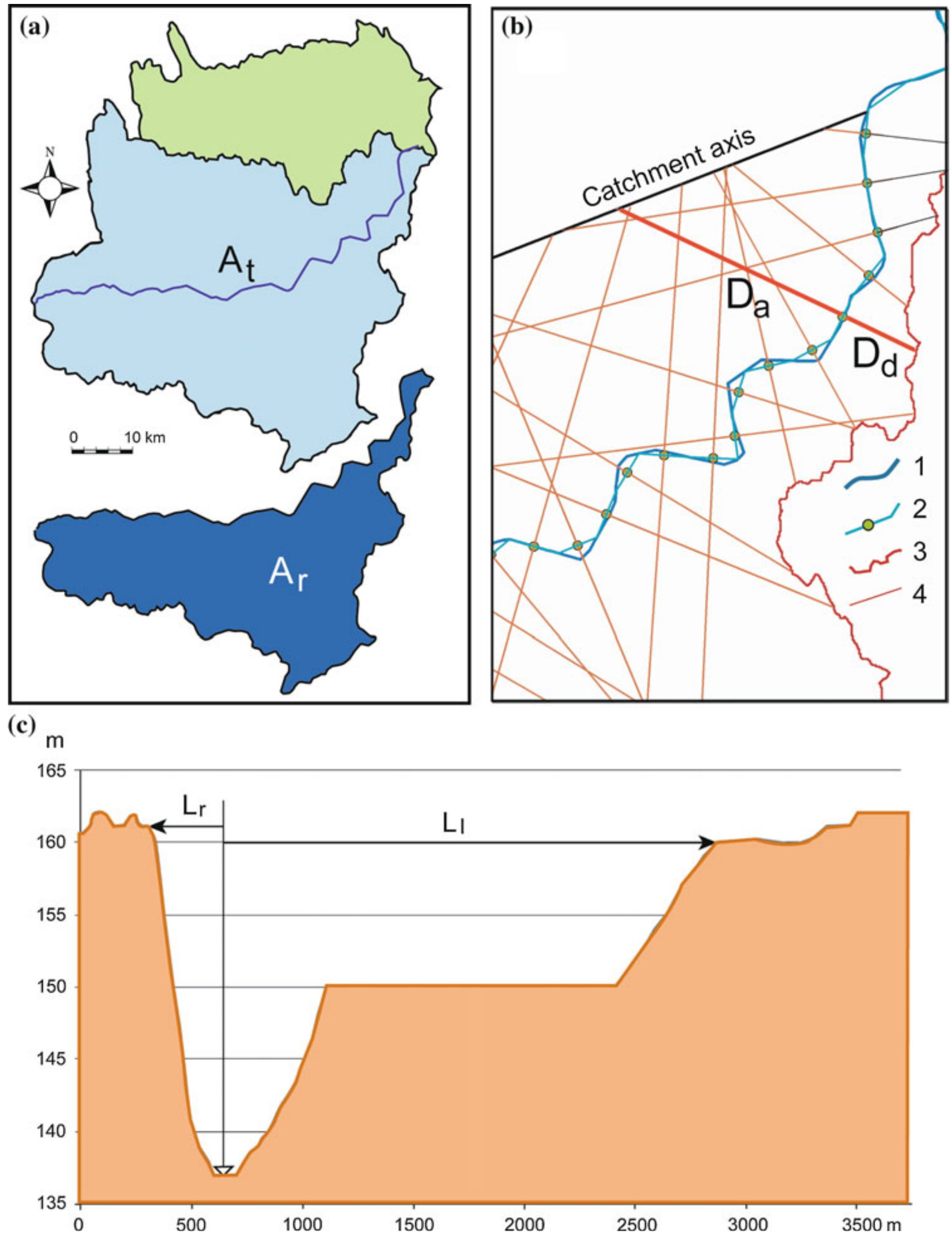


**Fig. 13.3** Comparison of the concepts of asymmetry indices. **a** AF index; $A_t$ Total catchment area, $A_r$ Right-bank catchment area. **b** T index; *1* Kapos channel, *2* Straight segment with centre point, *3* Watershed/divide, *4* Line perpendicular to straight segment. **c** DVM index. For calculation methods see the text

stream. For a regulated river, the straight segments are conceived as the axes of the pre-channelization meandering-anastomosing channel systems. Lines perpendicular to the segments are applied to measure the distances necessary for the calculation of the T index according to the formula

$$T = D_a/D_d,$$

where $D_a$ is the distance from the centreline of the drainage basin to the centreline of the channel (i.e. the axis of the channel systems) and $D_d$ is the distance from the centreline to the watershed/divide (Cox et al. 2001). T = 0 indicates perfect symmetry, while higher T values, approaching 1, point to increasing asymmetry. The azimuth angles of the lines can also be established and interpreted as vector fields. This approach is best suited for dendritic stream networks, particularly in loess-mantled areas, where stream segments have a full 360° range of orientations (azimuths). Identifying regional types of vector fields (as done by Cox et al. 2001 for the Mississippi Embayment), interpretations for and subdivisions of the study area by neotectonic characteristics become possible.

### 13.4.3 Distance to Valley Margins (DVM)

For the calculation of the DVM index (Fig. 13.3c) cross-sections are analysed which are constructed along the same lines drawn to establish the values of the T index. On the cross-sections the points of highest local elevation are identified on both banks and these are assumed to represent the valley margins. The map distance from the channel centreline (axis of the segment) to the first point with local maximum elevation on the left bank ($L_l$) is divided by the distance to the first point with local maximum elevation on the right bank ($L_r$) and hence provides the value for the DVM index:

$$\mathrm{DVM} = L_l/L_r.$$

Calculating the index values for each 2-km segment, valley asymmetry can be followed all along the course of the river, and the segment with maximum asymmetry can be identified. Another advantage of this index is that it is concerned with the elevation of valley sides and not influenced by the slope angles of valley sides, which are locally modified by slumps.

## 13.5 Morphometry of the Kapos Valley

The asymmetry of neotectonic origin is visibly manifested in slope inclination: slopes on the northern valley side with Pleistocene terraces are typically 2.5–3.5°, only exceptionally exceeding 15°, while slopes of 15–25° angle also occur on the southern valley side. Channel shifting to the right has steepened the southern valley margin before channelization. Asymmetry is also reflected in the shapes of tributary catchments in the north and south (Lóczy 2013).

To avoid a major bias when calculating asymmetry, it seems advisable to examine only the portion of the Kapos catchment upstream of the confluence with the largest tributary, the Koppány River (Fig. 13.2). This catchment area is 2,514 $km^2$. The AF index supplies a general picture of the asymmetry of the catchment. Since the right-bank catchment area upstream of the Koppány confluence is 1,224 $km^2$, AF = 52.4 %. This value is very close to 50 % and indicates no significant asymmetry at (sub)catchment scale. The T index, reflecting local valley asymmetry, however, shows gradual increase from the uppermost reaches to the point, where the valley turns in northeastern direction. Plotting the T and DVM index values against distance downstream, three segments of the valley can be distinguished using the DVM index (in good agreement with the typology of floodplain segments—Lóczy 2013). In the uppermost segment (0–35 km), although there are high peaks, the valley being so narrow, the asymmetry is not remarkable. The middle segment of the valley is symmetric (35–55 km), while downstream, at 55 km distance from source, asymmetry shows again maxima in the gaps, this time coupled with high AF and T index values (Fig. 13.4). The azimuths of the lines along which the T index is measured almost exclusively fall within 135 and 225° with a maximum at 180°. This corresponds to the general W to E alignment of the Kapos Valley. The deviation of the values, however, is very considerable and is also maximal on the lower segment (Fig. 13.4).

## 13.6 Neotectonic Explanations—Old and New

The drainage of the Outer Somogy Hills shows a marked chequerboard pattern (Fig. 13.6), which was previously directly attributed to geological structure (Szilárd 1967).

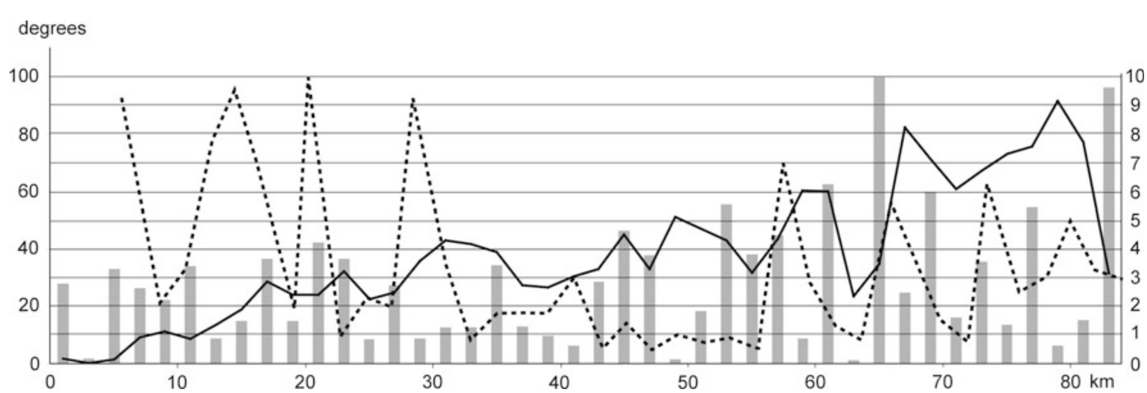


**Fig. 13.4** 10 × T index (*solid line*) and DVM index (*dashed line*) values for the Upper and Middle Kapos Valley. Deviations between the azimuth angles (*degrees*) of neighbouring lines used for the calculation of the T index (columns) are also shown by *bars*. Distances are from source

Over the past century there have been several explanations proposed for the regional structure of Southern Transdanubia (Magyari et al. 2005):

1. At first it was suggested that the units of the Neogene basement, called "tables" reached their present-day positions through normal faulting, vertical movements and tilting towards the south (Cholnoky 1918).
2. Others believed that the modern relief is primarily due to east to west progressing folding (Pávai Vajna 1926).
3. Still before the birth of the plate tectonic theory the concept was formulated that the topography results from upthrusts caused by NW-SE directed horizontal compression (Erdélyi 1961, 1962). Erdélyi enumerated pieces of evidence for Quaternary movements from the positions of various Tertiary and Quaternary deposits and also interpreted the location of divides within the valley floor, fairly common in Southern Transdanubia, as evidence for recent uplift.
4. In addition to sedimentological investigations, the hypothesis of east to west dextral strike-slip faulting in the Outer Somogy Hills was also based on gravity and magnetic anomalies (Némedi-Varga 1977).
5. Recently detailed structural morphological investigations were launched in Southern Transdanubia to distinguish between the manifestations of tectonic and gravitational movements (Csontos et al. 2005).

Following a very complicated Quaternary tectonic evolution (Magyari et al. 2005) with remarkable offsets (up to hundreds of metres), extension only locally occurs today and the predominant compression tectonics results in uplift over most of Transdanubia. Through the the identification of photolineaments from remote sensing sources it was found that the normal faults generated by extension are rejuvenated but now operate as upthrust planes (Síkhegyi 2008), constituting sections of the partially still active Mid-Hungarian Fault Zone (see Chap. 2). Even in the lack of detailed stress field reconstructions, Magyari et al. (2005, p. 60) claim that "the young/sub-recent age of most structures suggests that the Somogy Hills is the scene of active compressional and strike-slip type neotectonic processes", although whether these are dextral or sinistral movements, it is still debated (Bada et al. 2010). The overwhelming part of the Kapos catchment is affected by uplift (Joó 1985, 1998). Consequently, the valley is deepening in a relative sense and relative relief and stream power in the hills is increasing.

There is also seismological evidence for neotectonic activity. The traces of seismic events in the Tertiary and Quaternary can often be found as seismites in the exposures of the Outer Somogy Hills (Magyari et al. 2005; Csontos et al. 2005). In Hungary, the hypocentres of earthquakes lie usually less than 10–12 km below the surface (Tóth et al. 2008), which is explained by the thin lithosphere of the Pannonian Basin. Such shallow earthquakes suggest a direct relationship between tremors and geomorphic processes (Síkhegyi 2008). In historical times the epicentres of 16 earthquakes were located in the 5 km buffer zone of the Kapos River, with eight tremors of less than 3 M on the Richter scale since 1994 (the starting date of seismological monitoring in connection with the operation of the Paks Nuclear Plant) (Tóth et al. 2008). Also here the epicentres are not found directly above the main fault but shifted to the north. In Síkhegyi's opinion: "they are the surface projections of subsurface flower structures" (Síkhegyi 2008, p. 43).

## 13.7 Planform Floodplain Asymmetry

For the asymmetric valleys of tectonic origin in Southern Transdanubia the more or less regular alternation of constrictions (gaps) and relatively broad valley floor segments (embayments) is characteristic (Fig. 13.5). Although this planform feature is rather striking, previously no author could explain it in a satisfactory manner. According to Erdélyi (1962), embayments developed at the subsiding SSE corners of structural units, while at their NNW corners gaps were generated. A series of stream captures were assumed to be necessary to connect the embayment segments into a contiguous valley. It was suggested (Síkhegyi 2008) that lateral compression induced shears and horizontal displacements with rotation as well as compression and folding are jointly responsible for this 'sawteeth' pattern (Fig. 13.5). This pattern, however, is only typical for a shear zone some kilometre wide along the main fault-line, approximately to the village of Kurd. Downstream of Kurd the Kapos Valley turns into northeastern direction and then to the north. This

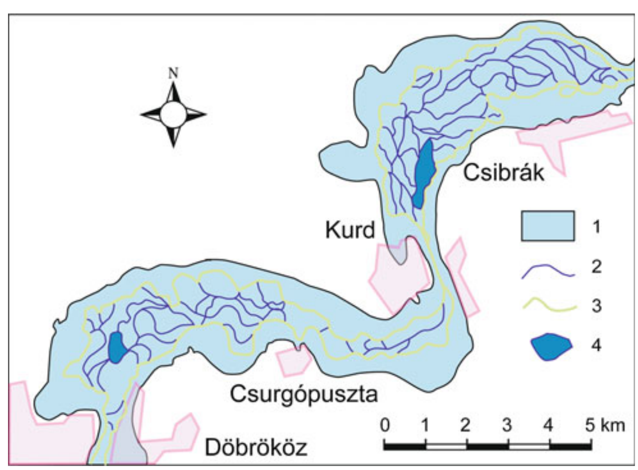


**Fig. 13.5** A segment of the reconstructed Kapos morphological floodplain with paleochannels, showing the characteristic "sawteeth" pattern. *1* Floodplain, *2* Traces of paleochannels, *3* Outermost paleochannel, *4* Previous lakes (mill-ponds) (by Péter Gyenizse)

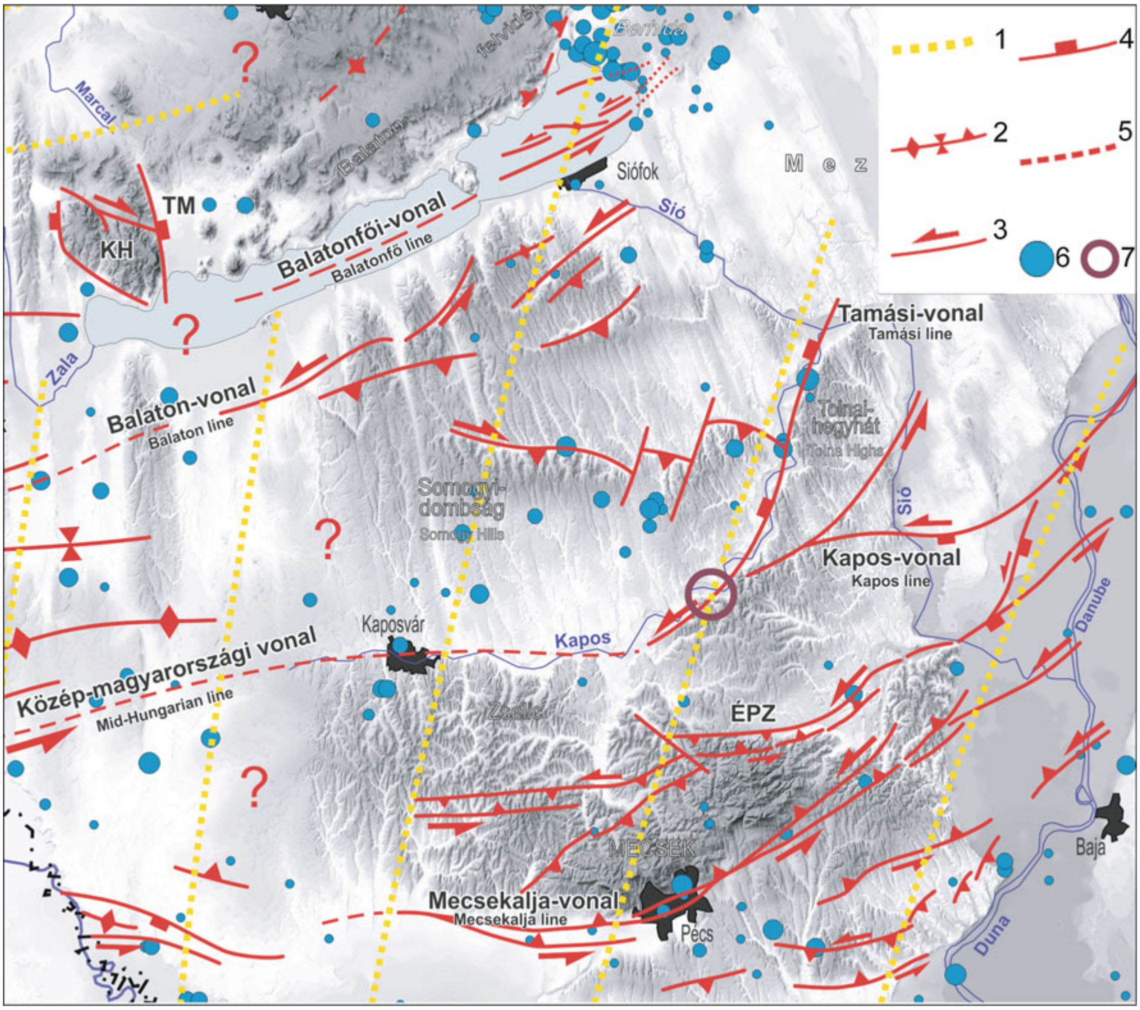


**Fig. 13.6** Map of structural geology of the Kapos catchment and environs (after Bada et al. 2010). *1* Maximal horizontal stress trajectories, *2* Folds and reverse faults, *3* Strike-slip faults, *4* Normal faults, *5* Uncertain faults, *6* Epicentres of earthquakes (in proportion to magnitude), *7* Fault "hinge" area

change in direction can be caused by the compression along the Tamási Line (Síkhegyi 2008), which is probably responsible for the recent uplift of the Tolna Hills on the eastern bank of the river (Fig. 13.6). The distances between gaps show a growing trend along the river (Lóczy 2013)—an observation that has not yet been explained either up to date.

As also mentioned by Erdélyi (1962), the strip of most intensive uplift does not coincide with the steep slopes of the hill ridge south of the Kapos Valley. He assumed that topography has been heavily modified by short run-out slumps and local channel shifts. In this strip the meadering stream pattern in tributary valleys becomes braided or anastomosing where they enter the Kapos floodplain. River action has also substantially affected the planform shape of the floodplain. Hillslopes undercut by channel wandering can easily be identified along both floodplain margins even today (Lóczy 2013).

As a consequence of the particularly marked valley asymmetry, the northern valley sides are usually free of true river terraces, only low escarpments without deposits are found. The alluvial fans of tributary valleys are attached to these surfaces. The north-facing slopes show a more varied microtopography with slump heaps, gullies and ravines. Relatively steep concave or straight footslopes are interrupted by the flat alluvial fans of major tributary streams and aprons of loess washed down and redeposited from the neighbouring hillslopes.

## 13.8 Conclusions

The principal factor behind the asymmetry of the Kapos valley is obviously neotectonic movements, locally exacerbated by river undercutting and slumping, often operating in combination. The morphometric indices do not confirm basin asymmetry (AF index close to 50 %), but underline valley asymmetry. The T index shows gradual increase downstream. The index called Distance to Valley Margins best indicates the most marked asymmetry between valley sides at and somewhat downstream the Kurd gap, which approximately coincides with the "hinge" of two main tectonic lines, the W-E Kapos Line and the SW-NE Tamási Line.

All the three morphometric indices applied justify the claim of the Kapos valley to be extremely asymmetric. Further investigations are needed, however, to reveal the exact forces and mechanisms of Plio-Pleistocene neotectonic movements behind and their control on the observed special morphometric properties.

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# Dunaszekcső Castle Hill: Landslides Along the Danubian Loess Bluff

14

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**Abstract**

Mass movements along the Danube banks are quite frequent events in the landform evolution of Hungary. A particular combination of geological, geomorphological, climatic and hydrological factors induces failures on loess bluffs. Clay layers of buried paleosols, steep slopes of loess bluff, water level fluctuations of the Danube and extreme precipitation events control the occurrence of landslides. The lateral erosion of the Danube has destroyed the edge of the Mezőföld loess plain and Tolna–Baranya Hills since the river began to shift westwards from its former course across the Danube–Tisza Interfluve ca 40 ka ago. Remnants of the Ancient Roman fortresses along the eastern frontier of the Empire indicate the time scale of bank retreat: ca 2–2.5 m per 100 years. The recorded dimensions of historical landslides reach 20–30 m in width and 100–1,000 m in length along the Danube. The largest landslide of the past decade occurred at Dunaszekcső Castle Hill in 2008 and in 2011 it endangered the village again. The lack of precise archeological reconstruction of the groundplan of *Castellum Lugio* encumbers to estimate the rate of bluff retreat. According to the latest studies Danubian floods dominantly control recent movements. The continuous monitoring of the landslide provides valuable information on the dynamics of such movements.

**Keywords**

High bluff • Landslide • Loess • River action • Geoarchaeology • Danube

## 14.1 Introduction

Mass movements are one of the most important and hazardous geomorphic processes to occur recently in Hungary. However, if compared with other recent processes, they have a localized impact on our environment and present a geomorphic hazard in limited areas (Szabó 1996). Among different mass movements landslides characterize the river banks in Hungary along the Hernád, Rába and Danube. Landslides have been triggered along ten sections on the right bank of the Danube, four of them located upstream and six downstream of Budapest (Fig. 14.1A). South of Budapest, a 20–60 m high loess bluff occurs that formed in the Late Pleistocene and Holocene due to river undercutting (Pécsi 1994; Lóczy et al. 2008). An infamous 1,300 m long landslide occurred in February 1964 at Dunaújváros, when a new city and a new steel works were hazarded. In the last fifty years, landslides were recurring in different sections along the Danube. Since 2006 mass movements happened regularly along the southern section of Danubian bluffs. At Dunaszekcső up to 20 % of the Castle Hill was destroyed (Újvári et al. 2009; Bugya et al. 2011). More than 20 % of this area is affected by recurrent landslides observed in spring of 2011,

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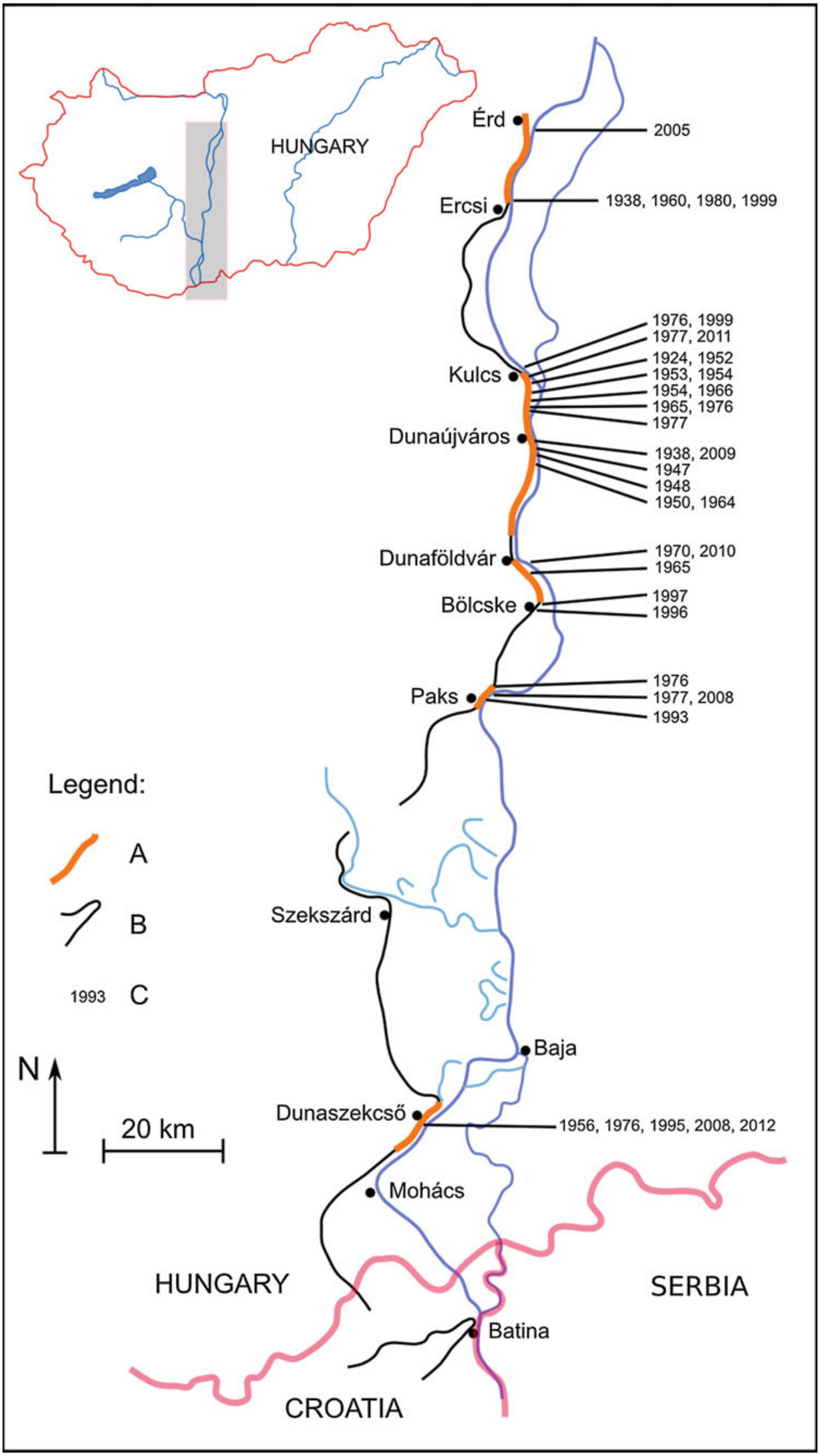


**Fig. 14.1** Bluffs along the Danube: active bluffs close to Danube (*A*), inactive bluffs away from river (*B*) and years of movements (*C*) (after Kleb and Schweitzer 2001, supplemented with new data)

when up to 40 cm vertical displacement, indicating an initial phase of a new movement, was measured by different geodetic and GPS methods (Fábián 2012; Bányai et al. 2013).

## 14.2 Geological Setting

Generally, one of the most important factors in landslide occurrence is bedrock lithology. Therefore, it is useful to review the last six million years of sedimentation in the wider surroundings.

The main subsidence period in the Middle Danube region was in the Late Miocene and Pliocene. Thickness of the lacustrine sediments (clay, marl and sand) is only half of the Hungarian average, ca 500 m. After the retreat of Lake Pannon, lacustrine sediments were eroded and dismembered by tectonic lines (Jámbor and Rálischné 2002). In the beginning of the Pleistocene it may have been a mildly differentiated surface slightly dipping south- and southeastward. Thus, the two main lithostratigraphic units of Pleistocene terrestrial deposits, discordant on the Pliocene surface of erosion, are thicker in the southern and southeastern part. The age of the older, variegated unit between the loess and the Late Miocene (Pannonian) deposits is around the Plio-Pleistocene boundary (ca 2.58 Ma). These old clay and clayey deposits have been described from Dunaföldvár, Tengelic, Paks and Szekszárd (Pécsi 1985; Schweitzer and Szöőr 1997; Kovács et al. 2011). According to Jámbor and Rálischné (2002) the Tengelic formation, Dunaföldvár series and Paks red clays, ($Pv_{1\text{-}5}$) belong to this unit.

Above the red clays a typical loess sequence was recorded, including paleosols and sand interbedding. Based upon their lithological properties and stratigraphic position, two main series of loess-paleosol sequences could be determined in Hungary (Pécsi 1993). The older and lower part of this sequence was named after the stratotype section in the Paks brickyard. The Paks series contains six loess and five paleosol horizons, although it has several erosional hiatuses. Loess from this series contains less $CaCO_3$ but it is strongly compacted with intercalated fluvial sands and reddish brown paleosols. The young loess series of Mende-Basaharc and Dunaújváros-Tápiósüly is sandy and four marker paleosols could easily be identified (Pécsi 1995). In spite of 60–70 years of research the chronology of loess deposition is still debated (Frechen et al. 1997; Újváry et al. 2010). According to Pécsi (1986), no loess in Hungary is older than 1 Ma and the age of the young series is maximum 110–130 ka. New results from the past decades found the Basaharc Double (BD) and Mende Base (MB) paleosols significantly older than ever thought (Horváth 2001; Thiel et al. 2014). Thus, the development of the young loess series started before MIS 9. Due to aeolian processes during glacial periods, a typical loess plain was formed in the western Great Hungarian Plain.

## 14.3 Landscape Evolution

In addition to the properties and structure of loess, the changes in the course of the Danube in the Last Glacial also affected geomorphic evolution. Erosional valleys dissected the loess surface of the Mezőföld, while gully incision moulded a hilly face for the Tolna-Baranya region. Tectonic uplift and periglacial processes (i.e. solifluction) were also assumed by Ádám (1969) as he explained the origin of the east-facing slopes of this hilly loess region (Fábián et al. 2006).

One of the largest alluvial fans was built by the Paleo-Danube in the middle of the Great Plain (Borsy 1990; Gábris and Nádor 2007). From the Pest Plain, following a NW–SE course, the Danube accumulated fluvial sediments in 600–700 m thickness around Szeged (Pécsi 1959). The river maintained an anastomosing system over the last glacial and postglacial periods controlled by changing climate. The present-day channel pattern is meandering and the river shows the tendency of moderate aggradation (Somogyi 1983).

Three major models have been proposed to explain the change of river courses from the ancient NW–SE to the recent N–S direction. Chiefly tectonic controls and partly hydrological influences and the Coriolis force have been mentioned in Hungarian literature. Recent earthquake hazard investigations indicate tectonic activity (e.g. earthquakes at Dunaharaszti in 1956 M = 5.3 and at Kecskemét in 1911 M = 5.6) along the Danube, focused on the NW–SE transversal faults in the Danube–Tisza Interfluve (Schweitzer et al. 2011). Tectonic movements are recurring from the Last Glacial period, together with the slow subsidence of Baja and Kalocsa–Paks depression in the Late Pleniglacial (Jaskó and Krolopp 1991; Hertelendi et al. 1991). These vertical displacements forced river channels to change their course and move westward. From the hydrological point of view, in the Late Pleistocene and Holocene climate change controlled the discharge of the Danube. In drier periods eastern channels were unable to transport sediment load, while western channels received recharge from the water reservoirs of the karst mountains of Transdanubia. Increasing discharge of the last termination period (i.e. the Pleistocene/Holocene transition) resulted in net incision (Gábris 2006; Gábris and Nádor 2007), reducing the number of anastomosing channels. Thus the Danube abandoned the interfluve, sliding down from its alluvial fan (Somogyi 1961). A few studies (Balla 2009) associate this well known westward shift with the Coriolis force. Although the impact of the Coriolis force on river systems has been debated for decades, the spatial distribution of alluvia along rivers is unequivocal in the Pannonian Basin. The left banks of all significant rivers are characterized by poor runoff and almost totally flat surface waterlogged almost permanently before flood control measures. On the right bank higher relief and bluffs or terraces occur as Karl Ernst Baer proposed it in the 19th century, but the geomorphology and sediment fill of Sárköz and the surroundings demonstrate a more western course of the Danube (Fig. 14.1) in the Holocene (Hertelendi et al. 1991).

All the different theories agree that the date of the direction change can be relatively determined because older terraces are missing from the Danube valley. Only the II/a level and floodplain terraces are identified at Tengelic, Szekszárd and Mohács (Pécsi 1991).

Two sites confirm that the extent of Transdanubian loess area was greater: two hillocks (Solti-halom, 124 m, and Tétel-halom, 112 m) rise from Danubian alluvium attesting to the lateral erosion of the river (Lóczy et al. 1989). This can be studied at Dunaföldvár or Dunaszekcső, where tributary valleys dissected the loess surface nearly parallel to the river and high floods of the Danube inundate the valley bottoms.

## 14.4 The Danubian Bluffs

South of Budapest six active bluff sections are found close to the river: at Érd, Ercsi, Dunaújváros, Dunaföldvár, Paks and Dunaszekcső (Fig. 14.1). At a distance from the river former bluffs are also identified, but there only minor landslides, gullies and sunken lanes characterize recent land formation.

From north to south, an increasing thickness of the loess and loess-like sediments is visible in the relative elevation of the bluffs, rising from 20 to 60 m. Interpreted from their geological and geomorphological settings, the following stability types of high bluffs can be distinguished: (i) banks directly eroded; (ii) banks with an unstable remnant of former landsliding material; (iii) banks protected by foreground deposits and fluvial sediments. In the case of type 1 and 2, the bluffs are unstable and have been considerably modified or endangered by mass movements. Type 3 bluffs are devoid of mass movements. Naturally, in many cases the types listed above are combined, resulting in mixed forms (Karácsonyi and Scheuer 1972; Fábián 2003).

For interpretation of the recurrence of landslides along the Danube, three principal reasons were proposed: tectonic events (i.e. earthquakes), humid periods and water-level fluctuations of the Danube. Only few authors (Bendeffy 1972) cite tectonic events as a trigger of landslides. Generally, landslides mostly occur during late winter or spring in years with precipitation above the average (Juhász 1999). Higher-than-average rainfall events, such as in 2009 and 2010, are particularly important factors in triggering landslides when they are preceded by a prolonged dry period. But along large rivers flood events are also considerable triggering factors and rivers have a more significant impact on groundwater levels than weather. Groundwater fluctuations are especially efficient to generate landslides along the Danube, because higher table reduces shear strength (Kézdi 1980).

Along the Danube, where large landslides have occurred, archaeological evidence can be analyzed to estimate the rate of bluff retreat. In the first centuries AD the frontier (limes) of the Roman Empire ran along the Middle Danube and a series of fortresses (castra) connected by paved roads with watchtowers were built for its defence. Aerial photograph surveys have been made to locate these structures and the evidence was evaluated in a monograph by Visy (2000). The interpretation of aerial photos allowed for the reconstruction of the original groundplans for most of the fortresses and it was found that in some cases major portions of them have been destroyed by landslides related to bluff retreat.

On the basis of the reconstruction of ancient Roman fortresses along the Danube and using historical maps from 18th to 19th centuries, different rates of bluff retreat have been suggested. According to Lóczy et al. (1989), three different rates are possible: (1) an overall rate on a geological time-scale (thousand years: ca 1–2 m per 100 years); (2) for periods of active undercutting (hundred years: ca 2–10 m per 100 years) and (3) a rate influenced by human interference (over the last 100 years: above 10 m per 100 years). Although there is no direct historical evidence for landslides from Roman times, the evaluation of archaeological data provides assistance in the reconstruction of undercutting and bluff retreat.

## 14.5 Dunaszekcső Castle Hill

The settlement is located on the marginal zone between the Great Plain and the Baranya Hills, at the confluence of the Lánka Stream with the Danube (Fig. 14.2). Since the Bronze Age it has been inhabited and the first residents were the Celts. Castle Hill was the most important point of the village from early times, due to its propitious position above the Danube. The Romans built *castellum Lugio* to defend the limes. It has been destroyed and only a manuscript map, a few historical documents and notes testify to its existence. A width of 175 m and a length of 102 m can be estimated for this fortress (Halász 1952; Visy 2000), but the original walls and towers of Lugio were destroyed by landslides. According to archaeologists, the width is acceptable for reconstruction as visible on Fig. 14.3. The flat-topped Castle Hill and its steep slopes and the network of streets were considered as basic features of this reconstruction.

The highest point of the Castle Hill is 142 m above sea level—almost 60 m above the mean water level of the Danube (82–83 m). The instability of the Castle Hill arose from the geology of the Pleistocene loess and subaerial series (ca 100 m thick) containing 23–25 paleosols developed on the Late Miocene lacustrine sediments (Fig. 14.4). In this section the loess series (ca 57 m) is subdivided into 14


**Fig. 14.2** A perspective of landslide affected area on the Dunaszekcső Castle Hill using helicopter aerial photography from east (**a**) (*source* Pazirik AreoArt) and from west (**b**) (*source* Civertan Graphic Studio). *SA* Study area for geodetic survey, *SB* Southern block of the landslide in 2008, *NB* Northern block of the landslide in 2008, *S* Scarp of the landslide in 2008, *FS* Former scarp, *NC* New dilatation crack, *RP* Rotational piling up, *P* Section for cross profile in Fig. 14.4

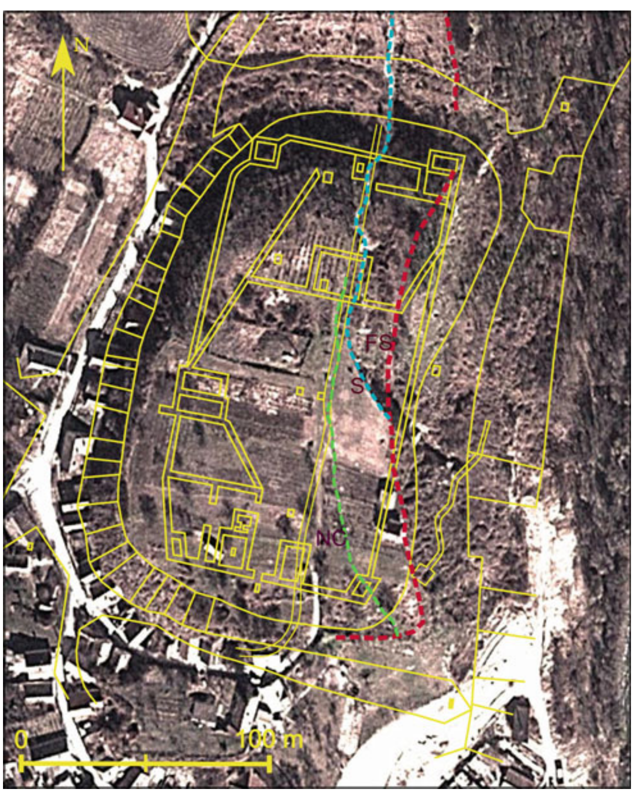


**Fig. 14.3** Supposed location of ancient Roman castellum Lugio (*source* Halász 1952) compared with the topography of the Castle Hill, Dunaszekcső. *S* Scarp of landslide in 2008 (*blue*), *FS* Former scarp (*red*), *NC* New dilatation crack (*green*)

paleosols and 1 sand strata. The underlying series (ca 40 m) is represented by sandy silts and clays and intercalated 12 paleosols and some erosional gaps (Pécsi and Schweitzer 1995). The floodplain of the Danube is very narrow or missing at the Castle Hill. The bluff consists of a 20–30 m high vertical loess wall above the 10–20 m high slopes on reworked loess from past landslides and fluvial silts and sands of the Danube (Fig. 14.4). Although training walls and groynes have been built during regulation measures to shift the current away from bank sections, the river intensively undercuts the slopes during each flood event.

Since 2006, when the first cracks were observed on the Castle Hill a slow vertical displacement characterized the movement. After a while, suddenly a 222 m long landslide occurred on the eastern part of the Castle Hill (Fig. 14.2) on 12 February 2008. The largest width of the sliding blocks amounted to 30 m and 0.3 million $m^3$ mass moved affecting an area of 5,086 $m^2$ (Bugya et al. 2011). Reactivation of this process was detected in spring of 2011 (Fig. 14.5). For monitoring the landslide different methods were established (Fábián 2012) or continued (Bányai et al. 2013). Preliminary results of monitoring show significant trade-offs between loess bluff stability and groundwater table influenced by the Danube floods. Based on the assumed dimensions of Lugio and recent coordinate measurements (Fig. 14.6) at least as rapid rate of bank retreat can be estimated as before the first century AD.

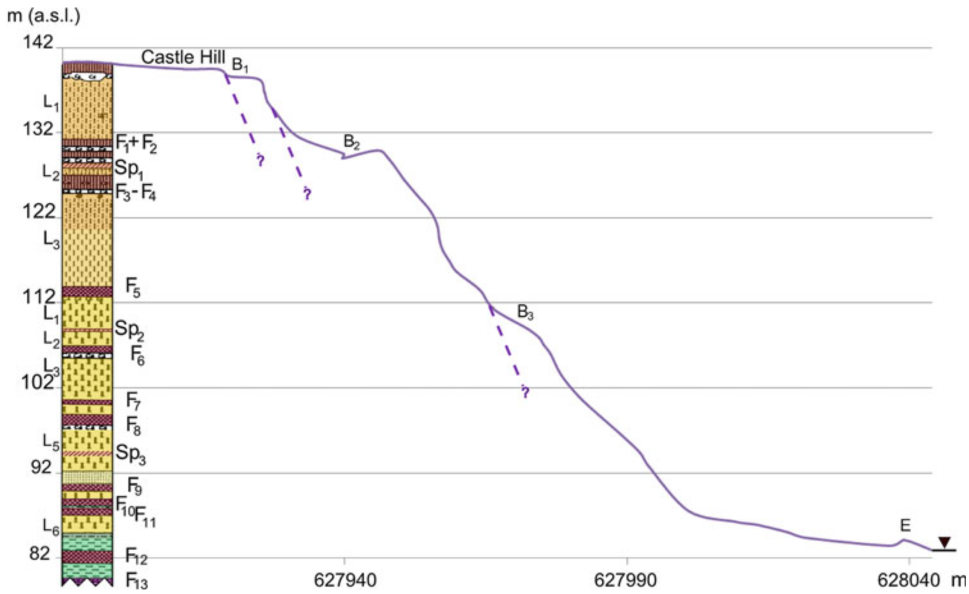


**Fig. 14.4** A differential GPS survey (coordinates in HD 72) of the cross section of a recent landslide on the Castle Hill, Dunaszekcső: $B_1$ Recent block of landslide, $B_2$ Sliding block in 2008, $B_3$ Old block(s), *E* Rock embankment. Source of loess section: Pécsi and Schweitzer 1995: $L_{1–3}$ Young sandy loess, $L_{4–6}$ Old loess, $F_{1–4}$ Steppe type soil, chernozem, $F_{5–12}$ Brown forest soil, $F_{13}$ Red clay, $F_5$ Mende Base marker soil complex, $Sp_{1–3}$ Semipedolite


**Fig. 14.5** Subsidence of the southern part of the recent Castle Hill landslide caused a few cm displacement (**a**) in the early phase (December of 2011) and developed into ca 30 cm high scarp (**b**) till January of 2014. In the northern part of the same landslide, the subsidence was more intense in early months (**c**) and a higher (ca 50–60 cm) scarp (**d**) developed till January of 2014

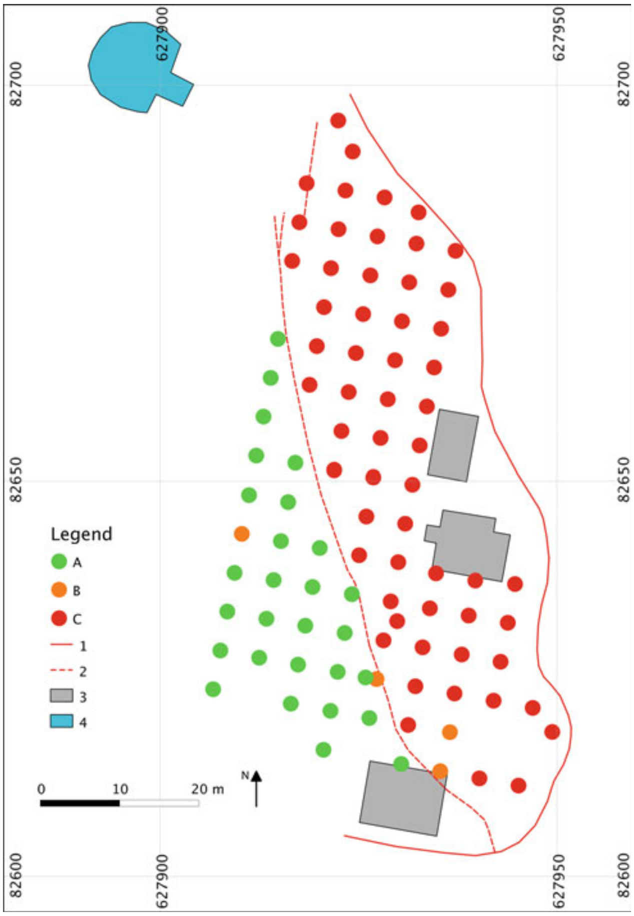


**Fig. 14.6** Geodetic survey network (coordinates in HD 72) to record vertical displacements with total station. Points were measured from October of 2011 to January of 2014 and show no subsidence (*A*), minor subsidence (*B*), or significant (up to 40 cm) subsidence (*C*). *1* Relict scarp, *2* Recent crack, *3* Cottages, *4* Former water tank

## 14.6 Conclusions

The eastern margin of the elevated Mezőföld Plain and Tolna-Baranya Hills has been affected by landslides for thousands of years. The Danube undercuts this loess-covered old Pannonian surface and washes away sediments, together with ancient and recent human structures. Recent landslide processes at the Dunaszekcső Castle Hill are contemplated as an 'in vivo' geodynamic experiment to understand the mechanisms of movement and to interpret landform evolution in the Danube valley of Hungary.

It is essential for the protection of the loess bluffs to drain the waters stored in aquifers and to protect the bank against high floods that obstruct subsurface water flow and moisten loess walls. It is necessary to emphasize that loess bluffs are inherently susceptible to collapse, and the mobility of its layers is associated with geological, geomorphological and hydrological factors. Geological and geomorphic settings, mostly the presence of Late Miocene clayey sediments, Pliocene–Pleistocene red clay and paleosols of the loess series, coupled with high and steep slopes are considered the basic factors for landslide hazard. But climatic and hydrological factors contributing to sliding were also identified. Mitigating this hazard and eliminating the related damage is a major challenge for engineering geomorphology.

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# Caves in the Buda Mountains

# 15

Szabolcs Leél-Őssy

**Abstract**
On the territory of Budapest, there are about 170 caves: mainly in the Rózsadomb (Rose Hill) area. The total known length of these caves (in the city) is more than 52 km. The caves of Budapest are hypogene (thermal karstic) caves, dissolved by mixing corrosion of ascending waters along tectonic joints. Therefore, the cave passages are totally independent of surface morphology, and there are no fluviatile sediments in the caves. The origin of the caves can be reconstructed from the careful reconstruction of underground circulation routes. The caves are characterized by varied morphological features: spherical cavities along corridors of various size, the walls and floors, sometimes even the ceilings, of which are well decorated with mineral precipitations (calcite, aragonite and gypsum, a total of almost 20 minerals), the most common being botryoids, but dripstones are also common. The cave passages are mainly formed in the Eocene Szépvölgy Limestone Formation, but the upper part is often in Eocene-Oligocene Buda Marl. The deepest horizon is sometimes in the Triassic limestone (Mátyáshegy Formation). Based on U-series dating of their minerals, the Buda caves are very young (between 0.5 and 1 Ma).

**Keywords**
Hypogene maze caves • Tectonic preformation • Spherical cavities • Botryoids • Mineral precipitations • Budapest

## 15.1 Introduction

Every cave of Budapest is situated (Fig. 15.1) in Buda, west of the Danube, whilst Pest lies in a plain on a few hundred metres of unconsolidated sediments. The Rózsadomb cave area in the Buda Mountains is only 5–6 km$^2$. At the end of 19th century nobody knew that there were big caves here. Only a few small (10–20 m long) caves were known inside Gellért, József, Mátyás and Vaskapu Hills.

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## 15.2 History of Cave Discoveries

Cave discoveries started simultaneously with limestone quarrying. First, the entrance to Harcsaszájú (“catfish mouth”) Cave was discovered in the Pál-völgy limestone quarry in 1902 and Pál-völgy Cave was found in 1904. This was the first big cave: more than 1 km of passages had been mapped by 1910. Until 2010 further 13 km of passages had been explored. Harcsaszájú Cave in the upper part of the quarry was 100 m long just like Hideglyuk (“cold hole”) Cave in the deepest point of the same quarry. In 1930 a small spherical cavity was found on the top of Szemlő Hill during house building. After a few days a 2.2-km-long cave very rich in precipitations, mainly botryoids, cave raft and gypsum crust was opened. Three years later, from another spherical cavity in Törökvész Street the Ferenc-hegy Cave was discovered during canalization works. By 2005 its explored length reached

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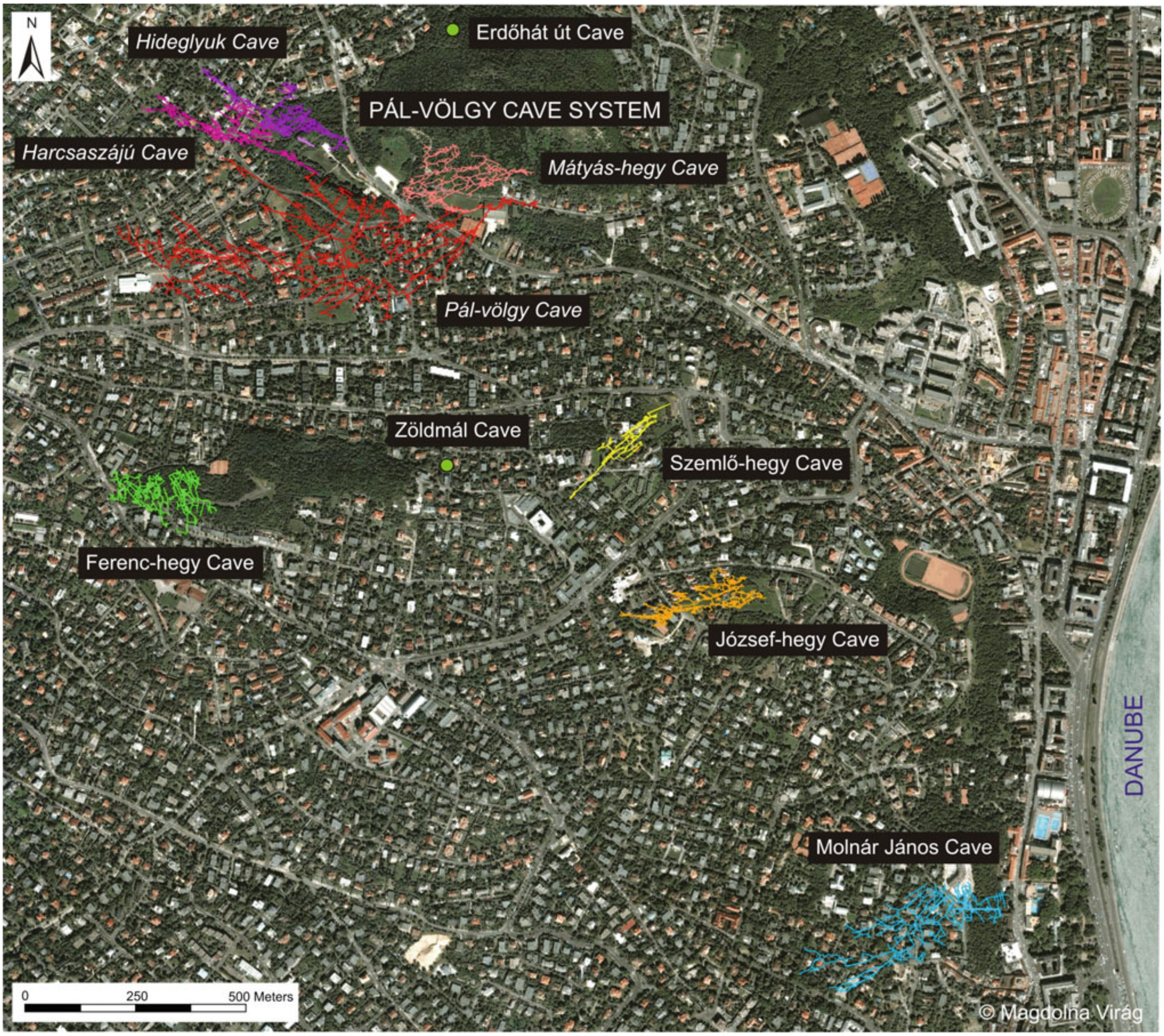


**Fig. 15.1** Cave-polygons displayed on the Google Earth (IKONOS) map of Rózsadomb (speleological data from the National Cave Register) (by Magdolna Virág)

6.5 km. In the 1930s two limestone quarries operated on Mátyás Hill and several small caves were found. The longest one, Futura Cave, as well as the Pál-völgy Cave, were used as a shelter during the Second World War. From the Futura Cave the 5.6-km-long Mátyás-hegy Cave was opened in 1948.

Next to Lukács Spa, behind the Tunnel Spring cave passages are filled with 17–27 °C spring water to the ceiling. The divers could swim into the cave in 1972, and in the next five years a 400 m-long cave was discovered. In 2002 the back part was found and explored to 6–7 km length by now. The whole cave (except the Kessler Hubert Chamber) is under water.

At a housing estate construction on top of József Hill in 1984 from the base pit vapour was emanating. József-hegy Cave of 6.5 km length, discovered at this site, is the most beautiful cave in Budapest, adorned by abundant calcite, aragonite, gypsum and barite crystals. In 2001 the connection between Mátyás-hegy Cave and Pál-völgy Cave sought for more than ten years was found. In 2005, work at the end point of Hideglyuk Cave started and it was extended by more than 4 km. In 2011 the connection between the Harcsaszájú–Hideglyuk Cave, and the Mátyás-hegy–Pál-völgy Cave was discovered and this new system, longer than 30 km became recognized as the longest cave of Hungary in 2013!

We can see that the caves of Rózsadomb were originally closed caverns, the caves could only be found by human work (quarrying, house building, canalization, drilling, etc.) (Leél-Őssy 2003). This also applies to the caves on Gellért Hill. Only a single cave, the Rock Church had been known before the 20th century. Everybody could see its big mouth (6 × 12 m) from the Danube. (The word “pest” means cave in old Slavic language: the capital was named after this cave!) It is only one big hall. In 1969 from this hall the Aragonite Cave was found by drilling. There are several active and inactive spring caves, but every cave is only a few metres long. In 2007, we discovered the Citadel Cave (in marl) from a base pit of a house building which is only 70 m long and 18 m deep only, but extremely rich in calcite, aragonite and gypsum crystals. The caves of Gellért Hill are in Triassic dolomite (Main Dolomite) or in marl (Buda Marl). Outside Rózsadomb and Gellért Hill there is only one major cave in Budapest, Bátori Cave, whose total length is 360 m, and depth is 50 m. It is formed in Triassic Dachstein Limestone and inhabited from the Neolithic. In the 15th century iron ore was mined there. The origin of another interesting cave, the Castle Cave, is similar to that of the caves of Rózsadomb, but here the host rock is very young (2–3 ka BP) freshwater limestone (travertine). The caves here were independent and very small: the heights of caverns were 0.5–1.5 m, but when the city was built (between 1243 and 1245) a lot of wells were dug. The 6–8 m thick travertine was cut through, and sometimes small caves were found. The limestone is underlain by 1–3 m thick loose fluviatile sediments, followed by Buda Marl with high clay content. The people expanded the caves: they created rooms and corridors for storage and shelter. During World War II a large shelter was established with toilets, kitchens and bathrooms for 10 thousand people!

## 15.3 Geological Setting

The oldest rocks in the area are somewhat older than 200 Ma. The sedimentation of the Budaörs Dolomite of 500–1,000 m thickness in the south started in the Ladinian, and finished in the Lower Carnian (mainly in the Middle Triassic). It is greyish white, well-stratified rock settled on a carbonate platform, rich in *Dasycladacea* algae (Esteban et al. 2009).

The Late Triassic Mátyáshegy Formation of 50–200 m thickness (Haas 2012) consists of two members: limestone and dolomite (chert in 5–10 cm nodules). Mátyás-hegy Cave and the lower part of József-hegy Cave is in intensively karstified limestone, while Erdőhát Cave and Tábor-hegy Cave are in Sashegy Dolomite.

The Carnian-Norian Main Dolomite (Hauptdolomit) of 1,500–2,000 m thickness (Haas 2012) is of grey colour, formed on a carbonate platform. As the rock is not prone to karstification, caves are very rare in this rock.

The Norian-Rhaetien Dachstein Limestone of 700–1,000 m thickness also deposited on a carbonate platform. It can be found on János Hill, the highest point of the city (529 m). Its colour is white and grey and it also abounds in lofer cycles and *Megalodontaceae* shells. A very well karstified pure limestone in the Alps, it only contains one or two minor caves in Budapest.

In the later Mesozoic the area was mainland. The Triassic formations are covered by Tertiary and Quaternary formations; thus there is a more than a 150-Ma-long hiatus in the sequence. The next rocks, Basal Conglomerate and Basal Breccia were deposited during the Eocene transgression (40 Ma ago) and are found in Sziklatemplom (“rock church”) and Zöldmál Caves.

The Szépvölgy Limestone, also of Upper Eocene age, is of brownish-grey colour, 50–100 m thickness, with large foraminifers (Nummulites, Dyscocyclina, Assilina, etc.), and was formed in deeper sea. These animalculeses (Protozoas) are 1–3 cm in diameter! There are also numerous shells and corals in this formation. Most caves of Rózsadomb were developed in this rock. The overlying Upper Eocene–Lower Oligocene Buda Marl Formation is yellowish brown and has a thickness of 100–300 m. It is well known from the caves: the upper parts of Rózsadomb caves (and the whole Citadel Crystal Cave) formed in this rock. The Buda Marl has variable, but always considerable clay content, which hinders karstification. The lowest part of the Buda Marl is called bryozoan marl.

The deeper caves are found in the thin (90–130 m thick) Tard Clay on Rózsadomb, Castle and Róka (“fox”) Hill. This grey-black, very well-bedded clay deposited in an anoxic environment is very poor in fossils (Báldi 1986).

In the Oligocene, after a short (2 to 3-Ma-long) hiatus, the Hárshegy Sandstone (often with small gravels, rather a conglomerate) of 20–200 m thickness and generally brownish-red colour was deposited (Báldi 1986). It is a really hard rock of quartz sand and gravel with chalcedony matrix. During the oncoming Oligocene transgression the grey Kiscell Clay with small foraminifers (0.1–0.2 mm), shells and snake fossils formed. At the end of the Oligocene and in the early Miocene there was a shallow sea in this area, where clays, sands and gravels deposited (Báldi 1986) in waters of normal or low salinity. In the Badenian, l5 Ma ago, a volcano erupted near the Buda Mountains. The Badenian Rákos Formation was the last to come about in water of normal salinity in Hungary. In the Pest side (in Old and Young Hill of Kőbánya) this limestone was quarried by deep workings, and a network of tunnels was created and used as a brewery after quarry closure. The next sediment, the Tinnye Formation (mainly limestone) settled in brackish water and was quarried in Budafok. The pits are used as wine and champagne cellars today.

At the end of the Miocene (5 Ma ago) sandstone deposited again. Finally, freshwater limestone precipitated from the springs (Kele et al. 2009) in many places (on Castle, and Gellért and Rose Hills) up to 200 ka ago (Fig. 15.2). Castle

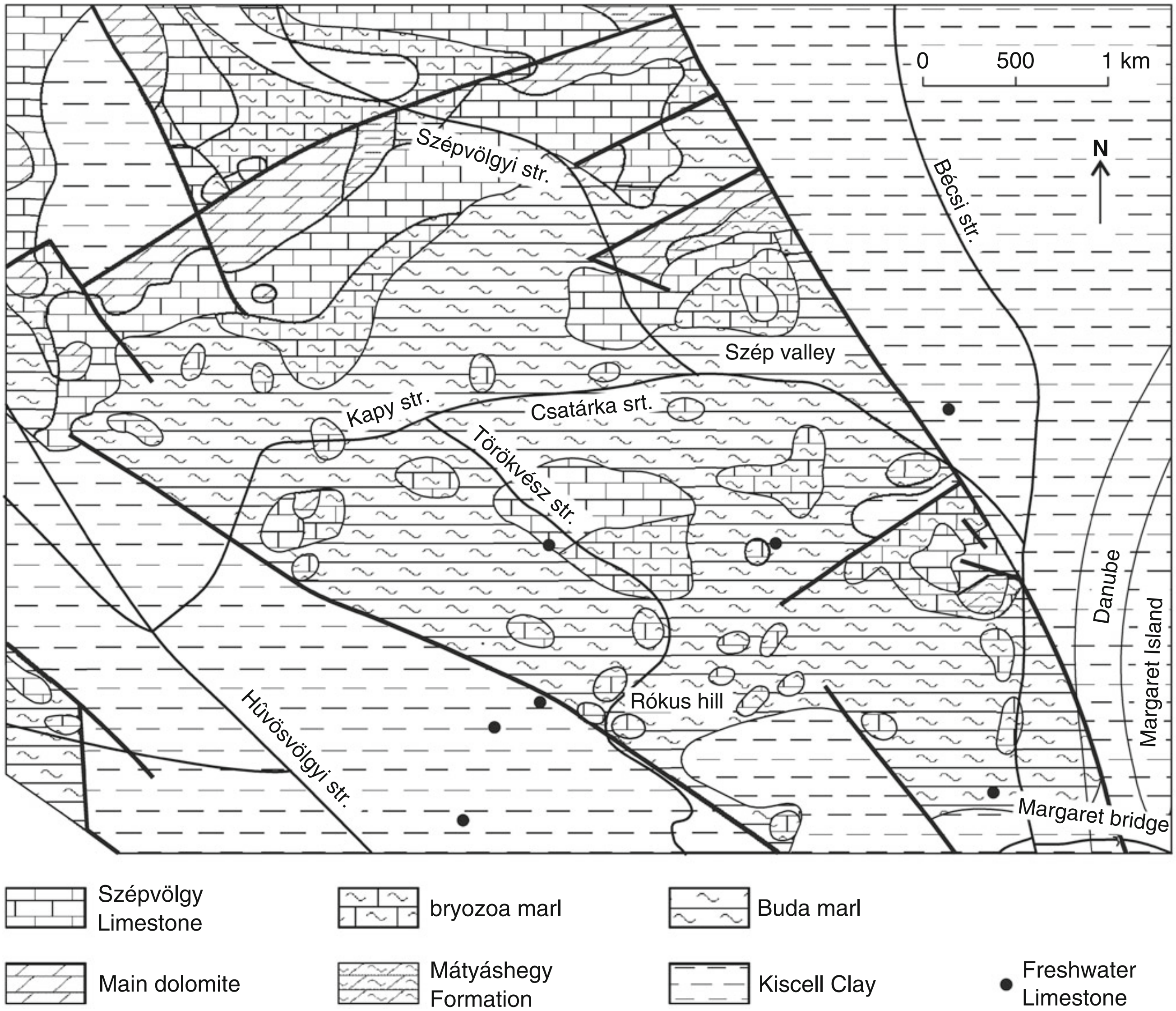


**Fig. 15.2** Geological map of the central part of Buda (after Wein 1977)

Cave was dissolved at the junction of the Buda Marl and freshwater limestone. It is now on the top of Castle Hill, 50 m above the Danube level, under 6–8 m thick travertine. Everywhere in the cave the noise of the traffic can be heard.

## 15.4 Origin of Caves

Around the cave area, in 60–70 km$^2$, rainwater infiltrates into the karstic rock, reaching after some time different depths and warming up to different temperatures (Jakucs 1977). There are local, intermediate and regional flow routes (Tóth 2009; Erőss et al. 2010).

Local water circulation reaches down only a few tens of metres into the rock. Water temperature in local circulation is almost the same as mean annual air temperature of the area: ca 11–12 °C. The intermediate route is maximum a few hundred metres long and its temperature is lukewarm: ca 15–25 °C. The warmest is the regional route, which reaches below 1,000–1,500 m and its temperature is between 60 and 80 °C. It is possible that the regional route has an additional section: some water arriving from below the Pest Plain. The infiltrating water is flowing in an eastern direction. Under Pest, where the Triassic karstic rocks are buried under impermeable deposits, it turns back and comes to the surface along the Thermal Fault System (again on the right Danube bank), along which several lukewarm and hot springs of balneotherapeutical importance issue. Budapest is therefore a world-famous spa city.

Along the tectonic joints waters of different type mix near to the surface, and dissolve limestones and marls. The theory of mixing corrosion (Bögli 1965; Runnels 1969; Plummer

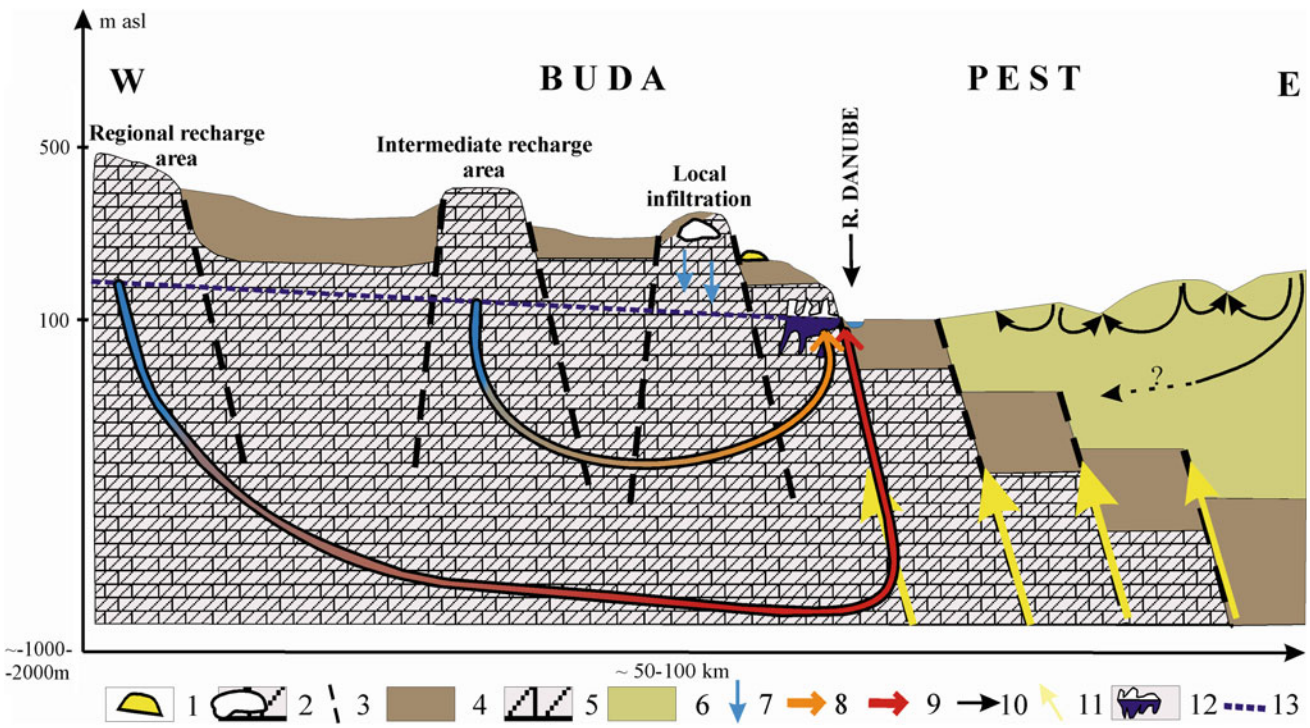


**Fig. 15.3** Revised conceptual model for the flow systems of the Buda Thermal Karst (modified from Erőss et al. 2010). *1* Travertine, *2* Inactive, dry cave, *3* Fault, *4* Oligocene clays, *5* Carbonates, *6* Neogene sediments, *7* Local infiltration, *8* Flow of karst water of meteoric origin, *9* Regional karst water component, *10* Gravitationally driven flow systems on the Pest side, *11* Basinal fluids driven by tectonic compression, *12* Hypogenic active cave, *13* Water table

1975; Palmer 1991; Ford 1995; Klimchouk and Ford 2009) appears to provide the best explanation to the origin of these caves (Müller and Sárváry 1977; Takács-Bolner 1989; Dublyanskiy 1995; Ford 1995; Leél-Őssy and Surányi 2003).

