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A comparison of Cenozoic Neo-Alpine tectonic evolution of the Western Carpathian and Himalayan orogenic belts (Slovakia – Nepal)

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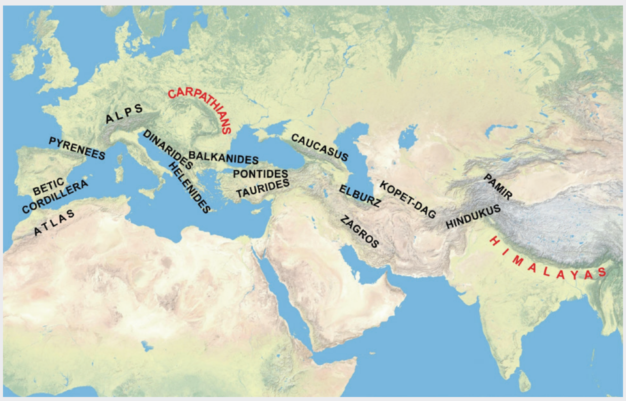
Abstract: The Western Carpathians and Himalayas belong to the same global Alpidic orogenic system, which is the result of Euroasian plate collision with the continental fragments of Gondwana after closing of the Tethys ocean. Having the field experiences from both distant mountain ranges of the same orogenic system and applying the principles of comparative tectonics, they were evaluated and compared in the paper. Generaly, they have the same collisional structural-tectonic style, but there are as well many peculiarities and differences resulting from the specific conditions of collision in the Western Carpathian and Himalayan areas. The Western Carpathians structure is a result of gradual alternation of Variscan (Hercynian; Paleozoic), Paleo-Alpine (Mesozoic) and Neo-Alpine (Cenozoic) convergent and divergent plate tectonic processes, while the Himalayas represents purely Neo-Alpine Cenozoic structure evolved during the continual long lasting and rapid plate convergence. Despite the geosutures from the earlier orogenic evolutions are known in the parallel north-located zone, too. As the most important factors, influencing character of collision, seems to be the geometry of converging plate margins and the rate of the ocean floor spreading/subduction, driving the orogenesis. Paper gives a brief overview of tectonic architecture and evolution of both mountain ranges and compares their common features and contrasts.

Key words: comparative tectonics, Tethyan Alpides, Western Carpathians, Himalayas

Graphical abstract



Highlights

- Paper compares tectonic architecture of two distant orogenic segments of Tethyan Alpides.
- The differences of W. Carpathians and Himalayas Cenozoic Neo-Alpine collisions result from the geometry of plate margins and rates of plates convergence during the Tethys ocean closure – W. Carpathians are characterized with oblique Cenozoic collision of strike-slip-orogen type, but Himalayas are a result of frontal collision, which in contrast with W. Carpathians still continues with high rates of recent movements, strong erosion and extreme seismicity.

1. Introduction

There are two prominent young and not yet denuded global orogenic systems of the world – a meridional pan-American Circum Pacific (South American Ands and North American Cordilleras) and equatorial Alpine-Himalayan (A-H) belts. They both evolved at active margins of converging lithospheric plates. The pan-American orogenic belt as a part of circum Pacific mobile zone is Andian and Cordillera type orogen, meanwhile A-H orogen is a collisional Alpine-type orogen (sensu Dewey & Bird, 1970). In terms of plate tectonics, the Alpine-Himalayan orogenic belt is a result of continent-continent collision, imprinting to collisional zone the tectonic style of extreme shortening and uplifting, produced predominantly by thrusting, with an important role of strike-slip tectonics as well. The Alpine-Himalayan global world collisional zone is composed of many mountain systems, listed from the west: North African Atlas, Betic cordillera, Pyrenees, Apenines, Western, Central and Eastern Alps, Western, Eastern and Southern Carpathians, Dinarides, Balkanides, Helenides, Anatolides (Pontides and Taurides), Caucasus, Iranides (Zagros, Elburz and Kopet Dag), Hindukus, Pamir, Karakoram and Himalayas. The southern continuation of this mobile belt is indicated by subduction zones of Sunda-Java trenches and Alpine mobile belts of Barma, Malaysia, Sumatra, Borneo, Java, Fiji and New Zealand (**Fig. 1**). Although all these segments of A-H belt have a common nature, there are particularities and some structural-tectonic differences between individual segments of this extended belt due to the local conditions and geometry of plate margins, type of collision, type and physical properties of lithosphere, rate of convergence, geological evolution, etc. In the frame of the Slovak research project APVV-16-0146 and in cooperation with the Department of Geology, Tri-Chandra Multiple Campus, Tribhuvan University in Nepal we realized the reconnaissance field research trip in april 2019 (**Fig. 6**). It was focussed on transect from Pokhara to Muktinath localities along Kali Gandaki river valley, crossing the zones of Lesser and Higher Himalayas in Nepal (Mojzeš et al., 2020). The Kali Gandaki river valley represents the deepest antecedent valley in the Himalayas. It provides a natural geological cross-section through the tectonic contact of the main Himalayan units. The main objective of our joint collaborative field work and review of relavant literature was to compare structural evolution and tectonic style of two distant segments of the global equatorial orogenic system – the Western Carpathians of Slovakia and the Nepal Himalayas.

Tectonic architecture and evolution of compared orogens

Western Carpathians

The Carpathians represent a part of the Mesozoic-Cenozoic Alpine-Himalayan fold and thrust belt – the result of collision of Gondawana plates with Euro-Asian plate (Laurasia) during the closure of Tethys ocean. They are divided into Western, Eastern and Southern Carpathians. The Western Carpathians, covering the whole territory of Slovakia (**Fig. 2, 4a**), represent eastern orographic continuation of Eastern Alps and further east they continue to Eastern and next to Southern Carpathians. Although the Alps and Carpathians belong to the same Alpidic system, there are differences in Neo-Alpine (Neogene–Quaternary) evolution of individual orogenic segments. The Alps represent a zone of shortening due to typical frontal continental collision with a very deep orogenic roots, while the Carpathians are the result of tectonic escape of microplates (Inner Western Carpathians (IWC), Pelsö, Tisia) from the Alpine domain (Doglioni et al., 1991; Ratschbacher et al., 1991a, b) to the area of subducting oceanic lithosphere of the Magura basin, creating the embayment in the Euroasian lithospheric plate (EP). It led to the oblique continent-continent (CC)

Fig. 1. Tethyan Alpides of the world.

collision of IWC microplate with EP in the western part of the Western Carpathians and tectonic arrangement of flysch sediments of the Magura basin into the pile of nappes forming the current Outer Carpathians accretionary wedge. The Carpathian loop was formed during two successive orogenic events. At the Paleo- and Meso-Alpine Jurassic–Late Cretaceous–Early Paleogene epoch (Plašienka, 1999, 2018a) the nappe architecture of pre-Tertiary units was formed far from their recent position. During the Neo-Alpine Cenozoic epoch (Kováč, 2000) units which were consolidated earlier during the Paleo-Alpine epoch removed to Carpathian space and were arranged in a new configuration – the nappe structure of Paleogene sedimentary complexes in front of the prograding Paleo--Alpine units was formed. Final neo-tectonic character was imprinted to orogen in the latest Pliocene-Quaternary stages of tectonic evolution.

The principal tectonic division of the Western Carpathians is derived from the youngest Neo-Alpine and mostly Miocene tectonic processes, when the flysch prism of the Outer Western Carpathians and the Pieniny Klippen Belt structure were created during collision of the Inner Western Carpathians block with the foreland. So, the Western Carpathians sensu Biely (1989) and Bezák et al. (2004) are divided to Inner and Outer Carpathians. The Inner Western Carpathians (IWC) are composed of the Tatric, Veporic and Gemeric Paleo-Alpine crustal basement nappe units and the Fatric, Hronic, Meliatic, Turnaic and Silicic detached superficial Mesozoic nappe units. The crustal basement units are formed of crystalline basement with incorporated fragments of Variscan (Hercynian) tectonic units, and covered by autochthonous Upper Paleozoic and Mesozoic formations. The Meliatic unit encompasses remnants of ophiolite suite of closed Jurassic ocean (Kozur & Mock, 1973; Kozur et al., 1996). The rare occurrences of Meliatic unit follows Paleo-Alpine collisional suture created after this Tethys-related Meliata Ocean closure (Plašienka et al., 2019). Another younger-Neo-Alpine

