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The interplay between tectonics, sediment dynamics and gateways evolution in the Danube system from the Pannonian Basin to the western Black Sea

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Abstract :

Understanding the natural evolution of a river–delta–sea system is important to develop a strong scientific basis for efficient integrated management plans. The distribution of sediment fluxes is linked with the natural connection between sediment source areas situated in uplifting mountain chains and deposition in plains, deltas and, ultimately, in the capturing oceans and seas. The Danube River–western Black Sea is one of the most active European systems in terms of sediment re-distribution that poses significant societal challenges. We aim to derive the tectonic and sedimentological background of human-induced changes in this system and discuss their interplay. This is obtained by analysing the tectonic and associated vertical movements, the evolution of relevant basins and the key events affecting sediment routing and deposition. The analysis of the main source and sink areas is focused in particular on the Miocene evolution of the Carpatho-Balkanides, Dinarides and their sedimentary basins including the western Black Sea. The vertical movements of mountains chains created the main moments of basin connectivity observed in the Danube system. Their timing and effects are observed in sediments deposited in the vicinity of gateways, such as the transition between the Pannonian/Transylvanian and Dacian basins and between the Dacian Basin and western Black Sea. The results demonstrate the importance of understanding threshold conditions driving rapid basins connectivity changes superposed over the longer time scale of tectonic-induced vertical movements associated with background erosion and sedimentation. The spatial and temporal scale of such processes is contrastingly different and challenging. The long-term patterns interact with recent or anthropogenic induced modifications in the natural system and may result in rapid changes at threshold conditions that can be quantified and predicted. Their understanding is critical because of frequent occurrence during orogenic evolution, as commonly observed in the Mediterranean area and discussed elsewhere.

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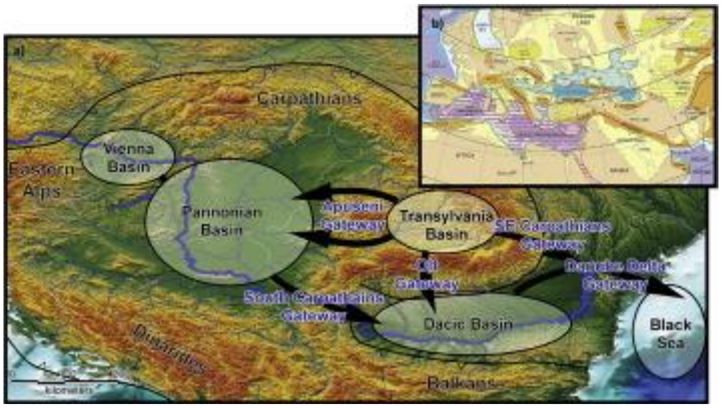
Graphical abstract



Highlights

► Tectonic and sedimentological background of human-induced changes in basin systems ► Focus on the recent past evolution of the Danube River–western Black Sea system ► Source to sink analysis of basins connectivity and gateway evolution ► Tectonic vertical movements, basin evolution and key events for sediment routing ► Model of understanding basins fragmentation during continental collision

Keywords : Source to sink, Gateways, Connectivity, Danube Basin, Black Sea

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1. Introduction

The distribution of sediment fluxes is linked with the natural connection between sediment source areas situated in actively uplifting mountain chains and deposition in plains, deltaic systems and, ultimately, in the capturing oceans and seas. This defines an integrated Source to Sink system. The Danube River – western Black Sea is one of the most active European systems in terms of sediment re-distribution that poses significant societal challenges and vulnerabilities (Figs. 1 and 2, Cloetingh et al., 2005; Matenco and Andriessen, 2013). The present situation is related to human changes superposed over a recent geological evolution that includes not only long-term processes such as general tectonic-induced vertical motions or fluvial sediment routing, but was significantly conditioned by events taking place in a specific geological situation.

Understanding Source to Sink systems that evolved during the last stages of mountain building is strongly dependent on understanding the mechanics of orogenic collision that fragments the associated sedimentary basins. These basins become gradually shallow and are filled by a combination between tectonic shortening/uplift and high sedimentation rates driven by rapid orogenic exhumation. Such collisional fragmentations associated with major deltaic processes filling up multiple basins separated by subaerial or submarine barriers are observed worldwide and make up the active geological background to human-induced changes in regions such as the Eurasian Paratethys system, the Mediterranean or the SE Asia arc- back-arc areas (Jolivet et al., 2006; Popov et al., 2006; Krijgsman et al., 2010; Pubellier and Morley, 2014).

One of the best examples of this type of interplay between collisional mechanics and recent evolution of a Source to Sink system with large impact into the active evolution of landforms is the Danube Basin - western Black Sea system, in particular the part including the Pannonian Basin and E-wards (Fig. 2). Major deltaic systems were active along roughly the trace of the present-day Danube River and its major tributaries that filled in the last 10Ma the 1300km wide Pannonian, Transylvanian and Dacian basins system (Krézsek et al., 2010; Jipa and Olariu, 2013; Magyar et al., 2013). At the same time, these relatively shallow water basins together with the surrounding mountain chains have suffered significant changes in terms of vertical movements, the driving tectonic and associated sedimentological processes being in most cases active at present and still conditioning the sediment transport and deposition. The coeval evolution of the western system of partly endemic Paratethys lakes and seas separated by the emerging Carpatho-Balkans, Dinarides or Anatolia mountains, hills or submerged barriers into the present day Danube River – Black Sea system (e.g., Rögl, 1999) is marked by key events actively re-shaping sediment routing. Such events included often situations of sediments trapping and starvation across barriers between basins, moments of basins connectivity (uni- or bi-directional) over (dis-)continuous gateways, rapid tectonic uplift and subsidence, large-scale active faulting, moments of rapid regressive basin fill and events of conditioning sediments discharge into large-scale turbiditic systems (e.g., Matenco et al., 2007; Matenco et al., 2010; Leever et al., 2011; Bartol et al., 2012; Munteanu et al., 2012; ter Borgh et al., 2015). The mechanisms driving this inheritance are of significant importance for understanding the impact into the present-day vulnerability (e.g., Stanica and Panin, 2009). A significant number of studies were published in recent years on the long term Miocene – Quaternary evolution (10^5-10^6Ma time-scale) of the various parts of the basins and mountain ranges located in the lower Danube River and adjacent part of the Black Sea (e.g., Cloetingh et al., 2006 and references therein). This provides an optimal timing for understanding the tectonic and sedimentological background of processes operating at higher resolution (10^3-10^5) and understanding the critical thresholds of the natural system impacted by human changes.

In this review we aim to analyse the tectonic and sedimentological background of the human-induced changes in the Danube Basin – Black Sea system by linking a number of key studies to understand both the long-term evolution and the geological events able to underline the mechanics of the sedimentary system and its regional specificity. This is obtained by analysing the tectonic and associated vertical movements background, the evolution of relevant basins and the key events affecting sediment routing and deposition. The unusual high rates of tectonic vertical movements driving the Miocene – Quaternary evolution of the Carpathians will be reviewed by looking both to the kinematics of the uplifting mountain chain and to the subsidence and associated sedimentary geometry of their foreland and hinterland basins. The moments of connectivity and gateways evolution will be reviewed by analysing the geometries of the connections between the Dacian Basin and the Black Sea as well as in between the Pannonian and Dacian basins. In particular, the onset and subsequent evolution of the massive sedimentation along the western Black Sea deep sea fans will be reviewed in particular in the context of major sea-level changes such as the Messinian Salinity Crisis of the Paratethys or the observation of often Quaternary sea-level fluctuations. The mechanics of the system will be described by reviewing and integrating a number of key numerical modelling studies applied for the evolution of the vertical movements in the Carpathians foreland as well as for the main moments of basins connectivity and gateways evolution.

2. Inherited tectonic configuration and vertical movements of the Source to Sink system

Continental collision is the moment when out-of-sequence contractional deformation becomes rather the rule than the exception, as commonly observed during the evolution of the Mediterranean orogens, driven by the rapid retreat of the Calabrian, Vrancea, Aegean or Gibraltar slabs that peaked in almost all situations during Miocene times (e.g., Faccenna et al., 2004; Jolivet and Brun, 2010; Ismail-Zadeh et al., 2012). This slab retreat was accommodated by coeval extension affecting the hinterland of the upper orogenic plate, which formed large extensional basins floored by either continental or oceanic lithosphere (such as the Pannonian and Aegean basins, Black Sea or Western Mediterranean). These basins are extensional back-arcs in terms of geodynamic evolution (e.g., Royden, 1993; Okay et al., 1994; Horváth et al., 2006; Doglioni et al., 2007) although their relative position behind a magmatic or island-arc (Uyeda and Kanamori, 1979; Mathisen and Vondra, 1983) is not always very clear. In some situations, the rapid evolution by translations and large-scale rotations is a local process that cannot be accommodated by the dominantly N-S oriented absolute plate motion of Africa relative to Europe (Kreemer et al., 2003; van der Meer et al., 2010; van Hinsbergen and Schmid, 2012), such as the rapid E-wards movement of the Carpathians and Apennines or the W-ward movement of the Betics-Rif system (Faccenna et al., 2005; Matenco et al., 2010; Vergés and Fernàndez, 2012). This type of local processes is always associated with large-scale mantle dynamics and evolution of slabs along laterally variable processes such as slab detachment or STEP faulting with significant impacts in terms of dynamic topography, rapid changes in fluvial systems and geohazards (Wortel and Spakman, 2000; Govers and Wortel, 2005; Syvitski et al., 2009).

One of the best illustrations of these processes is the Danube River and its major tributaries that crosses a number of mountain chains along gateways (such as the South Carpathians, Apuseni, Olt or Scythian gateways) on its way from the Pannonian and Transylvanian back-arc basins, passing thorough the Carpathians foreland Dacian Basin towards the main depositional sink of the system, the Black Sea back-arc basin (Fig. 2). The specific Miocene to recent geological evolution of this area was significantly influenced by the aftermath of the back-arc extension and collision that took place in the Carpatho-Balkans and Dinarides orogens (Cloetingh et al., 2004; Horváth et al., 2015). Of particular importance are a number of features that provided a continuous sediment routing background. Among them, the rapid differential motions taking place in the foreland of the Carpathians has provided a long-term trapping of sediments, which was ultimately surpassed only during their late stage evolution (Munteanu et al., 2012). The separation of internal Carpathians-Dinarides back-arc basins has resulted in the formation of a number of lakes or shallow seas in the larger Pannonian area that were ultimately filled by unusual high depositional rates during a regressive deltaic fill that included the paleo-Danube and a number of other hinterland rivers (Magyar et al., 1999b; Steininger and Wessely, 1999). The latter process is a very good analogue to understand the mechanisms driving sedimentation in more downstream Danube – Black Sea system in recent times.

The highly curved double loop of the Carpatho-Balkans orogen together with the Dinarides Mountains bordering to the south the Pannonian Basin (Fig. 2) formed in response to a Triassic to Tertiary evolution of continental tectonic units and intervening oceans. In a simplified terminology, two intervening oceans (Neotethys and Alpine Tethys) separated by gradual opening and enlargement five continental blocks (Dinarides, ALCAPA, Tisza, Dacia and the European foreland) and subsequently closed during Cretaceous – Miocene times by the subduction and collision that took place between these continental units (Fig. 1, Csontos and Vörös, 2004; Schmid et al., 2008). The Neotethys was the large ocean that separated Europe from Africa throughout most of their Mesozoic evolution, which opened in the studied area during Middle Triassic times between the Dinarides (Africa/Adriatic units) and Tisza-Dacia (Europe derived continental units) (Stampfli and Borel, 2002; Schmid et al., 2008). Tisza is characterized by mixed continental affinities because of its separation from Europe during Middle Jurassic time, movement southwards in a position adjacent to the Adriatic unit and re-alignment to European-derived blocks (i.e. Dacia) during the late Jurassic to late Early Cretaceous moments of closure of an eastern branch of the Neotethys Ocean. The entire Neotethys was ultimately closed by subduction and collision between the Dinarides and Tisza-Dacia during Cretaceous – Paleogene times (or East Vardar, Sandulescu, 1975; Vörös, 1977; Csontos and Vörös, 2004; Haas and Péró, 2004; Schmid et al., 2008). This collision was responsible the creation of a number of continental thrust sheets presently observed in the Dinarides orogen (Fig. 1, e.g., Dimitrijevic, 2001). The Middle Jurassic opening of the Alpine Tethys Ocean and its eastern Ceahlau-Severin continuation was kinematically connected with the rifting and enlargement of the Central Atlantic and in the study area it separated the Alps and Carpathians units from the European foreland (Favre and Stampfli, 1992; Csontos and Vörös, 2004). This ocean was almost completely closed by the convergence between Dacia and the European foreland during successive moments of late Early Cretaceous and latest Cretaceous subduction (e.g., Săndulescu, 1984; Stefanescu, 1995; Iancu et al., 2005). The ALCAPA (i.e. ALps-CArpathians-PAnnonia) unit is made up of far-travelled Adriatic-derived continental nappes, trusted northwards during the Cretaceous-Paleogene closure of the Alpine Tethys and the emplacement of the thick-skinned nappes of the Alps and Western Carpathians (Golonka, 2004; Schmid et al., 2004). At the end of this Cretaceous – Paleogene evolution the entire Alps, Dinarides and Carpatho-Balkans was an orogenic area characterized most likely by much higher topographic elevations than the ones observed at present. This orogenic area was separated from the European foreland by a thinned continental to possibly oceanic domain, i.e. the Carpathians embayment, formerly connected with the Ceahlau-Severin ocean (Fig. 2, e.g. Ustaszewski et al., 2008).

2.1. Exhumation of the Carpatho-Balkans and evolution of the neighbouring foredeep

The Cretaceous-Paleogene evolution was subsequently followed by the Miocene formation of the large internal Pannonian-Transylvanian basins system. This was coeval with the Miocene shortening that took place at the exterior of the orogenic area during the closure of the Carpathians embayment and the formation of the external foreland basin (Fig. 1). These coeval and kinematically contrasting processes took place during the rapid Miocene retreat starting at ~20Ma of a slab attached to the European continent (Cloetingh et al., 2006; Horváth et al., 2006; Schmid et al., 2008). This retreat resulted in a back-arc extension that took place in both intra-Carpathians units (Tisza-Dacia and ALCAPA), accompanying their opposite sense rotation that started earlier during Paleogene times (Balla, 1986; Csontos, 1995). ALCAPA was affected by counter-clockwise rotations and ENE-wards translations during its thrusting over the European foreland, Miocene extension of the Pannonian Basin and the coeval extrusion affecting the Eastern Alps (Ratschbacher et al., 1991; Tari et al., 1992; Csontos, 1995; Fodor et al., 1999; Krzywiec, 2001). Tisza-Dacia continued its clockwise rotation and accommodated variable amounts of Miocene extension decreasing SE-wards towards the connection between the Dinarides and Carpathians Mountains (Matenco and Radivojević, 2012). The Miocene subduction ultimately culminated during the final moments of collision and the emplacement of the frontal Subcarpathian nappe at around 11-8Ma. This event was followed in the SE Carpathians by subsidence and differential vertical motions during latest Miocene – Quaternary times, a process that is still active today (e.g., Leever et al., 2006; Matenco et al., 2007; Matenco et al., 2010; Ismail-Zadeh et al., 2012 and references therein). This was associated with the formation of a foredeep basin in the frontal part of the orogen with variable thicknesses along its strike (Matenco et al., 2003).

A large number of low-temperature thermochronological studies (fission tracks and U-Th/He, e.g., Gleadow and Duddy, 1981; Wolf et al., 1996) have become recently available for Carpathians Mountains (Merten, 2011 and references therein; Danisik et al., 2012) and have demonstrated that the Miocene – Quaternary exhumation of the Carpathian Mountains is generally limited to 5-6km (Fig. 3), with few exceptions in the Western Carpathians or at transcurrent contacts in the East Carpathians, where exhumation values are higher. The exhumation started during Oligocene – Early Miocene times and peaked during the last collisional stages ~15-8Ma due to deformation of the external Carpathians nappes and the subsequent contractional uplift of the orogenic core (Fig. 3a,d, Matenco et al., 2010). The same overall pattern is observed in the South Carpathians, although with significantly reduced values (Fig. 3c,d). In the SE Carpathians, the exhumation pattern is different, the fairly low values of Miocene exhumation in the internal part of the orogen contrasts with the much higher, up to 6km cumulative exhumation observed in the external nappes (Fig. 3b,d). In more details, higher resolution studies have demonstrated that the exhumation migrates towards the foreland, the ages decreasing from Cretaceous in the internal basement nappes to Miocene – Quaternary in the external part of the thin skinned wedge (Merten et al., 2010; Necea, 2010). In the latter, the gradually migrating exhumation increased from ~0.8mm/yr Miocene values, comparable with the ones in the East Carpathians, to much higher 1.6-1.7mm/yr values during the latest Miocene – Early Pliocene and Pleistocene exhumation events (Fig. 4). This has been explained by a transition from thin-skinned to thick-skinned deformation during the final phases of collision and the subsequent evolution driven by the evolution of the Carpathian slab (Fig. 4, Leever et al., 2006; Matenco et al., 2007; Merten et al., 2010). The combination between the post-Miocene exhumation of the SE Carpathians and the rapid and continuous subsidence of the Focsani basin, presumably driven by the pull of the underlying Vrancea slab, has created this unusual foreland geometry (Fig. 4) that is still active today in shifting the high topography towards the foreland and modifying the associated fluvial network (Leever, 2007; ter Borgh, 2013). The continuous subsidence created the thick Focsani basin in the SE Carpathians that reaches anomalous high values of up to 13Km (Figs. 3 and 4, Tărăpoancă et al., 2003; Tărăpoancă et al., 2004). This basin functioned a trap for the Carpathians sediment fluxes for most of its evolution (Munteanu et al., 2012).

The post- late Miocene collisional subsidence related to the evolution of the Vrancea slab affected the much larger area of the Moesian platform situated in the foreland of the SE and South Carpathians, which recorded subsidence and sedimentation gradually covering the external trusted units of the orogen (Rabagia et al., 2011; Krézsek et al., 2013). This Uppermost Miocene – Quaternary cover of the Moesian platform and adjacent external nappes is part of the Dacian Basin of the Eastern Paratethys (Jipa and Olariu, 2013). The thickness this basin gradually increase E-wards towards its Focsani part, where these sediments were partly tilted and exhumed by the Pliocene-Quaternary contraction recorded by the SE Carpathians (Leever et al., 2006).

The exhumation of the South Carpathians is significantly different (Fig. 3c). Available exhumation data indicate that the bulk of their last event of tectonic exhumation took place during (Late) Cretaceous – Paleogene times (Bojar et al., 1998; Sanders, 1998; Schmid et al., 1998; Sanders et al., 2002; Fügenschuh and Schmid, 2005; Merten, 2011), significantly predating the continuous Miocene –Quaternary exhumation recorded by the East and SE Carpathians. By the end of these Cretaceous contractional deformations, the clockwise rotations have brought the South Carpathians strike in a position that was likely sub-parallel with the Moesian margin (Fügenschuh and Schmid, 2005), which is also parallel with the direction of subsequent translations and rotations into the Carpathian embayment. This orogenic geometry resulted in the subsequent elongation of the South Carpathians (i.e. orogen - parallel extension) and the formation of the large-scale Cerna-Jiu and Timok fault transtensional systems cumulating ~100Km total dextral offset in respect to the Moesian margin (Fig. 1, Schmid et al., 1998; Răbăgia and Matenco, 1999; Tărăpoancă et al., 2007; Schmid et al., 2008). This extension-transtension has migrated E-wards during the Oligocene – Early Miocene along the foreland of the South Carpathians (Krézsek et al., 2013) and resulted in extension associated with subsidence that affected the core of the orogen (Matenco and Schmid, 1999). For instance Oligocene – Middle Miocene continental to shallow marine sediments are found at high elevation in the South Carpathians, deposited most likely due to coeval subsidence. These events were followed by the Middle – Late Miocene transpressional docking of the South Carpathians against the Moesian platform. This is in agreement with the much lower amounts of thrusting when compared with East and SE Carpathians, and with termochronology data, only few AFT Miocene ages as young as ~8Ma being obtained in the centre of exhumation (Fig. 3c,d). A larger number of AFT exhumation ages distributed throughout the Miocene have been obtained in the western part of the Danubian nappes (Bojar et al., 1998), near the connecting area with the Balkanides (Figs. 1 and 2).

