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New morphostructural subdivision of the Western Carpathians: An approach integrating geodynamics into targeted morphometric analysis

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ABSTRACT

A new quantitative method of morphostructure delimitation based on targeted morphometric analyses and multivariate statistical methods was applied to the Western Carpathians. Nine specific morphostructural regions and sixteen subregions were defined as an improvement on the preceding qualitative subdivision of the area. The integration of geodynamics into the targeted morphometric analysis represents a prerequisite for better interpretability of the delimited regions. The new subdivision of the Western Carpathians therefore reflects first of all the Pliocene–Quaternary geodynamics that controls the development of the present-day relief. The results also help to understand the timing of the basic dome-like morphostructural formation of the Western Carpathians (which began 4–6 million years ago, with the main stage continuing until the Late Pliocene and accelerated uplift taking place since the Middle Pleistocene), as well as the mechanism of its formation. The importance of the Middle Miocene extension for the development of the basin-and-mountain mosaic unique to the Western Carpathians is documented. The projection of the older structural boundaries into the new morphostructural regions and the increasing prevalence of the young morpholineament systems (N–S and W–E) on the southern and northern periphery of the Western Carpathians could be an indication of the gradual spreading of the Western Carpathians into the surrounding lowlands during the last stage of the morphotectonic development.

1. Introduction

The Western Carpathians are a discrete morphostructural unit of the Alpine–Carpathian mountain chain. They differ from their surroundings not only due to strong morphological individuality, but also by their specific geological development and tectonic pattern. Apart from the northern limit of the Western Carpathians, the margins of this morphostructural unit are typically gradational (Fig. 1).

The existence of the first-order Western Carpathians mega-morphostructure was mentioned and described long ago by Mazúr (1965) as a vague consequence of the 'dictiogenetic' (post-orogene) evolution. Lower-order morphostructures were traditionally perceived as geological (structural) regions. Morphologically-defined second-order morphostructural regions were firstly defined by Lacika and Urbánek (1998), but only for Slovakia and without exact explanation and interpretation. The most distinctive basic morphostructures—a mosaic of single mountain ranges and intramountain basins—used to be interpreted simply as block structures without any geodynamic explanation (Mazúr, 1979).

The new morphologically- and quantitatively-based approach presented here comprehensively defines and characterizes the basic second- and third-order morphostructures of the Western Carpathians by taking into account geological and geophysical data and geodynamic concepts, models and theories to build a synthetic picture. We show that the morphological character of the mountains can mirror the whole complex of young geological, geodynamic and geophysical features that do not have to be visibly reflected in the older geological structures. The suggested subdivision is an improvement on earlier qualitative, morphologically-based subdivision (Lacika and Urbánek, 1998; Urbánek and Lacika, 1998). This

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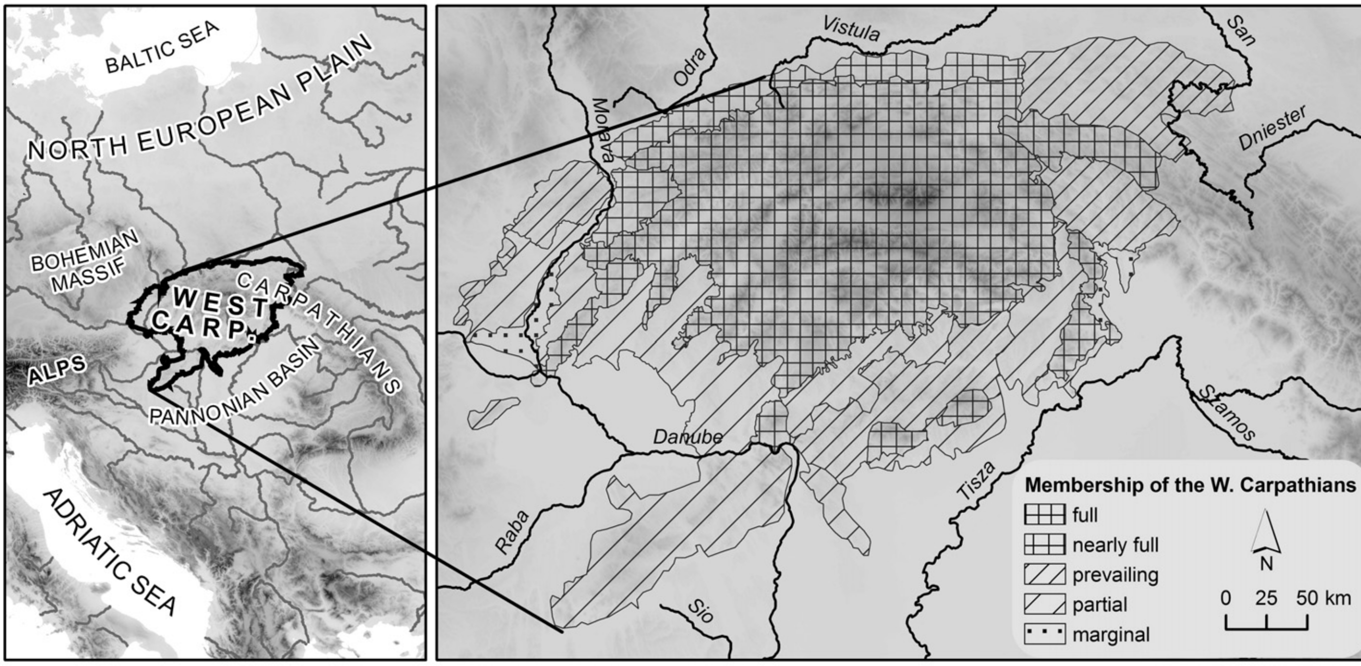

Fig. 1. Position and gradational nature of boundaries in the Western Carpathians.

improvement comes from quantitative targeted morphometrical analysis and cluster analysis that maximizes the internal morphological homogeneity of the delimited morphostructural regions.

2. Geological setting

2.1. Geodynamic development and structural pattern of Western Carpathians—an overview

The Western Carpathians form the northernmost, generally E–W trending oroclinal segment of the European Alpides and are linked to the Eastern Alps in the west and to the Eastern Carpathians in the east (Fig. 2). The northern Carpathian foreland is formed by the North European Platform that includes basement consolidated during the Palaeozoic and platform cover of the Bohemian Massif and Polish Platform in the NW and N. In the NE, the foreland is separated from the Russian Platform by the Teisseyre–Tornquist Line. A large part of the Central and most of the Internal Western Carpathians are covered by remnants of Palaeogene sedimentary basins and thick Neogene sedimentary and volcanic rock complexes that are related to the hinterland Pannonian Basin System. The present structural pattern of the Western Carpathians originated from the Late Jurassic–Cenozoic subduction–collision orogenic processes in a mobile belt between the stable North European Plate and drifting, Africa-related Apulian continental fragments. One of the most characteristic features of Western Carpathian evolution is a marked northward migration of pre-orogenic and orogenic processes, including Mesozoic rifting and extension of the epi-Variscan basement, compression and nappe stacking of attenuated continental crust and subduction of longitudinal oceanic basins (e.g., Maheľ, 1981; Birkenmajer, 1988; Plašienka et al., 1997a, b; Froitzheim et al., 2008), as well as transpression and transtension postdating the main shortening phases. The Western Carpathian orogeny ended during the Late Cenozoic after slab detachment terminated the southward subduction of the ocean–crust substratum of the External Carpathian Flysch Belt (Tomek and Hall, 1993; Bielik et al., 2004).

The Western Carpathians, like most of other Alpine collisional fold belts, have been traditionally divided into outer (northern, dominated by Tertiary deformation) and inner (southern, dominated by Mesozoic deformation) structural zones. This subdivision partly corresponds to the geographic–geomorphological subdivision of the Western Carpathians into the Outer and Inner Western Carpathians. Alternatively, a concept of triple division into the External, Central and Internal Western Carpathians has been proposed (Maheľ, 1986; Plašienka, 1999). The latter division emphasizes the presence of two diachronously closed oceanic sutures. The Early Alpine (Cimmerian–Late Jurassic) Meliata suture delimits the Internal and Central Western Carpathians. However, being largely obliterated by Tertiary cover, the precise position of this suture is not always clear. The Central and External Western Carpathians are divided by the Pieniny Klippen Belt, which is assumed to be one of the surface expressions of the Late Cretaceous to Early Tertiary closure of a supposed Penninic-related oceanic domain along the northern Central Carpathian edge.

The External Western Carpathians define the northerly convex, arcuate shape of the West Carpathian orocline. The External Western Carpathians comprise molassic sediments of the Late Cenozoic Carpathian Foredeep deposited on the margins of the North European Plate in Moravia and southern Poland, and the broad Flysch Belt composed of numerous thrust units. The Flysch Belt that forms the Cenozoic accretionary wedge of the Carpathian orogen is generally divided into two groups: (1) the Silesian-Krosno nappes in the north and (2) the Magura thrust system in the south. Layers of mostly siliciclastic Flysch sediments, which reach several thousand metres in thickness, are of Jurassic to Lower Miocene age and were detached from strongly-attenuated continental and, in some parts, probably also from truly oceanic crust (prolongation of the North Penninic). This oceanic crust was produced by Jurassic–Tertiary rifting and sea floor spreading. The basement substratum of the Flysch Belt was shortened and underthrust below the Central Western Carpathians.

The boundary between the External Western Carpathian Flysch Belt and the Central Western Carpathian basement-cover thrust units is formed by a steep, narrow zone with an intricate structure, the Pieniny Klippen Belt. This belt is formed exclusively of post-Triassic sedimentary formations sheared off their unknown basement substratum, which is interpreted to be a continental ribbon rifted off the North European Platform during the Jurassic and overridden by the Central Carpathian nappes during the Late Cretaceous. Geometrically, the basement to the Pieniny Klippen Belt corresponds to the

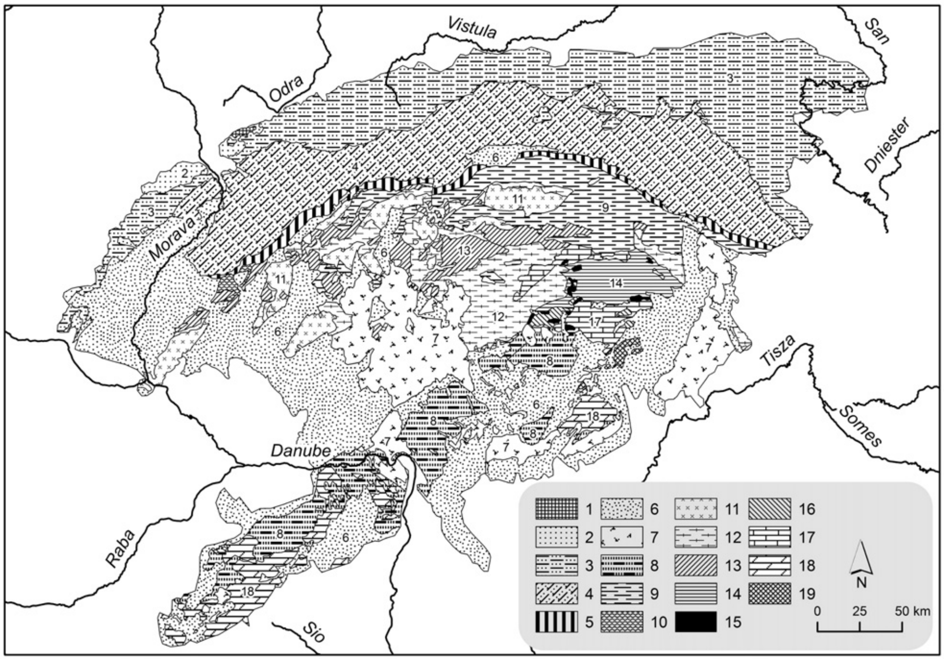

Fig. 2. Schematic tectonic map of the Western Carpathians: 1. Alpine–Carpathian foreland, 2. foredeep: Neogene sediments (unfolded molasse), 3. Silesian-Krosno units of the Flysch Belt, 4. Magura units of the Flysch Belt, 5. Pieniny Klippen Belt, 6. Neogene to Quaternary sediments of the Pannonian Basin system, 7. Neogene to Quaternary volcanic rocks, 8. Eocene to Early Miocene sedimentary rocks of the Buda Basin, 9. sediments of the Central Carpathian Paleogene Basin, 10. sediments of the Gosau Group (Late Cretaceous to Eocene), 11. Tatricum, 12. Veporicum and Fatricum, 13. Hronicum, 14. Gemericum, 15. Meliaticum, 16. Turnaicum, 17. Silicicum, 18. Transdanubicum and Bükkicum, 19. Uppony-Szendro Palaeozoic.