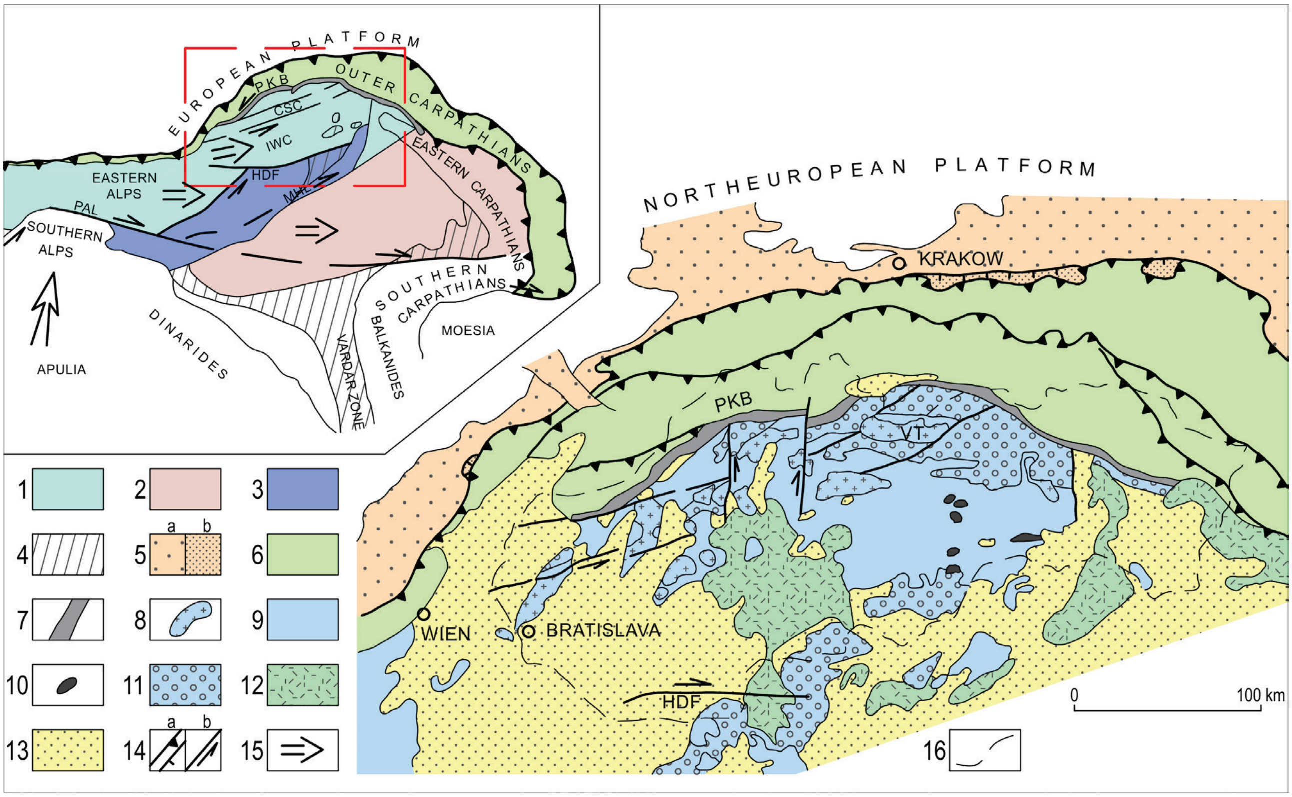


Fig. 2. Neo-Alpine architecture, dynamics and tectonic division of the Western Carpathians (compiled according to Kovács et al., 2000; Lexa et al., 2000; modified according Marko et al., 2017). Abbreviations: PKB – Pieniny Klippen Belt, VT – High Tatras = Vysoké Tatry Mts., IWC – Inner Western Carpathians, CSC – Carpathian Shear Corridor, HDF – Hurbanovo-Dijósjenö Fault, MHL – Mid-Hungarian Line, PAL – Peri-Adriatic Line. Explanations: 1 – ALCAPA micro-plate; 2 – Tisia micro-plate; 3 – Pelsö micro-plate; 4 – oceanic crust domains; OUTER (External) WESTERN CARPATHIANS: 5 – Miocene molasse sediments – a) autochtonous not deformed, b) mobilized, thrusted and folded; 6 – Neo-Alpine orogenic accretionary prism of pre-dominantly Paleogene flysch sediments; INNER (Internal) WESTERN CARPATHIANS: 7 – Pieniny Klippen Belt – suture zone of extreme shortenning and shearing; 8 – Paleozoic crystalline basement exhumed in core mountains; 9 – Paleozoic-Mesozoic complexes as a whole; 10 – Meliatic unit – ophiolites; 11 – Undeformed Inner Carpathian Paleogene sediments; 12 – Neogene syn- and post-collisional volcanites; 13 – Neogene back-arc and intra-arc sedimentary basins; 14 – a) Prominent thrust boundaries, b) Prominent strike-slip boundaries; 15 – Course of block extrusions; 16 – state border of Slovak Republic.

suture is represented by the Pieniny Klippen Belt (PKB) zone and cogenetic tectonic boundaries. PKB is described as a zone of extensive shortening and strike-slip shearing (e.g. Plašienka et al., 2020 and references therein). The sedimentary basins with the Upper Cretaceous, Paleogene and Neogene filling and neo-volcanic complexes represent the Neo-Alpine formations superimposed on Paleo-Alpine nappe system. The dynamic evolution of the Western Carpathians resulted in the Neogene sedimentary basins genetic variation. Depending on their geodynamic position within the orogenic belt, the fore-arc, inter-arc and back-arc basins are present (Kováč et al., 2016, 2017). There occur basins formed by lithospheric extension – thermal subsidence, flexure and strike-slip related basins (Vass, 1979, 1998; Kováč, 2000; Janočko et al., 2003a, b). The thin Penninic oceanic crust subducted during the oblique convergence, being melted in the upper mantle, thus providing a source for extensive subsequent volcanism, situated at the frontal edge as well as in the interior of the overriding crustal slab (e.g. Lexa & Konečný, 1998, Lexa et al., 2010).

The current morpho-structural character and shape of orogenic belt was to the Western Carpathians imprinted during the Neo-Alpine tectonic period. The shape of the Carpathian orogenic belt was constrained by the pre-collision shape of thin crust embayment of the flysch basin inside the stable North European Platform (NEP). The eastwardly prograding crustal segment of Internides (IWC) was broken into several different fragments, which underwent large translations, rotations, uplifts and subsidence, including tilting during the occupation of oceanic crust embayment in NEP (Marko et al., 2017; Bezák et al., 2020). This – with combination of astenolith arise and extension resulted in development of specific morpho-tectonic features, including alternating intra-montane sedimentary basins and core mountain horsts, structural bending, fan structures and robust Miocene volcanic activity; all peculiar particularly to the Western Carpathians.

The Western Carpathian part of the Alpine orogenic belt is recently generally inactive, because the driving force of collisional dynamics – the subduction and tectonic escape processes have already ceased in the Late Tertiary. This is the reason why the recent movements (max. first few mm/yr) and Neo-tectonic activity are very moderate, reflected in the weak intensity, character and distribution of earthquakes (Cipciar et al., 2016; Hók et al., 2016). The earthquakes are generated on the faults and fault zones, controlling relaxation post-collisional movements of individualized IWC blocks (Marko et al., 2017). Micro-earthquakes are prevailing, rare macro-seizmic events reach an intensity of M 2.9. The clustering of more important macro-earthquakes, related to large faults, has been recorded only in a few areas (Dobrá Voda, Žilina, Kolárovo, Komárno). The strongest recorded earthquake (1906) in the Dobrá Voda area had an intensity of M 5.7.

For the Western Carpathians, there is typical a distinctive polarity of the orogenic final overthrusts at the front of the Western Carpathians loop (Jiříček, 1979; Matenco & Bertoli, 2000). The active collisional front moved from the west to the east, and resulted in a complex, heterogeneous, polyphasic and diachronous structure of the Carpathian loop (Unrug, 1984). The same character has the Pieniny Klippen Belt structure (e.g. Andrusov, 1974; Birkenmajer, 1986; Plašienka, 2018b), on the border of IWC and OWC.

The crustal thickness (the Moho depth) of the Western Carpathians (Bielik et al., 2018 and references therein) ranges from 25 to 42 km. Its typical feature is that the thickness of the crust rises from south to north. While the southern parts of the Western Carpathians (IWC) are characterized by a thickness of only about 25 to 33 km, the northern parts (the Central and Outer Western Carpathians) by thicker crust (35–40 km). The thinnest crust of 25 km is observed beneath the Danube basin. On the contrary, the largest crustal thickness (Janík et al., 2011) in the Western Carpathians was measured northeast of the Vysoké Tatry Mts., which are the highest mountains of the Carpathians. In general, however, the Western Carpathian orogen is significant by crustal thickening also in comparison with Himalayan belt.

The crustal thickness of the Western Carpathians correlates very well with the thickness of the lithosphere-astenosphere boundary (LAB). The thickening of the lithosphere in the south-north direction can also be observed. The IWC are accompanied by a thinner lithosphere of about 100–120 km. A slightly thicker lithosphere can be observed in the northern part of IWC and Outer Western Carpathians. An interesting pattern of the Carpathian lithosphere is its thickening also along strike of the Carpathian arc, when in the Eastern Carpathians the LAB reaches up to 240 km (Zeyen et al., 2002; Dérerová et al., 2006).

Himalayas

Geomorphologicaly and structuraly the most spectacular segment of Tethyan Alpides is the Himalayan belt, one of the youngest gigantic mountain ranges of the world. This is an example of strongly polarized asymmetric, southvergent collisional orogen (**Fig. 3, 4b**). The high ranges of Himalayas were formed due to the Indian shield (a continental part of Indo-Australian plate) northward penetration into mega-embayment of the Tethys ocean in the Euroasian plate (e.g. Gansser, 1966; Golonka, 2000). The Alpine-Himalayan mobile belt is in various parts diachronous and heterogeneous, representing different final stages of Wilson cycle (sensu Dewey & Burke, 1974). Some parts are evolved between already collided continental plates, some parts are situated in segments, where the oceanic crust was not completely consumed by subduction and the convergence still continues (Java, Fiji, etc.). The Himalayas represent the peculiar terminal stage of the Wilson cycle. The Himalayas have evolved due to the closure of Meso-Cenozoic ocean floor – the process, which did not terminated by continent-continent collision, but continental lithosphere of Indian plate after subduction of Tethyan ocean floor and initial collision (ca 50 Ma ago) also subducted ca 700 km under Euroasian plate (Dadlez & Jaroszewski, 1994; Lyon-Caen & Molnar, 1983). It is a very rare Ampferer´s A-type subduction (sensu Bally, 1981), because continental lithosphere usualy does not undergo subduction, what is one from the basic paradigms of plate-tectonic concept. Nevertheless this continental plate subduction resulted in grandious crustal thickening, which is responsible for the highest uplifts, creating the highest mountains and plateaus in the world. From this point of view the Himalayas represent a speciffic collisional orogen, which is characterized by the continental crust duplexing of the underthrusting Indian crust and the overthrusting Euroasian crust (Yeats, 2012). Similar style is typical for the Alps.