The Balkanides units share a common Cretaceous tectonic history with the South Carpathians, a first late Early Cretaceous period of nappe stacking of the Srednogorie and Strandja units over their foreland being followed by reduced latest Cretaceous shortening (Fig. 1, Ivanov, 1988). The internal parts of the Srednogorie unit in the region of Panagyurishte and its lateral prolongation in Timok zone of the South Carpathians of Serbia was affected by late Cretaceous (~93-72 Ma) magmatism associated with the back-arc extension of the Neotethys subduction zone of the Dinarides-Hellenides (Ciobanu et al., 2002; von Quadt et al., 2005; Zimmerman et al., 2008). These deformations were followed by the emplacement of the Srednogorie and Central Balkan unit in the external part of the orogen that peaked in Eocene times (~42-40Ma) (Ivanov, 1988; Bergerat et al., 2010). Almost at the same time, the internal part of the Balkanides was affected by large scale extension, most likely driven by the extensional exhumation of the Rhodope core-complex, that was subsequently followed by the formation of a number of Oligocene – Quaternary basins superposed over the core of the orogen (Burchfiel et al., 2008; Bergerat et al., 2010; Kounov et al., 2010). Although there are almost no low-temperature thermochronological data over the external Balkanides, their geometry and sedimentary architecture suggest that these mountains were relatively stable and formed the southern margin of the Carpathains foreland basins during their entire Miocene evolution.

2.3. Extension of the Pannonian basin and its relationship with the Dinarides and Carpathians shortening.

Following a Late Jurassic – earliest Cretaceous period of obduction or otherwise emplacement of ophiolites over both the Dacia and Adriatic margin, the Neotethys (or Vardar) Ocean was subsequently closed in the Dinarides by subduction and collision during Cretaceous – Eocene times. This was associated with the formation of a thick-skinned nappe stack in the former Adriatic margin, deformation generally migrating from the internal to the external part of the orogen (e.g., Dimitrijević, 1997; Schmid et al., 2008; Ustaszewski et al., 2010). Field kinematics combined with thermochronology have recently demonstrated that the internal part of the Dinarides close or adjacent with the Pannonian Basin was affected by large scale extension starting with the Oligocene - Early Miocene, which peaked during Middle Miocene times. This extension took place along a series of detachments that separate large-scale crustal exhumation in their footwall from the formation of half-grabens in their hanging-wall. A large number of such detachments were mapped along the strike of the Dinarides separating Miocene (half-) grabens that are presently either juxtaposed over the Dinarides Mountains or are buried at depth beneath the subsequent cover of the Pannonian Basin (Tomljenović and Csontos, 2001; Fodor et al., 2008; Ustaszewski et al., 2010; Matenco and Radivojević, 2012; Stojadinovic et al., 2013; Toljić et al., 2013).

Northwards, the extension of the Pannonian Basin is observed along a large number of Miocene (half-)grabens. The interpretation of seismic lines in the SE part of the basin (Figs. 4 and 5) suggests that the extension was asymmetric and deformation migrated NE and E-wards in space and time across the basin, from Early Miocene to early Pontian (20Ma or older to ~8.5 Ma), ending at the same moment with the ceasure of collisional shortening in the East and SE Carpathians. The last, late Miocene (Pannonian-Early Pontian) stage of extension was associated with large scale mantle lithospheric thinning and the large astenospheric upraise that is presently still observed beneath the Pannonian Basin (Horváth et al., 2006; Matenco and Radivojević, 2012; Horváth et al., 2015). This extensional geometry has also been observed in the NW part of the Pannonian Basin, while the intermediate area is less clear due to the large-scale transcurrent movements affecting the contact between Tisza-Dacia and ALCAPA along the Mid Hungarian Shear zone (Tari et al., 1992; Fodor et al., 2005; Horváth et al., 2006). The Pannonian Basin was subsequently affected by a latest Miocene – Quaternary period of inversion that is still presently active, presumably driven by the indentation/subduction of the Adriatic promontory. This has created km-scale thrusting near the NW Dinaridic margin and similar size transcurrent movements observed more in the basin centre and to the NE (Tomljenović and Csontos, 2001; Fodor et al., 2005; Pinter et al., 2005; Bada et al., 2007). In the E-SE part of the basin, near the Dinarides, South Carpathians and Apuseni Mountains the effects of this inversion are fairly reduced, being limited to the formation of either positive flower structures with uplift in the order of few hundreds metres near the Dinarides (e.g., Toljić et al., 2013), or small scale transtensional structures with similar offsets near the Apuseni Mountains (e.g., Windhoffer and Bada, 2005).

Kinematic and exhumation data along a regional profile (Fig. 4) shows that the Miocene contraction and exhumation of the SE Carpathians was indeed coeval with the extension recorded in the SE part of the Pannonian basin. In this Tisza-Dacia sector of the chain, balanced cross-sections and various reconstructions have inferred a total amount of shortening in the external Carpathians in the order of 120-160 Km, increasing from north to south (Roure et al., 1993; Ellouz et al., 1994; Morley, 1996). The amount of the back-arc extension in the Tisza-Dacia sector of the Pannonian Basin is variable due to the coeval clockwise rotational kinematics, decreasing from the contact with ALCAPA at the Mid-Hungarian Shear zone towards the SE junction between the South Carpathians and the Dinarides of Serbia (Fig. 1). Available estimates indicate that near the Mid-Hungarian shear zone the total amount of extension is in the order of 140-180 Km (Lenkey, 1999; Ustaszewski et al., 2008), which is almost the same with the Romanian Carpathians shortening. These estimates mean that there was no large-scale absolute plate motion involved in the Carpathians Miocene shortening, contraction at the exterior of the orogen being entirely accommodated by back-arc extension. In other words, the continental unit composed by the Apuseni Mountains, Transylvanian Basin, East, SE and South Carpathians simply rotated clockwise and moved E-wards into the Carpathian embayment, driven solely by pull and sink of the slab roll-back, collapsing the Pannonian basin in the back and shortening the Carpathians in front. There was no other driving force pushing or otherwise moving this continental unit E-wards, in agreement with the overall N-ward movement of Africa relative to Europe during Miocene times. This process becomes obvious when comparing the Miocene structures on both sides of this continental unit (Fig. 4).

2.4. Dynamic topography in the Transylvanian Basin

One of the most striking European examples of dynamic topography (i.e. deep mantle driven topography) is the evolution of the Miocene Transylvanian Basin (Figs. 2 and 4). The up to 3.5km thick Middle – Upper Miocene sedimentary cover has an apparent symmetric geometry both in cross-sections and in map view (Figs. 1 and 4). The Miocene cover overlies an earlier orogenic evolution that includes late Early Cretaceous shortening and ophiolite emplacement, late Cretaceous extension, latest Cretaceous –Eocene shortening, Oligocene – Early Miocene denudation and the deposition of a foredeep wedge in the northern part of the basin during to the final ALCAPA thrusting over the Tisza-Dacia (Krézsek and Bally, 2006; Tischler et al., 2008; Tiliţă et al., 2013 and references therein). The onset of regional subsidence in the Transylvanian Basin took place during Middle Miocene times, continuing with accelerated pulses throughout the Middle – Late Miocene (Filipescu and Gîrbacea, 1997; Krézsek and Filipescu, 2005). The subsidence started near the East and SE Carpathians and gradually extended over the entire basin (Tiliţă et al., 2013). In parallel, the exhumation of the neighbouring Carpathians during Middle-Late Miocene uplifted the margins of the Transylvanian Basin in several pulses of movement and created forced regressive sequences with coarse deltaic deposition (Krézsek et al., 2010; Matenco et al., 2010). These vertical tectonic movements were interrupted by an eustatic sea level drop during Middle Badenian times (de Leeuw et al., 2010), which lead to the deposition of thick salt and other evaporitic sequences throughout the basin. These evaporites migrated during latest Middle – Late Miocene and created locally large exaggerated diapirs due to the overburden, contractional stresses and volcanic sagging (Krézsek and Bally, 2006; Szakacs and Krezsek, 2006; Tiliţă et al., 2013). Towards the end of the late Miocene times (~8Ma) the entire Transylvanian Basin was uplifted to the ~600m maximum topographic elevations of the sedimentary fill, being subsequently affected by significant erosion and local deposition of Pliocene - Quaternary continental sediments (Matenco et al., 2010).

None of these substantial Miocene vertical movements can be explained by the minor normal faults or thrusts observed in the Transylvanian Basin (Fig. 3). Both the initial subsidence and the subsequent regional exhumation are dynamic topography processes, related to the deep mantle evolution driven by the roll-back subduction of the Carpathians. It is rather unclear which dynamic topography process may be responsible from the wide variety inferred, although some are more likely, such as mantle thinning during extension and its subsequent response (Tiliţă et al., 2013). Furthermore, the evolution of the adjacent intramontane Pliocene – Quaternary Brasov and Tg. – Secuiesc basins (Figs. 2 and 4) and the associated alkaline and adakitic volcanism is also driven by dynamic topography processes. The formation of these extensional grabens with sediments averaging few hundreds to few tens of metres is related to the rise of the astenospheric mantle in the hinterland of the rapidly sinking Vrancea slab, as inferred by a wide array of deep geophysical observations and numerical modelling (Seghedi et al., 2011; Ismail-Zadeh et al., 2012). Therefore, the present day topography of the Transylvania basin and its smaller neighbouring intramontane basins is a response to active upwelling of the asthenosphere in the hinterland of the sinking Vrancea slab (Fig. 4). The effects in the topography are obvious by a general tilting and migration of the fluvial network that crosses the Apuseni Mountains and their northern prolongation (Fig. 2). This process appears to be still active at present, as inferred by the numerical modelling of the fluviatile network (Dombrádi et al., 2007; ter Borgh, 2013).

In fact, the postulation of Transylvania as a back-arc basin behind the volcanic arc of the East Carpathians is highly questionable. Transylvania is different when compared with the typical extensional back-arc geometries of the Pannonian Basin. Given the coeval low values of exhumation of the Apuseni Mountains (Fig. 4) and the pre-Miocene continuity of the orogen in the entire intra-Carpathians region, the Transylvanian Basin can be alternatively interpreted as an unusual large fore-arc basin overlying the frontal part of such a wide orogen. In other words, the current dilemma is: Transylvanian Basin is a small back-arc or a large fore-arc basin? A response would be eventually important for deciphering the underlying mechanisms driving its evolution.

2.5. Formation and recent evolution of the active Black Sea sink

The present-day sink area of the Danube River system in the Black Sea has a fairly long geological history of evolution as a large back-arc basin controlled by the subduction of the Neotethys Ocean beneath the Rhodope-Pontides arc (e.g., Letouzey et al., 1977; Okay et al., 1994). The Black Sea is made up of two sub-basins that opened gradually during Cretaceous – Middle Eocene moments of extension and is characterized by oceanic crust overlain by an unusual ~15km thick sedimentary sequence in the middle of both sub-basins (Starostenko et al., 2004; Shillington et al., 2008; Graham et al., 2013; Nikishin et al., 2015). The initial Cretaceous opening of the western sub-basin was continued by Eocene extension recorded in both sub-basins that enlarged the Black Sea and was subsequently followed by the onset of inversion at the scale of the entire basin starting with the late Middle Eocene times (Okay et al., 1994; Nikishin et al., 2003; Dinu et al., 2005; Munteanu et al., 2011). In the western part of the Black Sea the inversion continued and migrated N-wards throughout the late Eocene – Miocene times, cumulating a total shortening in the order of 30-40km. At the same time, the shortening changed polarity along the strike of the basin, from dominantly top to north thrusting in the west (thrusting of the Balkans-Pontides, Romanian offshore and Odessa shelf) to dominantly top to south in the east (thrusting of Crimea – Caucasus system over the Black Sea and in the eastern Pontides), as inferred by outcrop, regional and detailed seismic interpretation studies combined with analogue modelling (e.g., Munteanu et al., 2011; Munteanu et al., 2013).

In the western Black Sea basin, this tectonic history is the background of higher resolution depositional changes linked with often sea-level variations, rapid changes in sedimentation rates and deltaic sedimentation that were in particular important during the Pliocene – Quaternary (Popescu et al., 2001; Cloetingh et al., 2003; Gillet et al., 2007; Lericolais et al., 2009). These rapid changes are linked with the dominantly endemic and endorheic evolution of the Black Sea in the larger framework of the Eastern Paratethys, its present-day connection with the Mediterranean being very recent, likely Holocene (Ryan et al., 2003; Kerey et al., 2004; Hiscott et al., 2007). A large number of local and regional unconformities formed in response to high-frequency sea-level variations, in particularly well mapped in the Quaternary deposits of the deep-sea fans located in the front of major deltas, locally associated with large landslides and mass-wasting deposits (such as Danube, Dniepr or Sakarya, Dinu et al., 2005; Gillet et al., 2007; Oaie and Melinte-Dobrinescu, 2012; Dondurur et al., 2013; Lericolais et al., 2013). Among these unconformity, striking is the large-scale erosion observed at or near the time of transition between Miocene and Pliocene, which is also the time of the Messinian Salinity Crisis (MSC) of the Mediterranean. The erosion was observed by multiple seismic interpretation studies calibrated by biostratigraphy in exploration wells and locally reaches kilometres in size by incising along deep canyons the pre-existing sediments deposited in the Black Sea (e.g., Dinu et al., 2005; Tari et al., 2009; Munteanu et al., 2012; Dondurur et al., 2013; Nikishin et al., 2015). Such widely distributed large-scale erosion can only be sub-aerial, particularly well observed near the Turkish, Bulgarian and Romanian offshore.

3. The linked evolution of the Pannonian/Transylvania/Carpathians foreland basins and the separating gateways

The Oligocene – Early Miocene onset of extension in the Pannonian basin areas situated adjacent to the Dinarides and South Carpathians has resulted in the deposition of a continental alluvial to lacustrine endemic series. This was followed subsequently by larger subsidence during the second phase of Middle Miocene extension, resulting in connecting the Pannonian basin with the marine domain situated at the exterior of the Carpathians (Rögl, 1999; Horváth et al., 2006; Matenco and Radivojević, 2012). By continuing the earlier Cretaceous-Eocene orogenic processes of the Carpatho-Balkanides, the subsequent Miocene exhumation of the Carpathian Mountains has separated again the Pannonian and Transylvanian basins from the ones situated in the orogenic foreland. The complete isolation of the Pannonian and Transylvania basins was achieved with the onset of the endemic lacustrine fauna dominant during Pannonian times (e.g., Magyar et al., 1999a). Magneto-stratigraphic and absolute age dating suggested that this isolation took place between 11.6-11.3Ma most likely in two steps, at first by the separation of the Pannonian basin followed 330Ky later by the separation of the Transylvanian Basin (ter Borgh et al., 2013 and references therein). The isolation of the Pannonian basin ended towards the end of the Miocene times, when the uni-directional Pontian fauna transfer from the Pannonian Basin is recognized in the entire Paratethys. At higher resolution numerous short-lived events of marine fauna in endemic environment or endemic fauna in marine environments have been recognized in particular in the Carpathians foreland (Marunteanu and Papaianopol, 1995; Stoica et al., 2013).

This first-order connectivity interplay between the back-arc and foreland basins means at first connection by regional subsidence followed by isolation by orogenic uplift separating the basins. This was subsequently followed by one other event of connection either by subsidence or erosion of the barrier separating the basins. The earlier described thermochronological and kinematic studies indicate continuous shortening and exhumation through the Miocene-Quaternary times in the East Carpathians. Similarly, continuous exhumation was recorded in the SE Carpathians until around 8-9Ma, followed at first by latest Miocene - Pliocene subsidence and subsequently by Pliocene – Quaternary uplift. Exhumation studies indicated that the Apuseni Mountains were quite stable during their Miocene –Quaternary evolution and therefore have an unlikely role in the interplay at regional level. Therefore the only parts of the orogen that recorded significant differential vertical movements during this entire history is at first the South Carpathians during Oligocene - Miocene times (Oligocene-Middle Miocene extension and Middle – Late Miocene transpression) followed by the SE Carpathians during the latest Miocene -Quaternary times. The first area is potentially interesting for all three events while the latter for the last two events.

3.1. Opening and filling the SE part of the Pannonian basin

The SE Pannonian basin area adjacent to the South Carpathians and Dinarides was recently analysed in the regional study of Matenco and Radivojević (2012) and in the high-resolution study of ter Borgh et al. (2015) (Fig. 5). The formation of half-grabens was accompanied by large-scale extensional exhumation in the footwall of the previously described large-scale crustal detachments that affected the margin of the Dinarides. These detachments are laterally buried beneath the sediments of the SE Pannonian basin, where their hanging-wall geometry is depicted by seismic interpretation (such as the Bukulja or Morava detachments, Fig. 5c). Typically, they form half-grabens (e.g., Pancevo Depression) or more symmetric-looking geometries (e.g, Zagajica Depression, Fig. 5c). The latter is just apparent, as the much larger scale footwall exhumation along the main Morava Detachment has been subsequently followed by a late stage normal fault on the eastern flank of the basin. One other typical geometry is the formation of apparent symmetrical structures with low-angles dipping flanks (e.g., Morovici Depression, Fig. 5b). In more details, it is obvious that the syn-kinematic fill is asymmetric and indicates a migration in time of the normal faulting in the half-graben, accompanied with footwall erosion. Such symmetric low-angle structures with asymmetric basin fill are fairly common at the larger scale of the Pannonian Basin, where large scale crustal detachments were demonstrated near the Alps (area of the Little Hungarian plain, Tari et al., 1992) or speculated near the South Carpathians and Apuseni Mountains (such as the Mako-Tomnatec trough, Fig. 4, Tari et al., 1999; Magyar et al., 2006). In the latter area, the extension started during Early Miocene times near the Dinarides, continued everywhere in the basin during Middle Miocene and ceased during late Miocene times in an area close to the Apuseni Mountains and South Carpathians (Fig. 4). The first and last stages of extension were associated with the formation of detachments and half-grabens, while the second Middle Miocene stage of extension was more symmetric, resulting in the widespread formation of grabens across the basin.

The higher resolution has demonstrated that the syn-kinematic sedimentation can be genetically associated with phases of acceleration or deceleration of offsets along the normal faults by forming typical progradational-retrogradational sequences during a rift-cycle transition from initiation, climax to post-rift (Fig. 5b,c, see also ter Borgh, 2013). The large-scale progradational infill has started by building the first shelf-margin slope in the NW part of the Pannonian basin at 10Ma and has migrated 400km to the SE and S roughly along the present day trace of the Danube and Tisza rivers at a rate of ~70km/Ma slope advance until ~4Ma (Magyar et al., 2013). The high-resolution study in the SE part of the basin has demonstrated that this was indeed asymmetric, but was associated with shorter progradation patterns from the South Carpathians-Apuseni Mountains and the Dinarides in N, NW and W direction (Fig. 5a). The cyclicity observed in the progradation patterns shows that the fluctuations of the lake level had limited amplitudes being highly influenced by the pre-existing extensional geometry of (half-)grabens and uplifted areas, as well as the continuous subsidence driven by the extensional thermal sag, in particular large in the SE part of the basin (Fig. 5d). The outflow at the South Carpathians barrier and decrease of the lake level drop due to its erosion that started likely at around 8Ma was one other factor influencing sedimentation. When adding the significant climatic variations that took place during the progradation, the result is a complex interplay of factors influencing the accommodation space during the progradation (Fig. 5d). The rates of all these internal and external forcing factors were often in the same order of magnitude and the cumulated changes in the order tens to couple of hundreds of metres. The effects of the Messinian Salinity Crisis cannot be discriminated in this part of the basin in this fine interplay between multiple parameters given the usual 25-50m resolutions of reflection seismics. This is likely a conclusion that can be extended in the Pannonian domain, as the latest Miocene – Quaternary tectonic inversion that hampers the understanding of MSC sea-level variations elsewhere is reduced in this part of the basin.