Near to the surface, in the zone where different ascending (and also descending) waters meet, mixing is continuous. If this zone is stable for a long time, spacious passages form here. In the Late Pleistocene, when the water table sank quickly, it could only result in narrow corridors. Therefore, the caves of the Buda Hills constitute a multi-storey system. The tectonic preformation is very important in the genesis of these caves: the joints allow for the movement of descending and ascending waters. The examination of the fissures and the features on their surfaces provide new evidence on the origin of the caves (Fodor et al. 1991). In the multi-level caves there are some large halls several tens of metres long. Their size depends on rock properties and on the dissolution capacity of water. It is well known that the $CO_2$ dissolution capacity of cold water is higher than that of hot water; therefore, cooling can continuously dissolve limestone (Ford 1995). This process is taking place in ascending thermal water during the circulation by mixing corrosion (Fig. 15.3). The thermal water coming from the deeper parts is much warmer than its environment. During cooling, the water dissolves the wall of the fissures of the caves. Thus, the Rózsadomb caves can be considered fossil spring passages. The mixing zone was found in the Szépvölgy Limestone in the Middle and Late Pleistocene, when the caves were formed. Therefore, 80 % of cave passages are in this rock.

The cave passages entered into the next, destructive phase of their development when the water table sank below the cave. The cave halls and corridors were filled up with residuals from rock dissolution and collapsed material of the wall. The solution features of the cave form both below and above the water table. Mineral precipitation in the caves started earlier than the dissolution of cave passages, and continues to this day. In the Buda Hills, minor paleokarst caves are common in the larger younger caves (Müller 1989).

## 15.5 Morphological Properties

The caves in the Buda Mountains have seven typical properties which determine their character:

1. As passages were dissolved along tectonic joints, tectonic preformation is striking on cave maps.
2. The cave systems are multi-levelled and maze-like.
3. Since the galleries were dissolved by ascending and mixing water, they are absolutely independent from surface morphology.

4. The variability of cave size is high: sometimes there are only small catwalks between the 20–30 m long halls in every Rózsadomb cave.
5. Spherical cavities and corrosion niches very often adorn the walls and ceilings of caves. They are the predominant features of the Buda caves.
6. There are no fluviatile sediments in the corridors. Only Mátyás-hegy Cave (a member of the Pál-völgy Cave system) collects percolating waters, and some fluviatile debris is found.
7. Almost every cave is very rich in hydrothermal precipitations from cold water.

## 15.6 Minerals in the Buda Caves

The minerals of the caves can be classified into four groups on the basis of the time of their origin.

### 15.6.1 Preformed Minerals

Two minerals belong to this group: calcite and barite (Fig. 15.4). Calcite veins are rather common both in caves and limestone quarries. They regularly crosscut the rocks within which the caves were formed. Sometimes, if there is enough room in the fissure, the calcite forms scalenohedral crystals of a few centimetres. The largest white and sometimes yellow, often transparent "dogtooth spars" are found in Molnár János, Pál-völgy, Ferenc-hegy and József-hegy Caves. Brownish-yellow barite veins are also very common both in the Szépvölgy Limestone and Buda Marl. The length of the edge of independent rhombic crystals sometimes reaches 2–3 cm!

### 15.6.2 Minerals Directly Precipitated from Warm Water

Only calcite precipitates from the thermal water which had dissolved the cave earlier. After dissolution, when the water table started to sink, the pressure reduced, there was air above the water, and $CO_2$ was released from water (degassing). When the $CO_2$ level is dropping in water, precipitation begins because the rest of $CO_2$ is not sufficient to hold all the carbonate in solution.

Two main types of these precipitations are cave rafts and bedded calcite crust. Cave raft (Fig. 15.5) precipitates on the surface of warm ponds (Hill and Forti 1997) like ice. When dripping or moving water breaks up the thin crust, the broken raft sinks to the bottom, gradually forming concentrically thicker precipitation (for instance, in József-hegy Cave—Leél-Őssy et al. 2011). In the Buda Mountains Pál-völgy Cave is the richest in this mineral form, but it is known from many caves, e.g. from Tamara Cave and the underground Sáros ("muddy") Spring of the Gellért Bath. Cave rafts are the best source of data on cave genesis. This precipitation verifies that the Rózsadomb caves are younger than 1 Ma, became dry 65 ka ago, and that the water table sank 0.2 mm every year over the last 350 ka (Ruszkiczay-Rüdiger 2005; Leél-Őssy et al. 2011; Szanyi et al. 2012). The 2–6 cm thick bedded calcite crust, which consists of 5–10 layers precipitates maximum a few metres below the water table, often in cauliflower shape. In Szemlő-hegy and József-hegy Caves it covers the walls everywhere, while in József-hegy Cave basin or pool fingers hang from the almost vertical walls. Precipitations are up to 1–2 cm in diameter and 3–8 cm in length. The calcite sponges (in the Erdőhát út, Buda and Ferenc-hegy Caves) are of white or yellow colour, consisting of soft and crumbling fibres of 10–20 cm length and 1–2 mm width.

**Fig. 15.4** Barite crystals in József-hegy Cave (photo by Csaba Egri)

**Fig. 15.5** Cave rafts in Pál-völgy Cave (photo by Csaba Egri)

**Fig. 15.6** Aragonite needles in József-hegy Cave (photo by Szabolcs Leél-Őssy)

**Fig. 15.7** Gypsum flowers in József-hegy Cave (photo by Szabolcs Leél-Őssy)

### 15.6.3 Minerals Precipitated from the Vapour of Warm Pond

The pond in the cave hall evaporates and the vapour above the water precipitates on the colder walls. Mixed with the infiltrating water on the walls its dissolution capacity grows. Flowing down slowly, near the water table, on the warmer wall it again warms up, cannot hold the original $CO_2$ content, and botryoids form (Hill and Forti 1997) everywhere in Ferenc-hegy, Szemlő-hegy and József-hegy Caves. A single botryoid is smaller than 1 cm, of brownish-yellow colour, but in József-hegy Cave it is white. In these caves other types (angular, rose, coral, dripstone, draught and glass-ball botryoids) also occur. The aragonite (and dolomite) crystal needles (1–3 cm long and 1–2 mm thick) form similarly, regularly on the top of botryoids (Fig. 15.6).

### 15.6.4 Minerals Precipitated from Infiltrating Water

In the Buda Mountains the rocks in which the caves formed are usually overlain by the Buda Marl with high pyrite content. Therefore, the infiltrating-dripping waters in the cave have high sulphate and low carbonate contents. This is the reason why the caves of the Buda Mountains are rich in gypsum and poor in dripstones. Gypsum occurs in seven forms: gypsum chandelier, crust, flower or snake (Fig. 15.7), hair or gypsum grass, selenite needles, cave blister and gypsum stalagmites—all still evolving further. The largest gypsum chandeliers, hanging from the ceiling like stalactites and up to 1 m in length, precipitated from infiltrating waters (Palmer 2007). When the caves were discovered, many of them already lay on the floor, probably due to an earthquake or detached as a consequence of their own weight. In Szemlő-hegy, József-hegy and Citadel Crystal Caves (and locally in Mátyás-hegy Cave) there are numerous, 1–5 cm thick crystalline gypsum crusts with 1–2 cm long crystal needles. The gypsum flowers are between 5–15 cm in the József-hegy and Citadel Crystal Caves. These forms can be twisted many times (Ghergari and Onac 1995). The length of the largest gypsum hair crystals is ca 80–90 cm, while their diameter is less than a single hair. The crystals are moving when the cavers are breathing. They hang from the ceiling, on the wall, or lie on the floor. This form was described from the Szemlő-hegy Cave, but we found it in the Citadel Crystal Cave too. József-hegy Cave is the richest in such crystals and the selenite needles are 3–20 cm long, lying on the clay floor. In the third room of Citadel Crystal Cave there are many needles. We only saw cave blisters and gypsum stalactites in József-hegy Cave.

Among the recent crystals calcites predominates. It is true that generally dripping waters are very poor in carbonate, but where the cover bed is limestone (instead of marl), there are beautiful dripstones: in Pál-völgy, Harcsaszájú and Hideglyuk Caves, and locally in József-hegy Cave (mainly flowstones and stalactites, but also minor stalagmites). Calcite rombohedra (1–3 cm) precipitate from the small cold water pond in Pál-völgy and József-hegy Caves. Cold water slowly flows on the floor and builds miniature (1–2 cm high) rimstone dams. In Ferenc-hegy and József-hegy Caves glass-ball botryoids (absolutely independent from old warm water) are very common.

## 15.7 Conclusions

In the maze caves three major water circulation systems operate: local, intermediate and regional, which are all fed by rainwater. The waters spend a few thousand or even ten thousand years under the surface, the overwhelming part reaching 1–2 km depth and flowing from west to east (under the Danube). Water temperature is between 60 and 80 °C at the deepest point of the flow route. The ascending part of the circulation forms some active springs along the Thermal Fault System, on the right bank of the Danube (i.e. in Buda), at the foot of the hills.

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# Danube Bend: Miocene Half-Caldera and Pleistocene Gorge

16

Dávid Karátson

**Abstract**

The Danube Bend, a river curvature, called Visegrád Gorge in its deepest and narrowest part, is one of the most picturesque landscapes in Hungary. Its origin and relief evolution has been a long-standing problem in Hungarian earth sciences. A number of geomorphological theories have been put forward in explaining the valley with a U-shaped planform, which is incised in the surrounding Miocene volcanic mountains. In the past fifteen years, thanks to combined volcanological, geomorphological and geochronological studies, the landscape evolution of the river bend of 5 km diameter has been largely clarified. The present-day U-shaped loop is partly inherited from the horseshoe caldera morphology of Keserűs Hill volcano, a 15-Ma-old andesitic lava dome complex with an eroded central depression open to the north. The formation of the Danube Bend was initiated by river incision that started to remove post-volcanic sedimentary cover in middle or rather late Pleistocene times. These processes in turn were triggered by mountain uplift, climate changes, and drop of the remote erosion base level. The present curvature of the river was controlled by the exhumation of the horseshoe caldera as well as the surrounding resistant volcaniclastic rocks (Visegrád Castle Hill) and a hilltop lava dome (Szent Mihály Hill). Moreover, a previous meander may have also inherited. The accelerated Late Quaternary erosion and intense dissection has resulted in a “re-birth” of the volcanic relief that exhibits again steep slopes. At present, exposed spectacular rock formations (e.g. Vadálló-kövek) tower above the gorge that belongs to the Danube-Ipoly National Park in Hungary.

**Keywords**

Fluvial gorge • Miocene volcanism • Horseshoe caldera • Uplift • Incision • Danube

## 16.1 Introduction

In Northern Hungary, where the Danube river leaves the alluvial Little Hungarian Plain and reaches the Transdanubian Mountains, an arcuate, rocky, fluvial valley section can be found (Fig. 16.1). Its deepest part is named the Visegrád Gorge after the settlement of Visegrád and its castle hill that possess a rich ancient to medieval history. After passing the castle hill, the river turns southward, and starts to build islands of gravelly alluvium belonging to the Great Hungarian Plain.

The spectacular horseshoe valley section, geographically speaking the Danube Bend, is one of the most favourite tourist attractions for the inhabitants of Budapest (Fig. 16.2). It separates two mid-Miocene volcanic fields: the Börzsöny and Visegrád Mountains, which are among the oldest members of the Inner Carpathian Volcanic Chain (e.g. Pécskay et al. 2006). After pioneering research in Hungarian earth science, it has been widely accepted that the Danube River obtained its course between these mountains in Plio-Pleistocene times.

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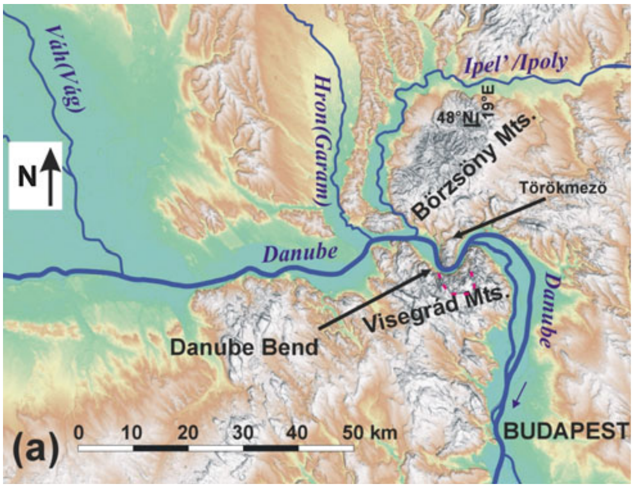


**Fig. 16.1** Location of the Danube Bend (Visegrád Gorge) between the Visegrád and Börzsöny Mountains (Karátson et al. 2006). *Red dashed line* indicates the present-day half-caldera rim of the mid-Miocene Keserűs Hill paleovolcano

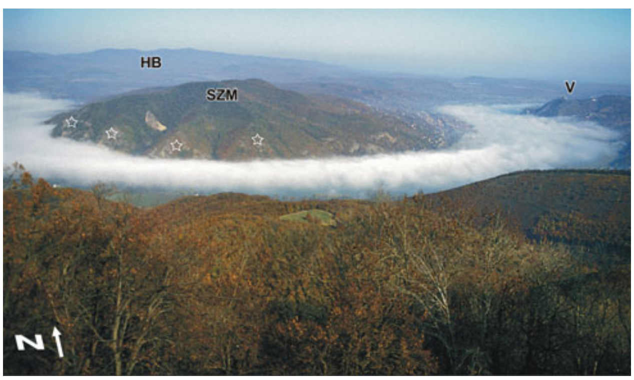


**Fig. 16.2** Aerial view of the Danube Bend. *HB* High-Börzsöny, *SZM* Szent Mihály Hill, *V* Visegrád Castle Hill. *Stars* indicate triangular facets (flatirons)

However, the fundamental question of how and why the 12-km long horseshoe-shaped section of the valley was formed has not adequately been addressed until recent studies in volcanic geomorphology (Karátson et al. 2006, 2007a, b) and Quaternary landscape evolution (e.g. Ruszkiczay-Rüdiger et al. 2005; Novothny et al. 2011). In this chapter, after summarizing the geologic-geomorphic background as well as the history of research ideas, the evolution of this spectacular region is presented.

## 16.2 Volcanic and Paleogeographic Evolution

Most of the Danube Bend belongs to the Alcapa microplate, which in this region consists of Paleozoic schists in the north (occurring as xenoliths in the volcanic rocks) and Mesozoic carbonates in the south. These old rocks are covered by sedimentary formations (sands, marls, clays, pebbles) deposited in Oligocene to Miocene epicontinental marine environments as part of the Paratethys (Magyar et al. 1999). The volcanic field of the Börzsöny–Visegrád Mountains developed in the mid-Miocene (Badenian) archipelago that covered a large area in the Pannonian (Carpathian) Basin (Table 16.1). Volcanism started prior to 16 Ma ago, and terminated ca 14.5 Ma in the Visegrád and 13.5 Ma in the Börzsöny Mountains (Karátson et al. 2000, 2007a).

A main edifice of the Visegrád Mountains, the 15 Ma-old Keserűs Hill volcano (Karátson et al. 2000, 2007a—Fig. 16.1), evolved as an amphibole andesite lava dome complex (Fig. 16.3). It showed a typical explosive character, producing collapsing lava domes that resulted in block-and-ash flows (e.g. breccia towers of Vadálló-kövek). In addition, at the end of its activity, the volcano was affected by a sector collapse producing one or more relative small debris avalanches toward the north (Karátson et al. 2006). A resultant, U-shaped caldera was formed similar to that of Mt. St. Helens (Fig. 16.4).

The post-volcanic evolution of Keserűs Hill, fundamental for the later Danube Bend, may have occurred as follows. The originally 1,300–1,500 m-high volcano was located in an archipelago of the Middle Miocene Paratethys (Harzhauser and Piller 2007). The last, isolated sea eventually became Lake Pannon, a large brackish water body covering the Pannonian Basin (Magyar et al. 1999). The volcanic relief was affected by intense erosion under the prevailing subtropical climate. However, erosion was retarded partly by a rapid infill of local erosion base (a shallow marine basin) due to ongoing volcanic activity in the vicinity, and partly by dense vegetation. At the same time, weathering led to pediplanation in higher areas resulting in flat ridges, that are conspicuous in the present-day relief (e.g. Dobogó-kő, Urak asztala, Vöröskő hills). These ridges have been uplifted later in the Pliocene and Pleistocene and further sculpted by intense areal erosion during glacial stages.

Based on erosion calculations obtained under similar climates, the rate of erosion in the mid-Miocene may have been 80–100 m/Ma, a rate that was dominant until ca 8 Ma ago (i.e. the mid-Pannonian) (Table 16.2). During that time, the Keserűs Hill volcanic edifice might have been lowered to ca 700–800 m elevation. It is interesting that correlative sediments, possibly containing volcaniclastic debris, cannot be found around the Danube Bend, implying that the removed rock should have been transported toward the internal parts of the Pannonian Basin, where intense subsidence started at that time (Kázmér 1990). During the Late Miocene, a significant (up to 300–400 m) burial of the lower parts of the Keserűs Hill volcano can be assumed (Karátson et al. 2006, 2007b). Such a sedimentation, by raising base level, might have moderated the rate of erosion (30–50 m/Ma). Less intense erosion

**Table 16.1** Major geological formations around the Danube Bend (Karátson et al. 2006)

| Age | Lithology | Facies, palaeogeography | Surface outcrops |
|---|---|---|---|
| Mid-Miocene | Leitha limestone | Reef limestone | Visegrád Castle Hill, Szent Mihály Hill |
| Mid-Miocene | Predominant amphibole andesite volcaniclastic and minor lava rocks | Lava domes and related block-and-ash flow deposits with interbedded volcanic-sedimentary rocks; subaerial volcanism | Central and N part |
| Mid-Miocene | Garnet-bearing (rhyo-) dacitic lava and volcaniclastic rocks | Lava domes, subvolcanic bodies and related pyroclastics and resedimented volcaniclastic deposits; shallow submarine to emergent volcanism | S–SE part |
| Lower Miocene | Sandstone | Shallow submarine (littoral) deposit | A narrow belt in the E periphery |
| Eocene-Oligocene | Clay, sandstone, gravel | Shallow submarine deposits | A few Oligocene localities mostly in the W |
| Upper Triassic | Limestone | Platform carbonates | S–SW of the Visegrád Mountains |

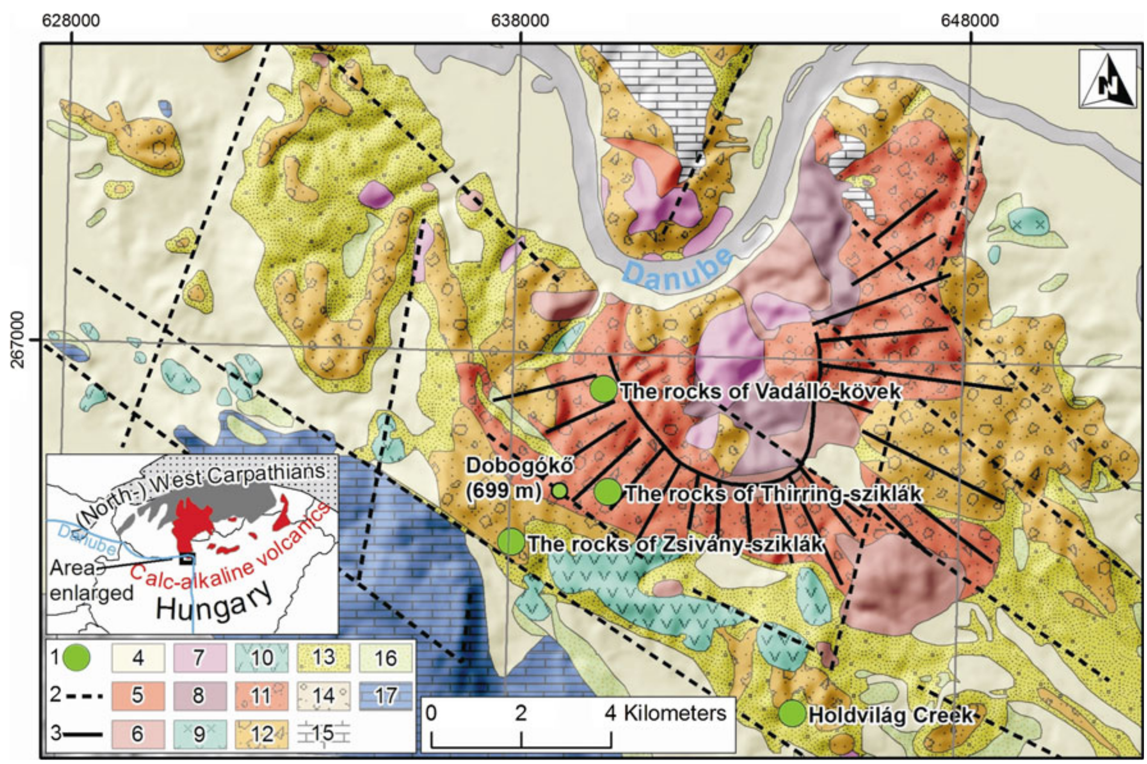


**Fig. 16.3** Volcanological sketch map of the Visegrád Mountains (after Karátson et al. 2007a, b). *1* Spectacular rock formations often with mushroom rocks, *2* faults, *3* erosional remnants of the horseshoe caldera rim and flanks of the Keserűs Hill volcano, *4* Quaternary cover, *5* pyroxene amphibole andesite lava rocks, *6* basaltic andesite, *7* pyroxene amphibole andesite, *8* biotite amphibole andesite, *9* garnet-bearing pyroxene dacite, *10* garnet-bearing biotite dacite, *11* monolithologic pyroxene andesite block-and-ash flow and debris-flow deposits, *12* heterolithologic, pumiceous, andesitic debris-flow, minor debris-avalanche and other volcaniclastic deposits, *13* pumiceous submarine/subaerial dacitic and andesitic mostly resedimented volcaniclastic rocks, *14* Mid-Miocene fossiliferous, pumiceous fine-grained submarine volcanic-sedimentary deposits, *15* Mid-Miocene cover limestone, *16* Eocene, Oligocene and partly Lower Miocene underlying deposits, *17* Triassic-Jurassic formations. Coordinates are in UTM projection

was also caused by the less warm, although still humid climate. By the end of the Miocene, the volcano was lowered to 600–700 m above sea level, but due to sedimentation in its surroundings, its relative height might have been only 300–400 m. In the Late Miocene, the infill of Lake Pannon accelerated (fed by the Paleo-Danube from the northwest: Magyar et al. 2013), and by the early Pliocene the infill became complete due to prograding river deltas from the Alps and West

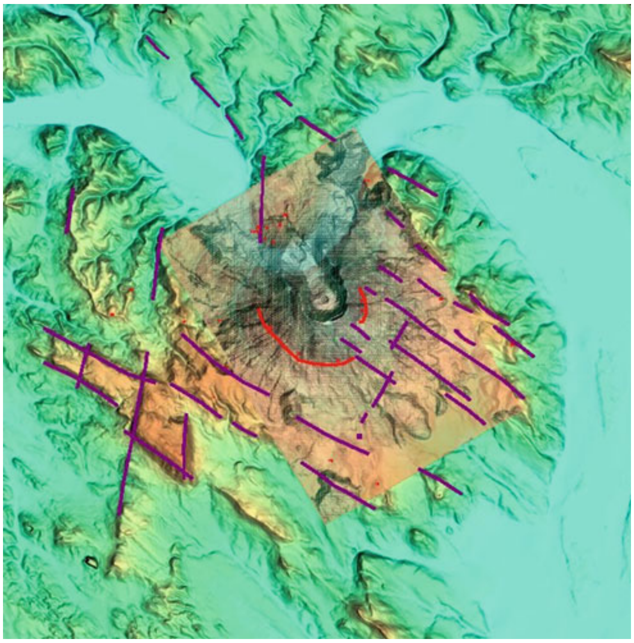

**Fig. 16.4** Superimposition of the Mt. St. Helens caldera on that of the Keserűs Hill volcano, with fault lines (*purple*; Karátson et al. 2006)

Carpathians. Around the Börzsöny-Visegrád Mountains, the precursors of a paleo-drainage appeared.

## 16.3 History of Geomorphological Research on the Danube Bend

As early as the mid-19th century, tectonic control on the Danube River was suggested by Szabó (1862). According to Salamon (1878), a late Tertiary "sea strait" might have existed in the Danube Bend similar to Bosphorus, inherited by the later Danube. The incision of a paleo-river (the presence of fluvial terraces) was first recognised by Strömpl (1915) and Schafarzik (1918). The latter author proposed a late Pliocene appearance of the Danube. This was agreed by Noszky (1935), who called attention to the importance of terrace dating. Importantly, gravel that is surely related to the Danube can only be found on the lower 3–4 terraces, making the classification of higher levels more problematic (cf. Pécsi 1959; Gábris 1994).

Subsequently, several authors explained the formation of the Danube as the result of river incision that kept pace with tectonic uplift (e.g. Sümeghy 1955; Pécsi 1959). Analysing the role of tectonic movements, Szádeczky-Kardoss (1938) was the first to suggest that the Paleo-Danube was diverted from the Little Hungarian Plain to the Börzsöny-Visegrád area around the Plio/Pleistocene boundary due to the uplift of the Transdanubian Mountains and the foreland of the Eastern Alps. This theory has been later widely accepted (e.g. Jámbor 1980). The uplift of the broader region was caused by a shift in neotectonic evolution from extension to compression (Horváth and Royden 1981; Horváth 1995; Fodor et al. 1999). Along the Danube Bend, the presence of arcuate terrace levels, highest on the middle section, was considered evidence of differential uplift (Pécsi 1959; Gábris 1994; Ruszkiczay-Rüdiger et al. 2005).

Apart from the formation of the gorge, the even more exciting question of why it is U-shaped in plan was addressed first by Kádár (1955). He argued for a regressional origin, namely, a paleo-river of the Börzsöny-Visegrád Mountains that flowed to the south (towards the Great Hungarian Plain) started to incise its bed northward, and captured another river flowing towards the Little Plain. Such a hypothesis, however, could not be proven: e.g. already the oldest sediment load of the Danube in the Great Hungarian Plain contains Alpine gravels (Pécsi 1959). Láng (1955) introduced the concept of epigenetic origin. Envisaging a former, high-level surface covered by post-volcanic sediments, he assumed that a meandering river developed, and the present-day U shape is but an inherited meander. Such an attractive hypothesis, however, was not verified either.

On the other hand, the reconstruction of volcanic landforms was not in fashion until the 1980s. The reason was partly the dominance of the concept of pediplanation: leading geographers (e.g. Láng 1955) claimed that the primary volcanic landforms could not be preserved against millions of years of erosion. Neglecting the possible role of volcanic landforms seems nowadays strange, especially when considering the widely accepted epigenetic origin; the nature of a paleosurface and paleo-landforms beneath the postvolcanic coverage obviously calls for studying. It was Csillag-Teplánszky and Korpás (1982) who first mentioned that "it is a matter of fact that volcanic structures had a role in determining the formation

**Table 16.2** Erosion, uplift and elevation changes of Keserűs Hill volcano which controlled the formation of the Danube Bend (Karátson et al. 2006, 2007b)

| Period (Ma) | Erosion rate (m/Ma) | Total vertical denudation (m) | Uplift or subsidence (m) | Volcano elevation (m) at the end of period | Relative height (m) above base level |
|---|---|---|---|---|---|
| Mid-Miocene (15–8) | 80–100 | 600–700 | −100 | 700–800 | 600–700 |
| Late Miocene (8–5.5) | 30–50 | 100 | −100 | 500–600 | 300–400 |
| Pliocene (5.5–2.6) | ≤30 | <100 | 0 | 400–500 | 200–300 |
| Quaternary (2.6–present) | 50–80 | >100 | +300 | 600–700 | 500–600 |

of Danube Bend". On that basis, Székely (1997) proposed that the Danube cut through the volcanic caldera (of Keserűs Hill), in his volcanic reconstruction completed also in the northern side, along Szent Mihály Hill (Fig. 16.2), which, however, as recent volcanological studies (Karátson et al. 2006, 2007a) revealed, is related to the Keserűs Hill volcaniclastic ring plain and a late-stage lava dome on top. Nevertheless, all these authors agreed that primary volcanic landforms could be preserved against long-lasting erosion.

Last but not least, explaining the peculiar shape of Danube Bend, some authors invoked faults determining this or that side of the bend (e.g. Czakó and Nagy 1977; Korpás 1998; Fodor et al. 1999; Csillag-Teplánszky and Korpás 1982). The most conspicuous is the fault controlling the western side of Szent Mihály Hill and another one southwest of the bend that may have controlled the actual river channel even in Holocene times (Székely et al. 2006). Nonetheless, faulting itself cannot give an explanation for the arcuate shape of Danube Bend.

## 16.4 Fundamentals of Paleodrainage

During the Pliocene (5.5–2.6 Ma), the broader area was occupied by a fluvio-lacustrine environment (Szádeczky-Kardoss 1938), characterised by poorly known paleo-rivers. The climate was cooler and more arid temperate continental (Kordos 1979; Hably and Kvacek 1998), and became unstable toward the Pleistocene. Lower amounts of rainfall, moderate elevation, and the changing, uncertain drainage system resulted in lower erosion rates (<30 m/Ma) than in Miocene times. By the late Pliocene, Keserűs Hill might have been 400–500 m above sea level. Since fluvial erosion around the volcano was not yet effective in removing the sedimentary infill, the relative elevation of the volcano was reduced to 200–300 m. At the same time, wind erosion and slope processes smoothed the volcanic relief to rolling hills.

This way, the paleogeography in late Pliocene was characterised by wide plains in the foreland of the Börzsöny-Visegrád Mountains, still infilled by marine sediments. Subsequently, these plains started to host water courses draining the Northwest Carpathians. Prior to the appearance of the Danube, an unstable drainage composed of the paleorivers of the Hron/Garam, Váh/Vág) and Ipoly/Ipel' system can be inferred (Prinz 1936; Szentes 1943). The rivers, some of them possibly meandering, flowed across the alluvial plains and may have appeared both to the south and/or to the north of Szent Mihály Hill (e.g. at Törökmező, Fig. 16.1). If the present-day Danube Bend represents an inherited meander, dimensions of the actual river curvature implies a much larger river having ca 6,500 $m^3 s^{-1}$ bankfull discharge (Karátson et al. 2006). Such conditions may have existed only sporadically (i.e. during certain interglacials). The increased erosional capacity of high-discharge river(s) and their tributaries obviously resulted in an accelerated removal of post-volcanic marine sediments.

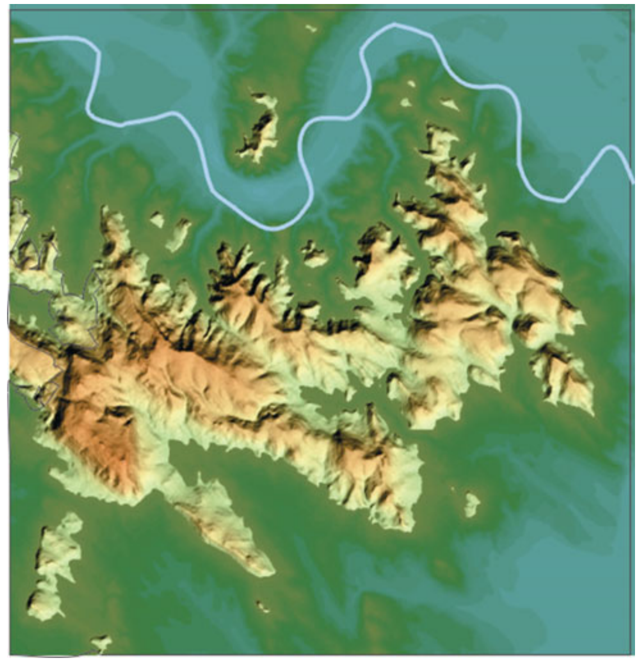

**Fig. 16.5** Digital elevation model of the Danube Bend infilling the terrain with 300 m sediments (as in Pliocene times). The *blue* and *green* areas may have been shaped subsequently by the Paleo-Danube

On the other hand, the volcanic landforms also seem to have controlled the course of the Danube. First, because the Keserűs Hill volcano had always been a positive landform, the enlarged remnant of its horseshoe caldera offered an embayment (e.g. meander) for the appearing river (Fig. 16.5). Second, to the north of the volcano, the resistant rocks of Szent Mihály Hill may have forced the river to pass round it. Finally, at the eastern end of the later Danube Bend, the Visegrád Castle Hill consisting of resistant breccia was also an obstacle, and the river should have flowed around that hill, taking its direction southward to the subsiding Great Hungarian Plain.

## 16.5 Origin of the Present-Day Visegrád Gorge

The early, more flexible (Pliocene?) drainage may have changed and long-term river channels formed when a very young uplift occurred, coupled with Pleistocene climatic oscillations (in particular, higher rainfall in the glacials) as well as the drop of base level at the Black Sea (see e.g. Georgievski and Stanev 2006). Well-developed, steep (30° to 40°) triangular slope surfaces (flatirons) on both sides of the gorge witness the latest-stage fast uplift and incision. The tips of such flatirons along the Danube Bend (both at Szent Mihály Hill and on the opposite side), located ca 150–200 m above the present river, constrain the net vertical movement of the youngest uplift.

Such rapid relief changes obviously resulted in accelerated erosion (Fig. 16.6). Based on analogues from the Alps, Pleistocene erosion rates should have been at least twice higher (50–80 m/Ma) than in Pliocene times and led to a further 100 m lowering. More importantly, fluvial processes have completely removed Pannonian cover deposits and a significant part of older (mid-Miocene) post-volcanic strata. The exhumed volcanic relief, on the other hand, started to become "mountainous" again, showing much steeper slopes and rocky ridges. Especially steep slope sections (>35° to 40°) are found inside the U-shaped erosion caldera of Keserűs Hill, on the southern part of Szent Mihály Hill, and on the northern side of Visegrád Castle Hill. As far as surface elevation is concerned, calculations on uplift, incision and denudation show that the trend of volcano lowering has changed during Pleistocene times, and by today the Keserűs Hill volcano has "grown" to a 600 to 700-m-high mountain, 500–600 m relative height above the present-day Danube channel (at ca 110 m).

Rapid mountain uplift is testified by a number of geomorphological surfaces. In relation to the differential tectonic movements, 5–8 river terraces, located at various elevations, were originally described in the Danube Bend (Kéz 1933; Bulla 1941; Pécsi 1959). As summarized above, the origin of the highest levels is uncertain, and they could be pediments rather than river terraces. Formation of the terraces were traditionally related to the glacials (Günz, Mindel, Riss, Riss-Würm, Würm) and there is an early Holocene terrace too. While the lower terraces are easy to follow and are commonly covered by alluvial material, the levels above the 4th terrace are scattered, pebble cover is scarce or absent, and their elevation greatly varies. For instance, terraces interpreted by Pécsi (1959) as of Günz to Riss age exhibit

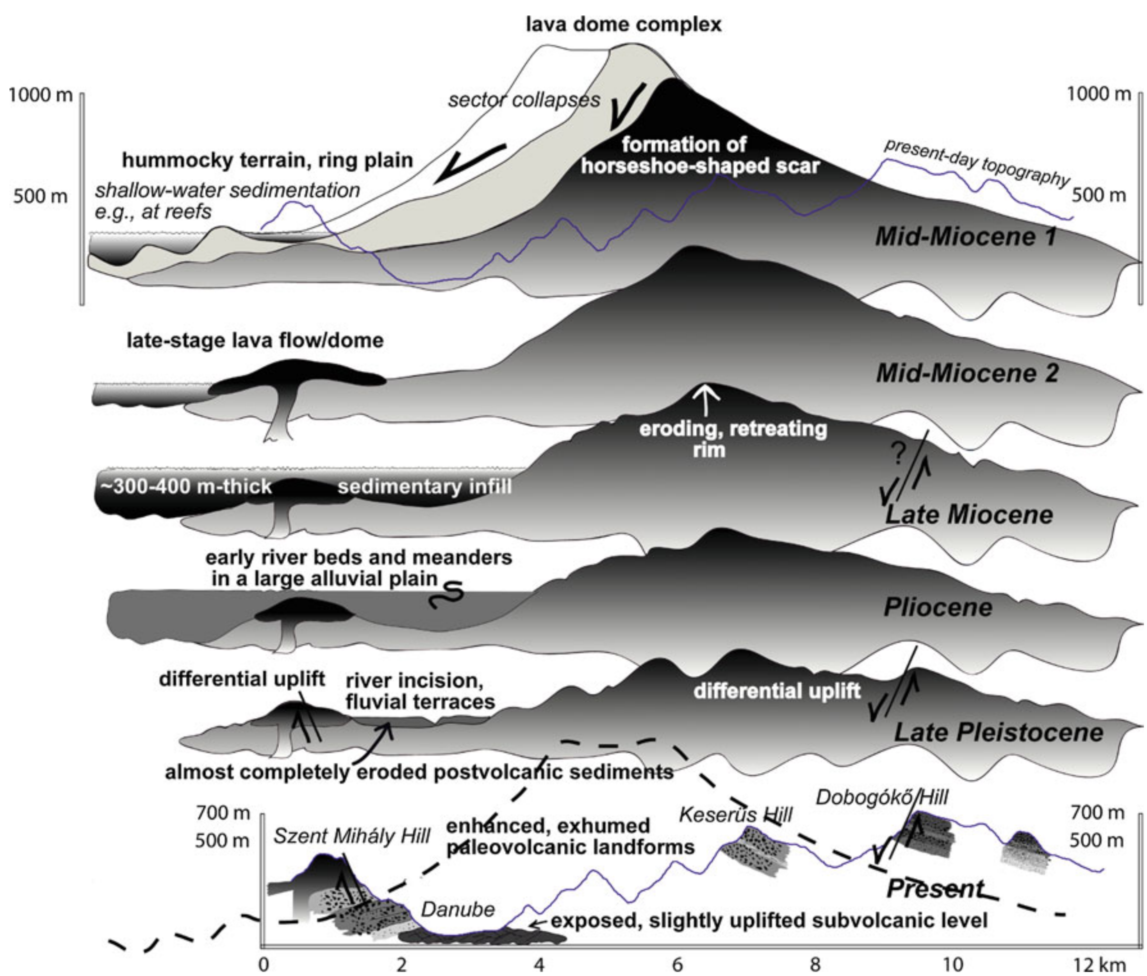


**Fig. 16.6** Geomorphic evolution of the Keserűs Hill volcano

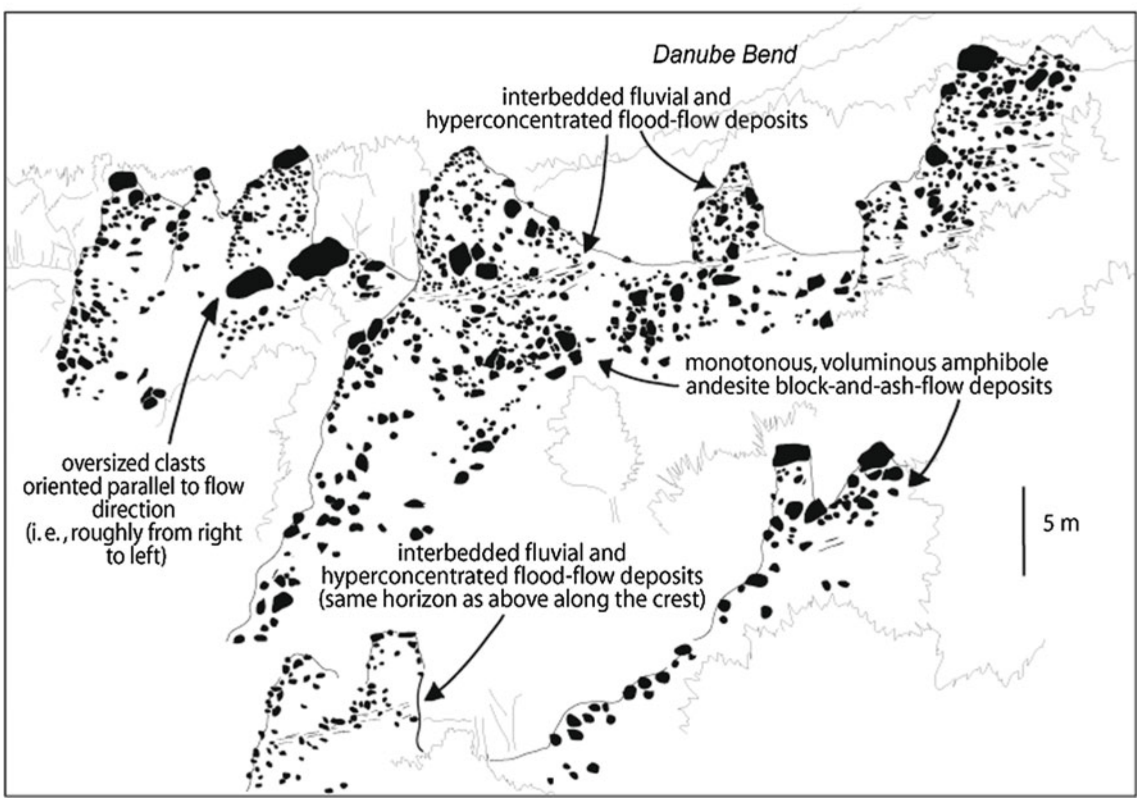


**Fig. 16.7** The Rocks of Vadálló-kövek reveal brecciated andesitic block-and-ash flow deposit

30–100 m vertical differences, reaching the maximum elevation (ca 300–350 m) in the Visegrád Gorge. As shown by exposure age dating of Ruszkiczay-Rüdiger et al. (2005) using in situ produced cosmogenic $^{3}$He, the highest fluvial terrace seems to be younger than ca 300 ka. This is verified by combined luminescence (post-IR IRSL and OSL: Novothny et al. 2011) and cosmogenic $^{10}$Be dating (Ruszkiczay-Rüdiger et al. submitted). The young Late Pleistocene age is in strong contrast to earlier estimates (e.g. Pécsi 1959; Kretzoi and Pécsi 1982) that assumed early Pleistocene terrace formation.

The findings by Ruszkiczay-Rüdiger et al. (2005) imply incision rates of ca 1.5–2.5 mm year$^{-1}$ for the Danube Bend area during the past 360 ka. After removing the postvolcanic sedimentary cover, the volcanic rocks have been exposed, and the former valley sections (including possible meanders) inherited, controlled by the volcanic landforms. Such an inherited fluvial geomorphology is not uncommon in Europe. Closest to the Danube Bend, the Bohemian Massif also shows a narrow valley segment, where the Danube, after removing the cover molasse, now downcuts into the crystalline basement (Ziegler and Dèzes 2006). In the Danube Bend, the river is now incised into less resistant volcaniclastic and fractured subvolcanic rocks (Karátson et al. 2006, 2007b). Incision has been characteristic of the tributaries as well. For instance, effective fluvial downcut dissecting the volcano flanks has resulted in the detachment of the foothill area of the Keserűs Hill volcano, which has “lost” its apron.

A nice example of complex landscape evolution can be observed along the brecciated rocks of Vadálló-kövek, mentioned above (Fig. 16.7), originally a 15-Ma-old andesitic block-and-ash flow deposit, which was emplaced on the lower flank of the volcano. However, due to long-term erosion as well as intense Pleistocene uplift, it can now be found at 500–600 m elevation, close to the erosionally enlarged and lowered horseshoe caldera rim (highest point 639 m), and on the other hand well above the deeply incised present Danube bed. It is worth noting that despite intense erosion, the remnants of the breccia—now forming mushroom rocks—still preserve the original flow direction (Fig. 16.7).

Such a spectacular landscape, along with the broader region of volcanic and even non-volcanic mountains, attracted national protection quite recently. The Danube-Ipoly National Park, established on 60,000 hectares in 1997, includes the Pilis (a karst landscape made up of limestone and dolomite), the Visegrád and Börzsöny Mountains. Apart from the caves of the former mountains revealing paleolithic finds, a number of ruins from the Bronze Age have been preserved in the Visegrád-Börzsöny area, erected typically on prominent hills. Although the popularity of the mountains (e.g. tourism and hiking), especially in the Danube Bend, is increasing, the dissemination of relevant geological, geomorphological, botanical and historical information is still a must.

## 16.6 Conclusions

The mid-Miocene volcanic field of the Börzsöny-Visegrád Mountains displays an eventful geomorphic evolution. The last chapter was the formation of the Danube Bend between the two mountains. At that area, the Paleo-Danube (along with the Hron/Garam, Váh/Vág and Ipoly/Ipel’ river system) may have appeared in Pliocene times, but at first the river was not necessarily confined to the present-day valley, but ran across a 5 to 8-km wide alluvial plain and exhibited possibly meandering sections.

Tectonic uplift, coupled with the drop of remote base level and climate changes, started in the Middle or Late Pleistocene. This resulted in a gradual and overall intense river incision in the Danube Bend, removing the still existing Late Miocene (Pannonian) and other post-volcanic sediments. Moreover, the earlier undulating hilly character of the strongly eroded volcanic region has changed due to long-term denudation; relief has become more enhanced, slopes much steeper, topography more dissected.

The fluvial incision, reaching the underlying volcanic rocks, eventually exhumed the mid-Miocene volcanic landforms. In particular, these were the horseshoe-shaped caldera of Keserűs Hill lava dome complex open to the north, and the Szent Mihály and Visegrád Castle Hill built up of resistant rocks. Although eroded, these still preserved landforms seem to have controlled the alignment of the Danube Bend valley section. In addition, the present U shape may be related to a previous meander, which in turn could be controlled by the horseshoe caldera. If so, a paleo-river with such a meander size (5 × 5 km) should have possessed a much higher discharge than today.

Fluvial incision has resulted in at least 4–5 river terraces at up to 300–350 m elevation over the past ca 300 ka. Uplift and incision are still in progress and landscape evolution is far from steady state.

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# 17 Anthropogenic Geomorphosites in the Karancs–Medves Region

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**Abstract**

The Karancs–Medves Region consists of two hilly-mountainous parts: the Karancs, the largest andesite laccolith in Hungary, and the Medves Plateau, composed of basalt sheets and necks surrounded by blocks and basins of Tertiary sediments. This region was made famous by its raw materials, including coal, basalt, andesite and refractory clay, which resulted in significant mining activities, especially from the second half of the 19th century. Based on the mines, considerable industry settled in the region and radically changed the landscape. Although by the end of the 20th century mining has stopped and the scale of industrial activity substantially reduced, in lack of reclamation abundant anthropogenic landforms were left behind, including extensive quarries, spoil heaps, slag cones, subsidence troughs caused by undermining, pseudocaves and others. It is worth visiting and studying the unique assemblage forms taking advantage of the nature trails established in the region.

**Keywords**

Anthropogenic forms • Quarry • Undermining • Subsidence • Spoil heap • Slag cone • Pseudocave

## 17.1 Introduction

Mining is one of the most powerful relief-altering human activities. The artificial removal of materials may significantly modify the topography of the surface (Nir 1983; Szabó et al. 2010). It is evident that landscapes can change especially due to mining and industrial activity, but less attention is paid to the post-mining and post-industrial period which may generate new surface-forming processes and alter the rates of others, resulting in the appearance of a different environment characterised by a new equilibrium. Based on reviews on the geomorphic role of mining made by Karancsi and Mucsi (1997), Dávid and Patrick (1998), Homoki et al. (2000), Karancsi (2000, 2002), Karancsi et al. (2007), Sütő (2008, 2010), it can be stated that the new landforms influence major ecological properties.

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## 17.2 Montanogenic Processes and Landscapes

In Hungary Erdősi (1966, 1987) introduced a new typology of the processes which cause significant landscape alteration. Among these categories (agrogenic, transportogenic, industrogenic etc.) he used the concept "*montanogenic*" for the

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processes related to mining activities and for the anthropogenic landscape resultant from these processes (Erdősi 1987). Such montanogenic processes were dominant in the Karancs–Medves region, Northeast-Hungary. In a relatively small area (ca 60 $km^2$), the region presents considerable impacts of basalt and coal mining and industrialization. Excavation of raw materials and the alteration of the surface dates back to the 18th century (moreover, basalt has already been extracted for erecting castles as early as the 13th century) and the major forms were created due to the growing demand for resources during the period of industrialization of Hungary under the Austrian-Hungarian Monarchy after 1867 (Karancsi 2002; Karancsi and Gaálova 2007; Horváth and Csüllög 2012). Considerable coal mining has existed in the region since 1848 and basalt mining since 1878, and after reaching the phase of maximum development in the 1960s, it has continuously declined and eventually ceased by the end of the 20th century. At the beginning, coal production attracted different industrial branches, among them iron, steel and glass industries were especially significant. The railway connection to Budapest built in 1868 promoted quick development. A former village, Salgótarján, in the centre of the region, became a major industrial centre. Due to underground and open-cast mining, big heaps of spoil came into being (Karancsi 2002; Dósa et al. 2006), and owing to the power station and the steel work, large amounts of slag were deposited (Angyal et al. 2004; Angyal 2007), resulting in the total transformation of the surface, especially close to these plants. In addition, cutting up original landforms and the disturbed hydrology launched semi-anthropogenic processes, and the rapid spread of traffic routes and further artificial structures resulted in the disturbance of relief and increased the degree of hemeroby (Csorba and Szabó 2009). Due to these processes, by the middle of the 20th century a montanogenic-industrogenic landscape was formed.

By the end of the 20th century, mining had become unprofitable, and after the political turn in 1989 the industrial activity proved uneconomic anymore either. The great industrial companies have finished their production and the factories were closed down. Due to the decline of mining and industrial activities, the region became a depressed area. Because of the lack of land reclamation, anthopogenic forms remained and even nowadays determine the physical landscape of the region.

## 17.3 Geological Background

The characteristic elements of the geological build-up (Gyalog 1999; Prakfalvi et al. 2007; Gaál and Horváth 2007) are Tertiary sediments, lying on a little known Proterozoic-Paleozoic crystalline basement. In the Paleogene mostly nearshore neritic sediments accumulated, forming mainly clayey marls, sandstones and aleurolites (schlieren). The main and thickest sediments are partly cross-bedded sandstones, particularly the Pétervására Sandstone Formation (see Chap. 19), and the schlieren, mainly the fine-grained Szécsény Schlier Formation. Due to marine regression at the beginning of the Miocene, a varied terrestrial relief came into being, enhanced by a heavy volcanic activity resulting in the thick ignimbrite-like Gyulakeszi (so-called "lower") Rhyolitic Tuff Formation (see Chap. 18). Following renewed transgression, brown-coal seams formed around Salgótarján. In the second half of the Miocene andesitic laccoliths intruded into the sequence, the largest one constituting Mt. Karancs, which later weathered out and now bears the highest point (729 m) of the region (Fig. 17.1). By the end of the Miocene continental conditions became dominant, resulting in heavy denudation. In the Pliocene, owing to intensive tectonic movements, the area was dismembered, certain tilted tectonic units began to rise and subside, accompanied by

**Fig. 17.1** Mt. Karancs (729 m), an andesite laccolith (photo by Zoltán Karancsi)

**Fig. 17.2** The basalt sheet of the Medves Plateau (*left*) and the neck of Somos-kő crowned by a medieval castle (photo by Zoltán Karancsi)

heavy volcanic activity. Basalts (Salgóvár Basalt Formation) erupted along deep-seated faults. Small lava cones and plateaus, explosion craters (tuff rings, maars), and weathered necks are the witnesses of this phase of volcanism. The different products (unweathered basalt, columns, laminas, scoria, xenoliths etc.) can be studied best of all along the short (0.5 km) Boszorkány-kő ("Witch Rock") Nature Trail.

During the Pleistocene vertical crustal movements resumed, and major climatic changes occurred, with humid and dry, as well as warm and cool epochs following each other. For most of the time the Karancs–Medves region belonged to the periglacial zone, where freeze-and-thaw alternations induced disintegration of rocks, solifluction (especially under 500 m elevation), soil creep and other mass movements moulding the surface.

As a consequence of young tectonic movements a typical horst-and-graben structure emerged, with steeply rising horsts with narrow grabens and deeply incised erosional valleys between them. However, the landscape is first of all controlled by rock properties. The enormous andesite laccolith of Mt. Karancs and other basalt outcrops, including necks such as Mt. Salgó, 626 m, and Mt. Somos-kő, 526 m (Fig. 17.2), and plateaus such as the Medves Plateau, 671 m (Fig. 17.2), are surrounded by landforms built up of sandstones and schlieren.

## 17.4 Montanogenic Landscape Transformation and Anthropogenic Landforms

Due to the boom of coal mining a considerable infrastructure has been built, resulting in great load on and an increasing disturbance of the natural landscape. The changes can be characterized by the fact that at the end of the 19th century the population increased 30-fold during 40 years. The alteration of the surface caused by coal mine activity can be divided into several stages. The first "mines" were only small, short adits or openings, with scanty barren materials; most of them filled up by now. After 1880 deeper shafts were opened, with cavity-working; for ensuring the wooden support, more and more forests had to be cut, resulting in significant landscape changes. In the third phase, due to increasing extraction, even larger spoil heaps accumulated and, owing to undermining, subsidence accelerated and collapses occurred at the surface. In addition, mining infrastructure (shaft-towers, loading platforms, narrow-gauge and cog-railway tracks, transmission lines, viaducts, ropeways, tunnels, coal breakers, engines etc.) has radically altered the landscape. During 150 years of mining more than 150 million tons of coal were extracted (Dósa et al. 2006; Karancsi et al. 2007), indicating indirectly how substantial were the changes which took place in the region. Moreover, mass movements are still active in the region despite the termination of coal mining.

Basalt mining was very probably active as early as in the 13th century, as suggested by the walls of the medieval castles erected from basalt columns at that time (Fig. 17.3). Nevertheless, considerable mining activity for satisfying the demands of building the infrastructure began only in the last quarter of the 19th century (Karancsi and Gaálova 2007). It was very advantageous that mining of the two raw materials (basalt and coal) used joint infrastructure. The quality of the basalt of the Medves Plateau was excellent. It was mostly used for railway construction and as cobble for paving roads. The basalt of the Medves Plateau was used for this purpose not only in Budapest but in Vienna and Paris as well (Karancsi and Mucsi 1997; Horváth et al. 2000; Karancsi et al. 2007). Due to the great demand, during 100 years—apart from the

**Fig. 17.3** The castle of Somos-kő is built of local basalt columns (photo by Zoltán Karancsi)

very small quarries which were characteristic at the beginning of the mining activity—more than 20 major quarries were opened at the rim of the plateau and total extraction reached 5 million $m^3$. It is not accidental that the biggest crushing plant in Central Europe was built here (Magyar Quarry) in 1927. Despite the good quality, also large amounts of spoil were extracted and accumulated in high heaps, especially in the 1950s, when production increased radically but the thickest strata of good quality in most of the quarries has already been extracted. Similar to the basalt mining of the Medves, considerable mining activity also took place in the andesite quarries of Mt. Karancs.

The extraction of spoil were even more characteristic for coal mining due to frequent faults caused by young tectonism. Along the faults the withdrawal of the thin coal strips could often reach several metres and therefore, to follow these strips required the extraction of huge amounts of spoil. In total, according to an inventory made by the Hungarian Geological Survey in 1990, there are 178 heaps in Nógrád County having altogether an area of 3 million $m^2$ and 32 million $m^3$ volume of spoil; more than 80 % of them in the Karancs–Medves region.

The largest basalt quarries along the rim of the Medves Plateau (Fig. 17.4) were active between 1910 and 1980 (Horváth et al. 2000). The Upper Quarry of 7.7 ha area is the largest (including 1.5 ha of spoil heaps). The quarries can be visited by the 3-km-long Nature Trail of Eresztvény Quarries, which presents more than 20 m high quarry walls with the different products of the basalt volcanism, variegated facies, and layers of thick unweathered lava, tuff, lava breccia etc. Owing to the pits and heaps the runoff conditions have changed. In the Middle Quarry (2.5 ha) between the 20 m high walls a 60 × 40 m big pond can be found (Fig. 17.5).

Some kilometres further south, close to Rónabánya village (a cultural value as a typical colony for miners), another spectacular basalt outcrop, the Szilvás-kő ("Plum Rock") rises. In lateral view the Szilvás-kő is similar to a three-humped camel, the biggest and highest "hump" (626 m) being the plateau-like central basalt sheet. An extensive basalt quarry was opened here, which exposes spectacular pentagonal and hexagonal basalt columns. In another smaller quarry perpendicular horizontal and vertical columns can be studied. Nevertheless, the most interesting and spectacular forms caused by anthropogenic effects are found on the top of the basalt sheet. The effusive basalt covered Tertiary sediments interbedded with brown coal seams. To extract the coal, the basalt sheet had to be undermined. Having finished the mining activity, the props have been picked out, therefore the galleries in 1912 suddenly collapsed and due to the collapse more than 30–40 m deep and several tens of metres long canyon-like chasms (Fig. 17.6) emerged and still exist, although partly filled up. Another geomorphic rarity are non-karstic, anthropogenic caves between the huge blocks of the collapsed material (Eszterhás 1993).

Undermining resulted in subsidence and mass movements in many other parts of the region. According to Karancsi (2002) on the Medves Plateau within an area of 10 hectares there are 268 doline-like or elongated negative forms, 1.5 m deep on average (Fig. 17.7). In general, in abandoned quarries disintegration of steep walls caused by freeze-thaw alternation is very common. Also frost shattering and frost heaving enhance these processes. Therefore, in front of the steep walls talus accumulates (Fig. 17.8). Other mass movements, mainly rockfalls, are also common.

Sometimes the horizontal galleries of coal mines function as water conduits. Probably due to the high $CaCO_3$ content of the underlying sandstone (see Chap. 19), these "water mines" build travertine terraces at springs. They can be well studied along the trail from Rónabánya to the Medves Plateau. Approximately 2.5 km from Rónabánya the spring of the Gortva River fed by a water mine is found, with remarkable travertine-like yellow-reddish precipitations.

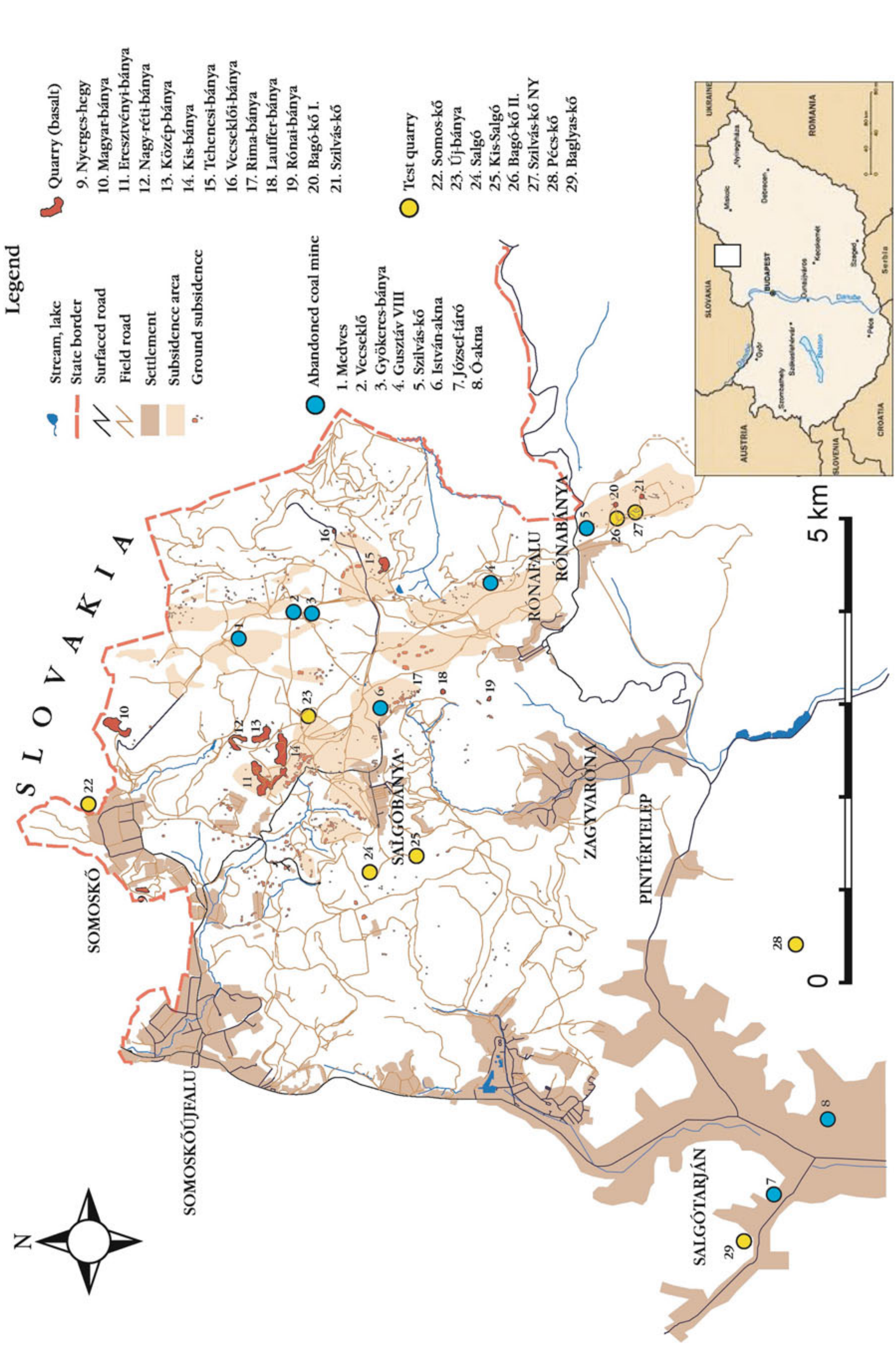


**Fig. 17.4** Map of mines and quarries (for production and testing purposes) in the Medves area (ed. by Zoltán Karancsi)

**Fig. 17.5** Pond in the abandoned basalt Middle Quarry (photo by Zoltán Karancsi)

**Fig. 17.6** Chasm on the collapsed surface of the basalt sheet of the Szilvás-kő (photo by Zoltán Karancsi)

As already mentioned, mining activity can modify runoff conditions, which can induce significant morphological changes. An excellent example can be studied several kilometres south of the Medves Plateau, close to Nemti village (see also in Chap. 20), where extremely quick ravine formation, triggered by human activity, took place. A quarry of refractory clay operated here in the last decades of the 20th century. As a consequence, the base level was lowered, natural runoff paths changed, and a new ravine emerged. Although mining activity is nowadays only occasional, during the past years the ravine has considerably extended and incised extremely quickly, reaching 15–20 m depth. Because of repeated mass movements, it is rather dangerous to visit this recent, hardly accessible ravine. Evorsion, piping processes, bank caving, and collapses take place frequently (Horváth and Karancsi 2011).

Human activity can result not only in negative but also in depositional landforms. Close to the settlement of Pintértelep (part of Salgótarján town), particular and spectacular anthopogenic forms can be seen. Looking like miniature volcanoes, two conical slag heaps have been piled up. The burnt slag originated from the former power station which was fired by the coal extracted in the region. Heaped up until 1973, both cones contained ca 400,000 $m^3$ of slag. The height of the slag heaps some years ago still exceeded 60 m, however, due to erosional processes they are getting lower year by year, especially the older cone. On the upper part of the cones heavy linear erosion occurs, while in the lower part mass movements are typical. According to Angyal (2007), the older cone has already lost more than 30,000 $m^3$ of material. Erosional rills generated in response to seasonal weather changes are much deeper and more active on the

**Fig. 17.7** Undulating forms on the basalt sheet of the Medves Plateau caused by undermining and ground subsidence (photo by Zoltán Karancsi)

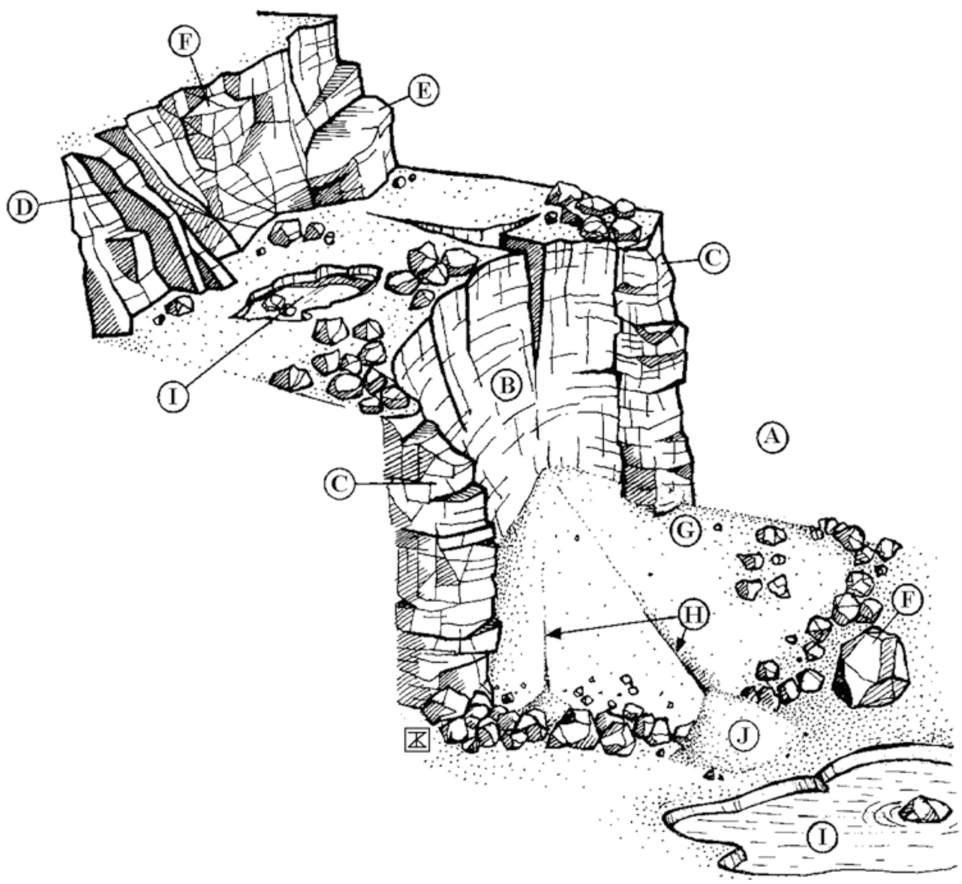


**Fig. 17.8** Schematic layout of a stone quarry (Karancsi 2000). *A* Quarry floor, *B* Quarry wall, *C* Pillar, *D* Rock buttress, *E* Rock bench, *F* Weathered rock, *G* Talus slope, *H* rill, *I* Depression with a small pond, *J* Talus cone

younger, higher cone (Fig. 17.9). The processes vary with slope inclination, vegetation cover and microclimate. Freeze-thaw alternations accelerate weathering. During wet periods, incision is much heavier and removal of the slag material can reach several m$^3$ in 2–3 days (Homoki et al. 2000; Karancsi 2002; Sütő 2008). It is promoted by the vegetation; these

**Fig. 17.9** Slag cone built up of the waste from a former power station (photo by Zoltán Karancsi)

slag-cones are covered either by very rare vegetation, or, quite the contrary, by dense, impenetrable bush, mainly composed of invasive species.