Briançonnais high that divides the Northern and Southern Penninics, i.e. it occupied a Middle Penninic position (e.g., Tomek, 1993; Plašienka, 2003). This basement probably forms the lower crust of the northern Central Western Carpathian zones (Tomek, 1993; Bielik et al., 2004).

The Central Western Carpathians consist of three principal crustal-scale super-units (from N to S): the Tatricum, Veporicum and Gemericum, and three detached cover nappe systems (Fatric, Hronic and Silicic; Fig. 2). Crustal units embody a pre-Alpine crystalline basement and its Late Palaeozoic–Mesozoic sedimentary cover. From the point of view of regional tectonics, the Central Western Carpathians consist of two belts: the Tatra-Fatra Belt of "core mountains" and the Vepor-Gemer Belt.

The Tatra-Fatra Belt is characterized by pre-Alpine basement-cored mountainous areas (the so-called "core mountains"), which are Late Cenozoic transtensional horst structures. From bottom to top, the following units are exposed in the core mountains: (1) the Tatric pre-Alpine crystalline basement and its Late Palaeozoic–Mesozoic sedimentary cover; (2) the Fatric (Krížna) Mesozoic cover nappe system; (3) the Hronic (Choč) cover nappe system; (4) Upper Cretaceous and Cenozoic post-nappe sedimentary and volcanic cover. The Tatra-Fatra Belt is separated from the Vepor-Gemer Belt by the so-called Čertovica Line, which is a moderately south-dipping, crustal-scale fault between the Tatric footwall and Veporic hanging wall basement-involved thrust sheets. Based on low-temperature thermochronology, the core mountains (from west to east: Malé Karpaty, Považský Inovec, Tribeč, Strážovské vrchy, Žiar, Malá Fatra, Veľká Fatra, Ďumbierske Tatry, Tatra) were uplifted during the Neogene in conjunction with subsidence of intervening intramontane basins (Kováč et al., 1994; Danišík et al., 2004).

The Vepor-Gemer Belt comprises: (1) the Veporic basement/cover, thick-skinned thrust sheet; (2) the Gemeric thick-skinned thrust sheet composed dominantly of low-grade Palaeozoic volcano-sedimentary complexes; (3) nappe outliers of Triassic–Jurassic volcano-sedimentary complexes of the oceanic Meliata Unit; (4) superficial nappes of the Silicic system; (5) Late Cretaceous–Cenozoic sedimentary and volcanic cover rocks. The Veporic basement and cover is characterized by palaeo-Alpine (Cretaceous) metamorphism that reaches amphibolite facies (e.g., Janák et al., 2001). Geochronological cooling data indicate uplift and exhumation of the Veporic metamorphic complexes during the latest Cretaceous (Kráľ, 1977; Kováč et al., 1994; Dallmeyer et al., 1996; Koroknai et al., 2001; Plašienka et al., 2007). Though partly rejuvenated during the Cenozoic, the Vepor-Gemer Belt represents the oldest exposed part of the Western Carpathians with remnants of the earliest relief (Lukniš, 1972).

The Internal Western Carpathians encompass Palaeozoic and Mesozoic volcanic and sedimentary, mostly carbonate, complexes in southern Slovakia and northern Hungary. The Internal Western Carpathians are represented by the so-called Pelso mega-unit that is exposed mostly in the north-Hungarian inselbergs (Transdanubian Range, the Bükk, Uppony, Szendrő and the Aggtelek–Rudabánya Mountains; c.f. Kovács et al., 2000). The supposed Central/Internal Western Carpathian boundary is indicated by the oceanic complexes of the Meliata Unit, but this suture is mostly obliterated by superimposed nappe units and overstep rocks. The Internal Western Carpathians are mostly composed of un-metamorphosed or low-grade Palaeozoic and Mesozoic complexes that form a south-directed thrust–fold belt. From the point of view of palaeogeography, the Pelso mega-unit exhibits close similarities to the South Alpine facies realm (Transdanubian Range), or even to the Dinarides (Bükk Mountains)—see e.g., Schmid et al., 2008.

2.2. Neogene birth of the Carpathian Chain and Pannonian Basin System

Formation of the present Carpathian Chain and Pannonian Basin System began at the end of the Early and beginning of the Middle Miocene (Mazúr, 1965; Kováč, 2000; Popov et al., 2004; Rasser et al., 2008). During this time, the Early Miocene microplates (Alcapa and Tisza-Dacia, forming a large part of the internal zones of the Carpathians and the whole basement of the Pannonian Basin) reached more or less their present position adjacent to the mid-Hungarian tectonic zone (Csontos et al., 1992; Csontos, 1995; Kováč et al., 1993, 1994, 1998).

The Early Miocene movement of the Western Carpathian lithosphere north-eastwards (due to tectonic extrusion of the Alcapa microplate from the East Alpine domain) led to oblique collision with the European platform. In front of the colliding palaeo-Alpine part of the Western Carpathians, the neo-Alpine accretionary wedge of the Outer Western Carpathians was formed. Flexure at the European platform margin, in front of the Carpathian overthrust, was flooded by foredeep basins. In contrast to the compressional tectonics in the accretionary wedge zone, the intra-Carpathian domains were subjected to stretching caused by the "slab-pull effect" of the subducting plate in front of the orogen (Konečný et al., 2002; Kováč et al., 1998; Royden, 1993a, b). Back-arc extension led to initial rifting and a large-scale back-arc basin system started to develop in the Pannonian domain (Horváth, 1993; Lankreijer et al., 1995).

The Middle Miocene geodynamic evolution of the Western Carpathians was still controlled by subduction at the orogenic front (Tomek and Hall, 1993). Termination of subduction was followed by gradual uplift of the accretionary wedge and palaeo-Alpine part of the Western Carpathians to the end of this period (Konečný et al., 2002; Kováč, 2000; Kováč et al., 2007; Kvaček et al., 2006). Sedimentation in the Carpathian foredeep retreated from its western part and continued in the eastern regions where the deepest basin depocentres developed during this time (Meulenkamp et al., 1996). In the Pannonian back-arc area, basin formation was influenced by extension. The synrift stage of basin development resulted in subsidence of individual depocentres that were filled by sediments transported by rivers flowing from the surrounding uplifting mountains (Horváth, 1993; Kováč et al., 1997, 1998). In the western and central part of the Pannonian back-arc basin, upheaval of an asthenospheric mantle diapir caused block tilting along deep-seated detachment surfaces (see Lankreijer et al., 1995; Kováč, 2000). The opening of the Danube Basin was associated with structural unroofing of the lowermost nappe units of the palaeo-Alpine structural pattern (Tari et al., 1992). The asthenospheric mantle uplift was followed by voluminous acid and calc-alkaline volcanism. Whilst this volcanic activity has been documented in the basement of the Danube Basin, it is mainly evident in the Central Slovakian Neovolcanic field at this time (Kováč, 2000; Konečný et al., 2002; Lexa and Konečný, 1998; Pécskay et al., 1995, 2006). In the east, back-arc basin formation was dominantly influenced by the pull of the submerging plate. This phenomenon was especially manifested by extension along the hinterland of the Eastern Carpathian accretionary wedge. The Transcarpathian Basin opened parallel to the orogen (Kováč et al., 1995, 1998; Kováč, 2000) and basin formation was initially associated with voluminous acid volcanic activity, and later with calc-alkaline to basaltic volcanism of island arc type, with a direct connection to subduction (Konečný et al., 2002; Pécskay et al., 1995, 2006).

The Late Miocene period represented a time of subduction retreat from the Western Carpathians eastward and south–eastward to the front of the Eastern Carpathians (Matenco et al., 1997; Linzer, 1996, Linzer et al., 1998). The end of subduction in the northern part of the Carpathian orogen and isostatic rebound led to accelerated uplift of the Western Carpathians at the end of this period. In the Pannonian Basin domain, the extensive rifting that initially took place was driven by the Eastern Carpathian subduction pull. After the Early Pannonian, moderate post-rift thermal subsidence was controlled by the development of individual basins that were mostly filled with deltaic deposits (Horváth, 1993; Lankreijer et al., 1995; Konečný et al., 2002). The Late Miocene basin formation was associated with basalt volcanism (Kováč, 2000; Konečný et al., 2002; Pécskay et al., 1995, 2006).

Tectonic inversion of the Pannonian Basin began in the Pliocene. This fact is well documented in the development of the Western Carpathians by the uplift of mountains and gradual termination of subsidence in the Neogene basins.

2.3. Deep-seated lithospheric structure of the Western Carpathians versus topography

An analysis and critical evaluation of studies related to the deep-seated lithospheric structure of the Carpathians and Pannonian Basin System showed clearly that the present topography in the region (Fig. 3a) is influenced significantly by recent structure, composition and geodynamics of the lithosphere (Zeyen and Bielik, 2000; Zeyen et al., 2002; Dérerová et al., 2006).

Lillie et al. (1994) showed changes in the topography, Moho depth, lithospheric thickness and gravity anomalies that relate to the current structure and tectonic history of the region. Their gravity models illustrated that the changes in the degree of continental collision in the Western Carpathian region, the result of plate convergence and stretching in the Pannonian back-arc basin, led to differences in lithospheric thickness. Thicker crust and higher topography in the Western Carpathians, along with the wavelength (about 50 km) and amplitude (about −60 mGal) of observed gravity anomalies, are consistent with about 50 km of continental crustal shortening and 4 km of isostatic rebound. Preservation of the thick Outer Carpathians Flysch Belt deposits and little isostatic rebound are attributable to the high-density, shallow mantle of the intact continent–ocean transition zone. Lillie et al. (1994) also concluded that the effect of the asthenosphere must be taken into account in modelling long-wavelength gravity anomalies in the Carpathian–Pannonian region. The first consideration of the lithosphere–asthenosphere boundary for interpreting the gravity field in this region was made by Šefara (1986).