3.2. Basin deformation and infill between the South Carpathians and Balkans

Although exhumation data are largely missing, all available geological information infer that the external part of the Balkans were relatively stable after their final orogenic phase of Eocene thrusting. Therefore the tectonic mechanisms driving vertical motions during the Miocene-Quaternary infill in the external part of the Carpatho-Balkans orogen are driven by the evolution of the South Carpathians segment and their foreland. A representative wide-angle (i.e. deep imaging) seismic reflection profile illustrating the mechanical interplay between the orogen and its foreland is available in the recent study of Krézsek et al. (2013) (Fig. 6). The profile illustrates the oblique emplacement of the external thin-skinned wedge of the South Carpathians, locally known as the Getic Depression, overlain by the post-tectonic sediments of the Dacian Basin. The total amount of shortening is in the order of 30km. The Middle - late Miocene thrusting of the thin-skinned wedge peaked at around 14Ma and at around 11Ma, as evident from the syn-kinematic wedge-shaped deposition. This was followed by the onset of regional subsidence affecting the entire foreland throughout the latest Miocene – Quaternary, resulting in a gradual transgression and sedimentation over the thin skinned wedge (Fig. 6). In this frontal part of the South Carpathians this subsidence gradually decreased with time (Matenco et al., 2003). The initial frontal emplacement was followed by out-of-sequence deformation that uplifted and eroded large parts of the centre and internal parts of the thin-skinned wedge. Note that instead of normal faulting in the internal part of the wedge (near the Govora fault in Fig. 6) the alternative interpretation of coeval thrusting deformation is more likely given the field structures and patterns of syn-kinematic sedimentation (Rabagia et al., 2011). This phase of out-of-sequence thrusting and/or transpression took place during the latest Miocene (late Sarmatian – Early Meotian at around 9-8Ma), which is roughly coeval with the last moment of thrusting and exhumation in the East and SE Carpathians. The last phase of tectonic deformation is recorded in the internal part of the thin-skinned wedge during Quaternary times, when Pliocene strata were tilted either along the flanks of wide-open folds or by the uplift of the South Carpathians (Fig. 6, see also Rabagia et al., 2011). This is coeval with the last much larger phase of contraction and exhumation affecting the SE Carpathians that is still presently active. Summarizing, four main moments of peaks uplift took place during orogenic shortening in the South Carpathians and their foreland (at around 14, 11, 9-8 Ma and during Quaternary) and were juxtaposed after 11Ma over the much wider subsidence that affected the entire Moesian platform.

The evolution of the accommodation space in the Moesian platform area adjacent with the South Carpathians was analysed by the seismic interpretation study of Leever et al. (2010). The transect of the study connecting two differently oriented areas of the South Carpathians has been chosen in such a way to avoid the effects of local tectonics and link the area deformed by significant Miocene – Quaternary shortening with the less affected one adjacent to the Pannonian Basin across the mountain chain (Fig. 7a). Incidentally, this profile is connected with the outlet of the Danube at its Iron Gate crossing over the South Carpathians. A large-scale progradation system filling two local depocentres is observed in the profile. This progradation is organized in two stratigraphic sequences (Upper Meotian – Lower Pontian and Middle Pontian – Quaternary) separated by erosional unconformities and their correlative conformities in the deeper part of the basin (Fig. 7b). These sequences were interpreted to reflect absolute changes in the water level, either variation in the local water budget of the isolated basin (i.e. the balance between water influx and evapo-transpiration) or regional changes in sea-level (i.e. affecting also the surrounding Paratethys basins). The onset of the second sequence during the Middle Pontian took place at or near the times of the

MSC of the Mediterranean and would imply an sea-level drop in the order of 200m that affected the Dacian Basin. By correlating with the roughly coeval moments of sea-level drop with different amplitudes proposed for the Pannonian Basin and the Black Sea, the study proposes a connectivity model in which these different amplitudes are driven by the separating barriers and their erosion during the sea-level drop (Fig. 7c). Once the sea level of the Black Sea had dropped below the separating Scythian gateway, the main parameter that controls the sea-level in the Dacian basin is the erosion of this barrier. The long period of sea-level drop during the low-stand would be explained by a positive water balance of the Dacian Basin. The same mechanism would be implied for the connection between the Dacian and the upstream Pannonian basin, the MSC sea-level drop of the latter being lower or equal. Interesting is the observation that, at the resolution of seismic lines, the progradation in the basin is asymmetric in respect to the position relative to the mountain chain. In front of the Danube outlet at the Iron Gates the upper Meotian – Lower Dacian progradation is ~70km long, while elsewhere in the profile is less than half in size. Our own observations indicate that the average progradation size in seismic lines elsewhere in the basin is in the order of 10-30km, confirming the asymmetry. This is valid even when compared with areas located E-wards in the front of the South Carpathians, which recorded higher tectonic-driven exhumation values during the Miocene –Quaternary shortening events. This suggests a much larger source area for the progradation located in front of the Danube outlet, implying that a connection with the Pannonian basin existed during progradation. This reasoning has obvious limitations, such as the resolution of seismic lines or mechanisms generating slopes for progradation.

The connectivity hypothesis has been tested in the area of the Danube outlet at the Iron Gates by the local high-resolution study of ter Borgh et al. (2014) (Fig. 8). This area was affected by the large-scale ~65km offsets along the Timok faults that were at first transtensional followed by transpressional (Fig. 8a). Based on observations in outcrops in the transitional area from the orogen to the basin combined with seismic interpretation, the study has demonstrated the significant link between vertical movements in the South Carpathians and depositional geometries and their evolution in the neighbouring Dacian Basin. A number of key events in the evolution of the basin were driven by tectonics and, to a lesser extent, by regional sea-level variations (Fig. 8b). A two-way connection between the Pannonian and Dacian basins existed already during Middle Miocene (late Badenian, until 11.7Ma) times. The area of the Dacian Basin near the Iron Gates area was affected by a relative sea-level drop and the formation of a late Miocene (Late Sarmatian – early Meotian) unconformity starting with 11.7Ma. This is the main moment when the Getic Depression was affected by shortening, at first by emplacement of the main nappe and later by out-of-sequence thrusting (Fig. 6). This is also when the main uplift of the East and SE Carpathians was recorded. The result was the formation of this regional unconformity, otherwise commonly observed along the Carpathian Mountains. The study infers subsequent moments of short-lived connection between the Pannonian and Dacian basins that were likely in one direction (i.e. from the Pannonian to the Dacian). In latest Miocene (Pontian) times, an initial transgression event was followed by a regional sea-level fall, whose amplitudes are in the order of 50-200m, although a realistic value is likely in lower part of this spectrum and, therefore, lower than the ones inferred by the study of Leever et al. (2010). The large-scale late Miocene – earliest Pliocene (Upper Meotian – Lower Dacian) progradation detected by the latter study in front of the Iron Gates, whose size is unique along the South Carpathians, does not appear to be accompanied by changes in depositional environment and fauna until Pontian times in the same area, when the fauna became common along the Central and Eastern Paratethys.

3.3. Modelling the connection between the Pannonian and Dacian basins.

The connectivity between the Pannonian and Dacian basins by a gateway located in the intervening area of the South Carpathians was modelled in the study of Leever et al. (2011) (Fig. 9). This study used a pseudo-3D (planform) forward difference code in which tectonics, surface processes and flexural isostasy are fully coupled (Garcia-Castellanos, 2002). The study was designed to account the interaction between surface processes and their lithospheric scale response in terms of vertical motions on large spatial and temporal scales. Therefore, it allows only first order predictions on the evolution of gateways or connectivity events between sedimentary basins. In the model setup (Fig. 9b), the Pannonian and Dacian basins are in restricted connection over a separating topographic high that simulates the South Carpathians. The Pannonian basin is restricted on both sides, while water circulation is allowed at the eastern end of the Dacian Basin (Fig. 9b). For further details on the model setup we refer to Leever et al. (2011). The model has tested a number of critical parameters, such as the difference between the water levels in the two basins, the barrier height, the flexural rigidity of the lithosphere underlying the basins controlling the accommodation space via sediment supply, or the balance between precipitation and evapo-transpiration.

This numerical modelling has demonstrated that the key factors controlling the sedimentary response are the upstream (i.e. Pannonian Basin) accommodation space and its elevation at the time of capture. One important parameter is the flexural rigidity through its control in accommodation space is of major influence in depositional geometries and sediment distribution (Fig. 9c). This controls both the timing and the rate of the bulk sediment shift from the upstream to the downstream basin. A strong lithosphere in the upstream basin will lead to a quick shift while in the case of a weak lithosphere the response is more gradual. The model infers that a uni-directional fluvial connection from the Pannonian to the Dacian basins should be marked by increased sedimentation rates in front of the gateway, even if this connection occurred before the Pannonian lake was completely filled. In other words, the model predicts that the large scale late Miocene sedimentation observed in front of the Iron Gates would reflect the connectivity and could have started earlier that the complete fill of the Pannonian basin at the end of Miocene times. The effects are largely dependent on the upstream basin accommodation space left at the time of the capture and the elevation difference between the water levels in the two basins at the time of capture (Fig. 9d). In all cases the time lag between the capture and bulk sedimentation shift is dependent on the underlying lithospheric strength (by the effective elastic thickness Te in Fig. 9d). By comparing the numerical modelling with the above-described observations in both basins, the modelling infers that at the time of the basin capture the accommodation space left in the upstream Pannonian Basin was low, the elevation difference between the two water-levels was small and the increase of sedimentation rates took place gradual after a longer time lag (panel lower left in Fig. 9d). The latter was controlled by the low flexural rigidity of the lithosphere underlying the Pannonian basin, resulting from the elevated thermal structure of this extensional domain. Obviously, this regional model cannot fit the much higher resolution of outcrop and seismic studies on both sides of the gateway, such as the prediction of a permanent uni-directional connection between the two basins that it at odds with the episodic influxes of fauna prior to its ultimate generalisation in both basins during Pontian times. Similarly, the observed moments of transgression in the Dacian basin or the fluvial sedimentation that kept balance with the on-going thermal subsidence in the Pannonian basin were not simulated in this model. But the overall concept of time lag is very important in understanding connectivity: bulk sedimentation shifts can significantly post-date the moment of establishing a connection. Even when the later decreases the accommodation space by eroding the separating barrier, the upstream basin needs time to fill up prior to bulk sedimentation shifts.

3.4. Recent subsidence and associated sedimentary processes in the SE Carpathians foreland

The external part of the SE and South Carpathians displays atypical foredeep geometries (Figs. 4 and 6). These show foredeep wedges localized only in the Upper Miocene (Middle – Late Sarmatian) and were overlain by subsequent thick latest Miocene – Quaternary mostly post-tectonic sedimentation of the Dacian Basin. This sedimentation was driven by the continuous subsidence of the Moesian platform in post-collisional times (Matenco et al., 2003; Matenco et al., 2007), being responsible for the low and flat topography that presently dominates the relief in most of the Dacian basin (Fig. 2).

The driving mechanisms of this subsidence cannot be quantified by the study of the upper crustal structural geometries of the Moesian platform, but it is certainly driven by the evolution of the Vrancea slab. Modern high-resolution local tomography and modelling studies (Martin and Wenzel, 2006) has demonstrated that the high-velocity anomaly that is often associated with the Vrancea slab is much larger, extending from the SE Carpathians in a SW-ward direction well beneath the Moesian platform and can be divided roughly in two segments. While the Vrancea segment is (barely) still attached, the anomaly beneath the Moesian platform is apparently detached from the Moesian lithosphere (Heidbach et al., 2007). This is a typical geometry that in other similar collisional settings has been interpreted as a STEP (subduction-transform edge propagator) fault. Such lithospheric-scale structures have been often invoked in curved roll-back systems, such as in the evolution of the Apennines Calabrian slab (Govers and Wortel, 2005). In the case of the South and SE Carpathians, this would mean that an lithospheric scale tear in the Carpathians subducting slab propagated along the South Carpathians during Miocene – Quaternary times, eventually arriving in the present day configuration. Such an hypothesis is still highly speculative and requires more research on the coupling between the mantle dynamics and near-surface deformation (Ismail-Zadeh et al., 2012).

Limited to the SE Carpathians area located between the Intramoesian and Trotus faults and their WNW-ward prolongation, one other mechanism is juxtaposed over the longer-term pattern of subsidence, the latter reaching extreme values in the centre of the Focsani basin (~6km of Upper Miocene – Quaternary sediments, Fig. 4). Starting with the late Pliocene - Quaternary, large-scale coeval differential vertical motions were recorded by ~5km of uplift of the external Carpathians nappes and up to 2km of subsidence in the neighbouring Focsani basin, which accommodated a total amount of shortening in the order of 5 Km (Leever et al., 2006; Matenco et al., 2007; Necea, 2010). This process is still highly active today: outside the well known intermediate mantle Vrancea earthquakes (Ismail-Zadeh et al., 2012 and references therein), there is a direct correlation between the crustal seismicity and the activity of recent faults derived from geophysical or neotectonic studies. For instance there is a direct correlation of crustal seismicity with the thick skinned thrusting beneath the external nappes of the SE Carpathians (Bocin et al., 2009), or with the activation of the large array of normal faults farther in the foreland linked with the 2013 active seismicity of the Galati area (Matenco et al., 2007). These studies are important for prediction of natural hazards, such as seismicity or flooding. For instance, the correlation of active faulting with the vertical movements derived from GPS studies resulted in predicting the evolution of the area threatened by flooding in the foreland of the SE Carpathians (Fig. 10). This prediction infers a clear pattern of acceleration of flooding and rivers instability, which is in agreement with existing geomorphological studies (Radoane et al., 2003).

4. The combined Carpathians foreland – western Black Sea system

One situation in which the concept of time lag can be best applied is the connected Dacian Basin – western Black Sea system during Miocene –Pliocene times. Numerous studies were previously focussed on the tectonic evolution and related sedimentation in both areas, but these were analysed individually (Matenco and Andriessen, 2013 and references therein), lacking a coherent integration of both areas in terms of distribution of sediments relative to the dominant Carpatho-Balkanides source area.

4.1. Connectivity between the Dacian Basin and the Black Sea

The structural evolution and sedimentary architecture of the connected Dacian Basin – Black Sea system was analysed recently in the study of Munteanu et al. (2012). Using the advent of a readily available detailed geometry and structure in the Carpathians foreland (Tărăpoancă et al., 2003; Tărăpoancă et al., 2004; Leever et al., 2006; Matenco et al., 2007) combined with a novel high resolution seismic stratigraphic interpretation in the western Black Sea, the study has quantified for the first time the moments and effects of connectivity between the two basins (Fig. 11). The analysis of a large scale cross section connecting the Carpathians and Focsani basin over the Dobrogea high with the western Black Sea (Fig. 2c) suggests that the main depocentre trapping sediments from the Carpathians source area was dominantly the Focsani basin during Miocene – Quaternary times. One exception is observed during the latest Miocene – earliest Pliocene (Meotian-Pontian), when massive sedimentation locally associated with large-scale progradation features is recorded in the western Black Sea (Fig. 11b,c). This suggests that the available accommodation space in the Dacian Basin was cancelled and the sedimentation was shifted towards the Black Sea. At the end of this period, sedimentation was again available in the Dacian Basin and the bulk of sedimentation shifted backwards in the Carpathians foreland. Large scale erosional features were observed in the entire Black Sea near the transition between the Miocene and Pliocene times, locally creating a spectacular paleo-relief depicted by the interpretation of seismic lines correlated with exploration wells (Gillet et al., 2007; Tari et al., 2009; Dondurur et al., 2013; Nikishin et al., 2015). The elevation differences of the buried topography reach locally more than 1km. This is obviously the result of sub-aerial erosion, this pattern being observed also in the deep-water part of the Black Sea. The unconformity formed likely during Pontian times (near the transition between the Miocene and Pliocene) due to a regional sea-level drop of the Black Sea. Whether or not the timing of this unconformity fits a higher resolution the Messinian Salinity Crisis of the Mediterranean is still a matter of debate. The sequential stratigraphic analysis of Munteanu et al. (2012) has demonstrated that the sea-level drop was more than 1Km in amplitude by quantifying coeval sedimentation patterns that change from erosion to mass-flows and turbiditic deposits in the deep-sea part of the Black Sea (Fig. 12). More important, this study links these large-scale erosional features observed in the Black Sea near the Miocene-Pliocene limit with a similar event in the Dacian Basin and proposes that the associated large-scale variations in the sea level controlled the accommodation space in the system of the two connected basins (Fig. 12c). In the case of a large sea-level drop, an outward sedimentation shift will occur from the trapping basin to the main sink being controlled by the height of the separating barrier (Fig. 12c). For a sea-level drop of 1.3–1.7 km and a barrier height of ca. 200 m, the sedimentation shift will occur after ca. 0.2 Ma, which is the time required for the accommodation space to be filled in the upstream Dacian Basin. An inward sedimentary shift will occur almost instantaneously, when the water level surpasses the barrier, and accommodation space is again created in the trapping basin (Fig. 12c). These observations demonstrate again the importance of sedimentation shifts in high-resolution studies and the fact that erosional unconformities driven by regional mechanisms are diachronous across connected basins due to the time required to fill available accommodation space. One other significant outcome of this study is the demonstration that large-scale erosional features formed during large sea-level drops condition the subsequent distribution of deep-sea sedimentation. The mass-and turbiditic-type of transport, presently active in front of rivers discharging into the Black Sea is an inherited effect of the large sea-level drop recorded near the limit between Miocene and Pliocene times (Fig. 12b). These deep-sea deposits were more sensitive to higher order sea-level variations than contemporaneous shelf sequences, thus capturing higher resolution changes on seismic data.

This concept of sedimentation shifts with application to the connected Dacian Basin – Black Sea system has been analysed by the means of numerical modelling in the study of Bartol et al. (2012) (Fig. 13). The analysis uses a pseudo-3D numerical modelling code that takes into account the tectonic evolution of basins in terms of subsidence and flexural distribution combined with sedimentation patterns and variations in the balance between precipitation and evapo-transpiration (Garcia-Castellanos, 2002). The modelling setup simulated the late Miocene – Pliocene situation of connected Dacian Basin – western Black Sea by allowing one source area feeding a trapping basin separated from the main sink by a submarine barrier (Fig 13a,b). By assuming a realistic erosion and sedimentation rate combined with an approximation of the natural scenario, the main parameters tested in the model were the barrier height, amplitude of sea-level variations, the lithospheric strength and the influence of climate. As a function of barrier height and cycle duration, the model predicts situations of no depositional shifts (no significant change in sedimentation rates), partial depositional shifts (change in sedimentation rates, but remains higher in the trapping basin) or full depositional shift (sedimentation rates are larger in the main sink between the shifts) (Fig. 13c). The occurrence of depositional shifts is controlled by the moment of sediment equilibration at the barrier, while their type being dependent of the time available for the system to respond to the equilibration. In the situation without a barrier, an outward and inward depositional shift occurs when the sea level falls below or rises above the floor of the trapping basin, cancelling or creating its accommodation space. When a barrier is present an outward depositional shift will be delayed due to the time required to reach the moment of sediment equilibration at the barrier. The time difference between the onset of a sediment shift in the situation without a barrier and the one with a barrier is called time-lag. Actually the time lag means delaying the sedimentation shift and the main effects of the sea-level drop and its crucial in understanding effects in connected basins. The time lag varies between 0.1-0.35Ma for the studied natural scenario, increases with the duration of the sea-level cycle and with lower amplitudes of the sea-level drop and it is not significantly influenced by the barrier height (Fig. 13d). In the studied natural scenario the modelling predicts that the time span recorded by the sub-aerial unconformities near the limit between Miocene and Pliocene should be in the order of 0.4–0.5 My in the Black Sea and the mountainous margin of the Dacic basin, and 0.2–0.3 My in the centre of the trapping Dacian Basin. Given this timing and the observed full depositional shifts between the Dacian trapping basin and the main Black Sea sink, the numerical modelling predicts magnitudes of sea level drop larger than 1000m. These modelling inferences are in agreement with the above described observations.

4.2. Higher resolution studies in the western Black Sea

The Miocene – Pliocene rapid changes in sedimentation patterns driven by changing conditions in external and internal forcing factors are also reflected in the recent evolution of the Black Sea, as depicted by shallower high-resolution studies.

A significant number of such high-resolution studies have been dedicated to the analysis of the dynamic sedimentation that took place during Quaternary times along the deep-water discharge fans situated in front of the main rivers of the western Black Sea (Fig. 14a, Popescu et al., 2001; Popescu et al., 2004; Popescu et al., 2006; Lericolais et al., 2009; Lericolais et al., 2010; Dondurur et al., 2013; Lericolais et al., 2013). Following the onset of this type of sedimentation near the limit between Miocene and Pliocene times, a significant number of stratigraphic sequences were detected in these turbidictic fans at various orders of resolution. For instance, a correlation of high-resolution shallow seismic lines and facies observed in wells suggests a direct correlation between stratigraphic sequences and the cyclicity of the last glacial maximum and deglacial to more recent times (Fig. 14b). The study of Lericolais et al. (2013) has inferred that the last glacial maximum deposits consist of thick mud turbidites overlain by the by the lobe deposits of the Danube deep-sea fan, sediments supplied by the Danube bypassed the shelf to the deep sea through the deeply incised Viteaz canyon. This deep sea progradation is linked with a significant change in the sediment supply of the Danube River combined with the deposition directly in the deep sea when the sea-level was close to the position of the shelf break. At high stands, this construction records only condensed pelagic sedimentation, a model that is in agreement with the general deposition of deep sea turbidites (e.g., Posamentier, 2003). Overlying these deposits in the Danube deep sea fan, the deposition of thick turbidites associated with reworked red continental clays suggests mixing of a Danube source with another source located more to the south during deglacial times (~18 – 15.5 ka BP). A period of carbonate rich deposition was possibly linked with a period of reversing water balance and sea-level drop to ~100m or lower below the present day position following the melt-water driven highstand. This was followed by the reconnection with the marine realm and the onset of deposition of sapropel deposits in the deeper parts of the Black Sea, correlated with a major change in magnetic minerals to greigite in particular at the lacustrine/marine transition (Strechie et al., 2002; Major et al., 2006). Interestingly, the oldest date of 7.4 ka of the first sapropel deposit does not coincide with the onset of marine incursion from the Mediterranean across the Bosporus strait, which is dated at ~9.4-9 ka BP. But it coincides with the time when the salinity rose to the sufficient level required for the depletion of bottom water oxygen and sapropel deposition. This is another type of time lag demonstrating yet again that connectivity events are not coeval with the onset of expected generalized conditions in sedimentation.