It has to be mentioned that these spectacular cones constitute only a very small portion of the slag-covered area, which can be followed along the main road between Salgótarján and its outskirt settlement, Zagyvaróna. Beyond the power station, steel manufacturing also resulted in the accumulation of a "slag mountain". As a new tendency, these slag areas are nowadays "secondary slag mines"; the material is utilized for sport camps or for covering dirt roads. Another special anthropogenic form can be studied close to the central hospital of Salgótarján, at the slopes of the hill, where sludge reservoirs form flat terraces above each other (Horváth and Karancsi 2011).

Finally, an interesting process is taking place under the surface, which is also triggered by human activities. The underground burning of the coal strata due to spontaneous combustion sometimes occurs. It could be studied in the past years on the Medves Plateau, in relation to mass movements. Namely, the collapse of abandoned deep coal mines generates fissures, which ensures supply of air and mostly oxygen necessary for burning. Along a narrow fissure several metres long hot air streams out melting the snow along the fissure, resulting in a striking colourful strip on the white snowy landscape. According to investigations on spoil heaps of coal mines, similar burning processes take place within the heaps; due to glowing burn, temperatures of several hundreds of degrees occur (Sütő 2010).

## 17.5 Conclusions

Human-induced landscape changes and landforms are increasingly common worldwide and also in Hungary. Although sometimes human activities (often inadvertently) result in striking or occasionally even spectacular landforms, these are rarely listed among popular tourist destinations. That fact motivated the authors of the present chapter, who try to draw attention to the interesting landforms which exist in a landscape fundamentally transformed by human impacts, especially due to mining and industrial activities.

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# 18 Rhyolite Badland at Kazár

Zoltán Pintér, Gergely Horváth, Gergely Jakab, and Zoltán Karancsi

**Abstract**

In North-Hungary, east of the Danube the hilly regions and middle mountains mostly consist of Oligocene–Miocene geological formations. Among them the early Ottnangian rhyolite tuff (Gyulakeszi Rhyolite Tuff Formation) is very typical. The tuffs are mainly ignimbrites originated from episodes of intensive volcanic eruptions repeated two to four times. The formation consist of pumice tuff (89 %), dust tuff (5 %) and coarse lithoclastic tuff (6 %, mainly pumice fragments), which determines the morphology of the outcrops. The rhyolite tuffs outcrop in greyish white cliffs and whitish barren patches. The barren rhyolite tuff surfaces are optimal for erosional processes, especially for gully development. Weathering, disintegration and sheet wash erosion intensively shape the landscape. Erosion is also influenced by granular structure and biogenic effects. Close to the village Kazár the erosional forms of the most extensive outcrop present a spectacular badland-like terrain. On other outcrops the formation and exfoliation of thin crusts, piping, the emergence of candle-like columns and earth pyramides produce a remarkable assemblage of micro and macroforms.

**Keywords**

Rhyolite tuff • Gully erosion • Piping • Badland • Erosional microforms

## 18.1 Introduction

At several places in northern Hungary odd greyish white cliffs and whitish barren patches can be seen—these are outcrops of rhyolite tuff. At a closer look the visitor is enraptured by interesting and attractive, mostly erosional, meso and microforms. The most extensive and beautiful occurrences are found near the village of Kazár (Fig. 18.1). The most picturesque among these sites is often mentioned as the only Hungarian badland.

## 18.2 Geological Background

The volcanic activity which resulted in rhyolite tephra production was due to horizontal movements of microplates during the early Miocene (see Chap. 2). As a consequence of extensional tectonics, NW-SE faults came into being, along them uplift and subsidence took place and magma intruded into the crust. At the beginning of the Ottnangian (20–19 Ma ago) heavy but episodic volcanic eruptions, mostly phreatomagmatic, were dominant, repeating two to four times either

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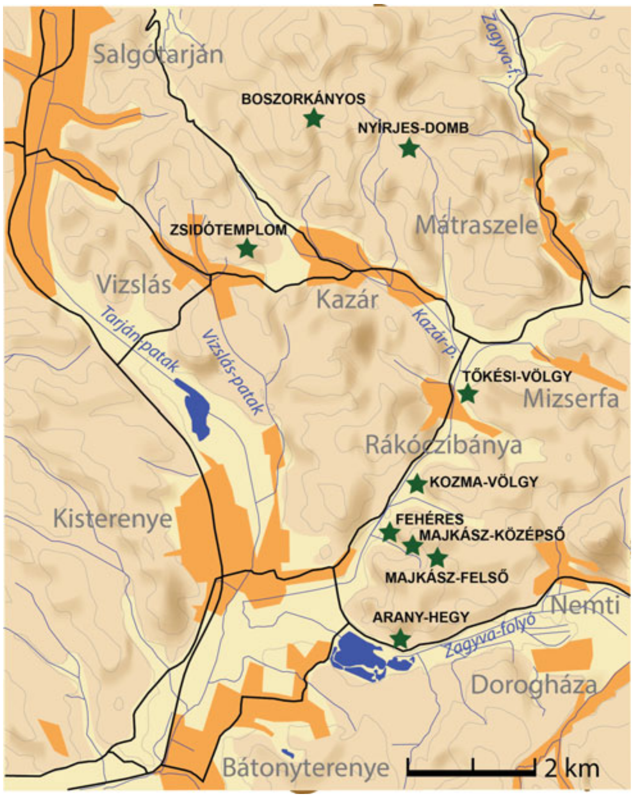


**Fig. 18.1** Location of rhyolite tuff outcrops close to village Kazár (by Zoltán Pintér and Zoltán Karancsi). *Stars* denote bigger outcrops

across single centres or along a series of craters. Enormous amounts of fallen pyroclasts consisting of volcanic ash and debris, or ignimbrites from pyroclastic avalanches resulted in the deposition of 10–30 m thick tuff in the region (Hámor et al. 1987). According to the official nomenclature of the Stratigraphic Commission of Hungary (Császár 1997) it is called the Gyulakeszi Rhyolite Tuff Formation.

## 18.3 The Rhyolite Tuff and Its Weathering

Hámor (1985) investigated the components of the tuff: the formation consists of pumice (89 %), dust tuff (5 %) and coarse lithoclastic tuff (6 %, with pumice fragments). The rhyolite tuff appears in two varieties: 'decomposed' tuff, characterized by strong argillation and welded 'glassy' tuff with a high pumice content. During the volcanic activity within the pyroclastic flows a major part of the rock fragments were welded, and, consequently, the ignimbrite-like rocks are often almost as hard as rhyolite lavas, stratified as hardened banks or blocks, with thicknesses varying from one or two centimetres to several metres.

The non-welded rhyolite tuffs are prone to weathering. Weathering processes of rhyolite tuffs were investigated by Török et al. (2005), who found that the minerals most exposed to dissolution are biotite and amphibole. Their quick argillation significantly influences denudation. Firstly, the swelling clay mineral contents of the decomposed rhyolite tuff and also the scaly structure caused by desiccation promote weathering and erosion. Secondly, if the argillizied rock is wetted, its resistance to erosion is weakened. As a consequence, hyperconcentrated (mud)flows move slowly downslope along fissures and rills. Thirdly, clay minerals play a significant role in surface evolution driven by cryogenic processes. Due to the ability of tuffs to absorb large amounts of water, regelation is highly active on such surfaces and the impact of frost processes is significant. In addition, salt weathering and biogenic crust formation also contribute to rapid rock decomposition.

In general, apart from the most resistant, silicified elements, the material of the rhyolite tuff disintegrates easily and quickly. Its rate also depends on weather conditions. During colder and drier periods frost-driven disintegration occurs, while during warmer and wetter periods chemical weathering and sheet wash are the most efficient erosive forces. The removal of the 'decomposed' tuff from the surfaces shows temporally different dynamics. After long dry periods, because of the high clay minerals content the surface of the tuffs hardens to a crust of mostly 1–2 cm thickness. Unless it spalls or becomes soaking wet, this crust protects the underlying rock mass. Under such conditions even a heavy shower can barely disrupt the surface. However, if this crust is soaked through and softens into a soap-like substance, considerable rill, gully and sheet wash erosion can ensue. The welded 'glassy' tuff is much less liable to disintegration and weathering and usually forms steep or vertical walls, cone-like or wigwam-like towers and other spectacular landforms.

## 18.4 Landforms on Tuff

The region can be characterized by variegated landforms on rhyolite tuffs (Pintér et al. 2009). The tuff is exposed at the surface due to Plio-Pleistocene tectonic movements, weathering and removal of cover sediments, and, in addition, owing to anthropogenic effects (deforestation, overgrazing; Horváth and Karancsi 2012). The barren rhyolite tuff surfaces

**Fig. 18.2** The wall of the Fehéres outcrop (photo by Zoltán Pintér)

show a remarkable assemblage of micro and macroforms, due to the fact that they are extremely prone to erosional processes. Nevertheless, the intensity of the erosion is strongly influenced by rock quality: the rate of cementation, porosity and the percentage of pumice are decisive factors.

The most beautiful erosional forms are found in the vicinity of villages Kazár, Vizslás and Rákóczibánya, along the valley of the Kazár Stream. The microforms are particularly noteworthy here, especially in the exposure of Fehéres ("whitish", Fig. 18.2). The upper part of the exposure is a bowl-like eroded surface, on which liquefaction of the 'decomposed' tuff wetted by rainwater is remarkable. The lower part consists of welded 'glassy' tuff; on its hard surface the overland flow from the upper tuff strata cuts erosional gullies with smaller rivulets, galleries, scoured walls and other features. Between the gullies varied buttress ridges and a series of columns rise, while the strips of welded tuff seem to be arranged in a horizontal and vertical order. Piping is very common: the water of overland flow disappears within the 'glassy' tuff as resurgent streams, forming horizontal and vertical tunnels under the crust (Fig. 18.3). In the course of time, however, these hidden conduits open up to the surface and generate intensive gullying into the tuff strata (Harvey 2004).

When the fluidized argillaceous substance dries up, at the end of the splits and gullies spectacular dripstone and cauliflower-like bulbs (Fig. 18.4b) develop, while on the surface of the tuff crack networks due to desiccation emerge, with tuff laminae of various thickness between the cracks. The result will be a mosaical pattern (Fig. 18.4a).

**Fig. 18.3** Piping on the wall of Zsidótemplom (photo by Zoltán Karancsi)

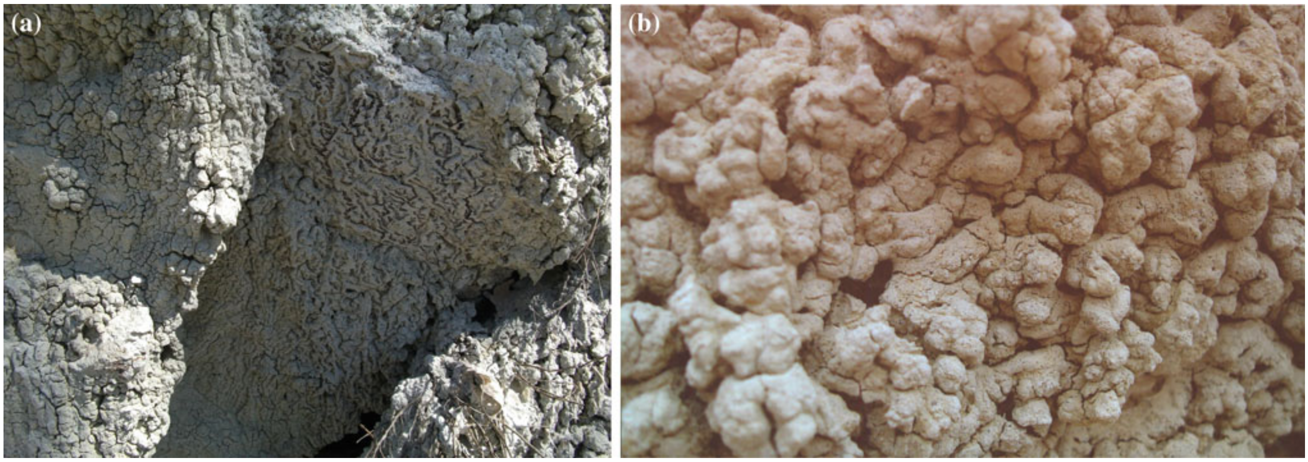

**Fig. 18.4** Mosaic pattern (**a**) and cauliflower microforms (**b**) on the wall of Zsidótemplom (photos by Gergely Horváth and Zoltán Pintér)

**Fig. 18.5** Incipient earth pyramids of Zsidótemplom (photo by Zoltán Karancsi)

More resistant tuff blocks separate from the tuff wall, constituting large buttresses. A very typical example can be found close to Kazár, in the vicinity of Vizslás village. The earth pyramids taking shape on the Zsidótemplom ("Jewish temple", Fig. 18.5) are also spectacular: on the wall there is a series of parallel candle flows (Fig. 18.6) formed by the dripping water which owe their protection to small surfaces of welded tuff on their top, which are hard as lava.

Argillization makes the Rhyolite Tuff Formation mostly impermeable and induces considerable runoff. As a consequence, overland flow is the main process of erosion. The direction of flow is determined by jointing and crumbling. Joints had already developed during cooling of ignimbrites, but their density was enhanced by exogenic processes, especially freeze-thaw alternations. The more fissured is the rock, the better it conducts running waters. Erosion caused by precipitation and meltwater starts essentially along the smaller fissures, later the fissures evolve into wider and deeper rills, which begin to retreat (regressive evolution), dissecting the surface into microforms on the larger buttresses, ridges and towers.

**Fig. 18.6** Candle flows on the wall of Zsidótemplom (photo by Gergely Horváth)

## 18.5 The Kazár Badland

On certain barren surfaces slope processes operating at rapid rates generate extremely dense gully systems. Influenced by structure, granularity and biogenic effects, disintegration, chemical weathering and erosion intensively shape the landscape, leading to the type of landscape best called badland. In general, badland formation is frequent in arid or semiarid areas on porous or collapsible sediments. Aridity inhibits the development of vegetation, which is the most efficient protection against soil erosion, but arid areas also suffer from occasional heavy rainstorms of great erosivity. An additional criterion is that parent rock should be unconsolidated and porous, generally without soil cover (Harvey 2004). However, Karancsi (2010) mentions badland formation under cleared rainforest and Veress et al. (2012) conclude that badlands can also develop on phyllite. Hungary is situated in the temperate zone with adequate yearly precipitation; consequently, badlands could only form in case of damage to the vegetation cover like exploitation, overgrazing, deforestation or forest fires, most often induced by human action. However, there are several small-scale erosional surfaces which are similar to badlands in the northern volcanic part of Hungary, i.e. in the Mátra, Cserhát, Börzsöny and Tokaj Mountains (Karátson 2006), each of them is framed by forests, which is not usual.

The largest, most impressive, only 'true' badland is situated 3 km north of Kazár village, on the Nyírjes ("birch") Hill (Fig. 18.1), on rhyolithic tuff, surrounded by a system of deep gullies of various length (Fig. 18.7). Sloping from ENE to WSW, the one hectare badland surface descends from 300 to 274 m elevation. Among the rhyolite tuff outcrops of the region gully erosion is the most intense here. First shallow rills form, later evolving into deep gullies incised into the less cemented, nevertheless relatively stable tuff surface, which develops into inter-gully ridges (Fig. 18.8). In some places the ridges have been divided into conical mounds by regression of tributary gullies, perpendicular to the master gullies. In the fine-grain matrix of the tuff the proportion of lithoclasts is about 15–20 %; their size can reach 2–3 cm and by mineral composition they mostly consist of biotite, feldspar and quartz. The majority of lithoclasts are angular and are detached from the matrix as sharp-edge grains, further rolled by sheet flow, which is an effective form of erosion. It is important to emphasize that the linear erosion process is caused by the very fragments which derived from the same rock; no fragments of other rocks are found on the valley floor!

In the badland, both downcutting and regression are active parallel processes. It is not uncommon that small rills are winding, do not show linear alignment, namely the water rills are deterred by more resistant grains, forming interesting" micromeanders".

Little is known about the age of the forms, but researchers agree that they were formed in historical times. Karátson (2006) presented the results of $^{14}C$ analysis of a buried log from the Kazár badland which indicates only 250 years. As a consequence of rare but extremely heavy rainfalls, a gully system may have developed within a short time span.

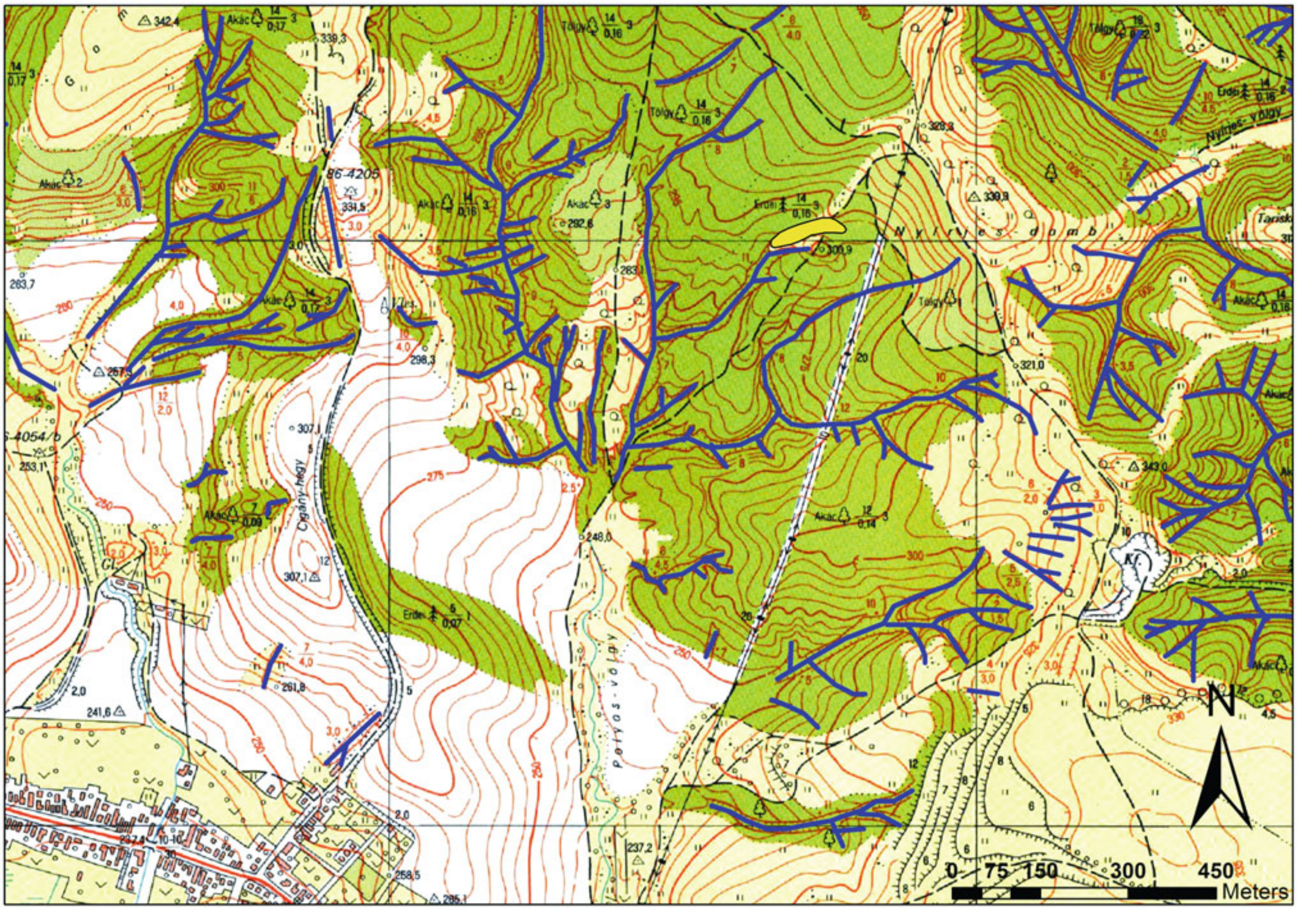


**Fig. 18.7** Map of the eroded surface north of Kazár. *Blue lines* show deep gullies and the *yellow* patch is the badland (surveyed and mapped by Gergely Jakab)

**Fig. 18.8** Rilled and gullied surfaces of the Kazár badland (photo by Zoltán Karancsi)

## 18.6 Conclusions

Rhyolite tuff formations are known from numerous places of the World, including some world-famous regions like Cappadocia in Asia Minor. Naturally, the much less extensive outcrops in the Carpathian Basin have no similar reputation. Nevertheless, they also have aesthetic values and should be considered as geoheritage (Chap. 32). Their vulnerability is high, because on the one hand under present climatic circumstances it is likely that such barren surfaces will be soon reforested, on the other hand the surface is very sensitive. Therefore, the wonderful erosional towers, cones, gullies, and especially the fine microforms can easily fall victim to swarms of visitors. Badlands are often regarded open-air geomorphological laboratories, where landform evolution takes place at much more rapid rates than elsewhere. Because of their unique geomorphological significance, it is desirable that badlands should remain in the present condition, which will be promoted by declaring them as nature reserves, including the above mentioned outcrops of the rhyolite tuff.

Nevertheless, it is fascinating that all over the World badlands are unwanted landforms, signs of land destruction and serious efforts are made to control or rehabilitate them. In Hungary, however, because of their rarity value they are protected and considered to be worthy of preservation in the future too.

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# 19 The Vajdavár Hills: Sandstone Cliffs, Hoodoos and Gorges

Gergely Horváth, Zoltán Utasi, András Hegedűs, and Zoltán Pintér

**Abstract**

The largest sandstone area of Hungary is lying in North Hungary, along the Slovakian border (and even extending into Slovakia). Tertiary sandstones are exposed in more than 1,000 km$^2$ area. As a whole, the sandstone landscape abounds in spectacular macro- and microforms. Due to the tectonic situation and the composition of the rock, high and steep walls with sandstone banks, rock ledges and "loaves" protruding from the walls are typical. Interesting weathering crusts and exfoliation can be studied. Owing to strong incision deep gorges are cut into the strata with characteristic cross-bedding on the walls. Very common forms are funnel-shaped niches and other cavities. Among the forms a big hoodoo is especially remarkable, invoking anthropomorphic resemblances. Among the microforms, precipitated travertine terraces and irregular mounds can be mentioned, originated from oversaturated solution of the carbonate content of the sandstones, also creating thin veneers and larger rimstone dams.

**Keywords**

Sandstone • Cliffs • Hoodoos • Gorges • Rock ledges • "Rock loaves" • North-Hungary

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## 19.1 Introduction

The Vajdavár Hills are a landscape of great diversity, located north of the North Hungarian Range, enclosed by the Ipel'/Ipoly, Rimava/Rima and Slaná/Sajó Rivers in the north and east, and in the west and south by the Cserhát, Mátra and Bükk Mountains (Katona 2006). The sandstone region of 1,000 km$^2$ area is built up of the Pétervására Sandstone Formation.

The contiguous, almost homogeneous sandstone area of low elevation lies between the towns of Salgótarján and Ózd (Fig. 19.1) predominantly with sandstone landforms, studied recently by geologists (Sztanó 1994, 1995) and mostly by geomorphologists (Horváth and Pintér 2001; Pintér 2003; Utasi 2004, 2010; Horváth 2007; Hegedűs 2008).

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**Fig. 19.1** Location of the Vajdavár Hills with distribution of the Pétervására sandstone formation on the surface (according to Sztanó 1994). *Brown spots* are major outcrops. The *red line* indicates the position of the Hungarian/Slovak border

The hilly region, situated 500–700 m lower than the neighbouring mountains, is part of a row of Northern Hungarian basins along the Slovak border and even extending into Slovakia, and shows an array of geomorphological surfaces. The central part is higher than 350 m (highest peak: Vajdavár 529 m) and it is surrounded by lower (240–280 m high) hilly zone of younger age and with less consolidated sediments. The most picturesque features, such as rock cliffs like on Nagy-kő (370 m) and Kő-hegy Hills (405 m), rise right on the boundary between the two zones.

## 19.2 Geological Background

The Late Oligocene to Early Miocene Pétervására Sandstone Formation, now partly exposed on the surface, formed in the embayments of the Parathetys. Littoral-neritic fine to medium-grained, often cross-bedded sands (subordinately gravels) turned by diagenesis into carbonate-cemented sandstone of yellowish grey or greenish grey colour and locally of several hundred metres thickness. The sequence is poor in fossils, only the uppermost strata include molluscs—at places as concentrated as in a lumachelle. At the same time, locally it is enriched in a greenish iron silicate, glauconite, which seldom forms a rock (Hámor 1985; Sztanó 1994, 1995).

The decisive property of the sandstone sequence is the carbonate cementation; average carbonate content is 14–15 % but locally rises to above 25 %. Iron contents are also considerable and variable. Strongly cemented strata, more resistant to erosion, are observable on both high sandstone cliffs and on any other barren surfaces and protrude as spectacular banks or rock ledges (Fig. 19.2), even more emphasized by the dip of these strata. On rapidly eroding strata hollows and niches develop.

Structural properties also contribute to the formation of rock ledges. The shearing of strata along fault lines is common and on shear surfaces steep slopes and intensified erosion induces the preparation of outcrops of strongly cemented and dipping sandstone beds in the form of bastions and towers. The general dip is to the north and northeast and, consequently, on the southern margins scarps developed, while towards the north more gentle surfaces formed along bedding planes prevail.

**Fig. 19.2** Beds with concretions at Ivád (photo by Zoltán Utasi)

## 19.3 Geomorphic Processes and Landforms

Due to tectonic movements and erosion processes, the landscape is minutely dissected by deep and narrow valleys with steep walls. Dissection resulted in high and steep (30° to 40°), often subvertical or step-like cliffs. The structural features (dip, tilting and downfaulting of beds) exert a fundamental control on geomorphic processes and the resulting landforms everywhere.

Characteristic features are the so-called "rock loaves". They are rounded, sometimes almost perfectly spherical concretions arranged in rows and protruding from less intensively cemented beds. The "loaves" are often detached by mechanical weathering from their original locations where cavities are therefore created. On the floors of long retreated erosional valleys with steep walls "loaves" of up to 30–40 cm diameter are scattered. Periodic heavy rainfalls induce floods which roll and round them further (Fig. 19.3). The ledges and "rows of loaves" are often situated in several levels above each other, reflecting major or even minor changes in the environment of the origin. Resistant sandstone caps protect the rock below and hinder the denudation of the latter. As a consequence, the cliff is dissected into protruding bastion-like rock formations.

The geological structure also favours mass movements which occur in high density in this landscape. Landslides are particularly common and slower movements mould typical derasional valleys on the sides of erosional valleys. Undermining of the surface during deep extraction of Miocene coal seams is locally also responsible for triggering surface movements.

Throughout the region the characteristic weathering processes of sandstone can be observed. Weathering produces thin but hard lamellar crusts, which are easily fractured and detached. In addition to climatic influences, this kind of exfoliation mainly arises from geochemical reasons. Due to environmental influences (e.g. acid rain) the fabric and mineral composition of the rock fundamentally changes, the $CaSO_4$ and $Ca(HCO_3)_2$ solutions rise to the rock surface, where gypsum and calcite crystals precipitate and form an ever thickening crust. Being more soluble, gypsum is removed by precipitation, the crust will be relatively enriched in calcite and becomes harder and more compact in fabric. Below the crust a thin veneer of stagnant water accumulates and generates freezing, dissolution and crystallization processes. As a result, the rock below the crust disintegrates into grains and the crust is gradually detached in scales over an ever increasing surfaces and is finally removed. The black stain of the gypsum-rich crust is explained by the contaminations contained as xenoliths in the gypsum crystals. Thus intensive weathering essentially results from the decomposition of carbonates (Török 2003).

It can be observed that on higher rock surfaces of good ventilation, exposed to wind, thick weathering crusts are uncommon, but they more typically occur in wind shelters, where the drying and eroding influence of wind is less efficient. On the other hand, in such spots weathering due to

**Fig. 19.3** "Loaf" tumbled out from the cliff above the floor of the Bugyizló Valley (photo by Gergely Horváth)

the presence of condensed water occurs since atmospheric vapour precipitates in the fissures and depression on the sheltered sides of blocks.

The biota also play a part in the lamellar exfoliation of sandstone surfaces. The colourful lichens weave through the surface by their hyphae, intrude to some depth and loosen the rock. In addition, lichens produce chelates (i.e. aromatic organic compounds which incorporate metal ions) and thus multiply the rate of weathering. After detachment of scales, on the fresh surfaces the same processes began to operate anew and ensure the apparently slow but steady denudation of sandstone surfaces.

The development of subvertical sandstone cliffs took place along structural lines in the warmer and more humid Pleistocene interglacials, when increased fluvial erosion broadened the valley floors and undercut valley margins. The bed outcrops on the resulting steeper slopes were further moulded by cryogenic processes under the cold climate of glacials and parallel slope retreat ensued (Utasi 2010). Equally picturesque are the steep-walled sandstone gorges, which actively develop even today.

## 19.4 Selected Landform Examples

The most spectacular landforms occur where the barren bedrock patches deprived of vegetation and soil after forest cutting are exposed on the surface. Among the best examples are the Noah's vineyard at Istenmezeje, the cliffs of Nagy-kő at Bükkszenterzsébet, Kő-hegy at Szentdomokos and Nagy-Lyukas-kő at Ivád, the Leány-kő and Morgó Gorge of Nemti and the Bugyizló Gorge of Zagyvaróna and these are treated below in more detail.

### 19.4.1 Noah's Vineyard

Noah's vineyard (Fig. 19.4), part of the Tarna Region Protected Landscape, is a barren and dissected sandstone cliff. It stretches ca 500 m along the main street of the village Istenmezeje and rises 50–70 m above the valley floor of the Tarna River, which is at 220 m elevation above sea level. Porous coarse-grained, strongly cemented greenish glauconitic sandstone banks of 2–8 m thickness protrude from the cliff as a series of rock ledges resistant to erosion. The formation was named after its concretions of special arrangement. Along cracks calcareous solutions infiltrated and precipitated. Their cementing effect was the most intensive at the junctions of joint systems, where hardened rock masses, some tens of centimetres to some metres across, resisted further weathering and erosion and stick out not only in horizontal rows, but are also arranged regularly in vertical direction. From the distance they look like regular rows of vinestocks on the hillslope (Utasi 2004; Selmeczi 2012). Popular legend explains that once two tired and thristy wanderers, Jesus and the Apostle Peter, were refused to be given grapes here and turned the vineyard to stone. In the 18th century the forest was cleared and the soil cover eroded, allowing for the development of the rock outcrops. The cliff is a valuable habitat for lichens of great diversity, endangered by the rapid spreading of bushes from the direction of the hill summit.

**Fig. 19.4** Noah's Vineyard at Istenmezeje (photo by Zoltán Utasi)

### 19.4.2 Nagy-kő

One of the most spectacular cliffs of the sandstone region is Nagy-kő (Fig. 19.5), 3 km NNW from Bükkszenterzsébet, on the southern margin of Vajdavár Hills. Its top (370 m) rises in a 70 m high and 200 m wide cliff above the surrounding terrain and provides an excellent view upon the Karancs, Mátra and Bükk Mountains. Its origin is also due to differential weathering. Through intense headward erosion high-gradient ravines and valleys dissect the surface of uplifted and tilted sandstone sequences into a series of rock bastions. Nagy-kő is one of such bastions isolated from an interfluvial ridge. Its thick banks have a northern (10°) strike and 30° dip. At a closer look, from the steep cliff of uniform appearance rounded "loaves" and flat concretions stand out in regular rows. In the late Bronze Age earthworks may have stood on the top.

**Fig. 19.5** Rows of concretions on Nagy-kő cliff at Bükkszenterzsébet (photo by Zoltán Utasi)

**Fig. 19.6** Niches on the Nagy-Lyukas-kő cliff at Ivád (photo by Zoltán Utasi)

### 19.4.3 Kő-hegy

North of Szentdomonkos, at 1.5 km distance, a characteristic triple cliff rises, with the highest part reaching 400 m in altitude and the south-looking walls up to 30–40 m high. Its sediments deposited in a quiet environment, indicated by the clearly visible thin bedding of northern (15°) strike and 30° dip. Lenses more prone to weathering have been removed and "rock loaves" dropped out leaving behind hollows, locally of considerable size (1–2 m across), which are deepening by further weathering. Exfoliation of sandstone is striking on the cliff of Kő-hegy.

### 19.4.4 Nagy-Lyukas-kő

In the valley of the Szénégető Stream, 3 km from the village Ivád, Nagy-Lyukas-kő is another spectacular cliff of sandstone which resulted from sedimentation in a quiet deposition environment. As a consequence, its surface is fairly smooth with small-size concretions and funnel-shaped niches in the wall (Fig. 19.6). The name, meaning "big stone with a hole", refers to a small cavity. It is interesting to note that on the top of Nagy-Lyukas-kő animals were offered to gods in pagan times. Grooves once conducting blood are still discernible on the rock.

### 19.4.5 Leány-kő

This typical hoodoo (rock pinnacle), one of the most beautiful landforms of the hills, is found 2 km northeast from Nemti (Fig. 19.7), rising above the Zagyva River graben. Differential weathering is typical of the southern slopes of asymmetrically tilted sandstone ridges of imbricate structure, where scarps of harder banks are better preserved than the intercalated softer beds (Pintér 2003). At the eastern end of such a ca 50-m-long vertical sandstone cliff, 60 m from the summit of Leány-kő Hill (357 m), a hoodoo of irregular oval shape, invoking anthropomorphic resemblances (suggested by the Hungarian name: "demoiselle"), rises to 4.5 m height from the direction of the wall and 2 m higher viewed from the direction of the steep slope. Originally it was a protruding part of the sandstone cliff, but slope retreat along the joints of the wall gradually isolated it from there. A cap on the top of the thick-banked sandstone with limonite zones

**Fig. 19.7** The hoodoo of Leány-kő at Nemti (photo by Zoltán Karancsi)

efficiently protects the less resistant beds below. The softer beds are eroded all around and this shows a trend towards the formation of a "puppet" driven by differential weathering. The relatively intact preservation of Leány-kő is due to difficult access which has so far precluded human impact.

## 19.5 Gorges

In spite of the high infiltration capacity of the rock, after intense rainfalls, cloudbursts, or prolonged rainy periods surface runoff intensifies and generates rapidly incising rills and gullies. The processes are supported by young uplift and heavy dissection and result in erosional gorges of variable sizes and planforms (Pintér 2001, 2003; Hegedűs et al. 2008; Utasi 2010). Valley development is intensive but typically cyclic: long periods of inactivity, without water flow, are suddenly succeeded by a flow episode after an extreme rainfall event. On such occasions the cross-sections of valley segments are completely remoulded: some are infilled in great thickness within a very short spell and, in contrast, deposits are transported from other valley segments.

One of the most spectacular ravines is the Morgó Valley of north to south alignment near Nemti. The broad valley narrows down after 300–350 m length and becomes a gorge of only 180 m length, 6–7 m depth and 0.5 m width at the narrowest section, with steps in the high-gradient floor. Its origin is due to a high-positioned sandstone block hindering rapid valley incision and retreat, but gradually cut across by the watercourse. During incision, remarkably steep and high sandstone cliffs came about. The water flow carves potholes on smooth surfaces and wavy scallops into the cliffs (Fig. 19.8), exposing cross-bedding and limonitic zones (Pintér 2003). The harder sandstone banks appear as scarps on the valley floor and divide it into higher and lower gradient reaches. The flat-floored reaches between steps are deepened into pools, which are filled by the abundant load transported by the watercourse during floods. Fill is often induced by the impoundment of the reach immediately upstream the step, for instance, by accumulating driftwood.

Another picturesque and much longer valley (with distributaries several kilometres long) is the Bugyizló between Rónafalu and Zagyvaróna, where intensive erosion formed a valley floor of highly variable width: in some V-shaped segments there is a no floor at all, while on others the broad floor reminds of wide corridor. Steps are visible on thicker and more resistant sandstone banks. In the narrow gorges of intermittent ravines steps of variable height (from tens of centimetres to 2 m), where waterfalls emerge in humid periods, undercut the base of higher and steeper steps and form minor potholes by evorsion. Evorsion results in episodic topples of undercut beds, inducing the retreat of steps and waterfalls. Large toppled sandstone blocks are observed below most of the channel steps. On the low-gradient segments the valleys are often filled by thick, loose and fresh deposits, indicating that collapses are also characteristic in other valley segments where valley sides are undercut by—often meandering—watercourses (Fig. 19.8). The detached sandstone blocks are gradually disintegrated and accumulated on a shorter valley segment in a terrace-like flat surface. After intense or prolonged rainfall the stream incises anew into this surface (Hegedűs et al. 2008).

## 19.6 Minor Features

The carbonate cementation involves the development of a range of minor features too. Similar to karstification, the equilibrium $CO_2$, which keeps $CaCO_3$ in solution is released from groundwater and from the oversaturated solution carbonates precipitate in the form of small terraces, irregular mounds, primarily at present or past springs. On the rock walls and their vegetation thin veneers of travertine are

**Fig. 19.8** Spectacular gorge of the Morgó Valley at Nemti (photo by Zoltán Pintér)

created. With time they are getting thicker and thicker and even large rimstone dams may take shape. Even at the bases of sandstone "loaves" there are dripstone-like precipitations and calcite crusts. On the cliffs cross-bedding originated by tidal deposition on the shallow and flat shelf of the Paratethys (Sztanó 1994) can be studied.

As seen above, differential weathering has commonly produced cavities of various size, including rock shelters carved by the lateral erosion of intermittent watercourses and caves of corridor and niche type which resulted from the combinations of different weathering processes (insolation, pressure release, crystal growth, solution, hydratation, hydrolysis, biogenic weathering etc.). Some of them were further deepened by humans. In some cavities of sandstones with higher carbonate contents dripstone formation is still active, while elsewhere a carbonate crust with lichens covers the walls. Cavities are generated by weathering due to condensed water. Water evaporated from overhangs precipitates or even freezes on the ceiling of the small cavity; as a consequence, scales are detached from the ceiling and the cavity gradually widens and deepens. On gentle walls and sandstone scarps the hollows left behind by detached "loaves" expand relatively rapidly compared to the neighbouring cliff surface and lead to the formation of tafoni. The merging of neighbouring tafoni results in the formation of small basins a few metres across.

## 19.7 Conclusions

One of the regions of Hungary least known and visited by tourists, the Vajdavár Hills are rich in geomorphosites. Nature conservation turns this disadvantage into an advantage. As the area is not a popular destination the physical environment can be preserved in an almost intact condition. Increasing tourism could put higher pressure on geomorphosites too. Only a small part of the local geoheritage is protected in the Tarna Region Protected Landscape (area: 95.7 km$^2$, established in 1993). More comprehensive conservation measures would be necessary. Fortunately, the cliffs (Nagy-kő, Kő-hegy, Nagy-Lyukas-kő) will preserve their features over longer periods, but in other cases (for instance, in Noah's vineyard) rapid plant succession is a hazard to spectacular rock formations, which could disappear within decades.

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# Baradla-Domica: Large Cave System on the Hungarian-Slovak Border

20

Márton Veress and Zoltán Unger

**Abstract**

In this chapter the longest stream cave in Hungary, located in the Gömör-Torna Karst (Slovak: Gemersko-Turniansky kras, Aggtelek Karst and Slovak Karst) is presented. The caves of this karstic system have been included into the UNESCO World Natural Heritage list. Among the most spectacular dripstone caves in Europe, Baradla Cave has two levels and is defined as an erosional through cave. Its upper level constitutes the Main Branch with several secondary branches, while the lower level includes the Lower Cave, at and below the karstwater table, controlled by the waters arriving from the level above. Among the major morphological features of the upper level are ponors, halls, debris mounds, gravel terraces, the bed of the Main Branch, and various speleothems (stalactites and rimstones). The ancient dissolution spongework was enlarged as it further developed due to the erosive effect of the water flowing from the nearby non-karstic terrain through ponors. At first, the ponors were fed from the north by the waters from a surface of southern exposures and later from the southern region (the Galyaság), which has a northern inclination. Two erosional segments are identified on the upper level of the cave with an accumulation wedge between them.

**Keywords**

Dissolution • Erosion • Accumulation • Ponor • Speleothems • Main branch • Lower cave • Gömör-Torna karst (Gemersko-Turniansky kras)

## 20.1 Introduction

Karst areas in Hungary extend over 1,350 km$^2$ (1.45 % of the total area). They mainly occur in the Transdanubian and in the North Hungarian Mountains in patches of variable size.

One of these island-like occurrences is the Gömör-Tornai karst in the northeast which also extends over the area of Slovakia. There are major karst features (dolines and uvalas), the international importance of this area, however, is due to the large number of its large-size caves (the Béke, Vass Imre, Kossuth and Baradla-Domica cave systems). It is because of the caves that it became UNESCO World Natural Heritage. For size, beauty and richness of features the Baradla-Domica cave system is of the greatest significance.

## 20.2 Cave Formation

Caves develop by dissolution and erosion. While dissolution caves are generated under the karstic water table, erosion caves are due to the abrasive effect of waterflow into the caves through ponors. Erosion caves are either inflow or through caves. The latter are accessible for humans from the ponor to source. If the base level of erosion sinks, cave

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ponors develop and the through cave is transformed into a multi-storey cave system.

Ponors emerge on the edges of allogenic karsts and a surface watercourse becomes a beheaded stream if its catchment area is small. Runoff into ponors increases the catchment area by backward erosion, and, consequently, water inflow will also grow. The width of the cave gallery is controlled by the amount of water influx. The cross section of the deepening cave gallery will acquire a triangle shape due to increased water flow (Jakucs 1977). On covered karst, the deepening valley will turn into an epigenetic valley in which ponors develop on the rock boundary. As the cave system is evolving, surface runoff expands its catchment area and the cavities take on trapezoid cross sections.

The water flowing through the karst produces minor features of both dissolutional and erosional origin. Solution or blind pockets form by mixture corrosion in caves developing along the karstic water tables. When the galleries are partly filled by sediments, the water is forced to the ceiling, dissolves the rock and forms anastomoses and meanders (ceiling channels) in that order. Anastomoses are parallel channels on the ceiling and meanders are overgrown anastomosing channels, some reaching even tens of centimetres' width. Cave karren, caused by flowing water, emerge on bare surfaces in caves, while scallops are pockets several (tens of) centimetres across on the wall, generated by either downflowing water (through erosion and corrosion) or by the dissolution of surfaces covered by sediments (Slabe 1995).

Streams in through caves cut channels into the bottom of the galleries. Vertical chimneys in the bedrock further develop into cave ponors. The scarps in the walls of the gallery are solution notches, due to dissolution either at the water table or at the fill level. If the water influx of the stream fluctuates during incision, scour grooves occur. If sediment input decreases and water flow intensifies, deposits are flushed out of the cave. On the walls of cave galleries the cemented deposits are left behind as terraces and terrace fragments. The broadened cave ceilings crumble, ceiling domes are created and the fallen debris accumulates in mounds. The gallery fragments are connected to form halls.

The development of erosion caves is facilitated by ponor formation in blind valleys at the boundary of the karstic terrain. It is quite common, also in the Aggtelek Karst, that paleodolines are filled during burial. Ponors can occur on the margins of paleodolines, from where they transport sediments into the karst and generate a closed feature, which is called depression of superficial deposit (Veress 2009). The ponors can be totally eroded, truncated or the galleries fill up with sediment and ponds occupy the depressions. In the cave the dripping water forms various stalactites and travertine and the cave stream builds rimstone dams.

## 20.3 The Karst Region

The Gömör-Torna Karst (area in Hungary: 200 km$^2$, elevation 300–500 m) is located in Northeast-Hungary, along the Slovak-Hungarian border, between the Sajó, Bódva and Tarna Rivers. Continued in the Szilice (Slovak: Silica) Plateau in Slovakia, the Hungarian segment is divided into the plateaus of Alsó-hegy ("lower mountain") and the dissected Aggtelek Mountain (Fig. 20.1), the plateaus of which host Baradla Cave. The mountains are rich in various karst features of considerable density and size compared to the small extension of the mountains. The Aggtelek Karst is built up of Upper Permian–Lower Triassic gypsum and anhydrite and Triassic carbonates, a part of the Szilice Nappe fragmented into klippen (Less 1998).

The northern plateaus (Haragistya, Alsó-hegy) are bare karst, dissected by structural valleys of the Jósva, Ménes and other streams and characterized by the presence of karren (rillenkarren, grike karren, root karren), dissolution dolines, uvalas, open shafts, dissolution and erosion caves, rimstone dams. Patches of covered karst (cryptokarst and concealed karst), important for the evolution of Baradla Cave, are preserved in the south, among the paleodolines of Galyaság. On cryptokarst patches ponors, depressions with superficial deposits and blind valleys occur, while on concealed karst subsidence dolines are found. Solution dolines are widespread on the bare karst patches.

## 20.4 The Cave

Numerous researchers have contributed to the exploration of the longest stream cave in Hungary: Vas (1831), Dudich (1930), Kessler (1933), Cholnoky (1935), Jakucs (1952, 1956, 1999), Szentes (1965), Szenthe (1970), Kovács (1970), Lauritzen and Leél-Őssy (1994), Bosák et al. (2004), Berényi et al. (2006), Veress (2012).

The cave developed in Middle Triassic Wetterstein and Upper Triassic Steinalm (a short section in Guttenstein) Limestones. Its length is 25 km (18.8 km in Hungary), catchment area is 40 km$^2$, encompassing partly covered (cryptokarst) and bare karst. The cryptokarst feeds karstwater through ponors and the bare karst does so by infiltration. The cave system has an upper and a lower storey. The upper level of the two-storey cave system is older and better developed. The height difference between the upper and lower level is at least 35 m, because Szenthe (1970) found a permanent stream under the Óriásterem ("giant chamber") ponor, which he considered to be part of the Lower Cave hydrological system.

The dye tracing by Szenthe (1970) proves that the lower level is split into two, probably communicating, caves: the Rövid-Alsó ("short lower") cave between the Sárkányfej

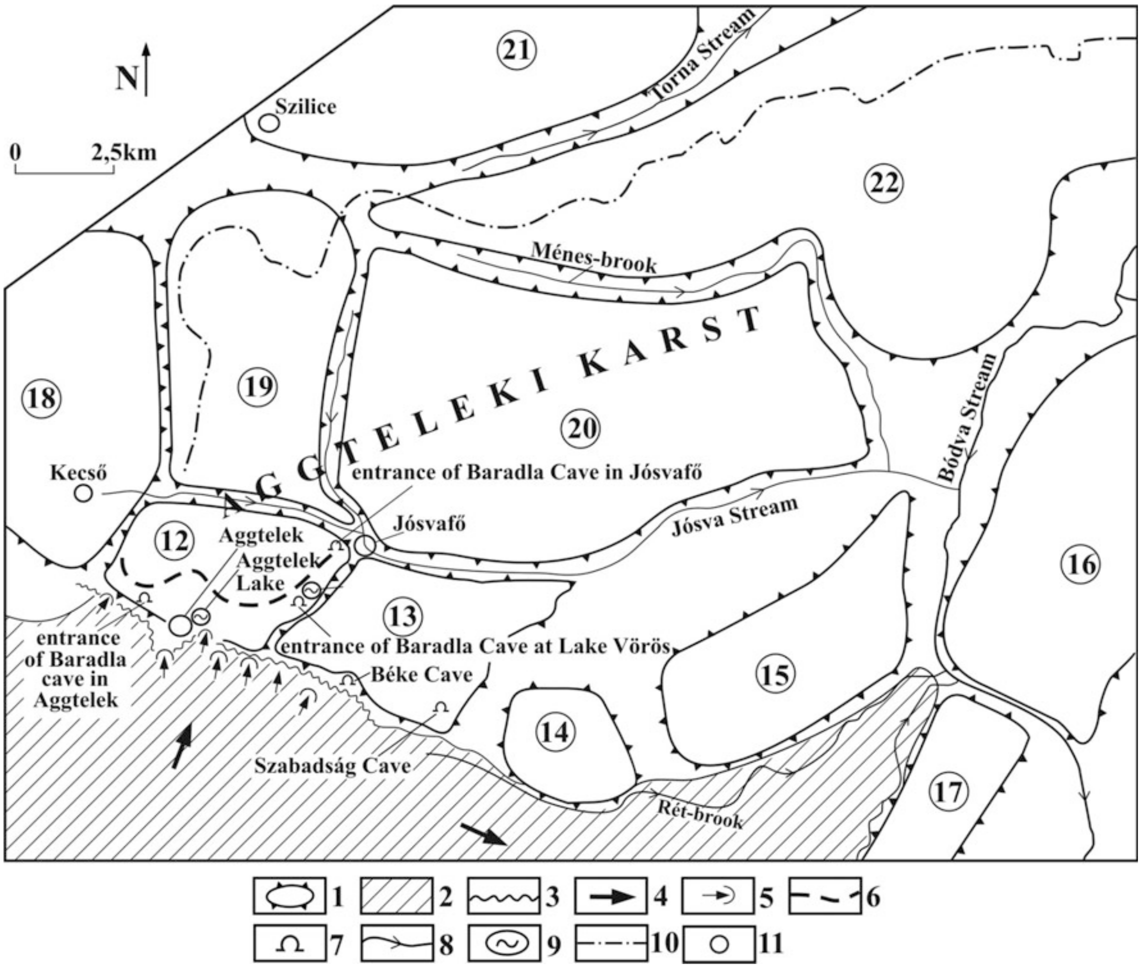


**Fig. 20.1** The Gömör-Torna karst and the line of subsurface capture (Veress 2012). *1* Karst plateau; *2* cryptokarst; *3* line of subsurface capture; *4* dip of cryptokarst; *5* ponor; *6* approximate site of Baradla Cave; *7* cave entrance; *8* stream; *9* lake; *10* national border; *11* village; *12* Aggtelek Plateau; *13* Mount Pitics; *14* Teresztenye Plateau; *15* Eastern Galyaság; *16* Szalonna Mountains; *17* Rudabánya Mountains; *18* Kecső Plateau; *19* Haragistya Plateau; *20* Szinpetri Plateau; *21* Szilice Plateau; *22* Alsó-hegy (Lower Hill)

("dragon head") Ponor and Baradla Lower Gallery and the Hosszú-Alsó ("long lower") cave between Négerkunyhó ("Negro's hut") Ponor and Jósva spring. The Lower Cave is active, partly or totally filled by water, and its base level is the Jósva Stream. Access to the Lower Cave system was only partially successful, since it is only possible if the water table is lowered by pumping. Its water is recharged from karstwater and from the upper level through ponors.

The upper level is composed of the 7-km-long Main Branch (Fig. 20.2) and secondary branches. The Main Branch of trapezoidal cross-sectional profile (Takács-Bolner 1998) has an east to west and, beyond Vörös-tó ("red lake"), a north to south segment. Its average width—without debris fall—is 10.5 m and passage height is 7–8 m, locally, for instance, in front of the Vaskapu ("iron gate"), reaching 50 m, and its width is 60 m in Óriások terme ("chamber of giants"). The cave stream is aptly called Styx.

Linked to the Main Branch from the south, the secondary branches (Jakucs 1984) are the 7-km-long Domica Cave in Slovakia, fed by Ördög-lyuk ("devil's hole"), Domica-lyuk, and Demnik-lápa (ponor). The Retek-ág ("raddish branch") is 2,748 m long, 4–5 m wide and was formed under the ponors Nagy- and Kis-Ravasz-lyuk ("big sly" and "little sly hole") (Ponors are called sly holes by locals.) The 1,125.6-m-long Török-mecset ("Turkish mosque") branch is the continuation of the Zombor hole (Fig. 20.3). The Oszlopok csarnoka ("hall of columns") branch (720 m long) was developed under the Little Baradla ponor just like the Denevér ("bat") branch (338 m long) under the Acheron ponor. The latter branch, connected to the upper storey of the Main Branch, is no more active. Similarly, the 318-m-long Róka-barlang ("fox cave") belongs to the upper level and no ponor is known from here. The Vörös-tó branch was damaged during the building of a new entrance. There is another inactive branch, situated 5,200 m

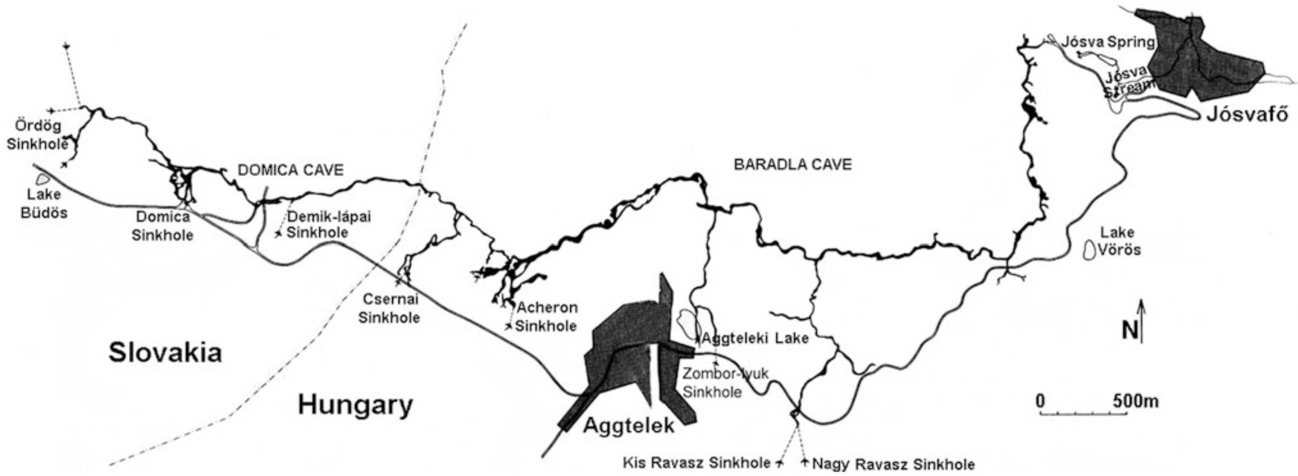


Fig. 20.2 Location of Baradla Cave (after Kordos 1984)

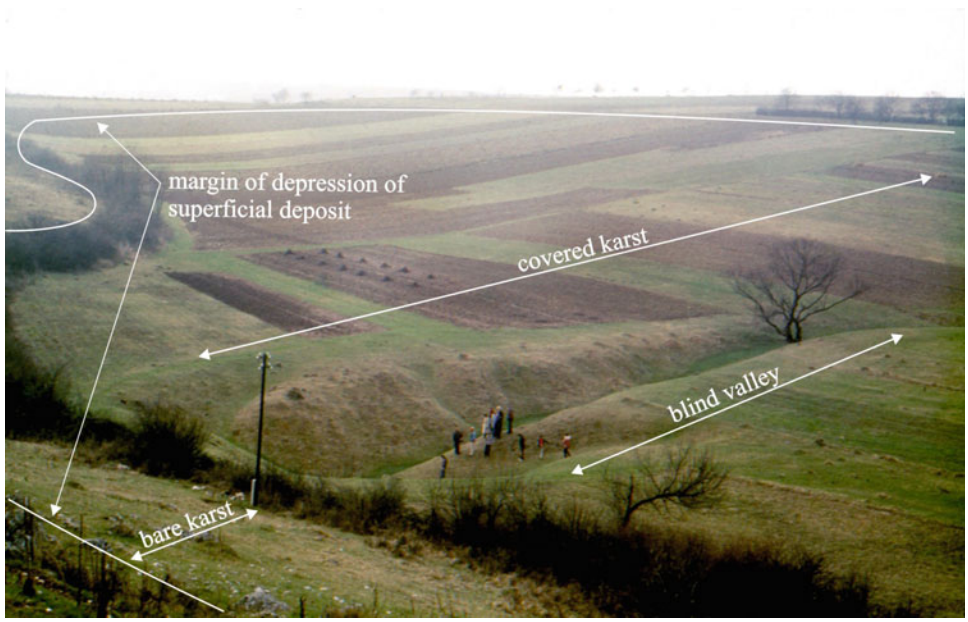


Fig. 20.3 The Zombor hole ponor on the boundary of the karstic terrain and its bearing depression with superficial deposits

from the entrance of the Aggtelek Cave: Arany-utca ("golden street") branch. The role of this branch has not been revealed yet; probably it was the continuation of the Main Branch, or it is here that Béke ("peace") Cave joins Baradla Cave (Jakucs 1975). The inactive Raisz branch with a mapped length of 120 m developed in the proximity of the Jósvafő entrance (Dénes 1970). Its ponor conducts water into the Lower Cave.

The cave system has several inactive ponors, both on the surface, such as Büdös-tó ("stinking lake") and Aggtelek Lake and underground in the Main Branch (see below). Except for Domica cave, the galleries of triangular cross section are narrow with pointed ending. The Retek-ág branch is outstanding because the upper part of the branch has a triangular profile, which turns trapezoid towards the lower levels (Jakucs 1975).

## 20.5 Cave Morphology

The minor cave features are scallops, solution pockets, chimneys of various sizes, ceiling channels and karren. The large and medium-size features are cave beds, cave halls, domes, debris cones, ponors, pebble terraces, scour grooves, pedestal rock and those with the smallest dimensions: precipitations. The most frequent minor features are the scallops (Fig. 20.4), found in various morphological positions. Solution pockets and blind chimneys are found on the roof of the Main Branch, but also occur in secondary branches, e.g. in the Retek-ág branch. Their origin is extremely variable in Baradla Cave. They could have formed by mixture corrosion or with the water pressed to the roof of the cave. According to

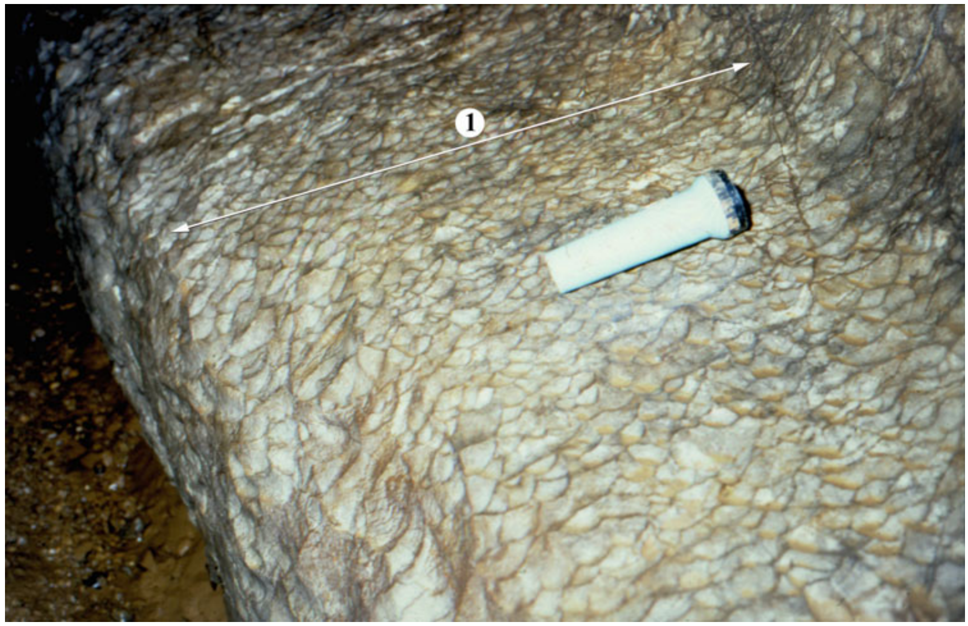


**Fig. 20.4** Scallops on the lower plane of a notch (*1*) (photo by Márton Veress)

Jakucs (2005), they are attributed to the solution effect of infiltrating waters, but they could have also developed from paleokarstic hollows which lost their fill. The ceiling channels are also features of the roof of the Main Branch, especially characteristic on the Aggtelek segment of the cave. On the walls three features occur locally: rinnenkarren (Fig. 20.5), wall karren, and meander karren (Gruber 1999). Gruber (1999) also described ceiling karren which derive from selective solution (Jakucs 2005). The fill of paleokarst hollows was more dissoluble than the homogeneous rock.

A pedestal rock is thinning downwards because of abrasion and erosion by flowing water. Scour grooves are characteristic for secondary branches (Fig. 20.6). In the cave, especially along the Main Branch, long channel beds developed. They were buried under debris along some sections. The main halls in order from the Aggtelek entrance are called Csontház ("charnel", 100 m), Fekete ("black", 150 m), Tánc ("dance", 200 m), Hangverseny ("concert", 300 m), Nagy ("big", 700–800 m), Hősök terme ("heroes' chamber", 1,100–1,200 m), Libanon hegye (Mt. Lebanon, 1,400–1,500 m), Reményi (1,500–1,600 m) halls, and finally Óriás terem ("giant hall"), Kafka and Vetődéses ("faulted") halls in the vicinity of Jósvafő. Chambers also occur in the Retek branch and the Lower Cave.

Ponors are common in the upper level, often occurring in pairs, one below the other. While those in higher position are not active, the lower ponors are active today. Major ponors are Vaskapu, Tündérvár ("fairy castle"), Pluto's organ. The elevation difference between the active and inactive members of ponor pairs is 1.5–2.5 m. The most important is the Óriásterem ponor, acting as the main drainage path leading to the Lower Cave.

**Fig. 20.5** Karren in the cave (photo by Márton Veress)


**Fig. 20.6** The mirror notch (*1*) from the Retek Branch (photo by Márton Veress)

**Fig. 20.7** Terrace remnant (photo by Márton Veress)

In the cave gravel terraces (Fig. 20.7) with upward refining gradation (Kovács 1970) are well developed, particularly under stalactite balconies or where the debris was cemented. The inclination of the gravel terrace surfaces is lower than that of the current bed. Among the wealth of speleothems, most spectacular and widespread are the carbonate precipitations, dripstones. The largest stalagnate, Csillagvizsgáló ("observatory") is 25 m high. Major rimstone dams (Fig. 20.8) occur at the end of Törökmecset branch (at 800 m from the entrance), another at 2,200 m, and at the end of Retek-ág branch (at 2,500 m).

## 20.6 Origin of Baradla Cave

As it becomes obvious in secondary branches, the galleries are preformed by fractures (Berényi et al. 2006), which control the alignment of the main branch too. According to Szenthe (1970), the faults forming the gallery strike N-S and NNE-SSW. At the junctions of several fractures dissolution intensifies and halls are formed where several cavities merged. Extensive debris accumulation (debris mounds) could only take place after the halls were formed and their expansion made their roofs unstable. Either ceiling domes, or —if entire rock strata were detached in fragments—bedding-plane roofs resulted. The spherical cavities on the roof of the Main Branch indicate dissolutional origin under the karst-water table.

According to Jakucs (1956), the surface watercourses running towards the north induced ponor formation at the contact of the karstic and non-karstic rocks on the Aggtelek

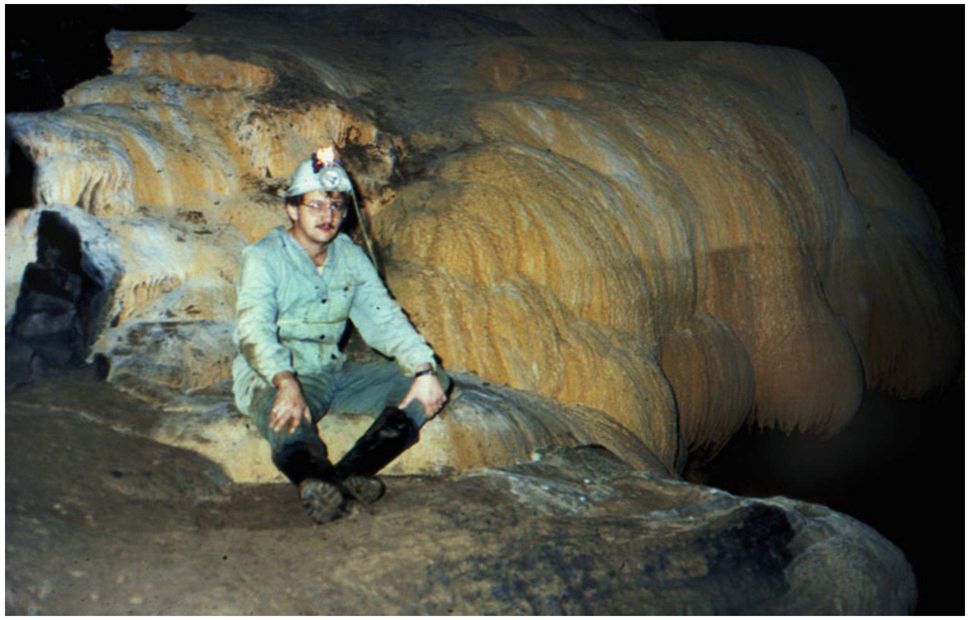

**Fig. 20.8** Sinter formation

Karst, facilitating erosion to replace dissolution in the evolution of cavities. The present-day ponor rows on the floors of epigenetic valleys resulted from repeated subsurface capture.

Further research proved that the karst region was tilted to acquire an inclination towards the SSW during the Plio–Pleistocene period (Sásdi 1990). On this surface the watercourses coming from the north accumulated a gravel sheet (Sásdi 1990) and incised into it. At that time the first generation of the ponors were formed along the present-day line of capture (Veress 2012). These ponors have already been eroded (truncated or filled up). According to Veress (2012), the current morphology is the outcome of Galyaság Plateau tilting towards the north. On Galyaság, covered karst could only be preserved in the area of paleodolines. Several paleodolines under the superficial deposits could be identified by our geophysical measurement—for instance, near Zombor hole—Fig. 20.3. In the area of paleodolines water inflow into the ponors created depressions with superficial deposition, constituting the catchment area of Baradla Cave.

Jakucs (1984) distinguished at least three phases of erosion and accumulation in the Main Branch of Baradla Cave. The ceiling channels show infilling and the gravel terraces prove erosion. Some segments of the Main Branch and the western part of the cave system are composed of at least two levels. This is evidenced by the fact that the Main Branch has locally two levels, in Oszlopok csarnoka ("hall of columns"), Denevér, Acheron, Róka ("fox") branches and in the Münnich passage (Fig. 20.2) (Kovács 1970). According to Berényi et al. (2006), the former corrosional cave or cavity system was filled up, subsequently the deposits were flushed out during the erosion phase. The investigations quoted do not exclude the possibility of repeated filling and erosion phases in the history of the cave.

Based on surface morphology, Veress (2012) divided cave development into four phases (Fig. 20.9):

- dissolution phase;
- the first generation of ponors followed by an erosion phase;
- accumulation phase;
- the tilting of Galyaság Plateau led to the formation of the second generation of ponors, which induced a second erosion phase, lasting to modern times, with the development of blind valleys on surface and depressions with superficial deposits in their environs.