Geophysical study of the lithosphere was based on 2D integrated lithospheric modelling that combined the interpretation of surface heat flow, gravity, geoid and topography (Zeyen and Fernandez, 1994). The new model of lithospheric thickness resulting from this modelling (Fig. 3b, c) indicates that the differences in recent topography of the Carpathian–Pannonian region can be explained by variations in the structure and thickness of both the crust and the lithosphere. These differences are also reflected in gravity and surface heat flow. The 2D modelling showed that the relationship between recent topography and geological and geophysical parameters changes significantly not only across the Carpathians, but also along the mountain chain (Fig. 3a, b, c).

From the theory of isostasy, it is well known that the high topography of mountain areas represents a topographic mass excess that is often compensated at depth by a mass deficiency (typically thick crust). In contrast, the relationships between topography, crustal thickness, topography and lithospheric thickness in the Western Carpathian region are different. While crustal thickening results in uplifted topography, lithospheric thinning leads to lowering of the topography (Dérerová et al., 2006).

Unlike the eastern part of the Western Carpathians and the Eastern Carpathians, no lithospheric thickening was observed in the western part of the Western Carpathians (the junction between the Eastern Alps, Western Carpathians and Bohemian Massiff). The lower topography (only about 300–400 m on average) is associated here with relatively thinner lithosphere (~110 km on average). This phenomenon might be explained by the oblique collision (dominated by strike–slips) of the Western Carpathians (Alcapa microplate) with the European Platform (the Bohemian Massif) margin (Ratschbacher, 1991a, b; Konečný et al., 2002).

The thin and warm crust (~26–28 km) and lithosphere (~60–90 km) in the Pannonian Basin System reflects stretching (extension) of the crust and associated uplift of the asthenospheric mantle. These low crust and lithosphere thicknesses are consistent with the low topography (150–200 m on average) in the Pannonian Basin region

(Fig. 3c). Here, the elevated Moho, which represents the mass excess, is isostatically compensated by the mass deficiencies of the Pannonian Basin infill (sediment thickness varies from 0 to 10 km) and the elevated asthenosphere beneath the Pannonian Basin. Regional geophysical cross sections (e.g., Lillie et al., 1994; Szafián et al., 1997; Zeyen et al., 2002 and Dérerová et al., 2006) indicate that lithospheric and crustal thinning is confined only to terranes of the Alcapa and Tisza-Dacia microplates (the region of the Pannonian Basin System and/or intra-Carpathian region). The elevation of the lithosphere–asthenosphere boundary and the Moho in the Pannonian Basin is thermally controlled (e.g., Čermák, 1982; Zeyen et al., 2002 and Dérerová et al., 2006).

3. Western Carpathian morphostructures

3.1. State of the art

The Western Carpathians comprise multiple hierarchically-ordered morphostructures. At the first level, they are an extensive, relatively flat and elliptical elevated area (dome). Its rise Mazúr (1965) attributed to repeated vertical movements conditioned probably by sub-crustal magma movement or lateral pressure in combination with an isostatic response.

Along its entire periphery, this elevated dome is bordered by the depressions of the Pannonian basin systems and Vienna basin, Carpathian Foredeep and the Transcarpathian depression. The surface of the Western Carpathian megaform is irregular, but a mosaic of discrete mountains (mainly horst and dome structures) and basins (mainly graben and flexures) creates distinctive patterns (Fig. 4).

The compactness of the megaform decreases from the NE to SW and, in the same direction, the area covered by depressions also increases. The SW margin is open to the Pannonian Basin where the extent of the elliptical morphostructure is only indicated by spurs of narrow, low mountains and slightly-elevated hilly lands in the Pannonian Basin. The average altitude of the megaform ranges from 300 to 1500 m, with the highest ranges located near the NE focus of the ellipse. In contrast, the altitudinal minimum lies near the SW focus on the gradational boundary with the Pannonian Basin.

While the highest-order unit (Western Carpathian dome) and the lowest-order units (discrete mountains and basins—basic geomorphic units) seem to be clear, the situation for intermediate units is ambiguous. Geomorphological subdivisions and morphostructural sketches (Pécsi and Somogyi, 1969; Balatka et al., 1971; Mazúr and Lukniš, 1978; Kondracki, 1978; Mazúr, 1979; Demek et al., 2007) tend only to reflect geological structures with small morphological modifications (Fig. 5). Therefore, geomorphological subprovinces and regions clearly correspond with geological structures (Fig. 5b) but their correspondence with topography alone is weaker (Fig 5a). Neglecting the neotectonic Pliocene–Quaternary development, they largely reflect the influence of pre-Neogene and Miocene lithospheric evolution.

However, various geological and geomorphological markers indicate the crucial role of Late Miocene, Pliocene and Quaternary evolution in the creation of recent relief. Fission track and radiometric data confirm the very young (since Middle Miocene) denudational history of many individual mountain ranges (e.g., Kováč et al., 1994; Baumgart-Kotarba and Král̆, 2002; Struzik et al., 2002; Bíl et al., 2004; Danišík et al., 2004, 2008). Increasing tectonic activity in the Quaternary can be deduced from the height of river terraces (Mazúrová, 1978). The neotectonic map of Slovakia (Maglay et al., 1999) stresses Pliocene–Quaternary tectonic activity. In the Outer Western Carpathians, the formation of planation surfaces and river terraces shows a number of narrow (15–20 km) but long (50–150 km) elevated and subsided structures. These structures originated in response to Pliocene–Quaternary relaxation of remnant horizontal movement within the Flysch Belt (Zuchiewicz, 1998). Young, active morphostructures are also described from various parts of the Western Carpathians (e.g., Mazúr, 1965; Činčura, 1969; Lukniš, 1973; Harčár, 1983; Lacika, 1990; Stankoviansky, 1993, 1994; Minár et al., 2003; Bizubová et al., 2005; Beták and Vojtko, 2009).

The most recent morphostructural subdivision of Slovakia (Lacika and Urbánek, 1998; Urbánek and Lacika, 1998) reflects the dominant role of the Late Miocene–Quaternary arched uplift of the Western Carpathians. The main morphostructural units are defined as parts of the Western Carpathians dome morphostructure (Central, Transitional and Marginal), complemented by southern depressed and elevated morphostructures. Although the link to the main morphostructural units of relief is evident, the purely expert-based approach without exact definition of regionalization criteria and methods raises doubts about detailed results (boundary location, delimitation of 3rd-order morphostructural units).

3.2. New method of morphostructural subdivision

Delimitation of morphostructures can be done in two principal ways. Looking for the morphological expression of geological structures is the traditional, more static and very simple approach (Fig. 6-a). However, this approach is limited in the case of complex tectonic evolution. The morphologically-based method used here arises from the application of targeted morphometric analysis and multivariate statistical methods to the delimitation of the most morphologically distinct and most clearly interpretable units (Fig. 6-b).

Targeted morphometric analysis can be considered as a part of the specific geomorphometry (Evans, 1972, 1986), where definition of analyzed map units, selection of generally used morphometric variables and definition of specific variables as well as procedures of their analyses and using are dependent on given target.

Therefore we used the basic geomorphological units that are morphologically defined but they also represent basic morphostructures—mostly individual horsts, grabens or flexures. Larger morphostructures can be delimited by examining the clustering of these basic units.

The basic geomorphological units (170 units, mostly 200–1000 km^2 in size) were characterized by mean, median and extreme values of three standard morphometric variables (altitude, slope and available relief—the difference between maximum and minimum altitude in a given area). A good morphostructural interpretation of all used morphometric variables is the first aspect of targeting of morphometric analyses. Altitude reflects the intensity of vertical tectonic movements. Uniform tectonic blocks tend to homogeneity of average altitude; fault boundaries manifest themselves by step changes in average altitude. Slope, together with available relief, defines the relief roughness that is influenced by rock resistance (passive morphostructure) and age of the morphostructure. Of course, altitudinal position (e.g. height of the tectonic uplift) also influences on roughness.

As exogenic processes considerably influence slope aspect, morpholineaments were used instead of aspect to express the spatial orientation of morphostructures. Moreover, the morpholineaments correlate with important structural boundaries and can be used for morphostructure delimitation. Morpholineaments were independently selected on DEM at two different scales: 1:3 000 000 (basic and oldest morpholineaments) and 1:1 000 000 (also reflecting younger development). To highlight regularities in the network of lineaments and to ensure compatibility with fault systems, parallel and orthogonal morpholineaments were preferred (Fig. 7).

Comprehensive knowledge of geological, geophysical and geomorphological models and concepts is a prerequisite for effective targeted morphometric analysis. This knowledge leads to more clearly interpretable morphostructural units. After considering previous work on the Western Carpathians (see above), we

