GEOLOGICA CARPATHICA, AUGUST 2015, 66, 4, 331–344 doi: 10.1515/geoca-2015-0029

Turbidites as indicators of paleotopography, Upper Miocene Lake Pannon, Western Mecsek Mountains (Hungary)

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(Manuscript received October 31, 2014; accepted in revised form June 23, 2015)

Abstract: The floor of Lake Pannon covering the Pannonian Basin in the Late Miocene had considerable relief, including both deep sub-basins, like the Drava Basin, and basement highs, like the Mecsek Mts, in close proximity. The several km thick lacustrine succession in the Drava Basin includes profundal marls, basin-center turbidites, overlain by shales of basin-margin slopes, coarsening-upward deltaic successions and alluvial deposits. Along the margin of the Mecsek Mts locally derived shoreface sands and deltaic deposits from further away have been mapped so far on the surface. Recent field studies at the transition between the two areas revealed a succession that does not fit into either of these environments. A series of sandstone a few meters thick occurs above laminated to bioturbated clayey siltstone. The sandstone show normal grading, plane lamination, flat erosional surfaces, soft-sediment deformations (load and water-escape structures) and sharp-based beds with small reverse faults and folds. These indicate rapid deposition from turbidity currents and their deformation as slumps on an inclined surface. These beds are far too thick and may reveal much larger volumes of mass wasting than is expected on the 20–30 m high delta slopes; however, regional seismic lines also exclude outcropping of deep-basin turbidites. We suggest that slopes with transitional size (less than 100 m high) may have developed on the flank of the Mecsek as a consequence of lake-level rise. Although these slopes were smaller than the usually several hundred meter high clinoforms in the deep basins, they could still provide large enough inertia for gravity flows. This interpretation is supported by the occurrence of sublittoral mollusc assemblages in the vicinity, indicating several tens of meters of water depth. Fossils suggest that sedimentation in this area started about 8 Ma ago.

Key words: Late Miocene, Lake Pannon, Mecsek, turbidites, slope, synsedimentary folds, soft-sediment deformation.

Introduction

Sandy turbidites are commonly found in bathyal water depths, namely below 200 m. On the other hand, flume experiments showed that development of turbidity currents is independent of water depth, as they are generated by density difference between the sediment laden current and the ambient fluid, therefore the only rule of thumb is that they form below the storm wave base (Walker 1984). Mud is usually transported over the shelf. It has been fairly commonly stated that prodelta sediments on shelves contain turbidites. Cases, however, where thick sand was deposited from gravity flows are far less common. Evidence from several prodelta to shelf regions proves that either storm-generated, wave-supported turbidity flows (Nelson 1982; Fenton & Wilson 1985; Myrow et al. 2002; Traykovski et al. 2007) or hyperpycnal flows related to highly concentrated river plumes during floods (Mulder et al. 2003; Pattison 2005; Lamb et al. 2008) are efficient enough in sediment transport. The former being of short duration produces rather thin, while the prolonged hyperpycnal flows may deposit relatively thick beds.

A more favourable situation for the deposition of thick turbidite sands in the prodelta is when the delta slope and the basin margin slope are united, as in the case of shelf-edge deltas. Depending on a complex set of external factors a wedge-shaped sandy turbiditic delta front can develop without reaching the base of slope or without generating major fans (Plink-Björklund & Steel 2005).

Both deltaic sediments and turbidites are common in the Upper Miocene lacustrine successions (sediments of Lake Pannon) in the Pannonian Basin (Bérczi & Phillips 1985; Juhász 1991; Lucic et al. 2001; Pavelić 2001; Saftic et al. 2003; Krézsek & Filipescu 2005; Vrbanac et al. 2010). Deltaic deposits have been studied both from cores (Juhász 1992; Korpás-Hódi 1998) and outcrops (Sztanó et al. 2013a). Their architecture was imaged by high-resolution seismic surveys (Horváth et al. 2010), unveiling only few tens of meters high delta slopes, which correspond to funnel-shaped portions mostly in gamma and/or SP well-logs, interpreted as coarsening upwards mouth bar (Juhász & Magyar 1992; Juhász 1994) or deltaic units (Sztanó et al. 2013a). Mudstones with only centimeter-scale fine sandy interbeddings with current and wave ripple laminations and shell lags have been reported from the few meters deep prodelta regions. They were formed mainly by storms (Sztanó et al. 2013a).