The different parts of the cave system are at different stages of evolution. The trapezoid profile of the Main Branch shows that the watercourse(s) which created the cave was/were captured subsurface at a relatively later date. Previously carving a valley on the surface, it/they consequently had had a significant discharge when captured subsurface. Therefore, the upper part of the cave also broadened. Some of the secondary branches developed in adjustment to the upper level of the Main Branch (Denevér branch). Other secondary branches (Törökmecset) were formed as allogenic types, characteristic for the karst margin in the prolongation of ponors. Based on their cross-sections, the watercourses that created these secondary branches had already been captured subsurface at the beginning of their development, contemporaneous with the erosion of the Main Branch or after its infilling. This facilitated sediment transport and the increase of recent erosion in the Main Branch. The present-day processes in the Lower Cave are most similar to the early evolution of the Main Branch. The surface streams are

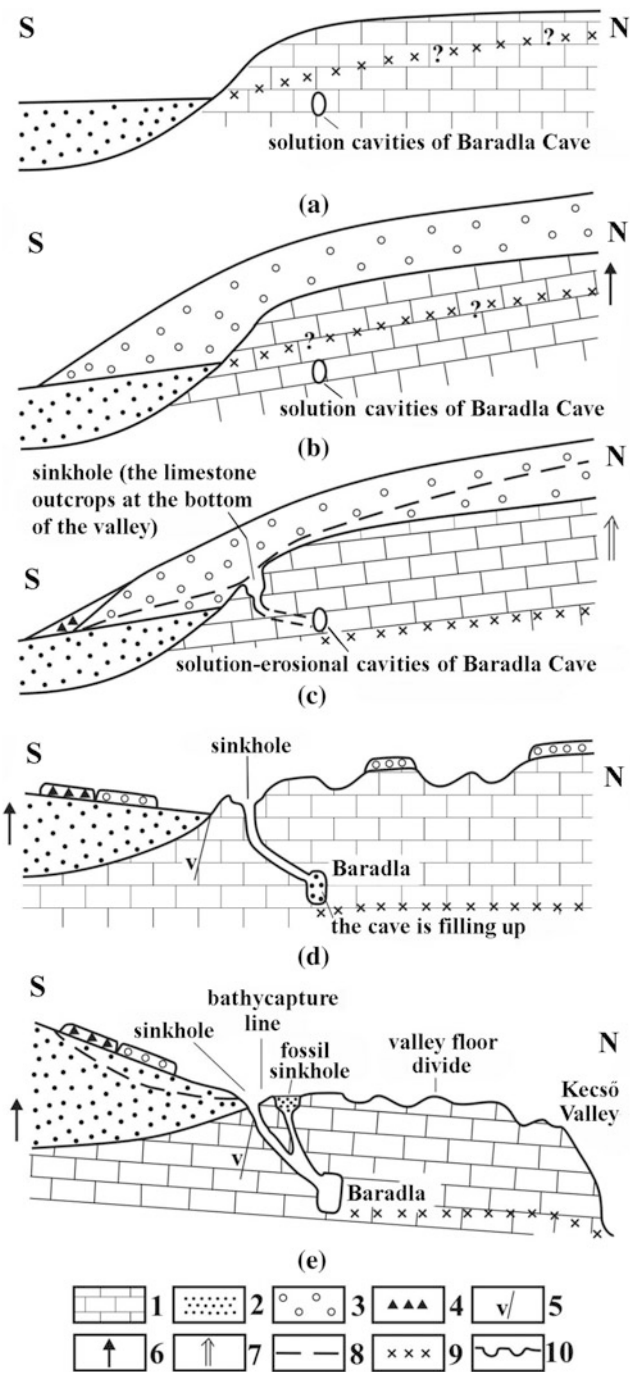


**Fig. 20.9** Evolution of Baradla Cave (Veress 2012). **a** Upper-Miocene: the Aggtelek Plateau is bare karst, Edelényi Formation developed south of the plateau. **b** Plio-Pleistocene: the plateau tilted to SSE, cryptokarsts are buried under the Borsodi Gravel Formation. **c** Pleistocene (?): the plateau is uplifted, valley formation, where the valley bottoms reach the limestone and karstwater table sank deeper than the valley bottom; (first generation) ponors develop. **d** Pleistocene (?): south of the plateau margin the present-day cryptokarst uplifted, and in the epigenetic valleys of the plateau solution dolines developed on valley bottom. **e** Upper Pleistocene and Holocene: tilting continues, plateau slopes to the north, second generation of ponors along the margins. *1* limestone; *2* gravel, sand, clay (Edelény Formation); *3* gravel (Borsod Gravel Formation); *4* redeposited gravel; *5* fault; *6* tilting; *7* uplift; *8* valley bottom; *9* karstwater table; *10* dolines and uvalas

drained into the karst cavities through the ponors of the Main Branch.

From the paleomagnetic analyses of deposits from Münnich passage (Bosák et al. 2004) and the dripstone analyses by Lauritzen and Leél-Őssy (1994), it is presumed that the infilling period encompasses 100 ka. Jakucs (1984), however, estimates the cave to be 1.5–2 Ma old.

## 20.7 Conclusions

Baradla is a stream cave of two levels; mostly of erosion type. The upper level is a through cave, formed from at least two different pre-existing levels. The Main Branch of the upper level shows subsurface capture processes at a later stage, while most watercourses of the secondary branches, as indicated by their cross-sections, suffered subsurface capture in earlier stages of their evolution. On the upper level there are numerous ponors. Debris mounds, halls, gravel terraces, and precipitations are prevailing features. The lower level is under transformation to an erosion cave at present. The upper level was developed from a dissolution cavity system turning into an erosion cave with two erosion phases separated by an accumulation period. Its vast dimensions, as compared to the extension of the karst area, are due to the large catchment area, earlier located to the north of the cave, and now shifted to the south. The deposits of the present catchment area occur in paleodoline fills.

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# 21 Uppony Mountains: High Geomorphological Diversity Within a Small Area

Gábor Demeter and Katalin Szalai

**Abstract**

Merely of 454 m maximum elevation, the geologically and geomorphologically diverse and still dynamic Uppony Mountains is relatively rich in geomorphic features (caves, pseudo-caverns, gorges, horsts, scarps, landslides, karren, natural outcrops) created by diverse processes (valley incision, derasion, mass movements, karst formation, periglacial, pedimentation). This small mountainous area with its intricate geological structure is ideal either for field investigations in order to obtain broader knowledge on the evolution of North-Hungary or to study both convergent and divergent evolution of forms generated on different rocks. Using a comparative approach, with an outlook to its environs, the quantification and numeric analysis of morphometric and lithological features combined with the traditional descriptive geomorphic approach promotes a better understanding of the area so important in the paleoenvironmental, paleogeographical reconstruction of North-Hungary.

**Keywords**

Paleozoic massif • Piedmont • Imbrication • Gorges • Landslides • Caverns

## 21.1 Introduction

The Uppony Mountains of 240 km$^2$ area are located in northeastern Hungary, in the Bükk Mountains northern foreland. It is mainly composed of slightly metamorphosed Paleozoic siliciclastic rocks and limestones with distant Alpine contacts, surrounded by semi-consolidated Neogene marine sediments. Its uniqueness is primarily due to the particular geology: both karstic and non-soluble, as well as highly diagenized and semi-consolidated sediments occur in the area. This geology induced diverse geomorphic processes (valley incision, derasion, mass movements, karst formation, periglacial), which created a variety of landforms (small caves, non-karstic pseudo-caverns, scarps, bedrock outcrops, rock crags). The folded-imbricate structure, with the south-easterly dipping reverse faults of the Paleo-Mesozoic massif (Pelikán 2005), resulted in scarps of different resistance to weathering and formation of fascinating cliffs. Owing to the imbricate structure, even the buried Paleozoic rocks of the neighbouring Bükk Mountains to the south (composed of Mesozoic limestone-nappes with southern vergence) can be studied here exposed on the surface. Along the abundant fault lines tectonically preformed, asymmetrically incised valleys and gorges were formed exposing the second oldest (Late Ordovician-Permian) rock assemblage of the uplifted continental floor in Hungary and thus creating an opportunity for paleoenvironmental reconstruction. The preserved small Neogene cover and the traceable erosion surfaces indicate well-dated transgression and exhumation periods. Thus, the post-sedimentary (orogenic) geomorphic history of the mountains can also be well detected. The origin of landforms, the effects of selective denudation on hard and loose rocks, slope evolution and downwearing processes can be studied within a few kilometres' distance (Fig. 21.1).

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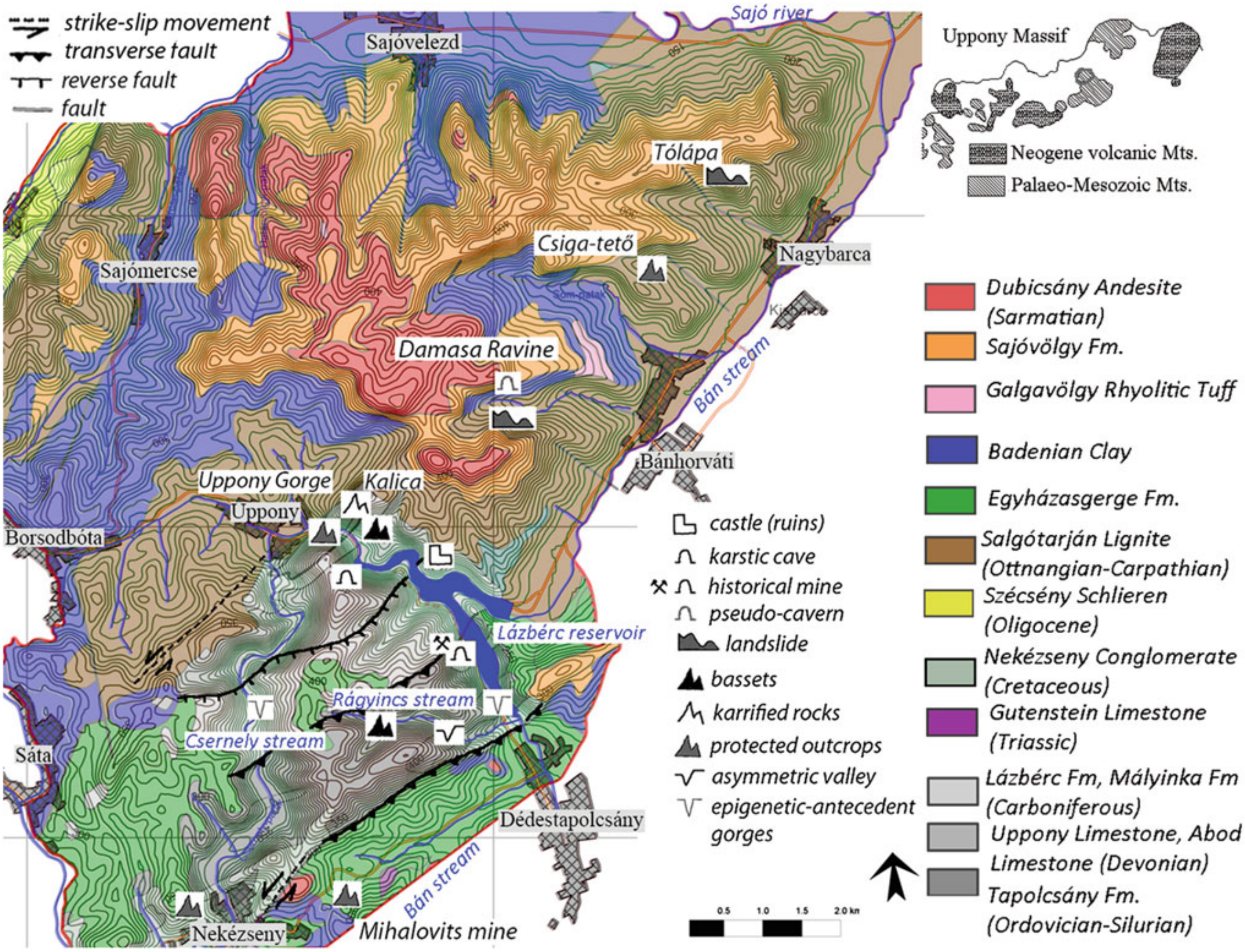


**Fig. 21.1** Geological and contour map of the area indicating sites of geomorphological interest (geology after Budinszkyné et al. 1999)

## 21.2 Geological and Geographical Settings

The Uppony Mountains are bordered by the Tardona Hills (E), the Bükk Mountains (S), the Ózd-Egercsehi Basin (W), the Vajdavár Hills (NW) and the tectonically preformed valley of the Sajó river (N). Located along the Darnó Fault Zone, it is one of the structurally most complicated regions of Hungary. It consists of two main parts. The 15 km$^2$ core area (massif) is a horst dominantly characterized by Paleo- and partly by Mesozoic rocks, with remnants of Tertiary marine molasse sediment caps on the surface proving former transgression, subsequent exhumation, and a 400-m uplift since the Late Miocene (Pannonian). In the north the Paleozoic rocks were thrust over the Oligocene and Lower Miocene foreland (Fig. 21.2), while in the south there is a complicated relationship with the rock formations of the Bükk Mountains along the Nekézseny reverse fault (Kovács 1994). To the west and the east the Paleo-Mesozoic core is buried under Tertiary littoral-paralic sedimentary series (Fig. 21.2). Similar rocks also occur beyond the Sajó River (Szendrő-Rudabánya Mountains and Putnok Hills), but with different scenery. As attested by the rocks of various age and origin eroded to similar elevations, both the massif and its surroundings had once constituted the northern Bükk piedmont, which was dissected by valley regression. Today only ridges of 350 m and 420–450 m elevation indicate the existence of former planation surfaces. In addition to periodically renewed uplift, selective denudation also caused differences in the height of ridges composed of various rocks (Demeter and Szabó 2008a).

Compared to the adjacent areas, the unit represents an upthrown block exposing the rocks of the basement and separates the Eastern and the Western Borsod Coal Basin from each other. The structure is dominated by folded ‘slices’ of NNW vergence dismembered along reverse faults (Kovács 1994). In addition to its different geology, the Paleo-Mesozoic Uppony Mountains with steeper slopes and deeper (but not graben-type) valleys, also contrast geomorphologically with its younger surroundings (Szalai 2004).

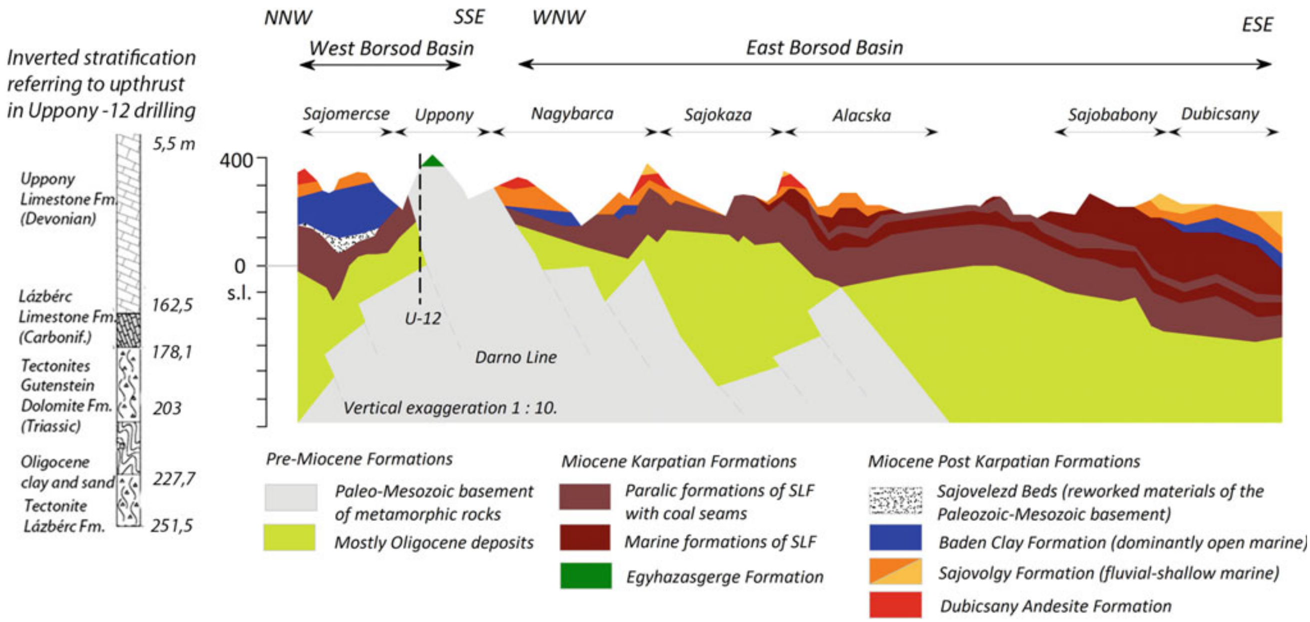


**Fig. 21.2** Deep borehole Uppony-12 indicating the upthrust of basement over Oligocene rocks and a geological cross-section crossing the Uppony Massif (modified after Püspöki et al. 2009)

The mountains belong to the catchments of the Csernely and Bán streams, the tectonically preformed, epigenetic-antecedent valleys of which expose both old resistant rocks and young unconsolidated deposits (Fig. 21.1).

There is a cold and moderately arid continental climate in Uppony, with slightly less hours of sunshine (1,850 h) and lower annual average temperature (8.5 °C) than common in Hungary. The average annual precipitation does not exceed 650 mm, and the dominant vegetation is oak and birch forest. The gorges, however, present particular microclimatic conditions.

The Lázbérc Landscape Protection Area, although established with the primary purpose to protect the drinking water basis of the Lázbérc reservoir (an artificial lake formed by impounding the valley of Bán stream) in 1975 (Fig. 21.3), bears a range of interesting geological, geomorphological and cultural sites, including the Damasa and Uppony Gorges and the Rágyincs valley, which are presented here in detail.

## 21.3 Geological and Geomorphic Evolution

The rocks of the Uppony Mountains are of regional significance for the evolution of the Carpathian Basin. Surrounded by Rudabánya-type Triassic and Lower Miocene sediments along the Uppony reverse fault to the north and by the Upper Permian–Lower Triassic rocks of the Bükk Mountains in the south along the Nekézseny reverse fault (Schréter 1945), the Uppony block is considered to be part of the Darnó Fault Zone (Péró 1997). Its imbricate structure with NNW vergence is marked by steep reverse faults of the Paleozoic rocks thrust over younger formations (Fig. 21.2)—also confirmed by the elongated NE-SW tracts. However, due to strong tectonics that rearranged the position of rocks the identification of the original settings is difficult. The imbricate structure (Fig. 21.4) had formed prior to the Upper Cretaceous (Alpine Orogeny), as the Nekézseny Conglomerate from this period in the SE part of the mountain settles discordantly on the underlying strata. Although the Variscan (Hercynian) Orogeny played a crucial role in the deposition of sedimentary units, the metamorphism of the Paleozoic sediments (foliation) took place during Alpine movements (Kovács 1994; Árkai et al. 1982) (Fig. 21.4).

During the first sedimentation cycle (from the Late Ordovician to the Variscan Orogeny) sands were deposited followed by radiolarites, indicating the deepening of the sedimentary basin below the carbonate compensation level (known worldwide as Silurian euxinic facies). Subsequent tholeiitic basalts represent open marine conditions with a rift valley. The Devonian limestones were deposited on carbonate platforms, while the subsequent flysch-like sediments of the Carboniferous Lázbérc Formation are composed of the reworked terrestrial material of the mountains in the north, uplifted during the Variscan Orogeny. The following shallow marine molasse unit (lydites and calcareous sandstones of Mályinka Formation) represents the closing of the Paleo-Tethys and the end of the first sedimentation cycle (Budai and Konrád 2011). The black Permian Nagyvisnyó

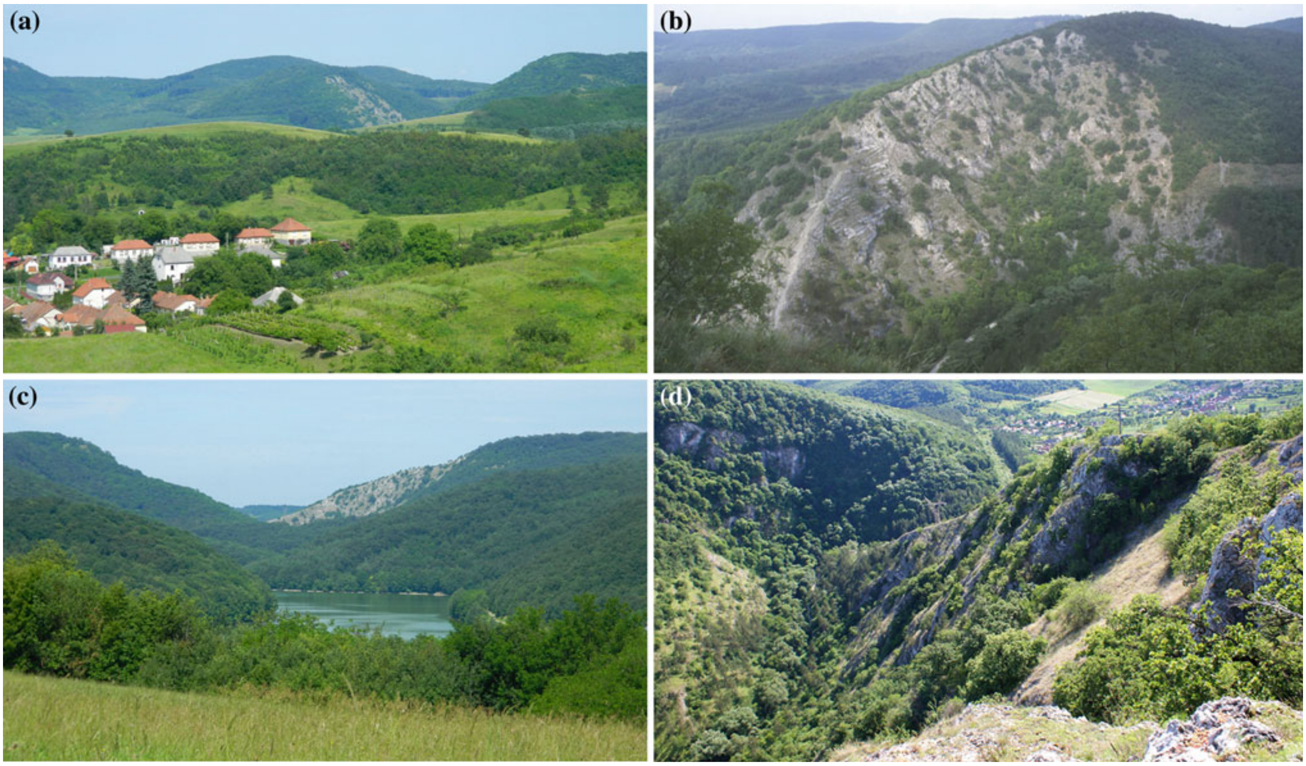


**Fig. 21.3** The Uppony Gorge (photos by Katalin Szalai). The Uppony massif and its semi-consolidated foreland (**a**). The karstic horst of Mt. Kalica composed of Palaeozoic limestones guards the entrance of the Uppony Gorge (**b**). The artificial Lázbérc reservoir is located in a tectonically preformed valley (**c**). Karren on limestone bassets and caves on the opposite side of the Uppony Gorge with Uppony village in the background (**d**)

Limestone on the fringe of the mountains deposited in poorly aerated lagoons, rich in organic material, also indicates regression (as seen in the Mihalovits quarry).

According to the paleogeographical reconstruction of the Carboniferous (Hercynian Orogeny), the Uppony Mountains, located on the margin of the ALCAPA plate of African origin (Pelso Unit), was linked to the Carnic Alps and the Graz Paleozoic. While the Gemer Carboniferous in the north represents a terrestrial lowland accumulation series with the crystalline rocks of the Western Carpathians in the background, the Lower Paleozoic sequence of the Uppony block is built up of coastal and continental shelf sediments (Péró 1997).

The shallow marine carbonate sediments of the opening Tethys—introducing the second sedimentation cycle—begin with remnants of the Triassic Gutenstein Limestone. In the Cretaceous, during the Carpathian Orogeny, an imbrication process took place resulting in tracts elongated in northeast-southwest direction. The elevated thrust sheets were then subject to erosion and the reworked and transported material was cemented into coastal conglomerate (Nekézseny Conglomerate Formation), a correlative sediment (exposed in outcrops as well). In the Oligocene this regression was followed by a transgression from the northeast induced by the subsidence of the floor in the northern foreland of the Carpathians, but the deposited material (up to 800 m thickness) was subsequently eroded from the Uppony Mountains.

During the Neogene three major transgressions affected the area. The first arrived from the southeast during the Ottnangian-Karpatian (17 Ma ago). Under Mediterranean conditions cross-bedded sand dunes, lidos and swampy lagoons developed with decreased salinity and abundant littoral (*Taxodium*) forests. These were buried by the transgressions generating 200–400 m thick coal-bearing seams under anoxic circumstances (Borsod Coal Basin), which are missing from the uplifted massif itself. Another transgression reached the peripheral zone of the mountains 15 million years ago and deposited the Badenian Clay Formation of clay marl indicating deeper shelf and open marine conditions (Püspöki 1998), exposed in the Damasa Gorge. Subsequent intensive uplift removed most of the Badenian sediments from the Tardona Hills and the Uppony massif. The third and smallest transgression followed in the Sarmatian (13 Ma ago). By then many klippen did exist and were exposed to marine abrasion, as affirmed by erosion discordance in Miocene sediments lying directly on Permian rocks in the Mihalovits quarry at Nagyvisnyó. Under Mediterranean

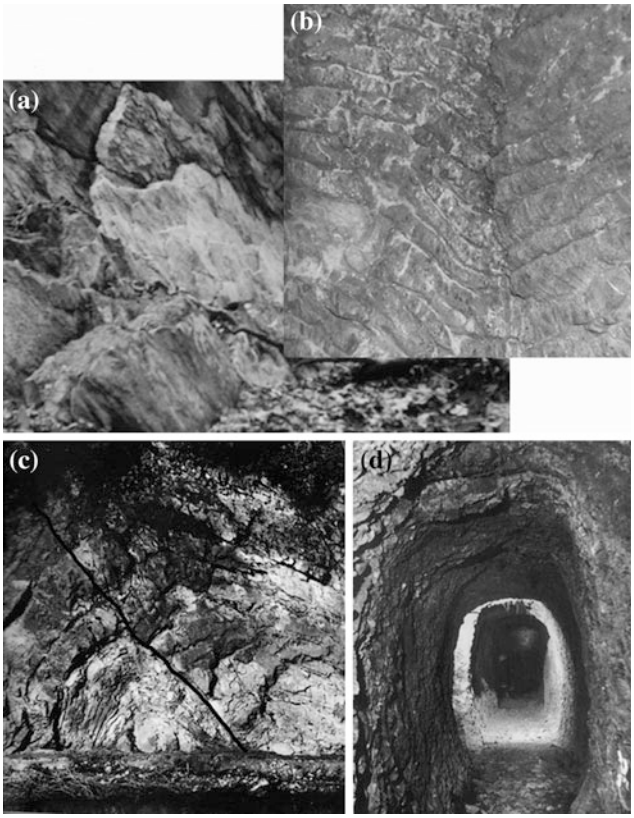


**Fig. 21.4** Geological features observed in outcrops (pictures of the bottom row from Fülöp 1994). Imbrication in an outcrop near Uppony (**a**). Tectonic movements resulted in the foliation of rocks (**b**). Folded Palaeozoic limestone and shale joining along a fault plane (**c**) Tunnel used to explore Fe and Mn ore (**d**)

conditions rivers removed gravels of metamorphosed Paleozoic rocks and Gemer-type Mesozoic limestones (Sajóvelezd Beds) from uplifting areas and redeposited them in the region of the Uppony Mountains (Fig. 21.2). Limestone gravels also prove the temporary exhumation of the Bükk Mountains and the post-Sarmatian formation of the tectonic valley of Bán which separated the Bükk, the Tardona Hills and the Uppony Mountains from each other (Hevesi 1986, 2006). With the Sarmatian tectonic movements andesitic volcanic activity also intensified (Dubicsány Andesite Formation).

During the so-called "Miocene geomorphic inversion", former sedimentary basins became uplifted, while ancient mountains subsided and seas retreated. The northern Bükk foreland with the Uppony Massif was elevated to 350–450 m. The former northeastern and latter northwestern rotational upthrust of the Bükk determined the direction of present-day epigenetic-antecedent incised valleys (Csernely and Bán streams), characterised by captures, exposing the underlying older rocks. In the Pannonian-Pliocene patches of red clays, stemming from weathered and redeposited Miocene tuffs, appeared as correlative sediments of periodical uplift, exhumation and pedimentation under subtropical conditions (7.5–3.7 Ma). In the Pleistocene both sheet wash and stream erosion were active under alternating warm-wet, warm-dry, cold-wet and cold-dry spells (Székely 1973). Stream erosion was predominant in warm-wet phases causing an increase in slope angles, while sheet wash became common in warm-dry periods. In periglacial cold-wet and cold-dry phases however, gelisolifluction processes and gravitational talus creep moulded the surface (Marton-Erdős 2001), reducing slope angles.

## 21.4 Landforms

### 21.4.1 Relief, Valley Density and Slope Inclination

In order to understand the sources of landform diversity, a structural geological and engineering geomorphological approach was combined with a comparative quantitative morphometric analysis of neighbouring landscapes. Based on the statements of Beavis (2000), who claims that geological structure determines the place, while lithology influences the extent of erosion, and Clayton and Shamoon (1998), who found that harder rocks statistically tend to support steeper slopes or higher landforms, we defined surface rock resistance to erosion (in MPa) and the relationship between rock strength (as a lithological feature) and the selected morphometric features of landforms (Püspöki et al. 2005). In general terms the horizontal and vertical dissection of a landscape may determine the character of positive landforms (Ollier 1967), and the calculated average values may be typical for the landscape. These two parameters, however, are not able to describe slope types (convex or concave), hence they cannot inform about the distribution of slope angles. Thus, apart from relative relief and valley density, slope angle frequency was also used as a distinctive morphometric feature (Demeter and Szabó 2008a) to estimate the extent of post-Miocene erosion in the Uppony Mountains and to compare the general character of landforms in different landscapes.

Although the correlations between valley density and relative relief ($r = 0.83$), relative relief and rock hardness ($r = 0.65$), sloping and rock hardness ($r = 0.73$), sloping and relative relief ($r = 0.83$) are not insignificant in the area, the relationship between rock hardness and valley density in the Uppony Mountains proved to be weak ($r = 0.4$) (Demeter and Szabó 2008b). Since higher values of valley density and increased relief indicate uplifting areas, the dissected surface of the Uppony massif can be regarded as one of the most dynamic regions in North-Hungary (Fig. 21.5a). Nevertheless, in certain cases low relief may occur with higher valley density and high relief with lower valley density, e.g. in the Aggtelek Karst.

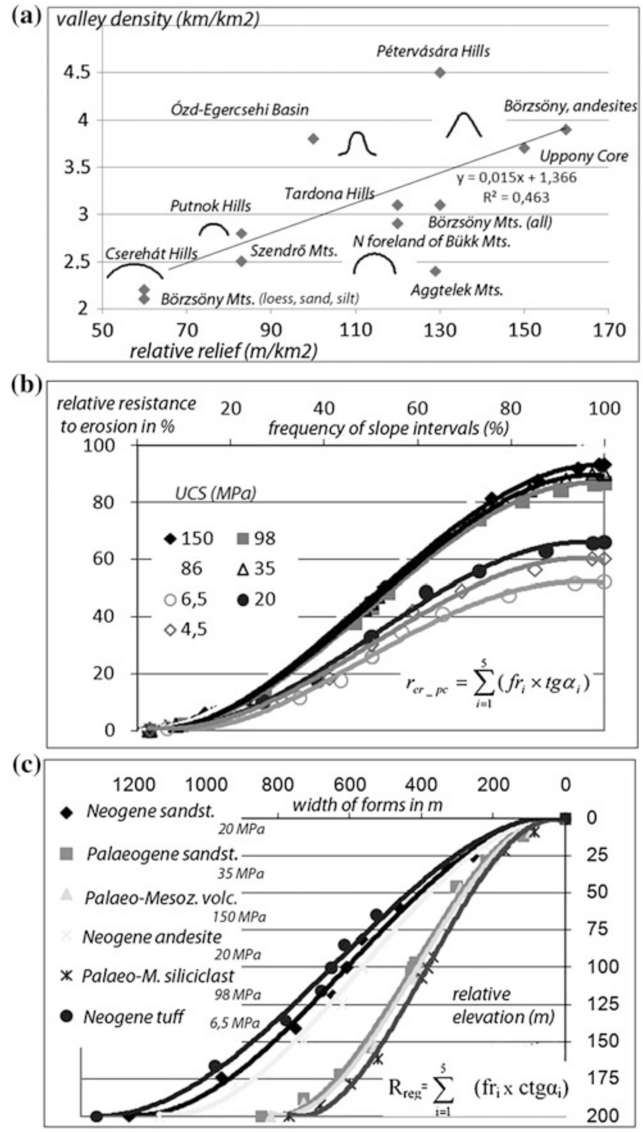


**Fig. 21.5** Comparative geomorphology: regional differences in the extent of horizontal and vertical dissection (A) and the role of selective denudation on different rocks (B – height and C – width of landforms)

Comparative studies show that the Szendrő Mountains, composed of the same rock types as the Uppony block, have lower valley density and relative relief (divergent evolution). Moreover, these values resemble those of the neighbouring Putnok Hills (composed of semi-consolidated Tertiary rocks), reflecting similar geomorphic settings (dissected planation surfaces in mountain forelands). The unconsolidated Pannonian sediments of the Cserehát (previously being an accumulation glacis of the Slovak Ore Mountains) form low ridges with wide valleys (mature phase). West of the Uppony Mountains, however, in the geologically homogeneous, but dissected Vajdavár Hills (Chap. 19), built up of consolidated Paleogene sandstones, higher elevations with steep slopes, similar to the features of the Uppony core, are predominant. Here convergent evolution (Lóczy 2008) based on rock strength can be suspected.

In the case of the Uppony massif a strong correlation was found between valley density and rock strength (r = 0.7) and also between rock strength and relative relief. But this relationship is not unambiguous. In the Börzsöny Mountains, for instance, there was no measurable correlation between relative relief and valley density either on schlieren or on andesites and tuffs, only on unconsolidated Pleistocene loess (Gábris 1987). Accordingly, both surface rock type and —beyond erosional valleys—the frequency of derasional valleys can influence this relationship. In addition, the increasing resistance of rocks led to the scarcity of derasional valleys ($r^2 = -0.93$) in Uppony.

High valley density in the Uppony massif reflects tectonic maturity. The horizontal dissection of the Paleo-Mesozoic-Neogene sedimentary area is similar to that of the North Hungarian volcanic ridge, namely the Visegrád (2.9 km km$^{-2}$), Börzsöny (2.8 km km$^{-2}$) and the Eastern Cserhát Mountains (3.1 km km$^{-2}$) (Ádám 1984). Nevertheless, if gullies (0-order valleys) are also included, even higher values are obtained for highly diagenized and semi-consolidated rocks (Paleo-Mesozoic siliciclasts: 4.3 km km$^{-2}$, Paleogene sands and silts: 4.8 km km$^{-2}$). The lower density (3 km km$^{-2}$) for Paleozoic limestones reflects continuous valley incision, while gullying remains to be of secondary importance. Terrains covered by old rocks are dominated by erosional valleys (Paleo-Mesozoic limestones: 66 %, Paleo-Mesozoic siliciclasts: 57 %), while gullies are predominant on Paleogene sandstones and Neogene marine sands and silts (46 and 37 %). Present-day gullying is considerable: a comparison of land-use maps in three settlements between the 1860s and 1960s indicates intensive valley regression (exceeding 30 m in 15 cases out of 51) and deposition in some instances (3/51) indicating derasional processes, the most common on loose Paleogene and Neogene deposits (constituting 37 % and 26 % of the valleys, resp.) (Demeter and Szalai 2004).

Several authors pointed out the tectonic character of valleys in the Uppony Mountains. According to Schréter (1945), the Lázbérc reservoir is located along a transverse (NNW-SSE) fault. Pantó (1954) recognised intensive tectonism and planes of weakness in rocks. Peja (1956a, b) regarded the Bán valley to be tectonically preformed. Dér (1957) proved that valleys are formed perpendicular to the strike on harder rocks (being the shortest way of runoff) and parallel to it on rocks of medium resistance. Láng (1953) identified asymmetrical valleys with usually shorter and steeper northwestern and longer and gentler southeastern slopes. Szalai (2004) confirmed that this phenomenon is regular in tectonically preformed valleys along the main faults where strata outcrop, but atypical in the case of transverse faults (see the Rágyincs Valley as an example) and that the directions of faults, fractures and valleys are in harmony (Fig. 21.6.).

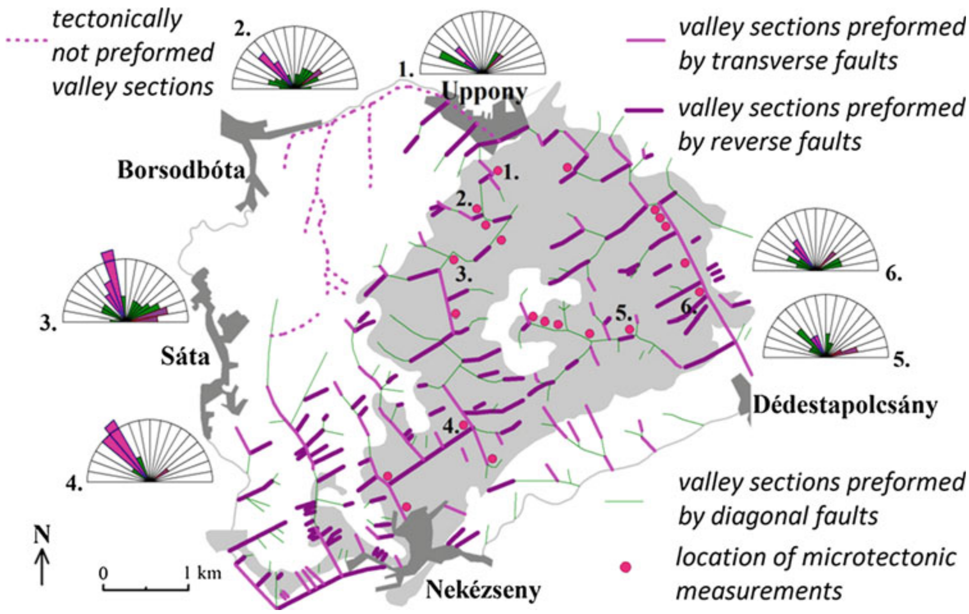


**Fig. 21.6** Relationship between valley directions and distribution of faults

According to Szalai (2004), almost all valleys in the Uppony block are tectonically preformed, with 35 % following major faults (60°) and steep thrust sheets with NNW vergence. Here in the massif the consequently parallel, northeast to southwest strike marks the direction of the reverse faults perpendicular to the direction of compression. To transversal faults (325°) 22 % of the valleys are connected. Due to spatial shortening (in addition to foliation) a diagonal fault system also appeared (43 %) in the region. On the Tertiary molasse sediments, however, valley directions vary more considerably because of the dispersed fault lines and widespread derasional processes.

The harmful effects of the erosion still apparent on the terrain were mentioned as early as in 1777 in a questionnaire (Csorba and Tóth 1991). Variations in rock type and erosional resistance influence cumulative slope steepness, which is high in the Uppony massif: above 1,500 points (out of 2,500), whereas it is around 700 in the Cserehát Hills, 1,100–1,200 in the Szendrő Mountains and the Putnok Hills, 1,200 points in the Aggtelek Mountains and the Pétervására Hills (Félegyházi et al. 1999). On Neogene andesites (20 MPa), Neogene silts and Paleogene sands and clays (7 MPa) most common slope steepness is around 10–14 %, while on Paleozoic rocks (80–150 MPa) values around 24–26 % are typical. To quantify the resistance of different rocks to selective denudation, downwearing or valley regression, hypothetical slope profiles are presented here using the frequency (%) of slope intervals and slope angle. The width and relative elevation of identical features developed on different rocks was also estimated (Fig. 21.5b, c).

### 21.4.2 The Uppony Gorge

The Csernely stream has been undoubtedly an important factor in geomorphic evolution of the region. After cutting across the whole mountains, it takes a sudden turn and carves the 500 m long, breathtaking Uppony Gorge (Figs. 21.3 and 21.7) and finally joins the Bán stream. It is one of the most spectacular and deepest (170 m) gorges in Hungary and stretches from Uppony to Dédestapolcsány, exposing the geological structure of the old block. The Paleozoic limestones, schists and clay shales are exposed on the shores of the Lázbérc reservoir. On a nearly 4-km long nature trail one can get a unique experience by learning about the paleoenvironment of 140 Ma (from the Silurian to the Carboniferous). The now abandoned manganese and iron ore mines were worked in the 18th century for the smelters established by Heinrich Fassola. Cut into the Paleozoic shales, they—beyond their cultural values—also reveal geological structure, tectonics and stratification (Fig. 21.4).

In the moderately soluble, steep limestone cliffs of the Uppony Gorge 25 small caves, rock shelters and hollows have been registered. From one of these cavities, only some metres in length, Uppony niche no 1, an abundant set of Pleistocene paleonthological finds (e.g. bats and rodents) was discovered (Pazonyi and Kordos 2004). The gorge is also extremely rich in rare plants and such a great biodiversity in a small area can hardly be observed elsewhere in Hungary. On the hot rock swards of the Uppony Gorge of southern exposure, species of Mediterranean flora live whereas in the northern-exposed cold grasslands, on debris

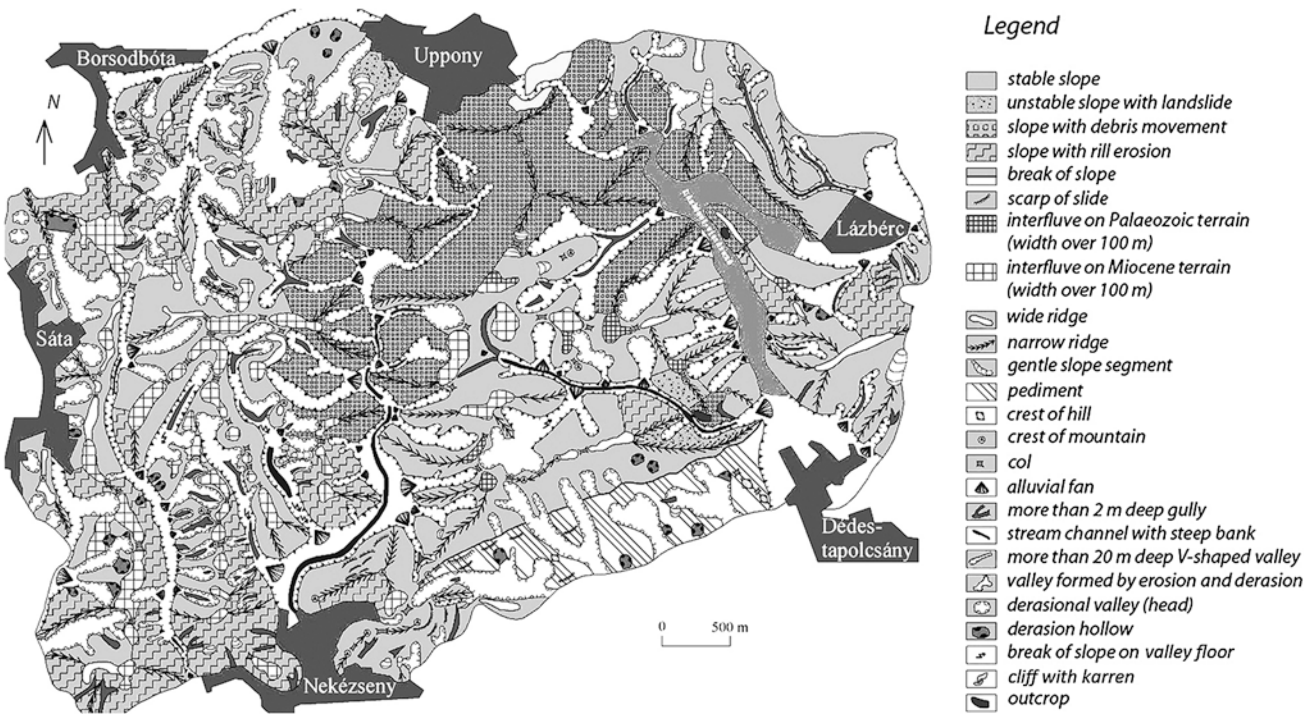


**Fig. 21.7** Geomorphological map of the Uppony Massif and its surroundings (after Szalai 2004)

slopes and in shrub forests an assemblage characteristic of cool mountainous areas has developed. Among the cultural monuments, on top of the picturesque cliff marking the entrance to the gorge (Kalica) a cross symbolizes the boundary between Protestantism and Catholicism, while the forgotten ruins of the Dedevár castle remind us of the Middle Ages.

### 21.4.3 The Rágyincs Valley

Located in the southeastern part of the Uppony block (near Dédestapolcsány) the Rágyincs Valley is an excellent example of accordant drainage evolution. Along an almost 3-km-long easy walk along the path in the valley, one can observe the most significant geological and geomorphological forms and processes of the Uppony Mountains. On the conveniently accessible catchment area of the Rágyincs Stream–starting from the mouth at the southern bank of the Lázbérc reservoir and moving towards the source—one can discover exposures demonstrating microtectonic fracture system and lithological characteristics of the Paleozoic Tapolcsány Formation, Lázbérc Formation and Rágyincsvölgy Sandstone (Fig. 21.8), together with different stages and morphofacies of valley development, not to mention a rock arch. The short stream changes its character more than once along the course, forming a valley rich in geomorphological features (alluvial fan, meandering, terraces, rock channel, gorge, rock cliff, debris cone etc.) (Fig. 21.8).

With the help of valley cross-sections in different rocks or structural units typical landform groups were classified into morphofacies and extended to other parts of the Uppony Massif (Szalai 2004). Valley sections along imbrications, in the foreground of reverse faults, are common in the Uppony massif and have asymmetrical valley profiles. Characteristically, the channel is asymmetrical and the stream undercuts the front of imbricate structures, uncovering the bassets, while alluvium is deposited on the dip slopes, bedding planes (Fig. 21.8). Relatively stable, gently sloping surfaces (sometimes with soil creep) occur here. Mainly asymmetrical terraces show steps. The fronts, in contrast, are being continuously denuded into steep valley walls and convex slopes shaped by mass movements often form near the valley floor (Fig. 21.8).

Deeply incised along transverse faults, streams could have been redirected from one structural strip to another. Rocks of considerable resistance withstand erosion for long periods causing alluvial deposition in the headwaters. However, after the waterfalls are cut through, fast downcutting and the sudden removal of the deposited alluvium occurs, accompanied by the formation of terraces, almost symmetrical valley profiles, weathered-out ridges and bedrock channels in the upper valley sections.

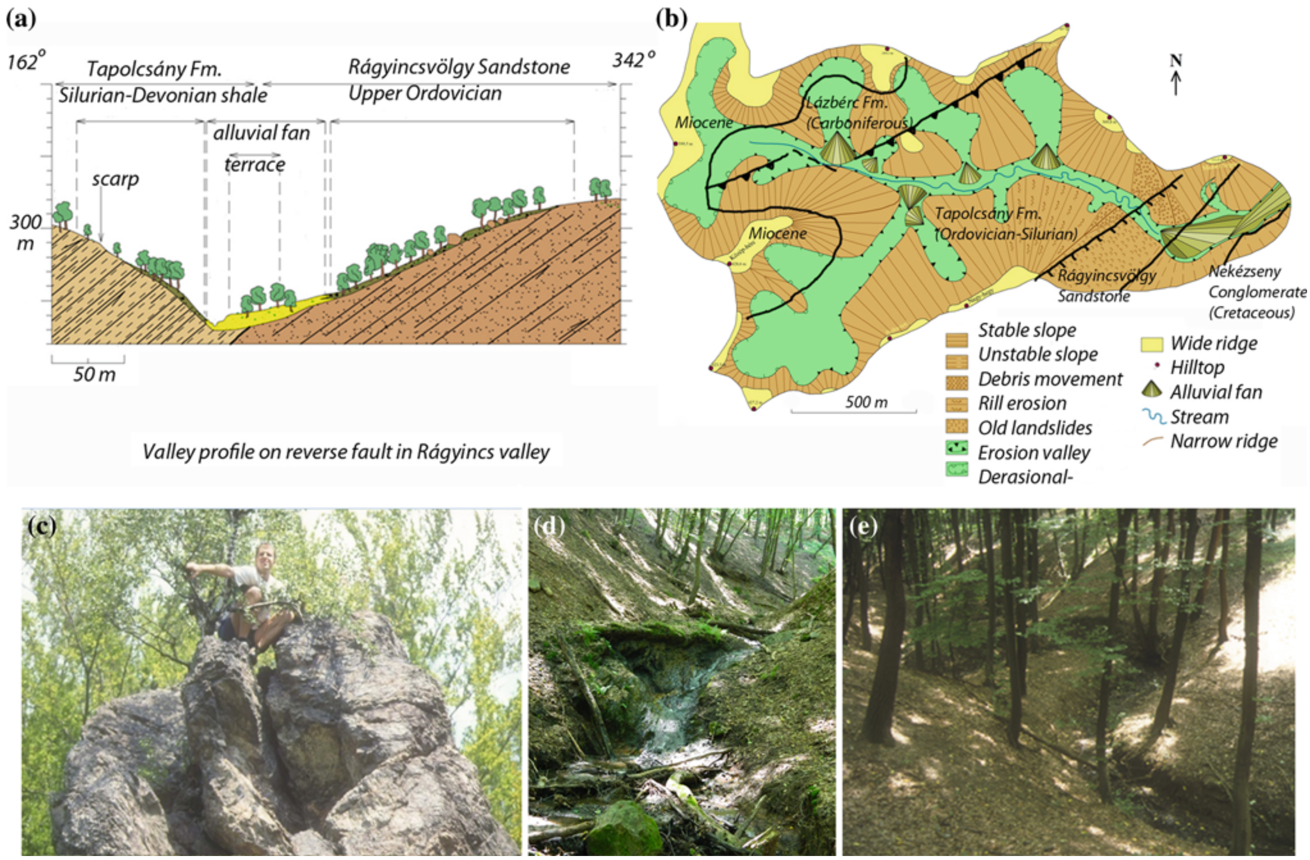


**Fig. 21.8** Geomorphological features of the Rágyincs Valley. Valley profile on reverse fault in Rágyincs valley (**a**). Geomorphology of the Rágyincs valley (**b**). Where the stream cuts into the hard Rágyincsvölgy Sandstone, a gorge-like valley section with spectacular cliffs and a steep bedrock channel were formed (**c** and **d**). On less resistant rocks however (e.g. alluvium) the stream is meandering (**e**)

Valley sections following the diagonal fracture system are transitional features between valleys preformed on the front of reverse faults and valleys running along transverse faults. This refers to the fact that stream erosion could easily cut diagonal fractures parallel with the main tectonic lines. Consequently, they could take part in preforming the imbricate structure or in maturing cut-throughs. The standard variation of symmetry index values is high.

### 21.4.4 The Damasa Gorge and the Nagybarca Landslides

For the density of mass movements, the Uppony Mountains do not exceed the Hungarian average with their 5–10 movements per km$^2$. In the surroundings only the Ózd-Egercsehi Basin showed higher values due to subsurface mining. Around 50–75 % of these movements were induced by human activity and similarly to the regional average, 25–50 % are considered to be active (Szabó 1996). The area of 50–75 % of all mass movement sites exceeds 250 m$^2$. Being tectonically preformed, composed of rocks of different consistency and surrounded by uplifting horsts, the valley of the Bán stream—leaving the Paleozoic block behind—is ideal for observing characteristic landslides. The Damasa Gorge near Bánhorváti was formed by blocky landslides (quite rare in the country) induced by successive earthquakes in the 18th to19th century (1763, 1829, 1834, 1848). Two major blocks of the Sarmatian volcanic andesite agglomerate nappe slid on the underlying unconsolidated Badenian clayey marine sediments (often impregnated by water) that functioned as a slip plane, and created a 170-m deep and long gorge, where the average air temperature does not exceed 0 °C even in summer (Hevesi 2006). The second biggest continuous network of non-karstic pseudocaverns of Hungary are found under these boulders. Additionally, rock shelters, outcrops and cliffs appear on andesitic rocks formed by exogenic processes (Fig. 21.9).

Near Nagybarca, close to the remarkable outcrop of Csiga-tető (where the Miocene cross-bedded coastal dunes and *Ostrea* fauna are exposed), an older and mature landslide can also be found. The so-called “Tó-lápa” was formed on unconsolidated sediments (Fig. 21.9). Intense valley incision led to more than 200 m km$^{-2}$ relative relief values in the vicinity. This landslide type is quite common in the

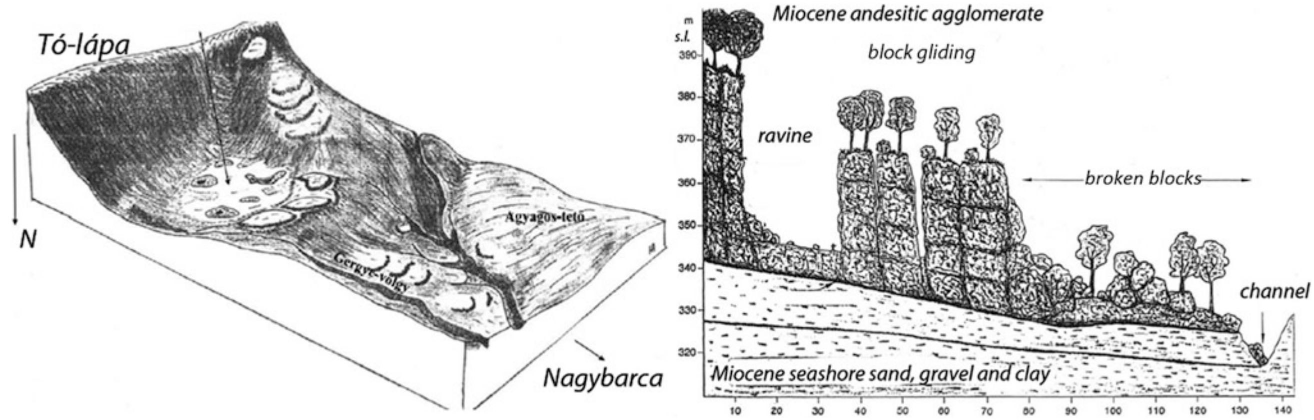


**Fig. 21.9** Sketch of the Tó-lápa landslide at Nagybarca and the Damasa Gorge near Bánhorváti (Hevesi 2006)

country, but rare in the Uppony region. Its considerable relative age is indicated by the fact that 3 out of 7 small sag ponds have disappeared by now confirmed by the vegetation cover and a comparison of old maps (Hevesi 2006).

## 21.5 Conclusions

The presented unique (both natural and semi-anthropogenic) features of geomorphological interest in the Damasa Gorge (landslides, pseudo-caverns), Rágyincs Valley (terraces, scarps, outcrops) and Uppony Gorge (karren, outcrops, cliffs, caves, mines, castle ruin) confirm the results of the general numeric and comparative analysis. High petrophysical diversity and an intricate geological structure contributed to the geomorphic diversity of the Uppony Mountains and reveal 300 Ma of cyclic sedimentary and orogenic history in situ. The numeric analysis proved that not only the harder rocks, but steeper slopes, greater valley density and higher relief make this small area—surrounded by gentle hills of similar elevation—resembling low mountains. The relationships between valley directions, slope morphology (valley asymmetry) and the occurrence of fault types were also verified. Prominent features built up of different rocks tend to have different shapes. Examples of convergent and divergent geomorphic evolution were also detected through a regional comparison of physical features, and the rate of recent valley regression was also measured.

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# 22 Bükk Plateau: A Karst Landscape

László Sütő, Richard W. McIntosh, Attila Hevesi, and Márton Veress

**Abstract**

The Bükk Mountains are among those with the most intricate geological structure and marked topographical contrasts in Hungary. They comprise the Big and Little Plateaus, the Bükkalja foothills and the northern foreland (Bükkhát). The unique rock sequence with highly variable hydrogeological properties suffered moderate regional metamorphosis and heavy folding over an extensive area during the Variscan and Alpine Orogenies. The accumulation of carbonates, which are affected by intensive karstification, began in the Carboniferous and lasted until the Middle Jurassic. The Big Plateau has an undulating surface of low rounded mounds and enclosed shallow valleys with landforms like doline fields in poljes, doline rows with ponors aligned along valleys and major individual ponors. Underground avens, chimneys and caves are found in large numbers, constituting a quarter of all caves in Hungary. The northern edge of the plateau is dissected by valleys with deep gorges. Along the marked southern margin, bastion-like limestone cliffs, called "The Rocks", form a prominent edge rising from the foreland of limestone and volcanic zones. Karst landforms (karren fields, opened caves) are also typical of "The Rocks".

**Keywords**

Karst plateau • Mesozoic limestones • Thrust structures • Poljes • Dolines • Ponors • Caves • North Hungarian Mountains

## 22.1 Introduction

Although karst rocks compose only a small portion of the territory of Hungary, the landforms developed on them are most diverse. The central Bükk Mountains of highest average elevation, the karstic Bükk Plateau of 115 $km^2$ area (Fig. 22.1), called the "Giants' Table" in tourism literature, is one of the least accessible inner areas of the Bükk National Park. The Big and Little Plateaus are divided by the valley of the Garadna Stream, regressing from the non-karstic areas in the east. Visitors marvel at its specific landforms described in detail by scientists (Tóth 1984; Hevesi 1986, 1990, 2002a, b, c; Bárány-Kevei 1997; Telbisz 2001, 2004; Martonné 2001; Dobos 1997, 2001 etc.). Various karst forms, rock formations produced by frost weathering, karstic and erosion valley types are geomorphological values worth protection (Dobos et al. 1999).

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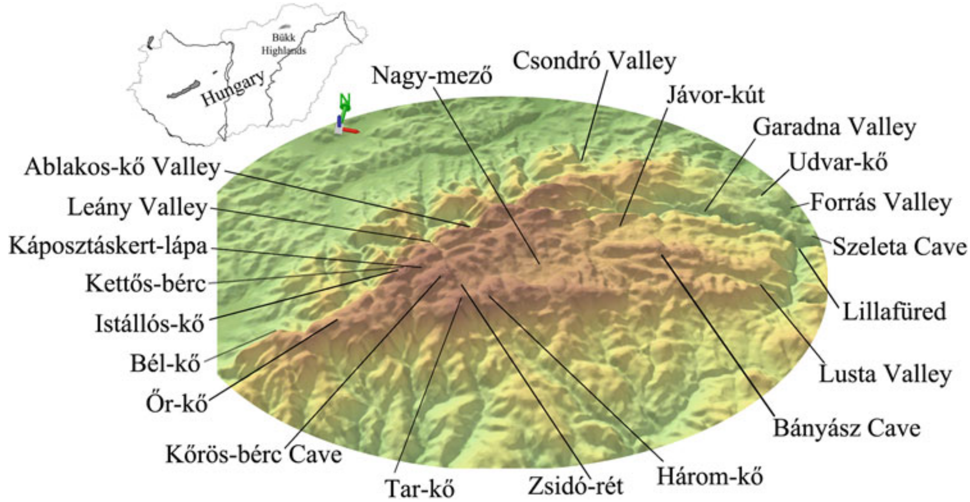


**Fig. 22.1** Digital elevation model of the Bükk Mountains (by Balázs Kohán and László Sütő)

Since the mountain is rich in karst features, most people believe it is primarily built of carbonates. Although limestone is dominant in the composition of the Bükk Plateau, even this area features old igneous formations (metamorphic andesite and basalt) and siliciclastic sedimentary rocks. Furthermore, the limestone formations are not uniform regarding petrography, structure and age. Petrographic variability together with complex structure provided the basis for the development of a romantic landscape with steep slopes and sharp turns of valleys and ridges.

## 22.2 Geological Setting

Between the 1960s and 2000 the mapping projects launched by the Geological Institute of Hungary brought much progress in the exploration of the mountains: proved the existence of Jurassic sediments, determined the degree of rock metamorphosis and resulted in new interpretations of the stratigraphy (Balogh 1964; Csontos 1988, 2000) and geological evolution (Pelikán 2005). Nevertheless, a number of episodes in the history have not yet been properly described and explained.

The oldest geological formations are located in the northeast part, in the North Bükk anticline (Fülöp 1994; Less 2002; Pelikán 2005). Only the Upper Palaeozoic is represented here by four formations. In the Carboniferous pelagic sediments were overlain by siliciclastic turbidites. With the infilling of the basin, a mixture of dark grey, black shale, slate, silt and sandstone with black limestone lenses formed in a shallow marine environment followed. The black slate in the northwestern part was slightly metamorphosed and, due to its excellent workability, was widely used on roofs. The Permian coastal sandstone and nearshore evaporites together with the later limestone formed in an extensive shallow anoxic marine environment during the Alpine orogenic cycle. The transition from Upper Permian to Lower Triassic sediments is continuous in two excellent outcrops of international renown (Bálvány and Gerenna-vár), where scientists researched for explanation of the great extinction at the end of the Palaeozoic.

Triassic shallow marine limestones and siliciclastic formations dominate in the eastern, central and western Bükk and in a smaller area on the southwestern edge. As transgression advanced, white and light grey carbonate platform sediments deposited in marine environments rich in oxygen, most extensively on the present-day Bükk Plateau. These "pure" limestone types hold diverse karst features and also give the quality raw material for cement production. One of the most spectacular outcrops of the Bükkfennsík Limestone Formation is the Bél-kő, significantly transformed by quarrying (Fig. 22.2). The almost vertical bedding and cleavage planes indicating strong structural deformation were exposed and cleared by the rock boulders rumbling down the steep slopes in order to disintegrate and enable miners to further process them. A recently constructed nature trail (Baráz et al. 2003) gives information not only on the geology of the limestone and the nearby Jurassic slate but also describes the history of quarrying and the application of the limestone both as raw material for the cement industry and as building stone in many of the local buildings like the Cistercian abbey, founded in the 13th century. Towards the end of the Triassic carbonates were mixed with silica and the resultant cherty limestone dominates in the southeast Bükk (Haas 2012).

The development of a continuous carbonate platform over the Triassic was interrupted by three episodes of volcanism (Szoldán 1990; Harangi et al. 1996; Pelikán 2005). Although the early volcanic activity mostly produced neutral volcanics considering their geochemical character, they are more similar to acid rocks in their petrographic appearance (Bodnár et al. 2014). In addition to effusive rocks, the formation is also rich in pyroclasts. The second phase of

**Fig. 22.2** Bare quarry walls of the Bél-kő with the Bélháromkút Cistercian abbey in the foreground (Photo by László Sütő)

volcanic activity produced mostly rhyolites scattered randomly in the Bükk today, while the third volcanic period resulted in basalts found in small occurrences across the mountains (Dobosi 1986; Szoldán 1990). The Late Triassic to Early Jurassic deep sea was part of a small rift trough (Pelikán 2005), associated with the break-up of Pangea (indicated by the ophiolite series with pillow lavas of the Szarvaskő Basalt, unique in Hungary). Mesozoic formations younger than the Jurassic pelagic shales, radiolarites and limestones are not known in the Bükk Mountains indicating that uplift and subsequent denudation in the Cenozoic removed all Upper Jurassic and Cretaceous formations. Uplift, however, was accompanied by structural activity resulting in the strong and complex deformation of rocks. This Mesozoic succession shows Dinaric resemblances (Balla 1987; Protic et al. 2000; Pelikán 2005).

Since the late Eocene transgression deposited gravels from the Bükk Mountains in the nearshore zone, it seems certain that the Bükk was a dry terrain before the late Eocene. Whether the advancing sea covered the entire mountains in the late Eocene or not remains unclear. Nevertheless, its gravels, marls and carbonate sediments can be found on the southern slopes of the mountains and in the northeastern Little Plateau. Oligocene marine clays and sands are similarly distributed as the Eocene sediments. As middle Miocene marine sediments directly overlie Triassic and Jurassic formations, it is probable that the carbonates were exposed to karstic processes again.

Extended volcanism in the Miocene affected the entire mountains (Hevesi 1989, 2002a, c; Pelikán 2005), with rhyolitic and dacitic tuffs. The middle Miocene sea also inundated the mountains. In the Late Badenian renewed uplift intensified karstification processes again over large areas: the covered karst became dissected by valleys and cave formation was initiated by increased relief. The covered, mixed, allogenic paleokarst was exhumed under Mediterranean climate and turned into a semi-covered, allogenic karst in the Sarmatian–Early Pannonian (Hevesi 1978, 1986).

## 22.3 Structure of the Bükk Mountains

The structure of the Bükk Mountains is arguably one of the most complicated among the Hungarian mountains. The apparent complex ductile and brittle deformation forming its nappe and imbricated structures developed during several deformation phases from the Cretaceous to the Cenozoic. The ductile deformation elements were formed mostly in the late Jurassic and early Cretaceous, when intense structural activity impeded the development or removed the sediments.

According to Balla et al. (1981), Balla (1987) and Csontos (1988, 1999) the structural elements and deformations are the result of nappe formation, including the areas of both the Little and the Big Plateaus. Other models (e.g. Pelikán 2005) consider continuous structural development with double phases of folding and thrust development resulting in imbricate structures. The complex interaction of two folding systems as a result of two stress fields is agreed by the alternative model of Kozák et al. (2002) and Kozák and McIntosh (2011). In their view the interaction of two deformation systems resulted in very complex structures and the orientation of most morphological elements (even of karst forms) like valleys and ridges is controlled by the dominant structures (Fig. 22.3). Similar results were obtained by Demeter and Szabó (2008) who studied the relationship between the geological conditions and morphology.

Most authors agree that the dominant structures in the Bükk have a southern vergence. This could be the result of backthrust near the surface during movement towards the north in a primarily compressional stress field (Kozák et al. 2002; Püspöki et al. 2012; McIntosh 2014). In this case, northward verging structures would be revealed in greater depth in the root and in the foreground of the Bükk.

The Bükk Plateau is distinct on the Digital Elevation Model of the Bükk Mountains (Fig. 22.1) by the steep slopes along its boundaries. The Big Plateau rises a few hundred metres higher than both its northern and southern foregrounds, composed of Palaeozoic rocks and Jurassic formations, respectively.

The alignments of the major valleys and ridges are correlated to three joint types occurring during brittle and ductile deformation caused by two partly interacting stresses (Fig. 22.3): $S_1$ acted from the SW, from an average direction of 220°, while $S_2$ acted from the SE, from 130°. Some of the valleys correlate to two different types of joints of the two stress fields (McIntosh and Kozák 2013; McIntosh 2014). Although relative relief in the interior of the plateau is limited (60–120 m), the orientation of geomorphic features here also correlates with the direction of major faults and fractures (McIntosh 2014).

## 22.4 Topography of the Bükk Plateau

The above structural properties make the central Bükk one of the most dissected landscapes of Hungary. Its valley density and relative relief (almost 200 m/km$^2$) are exceptionally high in Hungary. Although the average dissection of the karst plateau is lower, its marginal areas show even higher dissection due to headwater erosion of valleys and periglacial processes being strong on the cliffs (Martonné 2001) (Fig. 22.1).

The Bükk Plateau itself is an early Tertiary karstic peneplain (Pinczés 1968; Tóth 1984; Hevesi 1986, 1990, 2002c; Pécsi 1988). According to Hevesi (2002b), it is an exhumed autogenic and exhumed mixed autogenic-allogenic karst.