Based on tomographic inversion of regional earthquake data (Koulakov et al., 2015) and receive function image (Nábělek et al., 2009; Subedi et al., 2018) it was found

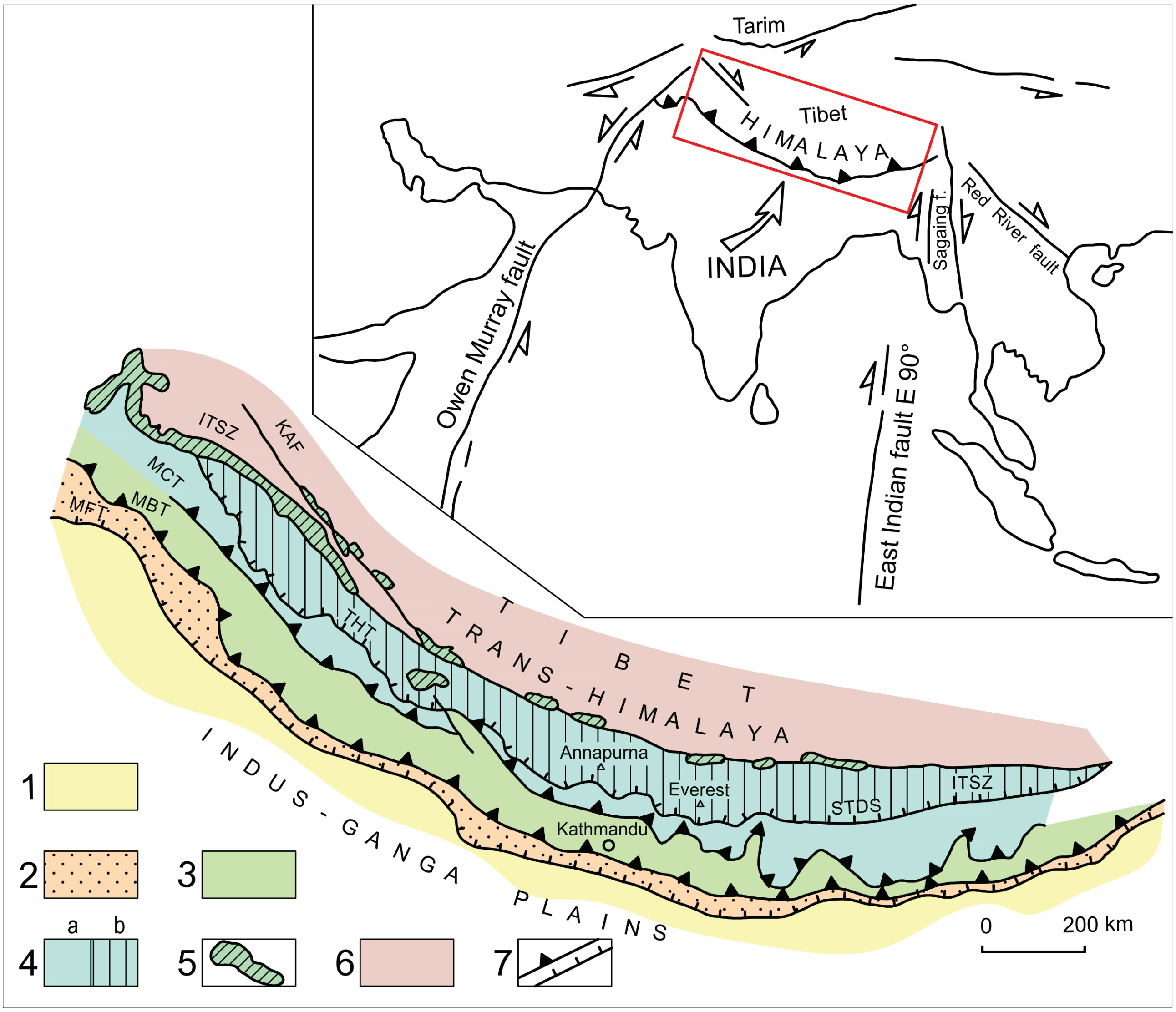


Fig. 3. Neo-Alpine architecture, dynamics and tectonic division of the Himalayas (after Valdiya, 1992, modified). Abbreviations: MFT – Main Frontal Thrust, MBT – Main Boundary Thrust, MCT – Main Central Thrust, THT – Trans-Himmandry Thrust (in India), STDS (in Nepal), ITSZ – Indus–Tsang-Po Suture Zone, KAF – Karakoram Fault. Explanations: INDIAN PLATE: 1 – Indian plains of Quaternary alluvial deposits covering ancient not mobilized craton; 2–5 – Part of craton mobilized in orogenic accretionary prism: 2 – Sub-Himalaya – Miocene-Pleistocene molasse sediments of Siwalik Group; 3 – Lesser Himalaya; 4 – a) Higher Himalaya, b) Tibetan Tethys Zone; 5 – Ophiolites; ASIAN PLATE: 6 – Trans-Himalaya (Tibetan Himalaya); 7 – Prominent thrust boundaries, detachments.

out that the thickness of the crust beneath the Nepal Himalayas varies from ~40 to ~75 km. It increases across the Himalayas from the south to the north. Below the foothills (the Himalayan Main Frontal Thrust – MFT) and the Lesser Himalaya the crustal thickness is about 40 km, but beneath the Higher Himalayan range and Central Tibet Plateau (Lhasa and Qiangtang Blocks) it reaches already 65–75 km.

Despite the fact that the results related to the position of the LAB in the Himalayan region differ (e.g. Zhao et al., 2010; Xu et al., 2011; Deng et al., 2014) it can be generalized that the depth of the LAB increases in direction of the underthrusting of the Indian plate beneath the Eurasian plate. However, a change in lithosphere thickness can also be observed along the Himalayan orogen. According to S receive function results (Xu et al., 2011), the LAB under the MFT and the Lesser Himalaya ranges at depths of ~80–120 km, while under the Higher Himalayan range and the Central Tibet Plateau it reaches values up to ~180–240 km. Deng et al. (2014) suggest that the lithosphere can have a thickness of up to 300 km below the Central Tibet Plateau.

Approximately 1500 km northward drift of Indian plate and its invasion to Euroasian plate embayment was controlled by the first order N-S transcurent dislocations Owen Murray and East Indian rift of 90° longitude. Owen Murray fault operated as a sinistral strike-slip and East Indian rift was reactivated as dextral strike-slip (Ollier & Clayton, 1984). These two block boundary faults – lateral ramp faults allowed invasion of Indian continental plate into the large Tethys ocean embayment within the Euroasian plate. Indian shield was separated from the Antarctica plate in the Early Cretaceous (ca 120 Ma) and thereafter drifted to the north; and Indian ocean floor subducted under the Euroasian plate (B-type subduction sensu Bally, 1981). Indian ocean floor subduction ceased beneath Indian–Euroasian collision zone during the peak of collision (40 Ma ago). Collision itself was diachronous (Searle, 1996), occurred earlier in the western part (Pakistan, 60 Ma), later in the eastern part (Southern Tibet). After initial collision of Indian shield with Euroasian plate margin (ca 50 Ma ago, Golonka et al., 2006) the Himalayan orogen nappe architecture formation has started and continental lithosphere of Indian schield subducted further under the Euroasian plate (A-type subduction sensu Bally, 1981). This process of nappes formation culminated in the Middle Miocene (Jaroš & Kalvoda, 1977) and it has operated continuosly up till recent and the Himalayas are recently still in the active zone of plates convergence (recently ca 50 mm/yr by Minster & Jordan, 1978; 36–40 mm/yr by Sockuet et al., 2006 ex Yeats, 2012). This extreme dynamics is responsible for high seismicity related predominantly to southvergent thrusts (MFT, less MBT) separating the main Himalayan nappe megaunits, less to tear wrench faults of these thrusts (Valdiya, 1992). The magnitudes of frequent and strong present-day and historical earthquakes reach average intensity M 7–9 (Valdiya, 1992; Yeats, 2012). As the strongest seismic events in the Central Himalayas are regarded historical superquakes 1505, 1835 and 1934 with estimated magnitudes M 8.2, 7.7 and 8.1 respectively (Pandey & Molnar, 1988; Yeats, 2012,). The last catastrophic earthquake with magnitude 7.8 affected Central Nepal Himalayas in 2015. The epicenters of earthquakes are generally clustered along the Himalayan front (MFT), Main Central Thrust and in Tibetan plateau along strike-slip faults controlling eastward escape of the Lhasa block. Currently the most seismically active is the Himalayan frontal area, where the plate boundary earthquakes are related to the Himalayan Main Frontal Thrust. For the earthquakes epicenters following MCT inside the Himalayan range is responsible the Main Detachement Fault (MDF) – the continuation of the Himalayan Frontal Thrust, which is situated deeper in the crust under MCT. MCT itself as well as MBT thrusts are recently inactive. Earthquakes are generated in depth due to frontal ramps on MDF, no seismogenic surface ruptures were described there. The Himalayas and Tibetan plateau are geneticaly tigthly related. The long lasting nortward movement of Indian plate into Euroasia lifted up the Himalayan range and affected Tibetan block, which one has been extruded eastwardly along the dextral Karakoram and sinistral Altyn Tagh strike-slip faults. These highly dynamic ruptures with the slip rate estimated to 30 mm/yr in average (Taylor & Yin, 2009) have been the source of strong earthquakes in Tibetan plateau.

The last gravitational nappes of Tethyan sedimentary sequences were thrust over the Great Himalaya crystalline basement in the Middle and Late Pleistocene. A very young Kathmandu nappe system – a Higher Himalayan crystalline slab was thrust over the Lesser Himalaya units. From the Holocene a vertical component of movements – uplifts has been prevailing (Kalvoda, 1978).

Generally, the tectonic activity during the Himalayan orogeny migrated southwards (Bogacz & Krokowski, 1983). The 50 Ma lasting collision and underthrusting of Indian continental plate result in grandious southvergent nappe architecture of Himalayan orogen and extreme uplifts (l. c.). The crustal shortening due to collision has been mainly absorbed by the northern margin of the Indian plate (Bogacz & Krokowski, 1985). In process of collision of the Indian and Asian plates, the northern continental margin (current Himalayan range) of the Indian plate was split into nappes and blocks by intracrustal thrusts and strike-slips. Continual propagation of Indian plate to the north resulted in great shortenning (estimated minimum 400–500 km by Kalvoda, 1976, 1978; 600–700 km by Le Fort, 1975; even 1500 km by Bouchez & Pecher, 1981) which was accommodated by thrusts and their tear faults. One of the largest movements were along MCT.