The other major source of sediment supply in the western Black Sea system outside the Danube/Dniepr system is the drainage system of the western Pontides along its northern fluvial system, such as the Sakarya and neighbouring rivers. Their influx has been recently analysed in the study of Dondurur et al. (2013) (Fig. 14b). The rapid environmental changes coupled with the active tectonics occurring in the Pontides by their thrusting over the Black Sea or the large-scale movements of the North Anatolian Fault have resulted in significant sedimentological variations and a high level of instability of the sedimentary wedge deposited over the passive continental margin towards the deep sea part of the western Black Sea basin. By analysing a number of unusual high quality shallow seismic profiles combined with multi-beam bathymetry and Chirp seismic data, this study has demonstrated the high instability of the southern margin of the western Black Sea. This is characterized by significant sediment erosion associated with major sediment slides and buried debris lobes in the offshore Asmara region (Fig. 14b). The sliding includes failures in the steep slope zones, smaller slides on the canyon walls, and relatively larger slides in the Asmara mass failure zone. Block - type sliding formed along listric faults due to gravitational loading over steep slope zones possibly triggered by local seismic activity. A number of large debris lobes were buried in Quaternary sediments, formed by excess pore pressures due to high sediment input and submarine fluid flow. The active tectonics triggered the sliding, facilitated by excess pore pressures in shallow sediments due to the submarine fluid flow possibly produced by gas hydrate dissociation. The latter is most probably the result of the penetration in the Black Sea of warmer Mediterranean water during the rapid transgression after the Last Glacial Maximum.

5. Discussion and conclusions

Our review demonstrates that the vertical movements recorded by the Carpathian Mountains have driven not only the formation and evolution of the associated basins but also their subsequent fill along roughly the present trace of the Danube. The least controlled connection between basins is the one that took place during Middle Miocene (Badenian) times when the onset of marine conditions is recorded in Pannonian and Transylvania basins and these become connected with the Carpathians foreland (e.g., Rögl, 1999 and references therein). Thermochronology demonstrates that the East and SE Carpathians were already affected by regional tectonic exhumation at the time of the connection, which started earlier during the Oligocene-Early Miocene. A similar situation is observed in their northern prolongation towards the Western Carpathians, although thermochronological data in this area are less available. We cannot exclude, obviously, the existence of individual corridors, but associated sedimentation connecting the basins is missing or was removed by subsequent erosion. Therefore it is likely that the connection formed via the South Carpathians, which leads us to an apparent paradox. Thermochronology indicates that the first order topographic patterns of the South Carpathians were established already during Cretaceous – Paleogene times (Merten, 2011). The few overprinting Miocene ages are focussed mainly in the western area of the South Carpathians adjacent to the Pannonian Basin and infer a period of pre-dating burial and coeval exhumation. The former is likely the effect of the Eocene orogen parallel extension and large-scale transtensional movements accommodating the movements into the Carpathian embayment. It is likely that the associated subsidence ultimately created a connection in this area between the Central and Eastern Paratethys. The subsequent severing of this connection during late Miocene times at 11.6-11.3My (ter Borgh et al., 2013) and the formation of the endemic Pannonian lake (Magyar et al., 1999a) took place by tectonic uplift of the same South Carpathians, but this uplift was restricted to ~2-5Km in such a way that the first order Cretaceous Paleogene topography was still preserved.

One other interesting situation is observed near the limit between Miocene and Pliocene times at the time of sedimentation shifts between the Dacian Basin and the Black Sea. In most of the Dacian Basin a clear sea-level drop associated with an erosional unconformity in proximal areas is observed. Such a regional unconformity is not obvious in the rapidly subsiding Focsani basin. The sea-level drop has been detected in its proximal parts, but seismic interpretation studies were not able to detect an erosional unconformity in the basin (Tărăpoancă et al., 2003; Matenco et al., 2007; Stoica et al., 2013) that may be associated with cancelling the accommodation space and shifting the sediments to the Black Sea. This is because of the rapid subsidence that kept the continental base level (the point of transition between continental erosion and deposition) at higher topographic elevations than the present remainder of the Focsani basin, highly eroded on its NW flank by the Quaternary uplift of the SE Carpathians (Leever et al., 2006). The same situation is observed at present: thick alluvial sediments are still being deposited in this basin, while the bulk of the sediments is transported by the Danube in the Black Sea (Fig. 8). A similar situation is observed in the SE part of the Pannonian near the Apuseni Mountains and South Carpathians during the thick deposition of Pliocene – Quaternary alluvial sediments that presently continues (Fig. 4, Juhasz et al., 2007; ter Borgh et al., 2015). A marked erosional unconformity is missing in this area although the bulk of the Danube sediments have crossed the South Carpathians on their way to the Black Sea. These observations demonstrate that simple stratigraphic observations are unable to quantify connectivity moments, which requires high-resolution sequence stratigraphic studies coupled with exhumation and landform evolution studies over the gateways separating sedimentary basins.

The Miocene-Quaternary evolution of the sedimentary basins presently filled and aligned along the lower trace of the Danube River system (Pannonian/Transylvania and Dacian) together with the coeval sedimentation patterns observed in the western part of the Black Sea are the result of rapid tectonic changes taking place in the Carpathians Mountains that still influence at present the transfer of sediments towards the main sink. The spatial and temporal scales of various processes is contrastingly different in the entire system. Long-term tectonic processes (in the order of 10^6My) drive the formation and evolution of the main basins, such as the back-arc extension of the Pannonian Basin, the sag subsidence of the Transylvanian Basin, the flexural subsidence of the Carpathians foreland or the inversion of the western Black Sea. However, threshold conditions that result in basin separation or connectivity events demonstrate that such processes take place in spatially restricted areas with rates of around one order of magnitude faster (in the order of 10^4-10^5My) than background tectonic processes. Relevant examples include the connectivity across the South Carpathians or in between the Dacian Basin and the western Black Sea. Locally, some tectonic or sedimentological processes can accelerate and encompass the scale of connectivity events, such as the ~70km/My deltaic slops advance in the shallow Pannonian lake or the 2.5km/My Quaternary exhumation rate in the SE Carpathians. The connectivity processes associated with geochemical or environmental changes act in an even narrower depth interval (tens of metres) and at higher temporal resolution (in the order of 10^3-10^4My). Relevant examples include environmental changes induced by the recent opening of the Black Sea gateways towards the Mediterranean or the rapid geochemical changes that were inferred for the Miocene connectivity events of the Mediterranean and Paratethys (e.g., Karami et al., 2011). Understanding basin evolution and parameters driving changes across basins in such a staggering range of spatial and temporal scale is very challenging, to say the least. The solution lies in controlling a wide range of parameters, from the tectonic and sedimentological background, threshold conditions to geochemical changes, which obviously implies collaboration between a wide range of disciplines.

Such situations of formation of isolated basins followed by connectivity and sedimentation shifts are rather common during orogenic evolution, in particular when its late stages is driven by slab retreat. The simple Miocene translation of the continental unit made up by the Apuseni Mountains, Transylvania Basin together with East, SE and South Carpathians, collapsing by extension the back-arc and shortening the foreland and successively filling them is not unique. It is also commonly interpreted in most of the other highly curved Mediterranean orogens. Examples include the W-ward movement of the Betics-Rif system and coeval extensional collapse of the Alboran Domain of the Western Mediterranean and its onshore Miocene basins, driven by the roll-back of the Gibraltar slab (Vergés and Fernàndez, 2012). This rapid roll-back was intimately related with the isolation and subsequent re-connection of the Mediterranean with the Atlantic over the Gibraltar gateway during and after the Mediterranean salinity Crisis (e.g., CIESM, 2008). Similar multiple basins opening events took place during the W-ward movement of the Calabrian slab accommodated by Oligocene-Quaternary shortening at the exterior of the Apennines and coeval back-arc extension of the oceanic Liguro-Provencal and Tyrrhenian (Faccenna et al., 2005; Faccenna et al., 2014). One other much larger example is the ~900km of S-ward translation of Aegean units that was accompanied by ~500km of N-S Europe – Africa convergence, while the amount of subducted material in the mantle is in the order of 1500 Km (van Hinsbergen et al., 2005; Jolivet and Brun, 2010). One other example is the gradual opening of the SE China Sea and its multiple moments of inversion and closure that still take place today (Morley and Westaway, 2006; Pubellier and Morley, 2014). All these situations contain (sub-)basins affected by moments of isolation and connectivity that are not well quantified, nor properly understood. The Danube – Black Sea system in the framework of the larger Paratethys can serve as a good example, whose mechanics and processes can certainly be ported elsewhere.

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References

Bada G, Horváth F, Dövényi P, Szafián P, Windhoffer G, Cloetingh S. Present-day stress field and tectonic inversion in the Pannonian basin. Global and Planetary Change 2007; 58: 165-180.

Balla Z. Palaeotectonic reconstruction of the central Alpine-Mediterranean belt for the Neogene. Tectonophysics 1986; 127: 213-243.

Bartol J, Matenco L, Garcia-Castellanos D, Leever K. Modelling depositional shifts between sedimentary basins: Sediment pathways in Paratethys basins during the Messinian Salinity Crisis. Tectonophysics 2012; 536-537: 110-121.

Bergerat F, Vangelov D, Dimov D. Brittle deformation, palaeostress field reconstruction and tectonic evolution of the Eastern Balkanides (Bulgaria) during Mesozoic and Cenozoic times. Geological Society, London, Special Publications 2010; 340: 77-111.

Bocin A, Stephenson R, Mocanu V, Matenco L. Architecture of the south-eastern Carpathians nappes and Focsani Basin (Romania) from 2D ray tracing of densely-spaced refraction data. Tectonophysics 2009; 476: 512-527.

Bojar AV, Neubauer F, Fritz H. Jurassic to Cenozoic thermal evolution of the southwestern South Carpathians: evidence from fission-track thermochronology. Tectonophysics 1998; 297: 229-249.

Burchfiel BC, Nakov R, Dumurdzanov N, Papanikolaou D, Tzankov T, Serafimowski T, King RW, Kotzev V, Todosov A, Nurce B. Evolution and dynamics of the Cenozoic tectonics of the South Balkan extensional system. Geosphere 2008; 4: 919-938, doi: 10.1130/GES00169.1.

CIESM. The Messinian Salinity Crisis from mega-deposits to microbiology - A consensus report. In: Briand F, editor. CIESM Workshop Monographs. 33. CIESM, Monaco, 2008, pp. 168.

Ciobanu CL, Cook NJ, Stein H. Regional setting and geochronology of the Late Cretaceous Banatitic Magmatic and Metallogenetic Belt. Mineralium Deposita 2002; 37: 541-567.

Cloetingh S, Bada G, Matenco L, Lankreijer A, Horvath F, Dinu C. Modes of basin (de)formation, lithospheric strength and vertical motions in the Pannonian-Carpathian system: inferences from thermo-mechanical modelling. Geological Society, London, Memoirs 2006; 32: 207-221.

Cloetingh S, Matenco L, Bada G, Dinu C, Mocanu V. The evolution of the Carpathians-Pannonian system: Interaction between neotectonics, deep structure, polyphase orogeny and sedimentary basins in a source to sink natural laboratory. Tectonophysics 2005; 410: 1-14.

Cloetingh S, Spadini G, Van Wees JD, Beekman F. Thermo-mechanical modelling of Black Sea Basin (de)formation. Sedimentary Geology 2003; 156: 169-184.

Cloetingh SAPL, Burov E, Matenco L, Toussaint G, Bertotti G, Andriessen PAM, Wortel MJR, Spakman W. Thermo-mechanical controls on the mode of continental collision in the SE Carpathians (Romania). Earth and Planetary Science Letters 2004; 218: 57-76.

Csontos L. Tertiary tectonic evolution of the Intra-Carpathian area: a review. Acta Vulcanologica 1995; 7: 1-13.

Csontos L, Vörös A. Mesozoic plate tectonic reconstruction of the Carpathian region. Palaeogeography, Palaeoclimatology, Palaeoecology 2004; 210: 1-56.

Danisik M, Kohut M, Evans NJ, McDonald BJ. Eo-Alpine metamorphism and the 'mid-Miocene thermal event' in the Western Carpathians (Slovakia): new evidence from multiple thermochronology. Geological Magazine 2012; 149: 158-171.

de Leeuw A, Bukowski K, Krijgsman W, Kuiper KF. Age of the Badenian salinity crisis; impact of Miocene climate variability on the circum-Mediterranean region. Geology 2010; 38: 715-718.

Dimitrijevic MD. Dinarides and the Vardar Zone: a short review of the geology. Acta Vulcanologica 2001; 13: 1-8.

Dimitrijević MD. Geology of Yugoslavia, 2nd edition. Belgrade: Geoinstitute, Belgrade, 1997.

Dinu C, Wong HK, Tambrea D, Matenco L. Stratigraphic and structural characteristics of the Romanian Black Sea shelf. Tectonophysics 2005; 410: 417-435.

Doglioni C, Carminati E, Cuffaro M, Scrocca D. Subduction kinematics and dynamic constraints. Earth-Science Reviews 2007; 83: 125-175.

Dombrádi E, Timár G, Bada G, Cloetingh S, Horváth F. Fractal dimension estimations of drainage network in the Carpathian-Pannonian system. Global and Planetary Change 2007; 58: 197-213.

Dondurur D, Küçük HM, Çifçi G. Quaternary mass wasting on the western Black Sea margin, offshore of Amasra. Global and Planetary Change 2013; 103: 248-260.

Ellouz N, Roure F, Sandulescu M, Badescu D. Balanced cross sections in the Eastern Carpathians (Romania): a tool to quantify Neogene dynamics. In: Roure F, Ellouz N, Shein VS, Skvortsov I, editors. Geodynamic evolution of sedimentary basins, International symposium Moskow 1992 Proceedings. Technip, Paris, 1994, pp. 305-325.

Faccenna C, Becker TW, Miller MS, Serpelloni E, Willett SD. Isostasy, dynamic topography, and the elevation of the Apennines of Italy. Earth and Planetary Science Letters 2014; 407: 163-174.

Faccenna C, Civetta L, D'Antonio M, Funiciello F, Margheriti L, Piromallo C. Constraints on mantle circulation around the deforming Calabrian slab. Geophysical Research Letters 2005; 32: L06311, doi:10.1029/2004GL021874.

Faccenna C, Piromallo C, Crespo-Blanc A, Jolivet L, Rosetti F. Lateral slab deformation and the origin of the western Mediterranean arcs. Tectonics 2004; 23: TC1012, doi:10.1029/2002TC001488.

Favre P, Stampfli GM. From rifting to passive margin: the examples of the Red Sea, Central Atlantic and Alpine Tethys. Tectonophysics 1992; 215: 69-97.

Filipescu S, Gîrbacea R. Lower Badenian sea-level drop on the western border of the Transylvanian basin: foraminiferal paleobathymetry and stratigraphy. Geologica Carpathica 1997; 48: 325-334.

Fodor L, Bada G, Csillag G, Horvath E, Ruszkiczay-Rudiger Z, Palotas K, Sikhegyi F, Timar G, Cloetingh S. An outline of neotectonic structures and morphotectonics of the western and central Pannonian Basin. Tectonophysics 2005; 410: 15-41.

Fodor L, Csontos L, Bada G, Györfi I, Benkovics L. Tertiary tectonic evolution of the Pannonian basin system and neighbouring orogens: a new synthesis of paleostress data. In: Durand B, Jolivet L, Horvath F, Seranne M, editors. The Mediteranean basins: Tertiary extension within the Alpine orogen. The Geological Society, London, 1999, pp. 295-334.

Fodor L, Gerdes A, Dunkl I, Koroknai B, Pécskay Z, Trajanova M, Horváth P, Vrabec M, Jelen B, Balogh K, Frisch W. Miocene emplacement and rapid cooling of the Pohorje pluton at the Alpine-Pannonian-Dinaridic junction, Slovenia. Swiss Journal of Geosciences 2008; 101: 255-271.

Fügenschuh B, Schmid SM. Age and significance of core complex formation in a very curved orogen: Evidence from fission track studies in the South Carpathians (Romania). Tectonophyiscs 2005; 404: 33-53.

Garcia-Castellanos D. Interplay between lithospheric flexure and river transport in foreland basins. Basin Research 2002; 14: 89-104.

Gillet H, Lericolais G, Rehault J-P. Messinian event in the black sea: Evidence of a Messinian erosional surface. Marine Geology 2007; 244: 142-165.

Gleadow AJW, Duddy IR. A natural long-term track annealing experiment for apatite. Nuclear Tracks 1981; 5: 169-174.

Golonka J. Plate tectonic evolution of the southern margin of Eurasia in the Mesozoic and Cenozoic. Tectonophysics 2004; 381: 235-273.

Govers R, Wortel MJR. Lithospheric tearing at STEP faults: Response to edges of subduction zones. Earth Planett. Sci. Lett 2005; 236: 505-523.

Graham R, Kaymakci N, Horn BW. Revealing the mysteries of the Black Sea. GEO ExPro 2013; 10: 58-62.

Haas J, Péró C. Mesozoic evolution of the Tisza Mega-unit. International Journal of Earth Sciences 2004; V93: 297-313.

Heidbach O, Ledermann P, Kurfeß D, Peters G, Buchmann T, Matenco L, Negut M, Sperner B, Müller B, Nuckelt A, Schmitt G. Attached or not attached: slab dynamics beneath Vrancea, Romania. International Symposium on Strong Vrancea Earthquakes and Risk Mitigation. Oct. 4-6, 2007, Bucharest, Romania, 2007, pp. 4-20.

Hiscott RN, Aksu AE, Mudie PJ, Marret F, Abrajano T, Kaminski MA, Evans J, Çakiroglu AI, Yasar D. A gradual drowning of the southwestern Black Sea shelf: Evidence for a progressive rather than abrupt Holocene reconnection with the eastern Mediterranean Sea through the Marmara Sea Gateway. Quaternary International 2007; 167-168: 19-34.

Horváth F, Bada G, Szafian P, Tari G, Adam A, Cloetingh S. Formation and deformation of the Pannonian Basin: constraints from observational data. Geological Society, London, Memoirs 2006; 32: 191-206.

Horváth F, Musitz B, Balázs A, Végh A, Uhrin A, Nádor A, Koroknai B, Pap N, Tóth T, Wórum G. Evolution of the Pannonian basin and its geothermal resources. Geothermics 2015; 53: 328-352.

Iancu V, Berza T, Seghedi A, Gheuca I, Hann H-P. Alpine polyphase tectono-metamorphic evolution of the South Carpathians: A new overview. Tectonophysics 2005; 410: 337-365.

Ismail-Zadeh A, Matenco L, Radulian M, Cloetingh S, Panza G. Geodynamics and intermediate-depth seismicity in Vrancea (the south-eastern Carpathians): Current state-of-the art. Tectonophysics 2012; 530-531: 50-79.

Ivanov Z. Apercu general sur l'evolution geologique et structurale du massif des Rhodopes dans le cadre des Balkanides. Bulletin De La Societe Geologique De France 1988; IV: 227-240.

Jipa DC, Olariu C. Sediment routing in a semi-enclosed epicontinental sea: Dacian Basin, Paratethys domain, Late Neogene, Romania. Global and Planetary Change 2013; 103: 193-206.

Jolivet L, Augier R, Robin C, Suc J-P, Rouchy JM. Lithospheric-scale geodynamic context of the Messinian salinity crisis. Sedimentary Geology 2006; 188-189: 9-33.

Jolivet L, Brun J-P. Cenozoic geodynamic evolution of the Aegean. International Journal of Earth Sciences 2010; 99: 109-138.

Juhasz G, Pogacsas G, Magyar I, Vakarcs G. Tectonic versus climatic control on the evolution of fluvio-deltaic systems in a lake basin, Eastern Pannonian Basin. Sedimentary Geology 2007; 202: 72-95.