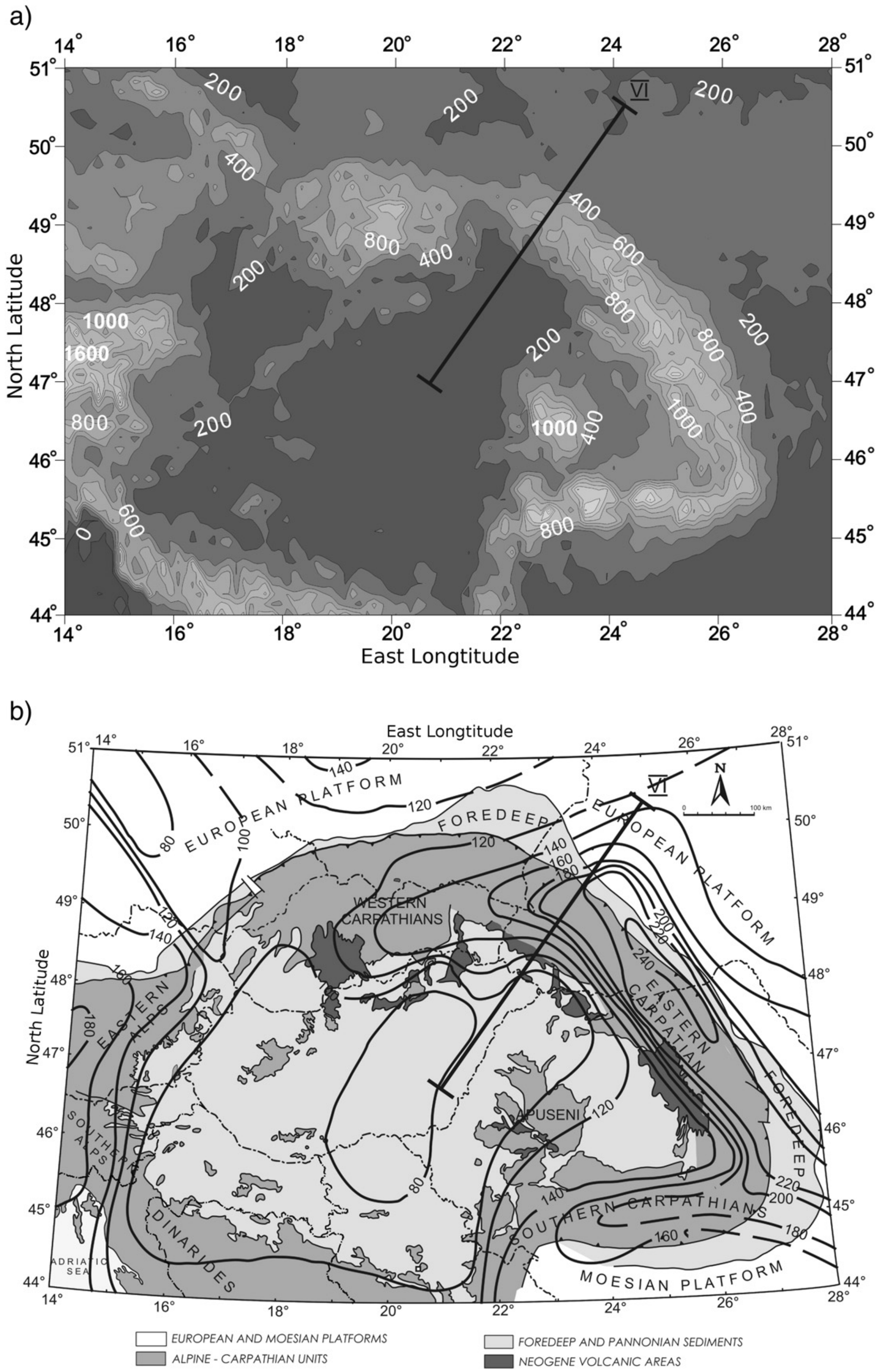

Fig. 3. (a) Topography (m) of the Carpathian–Pannonian Basin System region (from the GTOPO30 data set Gesch et al., 1999). (b) Map of lithospheric thickness (km) in the Carpathian–Pannonian Basin System region (modified after Déreróvá et al., 2006). Legend: VI—location of the transect shown in c. (c) Lithospheric model for profile VI (modified after Déreróvá et al., 2006). The dots correspond to measured data with uncertainty bars and solid lines to calculated values.

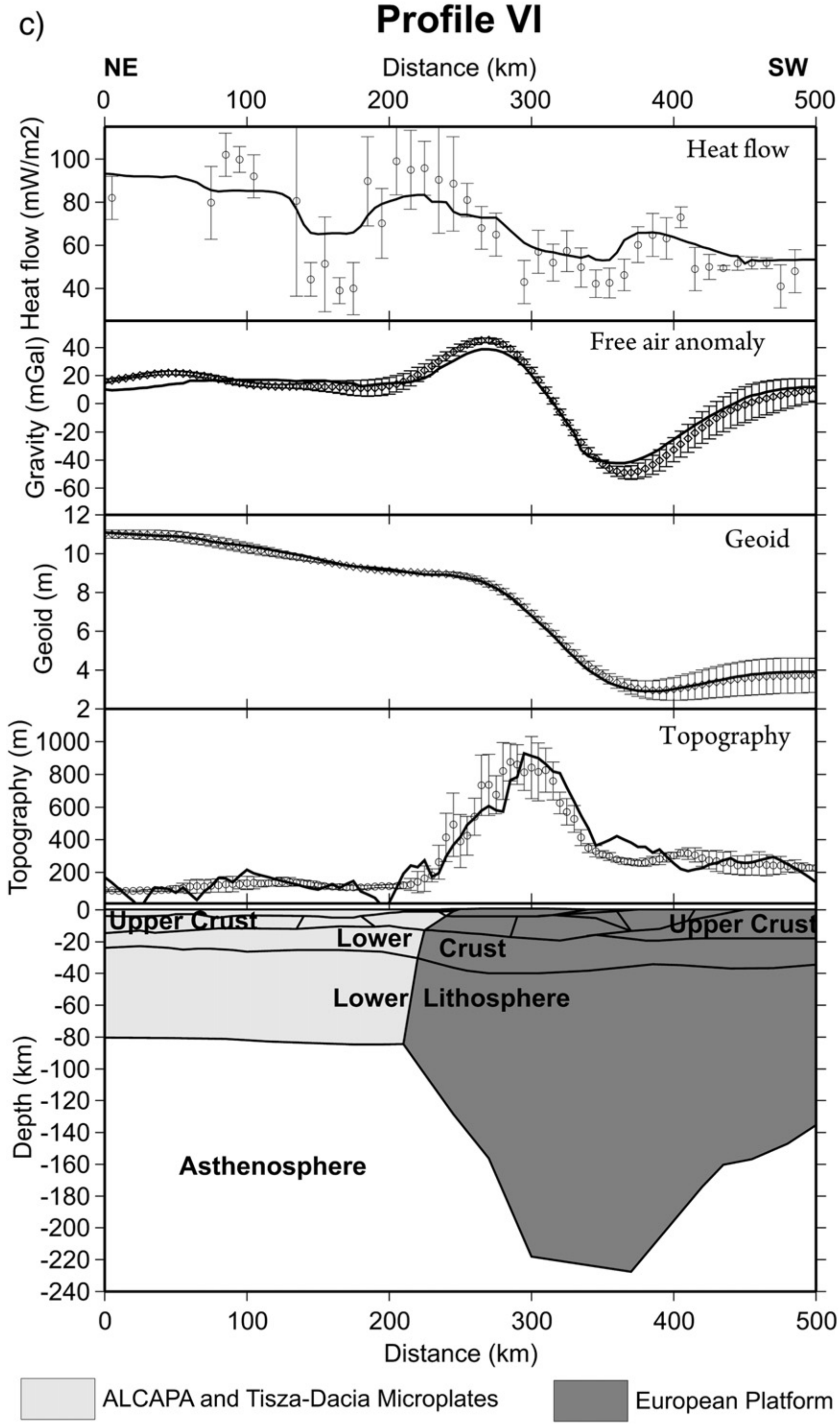

Fig. 3 (*continued*).

identified three crucial stages in the morphostructural evolution of the Western Carpathians important for targeted morphometric analysis:

a) Pre-Neogene processes that lead to the formation of a passive morphostructure, but also influenced the formation of morpholineaments that, after neotectonic reactivation, could later become the boundaries to active morphostructures.
b) Mostly extensional Miocene tectonics leading to creation of a mosaic of elevated and depressed areas that resulted from stretching of the overriding plate by the subduction pull effect.
c) Structural inversion connected with intensive Late Miocene–Quaternary uplift and creation of the recent Western Carpathian domal mega-morphostructure.

While the first two geotectonic stages are distinctly expressed in the mosaic of basic geomorphological units (lowest-order morphostructures), the influence of the last stage (c) is overprinted by the former (a, b). To identify a morphological expression in (c) we have to eliminate especially the strong morphological influence of (b).

Separation of positive (elevations) and negative (depressions) basic morphostructures and subsequent separate cluster analysis

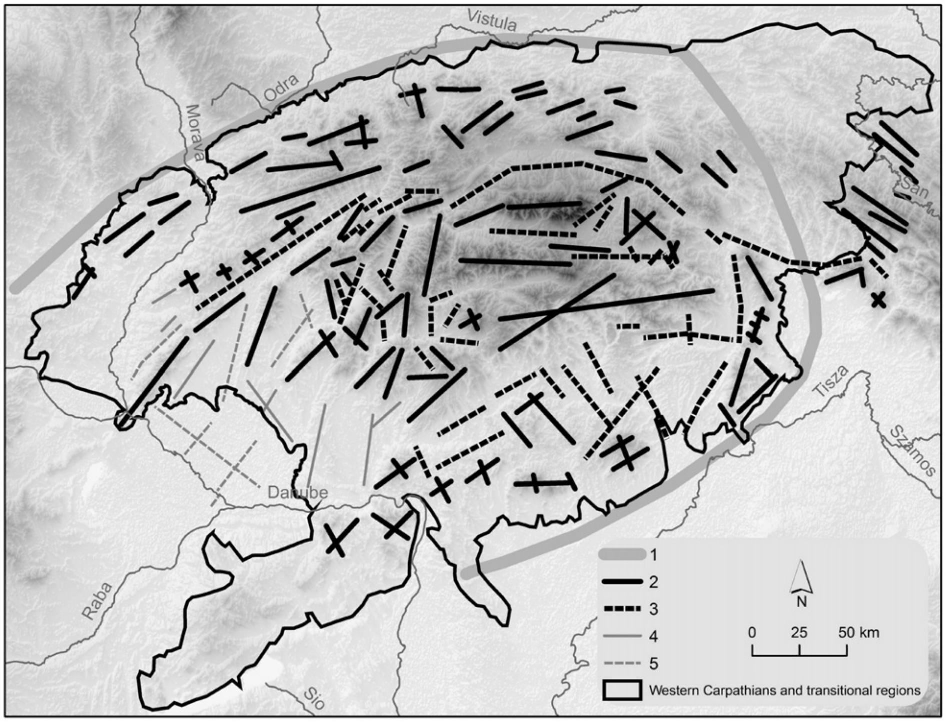

Fig. 4. Elliptical morphostructural dome of the Western Carpathians (1) and axes of positive (2—in the Carpathians, 4—in the Pannonian Basin) and negative (3—in the Carpathians, 5—in the Pannonian Basin) morphostructural units of the lowest-order (after Mazúr, 1976).

proved to be the solution. To achieve it a specific morphometric variable, relative height, was defined. Relative height along a boundary segment is defined as the difference in mean altitude of the units on either side (Fig. 8a). The affinity of a basic unit to a depression or elevation is calculated by averaging the positive and negative differences along each boundary segment, weighted by the length of each segment (Fig. 8b—the degree of affinity to elevation is expressed by positive values and affinity to a depression is expressed by negative values).

A limiting value for relative height of ± 50 m was applied to eliminate ambiguous units (with low affinity to both elevations and depressions). A set of statistical tests was separately applied to the clear elevations (63 units) and depressions (45 units). Input data are given in Table 1. The factor analysis used Varimax rotation and concentrated the morphometric information into a small number of independent and representative variables, or factors. Pairs of similar variables (mean and median altitude, total relief within a unit and 2-km relief or two boundaries characteristics) were selected in order to eliminate any statistical imperfection of the particular variables. More than 99% variance was contained in the first four factors (F1 to F4) for both elevation and depression sets (Table 2). Factors F3 and F4 mainly represent affinity to depressions/elevations and the total relief that we would like to eliminate as an expression of the older Neogene evolution. Moreover, only 7.6% (elevations) or 18.4% (depressions) of total variability is contained in these two factors, so we used only the first two factors in the subsequent analysis. Despite this, factors F3 and F4 show the relative significance of membership of the units in the sets of elevations or depressions (the same relative height is less important in a highly-dissected area than in a flat area) and can be used for the explanation of some abnormalities in the final regionalization.