The Himalayas are divided to several longitudinal tectonic zones – superposed megaunits separated by boundary thrusts (Mísař, 1987; Valdiya, 1992) listed below from the south towards the north (**Fig. 3**).

i) Indo-Gangatic Plain (Foreland Basin)

This zone represents the northern border of the Indo-Gangetic alluvial plain (Indus–Ganga lowland) and forms the southernmost tectonic zone of Nepal (Upreti, 1999). It is delimited by the Main Frontal Thrust (MFT) to the north, which is exposed at many places. At many places along this thrust, the Churia rocks are exposed over the Terai sediments. Terai plain gradually rises from 60 m above the sea level in the south to more than 200 m in the north. It is covered by Quaternary to recent sediments which are about 1500 m thick. The recent alluvium is mainly derived from the Churia Hills (Siwaliks) and also from the Lesser Himalaya by the river systems.

ii) Sub-Himalaya (Siwalik)

The Sub-Himalaya is an autochthonous unit formed by Middle Miocene-Early Pliocene to Pleistocene molasse sediments (Yeats & Lillie, 1991; Upreti, 1999) filling the Himalaya foredeep basin. These are a few thousand meters (up to 6 km) thick fluviatile sediments having the source area in uplifting Himalaya (Sigdel et al., 2011). Siwalik Group covers crystalline basement of Indian shield and itself is covered by hundred meters or even a few km thick fluvial Holocene sediments of Indus-Ganga lowland, and at the north it was in the Middle Miocene overthrust by the Lesser Himalaya nappe unit along the Main Boundary Thrust (MBT; Valdiya, 1992). MBT is recently reactivated as oblique dextral strike-slip (Yeats et al., 1992). Between the Sub-Himalaya´s deformed Siwalik Group and stable Indian plate (Indian plains) is also a tectonic boundary represented by the Main Frontal Thrust (MFT; Yeats et al., 1992). MFT is near the surface steeply dipping zone of southvergent reverse faults and thrusts, respectively. According to Bogacz & Krokowski (1985), the MFT thrust was in the later stages of collision reactivated in the western part as a dextral strike-slip and in the eastern part as a sinistral strike-slip. It was caused by the arcue shape of India–Tibet contact zone.

iii) Lesser Himalaya

This old mature, but recently rejuvenized terrain was thrust along MBT over the Outer Himalayan Siwalik Group in the Quaternary. Nappe mega-unit of Lesser Himalaya is composed of various metamorphic, as well non-metamorphic formations of wide stratigraphic range, from pre-Cambrian to Early Miocene. Lesser Himalaya is subdivided to three lithotectonic assemblages – superposed units:

a) the parautochthonous Proterozoic sedimentary rocks in the lower part, overthrusted by
b) sheets of low-grade metamorphics associated with ca 2 Ga old granites, which are overthrusted by
c) unit of medium-grade metamorphics intruded by ca 550 Ma old granitoid bodies.

The highest nappes represent isolated remnants – outliers of Higher Himalaya nappe unit, probably gravitationaly slided from the Higher Himalayas realm due to uplifts.

The Lesser Himalaya is separated from the Higher Himalaya by the Main Central Thrust (MCT).

iv) Higher Himalaya (Great Himalaya)

This is the huge tectonic slab of pre-Cambrian highly metamorposed and granitized crystalline basement (crystalline Tibetan slab, Himalayan gneiss zone) with its Tethyan Mesozoic cover sediments (Tibetan Tethys Zone) thrusted along Main Central Thrust (MCT) over sedimentary units of the Lesser Himalaya. These Indian plate margin deposits, where limestones dominate, are intensely folded to overturned, even recumbent folds and thrusted (Bogacz & Krokowski, 1983). Crystalline basement of Great Himalaya is composed of high-grade metamorphic rocks intruded by Mid-Tertiary granites, representing continental margin on which were deposited sediments of Tethyan sea. The basement crystalline complex is separated from the Tethyan Late Precambrian to Late Cretaceous sedimentary cover by detachment Trans-Himmandri Thrust (THT; Valdiya, 1992), South Tibetan Detachment System (STDS) respectively. The STDS is currently interpreted as a normal fault-shear zone. The Higher Himalaya with the southernmost part of Tibetan Tethys Zone is the best known and most attractive owing to occurence of highest peaks of the world. Some of them are composed of Tethyan Late Paleozoic–Mesozoic–Eocene non-metamorphosed sedimentary rocks (Everest – Sagarmatha, Annapurna, Dhaulagiri). Although the main thrusting events were pre-Eocene and pre-Miocene, Pleistocene reactivation of thrusting was recorded and present activity is confirmed by repeating seismic events related mainly to tectonic contact of the Higher Himalaya nappe with Lesser Himalaya nappe.

v) Tibetan Tethys Zone

The northernmost tectonic zone of the Himalayas occupies a wide belt consisting of sedimentary rocks known as the Tibetan Tethys Zone (TTZ). The Tibetan Tethys Zone lies between the South Tibetan Detachment System – STDS [Trans-Himmandri Thrust – THT (sensu Valdiya, 1992) respectively] and the Indus–Tsang-Po Suture Zone (ITSZ). STDS is interpreted as a north dipping normal fault-shear zone. TTZ has undergone very little metamorphism, except at its base where it is close to the Higher Himalaya Zone. The rocks of TTZ consists of thick and nearly continuous Lower Paleozoic to Lower Tertiary marine, highly fossiliferous sedimentary successions including slate, sandstone and limestone. The rocks are considered to have been deposited in a part of the Indian passive continental margin (Liu & Einsele, 1994). The Tibetan Tethys Zone formations are extensively folded (Fig. 4b) due to the extreme shortening in a proximity of the India–Asia contact zone (ITSZ). Folding in competent Tethyan sedimentary rocks accommodated a great deal of this shortening, while the rigid crystalline fundament of the upper crust was mainly thrusted. This explains for the controversy in describing the Tibetan Tethys Zone southern boundary as a thrust (Trans-Himmanry Thrust; Valdiya, 1992), or a normal fault (South Tibetan Detachment System; Burchfiel et al., 1992; Hodges et al., 1992). Both structures represent the same block interface, the southern boundary of TTZ.

The Himalayan orogenic belt represents a piggy-back thrust sequence where the younger thrusts propagated towards the front of orogenic belt. At the first stages of India–Asia collision the Tethyan sediments of Indian plate margin were folded, detached from the basement and thrusted, creating embryonal accretionary prism. At this period the southern tectonic boundary of TTZ operated as a thrust (Trans-Himmandri Thrust). Continual India–Asia convergence triggered development of younger thrusts (MCT, MBT, MFT). MCT accommodated crustal shortening by overriding of Higher Himalaya nappe, composed of crystalline basement, over the Lesser Himalaya formations. Long lasting plates convergence, producing crustal shortening of Indian plate, triggered gradual development of MBT and finally the MFT detachement zones, meanwhile the crustal slab of the Higher Himalaya was still pushed-up and extruded-up respectively. This dynamics of pushed-up Higher Himalaya terrane was controlled by the thrust kinematics of MCT and "normal" kinematics of THT, which is described currently as the South Tibetan Detachment System. So the southern tectonic boundary of TTZ operated as a thrust as well as a normal fault, in both cases in conditions of strong compression (Kellett et al., 2018).

vi) <u>Trans-Himalaya (Tibetan Himalaya)</u>

This terrane of Asian plate is uplifted plateau composed of Tethyan formations. Stratigraphical diapason of units is pre-Cambrian–Middle Eocene (sedimentary sequence itself is Early Paleozoic–Eocene; Jaroš & Kalvoda, 1978). Tibetan plateau (highland) is separated from the Great Himalaya by the most pronounced first order crustal dislocation Indus–Tsang-Po Suture. The Indus–Tsang-Po Suture represents tectonic contact of Gondwana and Euroasia plates, being a result of Himalaya–Tibet collision. Within the steeply dipping tectonic zone of recently already inactive Indus–Tsang-Po Suture Zone (ITSZ) the members of ophiolite formation are localized, representing the obducted remnants of subducted Tethyan ocean floor. The northern boundary of Tibetan plateau is represented by the Altyn Tagh strike-slip fault, controlling together with the Karakoram fault an eastward escape of Tibetan block and creating its contact with the Tarim basin.

Lateral extrusions, resp. tectonic escapes (Tapponier et al., 1986; Cobold & Davy, 1988) in collisional orogens result from the geometry of converging plate margins. This tectonic style, typical for Alpine type orogens, occurs in both peripheries of Himalayas (Pelzer & Tapponier, 1988). Collision and suturing of Indian plate to Asian plate triggered extensive strike-slip faulting in Asian plate (Knopp, 1997; Yeats, 2012). Due to movement of the Indian plate to the Eauroasian plate, the Tibetan plateau extruded along the sinistral Altyn Tagh and dextral Karakoram boundary strike-slip faults from collisional zone towards the east and the Indo-China and South China micro-plates were along the Red river and Arakan-Yoma strike-slip faults extruded southeastward and southward (Golonka et al., 2006). Blocks of the Pamir Mts. and Hindukus Mts. foothills plateau were extruded towards the west along the Quetta-Chaman and Herat strike-slip boundary faults (e.g. Tapponier & Molnar, 1977; Cobbold & Davy, 1988).

Himalayan range is affected by systematic faulting, developed after folding and thrusting period, which finished in the Middle Miocene (Bogacz & Krokowski, 1983). Faulting is genetically associated with uplifting of the Himalayan range. There are longitudinal and transversal fault systems. Longitudinal faults represent orogen-parallel strike-slips generated due to indentation of Indian shield and its CCW rotation. Transversal faults display also strike-slip component of the movement, but dip-slip normal movements are prevailing (Bogacz & Krokowski, 1983, 1985). Transversal rivers cutting the Main Himalayan ridge follows these fault damage zones, and subsidence of intramontane Plio-Quaternary sedimentary basins is controlled by this youngest population of normal faults (e.g. Fort et al., 1982). This is the case of the Kali Gandaki river valley controlled by Thakkhola fault system interrupting the main Himalayan ridge in between the Dhaulagiri and Annapurna massifs.