Karami MP, de Leeuw A, Krijgsman W, Meijer PT, Wortel MJR. The role of gateways in the evolution of temperature and salinity of semi-enclosed basins: An oceanic box model for the Miocene Mediterranean Sea and Paratethys. Global and Planetary Change 2011; 79: 73-88.

Kerey IE, Meric E, Tunoglu C, Kelling G, Brenner RL, Dogan AU. Black Sea-Marmara Sea Quaternary connections: new data from the Bosphorus, Istanbul, Turkey. Palaeogeography, Palaeoclimatology, Palaeoecology 2004; 204: 277-295.

Kounov A, Seward D, Burg J-P, Bernoulli D, Ivanov Z, Handler R. Geochronological and structural constraints on the Cretaceous thermotectonic evolution of the Kraishte zone, western Bulgaria. Tectonics 2010; 29: TC2002.

Kreemer C, Holt WE, Haines AJ. An integrated global model of present-day plate motions and plate boundary deformation. Geophysical Journal International 2003; 154: 8-34.

Krézsek C, Bally AW. The Transylvanian Basin (Romania) and its relation to the Carpathian fold and thrust belt: Insights in gravitational salt tectonics. Marine and Petroleum Geology 2006; 23: 405-442.

Krézsek C, Filipescu S. Middle to late Miocene sequence stratigraphy of the Transylvanian Basin (Romania). Tectonophysics 2005; 410: 437-463.

Krézsek C, Filipescu S, Silye L, Matenco L, Doust H. Miocene facies associations and sedimentary evolution of the Southern Transylvanian Basin (Romania): Implications for hydrocarbon exploration. Marine and Petroleum Geology 2010; 27: 191-214.

Krézsek C, Lăpădat A, Maţenco L, Arnberger K, Barbu V, Olaru R. Strain partitioning at orogenic contacts during rotation, strike–slip and oblique convergence: Paleogene–Early Miocene evolution of the contact between the South Carpathians and Moesia. Global and Planetary Change 2013; 103: 63-81.

Krijgsman W, Stoica M, Vasiliev I, Popov VV. Rise and fall of the Paratethys Sea during the Messinian Salinity Crisis. Earth and Planetary Science Letters 2010; 290: 183-191.

Krzywiec P. Contrasting tectonic and sedimentary history of the central and eastern parts of the Polish Carpathian foredeep basin - results of seismic data interpretation. Mar. Pet. Geol. 2001; 18: 13-38.

Leever KA. Foreland of the Romanian Carpathians, controls on late orogenic sedimentary basin evolution and Paratethys paleogeography. Faculty of Earth and Life Sciences. PhD. VU University Amsterdam, Amsterdam, 2007, pp. 182.

Leever KA, Matenco L, Bertotti G, Cloetingh S, Drijkoningen GG. Late orogenic vertical movements in the Carpathian Bend Zone - seismic constraints on the transition zone from orogen to foredeep. Basin Research 2006; 18: 521-545.

Leever KA, Matenco L, Garcia-Castellanos D, Cloetingh SAPL. The evolution of the Danube gateway between Central and Eastern Paratethys (SE Europe): Insight from numerical modelling of the causes and effects of connectivity between basins and its expression in the sedimentary record. Tectonophysics 2011; 502: 175-195.

Leever KA, Matenco L, Rabagia T, Cloetingh S, Krijgsman W, Stoica M. Messinian sea level fall in the Dacic Basin (Eastern Paratethys): palaeogeographical implications from seismic sequence stratigraphy. Terra Nova 2010; 22: 12-17.

Lenkey L. Geothermics of the Pannonian Basin and its bearing on the tectonics of basin evolution. Vrije Universiteit Amsterdam, 1999, pp. 215.

Lericolais G, Bourget J, Popescu I, Jermannaud P, Mulder T, Jorry S, Panin N. Late Quaternary deep-sea sedimentation in the western Black Sea: New insights from recent coring and seismic data in the deep basin. Global and Planetary Change 2013; 103: 232-247.

Lericolais G, Bulois C, Gillet H, Guichard F. High frequency sea level fluctuations recorded in the Black Sea since the LGM. Global and Planetary Change 2009; 66: 65-75.

Lericolais G, Guichard F, Morigi C, Minereau A, Popescu I, Radan S. A post Younger Dryas Black Sea regression identified from sequence stratigraphy correlated to core analysis and dating. Quaternary International 2010; 225: 199-209.

Letouzey J, Biju-Duval B, Dorkel A, Gonnard R, Krischev K, Montadert L, Sungurlu O. The Black Sea: a marginal basin: geophysical and geological data. In: Biju-Duval B, Montadert L, editors. Structural History of the Mediterranean Basins. Editions Technip, Paris, 1977, pp. 363– 376.

Magyar I, Fogarasi A, Vakarcs G, Buko L, Tari G. The Largest Hydrocarbon Field Discovered to Date in Hungary: Algyo. In: Golonka J, Picha FJ, editors. The Carphathians and their foreland: Geology and hydrocarbon resources. AAPGMemoir 84, 2006, pp. 619-632.

Magyar I, Geary DH, Muller P. Paleogeographic evolution of the Late Miocene Lake Pannon in Central Europe. Palaeogeography, Palaeoclimatology, Palaeoecology 1999a; 147: 151-167.

Magyar I, Geary DH, Suto-Szentai M, Lantos M. Integrated biostratigraphic, magnetostratigraphic and chronostratigraphic correlations of the Late Miocene Lake Panon deposits. Acta Geologica Hungarica 1999b; 42: 5-31.

Magyar I, Radivojević D, Sztanó O, Synak R, Ujszászi K, Pócsik M. Progradation of the paleo-Danube shelf margin across the Pannonian Basin during the Late Miocene and Early Pliocene. Global and Planetary Change 2013; 103: 168-173.

Major CO, Goldstein SL, Ryan WBF, Lericolais G, Piotrowski AM, Hajdas I. The co-evolution of Black Sea level and composition through the last deglaciation and its paleoclimatic significance. Quaternary Science Reviews 2006; 25: 2031-2047.

Martin M, Wenzel F. High-resolution teleseismic body wave tomography beneath SE-Romania - II. Imaging of a slab detachment scenario. Geophysical Journal International 2006; 164: 579-595.

Marunteanu M, Papaianopol L. The connection between the Dacic and Medirranean basins based on calcareous nannoplankton assemblages. Rom. J. Stratigr. 1995; 76: 169-170.

Matenco L, Andriessen P. Quantifying the mass transfer from mountain ranges to deposition in sedimentary basins: Source to sink studies in the Danube Basin–Black Sea system. Global and Planetary Change 2013; 103: 1-18.

Matenco L, Bertotti G, Cloetingh S, Dinu C. Subsidence analysis and tectonic evolution of the external Carpathian-Moesian Platform region during Neogene times. Sedimentary Geology 2003; 156: 71-94.

Matenco L, Bertotti G, Leever K, Cloetingh S, Schmid S, Tărăpoancă M, Dinu C. Large-scale deformation in a locked collisional boundary: Interplay between subsidence and uplift, intraplate stress, and inherited lithospheric structure in the late stage of the SE Carpathians evolution. Tectonics 2007; 26: TC4011, doi: 10.1029/2006TC001951.

Matenco L, Krézsek C, Merten S, Schmid S, Cloetingh S, Andriessen P. Characteristics of collisional orogens with low topographic build-up: an example from the Carpathians. Terra Nova 2010; 22: 155-165.

Matenco L, Radivojević D. On the formation and evolution of the Pannonian Basin: Constraints derived from the structure of the junction area between the Carpathians and Dinarides. Tectonics 2012; 31: TC6007.

Matenco L, Schmid S. Exhumation of the Danubian nappes system (South Carpathians) during the early Tertiary: inferences from kinematic and paleostress analysis at the Getic/Danubian nappes contact. Tectonophysics 1999; 314: 401-422.

Mathisen ME, Vondra CF. The Fluvial and Pyroclastic Deposits of the Cagayan Basin, Northern Luzon, Philippines - an Example of Non-Marine Volcaniclastic Sedimentation in an Interarc Basin. Sedimentology 1983; 30: 369-392.

Merten S. Thermo-tectonic evolution of a convergent orogen with low-topographic build-up: Exhumation and kinematic patterns in the Romanian Carpathians derived from thermochronology. PhD. VU University Amsterdam, Faculty of Earth and Life Sciences, Amsterdam, 2011, pp. 202.

Merten S, Matenco L, Foeken JPT, Stuart FM, Andriessen PAM. From nappe stacking to out-of-sequence postcollisional deformations: Cretaceous to Quaternary exhumation history of the SE Carpathians assessed by low-temperature thermochronology. Tectonics 2010; 29: TC3013.

Morley CK. Models for relative motion of crustal blocks within the Carpathian region, based on restorations of the outer Carpathian thrust sheets. Tectonics 1996; 15/4: 885-904.

Morley CK, Westaway R. Subsidence in the super-deep Pattani and Malay basins of Southeast Asia: a coupled model incorporating lower-crustal flow in response to post-rift sediment loading. Basin Research 2006; 18: 51-84.

Munteanu I, Matenco L, Dinu C, Cloetingh S. Kinematics of back-arc inversion of the Western Black Sea Basin. Tectonics 2011; 30: TC5004.

Munteanu I, Matenco L, Dinu C, Cloetingh S. Effects of large sea-level variations in connected basins: the Dacian–Black Sea system of the Eastern Paratethys. Basin Research 2012; 24: 583-597.

Munteanu I, Willingshofer E, Sokoutis D, Matenco L, Dinu C, Cloetingh S. Transfer of deformation in back-arc basins with a laterally variable rheology: Constraints from analogue modelling of the Balkanides–Western Black Sea inversion. Tectonophysics 2013; 602: 223-236.

Necea D. High-resolution morpho-tectonic profiling across an orogen: tectonic-controled geomorphology and multiple dating approach in the SE Carpathians. Faculty of earth and Life Sciences. PhD. VU University Amsterdam, Amsterdam, 2010, pp. 147.

Nikishin AM, Korotaev MV, Ershov AV, Brunet M-F. The Black Sea basin: tectonic history and Neogene-Quaternary rapid subsidence modelling. Sedimentary Geology 2003; 156: 149-168.

Nikishin AM, Okay AI, Tüysüz O, Demirer A, Amelin N, Petrov E. The Black Sea basins structure and history: New model based on new deep penetration regional seismic data. Part 1: Basins structure and fill. Marine and Petroleum Geology 2015; 59: 638-655.

Oaie G, Melinte-Dobrinescu MC. Holocene litho- and biostratigraphy of the NW Black Sea (Romanian shelf). Quaternary International 2012; 261: 146-155.

Okay AI, Celal Sengor AM, Gorur N. Kinematic history of the opening of the Black Sea and its effect on the surrounding regions. Geology 1994; 22: 267-270.

Pinter N, Grenerczy G, Weber J, Stein S, Medak D. The Adria Microplate: GPS Geodesy, Tectonics and Hazards (Nato Science Series: IV: Earth and Environmental Sciences): Springer, 2005.

Popescu I, De Batist M, Lericolais G, Nouze H, Poort J, Panin N, Versteeg W, Gillet H. Multiple bottom-simulating reflections in the Black Sea: Potential proxies of past climate conditions. Marine Geology 2006; 227: 163-176.

Popescu I, Lericolais G, Panin N, Normand A, Dinu C, Le Drezen E. The Danube submarine canyon (Black Sea): morphology and sedimentary processes. Marine Geology 2004; 206: 249-265.

Popescu I, Lericolais G, Panin N, Wong HK, Droz L. Late Quaternary channel avulsions on the Danube deep-sea fan, Black Sea. Marine Geology 2001; 179: 25-37.

Popov SV, Shcherba IG, Ilyina LB, Nevesskaya LA, Paramonova NP, Khondkarian SO, Magyar I. Late Miocene to Pliocene palaeogeography of the Paratethys and its relation to the Mediterranean. Palaeogeography, Palaeoclimatology, Palaeoecology 2006; 238: 91-106.

Posamentier HW. Depositional elements associated with a basin floor channel-levee system: case study from the Gulf of Mexico. Marine and Petroleum Geology 2003; 20: 677-690.

Pubellier M, Morley CK. The basins of Sundaland (SE Asia): Evolution and boundary conditions. Marine and Petroleum Geology 2014; 58, Part B: 555-578.

Răbăgia T, Matenco L. Tertiary tectonic and sedimentological evolution of the South Carpathians foredeep: tectonic versus eustatic control. Marine and Petroleum Geology 1999; 16/7: 719-740.

Rabagia T, Matenco L, Cloetingh S. The interplay between eustacy, tectonics and surface processes during the growth of a fault-related structure as derived from sequence stratigraphy: The Govora-Ocnele Mari antiform, South Carpathians. Tectonophysics 2011; 502: 196-220.

Radoane M, Radoane N, Dumitriu D. Geomorphological evolution of longitudinal river profiles in the Carpathians. Geomorphology 2003; 50: 293-306.

Ratschbacher L, Frisch W, Linzer HG, Merle O. Lateral extrusion in the Eastern Alps; Part 2, Structural analysis. Tectonics 1991; 10: 257-271.

Rögl F. Mediteranean and Paratethys. Facts and hypotheses of an Oligocene to Miocene paleogeography. Geologica Carpathica 1999; 50: 339-349.

Roure F, Roca E, Sassi W. The Neogene evolution of the outer Carpathian flysch units (Poland, Ukraine and Romania); kinematics of a foreland/ fold-and-thrust belt system. Sedimentary Geology 1993; 86: 177-201.

Royden LH. Evolution of retreating subduction boundaries formed during continental collision. Tectonics 1993; 12: 629-638.

Ryan WBF, Major CO, Lericolais G, Goldstein SL. Catastrophic flooding of the Black Sea. Annu. Rev. Earth Planet. Sci. 2003; 31: 525-554.

Sanders C. Tectonics and erosion, competitive forces in a compressive orogen: a fission track study of the Romanian Carpathians. Faculty of Earth Sciences. Vrije Universiteit, Amsterdam, 1998, pp. 204.

Sanders C, Huismans R, van Wees JD, Andriessen P. The Neogene history of the Transylvanian basin in relation to its surrounding mountains. In: Cloetingh S, Horvath F, Bada G, Lankreijer A, editors. Neotectonics and surface processes: the Pannonian Basin and Alpine/Carpathian System. 3, 2002, pp. 121-133.

Sandulescu M. Essai de synthese structurale des Carpathes. Bull. Soc. Geol. France 1975; XVII: 299.

Săndulescu M. Geotectonica României (translated title: Geotectonics of Romania). Bucharest: Ed. Tehnică, 1984.

Schmid S, Bernoulli D, Fügenschuh B, Matenco L, Schefer S, Schuster R, Tischler M, Ustaszewski K. The Alpine-Carpathian-Dinaridic orogenic system: correlation and evolution of tectonic units. Swiss Journal of Geosciences 2008; 101: 139-183.

Schmid SM, Berza T, Diaconescu V, Froitzheim N, Fuegenschuh B. Orogen-parallel extension in the South Carpathians during the Paleogene. Tectonophysics 1998; 297: 209-228.

Schmid SM, Fugenschuh B, Kissling E, Schuster R. Tectonic map and overall architecture of the Alpine orogen. Eclogae Geologicae Hervetiae 2004; 97: 92-117.

Seghedi I, Matenco L, Downes H, Mason PRD, Szakács A, Pécskay Z. Tectonic significance of changes in post-subduction Pliocene-Quaternary magmatism in the south east part of the Carpathian-Pannonian Region. Tectonophysics 2011; 502: 146-157.

Shillington DJ, White N, Minshull TA, Edwards GRH, Jones SM, Edwards RA, Scott CL. Cenozoic evolution of the eastern Black Sea: A test of depth-dependent stretching models. Earth and Planetary Science Letters 2008; 265: 360-378.

Stampfli GM, Borel GD. A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrons. Earth and Planetary Science Letters 2002; 196: 17-33.

Stanica A, Panin N. Present evolution and future predictions for the deltaic coastal zone between the Sulina and Sf. Gheorghe Danube river mouths (Romania). Geomorphology 2009; 107: 41-46.

Starostenko V, Buryanov V, Makarenko I, Rusakov O, Stephenson R, Nikishin A, Georgiev G, Gerasimov M, Dimitriu R, Legostaeva O, Pchelarov V, Sava C. Topography of the crust-mantle boundary beneath the Black Sea Basin. Tectonophysics 2004; 381: 211-233.

Stefanescu M. Stratigraphy and structure of Cretaceous and Palaeogene flysch deposits between Prahova and Ialomita valleys. Romanian Journal of Tectonics and Regional Geology 1995; 76: 1-49.

Steininger FF, Wessely G. From the Tethyan Ocean to the Paratethys Sea: Oligocene to Neogene Stratigraphy, Paleogeography and Paleobiogeography of the circum-Mediterranean region and the Oligocene to Neogene Basin evolution in Austria. Mitt. Österr.Geol. Ges. 1999; 92: 95-116.

Stoica M, Lazăr I, Krijgsman W, Vasiliev I, Jipa D, Floroiu A. Paleoenvironmental evolution of the East Carpathian foredeep during the late Miocene–early Pliocene (Dacian Basin; Romania). Global and Planetary Change 2013; 103: 135-148.

Stojadinovic U, Matenco L, Andriessen PAM, Toljić M, Foeken JPT. The balance between orogenic building and subsequent extension during the Tertiary evolution of the NE Dinarides: Constraints from low-temperature thermochronology. Global and Planetary Change 2013; 103: 19-38.

Strechie C, Andre F, Jelinowska A, Tucholka P, Guichard F, Lericolais G, Panin N. Magnetic minerals as indicators of major environmental change in Holocene Black Sea sediments: preliminary results. Physics and Chemistry of the Earth, Parts A/B/C 2002; 27: 1363-1370.

Syvitski JPM, Kettner AJ, Overeem I, Hutton EWH, Hannon MT, Brakenridge GR, Day J, Vorosmarty C, Saito Y, Giosan L, Nicholls RJ. Sinking deltas due to human activities. Nature Geoscience 2009; 2: 681-686.

Szakacs A, Krezsek C. Volcano-basement interaction in the Eastern Carpathians: Explaining unusual tectonic features in the Eastern Transylvanian Basin, Romania. Journal of Volcanology and Geothermal Research 2006; 158: 6-20.

Tărăpoancă M, Bertotti G, Matenco L, Dinu C, Cloetingh S. Architecture of the Focsani depression: A 13 km deep basin in the Carpathians bend zone (Romania). Tectonics 2003; 22/6: 1074, doi:10.1029/2002TC001486.

Tărăpoancă M, Garcia-Castellanos D, Bertotti G, Matenco L, Cloetingh SAPL, Dinu C. Role of the 3-D distributions of load and lithospheric strength in orogenic arcs: polystage subsidence in the Carpathians foredeep. Earth and Planetary Science Letters 2004; 221: 163-180.

Tărăpoancă M, Ţambrea D, Avram V, Popescu BM. The geometry of the South carpathians sole thrust and the Moesia boundary: the role of inherited structures in establishing a transcurrent contact on the concave side of the Carpathians. In: Lacombe O, Lave J, Roure F, Verges J, editors. Thrust Belts and Foreland Basins: From Fold Kinematics to Hydrocarbon Systems. Springer, 2007, pp. 492-508.

Tari G, Davies J, Dellmour R, Larratt E, Novotny B, Kozhuharov E. Play types and hydrocarbon potential of the deepwater Black Sea, NE Bulgaria. The Leading Edge 2009; 28: 1076-1081.

Tari G, Dovenyi P, Dunkl I, Horvath F, Lenkey L, Stefanescu M, Szafian P, Toth T. Lithospheric structure of the Pannonian basin derived from seismic, gravity and geothermal data. In: Durand B, Jolivet L, Horvath F, Serrane M, editors. The Mediterranean Basins: extension within the Alpine Orogen. Geol Soc London Spec Publ 156, 1999, pp. 215-250.

Tari G, Horváth F, Rumpler J. Styles of extension in the Pannonian basin. Tectonophysics 1992; 208: 203-219.

ter Borgh M. Connections between sedimentary basins during continental collision: how tectonic, surface and sedimentary processes shaped the Paratethys. Institute of Earth Sciences. PhD. Utrecht University, PhD Thesis, Utrecht, 2013, pp. 212.

ter Borgh M, Radivojević D, Matenco L. Constraining forcing factors and relative sea-level fluctuations in semi-enclosed basins: the Late Neogene demise of Lake Pannon. Basin Research 2015: n/a-n/a.

ter Borgh M, Stoica M, Donselaar ME, Matenco L, Krijgsman W. Miocene connectivity between the Central and Eastern Paratethys: Constraints from the western Dacian Basin. Palaeogeography, Palaeoclimatology, Palaeoecology 2014; 412: 45-67.

ter Borgh M, Vasiliev I, Stoica M, Knežević S, Matenco L, Krijgsman W, Rundić L, Cloetingh S. The isolation of the Pannonian basin (Central Paratethys): New constraints from magnetostratigraphy and biostratigraphy. Global and Planetary Change 2013; 103: 99-118.