The Bükk Plateau is enclosed by a cliff series, of more than thirty marginal peaks higher than 900 m, the famous "Bükk Rocks", which dominate the view from the south, the highest being Kettős-bérc ("Double peak", 961 m). Their origin is rather complex. Strömpl (1914) and Leél-Őssy (1954) described them as a fault scarp, while others identified them as planation (Pécsi 1988) or pedimentation steps (Tóth 1984). According to scientists today (Hevesi 2002a), the edge represents the thrust faults of nappes and imbrications along the boundary between Jurassic siltstones, shales, radiolarites and Triassic limestone formations of very different denudation parameters. Others (Baráz et al. 2009) believe that limestone formations continue across the cliff edges in the South-Bükk and the extreme relief can be associated with the denudation of heavily fractured strata

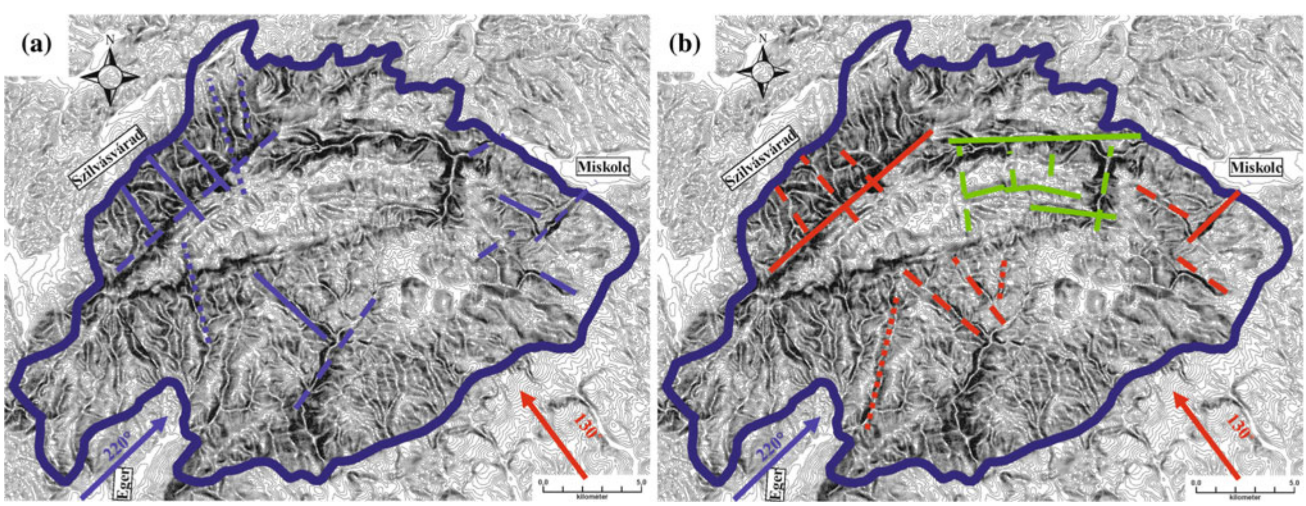


**Fig. 22.3** Correlation between the orientation of major valleys/ridges in the Bükk Mountains and the average direction of major faults/fractures (by Richard William McIntosh). **a** Correlation between valleys and main $S_1$ faults/fractures; *solid blue lines*: $S_1$ frontal fractures; *dashed blue lines*: $S_1$ transverse fractures; *pointed blue lines*: $S_1$ conjugate fractures. **b** Correlation between valleys and $S_{1\text{-}2}$ and $S_2$ faults/fractures; *solid green lines*: $S_{1\text{-}2}$ frontal fractures; *dashed green line*: $S_{1\text{-}2}$ transverse fractures; *pointed green line*: $S_{1\text{-}2}$ conjugate fractures; *solid red line*: $S_2$ frontal fractures; *dashed red line*: $S_2$ transverse fractures; *pointed red line*: $S_2$ conjugate fractures. *Arrows* indicate the direction of stresses (see text)

near the axis of the former anticline. Based on these characteristics (Dobos 2001) and our own structural analyses one may conclude that the two ideas complement each other rather well since there are zones where the role of thrust faults and others where that of the strata of the anticline structure are more important in the formation of the cliff series. Nevertheless, the less resistant siltstones and shales along the weakened rock boundary were denuded more intensively (Hevesi 1986, 2002a; Dobos 2001) when the plateau was uplifted. As a result of frost weathering in the Pleistocene glacials the marginal peaks were transformed into cryoplanation cliffs along the range of Bél-kő, Sándor-hegy, Oltár-kő, Hegyes-kő, Pes-kő, Tar-kő, Három-kő, the most spectacular of them are Tar-kő (950 m—Fig. 22.4), the quarried pinnacle of Bél-kő and Vörös-kő.

The northern edge of the plateau is dissected by valleys with deep gorges, which are inherited forms cut by rivers into the Early Miocene pediment, into the hard Devonian-Carboniferous and Triassic limestones after the Miocene molasses were eroded in the Sarmatian-Pannonian (Hevesi 1989, 2002a; Szalai et al. 2002). Diverse landforms occur in the valleys, dropping 200–300 m at places governed by the structural-petrographical properties. Steepest of the valleys in the NW is the Istállós-kő-lápa, which descends 200 m over a distance of 600 m. The Leány ("Girl") Valley is formed partly on cherty limestone and the Ablakos-kő (Rock-with-Window) Valley developed on diverse lower Triassic shallow marine sediments. Both abound in unique landforms and flora. In the Ablakos-kő gorge sunlight beams through the *"windows"* of 30–40 m high standing limestone "ribs" (blades) onto the scree at the foot of the pinnacles. The "windows" (holes) are either the remnants of former cave passages or niches produced by frost weathering. In the Csondró Valley screes cover the foot of the rock walls that bear the signs of former abrasion platforms under the half arcs of collapsed caves. Cryoplanation cliffs hiding collapsed caves (Odvas-kő) hang over the steep stream bed with waterfalls and travertine steps. In the lower sections klippen of Palaeozoic limestone lenses occur.

The plateau landforms developed not only along valleys perpendicular to the major structural lines but also parallel to them. Fine examples include the above mentioned Garadna valley running from west to east and dividing the plateau into two main compartments, the Lusta ("Lazy") Valley along a fold axis, and the Forrás ("Spring") Valley in the Little Plateau. Wide sections in non-karstic rocks alternate with narrow gorges rich in karst and frost weathering features in limestones. The Lusta Valley is enclosed by valley sides developed in different rock types, hence it is asymmetric in cross-section. The Forrás Valley joins the doline rows of the Little Plateau and its valley head is formed in a collapse doline, probably caused by the drop of the base level of the Sajó River (Hevesi 2002b, c).

## 22.5 Surface Landforms of the Bükk Plateau

It is probable that karst formation started earliest on the highest surface, the Bükk Plateau, after its cover of marine sediments and volcanic tuffs was removed. The remnants of the cover can be found in hidden dolines and collapsed caves of shallow karst valleys. On the uncovered parts of the

**Fig. 22.4** View of the Big Plateau from Tar-kő looking towards ENE with Három-kő in the foreground (Photo by Gabriella Szlatki)

**Fig. 22.5** Karren on the escarpment of Őr-kő (photo by László Sütő)

plateau karst processes may appeared as early as the Sarmatian (Hevesi 2002a, c). Based on the palaeoclimatological conditions and on the widely accepted presumption that the Bükk Plateau was exhumed at the end of the Cretaceous it can be suggested that subtropical cone karst landforms developed in the Eocene and Oligocene, however, their signs have not been verified undoubtedly yet (Hevesi 2002a, c). Typical macroforms of the plateau are peaks and valleys in which various karren fields, ponors, dolines and cave types were formed (Figs. 22.5 and 22.6). In contrast to the marginal cliffs, the peaks in the interior plateau area are gently sloping ridges with a relative height of 50–150 m. They are probably the remnants of the former denudation surface. Shallow valleys (in Hungarian "lápa") in between them were inherited from the former Miocene cover. In the strongly deformed rocks both karstic solution and frost weathering left their traces.

On the limestone surfaces of the peaks scattered crests of karren emerge. They are mostly formed under soil cover and are subsequently exposed to the surface when the soil cover is removed. Most beautiful among them are the karren of Cserepes-kő and Őr-kő (Hevesi 2002c).

The shallow karst valleys of Nagy-mező (Fig. 22.7) and Zsidó-rét formed at the junction of several inherited valleys are especially rich in minor karst landforms (Hevesi 1986, 2002a, b, c): dolines, twin dolines (uvalas), collapse dolines and valleys with doline rows generate a specific microclimate with associated unique endemic and remnant species in mountain meadows. The origins and dimensions of the dolines (occasionally 10 m) are very variable. The oldest are the shallow single hanging dolines near the top. These were probably ponors 30–70 m above the shallow valleys of the Pliocene paleodrainage (Hevesi 2002c). The denudation rate of the cover can be studied along the valleys with doline rows that follow the valley network pattern of the once covered surface (Hevesi 1986, 2002c; Ferenczy 2002). A fine example for the termination of the valley connection between the plateau and its foreground 2–3 Ma ago can be seen in the Káposztáskert-lápa, the continuation of which runs on the western plateau through the Vörös-kő-lápa towards the Tárkány Basin (Hevesi 2002a). The doline rows along the deepest line of the low-gradient wide valley show the retreat of subsurface capture simultaneous with the removal of the sediment cover. Most dolines are asymmetrical (Jakucs 1977) that can be explained by the gentle dip of the plateau towards the east—by structural reasons (Telbisz 2001, 2004) and partly by their ponor past (Hevesi 2002b)—although ponors are not present in all of the dolines. Bowl-shaped dolines are common (Zámbó 1970; Hevesi 1986) and result from merging of several smaller dolines (Hevesi 1986, 2002b, c; Veress and Zentai 2009). Unique features include the "quadruple" doline of Nagy-mező, the Tányéros-teber ('Plate doline') and the Mohos-teber, the largest in the area that may act as a ponor in the case of extreme amount of water (Hevesi 1986, 2002c).

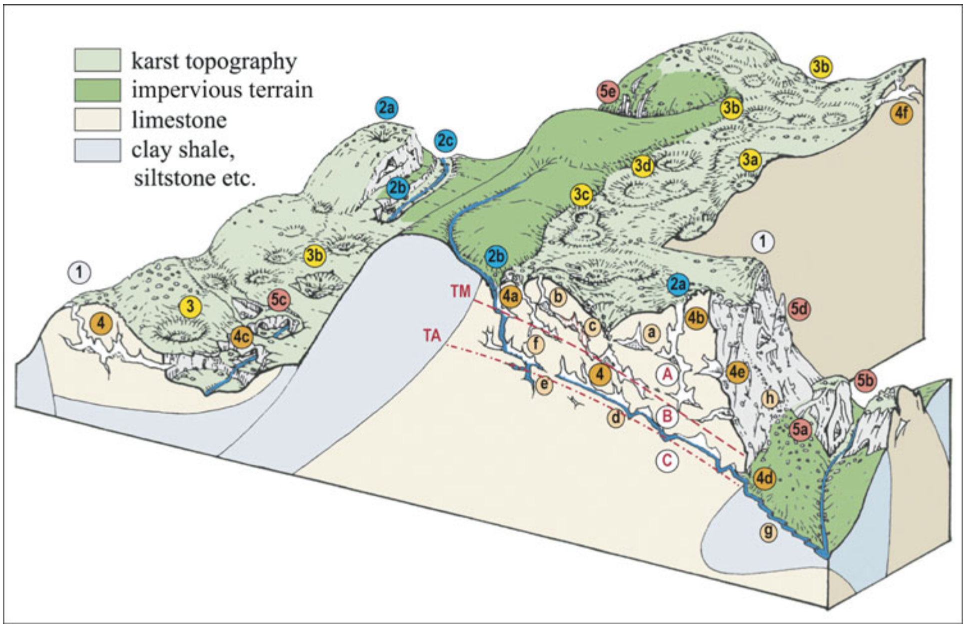


**Fig. 22.6** Karst landforms of the Bükk Plateau (by Csaba Baráz). *A* Aeration zone; *B* Epiphreatic zone; *C* Phreatic zone; *TA* low piesometric surface; *TM* high piesometric surface. *1* Karren field; *2a* Aven entrance; *2b* Ponor; *2c* Pocket valley; *3a* Hanging doline; *3b* Doline row in valley; *3c* Uvala; *3d* Polje; *4a* Sinkhole cave; *4b* Aven (inactive inflow cave) *4c* Stream cave; *4d* Spring cave; *4e* Eroded stream cave; *4f* Decrepit cave; *5a* Collapsed cave; *5b* Breakdown gorge; *5c* Collapse doline; *5d* Tube partition tower; *e* Hogback ("devil's rib"); *a* Cave hall with half tube; *b* Breakdown tube; *c* Stalactite level; *d* Siphon; *e* Internal sinkhole; *f* Rimstone pole; *g* Travertine dam; *h* Cone, debris slope

**Fig. 22.7** Shallow karst valley of Nagy-mező (photo by László Sütő)

## 22.6 Hydrodynamic System and Caves

The Bükk Plateau forms a uniform karst hydrodynamic system (Lénárt 1997, 2002). The precipitation entering the massive limestone through ponors issues into the surface again through the springs of the valley heads. Ponors of variable activity may form not only in the interior limestone areas but also along the boundaries of karstic and non-karstic rocks. In the latter case a blind valley forms on the non-karstic surface. Such a blind valley runs from Örvény-kő ending in Szamentu Cave through two separate sinkholes of the once single overground stream on the former covered karst: Jávor-kút and Bolhás Ponors. The caves of the ponors were connected under the surface 15 years ago (Hevesi 2002b, c).

Around 1,000 caves are found in the Bükk Mountains, out of the 3,500 caves in Hungary (Kordos 1984; Ferenczy 2002). The origin of the caves is bound to the multi-generation fracture system of the strongly deformed rocks and the diverse geological conditions (Ferenczy 2002). The collapsed caves and multi-level active caves were exposed at various altitudes from the top of the plateau to its margins with the uplift of the mountain and then opened by frost weathering during the Pleistocene. Their positions inform about the rate of the gradual uplift of the plateau and the descent of karst water table.

Caves at the highest elevations in Hungary are found in the Bükk. Scattered avens near the top belong to the oldest, Pliocene karst forms (Hevesi 2002b, c). Such is Kőrös-bérc cave (entrance at 932 m a.s.l.—Fig. 22.8), a member of the first spring cave generation (Sásdi 2003), and Kis-kőháti aven (915 m) that was probably a ponor 100 ka ago (Kovács 2003). Collapsed caves at the plateau edge were shelters for hominins in the Pleistocene. Best known are the young collapsed cave with a huge opening on Istállós-kő and the older Szeleta Cave, where artefacts of the Szeleta culture of international fame were found in the margin of the Little Plateau (Regös and Ringer 2003a, b).

The cauldron called Dante's Hell, 20 m across and 20 m deep presumably developed by the collapse of the roof of a cave hall. This closed depression surrounded by vertical cliffs has a specific microclimate (Hevesi 1986, 2002b, c, 2003).

Not only the highest located but the deepest cave is also found in the Bükk Plateau. The entrance of Bányász Cave can be found in a meadow south of the peak of Nagy-Hársas. The strictly protected inflow or sinkhole cave was first explored in 1964. Recent explorations revealed that Bányász Cave is the deepest cave in Hungary with 275 m reached in 2014. It is composed of several parallel vertical shafts with steeply dipping connecting sections. The cave was formed along NE-SW striking structural planes. Dripstone forms are abundant, however, apart from the characteristic dripstone flow many of them have been dissolved indicating that the shafts became inactive and then reactivated again (Rántó A, Sűrű P 2014).

The caves on the Bükk Plateau which formed not entirely in Triassic limestone are also special. They include Fekete Cave of diverse geology on the edge of the Big Plateau near the Garadna Valley. Its specific forms were mostly developed in dolomites, marls and shales as well as in Palaeozoic and Mesozoic limestones (Nyerges 2003). Also crossing porphyrite in addition to the plateau limestone, the Diabáz Cave, which opens at an elevation of 900 m, drives its water to the Garadna spring. The cave is rich in various dripstones.

Travertine precipitations equally occur subaerially and in caves (Hevesi 1972). One of the largest and best known examples are the brimstone dams of the Fátyol ("Veil") Waterfall near the head of the Szalajka Valley at the foot of Istállós-kő (Fig. 22.9). The travertine steps along the

**Fig. 22.8** Kőrös-bérc Cave, the highest-lying cave opening (932 m) in Hungary (photo by Richard W. McIntosh)

**Fig. 22.9** The Fátyol Waterfall in the Szalajka Valley, near Szilvásvárad (photo by Csaba Baráz)

Alsó-Sebes-víz on the margin of the Garadna Valley are also spectacular. The largest travertine dam in the Bükk—1,500 m long, 50 m thick, 300 m wide—with the highest (artificial) waterfall of Hungary (almost 20 m high) was formed on bedding planes on the Szinva Stream (Hevesi 2002c). The internationally known Anna Cave, formed by collapses and artificial connecting of several caves in travertine, was explored nearby in Lillafüred at the beginning of the 19th century. In the cave plant remnants are covered by fine calcareous crust (Hevesi 2002c; Lénárt and Hevesi 2003).

## 22.7 Cryoplanation Features

In addition to the prevalent karst features, the Bükk is also rich in periglacial landforms. Cryoplanation steps, frost-riven scarps (Pinczés 1983), comb ridges, shattered niches and talus cones, block streams and rock arches attest to the combined effect of Pleistocene climatic (the intensity of frost action) and lithological controls (rock properties such as porosity, jointing) (Dobos 1997). Differences in the products of frost weathering result from variations in rock type (Pinczés 1983; Dobos 1997, 2001). Periglacial features are particularly well-developed in the Southeastern Bükk, on the surfaces of dark grey shale alternating with resistant grey cherty limestone and sandstone (making up 70 % of the area) (Dobos 1997). In the typical solid limestone stone gates, classic steps and combs formed. Rock and structural boundaries induce the development of cliffs and rock pinnacles. Transitional forms are observed on the sides of the peaks. Along surfaces of weakness in limestones (joints, imbrications, thrusts) "devil's ribs" (blades, pinnacles) are separated from the cryoplanation cliff with stone gates whilst variable forms of scree develop on steep slopes (Hevesi 1986; Dobos 2001; Martonné 2001). Classic screes, almost 100 m long debris cones have various dimensions reflecting the multi-generation joint systems and bed thicknesses. They form blockfields of parallelepiped fractured debris and thinner limestone fragments (Dobos 2001). Slow creep is observed even today and gelisolifluction played a role in the development of slope landforms.

## 22.8 Conclusions

The few examples of the various types of karst forms discussed above are only an introduction to the geomorphology of one of the richest karst landscapes in Hungary. Karst processes active since the Eocene, the vast amount of mainly Triassic limestone formations susceptible to solution, interbedded non-karstic rocks, strong structural deformation and the gradual exposition of the covered karst all contribute to the diversity of the karst landscape. This is the reason why caves, inherited shallow valleys developed from small doline rows and gullies originated by cave collapse and erosion came about in the interior and at the edge of the limestone plateau. Hanging and valley bottom dolines evolved from ponors, whereas complex twin dolines and collapse dolines dissect the surface. Remains of spring and sinkhole caves and active passages at various elevations formed by the dissolution and erosion activity of descending water and cave streams—all derive from karst processes. The karstification of the Bükk Plateau has slowed down under the present climatic conditions, but it has not stopped. Visitors can observe and scientists can study the uninterrupted processes and spectacular products of karstic dissolution.

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# 23 Travertine Mound of Egerszalók: The Hungarian Pamukkale

Borbála Benkhard

**Abstract**

One of the youngest geological formation in Hungary is the travertine mound at Egerszalók, an indirect product of human activity. Despite its young age the travertine mound deposited from hot thermal water rich in dissolved mineral material erupted from a hydrocarbon exploring well has an impressive size. It is often called the "Hungarian Pamukkale" by journalists due to morphological similarities with the world-famous springs in Turkey. The landform assemblage is composed of a number of waterfalls, steps and terraces decorated by microforms like curtains, ridges and organs. Travertine formations are of geologically short duration. Although the travertine mound at Egerszalók increases steadily due to continuous and high discharge, it reacts sensitively to any human interference. Therefore, water supply has to be maintained carefully.

**Keywords**

Spring mound • Travertine • Thermal water • Rimstones • Bükk foothills

## 23.1 Introduction

Considering the freshwater limestone (travertine) formations of the Carpathian Basin, the mound of Egerszalók, North Hungary, is one of the most spectacular regarding its young age, significant size and beauty. Although the borehole drilled for hydrocarbon exploration on the edge of the Egerszalók settlement in 1961 was not productive, thermal water of 65.5 °C temperature (Lénárt 2011b; Szlabóczky 2012) gushed out of the borehole, which reached the thermal karst waters of the Bükk Mountains. The thermal water with significant dissolved material content running down the hillslope has been building the travertine mound ever since. The volume of the mound increased to 2,700–3,000 $m^3$ over 50 years (Lénárt 2011d). The accumulating carbonates formed tufa terraces with varied surfaces and basins (Dobos et al. 2005), extending over 900 $m^2$. The travertine mound of impressive appearance is called the "little Pamukkale", after the well-known UNESCO World Heritage site in Turkey—although Pamukkale is much older and of larger dimensions.

The calcium–sodium–hydrogen carbonate type of mineral water with sulphur content was qualified as medicinal water in 1992 (Lénárt 2011b). Due to its beauty and the healing effect of the water, the landform called "salt dome" by locals has become a visitors' attraction. As a consequence, the land (scape) is under significant human pressure.

## 23.2 Geographical, Geological and Hydrological Settings

The most extensive member of the North Hungarian Mountains is the Bükk, composed primarily of Mesozoic carbonates (Chap. 22). In the southern foothills of the Bükk Mountains volcanic tuffs cover the surface, however, sedimentary rocks are present deep underground (Fig. 23.1). Hydrocarbon explorations started in 1949 and more than 400 boreholes were drilled in the area (Szlabóczky 2012). In 1961

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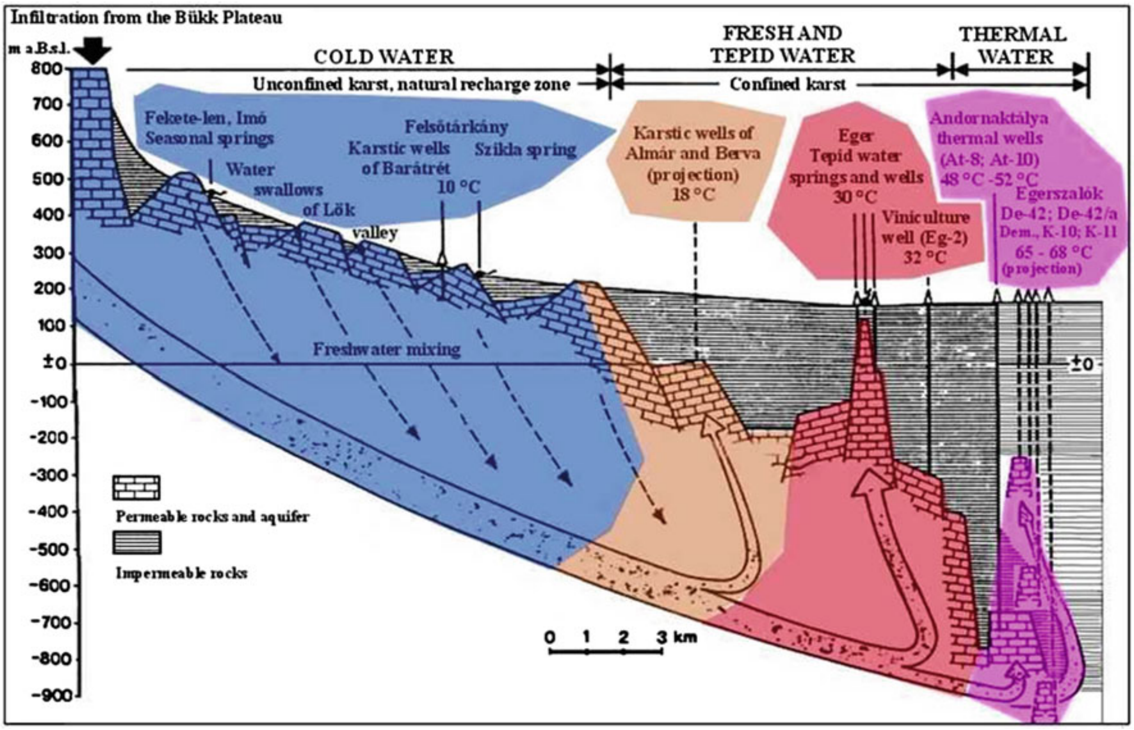



**Fig. 23.1** Thermal hydraulic conditions in the vicinity of Eger (Lénárt 2011a)

the exploration borehole in the Maklány valley in the Bükk Foreland, 5 km southwest of Eger, instead of hydrocarbons, brought thermal water of 65 °C temperature (Lénárt 2011b) from a depth of 407.5 m. Based on radiocarbon dating (Lénárt 2011d), the water from the Late Eocene lithothamnic limestone (Szépvölgyi Limestone Formation) (Lénárt 2011b; Dobos 2012; Szlabóczky 2012) is 27,300 years old.

Soon the utilization of the water for tourism became an issue, but the discharge of and precipitations in the well made the drilling of another well necessary. This was completed in 1987. According to the hydrogeological documentations, this well brought karst water to the surface from the depth of 426 m, from the Buda Marl Formation overlying the Early Eocene limestone (Szlabóczky 2012).

The Bükk region is one of the largest karst water extraction and utilization areas. Percolating cold karst water is heated by the heat flux of the Earth at greater depths. Thermal water of ever higher temperature can be extracted by boreholes from karstic rocks found at an increasing depth with an increasing distance from wells in various karst rocks (Fig. 23.1). As a result of favourable geothermal conditions of the Carpathian Basin, the geothermal gradient values are $G_g$ = 6.52–6.65 m/°C (Lénárt 2011a).

The southern foreland of the Bükk has a significant mineral water and medicinal water potential. Cold water karst springs with constant water supply are abundant as well as warm water (25–35 °C) and hot water (>37 °C) baths (Bogács, Andornaktálya, Demjén, Mezőkövesd) utilizing the water of thermal karst (Lénárt 2011a). The parameters of the water in the thermal well pair at Egerszalók are similar. The few differences (Table 23.1) can be explained by the different flow cross-sections of the filter pipe and production pipe, the depth of the bottom of the wells and the conditions in their environment (Szlabóczky 2012).

The thermal water belongs to calcium-sodium-hydrogen carbonate mineral waters containing sulphur (Lénárt 2011d), with significant metasilicic acid and metaboric acid content (Lénárt 2011b). The free $CO_2$ content of the water is variable (Lénárt 2011d). The water of the first well (De-42) was used for irrigation initially, however, with small success due to its high dissolved solid content. Water from the second well (De-42/a) was declared medicinal water in 1992 and its utilization for bathing and medical purposes was permitted by the Water Management Authority in 2001, limiting the maximum daily discharge to 2,200 $m^3$ (Lénárt et al. 2007). The well with healing water was later renamed Mária well.

## 23.3 Travertine Deposition

The significance of the water gushing to the surface at Egerszalók does not only lie in its healing effect! The high dissolved carbonate content (634–616 mg $L^{-1}$—Lénárt 2011d) induces significant precipitation, experienced when

**Table 23.1** Basic data of the thermal well pair at Egerszalók (*Source* Lénárt 2011d, 2003 and Szlabóczky 2012)

| | Egerszalók De-42 (1961) Vendel well | Egerszalók De-42/a (1987) Mária well |
|---|---|---|
| Temperature (°C) | 65–68 °C | 66.5–68 °C |
| Age (year) | 27,300 | 18,000–20,000 |
| Discharge of the free flow ($m^3$ $d^{-1}$) | 1,260–1,400 | 3,168–3,700 |
| Permitted discharge ($m^3$ $d^{-1}$) | 288 | 2,200 |
| Calcium (mg $L^{-1}$) | 150 | 151 |
| Sodium (mg $L^{-1}$) | 63 | 62 |
| Hydrogen carbonate (mg $L^{-1}$) | 634 | 616 |
| Total dissolved matter (mg $L^{-1}$) | 1,050 | 1,040 |
| Free carbonic acid on site | 460 | 427 |

the mineral water enters the surface due to the high difference in pressure and temperature. The “Hungarian Pamukkale” is evolving in front of our eyes and allows for the thorough observation of geomorphic processes.

The well constructed on the top of the hill in 1961 could not be sealed because of technical errors. The thermal water rich in mineral compounds flowed freely on the hillslope. After a while, local people dug an earth pond to collect water. From 1971 a concrete basin of 38 $m^2$ volume stored the thermal water and not only local people had a bath in it day and night. The water covered a distance of ca 50 m from the wells to the basin along the gentle hillslope of clay (affected by soil formation) on rhyolite tuff (Fig. 23.2). Waters from both wells jointly fed the travertine mound at times. Based on the measurements of Lénárt et al. (2007), 90 mg $L^{-1}$ of carbonate precipitation takes place along the route of the water parallel to the loss of carbon dioxide. This means 1 mm thick carbonate accumulation within a few days. As a result, the volume of the mound, composed mostly of travertine, is 2,700–3,300 $m^3$ today. A part of the mound is covered by vegetation (Fig. 23.3). Only minimum carbonate precipitation can be seen in the stream running off from the mound (Kele 2009).

Travertine cones may be regarded as the most typical products of continental carbonate sedimentation from springwater (Scheuer and Schweitzer 1985) (see also Chap. 8). Carbonates are transported dissolved in springwater; the depositional environment is primarily springwater; inorganic autochtonous precipitations occur in the immediate neighbourhood of a spring or away from the spring, depending on human influence. Lithification is typical and the material immediately hardens after precipitation (Scheuer and Schweitzer 1985). Slope-deposited freshwater limestones (Scheuer and Schweitzer 1986) are well-known from all parts of the world (e.g. the Mammoth Springs, Yellowstone National Park, USA; Pamukkale, Turkey). Although the travertine mound of Egerszalók was not deposited from springwater but from an overflow of drilled wells, its conditions of formation and appearance are similar.

**Fig. 23.2** The travertine mound prior to major investments in 2003 (rounded shape, snow-white terraces) (photo by Borbála Benkhard)

Spring discharge and consequently the rate of limestone deposition depends on the amount of precipitation in the catchment (Scheuer and Schweitzer 1985). On the contrary, the discharge and pressure of the wells drilled in the vicinity of Egerszalók can be regarded as stable (Lénárt 2011d). Therefore, limestone deposition is continuous and evenly distributed, being only modified by human interference. Controlling the discharge of wells, however, is inevitable in order to protect the quantity and quality of karst water in the

**Fig. 23.3** The sight has somewhat changed following hotel and bath construction and the channelling of water (photo by Dániel Puyau)

long term (Table 23.1). The Water Management Authority set the permitted amount of discharge for the first well (De-42) in 288 $m^3$ $day^{-1}$ in 2001 in order to sustain the development of the travertine mound. The water of the second well is utilized for bathing and medical purposes (Lénárt et al. 2007).

As we have seen, the Egerszalók Travertine deposits can be classified as a spring mound, which is a category close to the subgroup of autochtonous (in situ) travertines (Pentecost 2005). On the basis of hydrogeological observations and geochemical analyses (Kele 2008), it is regarded as transitional between thermometeogene and thermogene. The term "thermometeogene" means that the water circulates deep beneath the surface, becomes heated and rises to the surface as a hot spring, but contains only a meteoric component (Pentecost 2005). On the other hand, thermogene travertines are formed as massive deposits from fluids carrying $CO_2$ of hydrothermal origin.

## 23.4 Morphology

The 30-cm-deep basin (although artificially formed, it is covered completely by travertine today) around the drilled well (Kele 2008) is drained by artificial channels, changing their direction from time to time. Thus, the freshly precipitated snow-white deposit will cover an extensive area (Fig. 23.4). The maximum height of the mound is 8.3 m. Due to the topographic conditions and the channelling of outflow the thickness of the limestone is different at various places of the mound. Based on the measurements carried out in September 2013, the total area of the visible (uncovered surface) travertine mound is 822 $m^2$ and its diameter is 24 m today.

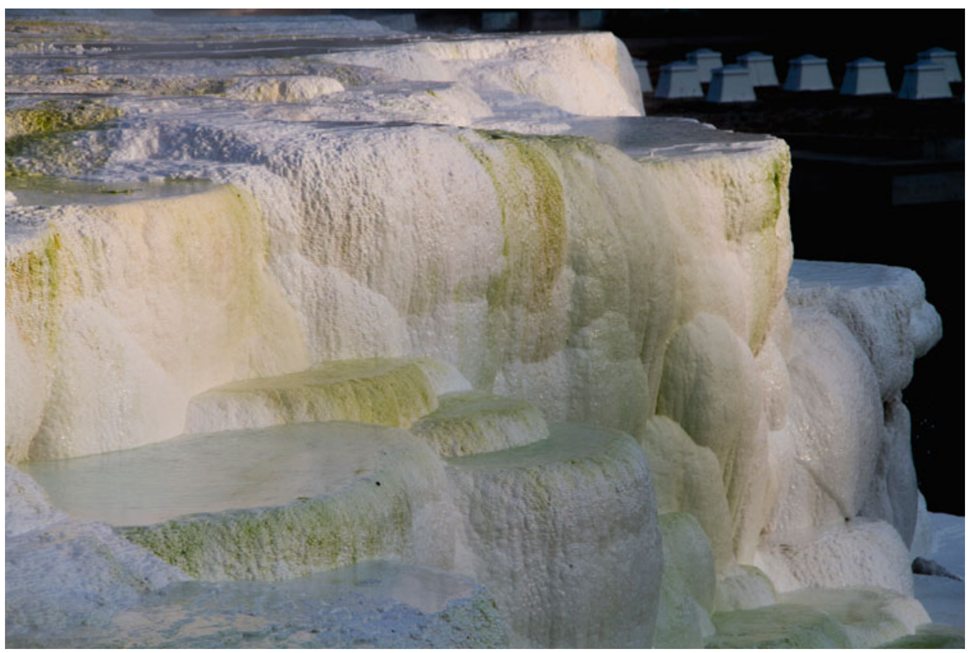
Smaller artificial ponds can be seen in the vicinity of the channels on the top of the mound with calcified bubbles at the bottom, which attest to prior gas release. Outflow from the thermal water channels pours down the proximal slope of an average 45° inclination, following the original relief (Kele 2009) Spectacular rimstones form, which are individual landforms composed of two parts: the dam and the basin (Scheuer and Schweitzer 1986). The dams are arcuate in planform and convex in cross section and steadily rise in height through deposition from overflowing waters.

The terraces and pools (basins) of variable size are located in higher or lower relative positions. Lateral cross-flow and deposition leads to complex structures. Due to human interference younger and older limestones are often located next to each other or may merge (Scheuer and Schweitzer 1970).

The depth of the ponds may reach 20 cm (Kele 2009). The heights of the risers of terraces vary between 0.1 and 1 m. From the waterfalls on the edges of the ponds characteristic crystalline crusts precipitate. These dams are similar to cascades but with impoundment of water behind a travertine crest (Pentecost 1995). Characteristic features of slope-deposited limestones are the rimstones (microtetaratas) which occur either individually or in association with macroforms. These tiny dams, a few cm wide, are formed by the waves of overflowing water almost everywhere (Fig. 23.5). Spectacular elements of the travertine mound visible from the distance are the vertically formed travertine features:

**Fig. 23.4** Water is conducted to appropriate places of the travertine mound through artificial channels in 2013 (photo by Dániel Puyau)

**Fig. 23.5** Microtetaratas on macroforms, 2013 (photo by Dániel Puyau)



curtains, organs, ridges and crests (Fig. 23.6). On the slope more distant from the well head (distal), with inclination below 15°, terraces and other features typical of slope-deposited travertines (ponds and waterfalls) are missing (Kele 2009).

The development of both microforms and macroforms is also influenced by the flora and fauna through chemical processes or vegetation presenting a physical barrier. Added to the variety of microforms, there are gas bubble ducts with a diameter of 3–5 mm (occasionally crystallizing) (Lénárt et al. 2007), the formation of which is induced by buried plant or animal remnants (including algae) in the way of water. These produce gases (e.g. carbon dioxide) in the course of their decay or metabolism making newly spouted water capable of local dissolution. From the water moving upward after dissolving the travertine additional amounts of carbonate precipitate in a circular shape forming pipes 3–5 mm across. Biogenic carbonate is manifested in several forms in Egerszalók, as stromatolites or shrub structures (Kele 2009).

**Fig. 23.6** Curtains, laces, organs, ridges and crests of calcium carbonate precipitations decorate the outer edges of the dams. Travertine precipitated onto the distal slope and coloured to brown by algae in the foreground (photo by Dániel Puyau)

Plants resistant to heat and saline environments are found on or even within travertine mounds (Lénárt 2011c). The presence of bacteria and algae is even more characteristic. Cyanobacteria (*Thiobacillus* sp.) oxidizing reduced sulphurous compounds are only able to live next to the thermal well, where the inner temperature is 67 °C (Dobos et al. 2010). In the two main artificial channels, where water temperature is 45–55 °C, photosynthesizing cyanobacteria formed microbial mats. Sites with lower temperatures can be spotted from a distance because the inhabiting bacteria and algae colour the snow-white terrain into green, brown and reddish brown.

## 23.5 Conclusions

The well pair and their surroundings in Egerszalók require special and careful protection due to their special hydrogeological and geomorphological values. The quantity and quality of the underground thermal water has to be maintained. Thermal water pumping is not permitted and almost the entire western and southwestern Bükk Mountains are declared a protected hydrogeological area These measures are also necessary to prevent detrimental effects of deep borehole drilling or mining (Hojdákné and Iván 2011; Koleszár 2007).

The travertine mound became protected in 1986 as a nature reserve of local significance (0.81 hectares in area). In order to maintain the landform and to allow its further growth, a discharge of 288 $m^3$ $day^{-1}$ is ensured by regulations. In case the travertine receives no or insufficient water, it becomes dry and grey and powdery. Erosion is intensified by disintegration through insolation weathering, deflation, sheet wash, frost shattering and human tramping (Dobos et al. 2005). Eroding spots have been occupied by vegetation locally.

Since the travertine mound of Egerszalók is a hydrogeological and morphological feature very sensitive to water recharge, it requires continuous and careful management. Just like other travertine mounds, it is threatened by both human trampling and breaking of precipitated microforms and water pollution. To prevent further damage, the travertine mound is fenced around, but, thanks to the trails built, the beauty of the "Hungarian Pamukkale" can be enjoyed from a few metres distance.

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# Beehive Rocks in the Bükk Foothills: A Landscape Born in Fire

24

Csaba Baráz

**Abstract**
The southern foothills of the overwhelmingly carbonaceous Bükk Mountains (a member of the North Hungarian Mountain Range) are mostly built up of silicic volcanic rocks. The majority of tuffs welded to various degrees (from soft tuffs to hard welded ignimbrites) are of rhyolitic composition and derive from Miocene Plinian or Ultraplinian eruptions—the type with the highest energy release in Earth history. Quaternary geomorphic evolution (fluvial incision and lateral erosion, aeolian action, sheet wash and weathering processes, primarily frost shattering) dissected tuff sheets into groups of conical mounds and ridges, into which the people of the remote past carved niches and transformed them into the so-called "beehive rocks".

**Keywords**
Rhyolite • Ignimbrite • Ultraplinian eruption • Miocene • Beehive rocks • Bükk Foothills

## 24.1 Introduction

In the Bükk Foothills (Bükkalja), from the Castle Hill of Sirok to the Kecske-kő of Kács, there are several groups of rocky ridges or conical rock mounds with niches carved in their sides. For this reason, such landmarks are popularly called "beehive rocks". Although there are numerous unanswered questions concerning their development, their volcanic origin and Miocene age are no longer debated (Fig. 24.1).

## 24.2 Intense Volcanism with Nuées Ardentes

Volcanic activities can be classified as effusive or explosive. In the latter category volcanic debris (pyroclasts) is ejected from the vent or fissure. Compared to explosions of moderate energy (Stromboli or Volcano types), Plinian eruptions are the most destructive, producing huge amounts of pumice also in the form of pyroclastic flows of dacitic, rhyodacitic or rhyolitic composition (Németh and Martin 2007). The volcanic formations of the Bükk Foothills result from such vehement Plinian eruptions. The gases released transformed the acidic magma into a scum—as clearly seen in the loose vesicular fabric of the pumice solidified after the explosion. According to the depth where the degassing occurs and the volcanic ash appears, two basic types of eruption can be identified (Karátson 1998):

(a) the classic Plinian eruption takes place at great depth, close to the magma chamber, in the lower section of the lava conduit, and produces a 20–30 km high eruption cloud spreading ash over areas of enormous extension;
(b) Ultraplinian eruptions occur at shallower depths (close to the crater or in a previously developed lava dome) through explosive degasification of high-density lava. The still hot nuée ardente ("glowing cloud") sweeps over the surface and causes widespread destruction. The collapse of the eruption cloud may also lead to such a pyroclastic flow (Fig. 24.2).

From the cloud of the classic Plinian eruption rhyolite tuffs originate. The pyroclastic material in the high-rising eruption cloud consists of volcanic ash, a mixture of fine

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**Fig. 24.1** Panorama of the Szomolya Beehive Rocks nature reserve with the South-Bükk Mountains in the background (photo by G. Bakó)

vitrous and crystalline debris of grain sizen smaller than 20 mm that can travel thousands of kilometres with the winds. The wind-blown pyroclastic material is winnowed and the rock deposited from it is well-sorted composed of debris of uniform grain size. This loose pumice is not welded and contains vitrous material and disrupted blocks of rhyolitic magma (Fig. 24.3).

The welded ash-flow tuff (ignimbrite) produced by the nuée ardente is similar to lava rock. The material of pyroclastic flows is partly re-melted at the high temperature (500–750 °C) and high gas content and flattened and compressed at the great thickness and pressure accompanying the eruption. In the welded tuff fiamme (flattened vitrous debris) are found.

The main characteristic of ash-flow tuffs is the chaotic arrangement of grains (poor sorting), which results from the whirling flow within the glowing clouds. However, higher-density solid debris (xenoliths entrained from the vent during the eruption and earlier solidified lava, lapilli) often deposit at the bottom of the pyroclastic flow and are overlain by lighter pumice blocks (Fig. 24.4). This layer is generally white or gray in colour and its texture is called sillar after the building stone quarried at the foot of El Mistí volcano in Peru (Kanen 2001). The hot gases trapped in the deposit are released through vertical conduits (Pentelényi 2005; Harangi 2011).

## 24.3 Plate Tectonics Background

The plate tectonic background to the formation of pyroclastic flows (recently also mentioned as pyroclastic density currents) in the present-day Hungary is well-known (Császár 2005; Harangi 2011). The ash flows are regarded to document the initial phase of the development of the Pannonian

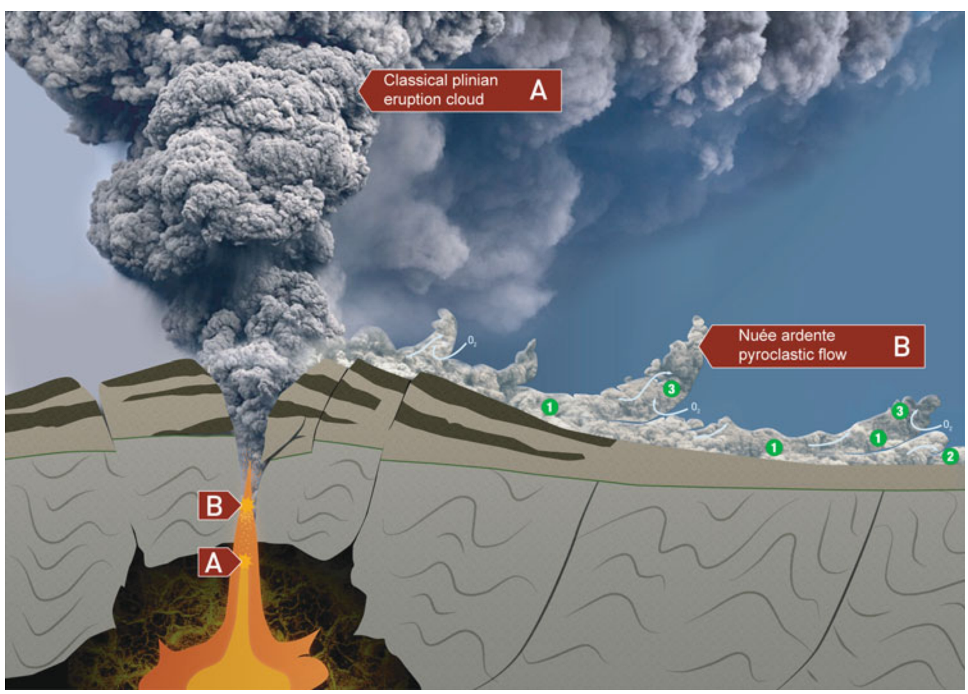


**Fig. 24.2** Eruption column and pyroclastic flow. *A* Classic Plinian eruption. *B* nuée ardente. *1* Fluidized pyroclastic flow; *2* turbulent pyroclastic surge; *3* ash cloud above nuée ardente

**Fig. 24.3** Exposure of the Gyulakeszi Rhyolite Tuff Formation at Szomolya, abounding in pumice (photo by Csaba Baráz)

(Carpathian) Basin, which happened through the collision of shifting lithospheric microplates (the Tisza and Alcapa) from the Middle Miocene (ca 50 Ma ago) on (Chap. 2). The movements involved crustal contraction estimated at 400 km on the average. The collision and subduction produced huge amounts of neutral to acidic magma, which erupted to the surface in the andesitic-rhyolitic Inner Volcanic Range of the Northeastern Carpathians.

Subsequently, between 20 and 13 Ma, on the floor of the seas occupying the Basin, highly explosive volcanic activity along circular tectonic lineaments was typical. The collapsing eruption columns gave rise to glowing clouds, spreading laterally along the hillslopes, setting to flames everything all along their way. The welded tuff (probably ejected from volcanic fissures) mantled thousands of square kilometres (Fig. 24.5).

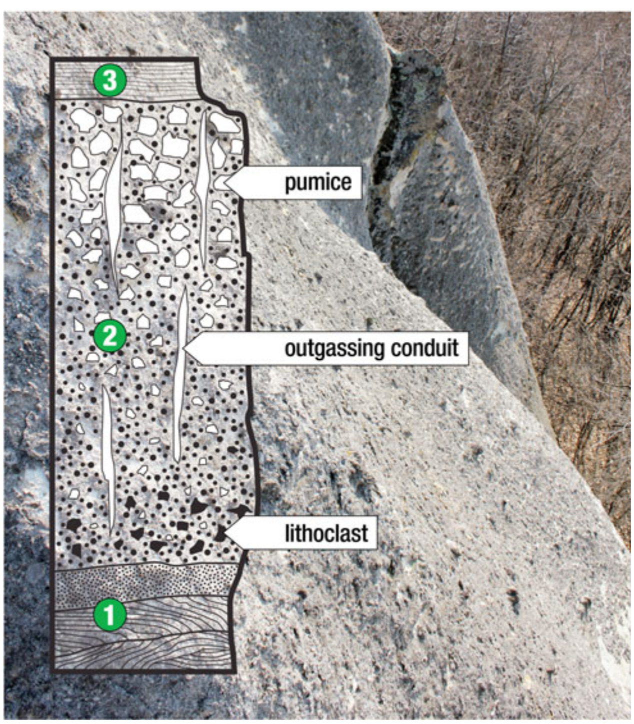


**Fig. 24.4** Cross-section of the pyroclastic flow (by Csaba Baráz). *1* Basal (surge) layer; *2* middle (sillar) layer; *3* upper (fallout) layer

## 24.4 Volcanic Sequences in the Bükk Foothills

The Miocene volcanic sequence of several hundred metres thickness can be subdivided into three series (Balogh 1964; Hámor et al. 1980; Varga 1981; Lukács et al. 2010):

(a) In the lowermost position is the oldest series, the Lower Miocene (Ottnangian) Gyulakeszi Rhyolite Tuff Formation (previously called "lower rhyolite tuff"). Most of it is fall-out tuff with high proportions of pumice, quartz and biotite. The welded tuff of the pyroclastic flow is identified as the Kisgyőr Ignimbrite Member. The varieties with perlitic-obsidian fiamme ("pseudofluvial rheoignimbrite") had long been classified as lavas.

The activity normally started with a Plinian eruption of local pyroclastic surge, tuff avalanche or with phreatomagmatic explosions. Ash flow was typically generated in the terminal phase of the eruption. The thickness of the volcanic formation ranges from 150 to 450 m. The eruption was centred on a fissure in the southeastern

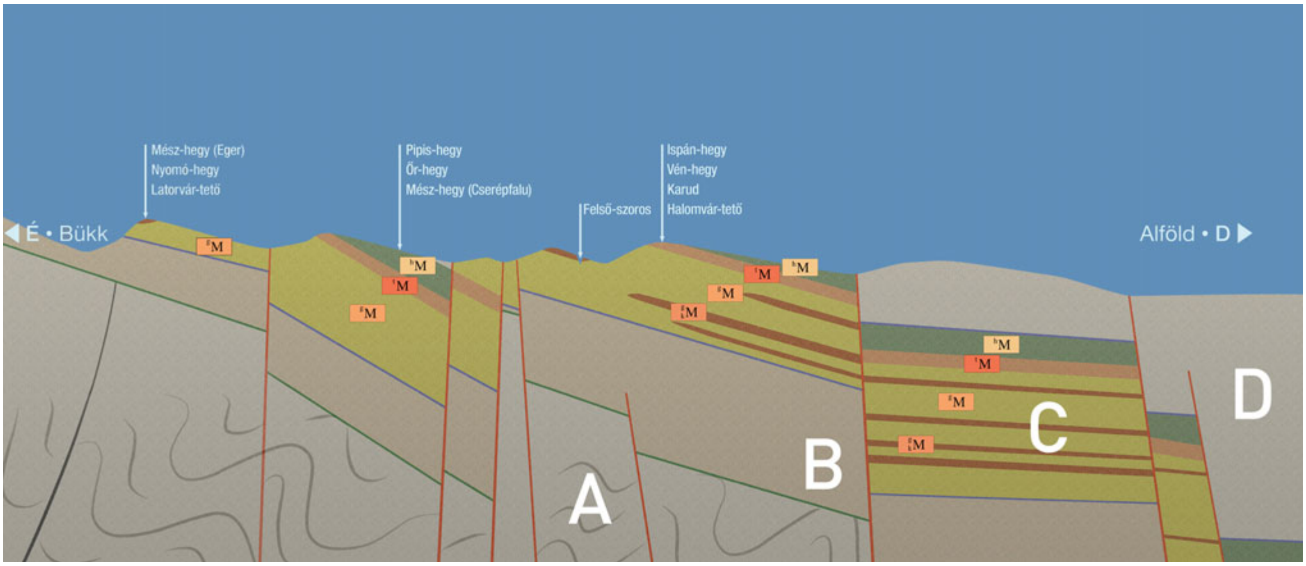


**Fig. 24.5** Geological cross-section of the Bükk Foothills (by Cs. Baráz). *A* Mesozoic formations. *B* Paleogene-Miocene prevolcanic sediments. *C* Miocene volcanites: *gM* Gyulakeszi Rhyolite Tuff; *gkM* its Kisgyőr Ignimbrite Member; *tM* Tar Dacite Tuff; *fM* Felnémet Rhyolite Tuff; *hM* Harsány Rhyolite Tuff. *D* Pliocene-Quaternary postvolcanic sediments

**Fig. 24.6** The Felső-szoros (Upper Gorge) cut into terrain built up of the Kisgyőr Ignimbrite Member (Photo by Csaba Baráz)

foreland of the Bükk Foothills—now buried under younger sediments. This tuff series was dated radiometrically to 21–18.5 Ma BP (Hámor et al. 1980) (Fig. 24.6). The Kisgyőr Ignimbrite Member outcrops in an inner (closer to the Bükk Mountains) and an outer range (closer to the Great Hungarian Plain).

(b) Towards the end of the Karpatian age another series of eruptions produced the dacitic ignimbrites of the Tar Dacite Tuff Formation ("middle rhyolite tuff"), where fall-out and reworked tuffs are subordinate components and pyroclastic flows and ignimbrite varieties welded to various degrees (Bogács Ignimbrite Member) are typical. Enriched in pumice, scoria, perlite and obsidian fiamme, the non-stratified series of 30–50 m thickness locally also displays lava character (rheoignimbrite). The radiometric dating showed ages of 17.5–16 Ma BP (Hámor et al. 1980).

The rhyodacitic-andesodacitic material is regarded by some as acidic and intermediary magma derived from the melting of the upper mantle (Póka et al. 1998), while others hold the view that it was generated by the mixture of calcalkali magma from the upper mantle and material from the lower mantle (Lukács 2000). The alternating dark andesitic (scoria) and the light-coloured rhyolite (pumice) of the Tar Dacite Tuff are undoubtedly the products of magmatic mixture. The magma chamber, situated at several kilometres' depth, had been developing for thousands of years and the crystallizing mass was differentiated. The melted silica-rich lenses, unaffected by crystallization, were remobilized by upthrusts which pushed them up to the surface (Lukács 2009). The ignimbrite member is also exposed in both the inner and the outer tuff ranges.

(c) After another period of quietude volcanic activity resumed in the Late Badenian period and lasted to the Late Miocene. The centre of eruption was far from the Bükk Foothills and predominantly ejected fall-out, phreatomagmatic (spherical concretions, tuff fetches) and reworked rhyolitic tuff, tuffite and diatomite, referred into the Harsány Rhyolite Tuff Formation ('upper rhyolite tuff'). The 150–300 m thick tuff series (even thicker in buried position) totally lacks welded ignimbrite varieties. According to radiometric dating, the formation is 14.6–13.5 Ma old (Hámor et al. 1980).

The Badenian and Sarmatian pyroclasts in the western part of the Bükk Foothillls, in the hilly region between the Mátra and Bükk Mountains, are called the Felnémet Rhyolite Tuff Formation (Pentelényi 2005).

## 24.5 Geomorphic Evolution

There is a marked topographical contrast between the Bükk, mountains mostly built up of limestones and shales, and the Bükk Foothills (Bükkalja) of primarily volcanic rocks. Running along faultlines, their boundaries are also clearly defined.

As early as the Late Pannonian (Late Miocene) Lake Pannon receded and pedimentation began on the southern margin of the uplifting Bükk Mountains. A double pediment was formed by sheet wash and subsequently dissected by stream erosion, particularly active during warmer and wetter interglacials in the Pleistocene. By the end of the Pliocene epoch the older pediment had only survived in isolated remnants. The intensive dissection of the foothills by a dense drainage network was also exacerbated by tectonic movements. Stream incision was accelerated by the Late Pliocene–Early Pleistocene subsidence of the Great Plain and resulted in a hilly region of relatively high relief by the Early Holocene (Fig. 24.7). In the NNW to SSE aligned broad main valleys Pleistocene and Holocene terraces occur locally. The SE to NE alignment of the tributary valleys and interfluvial ridges was controlled by tectonic lines and the strike of exposed welded ignimbrite zones.

In the colder phases of the Pleistocene glacials, however, particularly in times with short and cool summers as well as cold and dry winters, fluvial action was largely replaced by frost weathering (frost shattering, cryoplanation—Csorba 1982; Pinczés 1985) and mass movements.

In the valley sides built up of rhyolite tuff, the markedly conical beehive rocks were carved by weathering, aeolian and sheet wash processes in conjunction with gully erosion. The tuff towers show evidence of rainwater, wind and weathering action, while the steeper columns are mainly due to frost weathering. The spectacular "rock sacks" of the Felső (upper) gorge (Fig. 24.5) are detached from the margin of the ignimbrite plateau by frost action (Erdős 1972; Borsos 1991; Dobos 2002).

Selective erosion is assumed to spare the physically more resistant (harder) spots and create ridges and towers. In welded ignimbrites the turbulent, fluidized pyroclastic flow showed an inhomogeneous distribution of heat and this was reflected in the various degree of welding and resistance to erosion. The thin section analyses, however, indicate that the beehive rocks are not constituted of strongly welded tuffs, rather showing sillar texture. The most spectacular conical rocks of the Gyulakeszi Rhyolite Tuff Formation derive from pyroclastic fall, reworked by debris avalanches.

If we assume that higher resistance to erosion is not only due to processes active after rock formation, such as secondary silicification, carbonification and vitrification, emergent lithological differences could influence the formation of beehive rocks in the first place. It is important to note, however, that the tuff cones do not differ from their immediate environs either in mineral or rock composition. The microscopic investigation of the rocks (thin sections analyses) did not show the impact of any of the mentioned secondary petrological transformations (Borsos 1991).

What could then give protection to the rock formations against weathering and lowering? In the shaping of tuff cones the crusting of rhyolite tuff surfaces is assumed to play an important part. The crust is a product of interactions between rain, air and living organisms (Borsos 1991). Chemical weathering resulting from the action of rainwater and groundwater dissolves the rock, while sheet wash removes the clayey products, leaving behind resistant materials in the form of a hard crust, which develops further

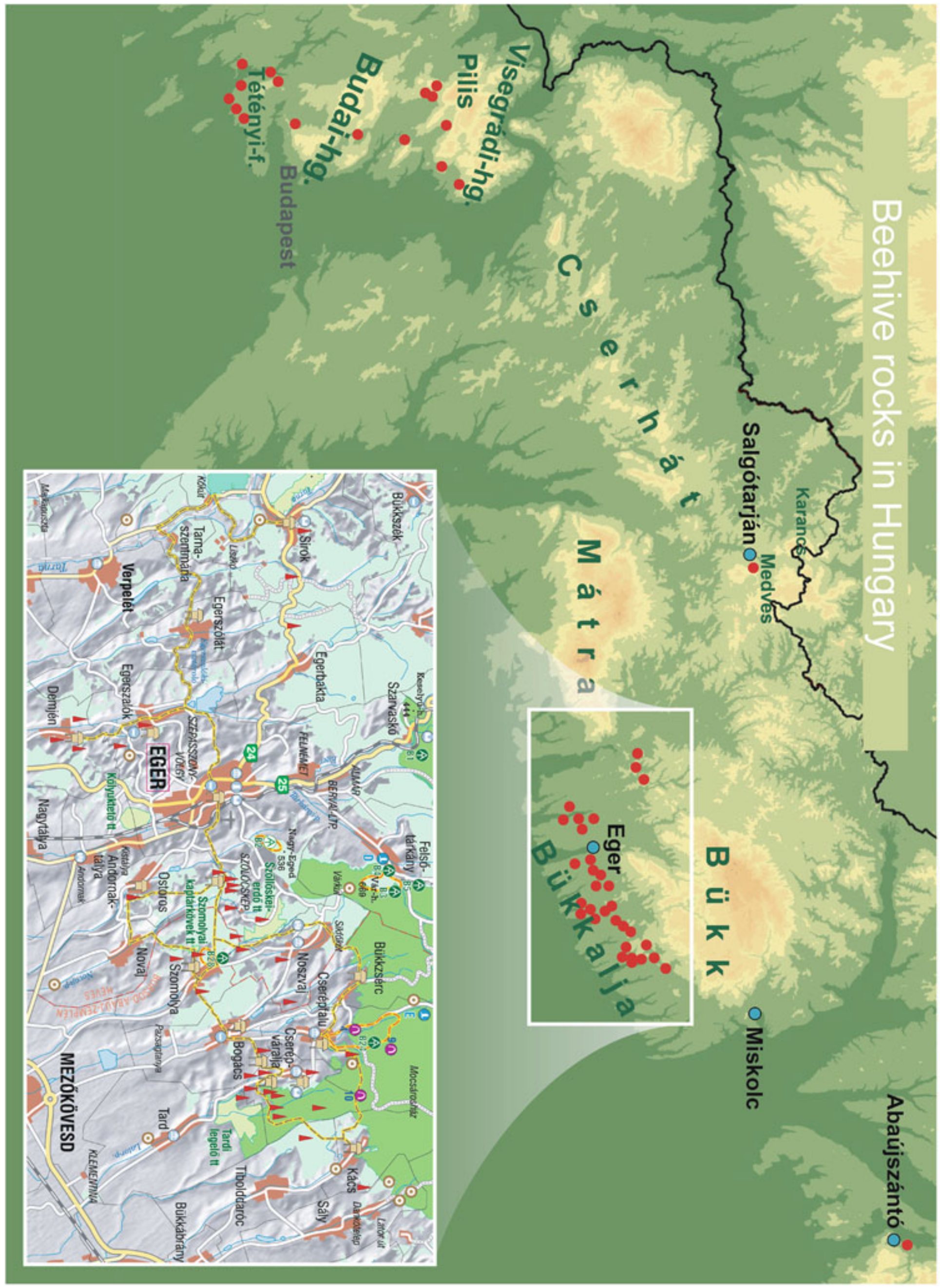


**Fig. 24.7** Map of distribution of beehive rocks in Hungary (*insert map* in the Bükk Foothills)

on barren exposed rock surfaces. Even if it is detached in patches, crusting resumes on the newly exposed surfaces and preserves the landform. Sheet wash on the barren rock surfaces of valley shoulders is replaced by rill erosion on midslopes. On the slopes of barely welded tuff dendritic rill networks develop, occasionally deepening into gullies (badland formation).

As we have seen, vegetation-free surfaces favour beehive formation. Such barren surfaces used to be common in the Pleistocene glacials, while in the Holocene wetter and warmer climate induced afforestation of hillslopes. As a consequence, the evolution of beehive rocks slowed down but did not stop, and it continues to our days. As they are vulnerable to destructive processes, their protection is an important task of nature conservation.

## 24.6 Types

The typology of beehive rocks reflects their tectonic or geomorphological settings. Where the rock mass is affected by faultlines, an irregular, rough rock face develops, for

**Fig. 24.8** The Ördögtorony (Devil's Tower) on the Mangó Plateau Cserépváralja (photo by Csaba Baráz)

instance, in the rock group of the southern slope of Mész Hill and on the western side of Nyerges Hill near Eger (see map in Fig. 24.7). Extremely dismembered cliffs (nos. IV, V and VII) are found on the southwestern slope of Vén Hill of Szomolya, in the Kaptár ("beehive") Valley. Their shapes are closer to irregular ridges than cones. Although faultlines are not necessary conditions to the development of beehive rocks, they exert fundamental morphological controls on them. Where no tectonic impact is observed, rainwater erosion carves out halves and quarters of cones from the rock face, for instance at rock no. IV of Cakó Plateau (H.2.a). Sheet wash causes gradual rockwall retreat as protruding sections are detached into isolated cones, only connected by a narrow "neck" (e.g. the Ördögtorony (Devil's Tower) of Mangó Plateau, B.4.i—Fig. 24.8). On gently sloping terrain rows of rounded hills with convex slopes form (rock no. I of Mész Hill, H.2.c).

While the material of the most spectacular beehive rocks shows no welding at all, the more resistant ignimbrite cones have rather irregular shapes (Vén Hill of Cserépváralja) or form rock benches (Pipis Hill of Noszvaj) or balanced rocks (Köves-lápa of Cserépváralja) (Baráz 2005).

Fine examples of the final stage of beehive rock formation are described from the Furgál Valley (Fig. 24.9). For cone no. IV, the neck connecting it with the slope is still recognizable, but the conical rock itself is low with a steep slope towards the hill. This neck is already missing in the cases of the beehive rocks nos. I, II and V, the Devil's Tower of Cserépfalu (Fig. 24.8) and the Csordás valley tuff cones (B.4.h. I and II).

## 24.7 Terminology and Distribution

In the Carpathian Basin, the Bükkalja shows the longest tradition of stone carving, building stone quarrying and the utilization of stones in vernacular architecture (Fig. 24.10). One of the oldest manifestations of this tradition is the conical rocks with niches, called "beehive rocks" in the village Szomolya, while elsewhere other names are used: "blind-window rocks" around Eger, churning rocks, devil's tower, saddle rock, pointed rock, goat rock, window rock, king's chair, stone dragon and several others. Of all these names, beehive rocks is used most commonly and, therefore, it is borrowed by the authors of scientific studies. The first investigator of rocks with niches was a priest in Eger, historian and archaeologist Gyula Bartalos, in the late 19th century. He used the terms "monument stone/cliff" most consistently, but also mentioned them as "carved cliff group", "monument rocks with blind window", "megaliths" or "churning rocks". It was Andor Saád who made the term beehive rocks widely accepted (Saád and Korek 1965). Later this became the standard denomination in the works of researchers who wanted to solve the mystery of beehive rocks (Bartalos 1891; Baráz and Mihály 1995–1996; Baráz 1999, 2000).

Although they also occur in the Pilis and Buda Mountains (particularly on the Tétény Plateau), the area richest in beehive rocks is near Eger, Szomolya and Cserépváralja and some others are found near the villages Sirok, Egerbakta, Egerszalók, Ostoros, Noszvaj, Bogács, Cserépfalu, Tibolddaróc and Kács (Fig. 24.7). In the Bükk Foothills 77 rocks at 40 sites have been inventoried (Baráz 1999; Baráz and Kiss 2010). There are altogether 482 niches carved in the rocks. They are 60 cm high on the average, 30 cm wide and 25–30 cm deep. On the still intact rocks, a frame lines the niches and occasionally even holes can be detected on it—indicating that the niches used to be covered and the cover slab was fixed by wedges hit in the holes.

The beehive rocks occur in prominent topographical positions: mostly in valley sides, close to summits or halfway downslope (Fig. 24.11). As far as slope exposure is concerned, in their majority they are found on southwestern slopes—often with no counterpart on the opposite slope. This fact suggests that solar radiation may be an influential factor which promotes desiccation, the loss of vegetation and soil on these warm slopes.


**Fig. 24.9** Types of beehive rocks. **a** Conical stone in the Furgál Valley (Cserépváralja). **b** Balanced or perched rock type (Karud Hill, Cserépváralja) (photos by Csaba Baráz)

**Fig. 24.10** Cave dwellings at Noszvaj (photo by Csaba Baráz)

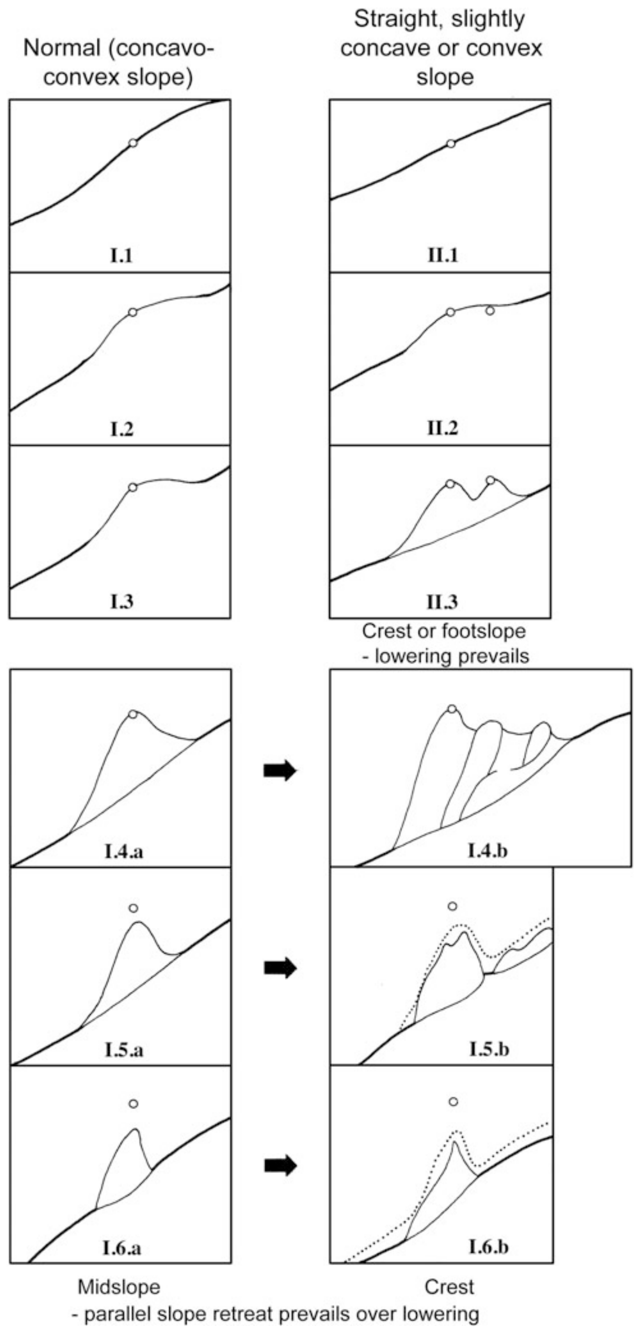


**Fig. 24.11** Development of various types of beehive rocks (by Csaba Baráz). Codes indicate successive stages of development. *Circles* allow the comparison of altitudes during denudation

## 24.8 Conclusions

The Miocene volcanic sequence of the Bükk Foothills (rhyolites, dacites and andesites) takes the form of a disrupted tuff and ignimbrite cover of rift origin. Petrological composition, however, is often not decisive for the resistance to erosion: unwelded loose tuffs, medium welded tuffs and hard ignimbrites equally occur in the area.