A common features and contrasts of the Western Carpathian and Himalayan tectonic architecture

Both orogens are the result of the Tethys ocean closure followed by diachronous continent-continent collision, started in the Himalayas during the Paleocene/Eocene boundary and in the Western Carpathians in the Early Miocene. The Western Carpathians are north-vergent, while the Himalayas are south-vergent (**Fig. 4**). The Western Carpathians represents the northern branch of the bilateral symmetric Paleo-Alpine orogenic wedge – mega-flower structure and Dinarides are its southern south-vergent branch. Carpathians and Dinarides are separated by the Pannonian central block (Dadlez & Jaroszewski, 1994). The Himalayan collision was frontal, while the Western Carpathian one was oblique, typical for the strike-slip orogens (Badham & Halls, 1975). Different is also the rate of ocean floor subduction, which was much higher in the case of Himalayas. Ocean floor spreading driving the northward drift of Indian plate was estimated as much as 20 cm a year, while plate motion rates in the Carpathian realm were an order smaller. It is reflected in geomorphology and mountain altitudes – the Higher Himalaya relief is the most extreme one over the World (**Fig. 5**). High plate motion rates should be responsible as well for the tremendous crustal thickening by continental crust duplexing, which is the rare phenomena, specific for Himalayas and the Alps. Except of tectonic reasons as piling of nappes, the intense isostatic movements due to crustal thickening caused the extreme uplift.

Common features of both orogens is their Neo-Alpine nappe architecture, though involving different lithological units. The Western Carpathians comprise thrusted Tertiary sedimentary sequences of orogenic accretionary wedge, while the Himalayan architecture involves also huge nappes of deep crystalline units creating the basement of the main Himalayan ridge and Mesosoic Tethyan units. We had an opportunity to observe these units and stuctures of both orogens in situ (**Fig. 6**). In the Western Carpathians, the basement nappes comprising crystalline basement are Paleo-Alpine (Cretaceous). Meanwhile tectonic evolution of Himalayas has been continual Tertiary-Quaternary process of plates collision, the Carpathians evolved during several tectonic stages divided by long lasting periods of extension and denudation. This is the reason of differences in structure, crustal thickness and morphology of both orogens.

The tectonic activity in both orogens displays a distinctive polarity of thrusting, which migrated from the internal to the external parts of the orogenic belt and from the west towards the east. In the Himalayas all thrusts are Neo-Alpine and active even in the Quaternary period till the present time. So the Himalayas are still active collisional zone/tectonic suture, which is evidenced by strong seismic activity and extensive recent vertical and horizontal movements. In the Western Carpathians the Inner block/terrane is formed by Paleo-Alpine (Late Cretaceous) nappes. The Neo-Alpine nappes, recently already inactive, create accretionary wedge of the Outer (External) Carpathians. The collision in the Western Carpathians has already ceased, recent moderate seismic activity is related to movements on relaxation faults. Recently active is the southern branch of European Alpine orogen – Dinarides, evidenced by strong seismic activity.

Shape of both orogenic arcs depends upon the pre-collisional geometry of foreland plate margins. It was produced by escape tectonics, controlled by strike-slip faults, which is a common feature of many segments of Tethyan Alpides, Carpathians and Himalayas including (Yin & Taylor, 2011).

Tectonic sutures after ocean crust subduction are traced in both orogens by occurences of ophiolites – the remnants of oceanic crust. Contrary to the huge Himalayan ophiolite belt following the Indus–Tsang-Po suture closed in Paleogene period, there are in the Western Carpathians the ophiolites preserved only rudimentarily (Meliata unit), but they are related to Paleo-Alpine subduction. Voluminously large occurrences of ophiolites comparable with Himalayan ones are situated in the Paleo-Alpine Vardar zone of the southern branch of the European Alpine belt.

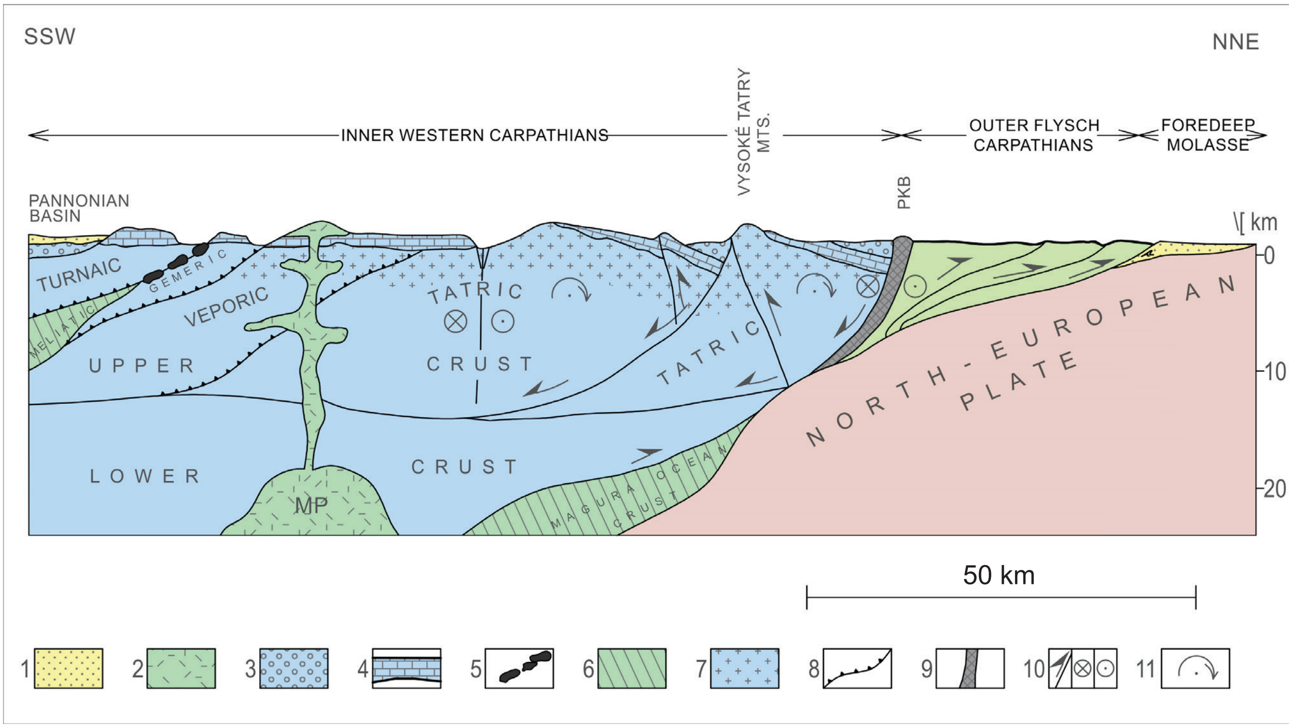
The noticeable differences are in dimensions and recent movement rates and magnitudes – all are an order higher in the Himalayas, where collision has not finished yet.

Difference is also in magnitudes and origin of seismicity in both orogens, resulting from the character of collision and its maturity. While the Himalayan earthquakes are strong, focussed in great depths and related to the syn-collisional thrust faulting, the Western Carpathians earthquakes, resp. micro-earthquakes are much weaker and they are related mainly to post-collisional relaxation strike-slip, less dip-slip faults with moderate slips.

Almost all values of orogenic parameters are lower in the Western Carpathians than in the Himalayas, except the volume of subsequent Neo-Alpine volcanism, which is extensively developed in the Western Carpathians, while in the Himalayan orogenic accretionary prism not. On the other hand, the northern terrains of Himalayas are massively intruded by the Miocene granites exhumed due to extensive erosion and the extreme terrain morphology; and massive Tertiary volcanism is situated in the Lhasa block of Asian plate. This magmatism in both orogenic belts is related to melting of subducted crust.

For the internal part of the Western Carpathians is typical Neo-Alpine basin and range structure controlled by faulting and related block rotations and tilting. Neogene intramontane sedimentary basins of this type were not developed in the Himalayas due to extreme uplift, deep erosion and lack of extension. A mantle astenolith is not developed in part of the Himalayan orogenic accretionary prism. However, terrestrial Mio-Quaternary sediments were deposited in narrow fault controlled deep grabens (e.g. Mustang graben) crossing the main Himalayan structural direction (Adhikari & Wagreich, 2011).