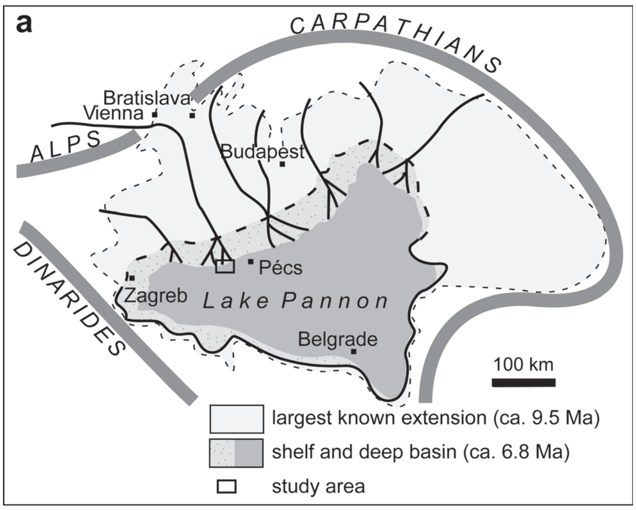
Although turbidites are exposed on the surface in uplifted peripheral subbasins of the Pannonian basin system, such as the Transylvanian Basin (Krézsek & Filipescu 2005; Sztanó et al. 2005; Tőkés et al. 2013) or the Zagorje Basin (Kovačić et al. 2004), from the central basins they have been reported only from cores, coming from positions several km deep in the basin interiors, where molluscs also point to profundal conditions (Juhász & Magyar 1992). Turbidites in the deep basin interiors are related either to the confined basin-center accumulations up to a thickness of 1000 m (Juhász 1992, 1994) or to the base-of-slope turbidite systems (Sztanó et al. 2013b; Bada et al. 2014). The 400–600 m (occasionally 1000 m) high and 8–10 km long basin margin slopes, bridging the morphological shelf and the deep basins, served as pathways of mass gravity flows feeding both types of turbidite systems (Pogácsás 1984; Magyar 2010; Magyar et al. 2013). The principal difference between the two types of turbidite systems is in the areal extent of the major sand bodies and in their thickness, the latter being in the range of hundred meters in deep confined basin centers or few tens of meters in the unconfined base-of-slope systems (Sztanó et al. 2015). The individual beds depending on their position in the turbidite systems are thin-bedded silty to very fine-grained sandy, or medium-bedded graded turbidites of well-developed Bouma sequences intercalated with shales or several meter thick amalgamated, commonly massive beds with various soft-sediment deformation structures (Sztanó et al. 2013b). In addition, features related to large slumps are common in the base-of-slope turbidite systems (Bada et al. 2014).

Recent field studies in the SW part of the Pannonian Basin revealed unusual lacustrine sediments on the surface, which show characteristics of turbidites. Formerly these were mapped as nearshore lacustrine deposits (Chikán & Budai 2005), although the occurrence of sublittoral mollusc faunas in the vicinity indicated that deeper environments did develop in the region. The aim of this paper is to document and interpret the depositional environment of this locality and to integrate it into its geological surroundings.

Geological setting

Sediments of the Upper Miocene Lake Pannon (Fig. 1) comprise the major part of the basin-fill succession in the Pannonian Basin, a classic back-arc basin shaped by several low-angle normal and strike-slip faults (Horváth & Royden 1981; Horváth & Tari 1999). In this way a fairly complicated topography had resulted by the end of the Middle Miocene, and it partly evolved further during the Late Miocene post-rift and intervening inversion events (Horváth & Cloething 1996). Due to these differential vertical movements the lake floor had considerable relief, including both deep sub-basins and elevated basement highs in close proximity.