The geological and geomorphological issues arising concerning the origin and evolution of beehive rock formation need further study. According to our present knowledge, the various rock formations cannot be viewed as stages in a single evolution sequence. A range of factors have influenced their evolution: mineral and rock composition, the degree of welding, secondary petrological processes (silicification, carbonification, vitrification) as well as geomorphic processes dependent—among others—on topographic position.

Equally intriguing questions can be raised concerning the function of the multiple niches on the rock surfaces (Baráz 2000; Baráz and Kiss 2010; Bartalos 1891). It is not known who, when and why carved them into the rock.

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# 25 Loess Features on Tokaj Hill

Attila Kerényi

**Abstract**

A variety of landforms have developed on the 5–30 m thick loess which mantles Tokaj Hill, a solitary elevation (516 m) of volcanic rock in Northeast-Hungary, on the right bank of the Tisza River. Following a brief history of loess research in Hungary, the chapter presents specifically the findings of investigations into the loess cover at Tokaj. Major erosional features on the loess mantle of Tokaj Hill include sunken lanes, gullies, pipes and subsidence hollows. Based on measurements carried out on a slope with cultivation terraces, the relationships between superficial and subsurface loess erosion are presented in detail. The role of human activities in the removal of the loess mantle is also considerable on the hill, which gave its name to the best-known historical wine-producing area of Hungary.

**Keywords**

Loess • Sheet wash • Piping • Gullying • Hollow roads • Sinkholes • Viticulture • Mitigation • Tokaj Mountains

## 25.1 Introduction

Being the most widespread surface deposit of Hungary, unconsolidated Pleistocene aeolian loess covers around 30 % of the territory with the most fertile soils of the country, chernozems. Deep loess mantles are equally found in the Great Hungarian Plain, in many hilly and piedmont areas, but loess features (hollow roads, pipes, erosion gullies and shaft-like loess wells) are the most diverse on the slopes of the North-Hungarian Mountains. They are also important from economic point of view as surface dissection impedes agricultural cultivation. On the loess mantle of a southern outlier, Tokaj Hill, numerous morphological features developed within a small area (around 20 km$^2$). The landform assemblages have been studied by generations of geomorphologists, including Pinczés (1968, 2000), Pinczés and Boros (1967), Boros (1977), Kerényi and Bálint (1986), Kerényi and Kocsisné Hodosi (1990) and Kerényi (1994).

Tokaj Hill is part of the Tokaj-Hegyalja wine-growing region (890 km$^2$ area), which has been the most important viticultural region of historical Hungary since the late 16th century. The centuries-old intensive cultivation of vineyards has exerted a decisive impact on erosion processes.

## 25.2 Environmental Background

Tokaj Hill (Fig. 25.1) is located at the southern edge of the Tokaj–Eperjes Mountains, the easternmost member of the inner Carpathian volcanic range in Hungary, which extends well into the territory of Slovakia in the north (Fig. 25.2). The main mass of the mountains was formed by volcanic eruptions in the Miocene (dated to Early Badenian, 10–9 Ma ago). Comprised mainly of dacite (Gyarmati 1974), the 516 m high Tokaj Hill is an almost perfect cone, a composite volcano slightly isolated from the main mountain range and connected to it in the northwest through a low col (128 m elevation) excavated in rhyolitic tuff (Boros 2002). Tokaj

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**Fig. 25.1** View of Tokaj Hill from the wetlands in the northeast (photo by Csaba Tóth)

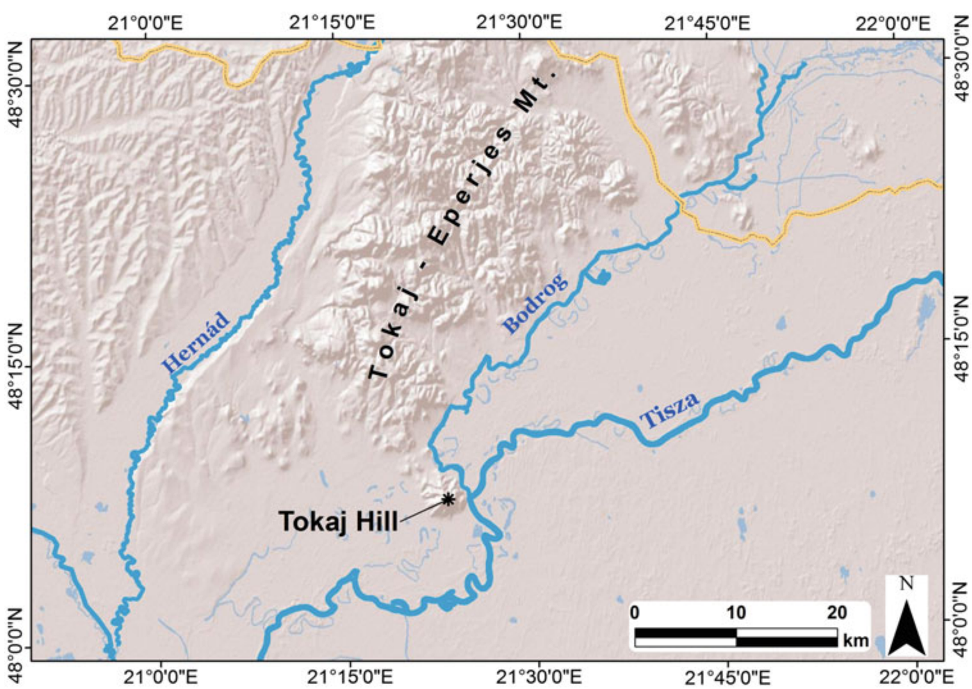


**Fig. 25.2** Location of Tokaj Hill

Hill rises steeply above the surrounding alluvial surfaces shaped by the Bodrog, Tisza and Takta Rivers. The base level of erosion is the Tisza River (94 m above sea level), whose channel runs at the foot of the hill. The southern slopes are steeper than the northern (Fig. 25.3): valley sides of southern aspect show slope angles between 20 and 30°, while general slope ranges from 15 to 17°. Similarly steep are slopes of eastern and western aspect. The average angle of northern slopes is not more than 10–12° and deeper-cut and steep-sided (20–25°) valleys are less common.

For the most part, the hill is covered by loess deposited directly upon volcanic rocks (Pinczés 1954, 2000). It is only absent from the very top of the hill; in downslope direction it first appears in patches below 450 m above sea level, then

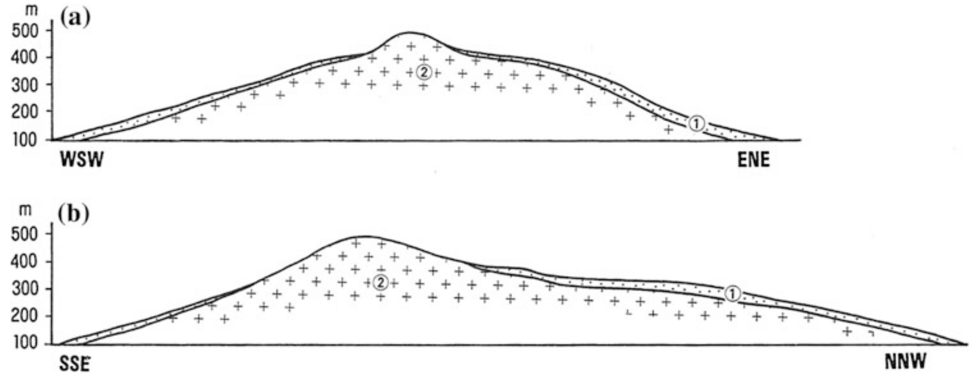


**Fig. 25.3** Loess thickness on Tokaj Hill (section vertically exaggerated). *1* Loess cover; *2* Miocene volcanic rocks

the cover becomes continuous and mantles the volcanic rocks in increasing thickness (Fig. 25.3). The loess mantle is thickest on the eastern slopes today (20–30 m), reaches around 15 m on the northern slopes and is thinnest with 5–10 m on the western and southern slopes. Thus, loess thickness varies with slope aspect, partly as a consequence of the fact that the southern and western slopes are the most suitable for viticulture and the establishments of vineyards contributed to intensified erosion. Loess cover was still uninterrupted 1,500–2,000 years ago; its rapid denudation began with deforestation, which started 1,000 years ago, and further intensified due to the spreading of viticulture. The development of a wide range of morphological features on the loess slopes has also resulted from other human activities associated with viticulture, like land drainage, terraced cultivation and dirt road construction.

## 25.3 Research History

Intensive loess research in Hungary began after World War II, and until the 1980s focused on the origin and chronology of loess (Pinczés 1954; Pécsi 1965, 1966, 1979; Hahn 1966; Pécsi and Szebényi 1971; Szilárd 1983). Other research fields comprised mapping of denudation forms in loess terrains and the mostly qualitative characterisation of these landforms (Sédi 1942; Ádám 1954; Pinczés 1968). Some of the works mentioned already included quantitative analyses as well. For instance, Ádám (1954) performed field measurements and calculations to estimate the intensity of loess denudation. The field measurement series by Pinczés (1980) and Pinczés and Boros (1967) primarily aimed at the determination of the rate at which loess was being removed from Tokaj Hill. Kerényi (1984) elaborated and tested a laboratory method for the investigation of the rate of soil erosion and applied it to study the dissolution processes in loess on Tokaj Hill (Kerényi and Bálint 1986). Some years later Kerényi and Kocsisné-Hodosi (1990) quantitatively studied rates of processes and the resulting erosional features on loess surfaces jointly applying field measurements and laboratory experiments in their study areas, principally on Tokaj Hill.

## 25.4 Major Erosional Features

As a result of erosion accelerated due to human activities, 90 % of the soil cover which existed 2,000 years ago have been eroded, and viticulture today utilizes bare loess surfaces. The highest rate of sheet erosion was measured at 250 m elevation proved by the thinning of the loess cover there. According to the interviews made with local owners, ca 0.5–0.7 m thick loess was eroded in the last 100 years. Landforms generated by cultivation developed along field boundaries and pseudoterraces attest to intensive denudation. The height of pseudoterrace risers depends on slope angle and the duration, frequency and intensity of soil cultivation—therefore, it varies in a wide range, mostly between 1 and 2 m.

Hollow roads or sunken lanes in loess (Fig. 25.4) are also man-induced landforms (Boardman 2013). In almost all cases, tracks on the dirt roads providing access to the vineyards develop into hollow roads over the centuries. Since the mechanical effect (increased pressure) under the wheels of carts intensified erosion by runoff along the roads, the depth of hollow roads has been influenced by both the frequency of traffic on the road and loess thickness. The deepest loess roads (10–15 m) are found on the eastern slopes of the hill, where the loess mantle is also the thickest.

Gullies and ravines are mostly natural landforms of considerable size. They are most common at elevations of 100–250 m above sea level and aligned radially on Tokaj Hill, which has an almost perfect conical shape. During heavy rainfalls large amounts of loess accumulate at the mouths of these valleys even today. According to measurements by Boros, the thickness of redeposited loess

**Fig. 25.4** Hollow road on Tokaj Hill (photo by Tibor Novák)

accumulated over 15 years is 1 m on average at the mouths of the dry valleys which run towards the southern and eastern forelands of Tokaj Hill (Pinczés and Boros 1967).

Smaller forms of concentrated erosion, i.e. erosion rills are widely distributed, especially in cultivated areas. Erosion rills formed as a result of single major downpours in a loess area prepared for planting grape vine can be seen in Fig. 25.5. Rills may develop into gullies within a few years if their deepening is not blocked by soil conservation measures.

Superficial depressions in loess are generated by both man-induced and natural processes. They mostly appear on flat or counterslope segments and particularly on the surface of man-made terraces. The oval depressions are 10–20 m long along their major axis, 3–8 m along their minor axis and 0.3–0.6 m deep. Their origin is due to the fact that the $CaCO_3$ content of a few metres' thick loess layer underground is reduced by the dissolution effect of infiltrating rainwater accumulated on the terrace surface. The dissolution effect is observed over a longer period of time (perhaps a few decades). As a consequence, the stability of the loose deposit is gradually reduced, its original structure is destroyed and subsidence ensues. Although this landform is not visually spectacular, the material loss is significant. According to the measurements of Kerényi and Kocsisné Hodosi (1990), 300 $m^3$ of material was lost from the depressions in a study area of 2 hectares in Tokaj Hill over 25 years. This was the highest value found for any of the studied landforms in the area. It has to be noted though that

**Fig. 25.5** Rill on a loess surface prepared for planting grape vine (photo by Tibor Novák)

the real loss of material is somewhat smaller since the destruction of loess structure involves no loss of material, however, it is rather the result of material removal. The amount of material removed, mostly in dissolved state, along underground pathways, cannot be determined accurately.

Sinkholes due to piping develop where runoff is repeatedly concentrated on poorly designed terraces. Dissolution processes start in loess first as a result of oversaturation at the time of heavy rainfall. Then mechanical erosion begins in the subsurface, pipes are widening and increase in size. The infiltrating water widens the vertical pipe, transforming it into a sinkhole, and then a pipe develops downslope (see also Chap. 13). Pipes are exposed to the surface either on terrace risers or on other steeper slope segments.

Collapsed pipes/tunnels are formed where the loess layer above a pipe becomes so thin that it caves in due to its own weight. This linear form differs from the erosion gully for only a short period of time. The difference is that the erosion gully mostly has a V-shaped cross-section, while the bottom of a collapsed pipe is covered by fallen loess blocks, and, therefore, its walls are vertical or even overhanging at places. Such forms develop on poorly designed terraces, frequently impeding the cultivation of vineyards.

## 25.5 Surface and Subsurface Erosion

Based on field and laboratory studies, interactions have been disclosed between the development of features of loess erosion on the terrace system in the Rákóczi valley of the Tokaj Hill (Kerényi 1994; Kerényi and Kocsisné Hodosi 1990). Terraces were constructed on bare loess surfaces, however, poor soil formation took place on the terrace flats due to regular tillage (applying organic manure every now and then). This was of importance since biological activity of the ploughed layer resulted in higher $CO_2$ concentrations in soil air than in the air of pores in the loess below the ploughed layer and in the atmosphere. Laboratory experiments on loess blocks proved that a thin soil layer on bare loess fivefold increased the calcium-solving capacity of distilled water seeping through the blocks (Kerényi 1994). This can be explained by the transformation of the water filtering through the soil layer into a weak carbonic acid due to the increased $CO_2$ concentration of soil air. The carbonic acid can dissolve more $Ca^{2+}$ ions (Table 25.1). It is worth noting, however, that the water enriched in carbonic acid percolating through the loess block also dissolves other ions, like $Mg^{2+}$, $K^+$ and $Na^+$ ions, in larger amounts than distilled water.

The formation of surface depressions, generated by subsurface processes, takes place in several stages (Fig. 25.6):

1. Electron microscopic images revealed that the microstructure of loess collapsed due to the effect of the acid and pores with 0.1 mm diameter were formed in it. Figure 25.6a presents this first phase of the process.
2. With further dissolution the number of pores of 0.1 mm diameter increases, then with the collapse of the separating walls pores with size visible to the naked eye develop quickly. After this point the process of mechanical erosion is accelerating. In the small sinkhole initially with no exit (Fig. 25.6b) all of the pores can be filled by water in a wet period and then this water begins to percolate in all directions, becomes enriched in $CO_2$, dissolves $CaCO_3$ intensively, and causes significant reduction in the mechanical resistivity of the sinkhole wall. In this way dissolution promotes mechanical denudation.
3. In the third phase of the process, the complete length of the pipe is developed by the mechanical process intensified due to dissolution effect of water enriched in $CO_2$ (Fig. 25.6c). In the vicinity of the pipe the structure of the loess collapses under its own weight and a depression is formed on the surface. The wall of the pipe dries faster following the wet period than the deeper parts of the loess. As a consequence, the difference in capillary potentials drives the water towards the pipe (Fig. 25.6c). Salts transported in the water accumulate on the pipe walls. In the first decade of pipe evolution the subsurface hole widens mostly due to mechanical erosion and chemical transformations in the wall of the pipe, mainly through the accumulation of chlorides and the relative impoverishment in carbonates.
4. In the fourth phase, dissolution and chemical transformations play decisive roles again. In the loess layers above the pipe capillary potential and gravitational effects support each other. Therefore, the flow of water

**Table 25.1** Ion concentrations (mg/L) in water runoff and infiltration into loess solums (T = 20 °C)

| Loess solum | | | | | | | |
|---|---|---|---|---|---|---|---|
| Horizon | Origin | Water | $Ca^{2+}$ | $Mg^{2+}$ | $Na^+$ | $K^+$ | $Cl^-$ |
| $A_p$ | From cultivated layer | From loess surface | 14.8 | 3.8 | 7.3 | 6.0 | 18.3 |
| | | From infiltration | 62.6 | 5.5 | 5.1 | 18.4 | 21.0 |
| C | From terrace riser | From loess surface | 14.5 | 3.7 | 0.31 | 0.001 | 10.6 |
| | | From infiltration | 66.5 | 28.9 | 3.5 | 0.001 | 29.9 |

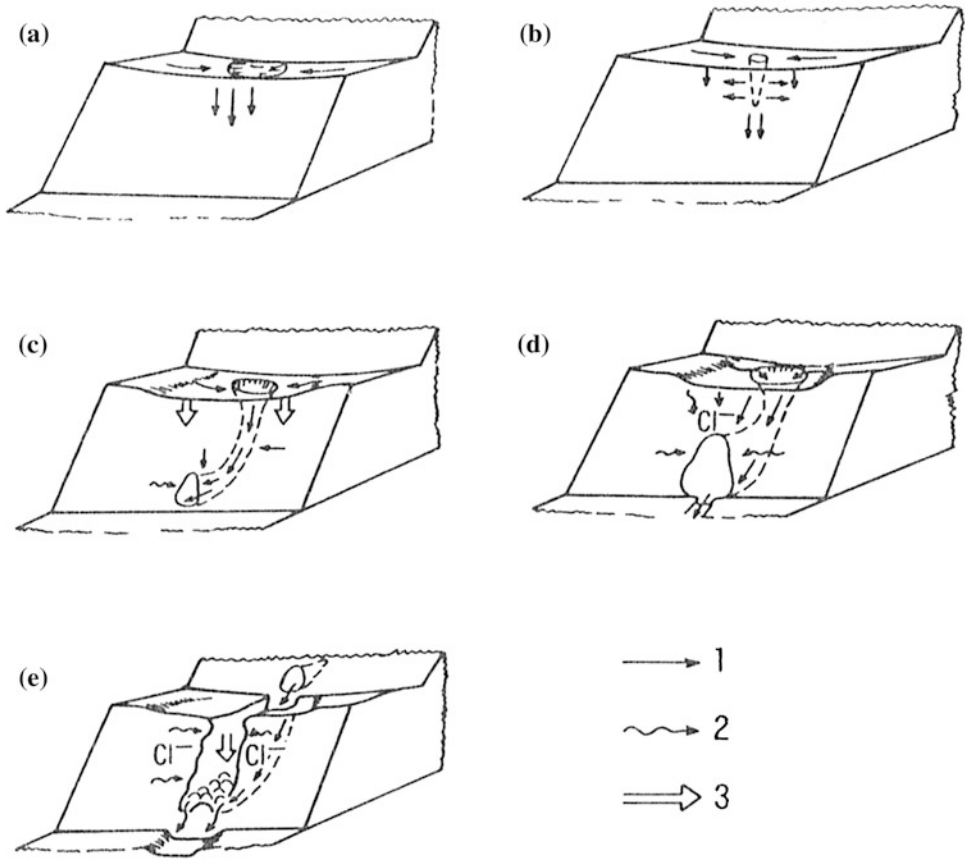


**Fig. 25.6** Stages of loess erosion on a terrace (for explanation see text). *1* Infiltration of rainwater; *2* lateral percolation to pipe; *3* loess collapse

towards the hole is the most intensive here (Fig. 25.6d). The loess layer above the pipe will include numerous widened larger pores and its strength decreases. This influence is complemented by the abundance of chlorides and the reduction of carbonates in the top loess layer, further reducing its stability.

5. As a result the top of the pipe will cave in due to its own weight. At this time the piping process stops at the given place and a new phase of loess denudation starts with the formation of erosion gullies (Chap. 12). Their walls are exposed directly to the wetting effect of rain, to wind and direct solar radiation (Fig. 25.6e).

## 25.6 Conclusions

A large variety of erosional forms can be found in a relatively small area of the loess mantle of Tokaj Hill. Increasing rate of loess and soil denudation started with deforestation and associated viticulture 1,000–1,500 years ago and is active even today. Specific to the development of erosional features on loess is the joint influence of subsurface dissolution processes and mechanical denudation by water erosion. Loess denudation also occurs in areas where soil conservation measures have not been implemented carefully enough. For instance, counterslope terraces were established where rainwater is collected and over-saturation induces pipe flow. In the final stage pipes cave in and erosion gullies form. The destructive process can be prevented by careful adjustment of technical soil conservation measures, observing the specific properties of loess, with special concern to the precipitation conditions of the given area and to the balanced water budget of terraces and slopes.

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# Megyer Hill: Old Millstone Quarry

# 26

János Szepesi and Zsuzsanna Ésik

**Abstract**

In the Sárospatak region of the Tokaj Mountains, Northeast-Hungary, thick Badenian ash flow deposits accumulated in submarine settings and later uplifted and suffered intensive hydrothermal alterations. The semi-circular range of Megyer and Király Hills was formed by differential weathering and selective erosion of rocks in the intensely altered and mineralized zones. The exploitation of raw materials has a long history in the region. The secondarily hardened rocks provided building materials for the surrounding settlements; the pottery was supported by clay minerals. The intensely silicified pyroclastic rocks were used as millstones as early as the 15th century. The legendary French-style millstones won the First-Order Medal of the 1862 World Expo in London The picturesque lake in the abandoned quarry was selected as Hungary's most beautiful natural attraction in 2011.

**Keywords**

Ash-flow deposit • Submarine volcanism • Silicification • Millstone quarrying • Cultural values • Tokaj-Hegyalja

## 26.1 Introduction

Tokaj-Hegyalja, the foothill of the Tokaj Mountains, is built up of ignimbrite sheets around stratovolcanic and acid extrusive centres (Fig. 26.1). The south-facing slopes on rhyolite tuffs coupled with soils and microclimate defined a special viticulture. The world's first closed wine region (since 1737) was declared a UNESCO World Heritage site, the Tokaj Wine Region Historical Cultural Landscape, in 2002. In addition to rows of cellars, the quarries (e.g. Mád, Bodrogkeresztúr, Sárospatak—Fig. 26.1) producing millstones, building and decorative stones also determined the landscape character. With its millstones of unique quality Sárospatak acquired a European reputation.

## 26.2 Volcanism in the Tokaj Mountains

Volcanism in the Pannonian (Carpathian) Region was active since the early Miocene through various phases in variable geotectonic and magmatic settings (see Chap. 2). The heterogeneity of the mantle source and crustal differentiation is manifested uniquely in the same amounts of rhyolitic and andesitic rocks deriving from the Badenian–Sarmatian–Pannonian period (15–9.4 Ma—Pécskay et al. 1995). The rare olivine basalt was emplaced as a final effusion. The Proterozoic to Mesozoic metamorphic and carbonate basement (Fig. 26.2) was subsided to form a north-south oriented graben-like structure hosting the volcanic sequences of the Tokaj Mountains (Gyarmati 1977). Extension processes were accompanied by basement subsidence and marine transgression, so the thick Badenian acidic (e.g. Megyer Hill

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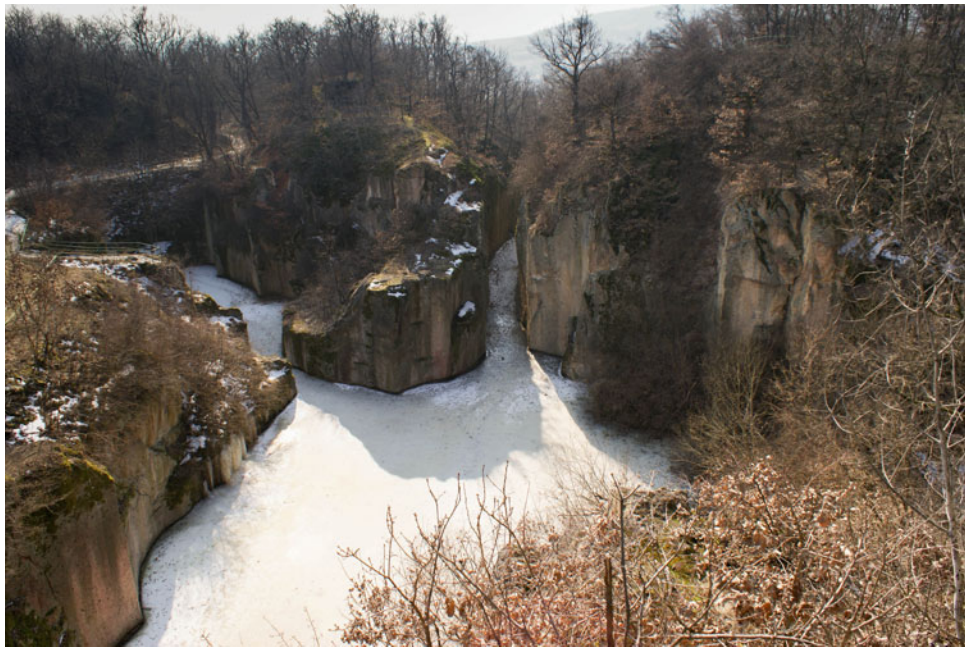

**Fig. 26.1** Panorama of the lake of Megyer Hill (photo by János Szepesi)

ash flow tuffs) and intermediate formation accumulated in submarine environment but the archipelagic nature became prevalent with the thickened volcanic sequence in the Sarmatian (Fig. 26.3).

Based on radiometric ages, the andesitic-dacitic stratovolcanic centres operated synchronously with pyroclastic flows and Plinian eruptions. With the reduction of explosion energy the formation of thick tuff horizons was followed by extrusion of coherent lavas (perlites, rhyolites), uniquely diverse in the Carpathian Volcanic Chain.

Postvolcanic activity reached its peak in the Sarmatian-Pannonian and ended in the Pleistocene. The thermal circulation of exogenic water along fractures was often promoted by the higher porosity of ash flows rich in pumice. Erosion revealed the mineralized zones (Pécskay and Molnár 2002). The deepest, K-metasomatic deposits are well known from the gold-silver bearing quartz veins (Telkibánya, Rudabányácska). The Király-Megyer Hill range is one of the best examples of the surrounding alunitic-kaolinitic zone.

The volcanism and hydrothermal activity generated raw materials, mainly non-metallic mineral resources, 13 special raw materials (including quartzite, kaolinite, bentonite, perlite) at 47 known occurrences (Mátyás 2005).

## 26.3 Geology of the Király-Megyer Hill Range

The Király-Megyer Hill range of 9 km$^2$ area along the northeastern boundary of the Tokaj Mountains, north of Sárospatak form a semi-circular range around a local basin opening to the south. The effective postvolcanic alterations result from the regional morphological-tectonic pattern. The tectonic lines of the basement were renewed by younger extensional processes. The main structural lines are north to south directed.

Sárospatak and its environs was one of the most frequent areas of geological research in the last 200 years. The city was known for millstone manufacturing, ceramics and glass production in the middle ages but geological knowledge only began to accumulate in the 19th century. After sporadic observations the first comprehensive works were published by Szabó (1867), who classified acid pyroclasts by their utilization (millstone, powder tuffs) and recognized the Badenian age of the silicified tuff of Megyer Hill based on mollusc fauna. Detailed raw material exploration accelerated after World War II. The geological, tectonical and geothermal conditions and the non-metallic resources (bentonite, kaolin) were characterized by Frits (1959, 1964). The cinnabar, as a characteristic mineral of zonal hydrothermal alteration was identified by Kulcsár (1968) based on the analogy of the Beregovo Hills in Ukrainian Transcarpathia. The comparative investigation of postvolcanic mineral paragenesis with the genetic relationships of the hydrothermal zonation was focused on the alunite resources of Király Hill (Mátyás 1969, 1977). The issue was re-investigated and clarified by new methodologies (K–Ar, fluid inclusion, isotope studies—Molnár 1993; Pécskay and Molnár 2002).

The Király-Megyer-Botkő group is one of the many hydrothermal centres aligned to the regional tectonic patterns. The Mesozoic basement has influenced recent and paleogeothermal activities. Although the presence of limestone xenoliths in the ash flow tuff had indicated a Triassic

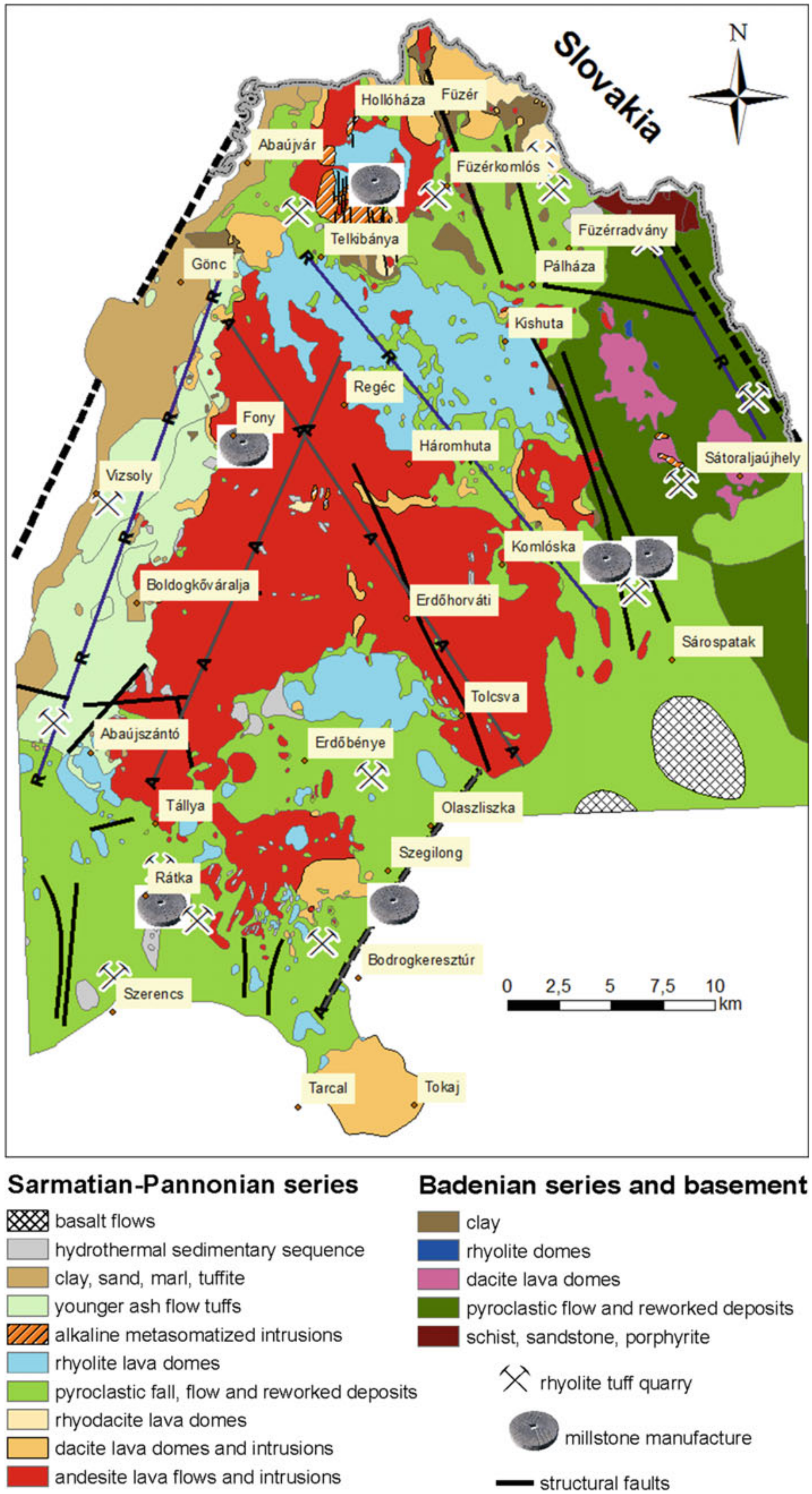


**Fig. 26.2** Geological map of the Tokaj-mountains with the main rhyolite tuff and millstone quarries (modified after a volcanological sketch by Gyarmati 1977)

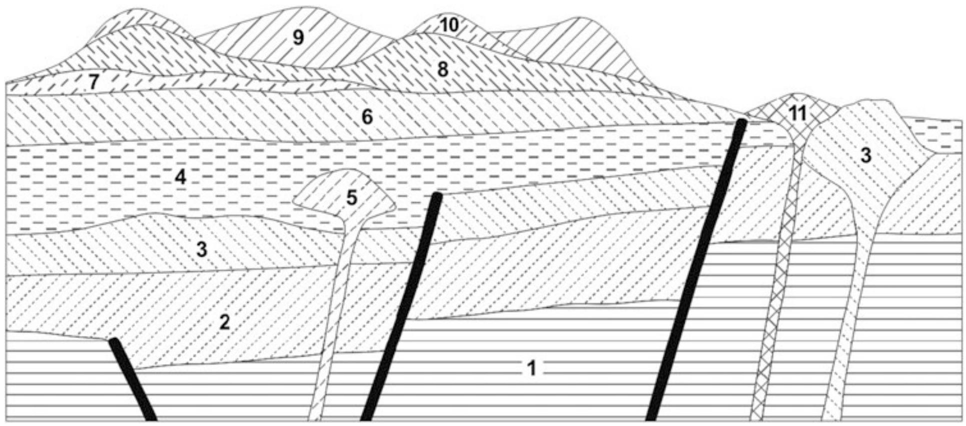


**Fig. 26.3** Schematic spatial and temporal relationships of geological formations in the Tokaj Mountains (Gyarmati and Szepesi 2007). *1* Prevolcanic basement; *2* upper Badenian pyroclasts with intercalated marine sediments (15–14 Ma); *3* upper Badenian submarine and subvolcanic intermedier volcanites (15–13 Ma); *4* lower Sarmatian pyroclasts with brackish sedimentation (13–11 millió év); *5* lower Sarmatian rhyolite lavadomes and flows; *6* Sarmatian andesitic and dacitic volcanites; *7* intercalated pyroclasts; 8, Upper Sarmatian intermediate lava flows (11–10 Ma); 9, Youngest (Pannonian) rhyolite and rhyodacite (10 Ma); 10, Youngest intermediate lava flows (10–9 Ma); 11, Olivine basalt (9.4 Ma)

**Fig. 26.4** The pumice breccia character of the ash flow tuff in the wall of the wagon road (photo by János Szepesi)

basement before deep drillings began, it was still surprising that a borehole reached the Mesozoic at only 225 m depth (ca 90 m above sea level) and the karst cavities supplied 40 °C thermal water with 2 $m^3$ $min^{-1}$ discharge.

The onset of the volcanism was accompanied by a general transgression and the first rhyodacitic submarine ash flow series are intercalated with older basement debris. On the eroded surface of Mesozoic limestone tuffaceous conglomerate formed. The welded parts and smaller eroded rhyolite lava dome necks emerged as small islands. Under similar paleoenvironmental conditions volcanism turned to rhyolitic character to produce the predominant millstone-rhyolite tuff unit deposited in a submarine caldera but subordinately bedded ash fall tuff from Plinian eruptions (at Király Hill) is also present. The abundance of angular and variable rounded pumice clasts defines a lithic (perlitic lapilli) pumice breccia facies (Fig. 26.4). The absence of the wave-generated bedforms are consistent with a below-wave-base environment, also supported by molluscs (Chlamys, Cardium, Isocardia—Szabó 1867; Kulcsár 1968). Large quartz phenocrystals, less common plagioclase and rare biotite are typical for Badenian pyroclast series.

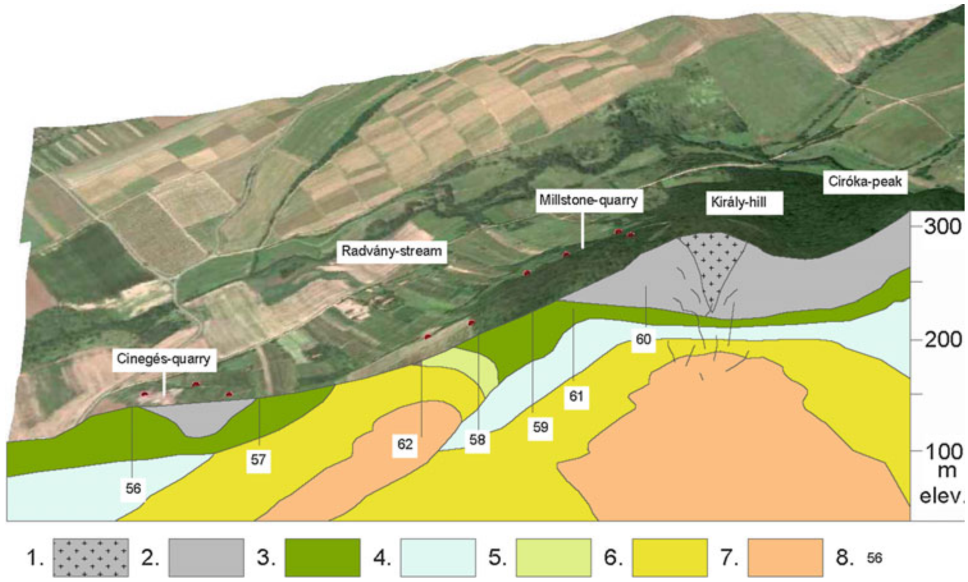


**Fig. 26.5** Geological cross-section Cinegés quarry–Király Hill tectonic line (Fig. 26.2) with the main alteration zones (modified after Molnár 1993). *1* Hydrothermal breccia; *2* quartz-opal-barite-cinnabar-hematite; *3* quartz-opal-alunite-kaolinite; *4* quartz-opal-kaolinite; *5* montmorillonite-illite-kaolinite; *6* illite-kaolinite-montmorillonite; *7* adularia-illite-hematite; *8* Sárospatak boreholes. The millstone quarry was deepened in alteration zone 2

Its thickness suggests that the sequence was generated by magmatic volatile-driven explosions. The subaerial pyroclastic flows crossed the shoreline, and transformed into eruption-fed subaqueous volcaniclastic density currents. The curved pumice clasts probably became rounded during the transportation.

At the end of Badenian and Early Sarmatian a further marine transgression occurred and rhyolite tuff and tuffite accumulation and reworking as well as shallow-water clay and sand sedimentation took place. The Lower Sarmatian series deposited in lagoons. Thinning layers are found at the foot of Király-Megyer Hill group (Cinegés quarry, Nagybotkő) and in boreholes. The accumulation was fed by the material of Plinian clouds, reworking of former Badenian series coupled with hydrothermal activity. The diverse series consisted of alternating tuffaceous clays, pumice breccia and limnic quartzite layers (I Perlaki 1989).

The strongest hydrothermal activity occurred along the Király and Megyer Hills, but the characters of alteration zones differ. Király Hill has the widest mineral paragenesis (Fig. 26.5). The first stage of the postvolcanic alteration probably was simultaneous with acid volcanism. Areal silicification affected almost the whole region and represents the sites of highest temperatures and strongest fluid exchange and leaching processes. The most silicified rocks formed along faults and fractures acted as conduits for hyrothermal fluids. The large proportion of the pumices dissolved with changing of hardness. Strongly silicified rock bodies cap Király Hill, Megyer Hill, Cinegés and Botkő Hill (Fig. 26.5). The concentration of silica in these rocks is above 95 % and the original tuff texture is totally destroyed, only quartz phenocrysts are observable (Gyarmati and Pentelényi 1973; Mátyás 1977) Cinnabar is present as dust-like encrustations in the cavities or dendritic pattern in the silicified rock (Molnár 1993). Sanidine phenocrysts and the groundmass of the tuff are altered to illite and kaolinite (Kulcsár and Barta 1969; Molnár 1993).

With reducing intensity of alteration the siliceous horizons were underlain by alunite and kaolinite alteration zones (Molnár 1993), which were moderately resistant to erosion and are exposed in lower positions (Fig. 26.5). The argillized rock is locally stained red, purple, brown or yellow depending on iron oxide content. Aggregates of fine-grained (0.1–0.5 mm) rhombohedral alunite crystals occur in the cavities formed by leaching of the pumice fragments of the tuff. Parallel with this the dominant $SiO_2$ mineral changes from quartz to opal and cristobalite. The illite and montmorillonite dominated and a potassic feldspar (adularia) bearing alteration zones formed beneath the kaolinite-alunite horizon at Király Hill (Molnár 1993) providing a typical low temperature hydrothermal alteration pattern.

## 26.4 Mining History of Sárospatak

The exploitation of rock varieties (rhyolite tuffs and rhyolite, perlite, obsidian lavas) produced by acid volcanism and hydrothermal activity has thousands of years of history. At the different level of social and technical development more and more raw materials were placed in the centre of interest from the Palaeolithic obsidians. Rhyolite tuffs show the widest distribution at Tokaj-Hegyalja and have been utilized as a natural building stone for several centuries as demonstrated by large numbers of abandoned quarries (Fig. 26.2). Data on ancient quarries were registered in the early domestic geological mining inventory (Schafarzik 1904) and also in recent databases (Atlas of European Millstone quarries, Historic Quarries, Hungarian Mineral Occurences).

The silicified zones of Megyer Hill were proper for quality millstones. After the first mentioning from the 15th century quartzite was a popular and precious product over six centuries. The industry was supported by the grindstones demand of precious metal mining at Telkibánya. The quality and spatial awareness of the stones had earned a reputation for Sárospatak.

The regional industrial activity (building stones and millstone) stimulated the development of clay mineral quarrying and ceramic industry, which had a golden age in the 1800s. Pottery, tile stove and pipe factories ("famous black pipe") were also operated (Mátyás 2005). The large variety of dish forms (bowls, plates, jars, food containers, jugs) was widespread in the villages of Tokaj-Hegyalja, Hegyköz and Bodrogköz. The kaolin resources (Megyer Hill, Végardó) excavated from five quarries in the late 19th century. The most valuable portion of the deposit was the snow-white dense upper parts but bentonitic ("greasy") kaolin was also mined. Unfortunately the production intensively decreased after the world war. The clay minerals and alunite stocks were re-investigated in the 1950–1980 years (Frits 1959, 1964; Mátyás 1977; Perlaki 1989). The Végardó clay deposit was excavated by underground working through a 76-m-deep shaft only between 1957–1959 (Izsó 2011). The kaolin resources of Botkő exploited from 1972, the alunite stocks from 1977. As a result of the raw material inhomogeneity and other risk factors the mining activities had to stop.

## 26.5 The Old Quarry of Megyer Hill

The silicified zones are easily recognized as resistant outcrops over the entire area of Tokaj Hegyalja (Szerencs, Mád, Sárospatak). Megyer Hill forms the eastern side of the arcuate hydrothermally altered range. The geographical position of quartzite deposits has controlled the shape and alignment of present valleys. The downcutting Suta Stream was forced northward by the greater resistance of low quartzite hills (Cinegés and Botkő). The silicified zones are surrounded by argillized rocks. The minor kaolinite reserve of Megyer Hill was mined for the Zsolnay porcelain factory of Pécs between 1887 and 1940 (Boczán et al. 1966).

The old millstone mine was deepened in the silicified cap where the permeable lapilli tuff and tuff breccia wallrocks promoted the hardening process. The quarry operated from 1,400 years and a three-level mine of 150 × 50 m was deepened during the 500 years of mining. The lowermost level is the "Óbánya" with the canyon-like narrow wagon road and the picturesque lake (Fig. 26.1). Cold air is accumulating at the bottom of the old quarry pit and a thick ice sheet covers the water in winter and early spring (Fig. 26.1). The vertical quarry walls are up to 70 m high above the lake. (That is why the lake is popularly called a "tarn".) The second level is situated 5–6 m above the water level with the mining buildings but the substandard millstones left behind are also characteristic. The third, topmost and widest, level where a two-roomed cave with windows, doors and stove was also carved in stone.

The excavation was carried out with the similar manual technique and toolkit for centuries, as attested by the curvy walls. The main users were grain (wind, water, dry and hand mills) and ore milling industry. Initially whole stones were mined, ca 300–450 pieces a year. First, circles were drawn around on the rock. The grinding surface of stone pairs was made from opposite rocks for the best fitting in the mill. The wheat grains slightly roasted during the milling which gave a pleasant flavour to the flour. The deepening of the quarry yard called for an easier way for transportation and a canyon-like wagon path was cut into the rock (Fig. 26.6). Unfortunately, productivity decreased at the end of 19th century and the Megyer Hill quarry ceased to operate in 1907.

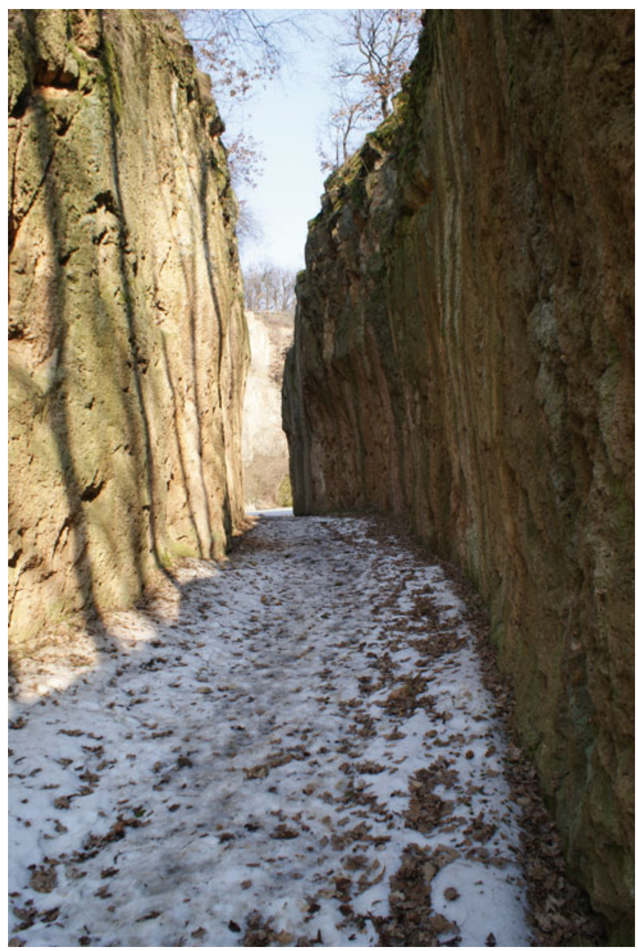

**Fig. 26.6** The vertical walls of the wagon road (photo by János Szepesi)

A picturesque lake was formed by rainwater accumulation in the quarry yard. The constant volume of water is ca 4,000 $m^3$ with a maximum depth about 6 m higher than expected from the annual rainfall. The absent of evaporation lines also prove that the hydrological cycle is not only climate related and must be an additional inflow along the fractures. The hydrothermal alteration pattern also plays an important role. The argillation zone with clay minerals (kaolinite, illite, monmorrillonite) forms an impermeable layer under the fractured silicified cap.

Beside the mining heritage the area has specific botanical and zoological values. The main plant community is acidophilus oak forest (*Genisto tinctoriae–Quercetum*), but the mining activity caused continuous degradation. Today, the scrubby-woody vegetation has been reclaiming the quarry walls and need control for the better visibility. The maple-oak woods (*Aceri tatarico–Quercetum)*) formerly covered the loess deposited foothill slopes but receded with the spreading of viticulture. Accumulating water promoted the establishment of waterside and aquatic plants. Half of the lake's surface is covered by small duckweed (*Lemna minor*) and submerged cross duckweed (*Lemna trisulca*) from spring to fall (Mercsák et al. 2007). Grapes are cultivated on the southern slopes of Megyer Hill for hundreds of years. The plantations are now mostly aged and often abandoned.

The fauna is rich in species despite of small size of the protected area owing to the forests and the favourable exposure of the quarry walls. The lake is important breeding and feeding sites of the amphibian species (frogs), but large numbers of singing birds waterfowl and reptiles (lizards, glides) are also specific.

The amazing 1.1 ha quarry yard with area of Megyer Hill was declared a nature reserve in 1977, which became a nature conservation area of national interest in 1997. It would be appropriate to extend the boundary of the protected area to the forests and Király Hill (in south-western direction) in order to maintaining the fauna richness of the natural reserve (Mercsák et al. 2007).

## 26.6 World Champion Millstones

The new chapter of Sárospatak millstone industry has already started in the middle of 19th century. The preparation of the so-called "French-style" millstones has begun in 1859 in the Botkő mine and 1864 in the Király-hill quartzite mine.

French-style millstone manufacturing was based on an increasing number of more efficient power mills at the expense of water, wind and dry mills in the mid-19th century. The faster rotation of these mills needed long-lasting, wear-resisting stones from harder rock. The French millstone manufacturing centres (La Ferté sous Jouarre, Margay, Epernon, Vernot, see Atlas of European Millstone quarries) dominated the world market for a long time. French millstone manufacturing in Hungary was born in the Tokaj Mountains. The millstone manufactory was founded at Fony in 1858 and first grinding stone was assembled in Budapest (Haggenmacker power mill). Similar plants started operating in the first half of the 1860s at Rátka and Szegilong, but the largest size and reputation was reached by the plant founded by K. Láczay Szabó at Sárospatak (Fig. 26.2). The production of French millstones from the quartzite of Botkő Hill started in 1859. Initially, French factory workers were also employed, and later worked with locals. The millstones won "first-order medal" at the 1862 World Expo in London and other shows (Prague, Vienna, Szeged, Székesfehérvár) also featured a great success: "the stones have achieved their goals and the French flint stones became unnecessary with breaking their price and sixty people have been employed. 42 millstones have been sold only at region of Trieste, the export reached the border of France, even longer grind in Switzerland" (Láczay-Szabó 1864). Additional high-grade material discovered at the Király Hill in 1864, which retained the sharpness through 8–10 days and foremost was suitable for wet grinding.

The production of the French-style millstone required complex activity. The mining was made primarily by hand tools until the 1950 years. The millstones were always done in pairs; one lower stone and a rotating upper block were cut to ensure uniform grinding and abrasion. The selected stones have the same hardness and porosity. The porosity defined two different rock types. The denser variety was suitable for lower stone and the more porous for rotating block. A millstone consisted of two parts: from the inner "heart stone" and the glued, outer "bricks" (12–16 pieces), which defined the French style. The bricks were suited around the "heart" in cement stuck and were left to harden at least 2–3 days. Finally, an iron frame placing around and angled slots were carved out to promote cooling and outward drifting of the grist.

The plant operated successfully in the first decades of the 20th century, although only 200–300 pieces were produced annually contrary to the past 400–500 millstones. The expansion of steel rolling mills heavily influenced manufacturing, but the stones were ordered and delivered even in 1944. The plant was nationalized in the early 1950. The production was about 150 millstones in the 1950–1960 years and 6–10 pieces in the 1970s. Major customers were the paprika mills and the Herend and Pécs porcelain factories. The last millstone was taken in 1979, when excavation in the

**Fig. 26.7** Millstone left behind in the quarry yard (photo by János Szepesi)

Király Hill quarry terminated. Some examples are still seen scattered in the yard of the quarry (Fig. 26.7). Afterwards only stone lining of drum mills was made from the Botkő quartzite (Hála 1993).

## 26.7 Conclusions

The landscape around Megyer Hill represents the typical natural and cultural heritage of Tokaj-Hegyalja. The lower slopes are covered by vineyards. The remnants of old volcanism and geysers were revealed by quarrying activity since the 15th century. The amazing quarry yard of Megyer Hill was declared nature conservation area of national interest in 1997. The main purpose of conservation management is the preservation of geological values and mining history. Specific botanical and zoological values formed by the accumulation of rainwater since the termination of mining activities. About 150 years later of the world champion winning millstones the natural and cultural heritage received another first prize: as Hungary's most beautiful natural attraction in 2011 (www.origo.hu), which underlines the great public interest in the preservation of this natural monument.

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# Hortobágy Puszta: Microtopography of Alkali Flats

27

Csaba Tóth, Tibor Novák, and János Rakonczai

**Abstract**

On soils with Natric or Vertic properties in the subsoil and easily erodible (e.g. Mollic) topsoil waters running downslope generate diverse erosional microtopography. This process is particularly well observed on the alkali (Solonetz and Solonchak) flats of the Great Hungarian Plain. To the effect of erosion beginning along the cracks of the alkali ground a *szikpadka* (salt berm) of some tens of centimetres height and various gradients forms between the berm top with intact soil profile and the so-called *vakszik* ("blind pan") with eroded topsoil. The extension of salt-affected soils and the accompanying berm erosion was substantially increased by groundwater table changes in the wake of the 19th-century river regulation and land drainage works. At the same time, previous research has proved that this type of erosion of extremely slow rate was an active geomorphic agent on alkali soils in the Pleistocene and in the warm semiarid stages of the Holocene.

**Keywords**

Salt-affected (alkali) soils • Vertic Solonetz • Alkali flat (salt pan) • Berm erosion • Alkali microforms • Hortobágy • Great Hungarian Plain

## 27.1 Introduction

In Hungary an estimated almost one million hectares of salt-affected soils are found; i.e. ca 10–11 % of the territory is covered by alkali soils. The Solonetz and Solonchak types of alkali soils with unfavorable properties are primarily located in the Great Hungarian Plain (Fig. 27.1), but they also appear in limited patches in the regions of West Hungary where groundwater levels are high. One of the most extensive contiguous alkali flat is the Hortobágy, the majority of which had been the floodplain of the Tisza River until the river regulations which began in the mid-19th century. In 1973 the first National Park of Hungary was established in the Hortobágy area to protect botanical, zoological and cultural history values. In addition, the region is also rich in pedological and geomorphological features. The salt-affected soils excluded from cultivation remained in their natural state and show special and characteristic erosion processes. The flat surface of the Great Hungarian Plain is deficient in macroforms, but abounds in microforms. The so-called *szikpadka* (salt berm) areas (ledges of saline depressions) with some tens of centimetres or up to one meter relief clearly appear even in vertical photography, but the best way to study the alkali pastures is by foot (Fig. 27.2). The erosional features and their respective soils and plant associations usually result in an exceptionally varied, mosaical landscape pattern and biodiversity at a microscale. For its preservation it is of crucial importance to

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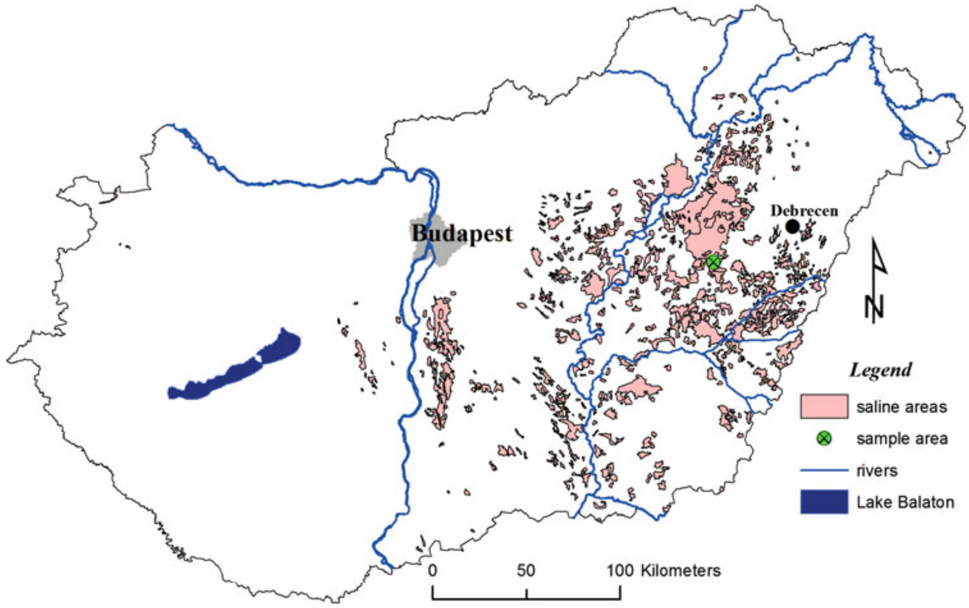


**Fig. 27.1** Saline areas in the Great Hungarian Plain (edited by Csaba Tóth)

**Fig. 27.2** Diverse microtopography on alkali soils in the Hortobágy: berm top covered by alkaline grassland, the bare berm edge and the berm slope (photo by Tibor Novák)

maintain the cultivation mode most fitting to regional characteristics, namely, the extensive stock breeding, which is typical of the Hortobágy, the biggest continuous natural grassland ("puszta") of Europe. In 1999 the Hortobágy National Park was included into the UNESCO Cultural World Heritage list.

## 27.2 Environmental Background

The development of the alkali (in Hungarian: *szik*) microforms is due to a combination of environmental factors. The most important are a heavy clay subsoil, resistant to erosion, and a more easily erodible topsoil. Such a profile is typical of Natric (sometimes simultaneously Vertic) subsoil and Albic or Mollic topsoil; therefore the microforms of berm erosion mainly appear on Solonetz soils. If texture differentiation between the Solonetz subsoils and topsoils is significant, it usually promotes erosion. On the other hand, according to some observations, the berms are not exclusive features of Solonetz soils, but in the case of Vertic type Chernozem and Solonchak erosion forms of similar appearance and origin also occur. These areas of Vertic subsoil in the Alföld region are primarily found in the transitional zones between the highest ridges (typically covered by Chernozem-type soils) and the Vertisols in depressions. Consequently, szik berms are primarily found not in depressions of but on the edges of elevated ground. A typical characteristic of the soils with Vertic-type subsoils is high swelling clay content (illite and montmorillonite), swelling when exposed to water, and shrinking and cracking when dry (A Nagy and Korpás 1956; Stefanovits 1981).

An essential factor in berm formation is the removal of topsoil that starts along the subsoil cracks. In the dry season the topsoil with low clay content begins to crack too. Mechanical impacts (e.g. trampling, wind erosion) result in the penetration of topsoil deep into the subsoil along further widening cracks. In the oncoming wet period rainwater and/or excess water (groundwater flooding) on the surface would quickly reach the subsoil through the cracks, where—due to the swelling of the clay minerals after water uptake—soil gaps close. The Natric or Vertic ground level is almost perfectly impervious and hinders further infiltration into the soil. On the other hand, no swelling occurs in the topsoil, and the surface cracks that were formed in the dry period become filled with water retained by the subsoil. This explains the extensive inundations in the wet season and/or when the subsoil is saturated. In the periods of flooding the cracking network of the topsoil becomes wider and deeper by water flowing towards the lowest point, ultimately completely removing the previous topsoil over ever larger areas. As a consequence, the Natric or Vertic subsoils—which, on the other hand, are more resistant to erosion—would be exposed.

The erosional microforms have developed most spectacularly at the edges of natural levees, where the surfaces of alkali soils tend to have a definite slope. In the Hortobágy berm formation usually takes place at 86.5–87 m elevation, because the Vertic Solonetz soils are heavily cracked during the dry period and prone to erosion. However, the crucial factor in the relationship between the terrain features and szik formation is never elevation but rather relative relief, i.e. height above the local base level.

Another important factor is the amount, intensity and uneven temporal distribution of rainfall. Especially the intense summer downpours (25–20 mm $h^{-1}$) are able to induce erosion, especially if the soil is already saturated with water. On the other hand, calmer but steadier, long-lasting rainfalls may also lead to some erosion soaking the soil over a longer period, and because of the solution capacity of rainwater.

Therefore, because of the erosive process the cracks—initially only a few centimetre wide and deep—may grow into several metre wide alkali rills and barren patches, the depth of which is determined by the vertical position of the Natric (or Vertic) type B-horizon, and/or the thickness of the more easily erodible A-horizon of the Solonetz soils, respectively. The exceptionally hard accumulational B-horizon with high clay content resists erosion. As a result, the berm top with intact soil profile and covered by closed alkali puszta lawn usually transforms into the lowest berm foreground of two sections: an usually 5–30 cm high erosional microcliff, berm edge, with various slope angles, and a lower-lying area with thinner (or even missing) A-horizon, berm slope (Tóth and Novák 1999) (Fig. 27.2). The latter area loses its upper 5–30 cm layer through erosion, and thus the B-horizon with unfavourable conditions and high salt content is exposed on the surface. Hydrolytic silica precipitations are often observed. With time the berm edge continuously retreats and the barren, eroded "*vakszik*" gradually extends in area.

The berm tops are covered by a closed saline pasture (usually *Artemisio Festucetum pseudovinae* and *Achilleo Festucetum pseudovinae*), while the vegetation of berm edges depends on their angle of slope (Tóth and Novák 1999). Gently sloping berm edges (less than 45°) are covered by the same vegetation as berm tops, but steeper (75° to 80°) edges are mostly bare, with only a few *Artemisia santonicum* and *Puccinellia limosa* growing on them. Species composition and coverage in berm foregrounds are completely different. The simple and undemanding, salt-tolerant *Camphorosma annua*, *Puccinellia limosa,* and perhaps *Matricaria chamomilla* are found here with low coverage value (1–5 %) (Tóth and Rajkai 1994). With concentrated, intensive animal trampling, alkali pastures have poorer and scarcer vegetation, which is unable to protect the surface against erosion. Thus, berm tops are eroded at higher rates.

The salt berm microforms can develop not only under natural conditions but also in the areas which are heavily exposed to human impact. Because of the abrupt increase of relative relief, along the edges of artificial depressions (rainwater drainage ditches, canals, borrow holes, etc.)

erosion process may accelerate by several orders of magnitude. But intensive erosion can also be expected on bare surfaces around the sheepfolds—due to the frequent trampling—as well as on the areas cut by wheel tracks of motor vehicles in wet periods.

## 27.3 History of Research on Alkali Microforms

In the early 20th century berm formation was explained by co-sedimentation with surface shrinkage, i.e. washing away of colloidal humus, colloidal clay and salt dissolution by rainwater (Treitz 1924; Strömpl 1931). Microforms like *szik* top, berm, slope and flat were identified. Pál Magyar emphasized both the mechanical impact of rainwater flowing in the cracks—in addition to its solving power—and the erosive effect of animal trampling (Magyar 1928).

Later berm formation was unequivocally called flatland soil erosion. In the opinion of soil scientists surfaces with no or scarce vegetation cover are attacked by surface waters, and so the process of erosion begins. The resulting mosaical structures are called complex and microcomplex soils, respectively, where berm tops and berm bottoms form complexes. In such areas several different soil types and subtypes may coexist simultaneously within 1 $m^2$ area (Arany 1956). In the pedological literature the formation of berm form groups is held as an attribute of structural alkali soils (Solonetzes), because their highly compact, thick and columnar B-horizon would enable the forms to "align on the straight". Their formation is prepared by rain-wash (A Nagy and Korpás 1956). Berm erosion along soil cracks was first mentioned in the literature in the 1960s. In winter, surface waters consume the cracks and cut ever widening channels into them. Therefore, the relatively fertile berm tops diminish, and parallel to that, areas of the barren bottoms increase (Fekete et al. 1967).

In the 1970s a new theory was born, according to which the alkali microforms were created by aeolian action (Keresztesi 1971). The geomorphic role of winds was already mentioned by Treitz (1924) and Strömpl (1931); however, they did not mean direct deflation, but rather they considered the possibility of indirect abrasion by waves of water flow on alkali flats, set in motion by the wind.

Infusion loess as a parent rock plays a significant role, as its high and finely distributed calcium carbonate content dissolves to $Ca(OH)_2$ when exposed to water. The basic pH caused by calcium hydroxide is favourable for the disintegration of silica in the soil, and as a result, clay minerals (illite and illite–montmorillonite mixed layers), amorphous silica and salts are produced. Through leaching the majority of colloidal clay and humus are washed to the deeper parts of the soil profile, and the uppermost, 5–15 cm deep, pale grey layer can be easily disintegrated by rain water (Székyné and Szepesi 1959). In addition to natural erosive agents, anthropogenic effects (animal herds' trampling, traffic) have been emphasized by more and more researchers (Stefanovits 1981; Tóth and Novák 1999; Tóth 2001). A number of researchers studied the connection between soils, alkali microforms and their vegetation. The striped distribution of soil types from the levee top to the base level and their respective special plant associations were described in detail (Tóth and Rajkai 1994; Tóth and Novák 1999).

Besides the agents inducing microform evolution, another important field of research was to determine the rate of alkali soil formation. On the Danube–Tisza Interfluve (at Miklapuszta) the destruction of berm tops on Solonchak-type soils were studied through comparing maps and vertical photographs taken at different times, and by high precision GPS measurements (Rakonczai 2000; Rakonczai and Kovács 2000). Berm recession of 20 m per 100 years was demonstrated (15–20 cm $year^{-1}$), equivalent to soil removal of 3 $m^3$ $year^{-1}$ $ha^{-1}$. Based on the geomorphologic investigations of Solonetz soils of the Hortobágy puszta, under natural conditions the berm recession is very slow (0.2–0.7 cm $year^{-1}$). On the other hand, with the effects of anthropogenic interferences—e.g. intensive pasturage, artificially increased relief energy, etc.—this process takes place by several orders of magnitude faster (5–15 cm $year^{-1}$) (Tóth 2003).

## 27.4 Age of Alkalization

The time scale of processes and the age of microforms have been estimated differently by various authors. According to previous views (Treitz 1924; Rapaics 1916), the majority of our alkali puszta and its vegetation is of anthropogenic origin. The formation of the alkali soils of the Great Plain (including the Hortobágy) was explained by the major 19th century interventions involving flood control and land drainage. This view is contradicted by the 18th century reports and physiographies written by travelers and explorers, like Robert Townson or Pál Kitaibel, who described the landscape, animal and plant species. It is clear that alkali flats already existed in the Hortobágy well before the flood control and drainage operations (Gombocz 1945; Nyilas 1999).

According to recent research, alkalization in the Hortobágy had already begun in the Late Pleistocene (Sümegi et al. 2000). On the eastern margins of the Hortobágy and the Hajdúság alkali soils formed before the Holocene (32–26 ka BP). At Nagyhegyes, 4.5–5 m below the surface, a light-coloured paleosol (szik soil on the verge of steppe-forming) was found (Szöőr et al. 1992). Others (Szabolcs and Máté 1955; Somogyi 1965) placed the beginning of alkalization in the Boreal stage of warm and relatively dry climate, when alkali areas (both Solonchaks and Solonetzes) could have been more extensive than currently. Solonchaks were developed in floodplains and lands of poor drainage. Solonetz soils were formed in more elevated areas, where the capillary zone did not reach the surface, and/or—because of the high clay content—cation adsorption surpassed the accumulation of water-soluble salts. Pre-Holocene alkalization has both direct and indirect (sedimentological and paleontological) evidence, such as the significant proportions of endemisms. The effects of the continental climate, river regulation and flood control, traditional land use (extensive farming) only contributed to stabilization and expansion of alkalization (Sümegi et al. 2000).