Specific tectonic development of the Western Carpathians and Himalayas is also recorded in the gravity field (Bouguer anomalies) difference. On the one hand, both orogens create a regional negative Bouguer anomaly, which is a typical accompanying phenomenon of collisional orogens. However, the difference between the low gravity value of the Himalayas and the Western Carpathians is extraordinary. While the gravity low of the Himalayas reaches a maximum amplitude of almost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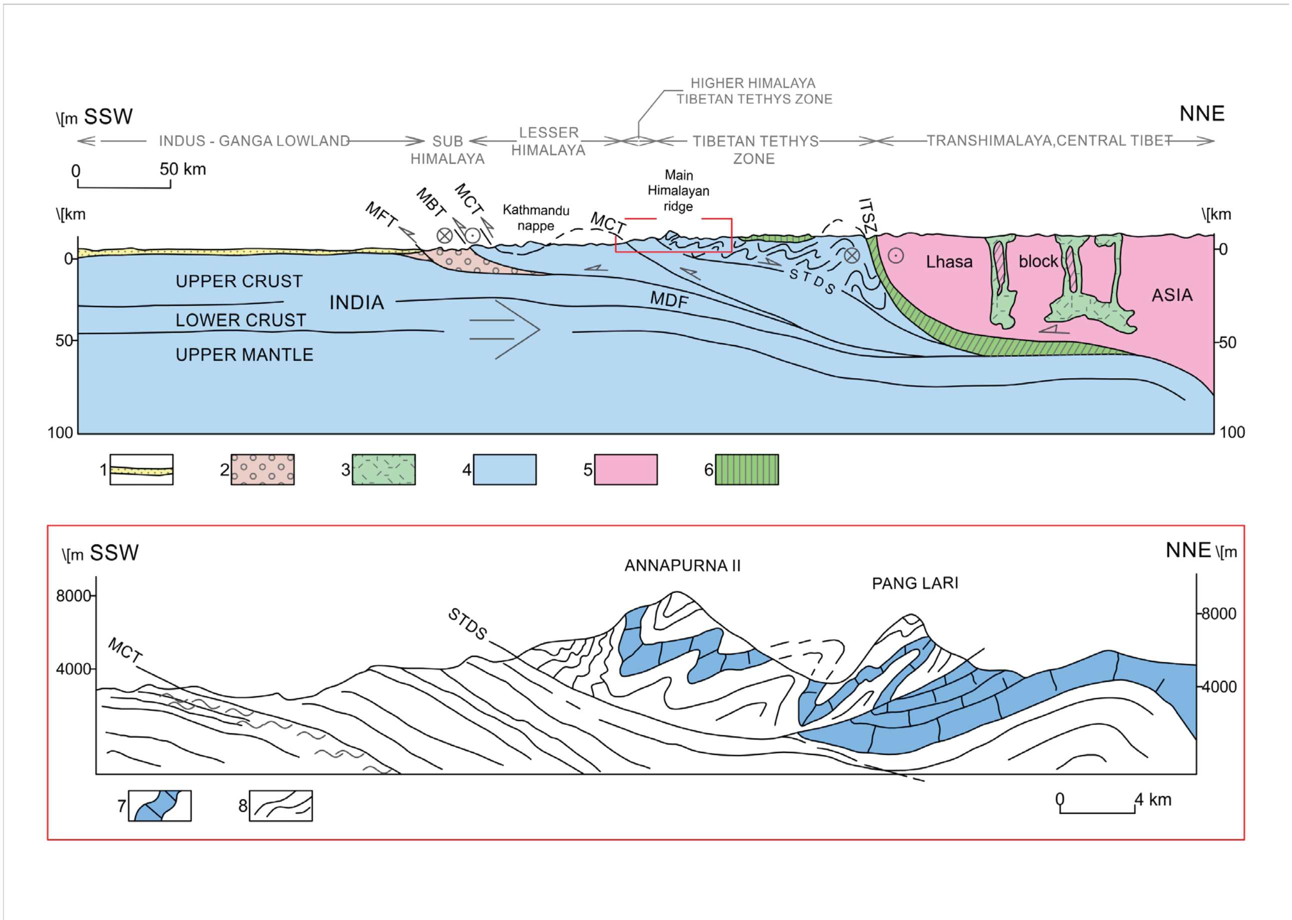
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HIGHER HIMALAYA
TIBETAN TETHYS ZONE
INDUS - GANGA LOWLAND
SUB HIMALAYA
LESSER HIMALAYA
TIBETAN TETHYS ZONE
TRANSHIMALAYA,CENTRAL TIBET
0
50 km
km
MFT
MBT
MCT
Kathmandu nappe
MCT
Main Himalayan ridge
ITSZ
STDS
UPPER CRUST
INDIA
LOWER CRUST
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STDS
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8000
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Fig. 4. A – Conceptual and generalized geological cross-section of the Western Carpathians. Scale is approximate and some structural phenomena are due to better readability exaggerated. Abbreviations: MP – Mantle plume (astenolith), PKB – Pieniny Klippen Belt. Explanations: 1 – Neogene sediments; 2 – Neogene volcanites; 3 – Inner Carpathian Paleogene sediments; 4 – Late Paleozoic-Mesozoic cover units and Paleo-Alpine superficial nappe units of Mesozoic sequences; 5 – Ophiolites – remnants of subducted ocean crust of Meliatic unit; 6 – Ocean crust slabs; 7 – Variscan (Hercynian) Paleozoic crystalline basement units; 8 – Paleo-Alpine thrusts boundaries of basement nappes; 9 – Pieniny Klippen Belt – a Neo-Alpine suture zone; 10 – Neo-Alpine (Miocene) structures: a) thrusts, reverse and normal faults, b) strike-slip faults; 11 – block tilting. **B –** Generalized geological cross-section of the Central Himalaya (modified, compiled according to Gansser, 1980; Nabělek et al., 2009; Yeats, 2012; Yeats et al., 1992; Le Fort, 1975; Valdiya, 1992; Yeats & Thakur, 2008; Bagacz & Krokowski, 1983; Tapponier et al., 1982, detail of Annapurnas segment after Pecher, 1976; Le Fort, 1981). Some structural phenomena are due to better readability exaggerated. Abbreviations: HFT – Himalayan Frontal Thrust, MBT – Main Boundary Thrust, MCT – Main Central Thrust, THT – Trans-Himmandry Thrust, STDS – South Tibetan Detachment System, ITSZ – Indus–Tsang-Po Suture Zone. Explanations: 1 – Holocene alluvial sediments of Indo Gangatic Plain; 2 – Miocene-Pleistocene molasse sediments of Siwalik Group; 3 – Cenozoic volcanites; 4 – Indian (Gondwana) Plate; 5 – Euro-Asian Plate; 6 – Ophiolite mélange – Tethys ocean crust incorporated to collisional suture (ITSZ), or obducted remnants; 7 – Tethyan Paleozoic limestones; 8 – Quartzites, schists and gneisses of Indian craton.

–600 mGal (Sandwell & Smith, 1997; Shin et al., 2007) whereas in the Western Carpathians it is only about –70 mGal (Bielik et al., 2006). Equally different are the values of the wavelengths of the gravity lows in both mountains. The Himalayan gravity low attains a value of about 500 km, while the Western Carpathian low gravity is characterized by 50–100 km. Lillie et al. (1994) calculated on the basis of a kinematic model of ocean basin closure and subsequent continental collision (Lillie, 1991) that the Western Carpathian narrow width of the gravity low suggests the continental convergence ceased soon after the ocean basin closure. So that only about 50 km of continental crustal shortening occurred in the Western Carpathians. The amplitude of the Western Carpathian gravity low further indicates small crustal root (on average 35 km with a maximum of 42 km). Taking into account the Lillie's model (Lillie, 1991) it can be suggested that the continental collision between the Indian and Eurasian plates was much stronger. The width of the Himalayan gravity low assumes that the Indian plate was after Tethys ocean closure underthrust beneath the Eurasian plate about 500 km. The amplitude of the gravity low indicates 70–80 km crustal root under the Himalayas. Which is in good agreement with seismic observations (e.g. Nábělek et al., 2009; Zhang et al., 2011; Koulakov et al., 2015; Subedi et al., 2018) and geophysical crustal models (e.g. Munt et al., 2008; Tenzer et al., 2015).

Discussion

Both compared distant orogenic belts of the same equatorial Alpidic orogenic system apparently belong to Intra-Pangea subduction-collisional zones (sensu Németh et al., 2016, 2017, 2018). Nevertheless, there are some pecularities and differences in their tectonic architecture (see Tab. 1). The most prominent difference is in complexity of their tectonic evolution.

The Western Carpathians have evolved during the three Wilson cycles (Variscan, Paleo- and Neo-Alpine), producing double collisional crustal thickening (Variscan and Paleo-Alpine), and related double unroofing during the post-collisional relaxations (Németh et al., 2016). So the WC structure is a result of multiple alternation of convergent and divergent geotectonic processes.

The Himalayas as a distinct mountain range is much simpler – they completely represent a Neo-Alpine structure, being the result of giant long lasting continual collision not interrupted by the relaxation extensional periods. The recent unroofing has taken place in Himalayas due to an extreme uplifting, related to the collision and isostatic forces, which have triggered the gravitatioanal nappes sliding during unroofing.

The distinct contrast between the Western Carpathians and the Himalayas during the Cenozoic Neo-Alpine evolution is the oposite vergency of subduction and following thrusting in orogenic collisional prism. While in the Western Carpathians the Neo-Alpine orogenic structure has evolved as forward thrusted nappes in the frontal rim and in the front of prograding overriding plate (IWC, ALCAPA respectively); in the Himalayas, the accretionary orogenic wedge evolved from backward thrusted (in relation to plate movement) tectonic slices detached from the subducting-underthrusting Indian plate.

Taking into account the wider regional relations – in the north-located zone of Tibetan Plateau, neighboring the Himalayas, there are known several parallel suture zones (cf. e.g. Chung et al., 2005; Zhu et al., 2013), whose geodynamic evolution could be parallelized with Variscan and Paleo-Alpine processes known in the western segment of Alpine-Carpathian-Himalayan orogenic belt, incl. Western Carpathians. This indicates a principle of pulsing (multiple repeated) divergent and convergent processes of tectonic evolution, valid in the whole orogenic belt of Intra-Pangea type (Németh, pers. com.).

Disregard the opposite vergency, concerning the Neo-Alpine processes, we can try to compare geometrically tectonic terranes of both orogens according to their

Tab. 1
A comparison of the Western Carpathian and the Himalayan geological features – a summary.

COMMON FEATURES		
PARAMETERS OF OROGEN	**WESTERN CARPATHIANS**	**HIMALAYAN BELT**
Affiliation to world orogenic system	Both orogenes belong to the same global Neo-Alpine equatorial Alpidic collisional orogenic system, which is the result of the Euroasian plate collision with the Gondwana continental fragments – microplates after closing the Tethys ocean.	
Regional Bouguer anomaly	Both orogens create a regional negative Bouguer anomaly, which is a typical accompanying phenomenon of collisional orogens.	
Symmetry of orogen	Structure of both orogenic belts is strongly asymmetric. In front of both orogens are well developed foredeep molasse basins feeded by clastic material coming from the growing orogen.	
General tectonic style	Convergent-collisional style, shortennig is accomodated by thrusting and folding; as well as extrusions controlled by wrench faulting.	
Polarity of orogen	Continent-continent collision in Alpidic orogenic system was diachronous, both orogens display distinctive polarity of tectonic activity – thrusting, migrated from the internal to external parts of orogen and from the west eastward.	
Syn-orogenic magmatism	Crust subducted during the plate convergence was in both orogens melted in the upper mantle, thus providing a source for extensive subsequent Cenozoic magmatism and volcanism.	
Pre-collision constraints of orogen loop geometry	Shape of both orogenic arcs depends upon the pre-collisional geometry of foreland plate margins – an embayments of an ocean crust situated within the Euroasian plate.	