To find the most easily interpreted results, various sets of input data and different factor and cluster-analysis procedures and settings

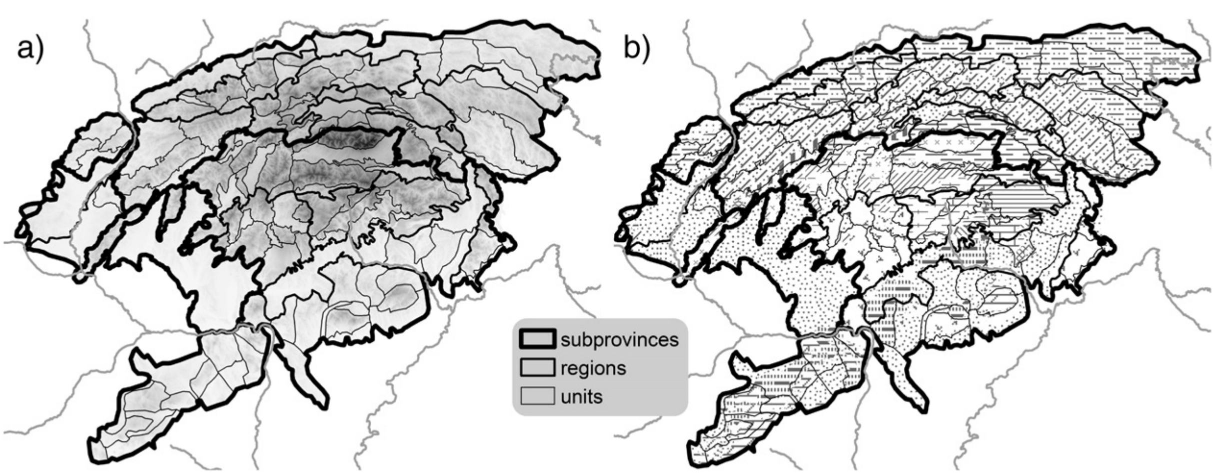

Fig. 5. Higher-order units of traditional geomorphological subdivision (synthesis of national geomorphological subdivisions: Mazúr and Lukniš, 1978; Balatka et al., 1971; Kondracki, 1978; Pécsi and Somogyi, 1969) in comparison with (a) topography and (b) geological structure (see Fig. 2 for details).

a)

Morphological specification and description

↑

Morphostructural units

↑

Geological units

b)

Interpretation of *Geological structure*

↑

Morphostructural units

↑

Multivariate statistics

↑

Targeted morphometric analysis

↑

Geotectonics models

Fig. 6. Procedures for morphostructure identification.

were tested. The most commonly used methods (Nearest Neighbour, Furthest Neighbour, Centroid, Median, Group Average, Ward's and K-means) and three distance metrics (Squared Euclidean, Euclidean, and Simple Manhattan) were applied. The results obtained were evaluated from the point of view of the stability and spatial continuity of the subdivision, as well as morphostructural interpretability. The best results were achieved using the Squared Euclidian metric as it most clearly singled out interpretable morphological domains. Ward's method was selected for displaying the clustering results shown in Fig. 9 and Fig. 10 because these results were required to minimal change in subsequent steps.

The clusters obtained ensure that there is maximum statistical difference between clusters in attribute space, but no maximal contrast between neighbouring units in geographical space. In ambiguous areas where various cluster methods offered different subdivisions, we tested the separation of geographically neighbouring units in the attribute space (F1, F2). Subsequently, we reclassified some boundary units so that we achieved maximum contrast on the geographical boundary of a morphostructure (Fig. 11a). A few units in the elevation set remain surrounded (in geographical space) by units that are classified differently. In all cases, we found reasons for these differences and reclassified them (Fig. 11b, see also next chapter). Both groups of reclassified units are depicted in Fig. 9.

An overlay of elevation and depression clusters (Fig. 10) correlates well in geographical space and was the basis for the final delimitation of morphostructural units. We selected five elevation clusters but only four depression clusters because the depressions do not include a unit like the Tatra Mountains with extreme elevations. A small shift between the character of elevation and depression clusters can have a systematic or local character and sometimes this was a reason for the definition of morphostructural subregions. Subregions were also defined on the basis of evident local morphological contrasts (distance in F1, F2 space) that were generally reflected in the alternative clustering (other methods, distance metrics or number of clusters).

The last step in our analysis was to make a schematic generalization of morphostructure boundaries. For better morphotectonic interpretation of delimited regions, the location of the boundaries was constrained by the nearest morpholineaments (Fig. 7).

3.3. New morphostructural subdivision

The delimited morphostructural units (Fig. 12) have significantly higher morphological homogeneity then traditional geomorphological units of a similar dimension. When compared to the 8 new regions (compounded from more than one elevation and one depression unit), the average homogeneity of the 15 traditional

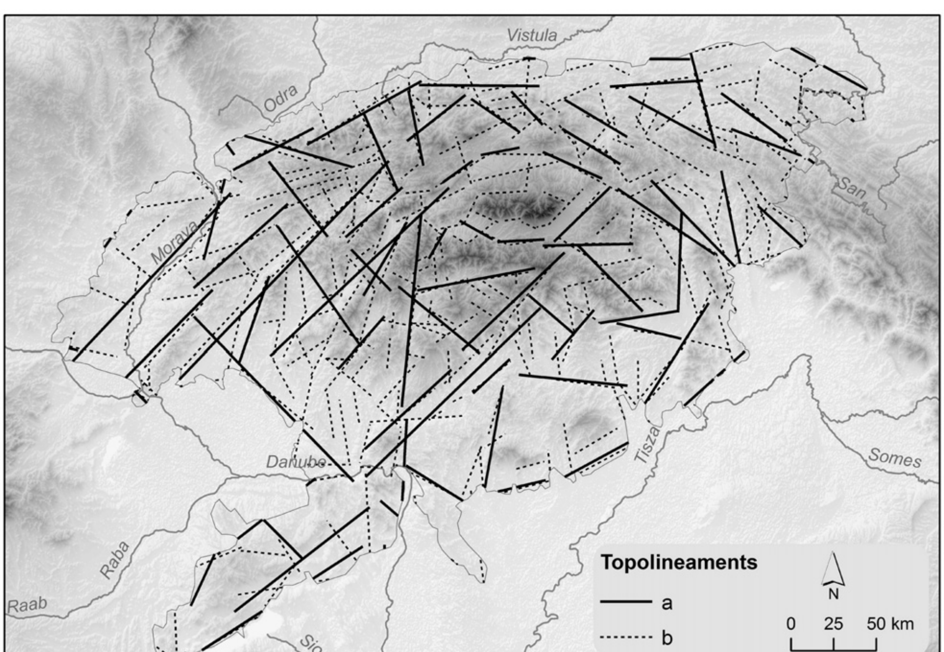

Fig. 7. Topographic lineaments derived at two different scales: a) 1:3000 000, b) 1:1 000 000.

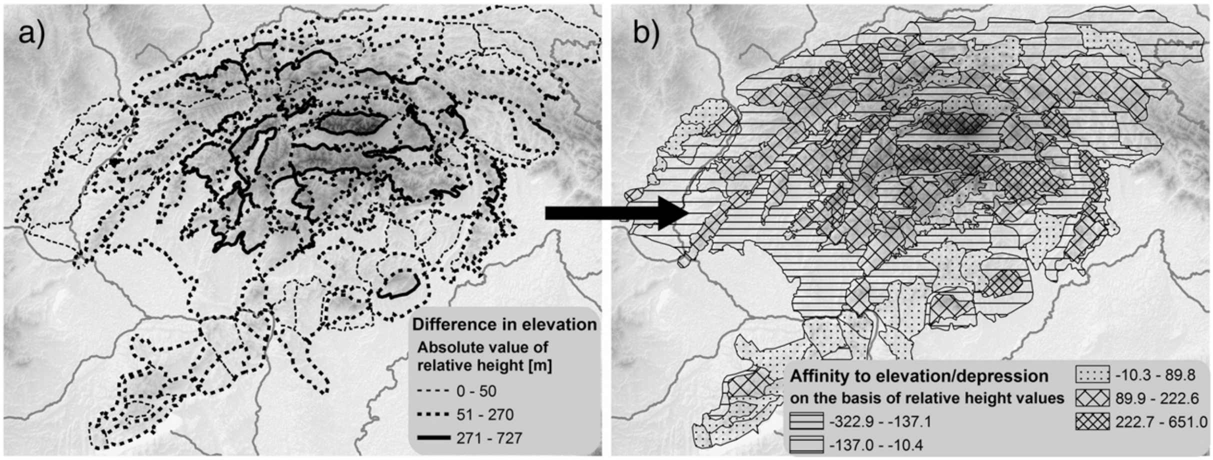

Fig. 8. Determination of the affinity of geomorphic units to elevations and depressions. Affinity (A) is expressed by the summary relative height (Rh) of its boundary segments $A = \sum_{i=1}^{n} Rh_i \cdot L_i$, where n is a number and L length of boundary segments.

regions (Fig. 5) is almost three times less for elevation units and more than two times less for depression units. This increase in homogeneity is doubled for 16 newly delimited subregions (Table 3). The following description of the new regions outlines their basic character and serves as a basis for more detailed morphostructural interpretation.