Tiliţă M, Matenco L, Dinu C, Ionescu L, Cloetingh S. Understanding the kinematic evolution and genesis of a back-arc continental “sag” basin: The Neogene evolution of the Transylvanian Basin. Tectonophysics 2013; 602: 237-258.

Tischler M, Matenco L, Filipescu S, Groger HR, Wetzel A, Fugenschuh B. Tectonics and sedimentation during convergence of the ALCAPA and Tisza-Dacia continental blocks: the Pienide nappe emplacement and its foredeep (N. Romania). Geological Society, London, Special Publications 2008; 298: 317-334.

Toljić M, Matenco L, Ducea MN, Stojadinović U, Milivojević J, Đerić N. The evolution of a key segment in the Europe–Adria collision: The Fruška Gora of northern Serbia. Global and Planetary Change 2013; 103: 39-62.

Tomljenović B, Csontos L. Neogene-Quaternary structures in the border zone between Alps, Dinarides and Pannonian Basin (Hrvatsko zagorje and Karlovac Basins, Croatia). International Journal of Earth Sciences 2001; V90: 560-578.

Ustaszewski K, Kounov A, Schmid SM, Schaltegger U, Krenn E, Frank W, Fügenschuh B. Evolution of the Adria-Europe plate boundary in the northern Dinarides: From continent-continent collision to back-arc extension. Tectonics 2010; 29: TC6017, doi: 10.1029/2010tc002668.

Ustaszewski K, Schmid S, Fügenschuh B, Tischler M, Kissling E, Spakman W. A map-view restoration of the Alpine-Carpathian-Dinaridic system for the Early Miocene. Swiss Journal of Geosciences 2008; 101: 273-294.

Uyeda S, Kanamori H. Back-arc opening and the mode of subduction. Journal of Geophysical Research 1979; 84: 1049-1061.

van der Meer DG, Spakman W, van Hinsbergen DJJ, Amaru ML, Torsvik TH. Towards absolute plate motions constrained by lower-mantle slab remnants. Nature Geosciences 2010; 3: 36-40.

van Hinsbergen DJJ, Hafkenscheid E, Spakman W, Meulenkamp JE, Wortel R. Nappe stacking resulting from subduction of oceanic and continental lithosphere below Greece. Geology 2005; 33: 325-328.

van Hinsbergen DJJ, Schmid SM. Map view restoration of Aegean - West Anatolian accretion and extension since the Eocene. Tectonics 2012; 31: TC5005.

Vergés J, Fernàndez M. Tethys–Atlantic interaction along the Iberia–Africa plate boundary: The Betic–Rif orogenic system. Tectonophysics 2012; 579: 144-172.

von Quadt A, Moritz R, Peytcheva I, Heinrich CA. 3: Geochronology and geodynamics of Late Cretaceous magmatism and Cu-Au mineralization in the Panagyurishte region of the Apuseni-Banat-Timok-Srednogorie belt, Bulgaria. Ore Geology Reviews 2005; 27: 95-126.

Vörös A. Provinciality of the Mediterranean Lower Jurassic brachiopod fauna: causes and plate-tectonic implications. Paleogeography Paleoclimatology Paleoecology 1977; 21: 1–16.

Windhoffer G, Bada G. Formation and deformation of the Derecske Trough, Pannonian Basin: Insights from analogue modeling. Acta Geologica Hungarica 2005; 48/4: 351-369.

Wolf RA, Farley KA, Silver LT. Helium diffusion and low-temperature thermochronometry of apatite. Geochimica et Cosmochimica Acta 1996; 60: 4231-4240.

Wortel MJR, Spakman W. Subduction and Slab Detachment in the Mediteranean-Carpathian Region. Science 2000; 290: 1910-1917.

Zimmerman A, Stein H, Hannah J, Koželj D, Bogdanov K, Berza T. Tectonic configuration of the Apuseni–Banat—Timok–Srednogorie belt, Balkans-South Carpathians, constrained by high precision Re –Os molybdenite ages. Mineralium Deposita 2008; 43: 1-21.

Figure captions

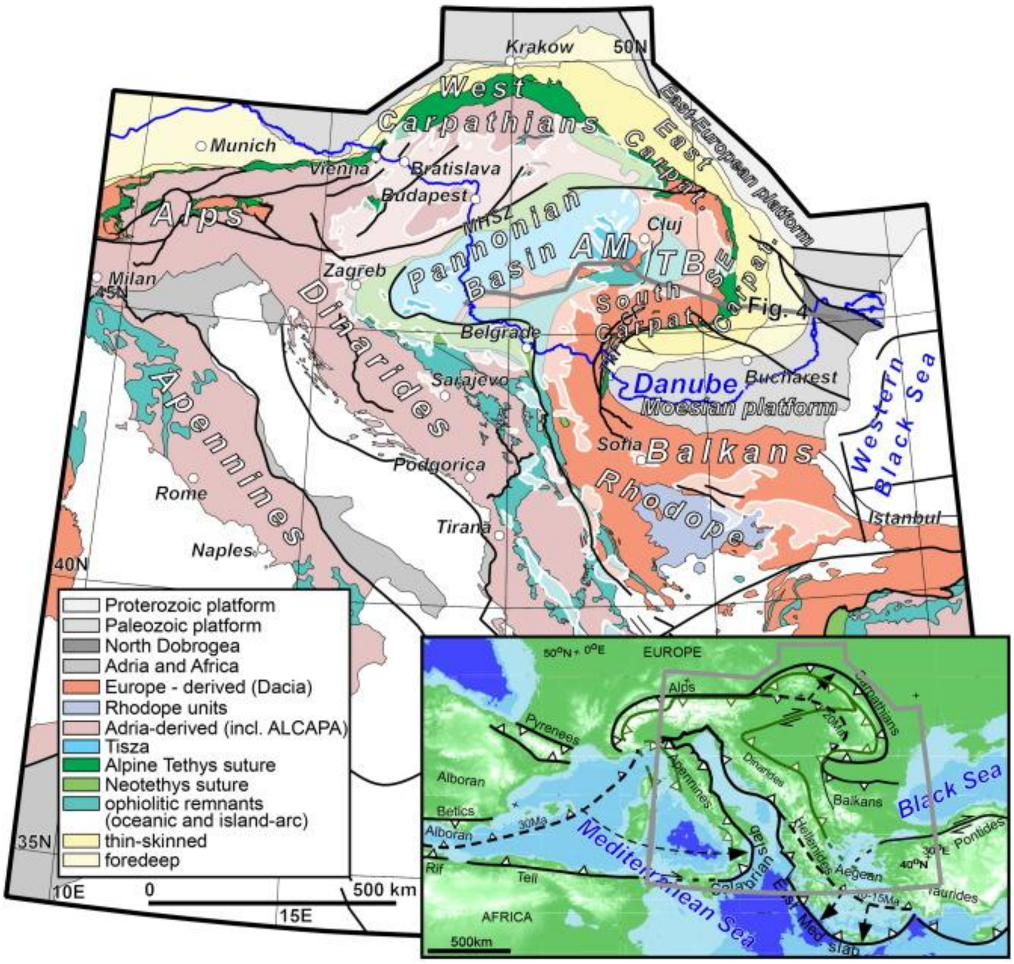

Fig. 1 – a) Tectonic map of the Alps – Carpathians – Dinaridic – Hellenidic system (simplified from Schmid et al. 2011) with the extent of the Pannonian, Transylvanian (white transparent background) and Dacian basins, as well as the Black Sea. The grey line is the location of Fig. 4. AM - Apuseni Mountains; TB - Transylvanian Basin. The lower inset is the location of the map in the system of European Mesozoic-Cenozoic orogens. Dashed black line is the position of the orogenic front prior to the onset of extension associated with the roll-back of the Calabrian, Aegean and Carpathian slabs.

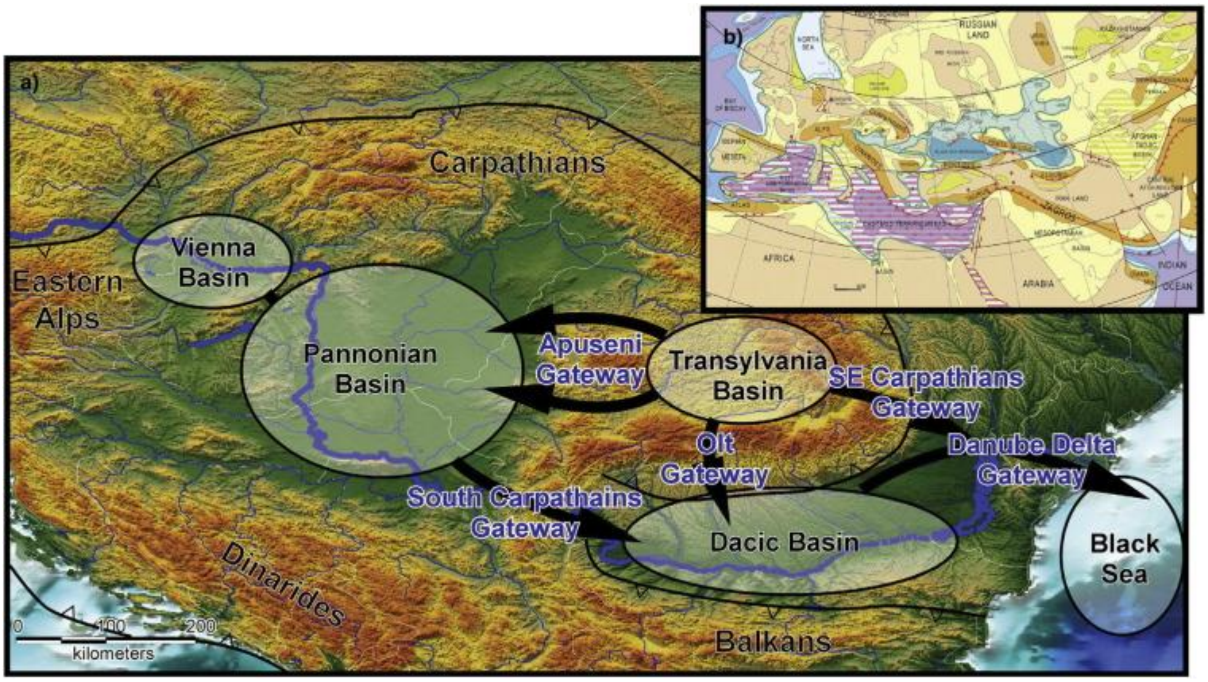

Fig. 2 - The concept of connectivity and gateways applied to the Miocene–Quaternary sedimentary basins in the Carpathians–Pannonian system together with the reconstruction of the (dis)connected Paratethys basins at the time of the Messinian Salinity Crisis (after Matenco and Andriessen, 2013).

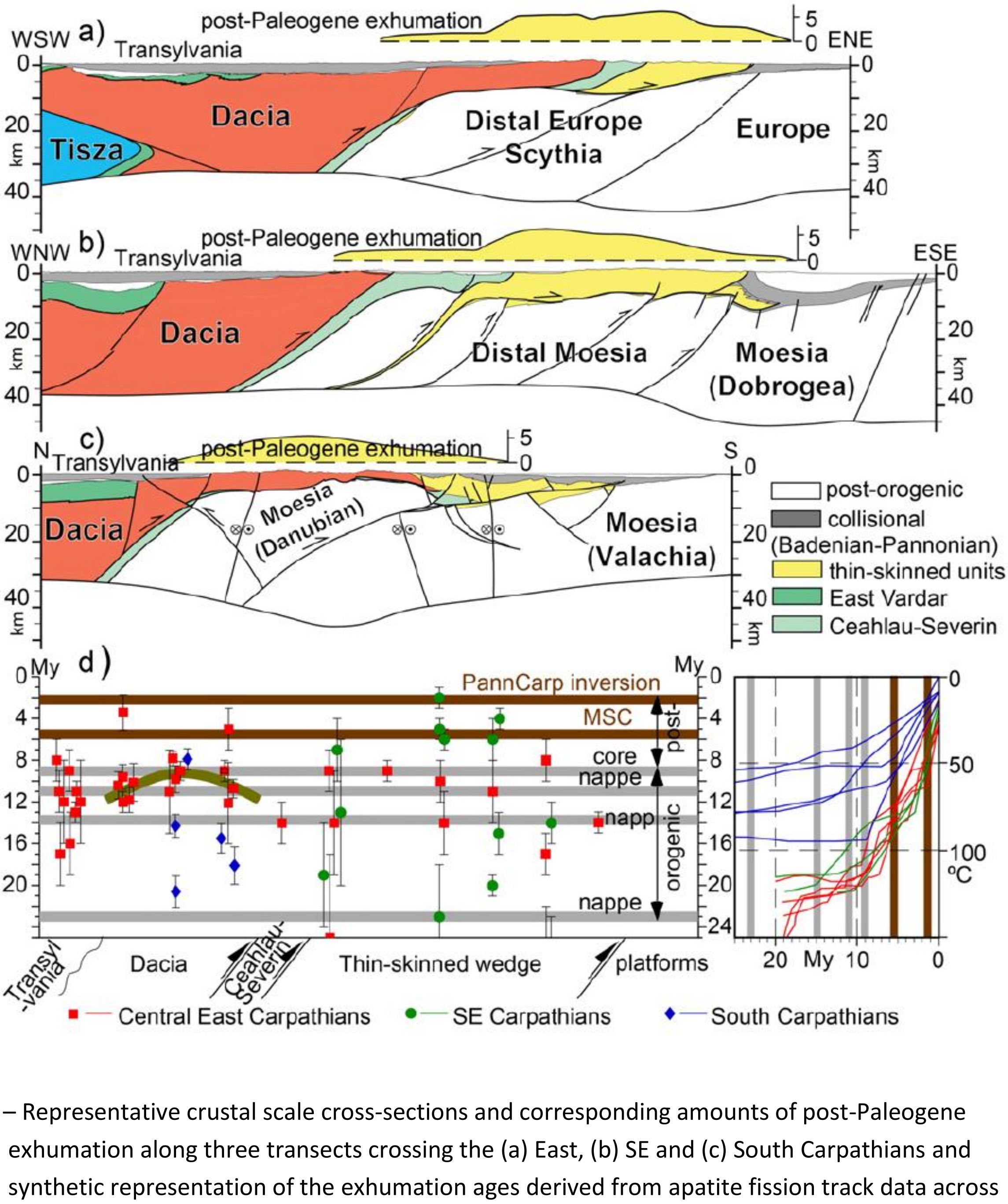

Fig. 3 – Representative crustal scale cross-sections and corresponding amounts of post-Paleogene exhumation along three transects crossing the (a) East, (b) SE and (c) South Carpathians and synthetic representation of the exhumation ages derived from apatite fission track data across the three areas an their thermal modelling (d). Note that only the Miocene ages are plotted (after Matenco et al. 2010).

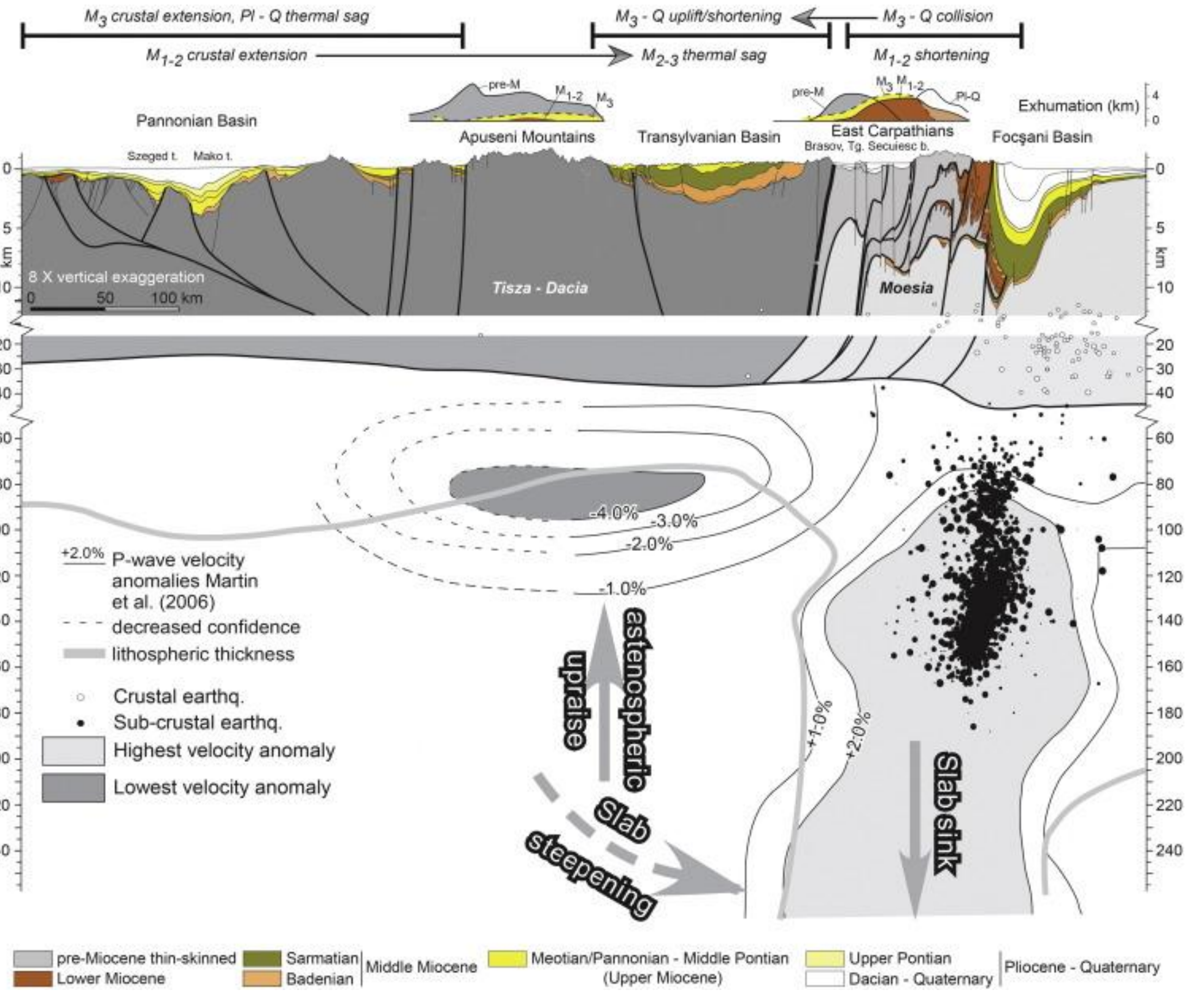

Fig. 4 - Simplified geological cross section across the SE part of the Pannonian Basin, Apuseni Mountains, Transylvanian Basin and SE Carpathians and amounts of exhumation over the Apuseni Mountains and SE Carpathians derived from low-temperature thermochronology (modified from Matenco and Radivojević 2012). The geological cross section displays only Miocene-Quaternary sediments geometries and faults patterns. All pre-Miocene structures were ignored. The location of the cross section is displayed in Fig. 1. pre-M = pre-Miocene; M_1 = Early Miocene; M_2 = Middle Miocene; M_3 = Late Miocene; Pl = Pliocene; Q = Quaternary. The lower part of the figure is the crustal and upper mantle structure beneath the western Pannonian Basin – Carpathian Mountains with underlying the seismicity and the anlomanies detected by high resolution, local teleseismic tomography. Note the dynamic topography associated both with the Vrancea slab and with the post-Miocene uplift of the Transylvanian Basin associated with the astenospheric upraise.