CONTRAST FEATURES		
PARAMETERS OF OROGEN	**WESTERN CARPATHIANS**	**HIMALAYAS**
Type of subduction	B-type subduction of the Magura basin thin lithosphere under progressing extruded Inner Western Carpathian micro-plates, followed by oblique continent-continent collision.	Benioff´s B-type subduction of the Tethys ocean lithosphere was after collision followed by the Ampferer´s A-type subduction – an underthrusting of the Indian plate continental lithosphere under the Asian plate resulting in the extreme lithospehere thickening due to the crustal duplexing.
Rate of plate motions driving Neo-Alpine orogenesis	Rate of the plates convergence in the Miocene is estimated up to 10 cm/yr.	Rate of the plates convergence in the Cenozoic is estimated up to 20 cm/yr.
Rate of recent plate motions	Recent plates convergence rate is almost zero.	Recent plates convergence rate is estimated ca 5 cm/yr.
Type of collision	Oblique continent-continent collision due to eastward extrusion of the Inner Western Carpathian crustal segments to embayment in NEP, typical for the strike-slip orogens.	Frontal continent-continent collision due to the India and Euroasia plate convergence, which resulted in the extreme shortening and crustal thickening accompanied by the strong isostatic movements – uplifts.
Duration and begining of collision	Miocene (22–12 Ma). It started in the Early Miocene, because continent-continent collision was oblique, it was gradualy prograding from the west eastward.	Paleogene – Recent (50 Ma – recent). It started in the Paleocene/Eocene boundary, collision – convergence of India and Asia is still in progress.
Age of collision related thrusting	Miocene, no Quaternary thrusting occurred.	Since the Paleogene to present day, extensive Quaternary thrusting is active.
Present day orogenic activity	Not active - the collisional orogenesis has already ceased.	Still active compressional orogenic belt, with active thrusting.
Orogen–thrusting vergency	North vergent	South vergent
Magnitude of crustal shortening	The narrow width of the gravity low suggests that the continental convergence ceased soon after the ocean basin closure. Estimated is only ca 50 km of overall crustal shortening.	The width of gravity low assumes extreme shortening ca 500–700 km, accommodated by the Indian plate underthrusting under the Asian plate as well as by thrusting and folding in frontal rim of the Indian plate.

Tab. 1
Continuation

CONTRAST FEATURES		
PARAMETERS OF OROGEN	**WESTERN CARPATHIANS**	**HIMALAYAS**
Structural style	For the internal part of the orogen (IWC) is typical Neo-Alpine basin and range structure controlled by faulting and related block rotations and tilting. Neo-Alpine nappe architecture is typical for the external part of orogen (OWC).	Whole mountain belt is composed of superposed Neo-Alpine compression and the rapid updoming as well gravitational nappes.
Continuity of tectonic evolution	Process of orogenesis is not continual. Western Carpathians evolved during several Wilson cycles – orogeneses (Variscan, Paleo-Alpine and Neo-Alpine), interrupted by long-lasting periods of extension and denudation.	Tectonic evolution of the Himalayan belt represented a continual Tertiary–Quaternary process of 50 Ma lasting plates collision.
Seismic activity – origin and earthquake magnitudes	Micro-earthquakes, rarely macro-seismic events reaching average max. intensity M 3–5 are related mainly to post-collisional relaxation strike-slip, less dip-slip faults with moderate slips.	Earthquakes are strong, numerous, generated mainly in great depths and related to the syn-collisional thrust faulting. Strong present-day and historical earthquakes reach average intensity M 7–9.
Units incorporated to Neo-Alpine nappe architecture	The Neo-Alpine accretionary prism of the OWC has incorporated only Tertiary (Paleogene – Neogene) sedimentary sequences. Mesozoic Tethyan units and their Variscan crystalline fundament consolidated by Paleo-Alpine tectogenesis create IWC.	Except the Neogene-Quaternary Siwalik Formation there are in Neo-Alpine orogenic accretionary wedge involved huge nappes of deep crystalline units and Mesozoic Tethyan units.
Synorogenic sedimentary basins	There are genetically various Neogene basins. Depending on their geodynamic position within the orogenic belt there are fore-arc, inter-arc and back-arc basins. There occur marine basins formed by lithospheric extension – thermal subsidence, as well as basins formed by tectonically-fault controlled subsidence.	Except the Siwalik foredeep basin, being the largest in the world and situated in the Himalayan belt, there is a lack of synorogenic sedimentary basins in the terrane of accretionary orogenic wedge due to the extreme uplifting accommodated by the extreme erosion. Subsidence of several transversal intramontane terrestrial Plio-Quaternary sedimentary basins has been controlled by the population of normal faults genetically associated with uplifting Himalayan range.
Character of gravity field (Bouger anomalies)	Gravity low reaches a maximum amplitude of only about –70 mGal.	Gravity low reaches a maximum amplitude of almost –600 mGal.
Crustal thickness	The amplitude of the gravity low indicates small crustal root (on average 35 km with a maximum of 42 km) under orogenic belt.	The amplitude of the gravity low indicates 70–80 km crustal root under orogenic belt.
Origin and tectonic position of synorogenic magmatic complexes	Volcanic complexes represent the Neo-Alpine formations superimposed on the Paleo-Alpine nappe system. Robust Miocene sub-volcanic and volcanic activity was except the subduction processes related as well to the astenosphere upwelling – mantle diapirs.	Northern terrains of Himalayan belt are massively intruded by the Miocene granites exhumed due extreme uplift, accompanied by extensive erosion, but forming the extreme terrain morphology. Massive Tertiary volcanism is situated out of the orogenic belt in the Lhasa block of the Asian plate.
Ophiolite complexes – remnants of oceanic crust	The ophiolites related to the Jurassic subduction are preserved only rudimentary (Meliata unit), they are tracing the suture after the Paleo-Alpine ocean closure.	There is a huge ophiolite belt related to the Neo-Alpine collision following the Indus–Tsang-Po suture closed in the Paleogene period.
Fault network	Faults are numerous, fault network affecting IWC is regular. Important role had the wrench faulting accommodating the extrusion of internal Carpathian rigid blocks to the embayment of the subducting oceanic crust in the North European plate.	Brittle fault network is much more simple, less numerous, dominate faults striking perpendicularly to the Himalayan structure, which have operated as a tear faults of thrusts, as well as normal faults accommodating the extreme upwarping of the mountain belt. Orogen parallel strike-slips at the orogenic root zone accommodate processes of tectonic escape produced by the India plate push.
Dimensions of orogen	Width of orogenic belt is ca 200 km, length of WC orogen loop is ca 630 km, the highest mountain summit has an altitude of 2650 m a.s.l.	Width of orogenic belt is ca 330 km, length of orogenic loop is ca 2600 km, the highest mountain summit has an altitude of 8848 m a.s.l.

geotectonic position, structural style, kinematics and the age of tectonic activity (see cross-sections in **Fig. 4a, b**). From a geometric viewpoint we shall compare individual units of orogens listed from their frontal zones towards their root zones:

- In described Cenozoic Neo-Alpine evolution the North European Plate (a foreland of the Western Carpathians) should geometrically correspond to the Indian plate;
- Western Carpathians foredeep basin corresponds to Sub-Himalaya (Siwalik);
- Outer (External) Western Carpathians (Flysch nappes) correspond to Himalayan accretionary wedge (Lesser Himalaya, Higher Himalaya, Tibetan Tethys Zone?);
- Pieniny Klippen Belt corresponds to Indus–Tsang-Po suture zone. Curious is, that PKB comprises no ophiolites. Ophiolites are known from the Meliata suture situated in the root zone of the Western Carpathians, but this structure is not Neo-Alpine, but one Wilson cycle older – the Meliata ocean was closed in Paleo-Alpine (Mesozoic) Wilson cycle;
- Inner (Internal) Western Carpathians correspond to Trans Himalaya (Asian plate). Neo-Alpine tectonics of IWC is represented by faulting. An important role in Neo-Alpine period had strike-slips, similar as in Tibetan block of Trans Himalaya.

Conclusions

This study inspired by own field experience from two orogens is focussed to comparison of these distant mountain ranges based on classical principles of comparative tectonics defined by Hans Stille (Stille, 1924), applying the up-to date plate-tectonic approach.

The Himalayas and Carpathians belong to the same global Alpine orogenic system, having similar tectonic style of shortening by thrusting and extrusions. There are several fundamental common features, but a lot of peculiarities and differencies in both orogens, too. Most noticeable difference is in the type of Gondwana microplates collision with the Euroasian plate. The Western Carpathians is a strike-slip orogen due to oblique collision, which already ceased after the full oceanic crust