1– *The Tatra region* is morphologically the most extreme region and is represented by a single elevation unit. Despite having a very small area, we had to separate it at the region level because of its extreme morphological contrast. The morphological distance from the nearest unit (Nízke Tatry Mountains) is greater than the distance range in all other regions and the region was distinct in all clustering experiments. The distinct topographic anomaly cannot be explained simply as a culmination of dome elevation—other concepts for explaining the creation of the Tatra Mountains horst must be considered (e.g., Bac-Moszaszwili, 1993; Marko et al., 1995; Sperner et al., 2002). In summary, these concepts can be described in terms of an upper-crustal, basement-involved compressive structure developed in response to late-Tertiary collision events along the northern Carpathian boundary. Related shortening was estimated to be 30% and the maximum vertical uplift of the Tatra Mountains is about 8 km (Janák et al., 2001).
2– *The Central Region* consists of 7 elevation and 6 depression units. Three elevations (Kozie chrbty Mountains, Skorušinské vrchy Mountains, Gubałówka Mountains) have poorly-fitting (significantly lower) relief (see Fig. 9a) that can be explained by younger uplift of the original basin areas. Lukniš (1973) and Jakál (1992) confirm changes in Quaternary river networks and uplift of the first unit, while Kondracki (1978) describes Late Miocene–Quaternary uplift of the other two units that were once part of the large Obniżenie Orawsko-Podhalańskie Depression. We will term them “delayed elevations” because they were raised as morphological elevations later than the surrounding mountains. Without them, the Central Region is quite homogeneous—all experiments separated it (or its two subregions) at the level of 4–6 clusters. The region is the second highest part of the Western Carpathian dome and is divided along a distinct N–S morpholineament (hereafter the Central morpholineament) into two morphologically different parts: Tatra Subregion (2a) and Fatra Subregion (2b). Their morphological distance is close to the distance between regions and the trend of the morpholineament system changes sharply from W–E (2a) to SW–NE (2b). Although the Chočské vrchy Mountains share morphological similarities with the Fatra Subregion, they are probably only a later-activated part (delayed elevation) of the Tatra Subregion, because of their exceptional structure (no outcrop of the core), morphology (many transverse disequilibrium canyons) and their position in relation to the surrounding depression units.
3– *The Transitional Region* encloses Central region and consists of 23 elevation and 11 depression units. Despite the highest number of units, a good separation of the region was achieved in the majority of the clustering experiments. Only five classified elevation units are specific (without clear membership to the transitional cluster). Four of these (Žiar Mountains, Dzialy Orawskie Mountains, Bachureň Mountains, Súľovské vrchy Mountains) are probably delayed elevations (c.f. Klimaszewski, 1972; Minár et al., 2004), while the last (Javorníky Mountains) have a transitional character on the region boundary. Six selected subregions are parts of an annulus that is lower to the west and east and that is broken up in a N–S direction by the Central morpholineament that also divides the Central Region. Subregions westwards of the Central morpholineament (3d, e, f) are on average 200 m lower, have a radial morpholineament system and the basic morphostructures have a fragmented pattern. The morpholineament system of the eastern subregions is more sparse and homogeneous and the subregions are more massive and compact. Excepting boundaries coinciding with the Central morpholineament, the separation of subregions is less distinct.
4– *The West Marginal Region* consists of 7 elevation and 5 depression units. The region is well separated from its neighbours except for the boundary with the Javorníky subregion (3e) that is a transition between the Transitional Region and West Marginal Region (some clustering experiments classified it as a part of the West Marginal Region). The Javorie Mountains (4b), a small subregion east of the Central morpholineament, is significantly higher, but local morphological distances expressly place it into this region. A major western subregion (4a) represents a low periphery of the fragmented western part of the Western Carpathians dome. The region has a character (morpholineament system and broken, small basic morphostructures) that is partially similar to the western part of the Transitional region.
5– *The North–East Marginal Region* consists of 3 elevation and 4 depression units with good separation of elevations from the surrounding regions in all clustering experiments. However, 3 of the depression units have a specific character (relatively low, but highly-dissected marginal hilly lands) that placed them into the cluster of the Transitional regions in many experiments (see Fig. 9b). The narrower western part (5a) represents the transition between the Transitional Region (3) and the North Foreland (6).

Table 1
Input variables to factor analyses. A digital elevation model (DEM) derived from SRTM data (Jarvis et al., 2008) with 90 m resolution and GRASS GIS (Neteler and Mitasova, 2008) was used for computation. Explanation of symbols see also in the text.

Variable	Definition	Used for:
$A1$	Arithmetic mean of altitude	Elevations + depressions
$A2$	Median of altitude	Elevations + depressions
A_{max}	Maximum of altitude	Elevations
A_{min}	Minimum of altitude	Depressions
TR	Total relief within a unit $= A_{max} - A_{min}$	Elevations + depressions
S	Arithmetic mean of slope	Elevations + depressions
AR	Arithmetic mean of 2 km available relief (2 km window)	Elevations + depressions
Rh	Arithmetic mean of relative height weighting by length of boundary segment	Elevations + depressions
% $Rh+$	% of boundary length with positive Rh	Elevations
% $Rh-$	% of boundary length with negative Rh	Depressions

This subregion consists of a series of dispersed smaller positive morphostructures and one significant tectonic depression (Kotlina Sądecka Basin). The broader and larger eastern part of the region (5b) represents a transverse depression through the Flysch Carpathians between the Sandomierz and Pannonian Basins (the morphostructure of intervening belt according to Mazúr, 1979). The region neighbours the Eastern Carpathians and has a transitional character. It is characterised by a classification that is different to the national geomorphological subdivisions (Mazúr and Lukniš, 1978 versus Kondracki, 1978), a morpholineament system typical of the Eastern Carpathians (the majority of the morpholineaments trend NW–SE) and reduced crustal thickness in the Transcarpathian depression (Hók et al., 2000; Lenkey et al., 2002).

6– *The North Foreland* consists of 7 only depression units. It represents the transition between the Carpathian foredeep on one side and, on the other side, the North–East (5), partially also the West (4) Marginal Regions and, in one portion, even a subregion of the Transitional Region (3f). The region was separated on the basis of basin unit properties and a lack of clarity in the subdivision was recorded only on the boundary with the West Marginal Region (depression units of both regions are morphologically similar and fall within one cluster—see Fig. 9b). The unit is orientated in a W–E direction and morpholineaments also dominantly trend in this direction. However, N–S and SW–NE trends are more evident for second-order morpholineaments. Though its southern boundary is mostly structurally and lithologically controlled (Krosno versus Magura nappe groups of the Flysch belt), the morphological difference with southern units also results from young tectonics (resulting in marked denudation of the neighbouring Moravian–Silesian Beskids).

7– *The South–East Region* has a specific character. It consists of 8 elevation units, 8 depression units and, in particular, a high number (9) of transitional units. The region is divided into two subregions: a higher South–East Marginal Subregion (7a) and a lower South–East Foreland (7b) that could be considered also as two independent regions. The subregions are distinct in the presented cluster classification (7a—Marginal regions, 7b—Forelands, c.f. Fig. 9), but in some clustering experiments they fitted into one cluster. Because division on two independent regions did not significantly improve the mean homogeneity of regions (see Table 3) we vote their distinguishing only on the subregion level. Moreover, the character of the Slanské vrchy Mountains is similar to the Transitional Region (3), but local morphological distances support our final classification. Another elevation unit of the marginal subregion 7a (Cerová vrchovina Mountains) fitted into the foreland cluster, but it is a clear delayed elevation (Lacika, 1990). Overall, this region shows the lowest individuality (mainly because its boundaries with regions 5 and 8 are ambiguous) and its morphological neutrality (i.e. many transitional units) connects it with the Transdanubian Mountains. Values of F1, F2 generally decrease from the NE to SW that confirms ambiguous gradient character of the region. The dominant morpholineament directions clearly change with their order (SW–NE for first-order morpholineaments; N–S for second-order morpholineaments). This change could reflect very young tectonic changes in the region.

8– *The South–West Foreland* consists of 8 elevation (and 3 parts of elevations) and 2 big depression units. The elevations create small spurs into young (Quaternary) uplifted parts of the Pannonian and Vienna Basins. Elevations are significantly lower than in the adjacent West Marginal Region (4). The southern parts of three elevation units (Považský Inovec Mountains, Tribeč Mountains, Štiavnické vrchy Mountains) were separated from the West Marginal Region and integrated into South–West 8Foreland. This solved an ambiguous classification mainly related to the Tribeč Mountains. The separation from other surrounding regions is also clear. Depression units of the region are morphologically less distinctive than those of the South–East Region and elevation units are more distinctive than those of the Transdanubian Mountains. The westernmost part was separated as a special subregion (8a) and is characterised by lower morphological factor-values and the dominance of transitional units. The absolute dominance of NE–SW first-order morpholineaments (higher in the 8a subregion) is replaced by a significant increase in the abundance of the N–S second-order morpholineaments (dominating in the 8b subregion).

9– *The Transdanubian Mountains* consist of only 3 clear elevation units, but these were expressly separated in all clustering experiments. Despite this, transitional units dominate both by frequency (10 units) and area. The transitional units are similar to the foreland regions by nature of their morphological neutrality and low relief, but they differ in the absence of depression units. A distinct morphological individuality results from the specific morphotectonic development of this region. This morphostructure comprises a series of indistinct horsts. In the Tertiary, the Transdanubian Mountains were subjected to differentiated faulting, burial and renewed uplift. The individual small horsts and grabens acquired different altitudes. In a geomorphological sense, the mountain range was only created by epeirogenic uplift that began in the Upper Miocene and intensified in the Pliocene and Quaternary. Thus, the range of planated horsts with graben-like basins is morphologically not a block mountain, but a young Alpine structure (Pécsi, 1996). Distribution of the first- and second-order morpholineaments is very similar to the neighbouring regions 7 and 8, but the inversion is more distinct and W–E directions dominate the second-order morpholineaments.

Table 2
Summary of the factor analysis. STATGRAPHICS Plus for Windows was used for computation. See text and Table 1 for detailed explanation.

	Elevations		Depressions	
Factor	Variance	Maximal loads	Variance	Maximal loads
F1	79.0%	$A1$ (0.90), $A2$ (0.89)	67.8%	A_{min} (0.95), $A1$ (0.90), $A2$ (0.89)
F2	13.4%	S (0.76), AR (0.72)	13.9%	S (0.90), AR (0.85)
F3	5.0%	TR (0.78), A_{max} (0.61)	12.0%	Rh (0.69), % $Rh+$ (0.62)
F4	2.6%	% $Rh+$ (0.77), Rh (0.63)	6.4%	TR (0.84)

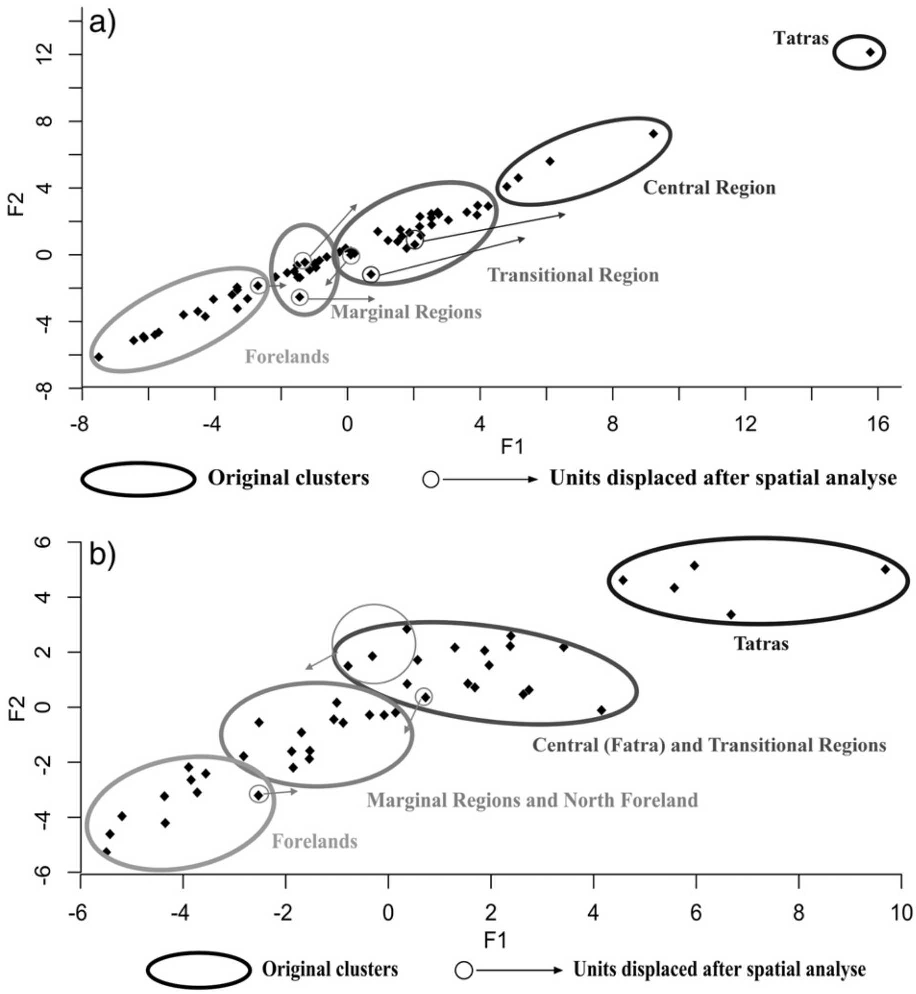

Fig. 9. Basic clusters of elevational (a) and depressional (b) units derived by using Ward's method and Squared Euclidean distance. Factors F1 and F2 are detailed in Table 1. Ambiguous (neutral) units are not depictured.