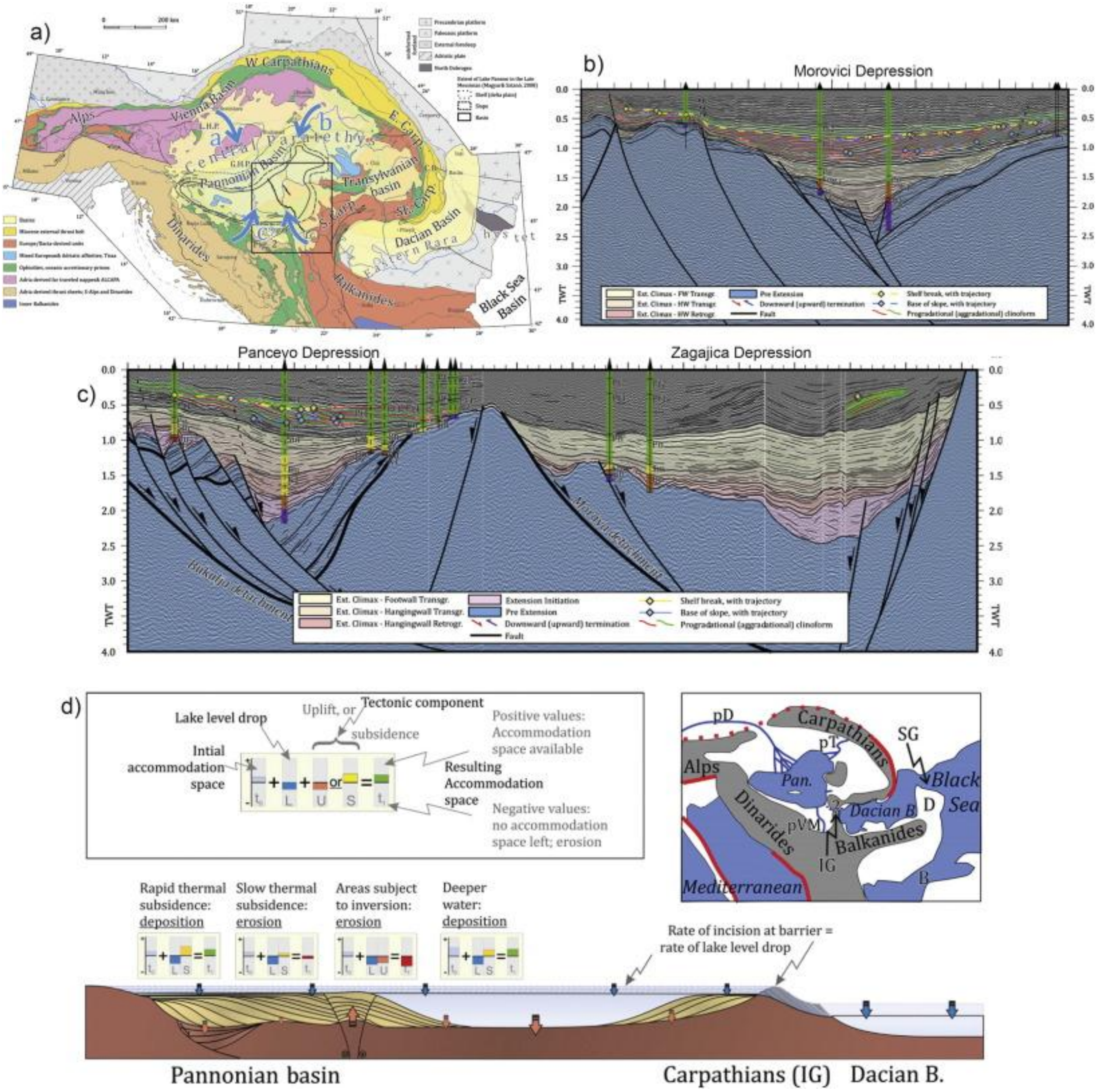

Fig. 5 – The evolution of the Miocene – Pliocene basin fill in the SE part of the Pannonian Basin, near the entrance in the South Carpathians gateway (after ter Borgh et al., 2015). a) Map of the Pannonian Carpathians area with the extent of the various basins and location of the studied transects; b) The late Miocene progradational sequence in the Pannonian Basin, near northern part of the Dinarides; c) The late Miocene progradational sequence in the Pannonian Basin, in the area connecting the Dinarides with the South Carpathians; d) The effects of forcing factors on the infill of the Pannonian Basin, derived from the high-resolution analysis of seismic data. Inset: palaeogeographic map of the region during the Early Messinian. For further details see ter Borgh et al. (2015) and Matenco and Radivojevic (2012).

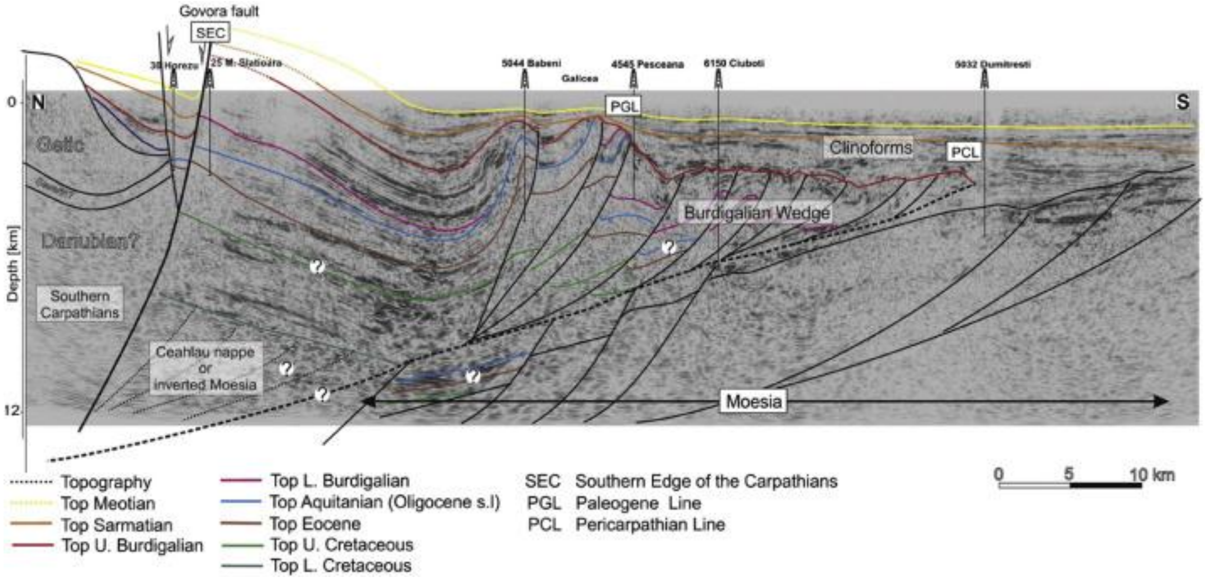

Fig. 6 - Regional transect over the central part of the South Carpathians, Getic Depression and Moesian foreland obtained by the interpretation of a wide-offset regional line (after Krezsek et al., 2013).

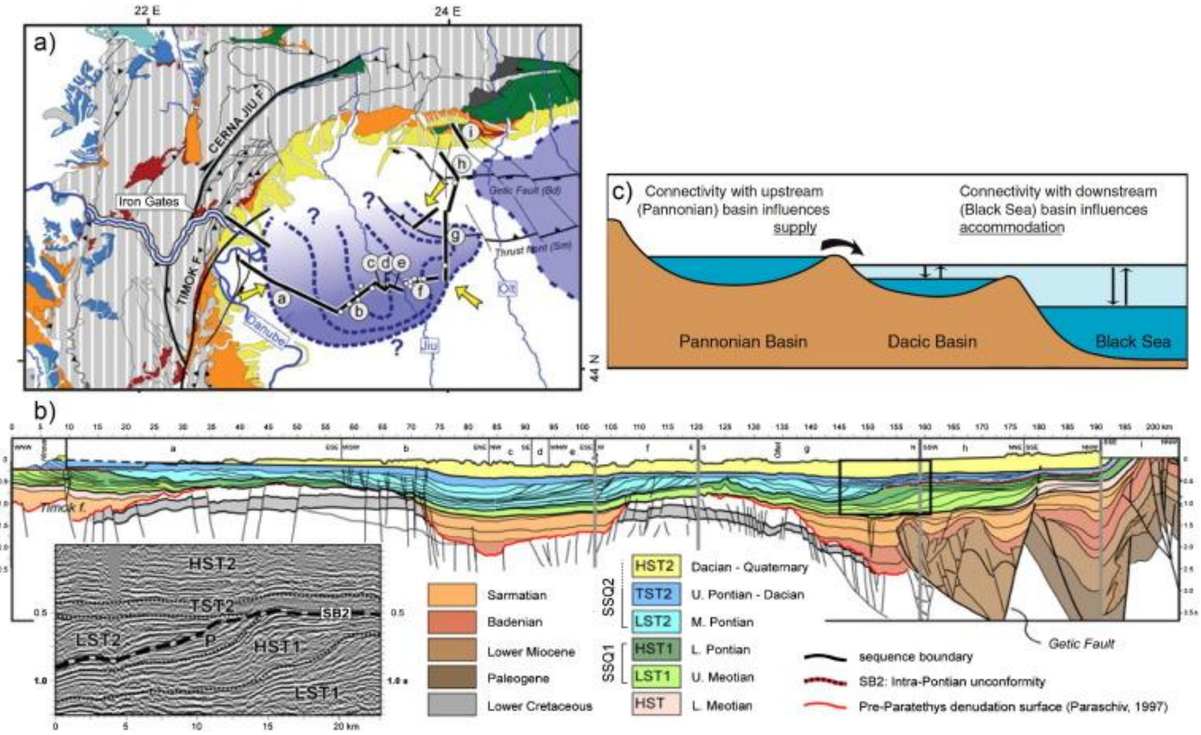

Fig. 7 – The sequence stratigraphic evolution of the Dacian Basin controlled by adjacent gateways (after Leever et al., 2010). a) Location maps with the position of the interpreted seismic transect in the western part of the Dacic Basin. Dashed lines within the remnant lake indicate the decreasing deep basin area due to the progressive infilling of the basin during the Middle Pontian lowstand (LST2); b) Sequence stratigraphic interpretation of a seismic section extending from the western to the northern margin of the western Dacic Basin; c) Cartoon showing the influence of connectivity on accommodation and sediment supply. Vertical arrows indicate base level fluctuations.

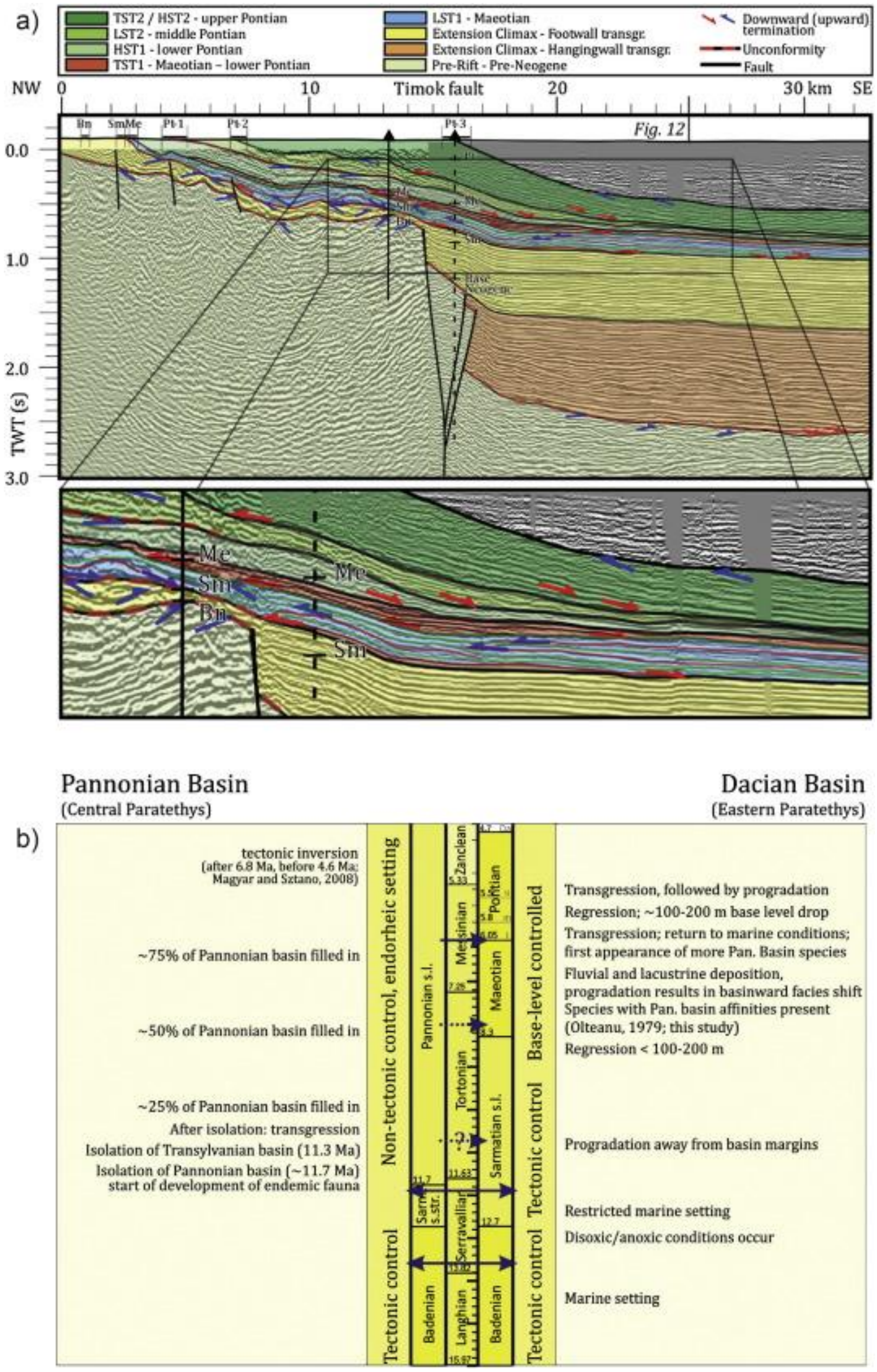

Fig. 8 – The Middle Miocene to Quaternary evolution of the westernmost part of the Dacian Basin, near the South Carpathians gateway (after ter Borgh et al., 2014). a) Interpretation of a seismic section in the westernmost part of the Dacian Basin. A significant vertical offset is evident across the Palaeogene–Sarmatian Timok strike–slip fault. A significant part of this offset of the fault predates the Middle Miocene; Bn — Badenian, Sm — Sarmatian, Me — Maeotian, Pt-1/2/3 — lower/middle/upper Pontian; b) Overview of major events affecting the Central and Eastern Paratethys during the Miocene and Pliocene, with specific regard on the connection between the Dacian and Pannonian basins.

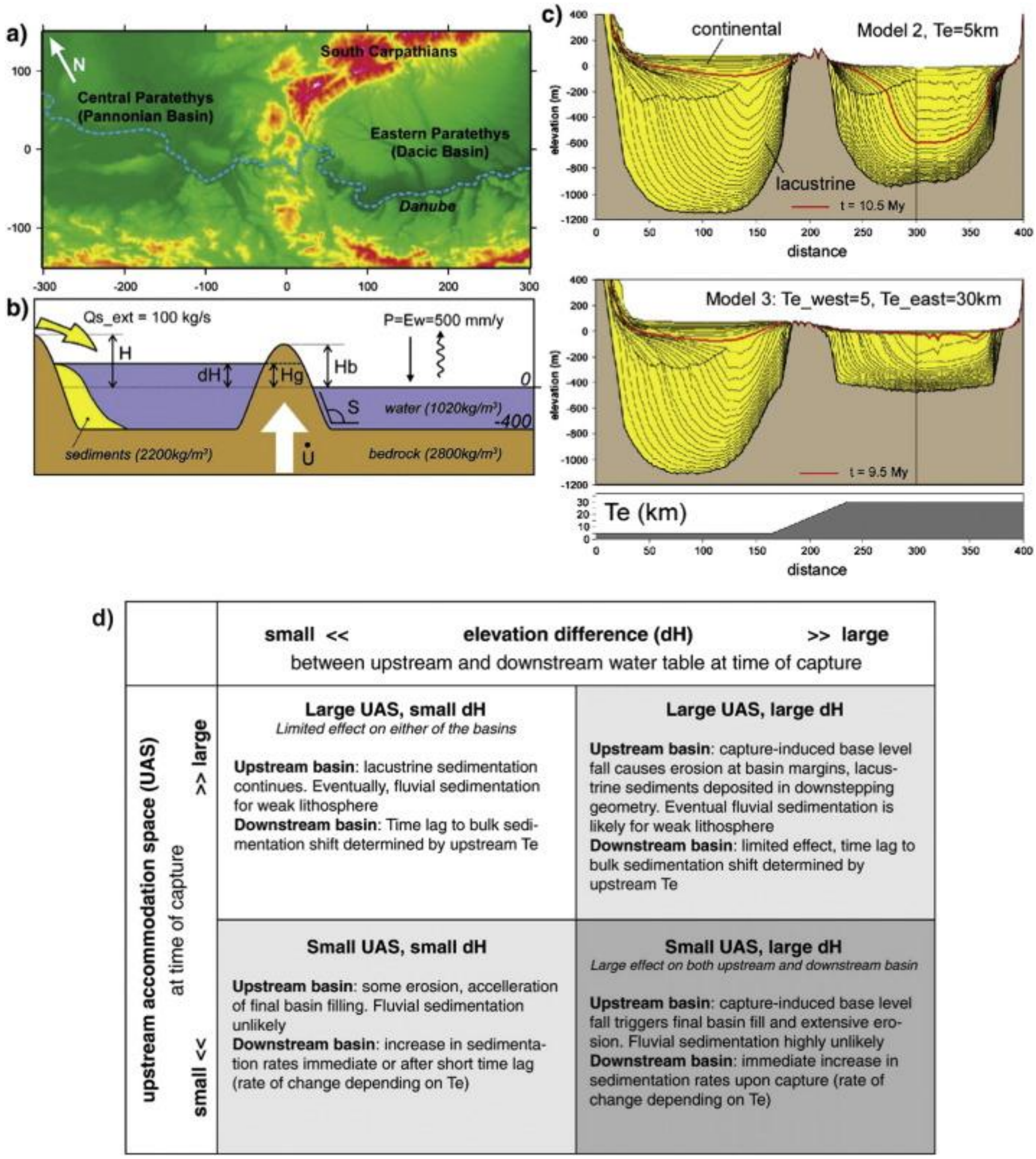

Fig. 9 - Numerical modelling of the connection between the Pannonian basin (Central Paratethys) and the Dacic basin (Eastern Paratethys) along the trace of the gradual basin fill of the Danube basins during late Miocene–Pliocene times (after Leever et al., 2011). a) 2 km resolution DEM of the studied area. b) Model setup and parameters in cross section. c) Effect of a barrier slope in the studied scenario of basins connectivity. The thick line is the corresponding time line preceding the capture of the upstream basin. The other highlighted (long dashed, dotted) time lines represent the end of base level fall and the final stage of lacustrine sedimentation in the upstream basin respectively; d) Factors controlling sedimentary response to lake capture.

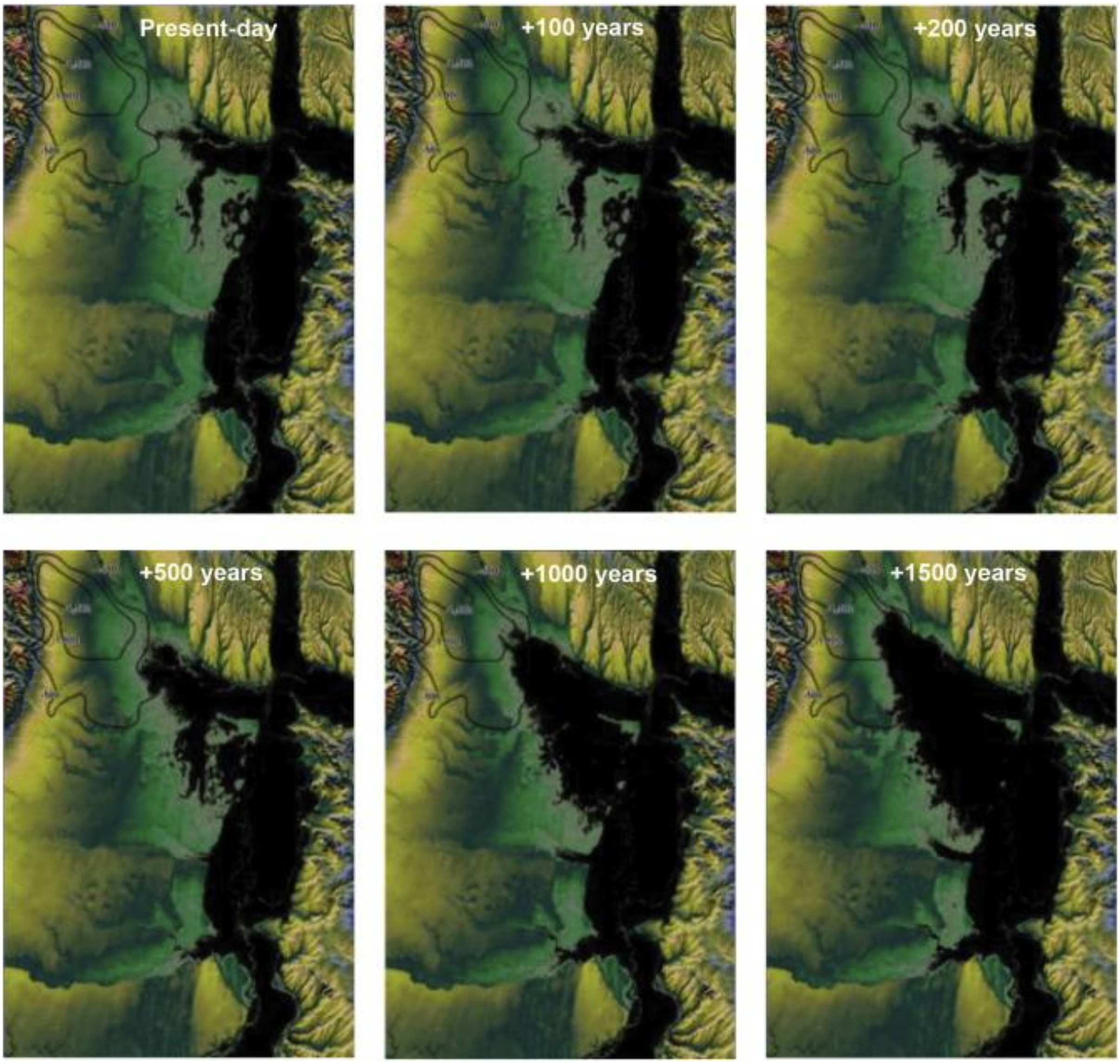

Fig. 10 – Model prediction of the evolution of the areas threatened by flooding in the SE Carpathians by correlating the vertical movements derived from GPS studies in the Carpathians foreland (van der Hoeven et al. 2005) with active faulting patterns (see Matenco et al. 2007 for further details).