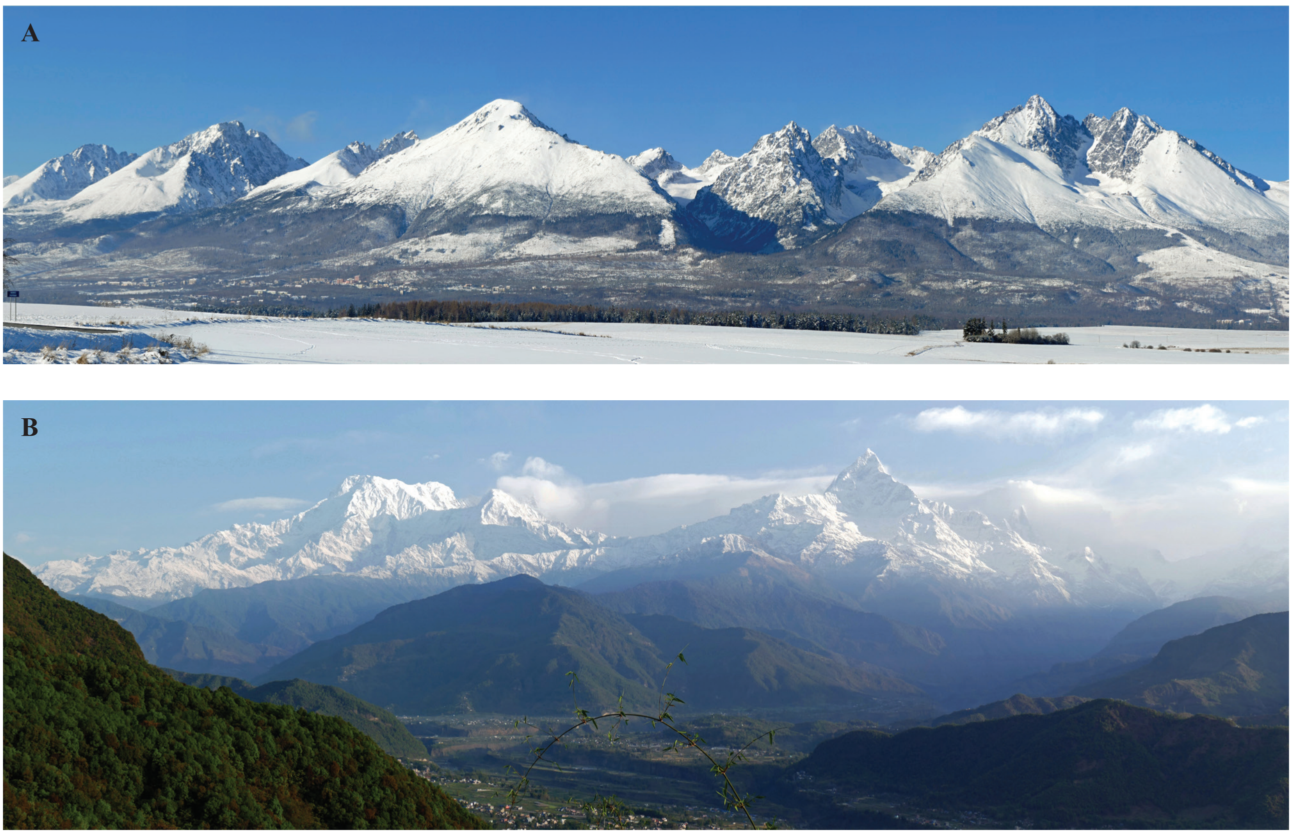

Fig. 5. A similar Alpine-type relief in the Western Carpathian and Himalayan mountain ranges. Difference is in dimensions and altitudes, the highest Gerlach peak in the High Tatras (Vysoké Tatry Mts.) in Western Carpathians reaches 2650 m a.s.l., while the highest peak of Himalayas Mount Everest has altitude 8848 m a.s.l. A panoramatic view from the south northward of: **A –** the High Tatras – Vysoké Tatry Mts., Inner Western Carpathians, Tatric unit. **B –** Annapurnas group (left) and Machhapuchchhre (right) of the main ridge of Himalayas (Great Himalaya) seen from the view point near the Pokhara village (both photographs by J. Madarás).

subduction. Himalayas are a result of frontal collision, which after the oceanic crust consumption has continued further by underthrusting of Indian plate continental crust under the Asian one. It led to formation of the most extreme shortening and crustal thickening, accompanied with a largest uplift in the world and creation of the highests mountains. The Western Carpathians branch of the Alpine mobile belt is currently inactive. Nevertheless, the convergence of India and Asia plates continues at present day resulting in high seismic activity presumably related to frontal Himalayan thrusts.

The Himalayas are purely Cenozoic Neo-Alpine structure, while in the Western Carpathians there is preserved Variscan (Hercynian; Paleozoic) and Paleo--Alpine (Mesozoic) nappe architecture in the internal part of the orogen as well as the Neo-Alpine (Cenozoic) fold and thrust belt, creating the external part of the orogen. Complex brittle fault network is Neo-Alpine, affecting mostly the IWC block of the Western Carpathians. Dominant role had wrench faulting, accommodating the extrusion of internal Carpathian block to the embayment of subducting oceanic crust in the North European plate. Himalayan brittle fault network is much more simple, there dominate faults striking perpendicularly to Himalayan structure, which have operated as a tear faults of thrusts, as well as normal faults accommodating the extreme upwarping of the mountain belt.

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Fig. 6. Participants of the field reconnaissance geological exploration trip along the Kali Gandaki river in front of the Departnent of Geology, Tri-Chandra Multiple Campus, Tribhuvan University in Kathmandu, listed from the left towards the right: Miroslav Bielik, Pavol Siman, Ján Madarás, Juraj Papčo, Ashok Sigdel, František Marko, Subash Acharya, Kamil Fekete and Andrej Mojzeš.

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Porovnanie kenozoickej neoalpínskej tektonickej evolúcie Západných Karpát a Himalájí (Slovensko – Nepál)

Na jar v roku 2019 kolektív autorov tohto príspevku absolvoval v rámci vedeckého projektu APVV-16-0146 krátku, ale programovo bohatú výskumnú terénnu cestu do nepálskej časti Centrálnych Himalájí. Jej cieľom bolo oboznámenie sa so stavbou tohto grandiózneho orogénneho pásma. Získaná terénna skúsenosť z najvyššieho pohoria sveta a rokmi nadobudnuté poznatky zo západokarpatských alpíd nás inšpirovali k zostaveniu porovnávacej štúdie tektonických štýlov oboch pohorí. Základy modernej komparatívnej tektoniky, z ktorej princípov sme vychádzali, položil už Stille (1924). Pri porovnávaní štruktúrno-tektonických čŕt oboch pohorí a tvorbe vlastného pohľadu na problematiku sme aplikovali princípy platňovej tektoniky a mohli sme sa oprieť aj o mnohé klasické práce zamerané na stavbu Himalájí a Západných Karpát.

Západné Karpaty aj Himaláje (obr. 2 a 3) sú súčasťou globálneho orogénneho systému tetýdnych alpíd (obr. 1). Generálne majú obe pohoria podobný tektonický štýl kolíznych orogénych procesov, ale nachádzame aj niektoré odlišnosti a špeciality. Tie sú výsledkom špecifických podmienok pri kolízii fragmentov Gondwany s eurázijskou platňou v európskej a ázijskej časti orogénej zóny, ktoré sa dnes nachádzajú v rôznych štádiách kolíznej fázy v rámci Wilsonovho orogénneho cyklu.

Najdôležitejším faktorom ovplyvňujúcim charakter a tvar orogénnej zóny je tvar konvergujúcich platní. Výsledkom šikmej kolízie je strižný (z anglického termínu *strike-slip*) typ západokarpatského orogénu (sensu Dadlez a Jaroszewski, 1994). Sformoval sa v neoalpínskej epoche extrudovaním rigidných mikroplatní do zálivu v severeurópskej platni tvoreného tenkou oceánskou kôrou, ktorá subdukovala pod prenikajúce mikroplatne karpatských jednotiek. Naproti tomu, himalájska orogénna kolízia je typickým príkladom čelnej kolízie. V Západných Karpatoch kolízia po konzumácii oceánskej kôry subdukciou vyvrcholila už v miocéne. V Himalájach tento proces pokračuje dodnes so všetkými sprievodnými znakmi, akými sú intenzívny výzdvih, erózia a extrémna seizmicita, generovaná najmä na rozhraniach nasúvajúcich sa príkrovov fundamentu. V Himalájach sa uplatňuje raritný typ platňovej konvergencie – kolízia typu A (Bally, 1981), pri ktorej sa po konzumácii tetýdnej oceánskej kôry indická kontinentálna platňa ďalej podsúva pod ázijskú kontinentálnu platňu, čím dochádza k extrémnemu zhrubnutiu litosféry. Na rozdiel od Západných Karpát, v Himalájach sú magnitúda a rýchlosť presunov príkrovov akrečnej prizmy orogénu aj dimenzie pohoria rádovo vyššie. Západokarpatský orogén sa sformoval superponovaním variských, paleo-, mezo- a neoalpínskych tektonických procesov oddelených etapami pokoja a denudácie. Variské, paleo- a mezoalpínske štruktúry sú zachované vo vnútrokarpatskom bloku južne od neoalpínskej orogénnej prizmy. Himalájsky orogén je výlučne neoalpínsky. Je výsledkom kontinuálnej, asi 50 mil. r. trvajúcej kolízie indickej platne s eurázijskou, počas ktorej bola a stále je generovaná orogénna akrečná prizma formujúca sa z jednotiek indickej platne. Špecifikom Západných Karpát je morfotektonický štýl striedania neogénnych sedimentárnych bazénov a hrastí kontrolovaný zlomami, vyvinutý vo vnútrokarpatskom bloku orogénu, porušenom hustou sieťou zlomov viacerých genetických systémov. Rozsiahle zaoblúkové neogénne sedimentárne bazény, generované dominantne termálnou subsidenciou sprevádzanou mohutným subsekventným vulkanizmom, sú vyvinuté v tyle západokarpatského orogénu. V Himalájach mladé neogénne molasové sedimenty, geneticky korešpondujúce so sedimentmi karpatskej čelnej predhlbne, sú vo veľkom rozsahu situované vo frontálnej časti orogénu (Siwalik). Vnútri himalájskej orogénnej prizmy sú len sporadické úzke priečne grabeny kontrolované poklesovými zlomami (napr. graben Mustang), vyplnené miocénnymi a kvartérnymi fluviálno-glaciálnymi sedimentmi. Významné smernoposunové zlomy subparalelné s himalájskym orogénnym frontom sú situované v tyle pohoria – v Transhimalájach, ktoré sú už súčasťou ázijskej platne. Tieto hlboké kôrové rozhrania prvého rádu sprostredkúvajú tektonický únik čiastkových blokov vyvolaný tlakom indickej platne. Je to proces, ktorým sa relaxuje časť energie konvergujúcich platní, indickej a ázijskej. Oba orogény sú výrazne asymetrické, no Západné Karpaty sú generálne severovergentné a Himaláje juhovergentné (obr. 4a, b). Oba orogény vykazujú polaritu kolízneho frontu postupujúceho zo západu na východ a z tyla orogénu smerom do čela, teda ide o nesené (*piggy back*) násunové sekvencie.

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