3.4. Discussion of geomorphic history and new morphostructural subdivision

The geomorphic history of the Western Carpathians, including timing, character and reasons for tectonic uplift leading directly to their present (modern) relief, are all reflected in the new morphostructural subdivision. The most important features include:

a) The previously-documented dome-like character of the Western Carpathians supra-region with its very young features.
b) The good conformity of river terraces and planation surfaces with the delimited morphostructural regions, but not with fission-track data (the scattered character of exhumation in individual regions indicates uplift these regions prior to the young morphostructures formation).
c) The cascade arrangement of the defined regions and the saddle shape of the Transitional region (3) that indicates the character of the young differential vertical tectonic movements.
d) The growth of altitudinal differences between depression and elevation units for the central parts of the supra-region.
e) The projection of the older structural boundaries into some newly-defined morphostructural regions, mainly in the northern and southern periphery.
f) The increased number of young (N–S and W–E trend) morpholineament systems, mainly on the southern, but also partially on the north periphery.
g) The distinctive change in character of the regions and subregions along the central morpholineament boundary.

The explanation for the dome-like character of the Western Carpathians supra-region with its very young features was mentioned only briefly in older literature. However, the reasons for this phenomenon were either not dealt with or the tectonic explanation was limited by the level of knowledge at that time (c.f. Mazúr, 1965; Klimaszewski, 1981). We suggest that the source of this uplift could be very similar to the Alps, naturally with some peculiarities.

In the Alps there are principally two concepts for explaining uplift. Some researchers argue that climate is/was important for the young uplift and the increase in erosion (e.g., Schlunegger, 1999; Kuhlemann et al., 2006). The second school argues for tectonically-driven uplift. In this group, a variety of mechanisms have been proposed to explain the uplift. Genser et al. (2007) assumes that the late-stage widespread uplift of the entire Alpine realm (at about 6 Ma), including the northern peripheral foreland basin, is most likely explained by distributed delamination and/or convective removal of over-thickened lithosphere and the northward spread of the mechanical decoupling between the mantle and crust of the subducted lithosphere. These processes are thought to be enhanced by erosional unloading. The experiments described by Willingshofer and Sokoutis (2009) depict such a situation during the late Miocene–Pliocene, where the foreland plate was thrust under the orogenic wedge, leading to crustal thickening and uplift of the internal part of the orogen and subsidence of the foreland plate. They argue that collisional mountain belts in general are expected to evolve from a decoupled to a coupled system, leading to a change in the dominant

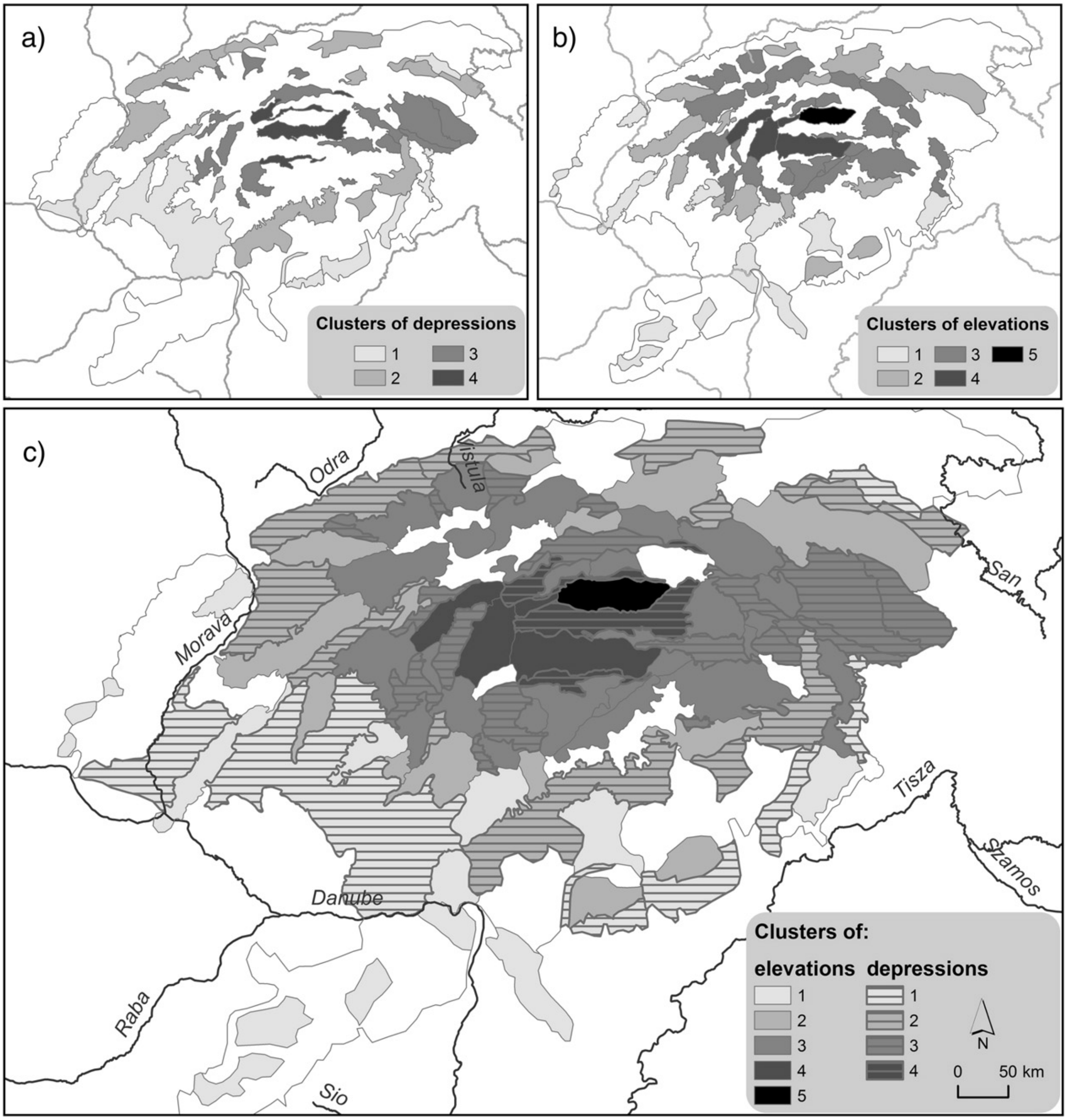

Fig. 10. Map view of the separate clusters (4 depression clusters, 5 elevation clusters) and their synthesis.

deformation mechanism as portrayed in the vertical motions of the orogenic wedge and the foreland.

The neotectonic rise of the Western Carpathian dome clearly and continuously deforms the initial planation surface (Lukniš, 1962)—'Mid-mountain level' (Mazúr, 1963) and the uplift should, therefore, be younger. In contrast, the exhumation age derived from fission-track data is very variable within the delimited regions (e.g. Kováč et al., 1994; Kováč, 2000; Baumgart-Kotarba and Král̆, 2002; Danišík et al., 2004, 2008; Struzik et al., 2002). This indicates that the detected exhumation history is older than the development of the dome. The youngest fission-track data from about 10 Ma determines the maximum age of the 'Mid-mountain level' that itself requires a

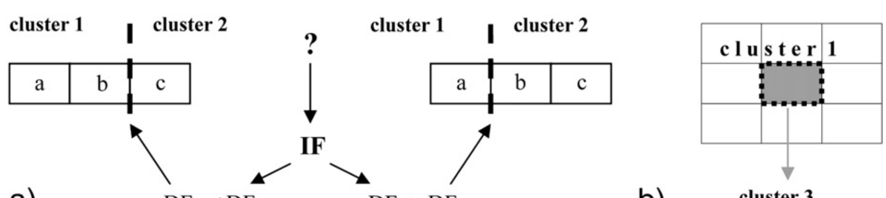

Fig. 11. Two reasons of unit reclassification: a) Ambiguous boundary unit classified by various experiments to various clusters; finally classified on the basis of local contrast. b) Indistinctive elevation unit surrounded by uniformly classified units; classified into cluster of neighbouring units. DF_{ab}—distance of units a and b in the attribute space (F1, F2).

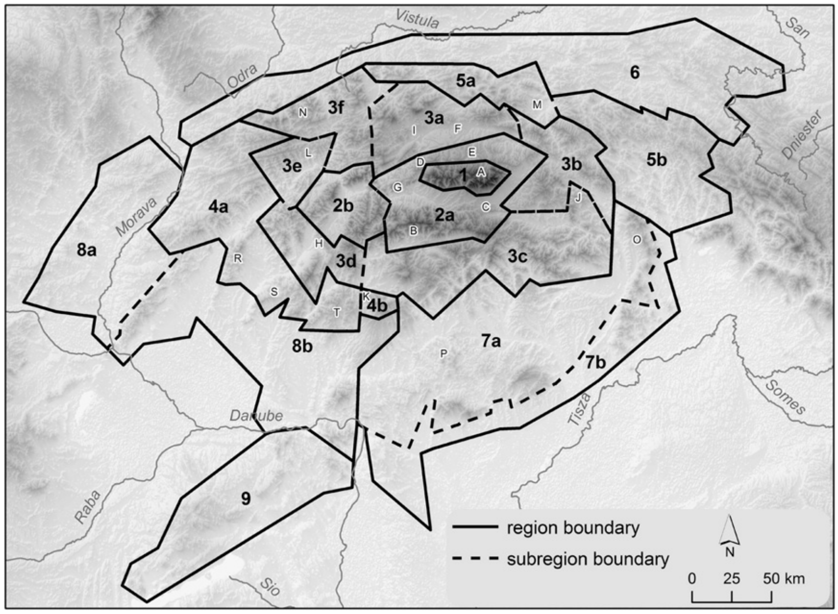

Fig. 12. Morphostructural subdivision of the Western Carpathians (numbers correspond with labelling on the map), regions and (in brackets) subregions: 1. Tatra Region, 2. Central Region (2a. Tatra Subregion, 2b. Fatra Subregion), 3. Transitional Region (3a. Beskid Żywiecki—Gorce Subregion, 3b. Beskid Sądecki—Levočské vrchy Subregion, 3c. Slovenské rudohorie Subregion, 3d. Strážovské vrchy—Kremnické vrchy Subregion, 3e. Javorníky Subregion, 3f. Moravian–Silesian Beskids Subregion), 4. West Marginal Region (4a. Biele Karpaty–Štiavnické vrchy Subregion, 4b. Javorie Subregion), 5. North–East Marginal Region (5a. Beskid Wyspowy Subregion, 5b. Low Beskids Subregion), 6. North Foreland, 7. South–East Region (7a. South–East Marginal Subregion, 7b. South–East Foreland), 8. South–West Foreland (8a. Vienna Basin Subregion, 8b. Danube Basin Subregion), 9. Transdanubian Mountains. Geomorphological units mentioned in the text: A Tatra Mountains, B Nízke Tatry Mountains, C Kozie chrbty Mountains, D Skorušinské vrchy Mountains, E Gubałówka Mountains, F Obniżenie Orawsko-Podhalańskie Depression, G Chočské vrchy Mountains, H Žiar Mountains, I Dzialy Orawskie Mountains, J Bachureň Mountains, K Javorie Mountains, L Javorníky Mountains, M Kotlina Sądecka Basin, N Moravian–Silesian Beskids, O Slanské vrchy Mountains, P Cerová vrchovina Mountains, R Považský Inovec Mountains, S Tribeč Mountains, T Štiavnické vrchy Mountains.