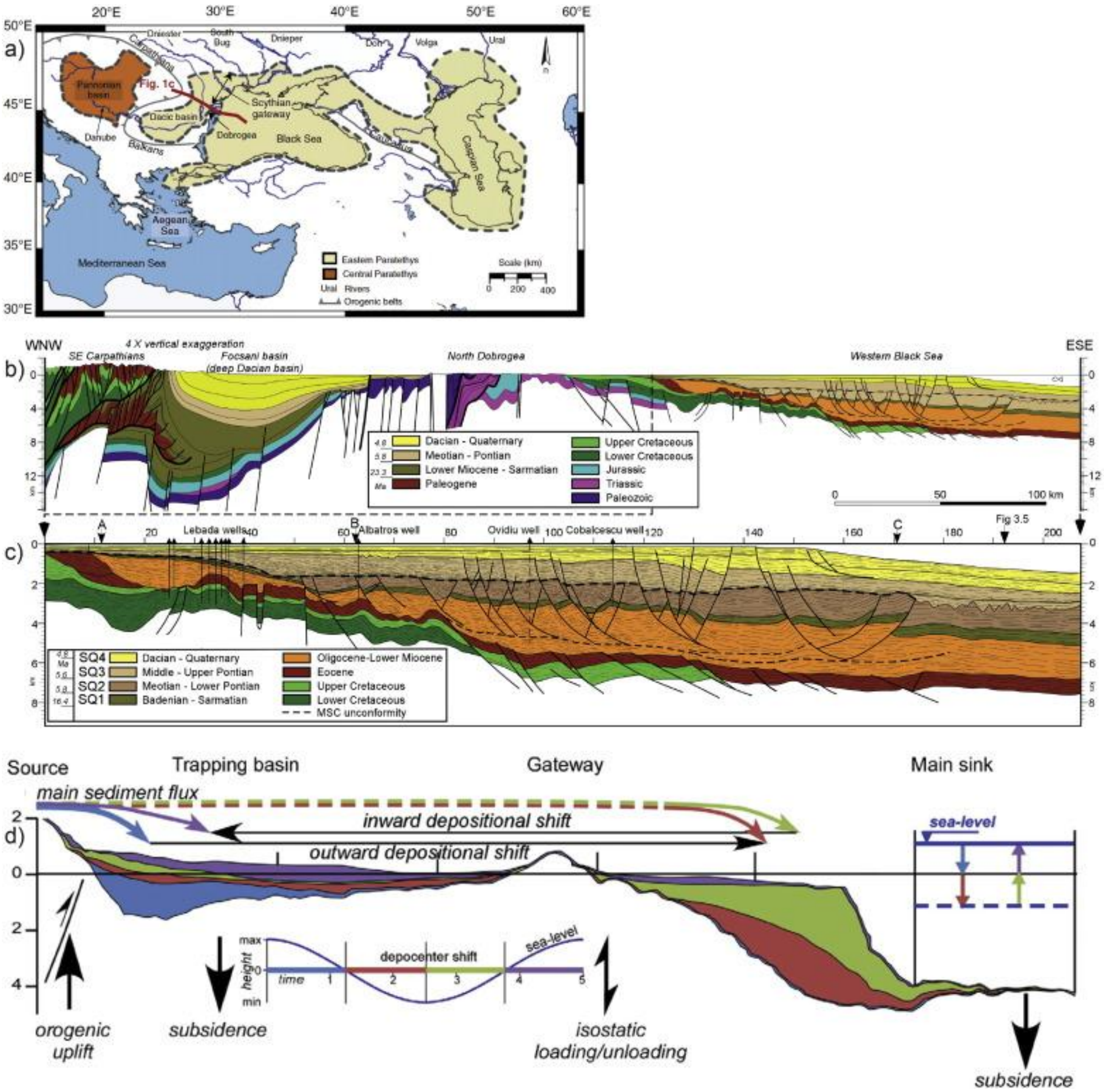

Fig. 11 – Application of the concept of connectivities and gateways on the link between the Dacian Basin and the Black Sea (after Munteanu et al., 2012; Bartol et al., 2012). a) Map of the Central and Eastern Paratethys during the Late Miocene; b) A regional cross-section (4X vertical exaggeration) spanning from SE Carpathians, Dacian Basin and Dobrogea highland to the deep-sea part of Black Sea. c) Detailed geometry of the western Black Sea part of the cross-section depicted in Fig. 2a, derived from the interpretation of regional seismic lines; d) Conceptual model of depositional shifts between two sedimentary basins separated by a submarine topographic barrier, specific for the Eastern Paratethys situation during the Messinian Salinity Crisis. The graph below illustrates the modelling of sea level change as a cosine function approximating a natural evolution of sea level, the cycle time and magnitudes being normalized. The decrease and subsequent increase in accommodation space in the trapping basin triggers outward depositional shifts (from the trapping basin to the main sink) or inward depositional shifts (from the main sink to the trapping basin), respectively.

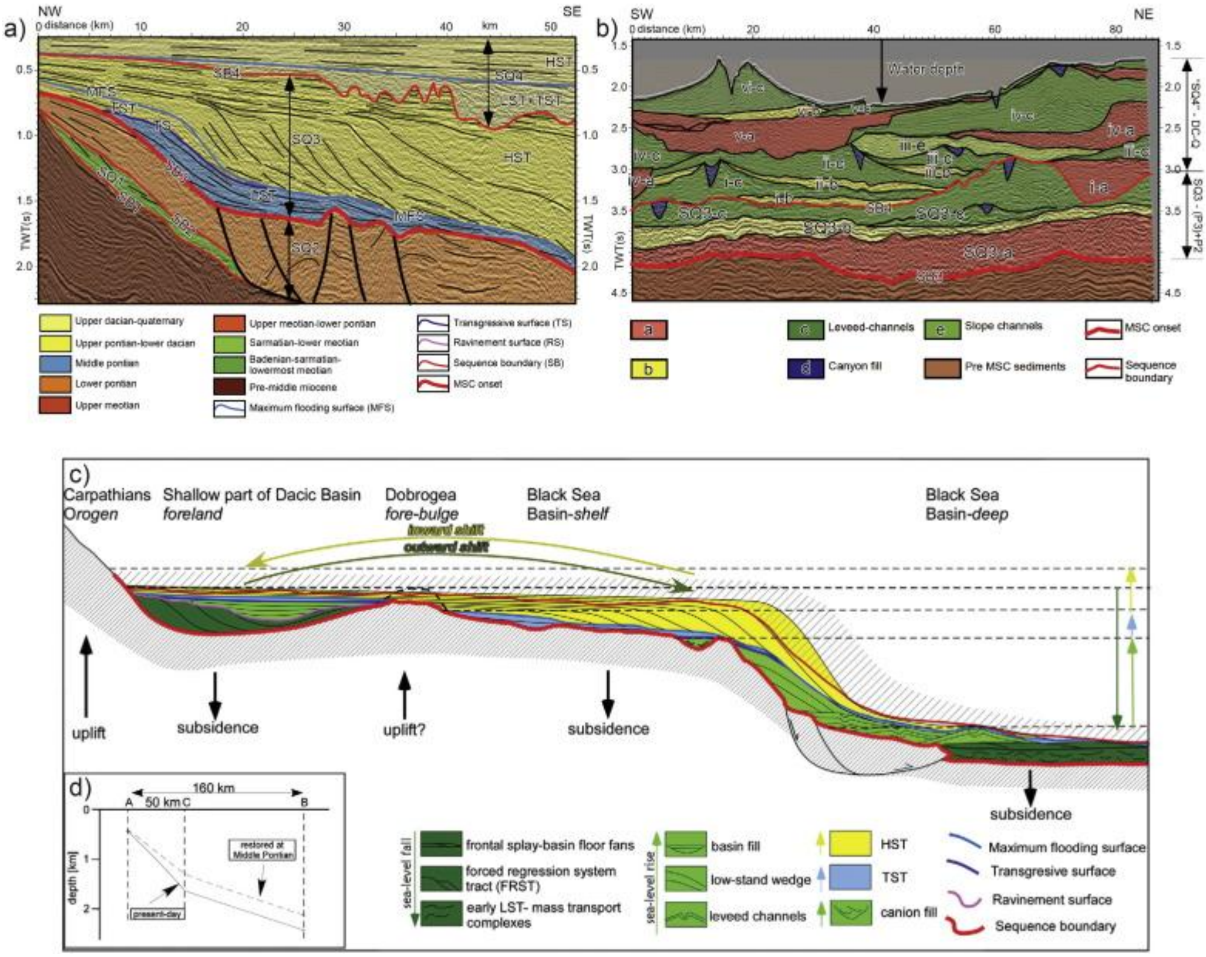

Fig. 12 – The connected effects of the Messinian Salinity Crisis sea-level drop and subsequent rise in the Dacian Basin – Black Sea system (after Munteanu et al., 2012). a) Interpreted seismic line illustrating the detailed internal geometry of sequences observed on the western shelf of Black Sea; b) Seismic interpretation of the detailed geometry of Middle Pontian – Quaternary deep-water deposits observed offshore western Black Sea. P2 – Middle Pontian, P3-Upper Pontian, Dc – Q – Dacian – Quaternary; c) Cartoon illustrating the genetic mechanism controlling depositional events during large sea-level falls in the Dacian–Black Sea basins system. The Dacian Basin provides a trap for sediments as long as accommodation space exists. When this basin is filled, sediments bypass into the Black Sea, where the sedimentation rate increases. d) reconstruction of the amount of sea-level fall during the Messinian Salinity Crisis in the Black Sea.

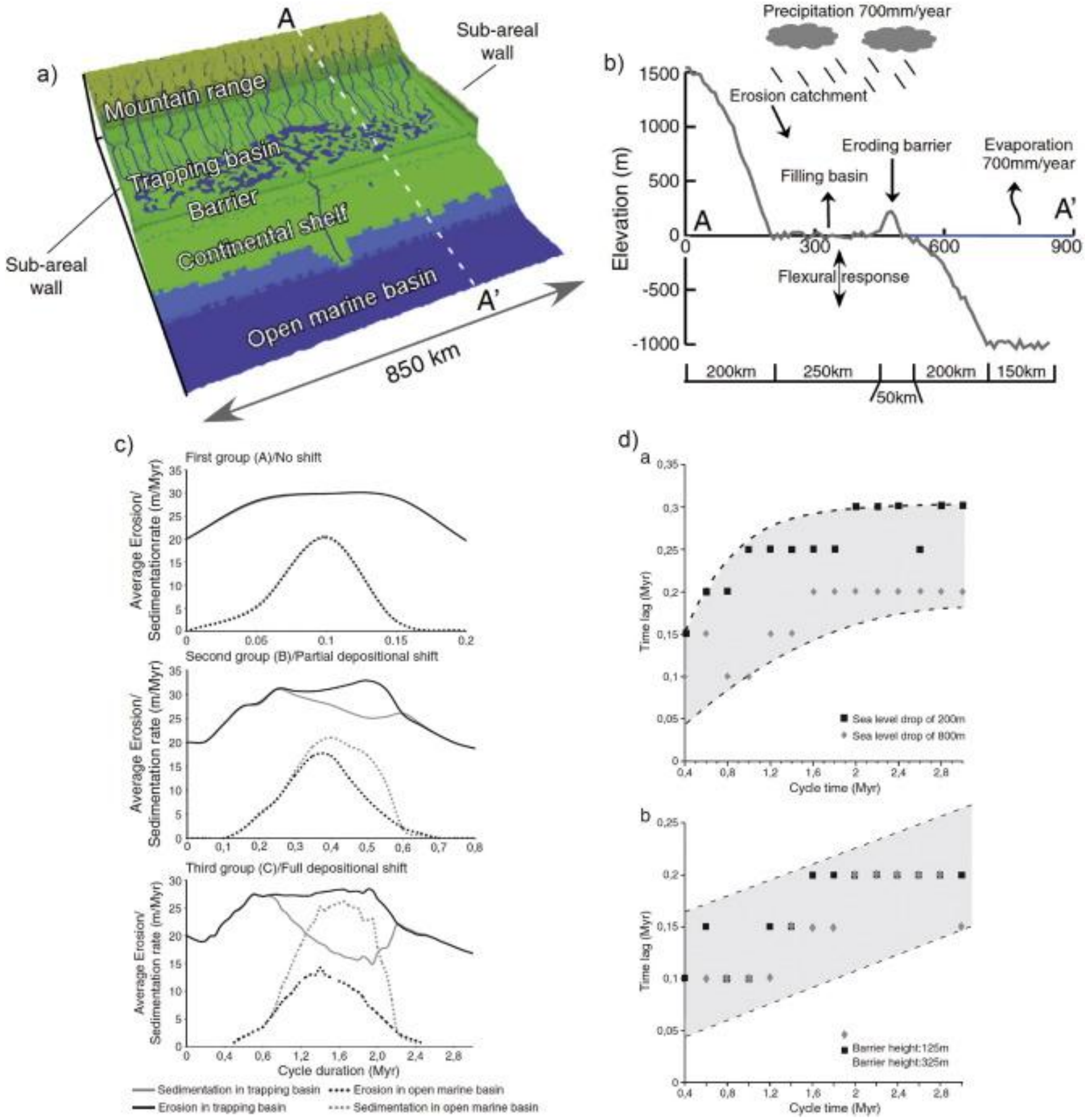

Fig. 13 – Numerical model of the Dacian Basin – Black Sea connectivity during the Messinian Salinity Crisis (after Bartol et al., 2012). a) Conceptual model for numerical setup. The trapping basin is enclosed with lateral walls to prevent sediments or water leaving the model. b) Cross-section illustrating the processes active in the model: erosion and sedimentation, flexural response, precipitation, evaporation; c) Quantifying the types of depositional shifts: A - no shift in deposition; group B - partial depositional shifts, sedimentation rate in the trapping basin remains higher than in the main sink; and group C - full outward and subsequent inward depositional shifts, sedimentation in the main sink basin exceeds the one in the trapping basin; d) Time-lag relationships. Upper panel - Relationship between the time-lag and duration of a sea level drop between 200 and 800 m, at a constant barrier height of 325 m. The time-lag increases with the duration of the sea level cycle Lower panel - Relationship between the time-lag and the cycle time for a barrier height of 125 and 325 m, at a sea level drop of 800 m. The time-lag does not appear to be dependent on the barrier height.

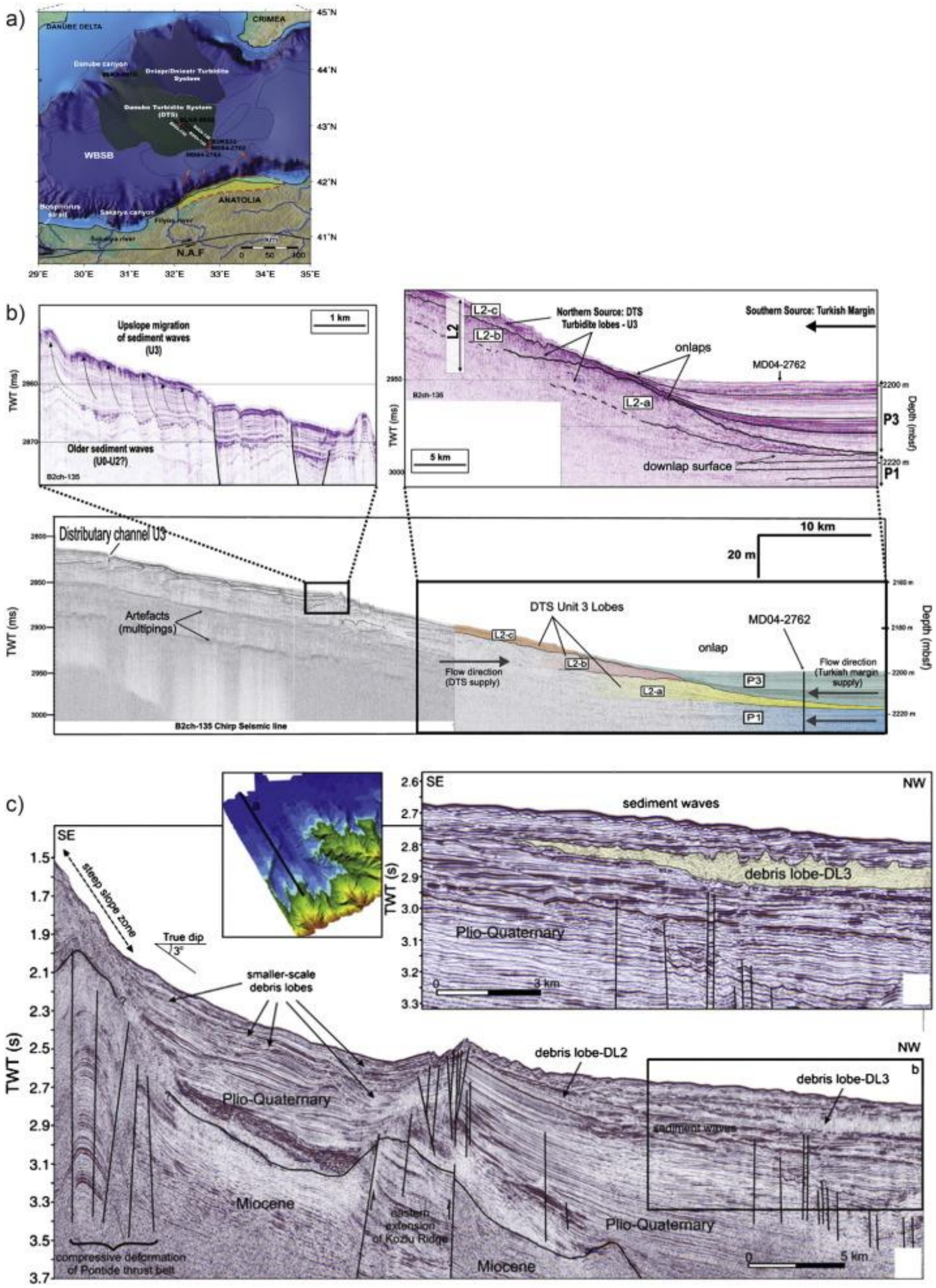

Fig. 14 – Comparison of the recent to active sedimentation in the western Black Sea between the Danube-Dniepr and the Sakarya deep sea sedimentation (after Lericolais et al., 2013; Dondurur et al., 2013). a) Location map with the extent of the deep-sea sedimentation; b) Chirp seismic profile and interpretation. Seismic units P1 and P3 are associated with fine-grained distal turbidites. Unit L2 corresponds to a chaotic to bedded continuous reflections package. L2-a, L2-b and L2-c are three sub-units of L2 corresponding to three individual depositional bodies; c) High-resolution shallow seismic line and interpretation demonstrating the high instability during the recent evolution of the Black Sea, offshore Turkey. The seismic lines are extending from shallower shelf to abyssal plain showing two buried large debris flow lobes DL2 and DL3. The upper panel is a close-up and interpretation of DL3 debris lobe.