The last natural period of alkali soil formation is presumed to happen in the Subatlantic stage of the Holocene, when, under drier and warmer climate, evaporation and capillary rise increased—primarily next to low floodplains. This repeatedly enabled the alkalization of subsoils and, thus, the formation of Solonetz-type soils (Rónai 1954; Somogyi 1965).

Current soil processes and their distribution have been influenced by dropping groundwater levels (by 10–100 cm; in the Hortobágy 50–80 cm) since flood-control measures were implemented (Szabolcs and Jassó 1961). Some Solonchak soils have been leached, ceased to exist, or transformed into Solonetz. On the other hand, in formerly flooded spots alkalization could begin, while some Solonetzes show a tendency towards steppe soil (Chernozem) formation (Szabolcs 1961).

## 27.5 Classification of Alkali Microforms

As it is clear from the above, alkali soil erosion requires very low relative relief of usually only 10–30 cm. This is still sufficient to establish distinguished "floors of the puszta". The following section describes alkali microforms from the highest level down to the base level.

- Natural levee: terrain rising 40–60 cm (rarely even 120–180 cm) above the base level of erosion, unaffected by erosion, with an intact soil profile, never waterlogged; most often with *Salvio-Festucetum rupicolae* or *Astragalo-Poetum angustifoliae* plant associations.
- Berm top: lower section of levee, sloping slightly towards the berm edge, covered by a closed saline pasture (*Achilleo-Festucetum pseudovinae*, *Artemisio-Festucetum pseudovinae*).
- Berm edge, erosional micro-cliff (in more strict sense the *szik* berm itself): a bare slope of 18° to 75° angle between the berm top and berm slope (*szik* slope); 5–30 cm high on the average. It is an abrupt, apparent landmark and the boundary between eroded and uneroded surfaces (Fig. 27.2).
- Berm slope: a very gentle slope between the berm edge and the edge of the bottom with accumulation in its highest parts and mainly erosion processes further away from the edge; soil profile not intact: A-horizon almost completely removed or only 1–2 cm deep. Its surface is usually bright white (silica accumulation). The B-horizon of salt accumulation is exposed on the surface or right below it, which renders it the least favourable habitat for the vegetation, appropriately called *vakszik* (blind alkali spot) in Hungarian. In winter it is temporarily waterlogged, but by the end of spring it turns to a dry and bare habitat. Only early summer rains induce some vegetation growth (mainly *Camphorosmetum annuae* and *Puccinellietum limosae* associations) (Figs. 27.2, 27.3, 27.4 and 27.5).
- *Szik* rills: flat features of various widths and depths, bordered by berm slopes; mostly dry, but after downpours and heavy rainfalls and during spring snowmelt become active; channel waters from higher landforms down to the base level, but they often end with no real drainage. *Szik* dolinas are smaller (a few metres in diameter), while *szik* flats are larger (several 10 m) closed depressions with temporary waterlogging. *Szikfok* or

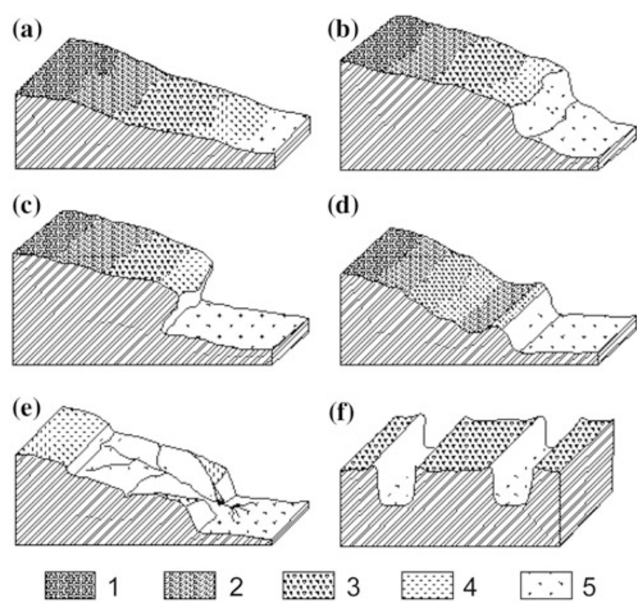


**Fig. 27.3** Geomorphological types of salt berms with plant associations. **a** gently sloping salt berm without berm edges, **b** stepped scarp with sharp berm edge, **c** stepped scarp with overhanging berm edges, **d** sloping, stepped scarp, **e** double stepped scarp, **f** berm formation caused by wheel tracks. Plant associations: 1 *Salvio-Festucetum rupicolae*; 2 *Cynodonti-Poetum angustifoliae*; 3 *Achilleo-Festucetum pseudovinae*; 4 *Artemisio-Festucetum pseudovinae*; 5 *Camphorosmetum annuae et Puccinellietum limosae*

**Fig. 27.4** Heavily cracked Solonetz surface in the Hortobágy (photo by Csaba Tóth)

**Fig. 27.5** Barren and eroded surfaces between intact berm tops on Solonetz in the Hortobágy (photo by Csaba Tóth)

berm foreground is a mixed (wet and dry) habitat type, steadily waterlogged in winter, and dry only by late summer. *Szik* bottom is an area permanently waterlogged, the deepest and largest portion of the *szik* landform assemblage.

The berm edges develop differently depending on soil properties, slope angle, vegetation cover and human impact—also manifested in their morphology. Alkali terrains can be classified into six evolution stages, also differing in their plant associations (Fig. 27.3):

1. *Initial stage Formation of a cracked surface* During the dry period the Natric or Vertic horizon (both rich in clay minerals) heavily cracks up, and the cracks are inherited from the subsoil to the topsoil—even if it is not prone to cracking. If the cracks would become 2–3 cm wide and 10–15 cm deep, they become permanent forms and the starting point of erosion—both horizontally and vertically (Fig. 27.4).
2. *Juvenile stage Development of szik rills and depressions without drainage* Rills, 25–30 m long, 1–2 m wide and 20–30 cm deep, continuously retreat by headward erosion along cracks and meander as small channels towards the base level, with remains of the original relief in the form of tiny islands. In many cases, on the lower levees with limited relative relief, no advanced *szik* rills are found. Along the cracks vertical material transport is much more dominant, and so the formation of undrained oval or elongated depressions are typical.
3. *Mature stage Widening and merging of rills* At this stage the *szik* rills become ever wider, and form a blind drainage area of various sizes. As a consequence, the size of the original, uneroded surface decreases significantly. From a top-view, these areas—the berm tops and *szik* flats—stand out with their exceptionally colorful, fractal-like patterns.
4. *Senile stage Formation of extensive alkali flats* In the parts of the natural levée where several *szik* rills merge and run into the final receiver, large (more 10 m$^2$) flat, barren and eroded tracts. On this devastated Solonetz topography a few residual islands resist erosion (Fig. 27.5).
5. *Human impact on the development of alkali flats* Alkali berms of anthropogenic origin primarily develop along artificial ditches, holes, or (in the wet periods) wheel tracks of motor vehicles, where the abrupt increase of relief could initiate erosion. Intensive animal trampling (sheep, cattle) could also start or amplify this process. Studies show that such features develop several orders of magnitude faster than under natural conditions.

## 27.6 Microforms in the Study Area of the South-Hortobágy

In the southern part of the Hortobágy National Park a 10 × 13 m test area was chosen for measuring the rate of erosion. The geomorphology of the experimental site is representative of the abandoned floodplain of the Hortobágy, where the drainage pattern often changed in the early Upper Pleniglacial time (30–20 ka BP). Since those times surface features had been transformed only by stagnant inundation originated mostly from precipitation accumulated on frozen, or water-saturated—and therefore impermeable—soils. Degraded point bars, meander scars and remains of river channels are observable on the terrain, having 1–1.5 m of relative relief per 1,000 m. To initiate erosion, relative relief of ca 110 cm between the highest point of a point bar remnant and the deepest point of a meander scar within 50 m was sufficient. The study area is situated on the western slope of a point bar, in the middle of a mature erosional field in which fractal-shaped curves divide areas of eroded, bare soils from non-eroded, grass-covered tracts. Here erosion processes were not influenced by ditches, dikes or any other human structures, but followed from traditional land use (sheep and cattle grazing). The soil type of the non-eroded pedon was classified as Vertic Solonetz (Albic, Humic, Siltic), according to the WRB system. Non-eroded terrain was covered by closed short grass vegetation (*Artemisio-Fetsucetum pseudovinae*), the eroded sections were bare, or with sparse *Camphorosma annua*, *Puccinellia limosa* and *Artemisia santonicum*.

Micro-scale erosion processes, resulting in complex structures of micro-cliffs, micro-channels and micro-alluvia, are initiated by deep cracking of the Vertic B-horizons through shrinkage. Although shrinkage is not typical of the topsoil, because it is not as rich in clay minerals as the subsoil, some substance of the A-horizon would fall in the cracks, and the topsoil is also disrupted in patterns reflecting the cracks of the B-horizon. After wetting in spring and early summer, clay minerals swell and the cracks of the B-horizon close, while discontinuities in the topsoil usually remain. If the surface is waterlogged (for instance, because of the subsoil being still frozen after snowmelt), these micro-channels are filled with water, flowing slowly towards the base level, and widen and deepen with time.

With such processes seasonally recurring, the devastated topsoil surfaces are gradually extending, while the grass-covered terrain with intact soils is shrinking. The two terrain types are divided by erosional micro-cliffs 15–30 cm high and of 60° to 80° average slope. Due to the high erosion resistance of the clay-rich (Natric, Vertic) B-horizon, the surface is not deepened below the boundary of the A and B-horizons. Although vertical incision is mostly inhibited, the horizontal spreading of the channel network continues until the relative relief is consumed entirely.

In the experimental field a semi-circular (in plan view), 22–25 cm high, micro-cliff was found, crossed by two smaller micro-channels (Figs. 27.6 and 27.7). In the

**Fig. 27.6** Autumn aspect of the study area at Ágota-puszta, South-Hortobágy. The gap in the rim of the depression appears on the lower part of geomorphological map (Fig. 27.7) (photo by Csaba Tóth)

foreground a bare, eroded surface and at the outflow of the channels two micro-alluvia occurred. Elevation changes could only be established through high-resolution geodetic surveys. Such field measurements were first carried out in 1998, and were repeated later, in 2012. In this 15-year time interval surface elevation typically showed minimal to negligible changes (Fig. 27.7). The retreat of micro-cliffs was found to be no more than 10–45 cm, and the change of slope decreased from 65°–68° to 45°–55°, with no significant change at many places. During this period, the alluvial accretion was between 1 and 2 m, and the erosion channels widened by 20–30 cm.

## 27.7 Conclusions

In the areas of the Great Hungarian Plain least favourable for agriculture (with Vertic or Natric subsoils, alkali soils of Solonetz or Solonchak type) rainwater erosion resulted in spectacular microtopography. The processes are proved to be active as early as the Late Pleistocene and the warm and semiarid spells of the Holocene—although much more restricted in spatial distribution than today. The area of alkali soils really began to expand in response to the effect of sinking groundwater levels caused by river regulation and land drainage in the second half of the 19th century. A contiguous alkali puszta, the largest in Hungary, is found in the one-time floodplain of the Tisza River, the Hortobágy. The findings of the geomorphological investigations performed here are summarized in the following:

1. The joint influence of several natural factors is responsible for the origin of alkali microforms: Vertic and Natric soils with high clay contents and easily prone to cracking, relatively high relief (margin of natural levees) and precipitation of proper amount and intensity.
2. Under natural conditions the rate of erosion is very low (some millimetres per year), manifested in the retreat of berm edges and sedimentation in the berm foregrounds.
3. Human influences (digging canals, wheel tracks, intensive grazing) increase the rate of evolution of microforms by orders of magnitude.
4. The extremely varied microtopography generated by berm erosion is coupled with a very dynamic mosaical pattern of soil types and associated plant communities, making the Hortobágy one of the most diverse microregions of Hungary at the microscale.

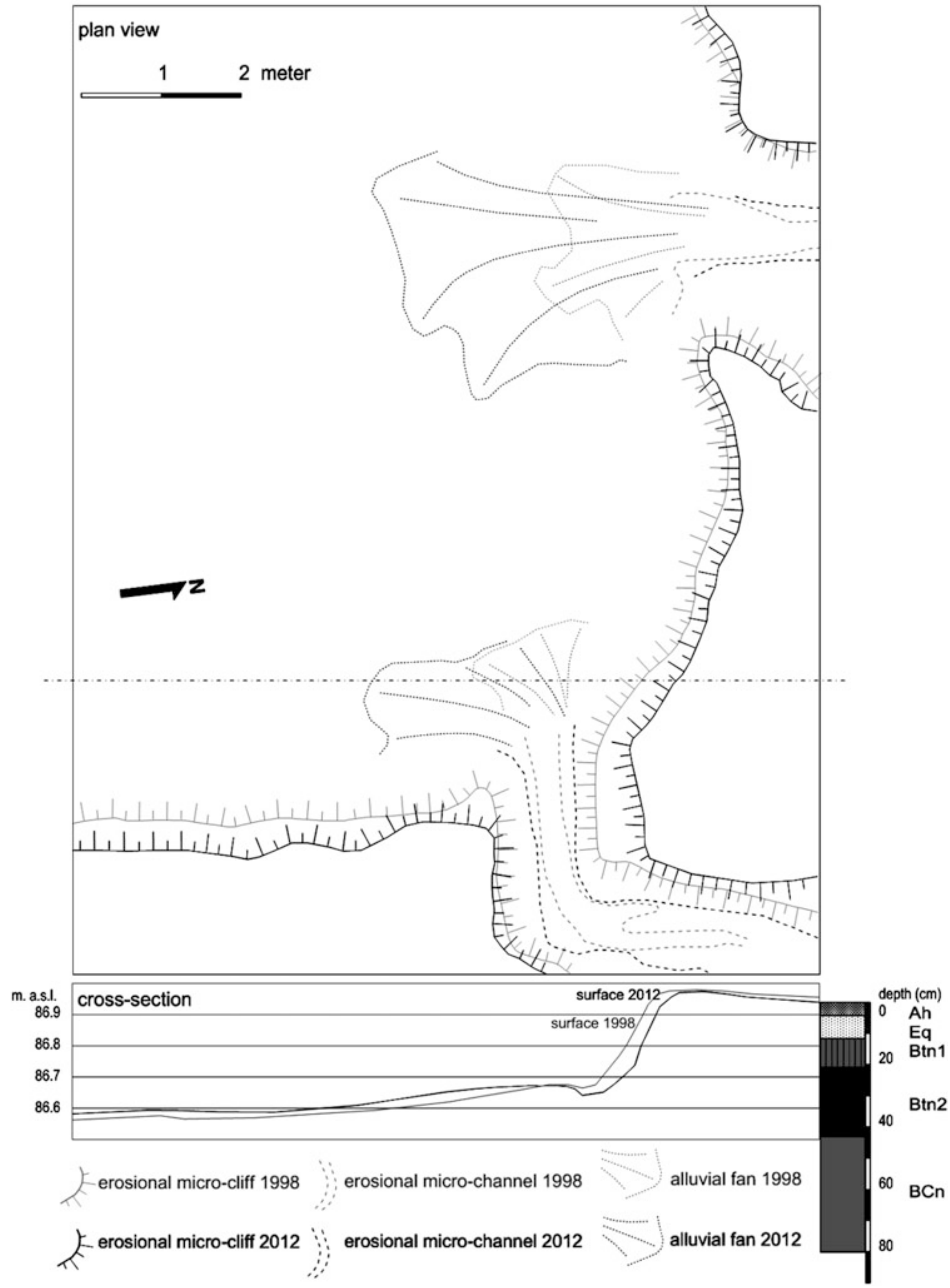


**Fig. 27.7** Geomorphological changes on the salt berm in the South-Hortobágy study area, 1998–2012 (edited by Tibor Novák)

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# The Crater Lake of Nagyhegyes

# 28

Dénes Lóczy, József Szabó, and Csaba Tóth

**Abstract**

In spite of an increasing security of operations, hydrocarbon exploration and extraction may involve accidents which sometimes cause localized but rapid and fundamental changes on the ground surface. In Hungary the scars generated by such accidents are still clearly detectable in many places, best visible in the outskirts of the village Nagyhegyes, where a major explosion took place in 1961. The resultant depression somewhat resembles crown holes (of mine subsidence origin), but has a 6.5-m-high circular rampart. The basin, sealed from below, soon filled up with groundwater and developed into the Crater Lake as we know it today. Although affected by erosion (minor landslips on steep slopes) and sedimentation, particularly in the first decade of its existence, the lake with the rampart has stabilized and survived. Now it functions as a recreational oasis encircled on all sides by the agricultural landscape of the Hajdúság loess-mantled alluvial fan of the Great Hungarian Plain.

**Keywords**

Natural gas • Blast crater • Mud volcano • Landslips • Siltation rate • Hajdúság • Great Hungarian Plain

## 28.1 Introduction

Man-induced landforms produced by diverse processes, including the extraction of various mineral resources or fossil fuels, are not uncommon worldwide (Bennett and Doyle 1997; Szabó et al. 2010). The surface manifestations of collapse of a subterranean cavity caused by shallow coal mining and rock quarrying are called crown holes. This is an unpredictable but gradual process. The zone of collapse may migrate upwards over several years, until the ground surface is suddenly affected (Braithwaite and Sklucki 1987; Blodgett and Kuipers 2002). An example can be cited from the English Midlands, where the improper infilling of Upper Wenlock Limestone workings on the Dudley Cricket Ground led to crown hole formation on 25 May 1985 (Braithwaite and Sklucki 1987). Closer in space and time, in the Nitra mining area of Slovakia, at Nováky, on 16 November 2006 a crown hole of 11 m diameter was generated above an active coal mine about 200 m underground in the middle of an arable field (ČTK 2007). The collapse of the overburden involved ca 2,500 $m^3$ of gravel and soil saturated with water and caused the death of four miners, who according to rescue workers had no chance to survive.

Equally, hydrocarbon extraction from moderate depths may also be responsible for major surface modifications. Wetland losses in the US Gulf Coast are largely associated with accelerated land subsidence and fault reactivation (Morton et al. 2006). The processes are induced by dropping reservoir pressures as a consequence of prolonged and

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improperly implemented extraction of natural gas or mineral oil. Evidence, among others, include measurable offsets of near-surface strata and accelerated subsidence rates near producing fields (Morton et al. 2006). To cite another classical American example, the average rate of subsidence for the Wilmington Oil Fields at Long Beach, California, was found to be around 0.75 m year$^{-1}$ (Mayuga and Allen 1970), while more recent inferometric analyses measured rates up to 1 mm day$^{-1}$ (Fielding et al. 1998).

Numerous further examples may be cited to show that gas explosions and groundwater effects are able to cause abrupt or even catastrophic geomorphic changes above mine galleries and hydrocarbon extraction leads to ground subsidence and cracking, but a blast crater formed by an accident during hydrocarbon drilling and developed into a permanent lake—as it happened in Hungary—is regarded unique—or at least not known to the authors from on-line sources.

## 28.2 Geographical Setting

The crater is located in the Hajdúhát mesoregion, an elevated old alluvial fan of the Tisza River, covered by loess and loessy silt (Szabó 1964; Szabó et al. 1999). Among the physico-geographical conditions, the properties of the soil and the underlying parent sediments are most relevant.

The soil at the explosion site is described based on the analysis of a soil pit ca 1 km southeast of the crater lake (Ébényi and Schmidt 1939). The analysis was made in 1936—almost exactly 25 years before the crater came about. A fertile pseudomiceliar, slightly leached and acidic (pH 6.5) Chernozem (in WRB classification: Luvic Chernozem) was identified. It has (clayey) loam texture and crumbly structure with humus layer 80–100 cm deep, 3–4 % humus and high nutrient content (P: 0.15–0.2 %, $K_2O$: 0.2–0.3 %).

The Chernozem is underlain by calcareous (14–16 %) loess, replaced by sand below 5.5 m depth. From the borehole at the Nagyhegyes school (Ébényi and Schmidt 1939) it is assumed that the 28-m thick sand deposit is underlain by 47 m of clay, 50 m of silt, 75 m of clay with calcretes ($CaCO_3$: 15–20 %), and alternating thin clay and sand layers. During the summer survey the groundwater table was found at a position exceptionally low for the Great Plain: at −7.5 to 8 m. The groundwater table probably rose during the wet period of the 1940s and 1950s (Csordás and Lóki 1989).

## 28.3 The Explosion

In the 1950s natural gas reservoirs were first explored in the southern Great Plain. Today half of the more than 7,000 hydrocarbon wells in Hungary, i.e. ca 3,500 of them, are

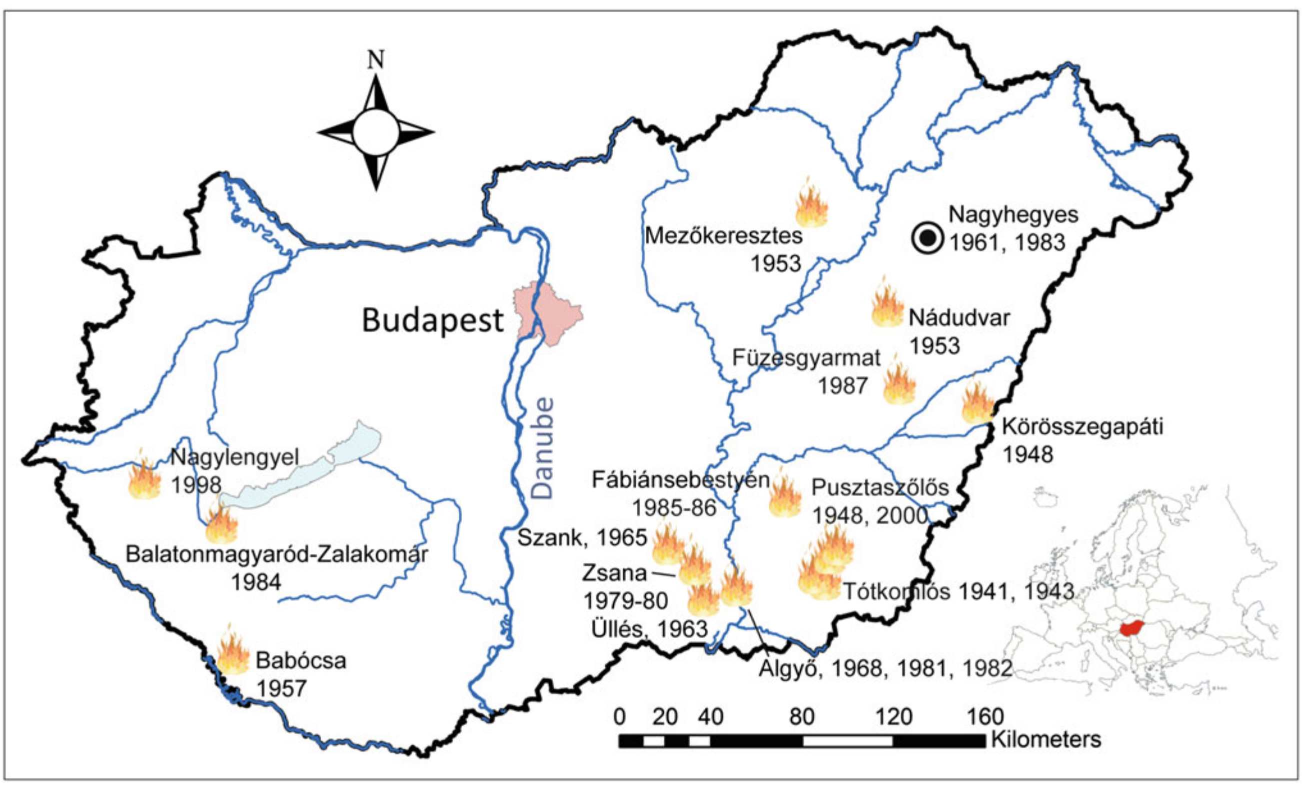


**Fig. 28.1** Location of the Nagyhegyes crater in Hungary (with sites of other hydrocarbon mining accidents) (by Dénes Lóczy)

**Table 28.1** Major accidents during oil and gas explorations or extraction in Hungary (compiled by D. Lóczy from various sources)

| Date | Site | Oil or gas loss per day | Diameter of area with topographical change | Remark |
|---|---|---|---|---|
| 4–15 August 1941 | Tótkomlós | Oil, saltwater, debris | Large crater | Borehole depth: 1,613 m |
| 18–22 June 1943 | Tótkomlós | 0.088 million $m^3$ of gas and water | 40 m, permanent pond: ca 30 m | |
| 1948 | Pusztaszőlős | 0.88 million $m^3$ of gas, oil and water | 35 m, temporary pond | |
| 1948 | Körösszegapáti | N/A | 72 m, permanent pond | |
| 1 July 1953 | Nádudvar | N/A | N/A | 21 days of fire |
| 11 August 1953 | Mezőkeresztes | N/A | N/A | |
| 1957 | Babócsa | N/A | ca 35 m | |
| 24–29 August 1961 | Nagyhegyes | N/A | 200 m, permanent lake: 58 m | Detailed in this chapter |
| 1963 | Üllés | N/A | 26 m | |
| 1965 | Szank | N/A | N/A | Site not detectable |
| 19–26 December 1968 | Algyő | 110 t of oil; 0.8 million $m^3$ of gas | 70 m | |
| 24 January 1979–15 February 1980 | Zsana | 4.4 million $m^3$ of gas | N/A | |
| 29–31 December 1981 | Algyő | N/A | ca 15 m | |
| 7–17 August 1982 | Szeghalom | N/A | 16 m | |
| 30 January 1983–2 February 1984 | Nagyhegyes | N/A | N/A | |
| 22–23 March 1983 | Battonya | N/A | 75 m, temporary pond: 43 m | |
| 14–17 October 1982 | Algyő | N/A | 40 m, permanent lake: 27 m | |
| 18–21 June 1984 | Balatonmagyaród-Zalakomár | N/A | N/A | |
| 16 December 1985–31 January 1986 | Fábiánsebestyén | Hot water + gas | N/A | Borehole depth: 4,239 m |
| 24–25 January 1987 | Füzesgyarmat | N/A | N/A | |
| 14–17 November 1998 | Nagylengyel | N/A | N/A | Oil eruption, inhabitants from three villages evacuated |
| 18 August–16 November 2000 | Pusztaszőlős | 50–70 million $m^3$ of gas | 83 m | Damaged pipeline 300 m underground exploded, the longest gas eruption in Hungary |

found in the Great Plain. The history of hydrocarbon exploration is accompanied by accidents, gas and oil eruptions (Fig. 28.1), which involved topographic changes at various scales as well as serious damage to industrial infrastructure and economic losses (Table 28.1). In each case the causes and circumstances of an accident were different.

In the environs of the town Balmazújváros, in the outskirts of a recently (in 1952) established settlement, Nagyhegyes, the centre of a network of hundreds of scattered farmsteads (Fig. 28.1), the natural gas reserves, estimated at 33,000,000,000 $m^3$, were localized at 670–1,210 m depth in 1958–1959.

The accident happened at dawn, on 24 August 1961, when, after drilling reached 1,391 m depth and the placement of well casing was ready down to 390 m (Borsy 1967), the pressure in the natural gas reservoir exceeded the weight of the mud column, supposed to counteract gas pressure. Although this risk had been known before, no countermeasures were taken. The gas exploded and soon destroyed the 45-m-high derrick together with the fragments of the well casing. The explosion set the emitted gas onto fire (Fig. 28.2). Fortunately, no intervention was needed to extinguish the fire: soon the accumulating water-saturated sludge clogged the vent of the gas eruption and to some extent sealed the floor of the future lake.

**Fig. 28.2** Flames shooting 150 m high of the erupting gas well on 25 August 1961 (archive photo by László Kádár)

**Fig. 28.3** The Crater lake from the rampart in August 2013 (photo by Csaba Tóth)

**Fig. 28.4** The forested rampart and the lake shore in October 2006 (photo by Csaba Tóth)

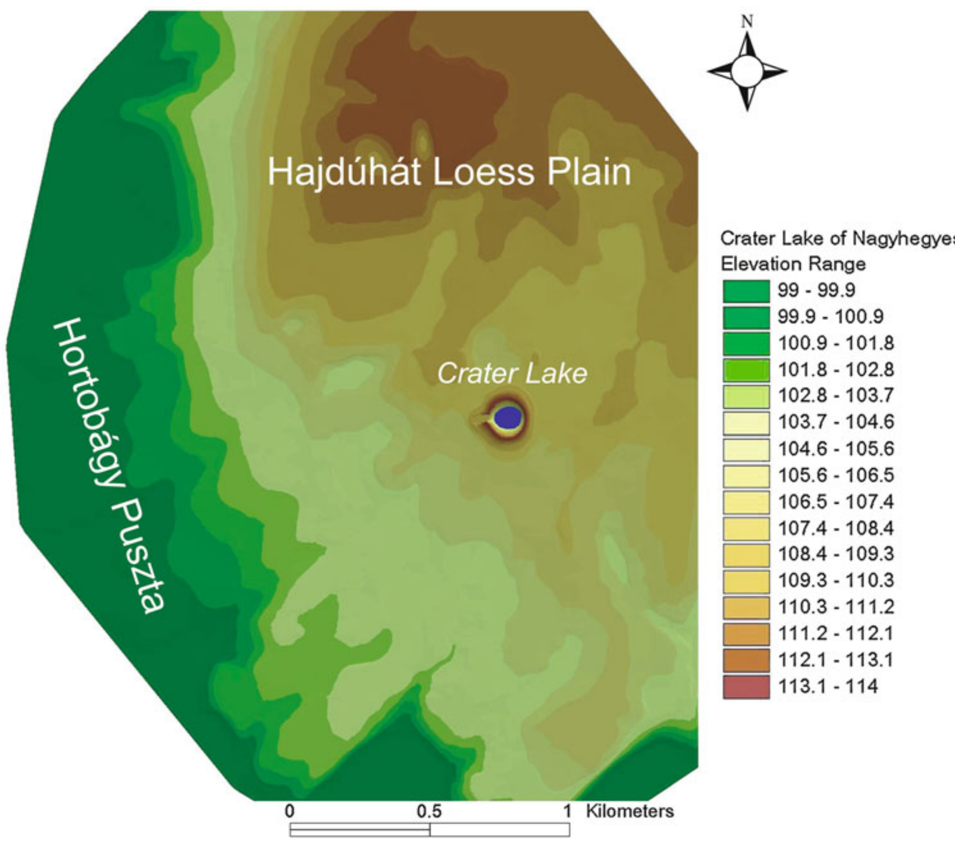


**Fig. 28.5** Digital elevation model of the Crater lake based on topographic map at 1:10,000 (by Csaba Tóth)

## 28.4 Geomorphological and Geoecological Changes

The explosion fundamentally changed the local landscape from the first moment on. The total amount of sediments (loess and silt) ejected reached 172,000 $m^3$. Most of it was accumulated in a regular circular rampart around a crater of almost 200 m diameter (Figs. 28.3 and 28.4), but the rest was scattered within 1.5 km distance from the eruption site (Borsy 1967). According to the theodolite measurements by Borsy (1967) and his students, in 1965 the rampart rose 6.5 m above the level of its environs (107 m above sea level), while the floor of the depression was 13.3 m below this level (Fig. 28.5). At present (November 2013) the top of the rampart of unequal height is at 112.5–113.5 m elevation. A secondary crater of ca 50 m diameter also took shape on the northern margin of the big crater but has been filled up since.

The suddenly released pressure generated ruptures and dislocations in bedrock (Borsy 1967). The length of ruptures reached more than a 100 m. After the extinction of the fire the accumulating pressure thrust large amounts of mud into the air to 8–10 m height. Borsy (1967) described the process that involved the periodical ejection of saturated silt from long cracks as a mud volcano activity. After clogging of the vent and cessation of gas supply, a small lake of "endogenic origin" (Borsy 1967; Sütő 2010), rather unique in Hungary, was generated. In some other cases, too, temporary ponds originated, like at Pusztaszőlős and Körösszegapáti—see Table 28.1.

During his field studies in 1965, Borsy found some asymmetry in the shape of the rampart: it was 47–55 m wide at its base on the northern side, while only 31–34 m on the southern one (Borsy 1967). Assuming that the ejection of materials was uniform in all directions during the eruption, he explained this asymmetry by minor landslips on the very steep (35° to 55°) inner slopes of the rampart (outer slope angles did not exceed 25°). Locally even vertical slopes with topples could have been observed on the yet unvegetated surface. The landslip tongues have been partly planated in 2006, when the lower trail was established.

Few data are available on the long-term siltation rate of the lake, which has now a diameter of 58 m. After the basin was filled up with groundwater, the resulting lake was 9.3 m deep. Barometric GPS measurements show that the present (November 2013) lake level is at 99.8 m elevation. Radar measurements indicate that the present (2012) water depth is 5.9 m (personal communication from Ferenc Luka). Assuming that lake level sank 3.2 m due to the drought periods after the 1960s, this means a long-term siltation rate of ca 0.1 cm $year^{-1}$.

Over the decades a particular ecotope has been created within the crater: a poplar wood (*Populus alba var. canadensis*) with trees of almost uniform age (ca 40 years) and height (ca 20 m) encircles the lake. The moist but warm microclimate, also noticed by Borsy 1967 and further enhanced by dense vegetation, which is in sharp contrast with the arable land around, is partly due to heat storage by the water mass of the lake and partly to the wind-sheltering and humidity-controlling influence of the poplar wood.

## 28.5 Conclusions

A double message can be drawn from the history of the Crater Lake at Nagyhegyes:

1. Naturally, the accident and the generation of the lake could have been prevented. A more careful appreciation of the geological conditions could have revealed that the reason for the silt deficit is the well-developed calcrete and the drilling technology should have been adjusted to the situation.
2. On the other hand, once the lake had formed, the benefits offered by the existence of this "oasis" have to be optimally exploited. The wetland is a refuge for aquatic animals and birds and, in addition, it should serve multiple human functions: a recreation (picnic) area with pleasant microclimate, a spot for anglers, and also serves demonstration purposes. It shows what kind of geomorphic processes can be induced by a drastic human intervention into the natural environment and how nature recolonizes surfaces created by human action.

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# Lyukas Mound: One of the Many Prehistoric Tumuli in the Great Plain

# 29

Csaba Tóth, Katalin Joó, and Attila Barczi

**Abstract**

Burial, residential, watch and border mounds called collectively "tumuli" (also "Cumanian hillocks" or "kurgans") have been under protection in Hungary since 1996. They are special man-made landmarks of the steppes of Eurasia, including the Great Hungarian Plain. From an archaeological point of view, in many cases they preserve valuable remains of one-time loess steppes, and through studying the buried soils, their stratigraphy, and palaeoecology new data can be obtained on the ancient environment, flora and soil formation in the past. The buried soil profiles under burial mounds are the messengers of ancient landscape-forming factors, soil processes and palaeobotanical patterns. Results from palaeopedology, soil micromorphology, palaeobotany (phytolith analysis and palynology) and archaeology are summarised in order to understand how tumuli were constructed and what their former palaeoenvironment was like.

**Keywords**

Cumanian hillock • Kurgan • Palaeoenvironment • Palaeosols • Palaeoecology • Great Hungarian Plain

## 29.1 Introduction

In the lowlands of the Carpathian Basin, particularly in the Great Hungarian Plain, on flood-free river terraces, flood-banks, and less frequently on shifting sand dunes there are thousands of artificial mounds 3–12 m high. The mounds of various functions were built from the late Neolithic through the early Migration Period to the early Middle Ages. In addition to material finds important for cultural history, the man-made mounds possess a number of natural values.

In Hungary, Cumanian hillock is the collective name of all prehistoric mounds of different types and ages. This term suggests—somewhat incorrectly—that such mounds were built by Cumanian people in the 13th century. In contrast, archaeological excavations and datings clearly demonstrate that a vast proportion of the mounds is older than the graves of the Cumanian people (Fig. 29.1).

The mounds are considered potential archaeological objects, and through their exploration we can learn a lot about the life and customs of human cultures. In addition to their archaeological significance, the mounds may carry many other cultural and scientific values, so it was definitely justified to protect them *ex lege* (Act No. LIII. of 1996 on Nature Conservation in Hungary). The national inventory, completed in 2002, identified and registered ca 1,900 mounds. Based on cartographic sources the total number of mounds could be some ten thousands, but it had been significantly reduced by the mid-20th century and their condition had drastically deteriorated. During the 19–20th centuries

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**Fig. 29.1** Typical kurgan of the Carpathian Basin (photo by Attila Barczi)

hundreds of them were demolished and ploughed mainly due to the expansion of large-scale agriculture subsequent to river regulations. For many tumuli even the name fell to oblivion.

From the aspect of geomorphology, landscape and nature conservation and the protection of cultural historical values, the salient importance of the remaining mounds is explained by the following:

- Cumanian hillocks are essential landmarks in the Hungarian Plain. In many cases their sight and the view upon the surrounding landscape from their tops give us an aesthetical experience.
- In the intensively cultivated agricultural area of the Great Hungarian Plain only a few mounds, left out of the arable farming for some reason, preserve the original loess vegetation. Even in degraded state they are the last refuges of the once characteristic plant associations and animal communities.
- The environmental conditions at the time of origin of the mounds can be reconstructed by geomorphological, pedological, palaeoecological and biogeographical methods in order to understand landscape evolution.
- The tumuli had been important sacred places for the cult of the dead and various pagan ceremonies even before Christianity. A large number of churches, chapels and calvary stations confirm the fact that after the emergence of Christianity these mounds retained their central importance as cult sites for smaller communities.

## 29.2 Distribution of Mounds

The conditions offered by the sites of tumuli were favourable for the settlement of oriental nomadic ethnical groups engaged in large livestock breeding. As they were used as burial sites from the end of the Copper Age, most of the mounds in Hungary belong to the kurgan type. A smaller proportion of them was built during late Neolithic and Middle Bronze Age by settled human cultures. These residential mounds, called tells, are rich in archaeological finds.

Most of the tumuli are located in the Great Hungarian Plain, and a smaller portion of them are found in the neighbouring hills and mountains. On the sand dunes areas of the Plain (Nyírség Region, Danube-Tisza Interfluve) the mounds are virtually absent (Fig. 29.2). We may find the highest density of mounds in alluvial and loess plains rich in active and abandoned river channels, preferably on natural levees (Fig. 29.3). The geographical distribution of mounds in the Great Hungarian Plain clearly shows certain regularities both horizontally and vertically. When mapping the mounds a linear arrangement was observed: they occur along curved lines (Tóth and Tóth 2011).

The absolute and relative altitudes of the mounds of the Great Hungarian Plain show regularity in their vertical arrangements too: the tumuli occur only above a certain elevation, i.e. that of the flood-free levees. For example, the majority of mounds of the Hortobágy and Nagykunság area are at 90 m above sea level, on levees and small aeolian dunes. Mounds can be rarely found in lower-lying areas, and there are no mounds at all below 86.6 m. This area has been permanently covered by water before river regulations (Bukovszki and Tóth 2008). The national average of the relative heights of kurgans in Hungary is 3.2 m, but there are significant regional variations. The low floodplain wetlands of the Hortobágy were less suitable for human settlement, therefore more low kurgans (average height 2.6 m) were erected there for this purpose. Small and large clusters are mainly concentrated along streams in this region. However, in favourable levee locations (the loess-covered Békési-hát and Hajdúhát), heights are above the national average (even 8–10 m)—although there are less than one hundred of these impressive mounds. The highest prehistoric mound in Hungary (the Gödény Mound) is found on the Békési-hát with 12.2 m relative height.

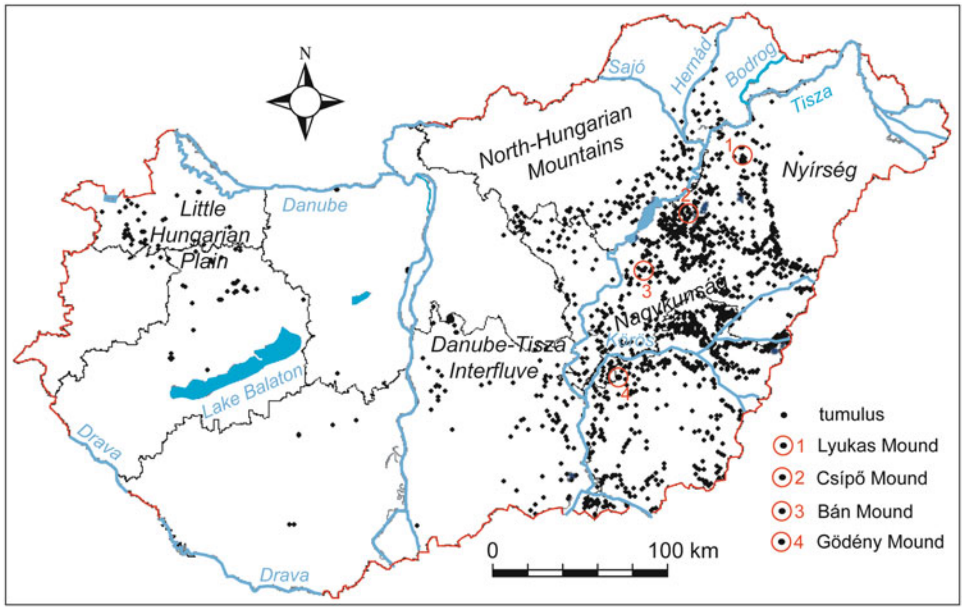


**Fig. 29.2** Distribution of tumuli in Hungary (*source* Ministry of Environment and Water Management, National Inventory of Tumuli)

**Fig. 29.3** Hegyes Mound is situated on the flood-free bank of the Kösely Stream (Nádudvar) (photo by Csaba Tóth)

## 29.3 Research History

The origin of tumuli was interpreted differently by Hungarian geographers, archaeologists and ethnographers. Before the advent of research in archaeology and soil science the mounds were thought by some researchers to be natural formations created by seawater surge. The geologist József Szabó listed many natural mounds on both the right and left banks of the Danube and the Tisza Rivers and described them as reef-like formations, but after geomorphological and stratigraphic analyses of the mounds from around Isaszeg and southern Baja, he underlined their human origin (Szabó 1868). At the international congress of the prehistoric archaeologists and anthropologists held in Budapest in 1876, Flóris Rómer drew attention to the necessity of archaeological research of the mounds. It was he who made the first comprehensive study and inventory of the mounds (Rómer 1878). As a result of archaeological excavations which began in the late 19th century, an anthropogenic origin of the mounds became clear, and their age and the primary functions were also determined. Analysing archive map sources, the number of such mounds in the Carpathian Basin was estimated at 1,200.

Several kurgans fell victim to constructions (mainly roads) from the mid-20th century, but professional archaeological excavations and soil analyses enriched our knowledge on Cumanian hillocks (Ecsedy 1979; Raczky et al. 2002; Csányi 2003; Dani and Nepper 2006). Mapping, field survey and inventory of mounds in the Hungarian Great Plain began in the last decades of the 20th century. The database compiled after the national inventory in 2002 showed that the remaining mounds were in very poor conditions. Nearly half of the mounds are under intensive arable farming, 40 % of them are disrupted with their bodies damaged, and one-fifth of them are wooded mounds overgrown with weeds (Tóth and Tóth 2011).

The ambitious inventory project was followed by a series of scientific investigations for various (geomorphological, archaeological geology, soil science, island biogeography, palaeoecological etc.) purposes. Thus, for instance, geoarchaeological and palaeoecological studies were conducted on a Bronze-Age mound in the northern Great Hungarian Plain near Szakáld (Sümegi et al. 1998). Then the fossil soils buried by the mounds and the quarried materials became the focus of attention. Morphological, soil chemistry, malacological and phytolith analyses of buried soils and radiocarbon datings were applied to reconstruct the ancient environments of three kurgans (the Csípő, Lyukas and Bán Mounds) and to clarify the circumstances of mound construction (Barczi et al. 2006a, b, 2009a, b; Molnár et al. 2004). The investigations employing the principles of island biogeography for the species-rich loess grasslands remaining on some mounds focused on environmental effects affecting the living communities beside the description of the precious flora and fauna species and their isolate dynamic tests (Novák et al. 2009).

## 29.4 Functional and Geomorphological Classification

In Hungary three classes of mounds are distinguished in archaeology by function: residential mounds (tell mounds) with remnants of settlements from the late Neolithic and Bronze Age; burial mounds (kurgans) (Fig. 29.4)—Scythian, Sarmatian, etc. burial sites from the Copper Age; watch mounds with no archaeological finds. The mounds can be also classified from a geomorphological point of view, observing their arrangement related to each other and shape (symmetry or asymmetry) (Table 29.1).

The three primary functions of prehistoric and ancient mounds changed significantly in later centuries. During the Middle Ages and modern era most of them formed into border mounds, jurisdictional or scaffold mounds, and newfangled cultic mounds (Fig. 29.5). On the summits of almost every mound geodetic signs were installed from the mid-20th century, and arable farming was extended over most of them.

## 29.5 The Lyukas Mound

We can clearly see from the history of tumuli research that the first isolated investigations were followed by ever more complex, synthesizing studies seeking correlations. The kurgan called Lyukas Mound was highly disturbed in the spring of 1993, its considerable portion was removed, and subsequently the kurgan was strongly and continuously looted on many occasions. For the prevention of further disturbance or deterioration an interdisciplinary research team (including experts in archaeology, soil science and landscape research, palaeoecological investigations, radiocarbon tests etc.) was established at the end of spring 2003. The primary task of the

**Fig. 29.4** Csípő Mound in the Hortobágy (photo by Attila Barczi)

**Table 29.1** Functional and geomorphological types of mounds

| Functional types | | |
|---|---|---|
| Type | Age | Characteristics |
| Settlement mounds (tells) Tell (Arabic), meaning: hill | Late Neolithic (4000–3500 BC) early to middle Bronze Age (2600–1500 BC) | Mature, oval in planform, 6–8 m high, 80–90 m long and 40–50 m wide; layered settlements reached their present size gradually, over several centuries |
| Burial mounds (kurgans) Kurgan (Turkic-Mongolian), meaning: tumulus | Copper Age (3500–3000 BC) Scythian, Germanic, Sarmatian, Hungarian Conquest era and Cumanian burials | Conical, 3–11 m high, 25–40 m across; after initial burials in the Copper Age, used by many ethnic groups as burial sites too |
| Watch (sentinel) mounds | 4000–3500 BC and 2600–1500 BC | Low mounds with rare archaeological artifacts, used for information transfer between tells using light or sound by sentinels |
| Border mounds | Middle ages | Some kurgans and sentinel mounds used as double, triple or even quadruple border mounds on administrative boundaries |
| Scaffold mounds | Middle ages | Ancient mounds; scenes of executions in the Middle Ages (hanging mounds) |
| Cultic mounds | Middle ages | Churches and chapels of the Árpád Age and medieval villages built on ancient mounds |
| Geodetical mounds | 19–20th centuries | Triangulation points installed on almost every higher mound, most of them are under arable cultivation |

| Geomorphological types | | | |
|---|---|---|---|
| Field appearance | Shape | | |
| | Symmetrical mounds | Asymmetrical mounds | |
| Single or stand-alone mounds<br>• Double or twin mounds<br>• Triple mounds<br>• Mound rows<br>• Mound groups (mound grave field) | Round-shaped mounds (built on natural levees)<br>Elliptical shaped mounds (built on dunes) | *Natural asymmetry* Opened mounds (lateral stream erosion) | *Anthropogenic asymmetry* (quarried, carried off etc.): Opened mounds (20 % of mass missing)<br>Fractured mounds (>50 %)<br>Demolished mounds (>90 %) |

research team was to prepare a west-east cross-section after looting. Archaeological excavation campaigns were performed in 2004 and 2009. The case of Lyukas Mound also proves that geomorphological and palaeoecological studies coupled with archaeological and soil analyses may lead to a more complete reconstruction of the palaeoenviroment.

### 29.5.1 Location and Description

Lyukas Mound is a typical Great Plain tumulus located on the boundary between the Hortobágy alkali alluvial plain and the Nyírség sand region, near Tedej in the northern outskirts of the town Hajdúnánás, on the road leading to Tiszavasvári (Tóth 1999; Barczi et al. 2006b). Its elevation above sea level is 103.3 m and relative height is 6.5 m. The mound is of relatively large size (groundplan area: 2,200 $m^2$) and well preserved. The total area disturbed in relation to the kurgan construction could be ca 14,700 $m^2$.

On the First Military Survey Map from the late 18th century the tumulus appears as Szántó-halom, which means “ploughed mound”. At present, in the vicinity of the kurgan cultivation takes place, but the kurgan surface itself is not ploughed. However, on the mound body there is no natural vegetation, the secondary vegetation is especially young black locust (*Robinia pseudoacacia*). In cross-section large foxhole chambers and passages could be seen, which heavily ravaged the upper layers of the kurgan, especially in the east. From the north and west the mound is bordered by a dirt road, which compacted and cut into its original surface. Traces of human intervention can be found in the form of deep robber pits incised in the top part of the kurgan’s body.

### 29.5.2 Structure and Palaeoenvironment

The following layers are found in the mound (from bottom to top): bedrock (C layer), on which a multi-level, advanced climate zonal palaeosol formed, which represents the walk level of former surface ($A_{paleo}$ layer that grades into $B_{paleo}$ layer of the palaeosol) and on which the kurgan was built from the material of the first alluvium ($K_1$ cultural layer). In the break of constructions a thin younger soil was formed on this alluvium (second walk level, $A_{paleo2}$ layer). The second

**Fig. 29.5** On the prehistoric Zelemér Mound (Hajdúböszörmény) the ruins of a late Gothic (14th century) church have survived to the present day (photo by Csaba Tóth)

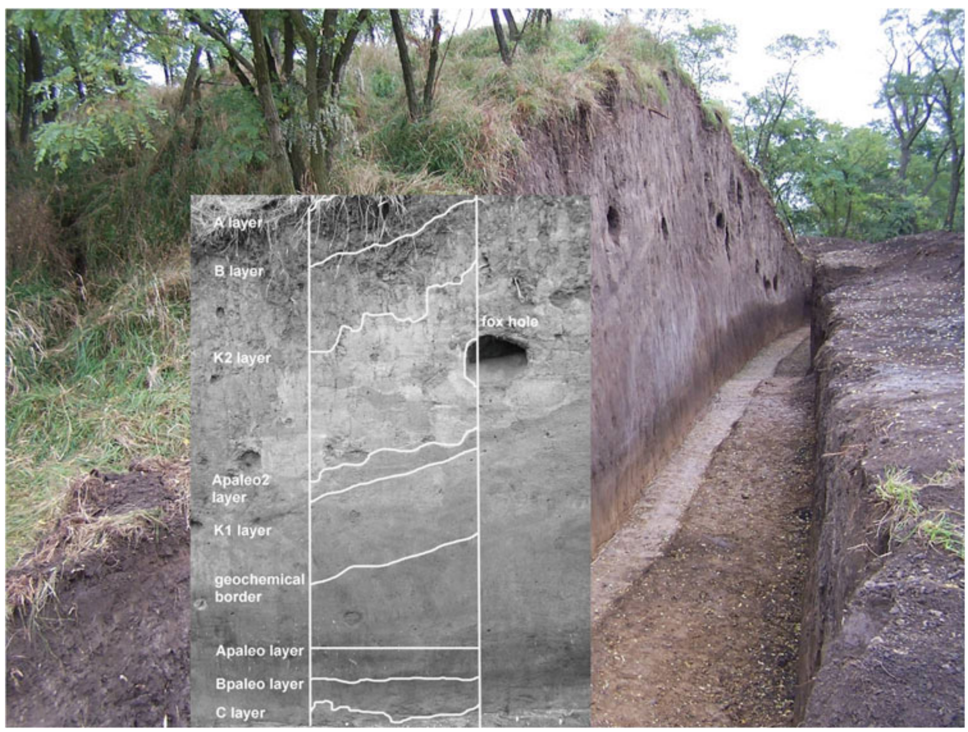


**Fig. 29.6** Cross-section and layers of Lyukas Mound (photo by Attila Barczi)

alluvium ($K_2$ cultural layer) covered the second palaeosol. On its surface recent A and B horizons layers of the modern soil developed (Fig. 29.6).

Summing up the existing data and information (Barczi et al. 2006b, 2009a, b), the loess type bedrock could be generated from falling dust on a periodically flooded landscape. By soil testing the parent material of the buried soil underlying Lyukas Mound is light-coloured, substantially calcareous, loess-type sediment with low organic matter content as usual for bedrock. The biomorphic analysis of the bedrock revealed indicators in small percentages (intact diatom skeletons and fractured/broken spines of sponge species) which are typical of surfaces developing under hydromorphic effect. Thus, excess water impact also prevailed in the formation of bedrock, which can either be the result of periodic water coverage, fluvial aggradation or falling dust deposited in aqueous environment.

The crumby structure forming under the grass, arid steppes (typical Chernozem soils) has been preserved in the palaeosol, in the morphology of the $A_{paleo}$ layer until today. Micromorphology is characterized by steppe micro-aggregate formation and significant organic matter reserves. The carbonate content also meets the requirements of a Chernozem topsoil. Mainly crop residues, predominantly steppe species, characterize the micromorphological facets. The sample which contained the highest amount of biomorphs comes from the humus layer of the buried soil profile. Traces of a highly productive vegetation biomass are found. The large amount of forms indicated grasses of steppe vegetation dominated by ancient *Stipa* species. Their abundance marks the former ground surface (walk level), which confirms the morphological and soil properties identified at the former palaeosol location. Thus, the soil morphological and micromorphological tests with the biomorphic analysis and physical and chemical properties of the soil show Chernozem soil formation. The $B_{paleo}$ layer is extremely rich in animal passages, it contains a large amount of carbonates, which is also characteristic of Chernozem-type soil formation.

In summary, it can be claimed that the ancient environment of the early metal-age kurgan is characterized by riparian forests, wooded grasslands, muddy and saline areas, above which higher loess hillocks emerged with Chernozem soils.

### 29.5.3 Reconstructed Building History

The kurgan was built in two phases. In the first alluvium ($K_1$ cultural layer) the source of the construction was the upper softer humus layers of soils from the immediate environment. The lower quantity of biomorphic particles and their qualitative heterogeneity is a consequence of anthropogenic origin. The existence of xerophile species is supported by the fact that a number of long-cell phytoliths were found in the sample (signs of xeromorphism). According to soil tests, the second palaeosol ($A_{paleo2}$) layer is a true soil layer with vestigial phytoliths, indicating heavy degradation.

Under the dry steppe vegetation of the upper kurgan surface a relatively short-term humus formation could start. It is likely that the thin soil layer is the product of this process. Nevertheless, we must reckon with that fact the accumulation of phytoliths takes a long time, so the little phytolith found in the layer can refer to its age too. However, the data obtained in micromorphological analysis of the layer clearly support our hypothesis that this level was also walk level and its material was affected by compaction. Animal (grazing) or human impact is assumed to have caused the compaction and also the parallel degradation of vegetation. Further heightening occurred relatively rapidly and prevented the formation of a more marked soil. The $K_2$ cultural layer is more mottled and more heterogeneous than the previous horizons, it probably carries the materials of subgrade levels of the environment. On top of the mound a young Chernozem soil formed (recent soil), which is typical of arid loess steppes. However, as a consequence of human impacts the surface of Lyukas Mound is characterized by a degraded locust grove with grass vegetation today.

The results of the archaeological research harmonize with the pedological, geological, botanical and palaeoecological studies (Horváth 2011). It is also demonstrated archaeologically that there was an interruption between the first and the second phase of construction, manifested in natural soil formation (development of the $A_{paleo2}$ soil). The ceramic fragment found in this level can be classified into the 3rd phase of Coţofen culture. Besides the Coţofen pot the fractions of two rough elaborated, broom decorated dish (pot) were also unearthed, which are certainly dated to the Early Bronze Age.

These ceramic fragments provide a clear starting point for the dating of the $A_{paleo2}$ layer of the mound. According to relative chronology this time was the end of the Late Copper Age and the beginning of the Early Bronze Age (switching period) in the northeastern part of the Carpathian Basin. This is largely confirmed by the reservoir corrected TOC $^{14}C$ soil age of the $A_{paleo}$ layer (original palaeosol) from Lyukas Mound: 2900–2300 cal BC.

The spot of the formal basic burial emerged from the geometrical centre of the mound digging in the $A_{paleo}$ layer. The grave is 140 cm wide and 200 cm long. According to archaeological and anthropological data, a Cro-Magnon type, 23–39-year-old adult of robust physique, whose height was calculated as 176 cm, was buried there. The $^{14}C$ dating from human bones is 4210 ± 35 BP, 2820–2670 cal BC (1σ). The remnants of animal bones unearthed not far from the centre of the mound and the worn household pottery fragments found in numerous fox holes from warren soil are identified as the remains of a Neolithic settlement. Scanty traces of the probably same Neolithic settlement are identified in the southern part of the mound, on a narrow dirt back. It is claimed that the loess ridge in the base of the mound has preserved traces of human cultures which existed prior to the construction of the kurgan.

## 29.6 Conclusions

It is not yet known what causes or antecedents brought to life the pit grave steppe formation in the area of the former Soviet Union. However, the fact that in terms of development of Yamnaya (=pit graves) culture and its environment with its metalwork, in those days it was just as dominant a phenomenon as in Mesopotamia. Interestingly, nomadisation as a lifestyle has survived continuously from very early times to the present day. As part of the migrations started at the end of the Copper Age with the westward movement of peoples herding large livestock, it is assumed that the first immigrant groups arrived into the Carpathian Basin from the south and the second wave of ethnic groups form the directions of Ukraine and Romania. This hypothesis is confirmed by the Coţofen pottery fragments found in the second zone of the kurgan body. Whereas the nomads primarily populated areas suitable for their lifestyle, they also occupied floodplains and saline grazing areas in the Carpathian Basin.

The complex research of tumuli also shows that Holocene climate changes were more subtle and mosaical in the Carpathian Basin during the metal ages, as it had been assumed earlier. Instead of extensive oak or beech forests, there were mosaics of alkaline dry steppe patches, loess and sand ridges, waterlogged floodplains and riparian areas and other wetlands in the area of Great Hungarian Plain during the Holocene.

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# Parabolic Dunes of the Southern Nyírség

30

Tímea Kiss and József Lóki

**Abstract**

A series of alluvial fans extend into the Carpathian Basin from the encircling mountains, and one of the biggest among them is the Nyírség, built by the Tisza River and its tributaries. In the late Pleistocene and Holocene aeolian processes reworked its surface, therefore sand dunes are the characteristic forms today. Approximately one quarter of the area of the Southern Nyírség is occupied by dunes, which form dune fields separated by former river courses. Although previous studies held parabolic dunes as the typical forms, only 6 % of the total dune area covered by filled, partially filled or unfilled parabolic dunes. Most of the dune area (ca 40 %) belongs to the valley-marginal dunes, which are associated with waterlogged paleo-courses and depressions. The rest of the features are transitional between straightened valley-marginal dunes and parabolic dunes. The present dune assemblage developed during the late Pleistocene, although due to human impact the sand was mobilized time to time, and smaller features (e.g. blowouts, hummocks) developed.

**Keywords**

Parabolic dune • Valley-marginal dune • Human impact • Alluvial fan • Nyírség

## 30.1 Introduction

After the regression of Lake Pannon the rivers entering from the mountains to the plains built extensive alluvial fans in the Carpathian Basin. The rivers deposited mostly fine sediments (sand, silt and clay) and frequently changed their courses, thus the alluvial fans gradually rose above the surrounding floodplains. The Nyírség extends over 4,600 km$^2$, having relative altitudes of 20–50 m, and is the second largest alluvial fan of the Great Hungarian Plain. In dry periods the surface of the Nyírség was shaped by aeolian action, which created parabolic dune assemblages—typically on sparsely vegetated sandy surfaces. In the Southern Nyírség characteristic forms of parabolic dunes were formed, not typical in other sand dunes areas of Hungary.

## 30.2 Landscape Evolution

A composite alluvial fan was built by the Tisza and its tributaries in the foreland of the Northeastern Carpathians in the Pleistocene (Sümeghy 1944; Borsy 1961; Püspöki et al. 2013). At the beginning of the Würm (ca 80–70 ka ago) the rivers of the Northeastern Carpathians and Northern Transylvania ran in a north-south or northeast-southwest direction (Fig. 30.1a) towards the subsiding area of the Körös Basin (Berettyó-Körös Plain in Fig. 5.1). The first major change of river courses happened ca 50–45 ka ago, when the Tisza and Szamos Rivers abandoned the alluvial fan of the Nyírség and shifted into the Ér Valley, on the present-day Hungarian-Romanian border (Fig. 30.1b). Later the northern

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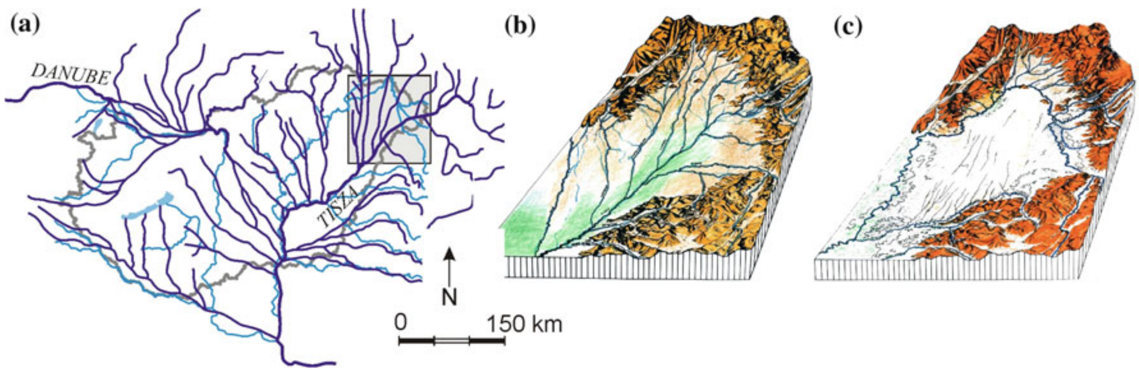


**Fig. 30.1** The alluvial fan of the Nyírség, built by the tributaries of the Tisza River. **a**. Pleistocene drainage network (*in purple*) superimposed upon the present-day network (*in blue*); **b** The river system of the alluvial fan at the beginning of the Pleistocene, when the Ancient Tisza followed the valley of the present-day Ér River (flowing approximately along the Hungarian-Romanian border); **c** The river system in the late Pleniglacial (mid-MIS 2) (*source* Borsy and Félegyházi 1984)

part of the Szatmár Plain (Fig. 30.2) started to subside, and at the same time the southeastern Nyírség and the Ér Valley became (relatively) uplifted. Due to these tectonic movements the Tisza abandoned the Ér Valley and after leaving the Carpathians it turned to northwestern direction (Fig. 30.1c). The exact age of this avulsion is controversial, as according to Borsy et al. (1989; Félegyházi 1998) it happened ca 20 ka ago, but Nádor et al. (2007, 2011) and Gábris and Nádor (2007) dated it to ca 14 ka BP. After the rivers had abandoned the alluvial fan, aeolian activity became

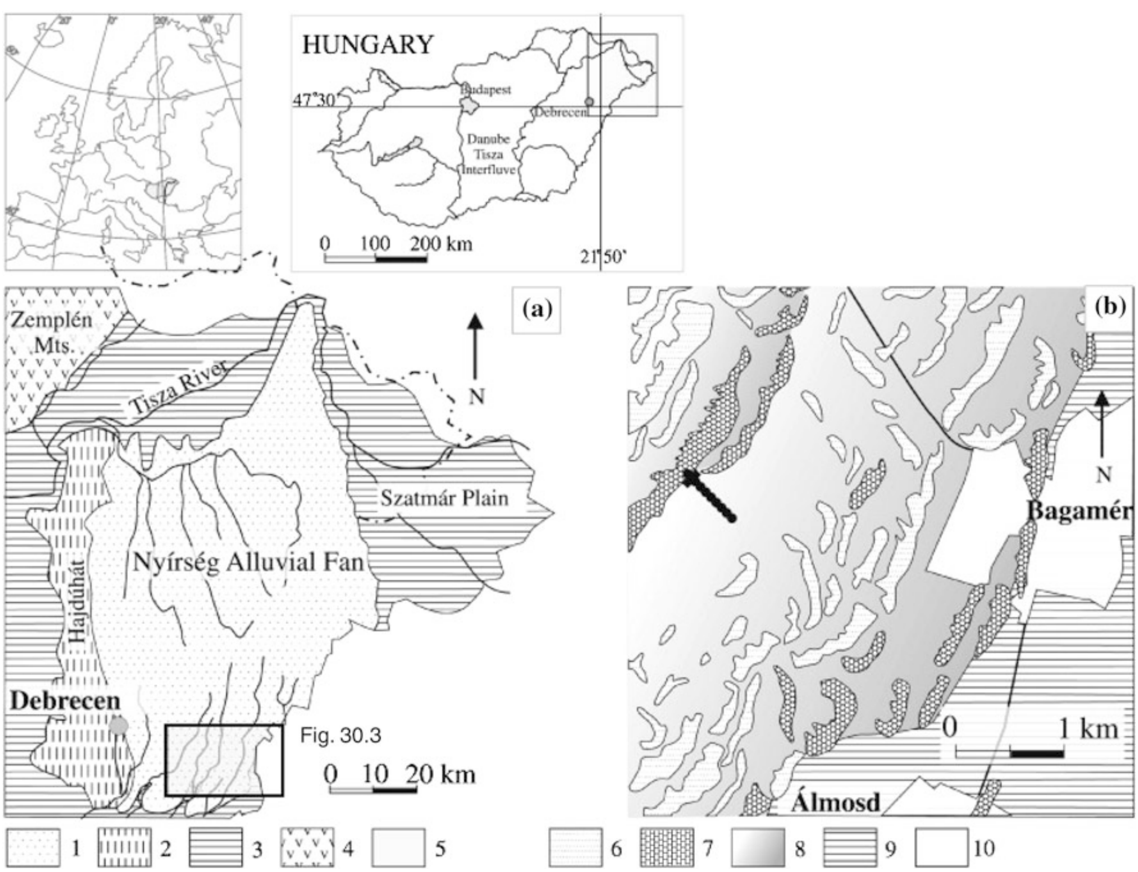


**Fig. 30.2** Location of study area in the Southern Nyírség alluvial fan (Kiss et al. 2012). **a** Topographical regions of Northeast-Hungary. *1* Aeolian sand; *2* Loess; *3* Alluvium; *4* Volcanic rocks; *5* Study area. **b** Landforms of the dune field at Bagamér. *6* Sand dune; *7* Highly eroded sand dune; *8* Paleo-valley with dunes; *9* Alluvial surface; *10* Built-up area

dominant, although some evidence also exists pointing to the co-existence of fluvial and aeolian landforms (Kiss and Bódis 2000). The sand dunes mostly accumulated on fluvial sand or silty sand. In the deeper blowout depressions, between the wings of the parabolic dunes and on the flat sandy areas the blown-sand veneer is quite thin, fluvial sand is very often exposed (Borsy 1961).