Table 3
Mean distance of elevation (*E*) and depression units (*D*) of new and traditional morphostructural subdivision in the attribute space F1, F2. Distance was not computed if only one clear elevation or depression unit create a region.

New regions and subregions	*E*	*D*	Traditional subprovinces and regions	*E*	*D*
1 Tatry Region	–	–	Inner Western Carpathians	5.51	3.73
2 Central Region	0.95	2.45	Fatra-Tatra Region	10.11	1.59
2a	–	0.20	Slovak Middle Mountains	1.44	0.82
2b	0.16	0.79	Slovak Ore Mountains	1.51	–
3 Transitional Region	0.40	0.75	Lučenec–Košice depression	–	0.55
3a	0.10	0.25	Matra–Slansky Region	0.35	0.40
3b	0.08	1.01	Outer Western Carpathians	2.72	2.95
3c	0.09	0.28	Austrian–Moravian Carpath.	0.02	–
3d	0.10	–	Middle Moravian Carpath.	–	–
3e	0.17	–	Slovak–Moravian Carpath.	0.34	0.05
3f	0.12	0.16	Western Beskids	0.76	0.16
4 West Marginal Region	0.17	0.16	Middle Beskids	1.22	3.89
4a	0.11	0.05	Eastern Beskids	0.11	5.08
4b	–	–	Low Beskids	–	1.61
5 North–East marginal region	0.10	0.52	Western Beskids Piedmont	–	0.29
5a	0.08	–	Eastern Beskids Piedmont	–	0.96
5b	–	0.20	Podhôĺno–Magura Region	–	3.66
6 North Foreland	–	0.41	Transdanubian Mountains	0.35	–
7 South–East Region	1.63	0.51	Small Danube basin	–	–
7a	0.17	0.17	Danube Lowland	–	–
7b	2.78	0.22	Vienna Basin	–	–
8 South–West Foreland	0.58	0.21	Záhorská Lowland	–	–
8a	0.23	–	Moravian Basins	–	–
8b	0.05	–			
9 Transdanubian Mountains	0.20	–			
Mean regions	0.58	0.72	Mean subprovinces	2.86	3.34
Mean subregions	0.33	0.33	Mean regions	1.62	1.59

few million years to form. Generally, fine-grained Late Pannonian and Pontian correlative sediments in the Pannonian Basin and Western Carpathians intramountain basins are also indicative of the formation of the 'Mid-mountain level'. The general character of sedimentation was demonstrably changed in the Pliocene (coarse sediments substituted fine sediments). Consequently, the dome probably first rose sometime during the last 4–6 million years. However, while the altitude of the 'Mid-mountain level' corresponds with the mean altitudes of the delimited regions, the younger 'River Level' (Upper Pliocene–Early Quaternary pediment after Mazúr, 1963) and Quaternary river terraces differ far less between individual regions. This indicates that the main stage of the dome formation occurred in the Pliocene and that both the 'River level' and the river terraces were formed in the existing dome.

While the isostatic component of the accepted model described above is clearly confirmed by lithospheric thickness modelling, the annular form of the regions and the saddle shape of the Transitional region are well explained by the influence of pressure from the encroaching Adriatic plate. This influence began in the Western Carpathians after relative cooling and strengthening of the Pannonian Basin lithosphere. Moreover, the time elapsed between the end of the extensional tectonic phase, with an elastic character reflecting overheated lithosphere (Sarmatian–Early Pannonian), and the relative strengthening of the Pannonian Basin lithosphere in a postrift/tectonic inversion state (Pontian–Early Pliocene), offers sufficient time for the formation of the 'Mid-mountain level' (its character, timing and extent represent an important part of regional geomorphology that has been intensively discussed during the last decade—see e.g. Činčura, 1998; Urbánek, 2001; Minár et al., 2004; Bíl et al., 2004; Danišík et al., 2008; Beták and Vojtko, 2009).

Middle Miocene extension in the Central and Internal Western Carpathians was connected with back–arc extension and active elongation of the Western Carpathians orogen in response to subduction roll back in front of the Eastern Carpathians (Csontos, 1995; Kováč, 2000; Konečný et al., 2002). This extension played an important role in the development of the basins–mountains (depressions–elevations) mosaic specific to the Western Carpathians. This mosaic was later amplified on the younger rising (uplifting) domal mega-morphostructure with an extensional regime on the surface. The influence of different unloading of the basins and mountains created conditions suitable for the observed Pliocene–Quaternary growth of the altitudinal differences between the depression and elevation units, mainly in the central parts of the Western Carpathians (c.f., Genser et al., 2007; Willingshofer and Sokoutis, 2009).

The projection of the older structural boundaries into our newly-defined morphostructural regions and the increased abundance of young morpholineament systems (N–S and W–E), mainly on the southern and northern periphery, also confirm and complete the concept outlined above. The first factor suggests that while the influence of the older structure in the central part of the dome was markedly reduced by dome uplift (because uplift is more intensive and probably older), the periphery preserves the influence of the older structural elements because of less distinct and younger uplift. The younger character of the periphery is confirmed by the abundance of the young morpholineament system, but also by the change in 'River level' character from the pediment (in the central part of the Western Carpathians) into a pediplain that truncates the youngest Neogene (Pliocene) sediments. However, a high abundance of the young morpholineaments could also reflect: (a) the presence of thin-skinned crustal bodies (thin-skinned tectonics) in the peripheral regions that are more sensitive to the most recent tectonic pulses; or (b) southward fading of the oroclinal shape influence on the recent stress field (the recent stress field indicates compression orthogonal to the Western Carpathians arc and extension parallel to it—c.f. Hók et al., 2000; Cloetingh et al., 2002; Vojtko et al., 2008).

The distinctive splitting of the Western Carpathians dome along the N–S oriented Central morpholineament (Central Slovak Fault System) can be explained by geophysical data that suggest rapid thinning of the crust and lithosphere to the west that led to lowering and disintegration of morphostructural units. However, the influence of crust and mantle thinning diminishes in the south (where the lithosphere is thinner to the east) and the young dynamics related to the recent approach of the Adriatic plate provides an alternative explanation (the Danube River has followed the Central morpholineament only since the Middle Pleistocene). Together with the abundance of young morpholineaments found frequently in the youngest Neogene sediments in the south, increasing river activity since the Middle Pleistocene (based on the height of the river terraces) and possibly also young elevated and subsided structures of the Outer Western Carpathians (c.f. Zuchiewicz, 1998) could indicate that the most recent and more active stage of the morphotectonic development started in the Middle Pleistocene.

4. Conclusions

- Original targeted morphometric analysis enabled the delimitation and basic description of nine specific morphostructural units (regions) creating parts of the Late Miocene–Quaternary uplifted dome-like mega-morphostructure of the Western Carpathian supraregion: 1. Tatra Region, 2. Central Region, 3. Transitional Region, 4. West Marginal Region, 5. North–East Marginal Region, 6. North Foreland, 7. South–East Region, 8. South–West Foreland, 9. Transdanubian Mountains.
- The morphologically-oriented morphostructural subdivision of the Western Carpathians is an alternative to older structurally-based subdivisions (Pécsi and Somogyi, 1969; Balatka et al., 1971; Mazúr and Lukniš, 1978; Kondracki, 1978; Demek et al., 2007) and improves and completes the first morphologically-oriented subdivision (Lacika and Urbánek, 1998) in a quantitative manner. The morphostructural subdivision is established not only from information contained in relief characteristics, but also on the basis of the geodynamic evolution of the orogen. The use of consistent regional elements of geomorphological, geological and geophysical theories and models, as well as the quantitative basis for regionalization, all offer sound data for a morphostructural interpretation that is more objective than the preceding qualitative subdivisions.
- The targeted morphometric analysis (separation of the elevation and depression units) and subsequent cluster analysis used here (maximizing the internal morphological homogeneity and external differentiation of regions) lead to delimited regions whose morphological homogeneity is several times higher than for traditional divisions (Mazúr and Lukniš, 1978).
- The morphostructural plan of the Western Carpathians presented here shows a certain degree of correlation with geological patterns, but mainly with the geodynamic evolution of the Western Carpathians. While the Palaeozoic and Mesozoic geological pattern rarely influenced the morphostructural regions (it can only be detected in some boundary morpholineaments), the Neogene to Quaternary geodynamic processes expressed in the young active fault structures determined the limits and character of the delimited morphostructures. Therefore, all delimited Western Carpathian morphostructural units can be regarded as active Neogene–Quaternary morphostructures (endogenous landform elements in terms of Gerasimov and Meščerjakov, 1967). Passive (older) morphostructures (reflecting the resistance of the rocks affected by exogenous processes) do not significantly manifest themselves in the morphometric variables used.
- The modern synthetic hypothesis of Western Carpathian relief formation was outlined in connection with the presented results. Particular characteristics of the delineated morphostructural units can contribute to the determination of the main stages of modern Western Carpathian relief formation: the planation of a large part of the Western Carpathians at the Miocene/Pliocene boundary; the beginning of the dome morphostructure formation before 4–6 million years; the main stage of the uplift to the Late Pliocene; and the last, more active stage starting in the Middle Pleistocene.
- The new morphostructural subdivision is also in good agreement with the probable mechanism of the Western Carpathian uplift (distributed delamination, convective removal of over-thickened lithosphere, spreading of the mechanical decoupling between the mantle and crust and erosional unloading preceded by an extensional roll-back effect). The subdivision enabled us to define various spatial specifications and their consequences (e.g. young spreading of the Western Carpathians into the Pannonian Basin realm and the specific character of morphostructural units that are divided by the N–S Central morpholineament).
- All the outlined interpretations create a new holistic framework for subsequent detailed regional studies focused on the consistent geomorphodynamic explanation of all morphological features within the delineated morphostructural regions.

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