While along the rivers the surface was wet, strong western and northern winds could start aeolian processes on the more remote and dry surfaces not protected effectively by the sparse vegetation (Kiss et al. 2012). According to Borsy (1991), the most intensive period of sand movement was between 26 and 22 ka BP. In the southern Nyírség aeolian activity was very intensive until the Bölling Interstadial, when the mild and wet climate and the denser vegetation put to it an end. In the Dryas cold phases the previously developed forms were reshaped (Lóki et al. 1994). By the beginning of the Holocene aeolian features became fixed by vegetation. Under warmer and more humid climate steppe communities covered the dunes and mixed forests occupied the waterlogged interdune areas (Kiss et al. 2012). In the Boreal phase, when the climate turned drier, blown-sand movement resumed (Borsy 1991). In the wet phases of the Holocene the dune surfaces were eroded by surface runoff, and the eroded material was deposited in the interdune areas. However, even during wet periods blown-sand movement could have occurred due to human impact, spatially and temporally adjusted to its intensity.

## 30.3 Description of the Southern Nyírség

The Southern Nyírség is densely covered by sand dunes (2.4 dunes per km$^2$), which cover approximately one quarter (24.4 %) of the area (Kiss 2000). Reflecting the original slope conditions of the alluvial fan the Southern Nyírség slopes are facing towards the southwest. Thus, the north-eastern part is the highest (150 m above sea level), whilst in the southwest elevation is the lowest (ca 100 m). The relative height of the dunes varies, but in general it increases towards the southeast. The dunes form dune fields with the paleo-courses of former rivers between them.

Within a dune field the relative height of the dunes shows a typical pattern, as the highest dunes are always located along the eastern margin of the dune field next to the paleo-channels, and within the whole region the relative height of the dune fields themselves increase toward the east (Fig. 30.3). The reason for this phenomenon is found in the evolutionary history of the dunes. During the phase of aeolian activity the youngest paleochannels were also active or the groundwater table was closer to the surface, therefore, the channels could effectively prevent the parabolic dunes from migrating eastwards. Since their migration stopped, the dunes along paleochannels became ever higher as a result of continuous sand supply from the dune field. In this way the parabolic dunes transformed into valley-marginal dunes (Kiss et al. 2009).

## 30.4 Aeolian Features

The dunes of the area constitute parabolic dune assemblages, parabolic dunes and all other aeolian landforms which develop on wet sand (see David 1977). Most of the dunes were identified by Borsy (1961) as parabolic dunes with undeveloped western wings (Fig. 30.4). He rarely found symmetrical parabolic dunes, but he described the valley-marginal dunes. He also identified small blowouts and larger blowout depressions, hummocks, residual ridges and combinations of different parabolic features. Applying better

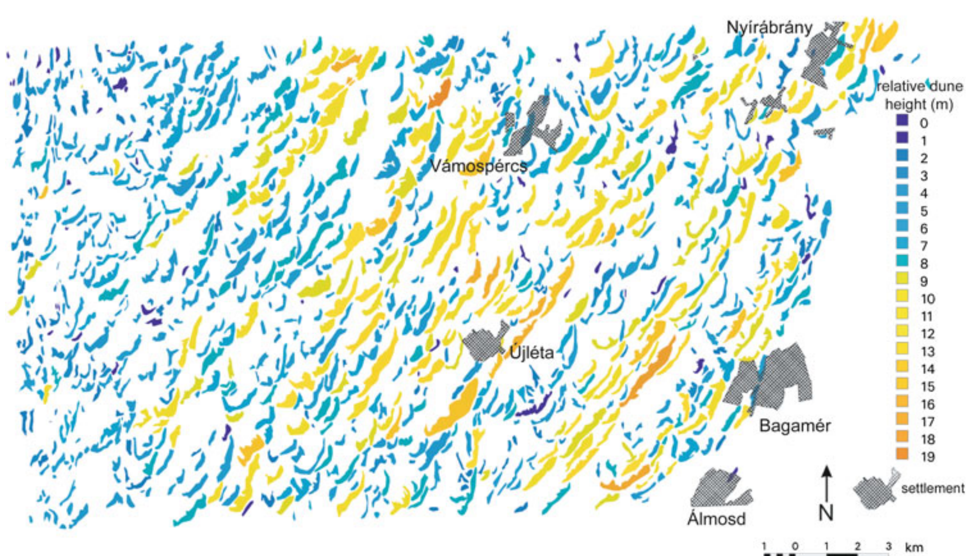


**Fig. 30.3** Increase in the relative height (m) of the dunes of the Southern Nyírség towards the east (Kiss et al. 2009). *Different colours* indicate different dune heights

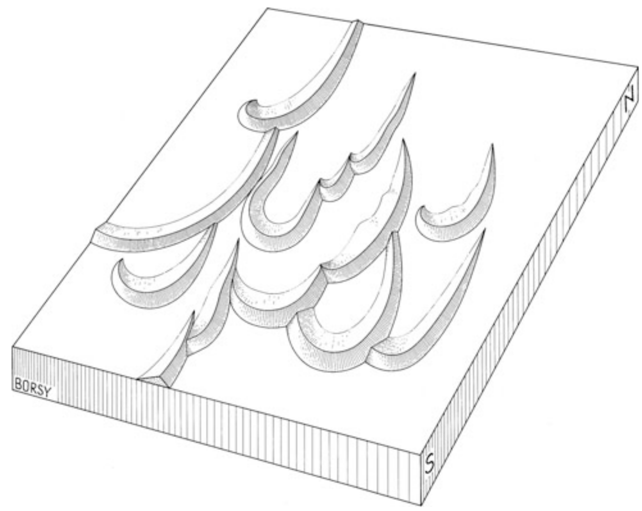


**Fig. 30.4** Asymmetric parabolic dunes described by Borsy (1961)

resolution maps the features are classified, based on their morphology (Fig. 30.5), into filled, partially filled and unfilled parabolic dunes, and valley-marginal and transitional forms are identified too (Kiss et al. 2009).

### 30.4.1 Parabolic Dunes

Parabolic dunes are built of the material removed from blowouts. At first, sand accumulated in hummocks, and during prolonged aeolian activity the hummocks started to migrate and coalesce (Borsy 1961, 1974, 1980, 1991). In the area Borsy (1974) identified parabolic dunes with undeveloped western wings and explained them with (1) the lack of space, i.e. the wings located too close to each other merged, and (2) the strong northwestern winds eroded the wings perpendicular to the wind direction. Although Borsy (1961) described the parabolic dunes as predominant forms of the Southern Nyírség, we found that only 6 % of the total dune area is covered by filled, partially filled or unfilled parabolic dunes (Kiss and Sipos 2007).

Unfilled parabolic dunes with no sand accumulation between the arms originated when the sand supply was limited and the vegetation was sparse. They mostly appear east of the paleochannels and on flat areas where dunes are rare. Their N–S axis—in contrary to the NW–SE axis of other types—indicates that they were probably formed in a period when the prevailing wind direction was different (Kiss and Sipos 2007). As there are no post-genetic blowouts on the unfilled parabolic dunes it shows that they are younger formations or their sandy surfaces were more stable than those on higher and drier grounds.

Between the wings of the filled and partially filled parabolic dunes there are no blowout depressions, but the space was filled up by sand, indicating abundant sand supply during their formation. They appear in groups on the east side of paleo-valleys (western edges of dune-fields). They show a slight asymmetry, as the axial low is located on the lower third of the brink-line. Thus, they were probably formed by bidirectional winds, though the prevailing wind direction was northwestern. Besides abundant sand supply, the formation of these dunes was probably enhanced by sparse vegetation and low groundwater table, allowing the surface to be easily mobilized by winds.

The dimensions of parabolic dunes are quite large, as the length of their arc (along the brink-line) is over 0.9 km. Their height depends on the rate of sand supply, thus the average height of unfilled dunes is 7.2 m (maximum: 11 m), while the filled and partially filled parabolic dunes are higher (average: 8.2 m, maximum: 15 m). Their windward slopes are straight and gentle (4° to 8°), while lee slopes are complex and steep (12° to 25°), the steepest segment being just below the brink-line. Windward slopes are longer, the length proportion of luv to lee sides is 2:1 or 3:1. The original slope morphology of the dunes were altered by surface runoff and accelerated soil erosion caused by human action. Therefore, the early Holocene soils are frequently buried and

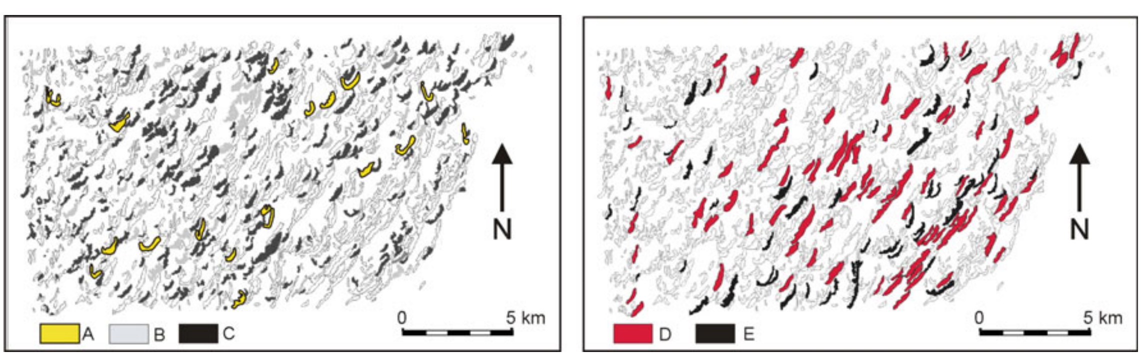


**Fig. 30.5** Dune types of the South Nyírség: unfilled parabolic (*A*), partially or totally filled parabolic (*B*), transitional parabolic (*C*), valley-marginal (*D*) and transitional valley marginal dune (*E*) (Kiss et al. 2009)

the slopes of the dunes are gentler. In other places human impact resulted in resumed aeolian activity, thus blowouts were formed on the windward slopes of the parabolic dunes and hummocks deposited at the crest of the dunes. In these cases back-ridges also appear indicating dense vegetation during the formation of the blowouts. Frequently the WNW–ESE axis of the blowout-hummock complexes do not overlap with the axis of the parabolic dune (NW-SE), indicating different wind directions and different time of origin.

### 30.4.2 Valley-Marginal Dunes

This dune type develops if the southward blown-sand drift is impeded by waterlogged paleo-channels or marshlands. If a parabolic dune reaches these wet surfaces, its eastern wing becomes fixed. However, the head and the western wing moves further south, the parabolic dune straightens into a valley-marginal dune (Fig. 30.6; Kádár 1956; Borsy 1961, 1974). The eastward migration of the straightened dune ceases after reaching the riverbed, but with still abundant sand supply, however, its height is increasing.

Valley-marginal dunes have a characteristic spatial distribution, as they are located on the western margins of the paleochannels of the alluvial fan, and very often they are densely queued behind each other. In the Southern Nyírség most of the valley-marginal dunes are found in the eastern dune-fields, and their number decreases towards the west. This spatial pattern points to the activity of the fluvial system during dune formation: the more active and wetter a channel had been, the most effectively it could stop the migration of dunes (Kiss and Bódis 2000). The valley-marginal dunes are quite typical forms of the Southern Nyírség, as they cover ca. 40 % of the total dune area. They are long forms (brink-line length: 0.9–3.5 km), their average height is 9.3 m (maximum: 18 m). Usually, the valley-marginal dunes along the paleochannels are 1.5 m higher than the group average. Behind the valley-marginal dunes transitional forms between parabolic and valley-marginal dunes could be identified (ca 40 % of the total dune area). These transitional forms are shorter (brink-line length: 0.9–2.0 km) and smaller (average height: 6.8 m; maximum: 17 m). The surface of both forms is dissected by blowout hollows, but their axes differ. Consequently, the blowouts probably developed after the migration of valley-marginal and transitional forms stopped.

### 30.4.3 Other Aeolian Features

Hummocks, residual ridges and abandoned wings appear in large numbers, but have very small dimension. Therefore they occupy just ca 15 % of the total dune area of the Southern Nyírség in sporadic distribution, although are predominant in the western half of the Southern Nyírség. Their presence can be explained with the finer grain size of the sand and intensive human impact in the vicinity of the city of Debrecen. Here their axes are aligned N–S as they were probably formed simultaneously with the unfilled parabolic dunes.

**Fig. 30.6** Valley-marginal dune at Bagamér (photo by Tímea Kiss)

### 30.4.4 Interdune Areas

Most of the Southern Nyírség (75.6 %) is free of sand dunes. These interdune areas have very gentle (≤1°) slopes. They could be classified according to their origin, location compared to the neighbouring dune, and runoff conditions. If interdune areas—as the name suggests—are enclosed by dunes, there is no surface runoff from them. If they are located in front of dunes, they are often remains of former channels, and they have overground runoff.

The enclosed interdune areas appear in the central part of the dune fields. Some of them were deflational (erosional) areas during phases of aeolian activity, thus the sandy material from these surfaces was accumulated in the dunes. Other enclosed interdune areas served as transportation routes where the dunes migrated. Recently, accumulation dominates in interdune areas and the deposits include materials eroded from the dunes and the organic matter of swamps.

The interdune areas in front of dunes are located along the paleochannels of the southern border of the Southern Nyírség. These areas were mostly waterlogged surfaces during the period of aeolian action. Since drifting sand dunes could not cross them, valley-marginal dunes were generated along their western side. From the top of the dunes sand-sheets could extend into these areas, typically in the Late Glacial and—as the result of human impact—also in the Holocene (Félegyházi and Lóki 2006). As most of the interdune areas in front of dunes follow the former paleochannels, they run from NE towards SW and they are often drained by minor watercourses. Therefore, besides the eroded material of the dunes and autochthonous organic material, allochthonous sediment deposited in these areas too. Iron pans (limonite) and vivianite lenses are also found in their sedimentary sequences (Székyné-Fux 1942; Borsy 1961).

## 30.5 Human Impact

The form of the dunes had been altered by human activity at several places (Lóki et al. 2008). The advent of human impact in the South Nyírség region dates back to the Atlantic phase, when the climate was humid and the dunes were stabilized by mixed forests. These environmental conditions were not favourable for wind action. During the Neolithic and Copper Age colonization of the region (ca 6.6–5.4 ka BP), however, aeolian activity reshaped the dunes (Borsy 1990; Kiss et al. 2012). The tribes burnt the forest and practiced slash-and-burn agriculture and animal husbandry. As the dunes lost their native vegetation the sand was mobilized by the wind (Kiss et al. 2012). During the Subboreal phase the climate turned cooler and more humid, thus beech and hornbeam forests were established on the dunes. At this time Bronze Age people also practiced animal husbandry. Overgrazing was typical, therefore the sand became mobilized. In the Subatlantic phase, despite drier climate, oak forests were still common the Southern Nyírség. The Celtic tribes started forest clearances and created ploughfields and pastures (Szabó 1971). Thus, ca 2.2 ka ago the tops of the dunes were mobilized by strong winds. There is also multiple evidence of aeolian activity in the Migration Period and in the early Medieval Times (ca 1.2–1.0-ka ago). From the 14th century on written documents exist on land use: vast forests had survived until the development of the settlement structure in the 12–13th centuries. At that time (ca 840 years ago) due to forest clearances and intensive agriculture aeolian activity was typical on the bare sand surfaces. As population density increased, the cultivated area extended, the forests were replaced by vineyards and pastures. Agricultural activities were the most intensive in the 17–19th centuries, when repeated aeolian erosion was recorded (Borsy 1990; Braun et al. 1992; Kiss et al. 2012). Most human activities resulted in the development of blowouts and small hummocks. Thus, the uniform straight slopes of the dunes were altered and some of them became dissected (Borsy 1991).

## 30.6 Conclusions

Most of the area of the South Nyírség is covered by members of parabolic dune assemblage, i.e. by parabolic and valley-marginal dunes and their transitional forms. Valley-marginal dunes developed where waterlogged surfaces (e.g. abandoned river channels) could effectively stop the migration of parabolic dunes. As NW and N winds prevailed during the main period of aeolian activity, the valley-marginal dunes are always aligned with the western edges of the wet paleochannels or depressions. Symmetrical parabolic dunes are rare, but their spatial distribution reflects that they mostly developed in the transportation zones of dune fields. The shape of the parabolic dunes points to the rate of sand supply during their formation.

Though more and more data are accumulated on the landform development in the region, it still not determined exactly whether there is any difference in the age of development of the different dune types—especially if we consider that some show the effect of a different wind direction. The spatial distribution of aeolian activity during the Quaternary climatic phases is also unknown.

The aeolian landforms themselves are not under protection, but nature reserves in the region aim at the protection of living organisms and their habitats, mostly in the interdune wetlands. The dunes are sensitive landforms: if they lose their vegetation, aeolian activity could resume in dry years, endangering agricultural areas and settlements.

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# Mártély Lake: An Oxbow of the Lower Tisza River

31

Tímea Kiss and György Sipos

**Abstract**

Oxbows are common elements of fluvial landscapes in Hungary. The aim of this paper is to introduce their origin, development and future perspectives. Oxbows have been formed either naturally or artificially. Natural oxbows, or rather paleo-channels have silted up by now, but have got a key importance in the reconstruction of Late Pleistocene and Holocene landscape evolution and natural floodplain aggradation. Man-made oxbows, resulted from cutoffs during the regulation works of the 19th century, are on the other hand experience recent environmental and land use changes, threatening their future sustainability. Problems and processes affecting them highly depend on their location with respect to the post-regulation active floodplain and artificial levees. Main issues are water recharge and retention, increasing sedimentation, spread of invasive species, improper landscape management and conflicting utilization interests. The exemplary Mártély Lake, an oxbow of the Tisza River, is one of the largest such forms in Hungary. Being on the active floodplain it has a great ecological potential, but meanwhile it is seriously affected by silting up and also has a diverse utilisation with conflicting interests. In order to sustain or even improve its status a complex management strategy has to be implemented in the future. This is true for other oxbows as well, being highly sensitive but at the same time extremely valuable elements of the Hungarian landscape.

**Keywords**

Floodplain • River engineering • Oxbow lakes • Sedimentation • Land-use management • Tisza River

## 31.1 Introduction

As the lowlands of Hungary have been primarily formed by rivers both in the past and present, oxbow lakes are common elements of the landscape. Numerous meanders and palaeochannels have been left behind by the actively migrating alluvial rivers, such as the Tisza, Danube or Hernád (according to Blanka 2010, for instance, 10 natural cutoffs occurred on the Hernád in the past decades). In the 19th and 20th centuries human interventions leading to artificial cutoffs have become the key processes behind oxbow formation. Let they be naturally or artificially developed, oxbows are very important landmarks of the alluvial landscape. Most of them are situated along the highly engineered Tisza and Körös Rivers, but practically they can be found anywhere on the plains. Their total number is estimated to be around 500 (Molnár 2013).

Unfortunately even those formed recently have silted up in the past centuries and started to disappear. Consequently, most of these lakes and marshlands are under strict protection and not just because of their geomorphological and hydrological importance, but also because they provide high-diversity refuges and important corridors for the

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continuously shrinking natural flora and fauna. As the area of wetlands in the Carpathian Basin decreased dramatically as a result of intensive river regulation and drainage works during the 19th to 20th centuries (from 57 to 2 %, Gábris et al. 2004), oxbows are almost the only still existing witnesses of the once flourishing floodplain ecosystems. Oxbows are very sensitive to climate change and intensified human impact, thus the area of their open water surfaces is decreasing, and at the same time their water quality is also deteriorating. In order to preserve these landforms for the future several problems need to be tackled to maintain their hydrology and water quality and to prevent further siltation and disturbance (for instance, through the spreading of invasive species). A well-designed management would also serve economic interests, since oxbows are significant water reservoirs, and can be used for water retention, irrigation or, in special cases, to extract drinking water. Their use for angling, fishing and summer tourism is also increasing.

## 31.2 Environmental Background

### 31.2.1 Natural Cutoffs

It is a well-known feature of meandering rivers that they continuously develop their channels and leave behind overmatured bends. A natural cutoff will occur when sinuosity exceeds a threshold value where at the given slope and stream power conditions the river cannot maintain its meander further (Hooke 2004). A natural cutoff can develop in two ways. If the river finds its shorter track along point bars or on the floodplain, a chute cutoff, the more common type according to Knighton (1998), occurs. However, on the Tisza River and its tributaries neck cutoffs are more characteristic. In this case two downstream migrating meanders in the same phase get so close to each other that during an erosive, high-energy event (flood) the neck of the enclosed bend is broken through, and its limbs are blocked by the sediments of the rapidly developing natural levee.

### 31.2.2 Artificial Cutoffs—Regulation Works on the Tisza River

Prior to the 19th century regulations the rivers of the Hungarian Great Plain were highly sinuous and their channel slopes were very low. Therefore, floods inundated vast, potentially arable lands for 5–6 months in almost every year. Rivers also functioned as the main routes of commerce, since boats provided practically the only means of transportation in the lowland, covered by extensive swamps and marshlands. Therefore, the need for flood control and safe navigation facilitated the elaboration of regulation plans in the beginning of the 19th century, and by the end of the century river training works were more or less completed.

One of the most important aims of these regulations was to increase slope and the rate at which flood waves pass. This was achieved through making numerous artificial cutoffs (Fig. 31.1). Cutoffs were actually narrow conductor channels made usually at the neck of meanders, while the excavated material was deposited 8–10 m away from the new banks. When the river was captured by the cutoff

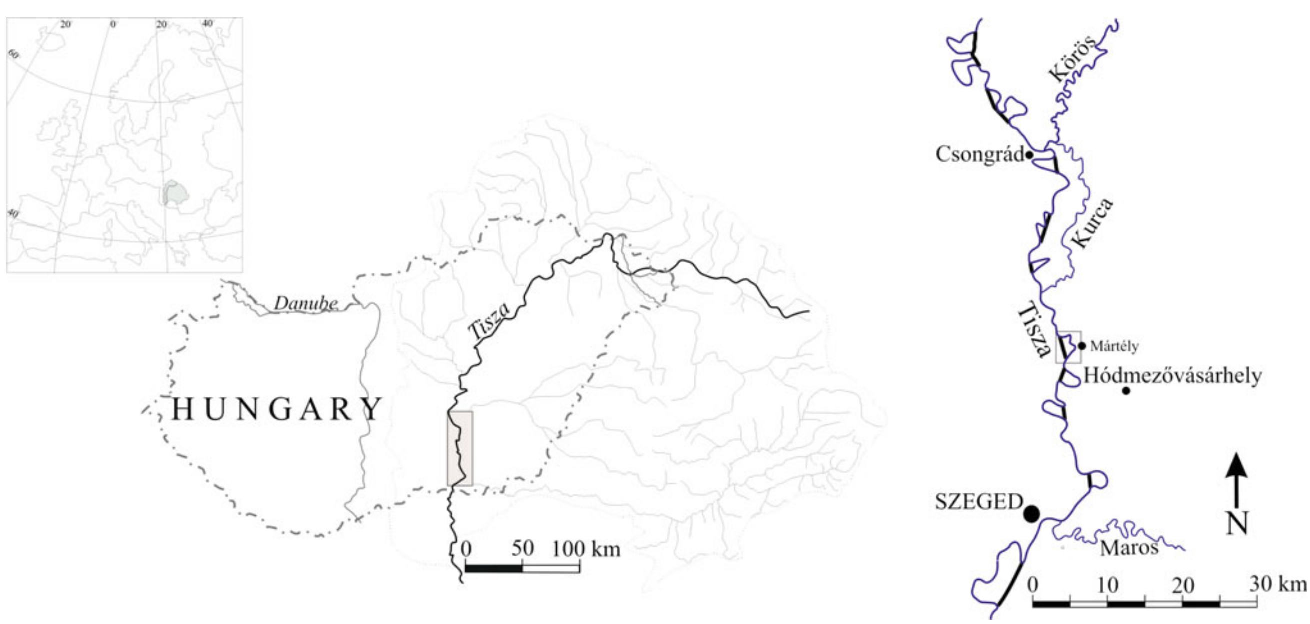


**Fig. 31.1** Location of the Tisza catchment and the exemplary Mártély Oxbow Lake

channel, it could develop its new cross-sectional geometry in accordance with its increased slope and energy (Ihrig 1973). This way the cutoff channel could naturally turn into the main channel of the river, saving a considerable amount of effort for engineers. However, the procedure was sometimes more complicated, especially at longer cutoffs and on the lower sections of rivers (Fig. 31.1), where the main channel was embedded in clayey-silty sediments. In these cases cutoff channels had to be deepened and widened, and had to be dredged from time to time to make the river finally occupy its new course (Károlyi and Nemes 1975; Lászlóffy 1982).

The first cutoff, ending with successful diversion and fixation of the main channel, was finished in 1846. The excavation of cutoff channels was usually completed quickly, e.g. on the Middle Tisza in less than 20 years, however, it took a much longer time to capture the thalweg of the river (Iványi 1948). It has to be emphasized though that all works were made by using only human power, no machines being available at that time (Dunka et al. 1996).

Along the Tisza River 114 meanders were cut off, shortening the river course from 1,419 to 966 km, and increasing its slope from 3.7 cm km$^{-1}$ (0.000037) to 6 cm km$^{-1}$ (0.000060). In total approximately 1,000 cutoffs were implemented on Hungarian rivers (Somogyi 2000).

In general the slope of rivers doubled, which initiated a series of geomorphic processes, though responses were different. Energy and slope increase usually resulted in incision, channel widening, increased sediment production and in certain cases pattern change. For example, in case of the meandering and anastomosing Maros River, the largest tributary of the Tisza, the whole process could be identified, and the river turned to be braided (Kiss and Sipos 2007). In the meantime the Tisza experienced a 3–5 m incision (Kiss et al. 2008), which resulted a 300–400 cm decrease in the absolute level of low waters (Rakonczai 2000) and the sinking of groundwater level along the river. Consequently, oxbows became relatively elevated, and only the greatest floods could recharge their water naturally, thus open water surfaces can only be preserved by human intervention.

Enhanced floodplain aggradation was another direct and also indirect outcome of cutoffs, which necessarily lead to the silting-up of oxbows as well. During the capturing of thalwegs by cutoff channels extra sediment entered the river systems directly. Subsequent incision and related bank failures and slides still supply further material to the channels from time to time (Kiss et al. 2008). These processes also lead to intensive sedimentation (1.5–2.0 m) on the narrow, artificial floodplain bordered by levees constructed for flood control purposes in the 19th century. The process is unfavourable not just for oxbows and geomorphological diversity but also from the aspect of increasing flood levels and flood risk (Lóczy and Kiss 2009).

## 31.3 Research History

The investigation of oxbows and palaeochannels is an important field of Hungarian geomorphological research. During the geomorphological mapping of the Tisza-Körös confluence zone, with numerous oxbow lakes, Schweitzer (2006) has identified several types based on the degree of sedimentation. A similar mapping was prepared along the Middle Tisza (near Vezseny) at 1:10,000 scale by Balogh et al. (2005), however actively developing forms (e.g. present-day point bars) were not indicated. For the Middle Tisza Region Tóth et al. (2001) had shown the possibility of mapping and classification of oxbows, also emphasizing the necessity of landscape rehabilitation and water retention.

The geomorphological mapping and absolute dating of channels on the now inactive floodplain also provides an opportunity to reconstruct the evolution of alluvial rivers. Analyses of this kind have already been made on the Sajó-Hernád (Nagy and Félegyházi 2001), Hortobágy (Félegyházi and Tóth 2003) and Maros (Katona et al. 2012; Kiss et al. 2014) alluvial fans, and along the Körös (Nádor et al. 2011) and the Middle Tisza Rivers (Gábris et al. 2001).

In the Upper Tisza Region detailed analyses of Pleistocene and Holocene palaeochannels revealed not only the pattern of landform evolution, but also the rate and timing of floodplain and oxbow sedimentation. For instance, near the Tisza-Bodrog confluence channels are silting up significantly faster (1 mm year$^{-1}$) than general floodplain aggradation (Borsy et al. 1989). However, there was a significant variation in the rate of sedimentation, being quite low during the Late Glacial and Preboreal Phase (0.2–0.3 mm year$^{-1}$), getting faster during the Atlantic Phase (1–2 mm year$^{-1}$) and lower again during the Subboreal Phase (0.8 mm year$^{-1}$) (Csongor et al. 1982). Based on palynological and radiocarbon data the palaeochannels on the Hernád floodplain silted up at a similar rate (0.4–0.5 mm year$^{-1}$) in the Subboreal Phase. However, during the past 2,000 years sedimentation increased (to 1 mm year$^{-1}$) and accelerated further in the past 300 years (8 mm year$^{-1}$) (Szabó 1996).

Depending on their location, the oxbow lakes which resulted from regulation works developed individually. Somogyi (2000) described those beyond levees as living water lakes of different status, while those situated between levees as forms completely silted up by the sediments of post-regulation floods. Although the later remark is not generally applicable, there are spectacular examples, for instance, along the Maros River, which transports a considerable amount of suspended load and has filled up all oxbows along its course by now (Kiss et al. 2011).

The sedimentation rate of Tisza River oxbows was investigated by Braun et al. (2000, 2003), using $^{137}$Cs and heavy metal markers. They found a 2–6 cm year$^{-1}$ accumulation on the average, though for instance in the case of a representative Upper Tisza oxbow, experiencing a 400 cm accumulation (ca 3 cm year$^{-1}$) since its cutoff in 1860, the rate of silting up was decreasing from 5 cm year$^{-1}$ (from the 1920s till the 1970s) to 2 cm year$^{-1}$ through time (Braun et al. 2000).

The pollen of adventive species (e.g. ragweed, *Ambrosia artemisiifolia*) were applied by Kiss et al. (2011) to study the sedimentation rate of Maros River oxbows. The oxbows located on the artificial floodplain silted up rapidly, at a rate of 1.3–2.6 cm year$^{-1}$, and water vanished from them within 50–70 years following cutoff. The analysis of several forms indicated that the rate of sedimentation was uneven in time and it was affected by several factors (Kiss et al. 2011). For instance, an increasing accumulation rate (from 2.5 between 1842 and 1960 to 3.5 cm year$^{-1}$) was detected in a representative oxbow as a consequence of longer inundation in the 1970s. The sedimentation rates in oxbows were primarily controlled by their location relative to the alluvial fan and their distance from the active river channel.

## 31.4 Classification

Based on the above, oxbows can be classified in four ways: by origin, location, degree of degradation and utilisation. As we have seen above, oxbows can either result from natural or artificial cutoff. From the aspect of water management and conservation, however, more recent artificial oxbows are more important, as many of them still have a permanent open water surface (Molnár 2013). Concerning their location the most important types are those located on the active floodplain and those beyond the flood-control levees.

As it was shown earlier, location primarily affects the degree of sedimentation and degradation. Water managers and conservation specialists identify three types of oxbows in this respect (Pálfai 2001). So-called "sanctuary" oxbows are resembling natural ecosystems. They are not under human use and have not silted up. These are usually under strict protection and managed by national parks. Oxbows of "wise utilization" are lakes with a certain economic use, slightly degraded, but their different uses can be harmonized. The third group consists of highly degraded oxbows, usually of minor natural value or silted up almost completely.

In general there are four main types of human use, which are the following according to Pálfai (2001). Use for water management purposes includes flood or excess water storage, drinking, irrigation and industrial water storage, or water quality improvement. Production-related uses are fishing, fowl breeding and reed growing. Recreational uses include bathing, tourism, water sports and angling. Finally, the fourth type of utilisation is in relation with nature and landscape conservation. Most of the lakes are naturally under a mixed use, which generates several land-use conflicts between different stakeholders.

## 31.5 The Oxbow of Mártély

The Mártély Oxbow was cut off from the main channel of the Tisza River between 1889 and 1892 (Fig. 31.2). The length of the Mártély Oxbow is 4.6 km, its average width is 100 m, its area is 46 hectares, from which 33.5 hectares are open water (Fig. 31.2). Average depth is 2 m, though at places it can be as deep as 6.5 m (Fig. 31.2). The oxbow is connected to the Tisza at its downstream end with a feeder canal and a lock (Pálfai 2001). Nevertheless, due to the incision of the Tisza, natural water supply is limited to flood periods. At lower stages water can only be recharged by pumping. The water of the lake is partly used for irrigation, the outlet is situated near the midpoint of the oxbow. Artificial pumping and simultaneous draining ensures at least some water circulation, though affecting only the southern limb of the oxbow, the northern limb lacks oxygen and has gradually turned into a swamp (Fig. 31.3). Due to dredging in 2003, however, water quality has improved considerably (Molnár 2013).

Although during the regulation works a localisation dam was constructed along the bank of the oxbow, the final levees were built on a different track, resulting in a fairly wide active floodplain (Fig. 31.2). Riparian forests and meadows are under nature conservation (protected landscape) and the oxbow itself is a Ramsar site (Fig. 31.4). The lake therefore has a mixed use. The main conflict is related to recreational use, since for over 100 years a bathing place is situated on the eastern shore of the oxbow, and an 18-hectare resort village has been growing around it (Molnár 2013). At present ecotourism is facilitated by a new visitor centre and several hiking and educational trails.

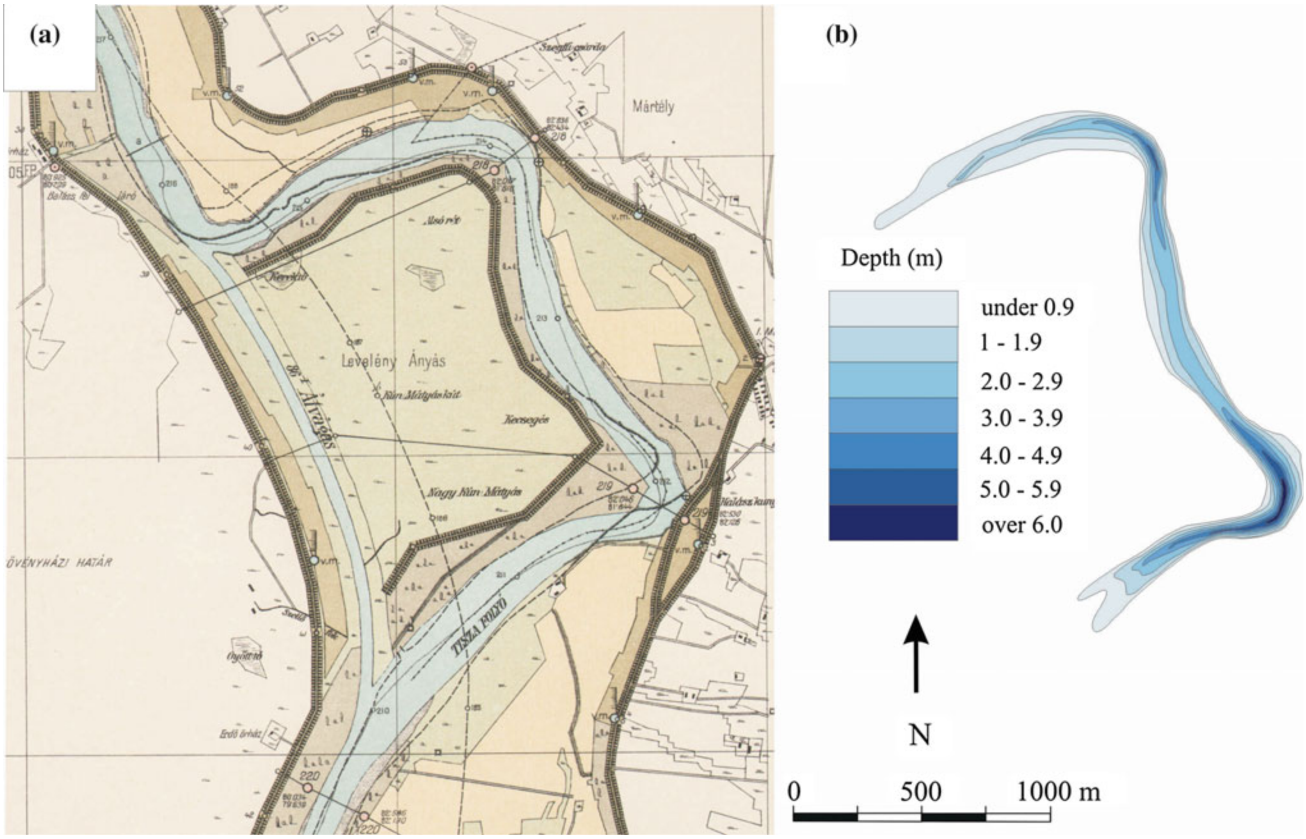


**Fig. 31.2** **a** The Mártély Oxbow Lake during the regulation works. The cutoff is completed, however the conductor canal has not been captured by the main channel (*source* Tisza Regulation Map Series). **b** Bathymetric map of the Mártély Oxbow Lake (*source* Bártfai 2011)

**Fig. 31.3** Both ends of the Mártély Oxbow Lake are filled up by now

**Fig. 31.4** A newly built board trail over the swamp

## 31.6 Conclusions

The oxbows of the Great Hungarian Plain, and especially those of the Tisza River, have exceptional natural and geomorphological values. They preserve something from the pre-regulation character of the floodplain both in terms of ecology and geomorphological processes. Their conservation is a difficult task, as they are seriously affected either by climate change, sedimentation and/or human use.

Future research should focus on factors determining the sustainability of these lacustrine and wetland systems. A key question in this respect is water recharge or water retention, which is most problematic for oxbows beyond the levees. The preservation of oxbows would also increase the resistance of landscape to climate change. Retention, however, also imposes water quality issues, becoming critical in the future.

The long-term dynamics of sedimentation varies with time and space and mostly affects oxbows on the active floodplain. To reconstruct the general pattern of changes further research is necessary, along with monitoring of present-day sedimentation. These investigations are of key importance for rehabilitation and conservation, and to determine, for example, the necessary extent of dredging.

Another very important sphere where earth sciences can address the management of oxbows is land-use mapping and related conflict and risk assessment. Over the past century land use around oxbows and in the floodplain has changed considerably. Main issues on the active floodplain are the lack of land management and the disappearance of traditional land-use techniques, which lead to the advance of adventive species and alteration of biogeomorphological processes. Meanwhile, oxbows beyond the levees have to face the effects of intensive agriculture, manifested in increased pollution and modified ecology. Conflicts, as seen in the case of the Mártély Oxbow Lake, are more profound if there are several interests of utilization, motivated by recreation, nature conservation or water management.

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# Part III

# Geoheritage

# Geoheritage, Geoconservation, Geomorphosites in Hungary

# 32

Gergely Horváth and Dénes Lóczy

**Abstract**

In addition to geological sites of scientific interest, which appear on the list of key sections compiled by experts of the Hungarian Geological Institute, geomorphosites constitute an important group of protected earth sciences spots. In Hungary the Act on Nature Conservation of 1996 ensures the protection of objects of scientific, cultural, aesthetic, educational, economic or other public interest in several categories and at different hierarchical levels (National Parks, Protected Landscapes, Nature Reserves and Natural Monuments of local importance). UNESCO Geoparks represent an additional category, which serve above all the aims of regional development and intend to increase public awareness to the environment and earth sciences. For most of the geomorphosites described in the present volume nature trails have been established. The foremost routes advisable for people who intend to visit geomorphosites are tourist trails, of which the Countrywide Blue Tour and the Rockenbauer Blue Tour are the most important.

**Keywords**

Nature conservation • National parks • Protected landscapes • Nature reserves • Natural monuments • Geoparks • Nature trails • Itineraries

## 32.1 Introduction

Parallel to early efforts aiming to protect wildlife, the conservation of geological (including geomorphological) monuments dates back to the mid-19th century. As early as in 1839 a local regulation prohibited the destruction and collection of dripstones in Baradla Cave. Later geologists (primarily of the Royal Geological Institute of Hungary, established in 1869) significantly contributed to the elaboration of the concept of nature conservation and its implementation. In 1974 the Geological Institute of Hungary proposed the following classification for items of geoheritage:

- (geological) landscapes of outstanding scenic value;
- caves;
- exposures reflecting relationships between the human and geological environment;
- geomorphological objects;
- springs unique for their chemical composition or historical character;
- exposures important for scientific research at a national or international level, i.e. key sections of geological formations, exposures of outstanding value for palaeontology or stratigraphy, mineralogical, petrographical and/or sedimentological sites.

A major step in the protection of geological sites was the identification of key geological sections (Császár 1997; Trunkó 2000), a programme that included the survey of the

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most typical occurrences of various stratigraphical and lithological formations. The introduction of the concept of geomorphosites (Panizza 2001) gave an impetus to the inclusion of landforms into geoheritage. Panizza (2001) defines geomorphosites as "landforms which acquired scientific, cultural/historical, aesthetic and/or even social/economic values due to human perception or exploitation". Geomorphosites are either single objects (e.g. rock formations, tumuli, oxbows) or broader landscapes (e.g. plains, hills, mountains) (Reynard et al. 2009). Against this theoretical background, the recently published guides and the present volume provide map representations with illustrations and brief descriptions in popular style (Budai et al. 2002; Budai and Gyalog 2010; Palotai 2010).

## 32.2 Nature Conservation

In Hungary the Act on Nature Conservation (Act 53/1996), in force since 1996, provides the legal basis for protection of botanical, zoological, geological, hydrological, scenic and cultural historical assets of outstanding importance, to sustain biodiversity and the sound functioning of natural systems, promote education, scientific research and recreation (Bolner-Takács et al. 2012). As far as geoheritage is concerned, the law ensures protection for objects which are of scientific, cultural, aesthetic, educational, economic or other public interest, including

- geological formations and key sections, mineral associations and fossils;
- major localities of rare minerals or fossils;
- natural landforms and the terrain above caves;
- man-made landforms (tumuli, earthworks);
- typical and rare soil profiles.

In addition, all caves with longitudinal axis longer than 2 m are *ex lege* protected (Székely 1989, 2003). Springs with discharges exceeding 5 L $min^{-1}$ and in karst regions dolines which are hydrologically active (at least temporarily) also qualify for protection. In the listed categories altogether 184 objects were found to be eligible for protection.

Within the national Key-Section Programme, launched in 1980 and implemented within 10 years, 265 surface exposures and some 600 boreholes throughout the country were systematically selected, studied and documented by the Geological Institute and the Stratigraphical Commission of Hungary (Nagy et al. 1987, 1989; Természetvédelem 2013). Finally the Commission identified 485 surface key sections of outstanding scientific significance—some of them are closely associated with landforms (for instance, the Tar Dacite Tuff Formation in the Bükk Foothills—see Chap. 24).

The protection of geomorphological objects is organised at different hierarchical levels, mostly in harmony with those defined by the International Union for the Conservation of Nature (IUCN—Dudley 2008): in National Parks, Protected Landscapes, Nature Reserves and Natural Monuments of local importance (Fig. 32.1). In 2013 the 211 protected areas of national importance in all categories covered 846,752 ha and those of local importance 51,306 ha—altogether ca 9.6 % of the territory.

### 32.2.1 National Parks

Geomorphosites are seldom protected at the highest level of the IUCN categories (Dudley 2008), although critically endangered sites may be grouped with Strict Nature Reserves or Wilderness Areas in other parts of the Earth. Being on the top of the Hungarian hierarchy, National Parks are defined as extensive areas where the natural environment has not been significantly altered (Act 53/1996). According to the IUCN definition, they are large (near) natural areas where species and ecosystems are protected, but also serve as the foundation for environmentally and culturally compatible spiritual, scientific, educational, recreational and visitor opportunities (Dudley 2008).

In at least four of the 10 National Parks in Hungary geomorphosites of outstanding significance are found (Bolner-Takács et al. 2012). The caves of Aggtelek National Park (see Chap. 20) were included in the World Heritage List in 1995. The Balaton Highland National Park is also rich in geological sites of European significance out of which the Tapolca and Kál Basins and the Tihany Peninsula are presented in this volume. The Tihany Peninsula was awarded the Europe Diploma in 2003. Bükk National Park is famed for its spectacular karst features (Chap. 22); whereas the Danube–Ipoly National Park holds a complex of karst and volcanic formations, including the calderas of the Danube Bend (Chap. 16). In addition, the sand dunes and saline lakes in the Kiskunság National Park or the active salinization processes in Hortobágy National Park (Chap. 27), the first in Hungary, established in 1973, are also worth mentioning. The geological environment, including geomorphosites, are organic parts of the unique ecosystems of these Parks.

### 32.2.2 Protected Landscapes

In extensive areas particularly rich in natural values and landscapes of special ecological, scenic, cultural and natural character marked man/nature interactions are incorporated and need protection. Among the 38 existing Protected Landscapes, Ság Hill, on the margin of the Little Hungarian Plain, exposing the interior of a 5-Ma-old volcano, was established with the exclusive purpose of conservation of its geological assets; but geomorphosites occur in sixteen other Protected Landscapes. These include, for instance, weathering features

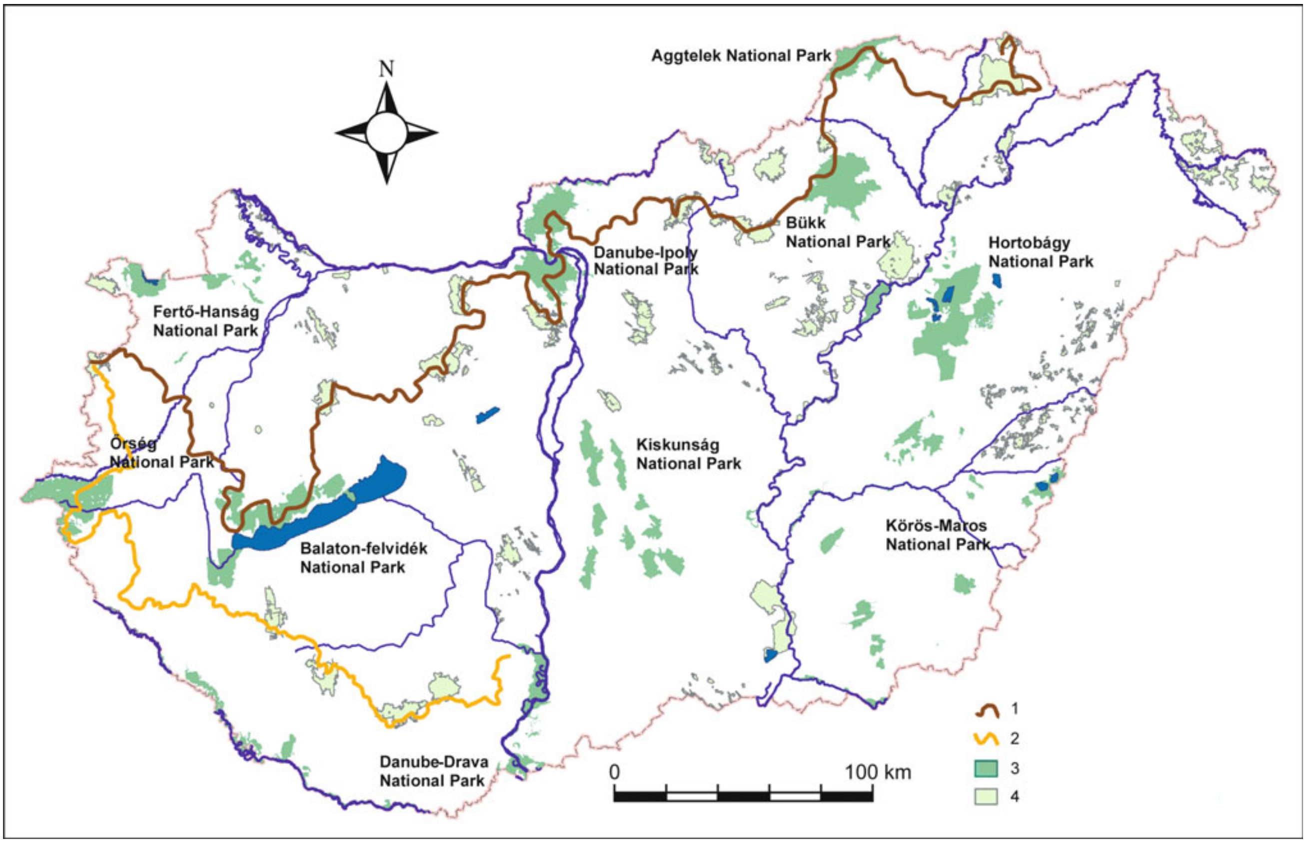


**Fig. 32.1** Nature conservation in Hungary (*Source* Nature Conservation Information System, Ministry of Rural Development, 2014) and itineraries for geomorphosites. *1* Countrywide Blue Tour; *2* Rockenbauer Pál Blue Tour of Southern Transdanubia; *3* National Parks; *4* Protected Landscapes

in the Kőszeg Mountains (Chap. 6); karst in the High-Bakony (Chap. 7); limestone formations and caves in the Gerecse and Buda Mountains; dolomite formations in the Vértes; a variety of karst features in the Western Mecsek; volcanic and man-made landforms in the Karancs–Medves (Chap. 17), Mátra, and Zemplén areas; a limestone gorge at the Lázbérc reservoir (Chap. 21); features of sandstone denudation in the Vajdavár Hills in the Tarna Region (Chap. 19); loess erosion features in the Tokaj-Bodrog-zug (Chap. 25) and oxbows in the Mártély Protected Landscapes (Chap. 31).

### 32.2.3 Nature Reserves

Smaller areas rich in outstanding natural assets (often the habitats of endangered plant or animal species) worth of protection are Nature Reserves. Most of the 162 Nature Reserves of national importance in Hungary have been established for their flora and/or fauna (some of them Ramsar sites—Tardy 2007). There are 15 sites declared as protected specifically for geological (geomorphological) values: stratigraphical (6) and palaeontological (4) sites, terrains above caves (3) and groups of spectacular cliffs (2). The fossil site at Ipolytarnóc (awarded with the Europe Diploma in 1995) and the hominid finds at Rudabánya and Vértesszőlős are well known beyond the borders of Hungary, too.

Most of the beehive rocks of the Bükk Foothills (Chap. 24) are located in the Szokolya Nature Reserve, the balanced rocks of Pákozd (Velence Hills—Chap. 11) and the travertine mound of Egerszalók (Chap. 23) also constitute Nature Reserves of limited areal extension.

### 32.2.4 Natural Monuments

The specific Natural Monuments are of very limited spatial extension, most often arboretums or parks, but sometimes also landforms or caves. These protected sites of local significance are often found within the built-up areas of settlements (Tardy 1996). For instance, the unstable bluff along the Danube is within the built-up area of the village Dunaszekcső (Chap. 14). The balanced rocks around Sukoró in the Velence Hills (Chap. 11) and some beehive rocks in the

Bükk Foothills (Chap. 24) also fall into the category of Natural Monument of local significance.

## 32.3 Local Protection Measures

Some local monuments of geomorphological interest do not enjoy protection. As part of the efforts to encourage tourism, however, some local authorities pay increasing attention to the maintenance and development of potential touristic attractions.

In some cases conservation is restricted to minimal measures. The "Hungarian Pamukkale", the travertine mound (see Chap. 23) is now incorporated into a large hotel complex serving thermal spa tourism. This way the trampling, detachment and contamination of travertine precipitations are precluded. The original physical environment of the formation, however, is lost for ever.

Another good example of careful management is the Crater Lake of Nagyhegyes (Chap. 28), a unique site, which, after the explosion, consolidated into a peaceful oasis in the middle of agricultural land. It has no legal protection as a natural asset. Nonetheless, in 2006 landscaping took place around the lake: two trails were constructed, one on the top of the rampart and a nature trail half way down to the shore. The information tables show the origin of the lake and its wildlife. In addition, picnic sites and angling spots were established to take advantage of the fish fauna, represented by the species common in Hungarian fish-ponds (carps, catfish, bream). Water quality is preserved by the filtering function of the reed-beds which almost completely surround the lake.

Different management is required for other geomorphosites which are also without protection. In the asymmetric Kapos Valley (Chap. 13) the maintenance and partial restoration of traditional land use (with the introduction of new land uses compatible with the floodplain environment, such as energy crop plantations) are sufficient measures. Other sites present immediate environmental hazard and call for mitigation measures. The Somogybabod gully (Chap. 12) has to be consolidated or even obliterated to prevent it from functioning as a route of channelized runoff of dangerous discharge. At Dunaszekcső the collapsing bluff (Chap. 14) has to be stabilized through soft engineering measures (draining surplus waters, establishing protecting vegetation cover etc.).

## 32.4 Nature Trails

In 1984 the first nature trail of Hungary was established in the Tihany Peninsula (Chap. 9), where sites of geomorphological interest also occur. The first geological trail was created on Ság Hill, on the margin of the Little Plain, 5 years later. Among the currently existing 147 nature trails, 43 display exclusively or mainly geological, 11 geomorphological and 5 pedological features. Further trails also include stops where earth science information is provided. In full, according to Budai and Gyalog (2010) there are 95 noteworthy trails of geoheritage in Hungary, the most relevant to the topic of this volume are

- Kuruc Ring, Badacsony (4 km long, featuring basalt forms: sack-like boulders, scree—Chap. 8);
- Boroszlán, Bakonybél (7 km, limestone, karstic forms, cave);
- Road of Giants, Bozsok (7 km, rock-forms in metamorphic rocks, the hoodoo Hat Rock—see Chap. 6);
- Ság Hill, Celldömölk (2 km, basalt quarrying, "core of a volcano", volcanic forms and history of volcanic activity —see Fig. 8.4);
- Gaja Gorge, Bodajk (3.5 km, rock niches);
- Jakab Hill, Cserkút (6 km, "Puppet Rocks", spectacular sandy-pebbly rock formations);
- Bauxite Geological Park, Gánt (1 km, abandoned bauxite mine);
- Borókás Ditch, Ipolytarnóc (1 km, sandstone covered by rhyolite ignimbrite; ancient footprints and fossils, petrified tree trunk);
- Theodora Wells, Kékkút (8 km, mineral wells, blockfield —see Chap. 10);
- Monolith Park at Hegyestű, Monoszló (0.5 km, abandoned basalt quarry, columnar basalt);
- Balanced Rock, Pákozd (10 km, granite weathering—Chap. 11);
- Eresztvény Quarries, Salgótarján-Salgóbánya (3 km, abandoned basalt quarries—Chap. 17);
- Witch Rock, Salgótarján-Salgóbánya (0.5 km, basalt lava, scoria—Chap. 17);
- Szilvás-kő, Salgótarján-Rónabánya (4 km, columnar basalt, deep chasms and pseudocaves of a collapsed basalt surface caused by undermining—Chap. 17);
- Sámsonháza exposure (3 km, abandoned quarry, andesite lava and pyroclasts, covered by limestone);
- Millstone Trail, Sárospatak (12 km, silicified rhyolite tuff, former millstone production—Chap. 26);
- Szalajka Valley, Szilvásvárad (4 km, karst springs, cave, travertine features, waterfall—Fig. 22.9);
- Beehive Rocks, Szomolya (4 km, rhyolite tuff, postvolcanic silicification—Chap. 24);
- Calvary Hill, Tata (1 km, different limestones, fossils, cave, chert pit);
- Lóczy Geyser Mound, Tihany (18 km,—thermal spring cones, basalt tuff—Chap. 9);
- Amber Trail, Tokaj (6 km, andesites);
- Csárda Hill, Úrkút (0.5 km, paleokarst, former manganese pit).

These nature trails either have information boards along the trail, digital information on internet or only posts with numbers. In the latter case the information corresponding to the given number can be read in an accompanying leaflet. The main problem in the dissemination of knowledge is, as always, how to draw up a text which is equally satisfactory for both scientific and popular demands. Other problem is vandalism; information tables must be often replaced. Notwithstanding, the number of nature trails is quickly increasing. In addition to the work of National Park directorates, the new developments are mostly due to the local governments of cities and villages.

## 32.5 Geoparks

Geoparks are unified areas with geological heritage of international significance (Dingwall et al. 2005). The purpose of the ever expanding network of geoparks is to increase public awareness to environmental values, sustainability, dynamics and hazards, promoting education and ecotourism, in addition, to create employment and to promote regional economic development (Kiss and Horváth 2006; Horváth and Csüllög 2011). A geopark

- is a territory encompassing one or more sites of scientific importance, not only for geological reasons but also by virtue of its archaeological, ecological or cultural values;
- will have a management plan designed to foster sustainable socio-economic development (most likely to be based on geotourism);
- will demonstrate methods for conserving and enhancing geological heritage and provide means for teaching geoscientific disciplines and broader environmental issues;
- will be proposed by public authorities, local communities and private interests acting together;
- will be part of a global network which will demonstrate and share best practices with respect to Earth heritage conservation and its integration into sustainable development strategies (UNESCO).

Thus, a geopark is not a category of protection, nevertheless, it promotes geoconservation; the geological heritage has to be safeguarded and sustainably managed. Recently the number of the recognized geoparks is increasing year by year and there is heavy competition to obtain this distinction.

In Hungary, the first geopark was also an international venture, the first cross-border geopark in the world (Horváth and Csüllög 2011). A Hungarian-Slovakian joint application for the Novohrad–Nógrád Geopark of 1,587 km$^2$ (1,251 km$^2$ in Hungary and 336 km$^2$ in Slovakia) was submitted and passed at the 4th UNESCO Conference on Geoparks in 2010 (Bolner-Takács et al. 2012). It comprises 91 settlements—63 in Hungary and 28 in Slovakia. There are 76 enlisted geosites within the area, among others the above-mentioned world-famous geosite at Ipolytarnóc, the curved columnar andesite in Bér, basalt sheets and necks close to Salgótarján town, sandstone cliffs and gorges at Nemti village (Chap. 19), the badland on rhyolitic tuff at Kazár village (Chap. 18) and others (see Enclosure 5 to the application dossier for nomination as a European Geopark, http://ipolytarnoc.kvvm.hu/uploads/File/pdf/NNG_Enclosure_5.pdf). Also historical and cultural values like medieval castles, literary memorials, and ethnographic features (like dresses and folk customs of the Palots ethnic group) are attractions of the geopark. Building up of the infrastructure of the geopark, including new nature trails, is in progress.

The second geopark of 3,200 km$^2$ area, approved by UNESCO in 2013, encompasses parts of the Balaton Highland National Park and the High-Bakony Protected Landscape with numerous sites of great significance in earth sciences (Korbély 2006; BFNP 2011). Beyond the outstandingly rich geological heritage considerable cultural, historical and ethnographic values also characterize the geopark. Located next to Lake Balaton, a popular holiday area, the geopark offers excellent opportunities for geotourism. Basalt cones and sheets, karstic phenomena, thermal spring cones, and raised beaches can be visited. The centre of the geopark is the Hegyestű Geological Interpretive Site, in an abandoned basalt quarry with spectacular basalt columns and excellent view from the top. Another gate to the geopark is Lavender House Visitor Centre at Tihany, at the Tihany Peninsula, which has volcanic origin. Evidence is provided e.g. at the Monks' cells, where series of explosions rapidly following each other resulted in pyroclastic flows, or at the Aranyház (Golden House) Rock, where cones of former hot spring activity can be found (Chap. 9). Further elements of geoheritage include the Tapolca Lake Cave, the basalt-covered buttes of the Tapolca Basin (Chap. 8), the paleontological heritage of the Bakony Mountains, including ammonites (sometimes reaching 1–2 m in diameter) and the recently discovered dinosaur finds. In the chapters of the present book, some of the most prominent sites, like Kab Mountain (Chap. 7) in the Bakony, the Tihany Peninsula (Chap. 9), the Kál (Chap. 10) and the Tapolca Basins (Chap. 8) of the Balaton Uplands are described.

## 32.6 Itineraries for Geomorphosites

In Hungary the most famous path for hiking is the 1,128-km long Countrywide Blue Tour trail, which runs from the Kőszeg Mountains on the western border to the northeastern corner, Hollóháza, in the Tokaj-Eperjes Mountains, along the Hungarian Middle Mountain range (Cartographia 2012, 2013a—Fig. 32.1). Receiving its name after the marking, horizontal blue stripe between two white stripes, the trail was first established in 1938. After World War II it was extended and maintained by the civil organization Friends of Nature.

### 32.6.1 Countrywide Blue Tour: Western Section

The Írott-kő (Geschriebenstein) to Visegrád section of the Blue Tour (Fig. 32.1) abounds in geosites (Cartographia 2013a). First, the geomorphosites of the Kőszeg Mountains are visited, then the Tour crosses the Alpine Foreland and touches upon the maar ruins of the Little Hungarian Plain margin (in the Sitke-Gérce area) and arrives into the Keszthely Mountains at the Sümeg Castle Hill, a steep-gradient isolated block of Lower Cretaceous crinoidal limestone, squeezed out during Miocene tectonic movements (Budai and Gyalog 2010). The Tour continues in the mountains between castles on top of Upper Triassic dolomites and Pliocene basalt formations (Basalt Street). The next stop is Lake Hévíz, the famous thermal spring crater of 4.44 ha, where water of 40.5 °C temperature is produced from the mixing of warm water deriving from a basin in the west and cold water from the thick Upper Triassic Main Dolomite sequence. Crossing the Tapolca Basin of basalt-capped Pliocene residual hills (Chap. 8), one of the most spectacular landscapes of Hungary, the route follows the northern shore of Lake Balaton into the Kál Basin (Chap. 10) with gigantic boulders in the blockfields of Upper Miocene sandstone and on to the Bakony Mountains. It climbs up to basalt-capped Kab Mountain (Chap. 7) and descends to the Csárda Hill paleokarst at Úrkút in light-red Lower Jurassic limestone, exposed by one-time manual mining of oxidic manganese ore. The valley of the Torna Stream is reached just upstream the site where one of the greatest environmental disasters, the breach of the Ajka red sludge reservoir, happened on 4 October 2010 and contaminated the stream. The Tour continues to the highest point of the Bakony Mountains (Kőris Mountain, 709 m) of Lower Jurassic nodular limestone. Towards the east, at Bakonynána, the Tour runs in the Gaja Gorge, cut into Lower Cretaceous neritic limestone forming the rapids and waterfall of the Roman Bath and again in another gorge of the same stream with rock niches at Bodajk, then crossing the tectonic Mór Graben (where the largest earthquakes in Hungary occurred). The Vértes Mountains offers important caves with finds of the Miocene savanna fauna. In the much lower Gerecse Mountains (or more properly hills) quarries of red Jurassic metamorphosed limestone (popularly called “red marble”), shaft caves, fossil localities and asymmetric Dachstein Limestone cliffs (for instance, Gellért Hill, Buda Mountains of Eocene nummulitic limestone and Eocene-Oligocene bryozoan marl) draw the visitors’ attention. The Pilis and particularly the Buda Mountains are again famous for their caves (Chap. 15) and cliffs. With the caldera of the Visegrád Mountains (Chap. 16), where further wild gorges, hoodoos, steep cliffs carved of pyroclastic flow deposits can be seen, the Transdanubian section of the Blue Tour ends on the right bank of the Danube.

### 32.6.2 Countrywide Blue Tour: Eastern Section

Beyond the Danube the Tour continues in the Börzsöny Mountains (938 m) of Miocene (Badenian) andesite (Cartographia 2012), where periglacial landforms (cryoplanational steps, rock pinnacles, rock streams, solifluction slopes) and block slides abound. Descending from the eastern slopes the path climbs up to the Castle Hill of Nógrád, a solitary parasite (biotite dacite) vent of the Danube Bend volcano. In the destroyed volcanic edifice of the Cserhát Mountains (Szanda Hill, 529 m) andesite dyke systems are exposed, ruined lava mantles and columnar andesite (Bercel Hill, 476 m) can be studied. In the eastern Cserhát the abandoned quarry of Sámsonháza presents the cross-section of a stratovolcano (alternating pyroclast and lava layers—Budai and Gyalog 2010). Crossing the tectonic Zagyva Valley, the Tour arrives at the highest mountains of Hungary, the Mátra, where it follows the narrow crest with eruption centres (Galya-tető, 964 m; Kékes, 1014 m; Szár-hegy, 743 m—Karátson 2007). In the eastern Mátra the hoodoos called “Monk” and “Nun” on the Castle Hill of Sirok are carved by sheet wash and aeolian erosion from silicified rhyolitic tuff of 15–12 Ma age. If we continue on the Blue Tour in eastern direction, we reach the Bükk Mountains (Chap. 22) at the famous exposure of the basalt and contact slate related to Jurassic submarine eruptions (pillow lava) in the Szarvaskő Gorge. Only then does the path ascend to some of “The Rocks” bordering the Bükk Plateau (Chap. 22). Turning to the north, it runs along the southern shores of the picturesque Lázbérc Reservoir of the Uppony Hills (Chap. 21), crosses the Sajó River valley to reach the strictly protected Kelemér bog, which preserves unique vegetation from postglacial times, when its depression originated from landslide damming. The next geomorphosite worth visiting is the Aggtelek Karst, a UNESCO Natural Heritage site, where the huge Baradla-Domica Cave system (Chap. 20) is found. A southeastward heading section follows across the Cserehát Hills, a Pliocene gravelly-sandy glacis surface, minutely dissected by stream valleys. Beyond the terraced valley of the Hernád/Hornád River, the first site is Fortress Hill of Boldogkő (a hill of silicified rhyolite tuff, moulded by rain, wind and frost) in the Tokaj-Eperjes Mountains of andesite and rhyolite volcanoes and Tertiary sedimentary basins. A blockfield of Sarmatian columnar andesite lies nearby. The Tour passes by the Megyer Hill of Sárospatak (Chap. 26), where millstones used to be produced. Turning to the north the trail leads to

Füzér Castle Hill, a steep dacite neck, and on to the northernmost point of Hungary, Nagy-Milic Mountain (894 m), an assumed erosional caldera. The Blue Tour ends in Hollóháza, one of the centres of Hungarian porcelaine industry.

### 32.6.3 Rockenbauer Pál Blue Tour

Pál Rockenbauer (1933–1987), a geographer and director of popular travel films for television in the editorial section called Natura, hiked with his team along the Blue Tour and presented it in television series of huge success. A more recent Blue Tour of 560 km length (Fig. 32.1), created in 1989, is dedicated to his memory. The Rockenbauer Blue Tour of Southern Transdanubia (Cartographia 2013b) also connects a number of geomorphosites, partly also included in this volume.

The starting point is the same as for the Blue Tour, Írottkő/Geschriebenstein Mountain in the Kőszeg Mountains. This trail, however, is directed southward on the foothills of the Eastern Alps, not far from Hat Rock (Chap. 6), and climbs up to the Kemeneshát Hill Range, a remnant of the gravel mantle beyond the Rába River. Making several bends in the Őrség and Göcsej Hills with loam cover and high drainage density, the trail turns to the east to the sand area of Inner Somogy and loess area of Outer Somogy with a wealth of erosional features (described in Chap. 12). The Tour follows the crest of the Mecsek Mountains (684 m) of complicated structure (eroded anticline in the west and syncline in the east) and lithology (Permian sandstone, Triassic limestones, dolomites, Jurassic marls and coal seams, Miocene volcanics). The hoodoos of Lower Triassic conglomerates (Puppet Rocks) on the southern slopes of Jakab Hill are spectacular geomorphosites. To the north is the asymmetric valley of the Kapos River (Chap. 13) and to the south the Danube bluff (Chap. 14), where detours can be made. The route ends in the Szekszárd Hills with steep-walled loess gorges, sunken lanes and slump heaps.

## 32.7 Conclusions

In Hungary, although the conservation of geological monuments dates back to the 19th century, for most people nature conservation only means the preservation of living organisms, plants and animals. For about two or three decades geoheritage has also become a central concept of the conservation; however, it has not become firmly rooted in people's mind. Therefore, it is very important, on the one hand, to demonstrate the geological-geomorphological values of the national parks, protected landscapes, nature reserves and monuments, and, on the other hand, to establish new geoparks, nature trails and other informative tools for introducing these values to the public. In Hungary, great efforts have been taken in the past years to fill this gap, and, as a consequence, the knowledge on and an interest in geosites and geomorphosites noticeably increased. Hopefully, this book too will draw more attention to the valuable geomorphosites of Hungary.

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