Development of the deep basins reflected by their sedimentary fills follows a uniform pattern, only their initial relief (depth) and local rates of subsidence may have been different. This succession includes profundal marls (Endrőd



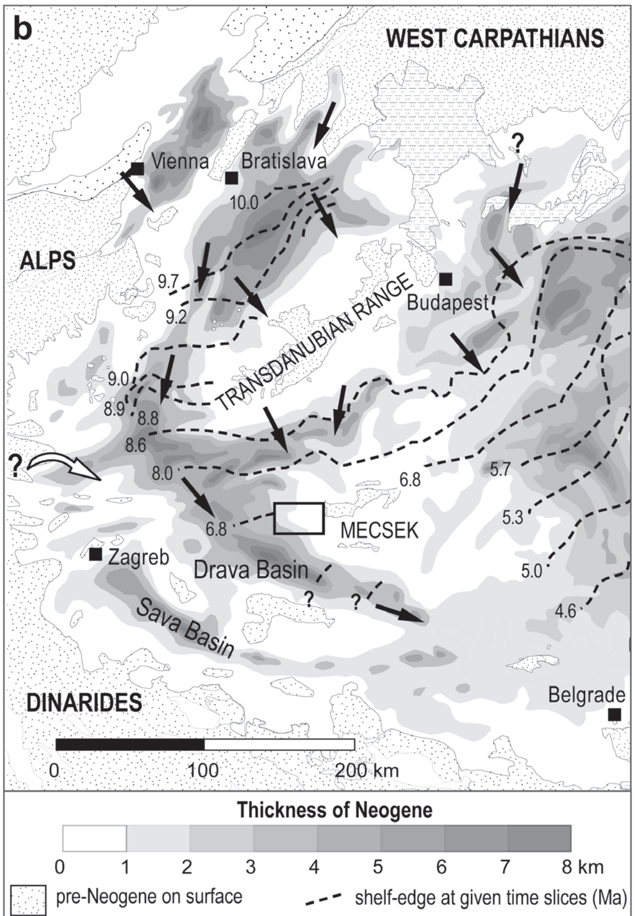


Fig. 1. a — Simplified paleogeographic sketch of Lake Pannon within the Pannonian Basin about 6.8 Ma ago (drawn after Magyar et al. 1999). The north-western part of the basin had already been filled up with sediments. Slope and overlying deltaic sediments were accumulating in the study area; **b** — Present-day depth of the pre-Neogene basement shows the major depocenters which accumulated the most complete lacustrine sedimentary successions from profundal marls to alluvial deposits up to a thickness of several kilometers during the Late Miocene. Most of the present-day hilly areas were parts of basement highs which got flooded only during the late Late Miocene, and hosted only some hundred meters thick, mostly relatively shallow-water lacustrine sediments. Delta progradation over these elevated areas is proven by sediment transport directions among others. Shelf edge positions after Magyar et al. 2013.

Formation), basin-center turbidites (Szolnok Formation), slope shales (Algyő Formation), stacked deltaic successions (Újfalu Formation) and finally alluvial deposits (Zagyva Formation), and reflect gradual fill-up of these basins due to high sediment supply from Alpine-Carpathian source areas (Fig. 1) (Magyar et al. 2013). This basin-type succession was also described from the Drava Basin (Fig. 2) near the study area (Saftić et al. 2003).

In contrast, the basement highs, some emerging above water level as islands or peninsulas during the early Late Miocene, may have become inundated — partly or fully — only later, at varying time points. These areas were marked by shoreface or deltaic deposits of local origin, usually overlain by shales (Csillag et al. 2010). The fauna of these shales may point to water depths of either less than 100 m (Cziczer et al. 2009) or few hundred meters (Magyar et al. 2004) showing great spatial variations. As the deltaic to alluvial feeder systems from remote Alpine-Carpathian source areas reached these locations, the depositional environment changed, depending on the water depth of the given location. If water depth reached a few hundred meters, slope shales and related thin turbidite accumulations may have followed, but these areas lack thick accumulations of both profundal marls and turbidites. If water depth was shallow, deltaic successions followed without underlying slope deposits (Sztanó et al. 2013a). This latter situation was perfectly visualized by high-resolution seismic profiles acquired on Lake Balaton (Sacchi et al. 1999; Horváth et al. 2010). Since the Pliocene these basement highs have been uplifted and partly eroded (Horváth & Cloetingh 1996; Konrád & Sebe 2010), therefore in their vicinity various Upper Miocene lacustrine and older Neogene sedimentary units are exposed today (Fig. 3).

The study area west of the Mecsek Mts (Fig. 1) is transitional between the Mecsek, an emergent Paleo-Mesozoic basement unit, and the Drava Basin, where Neogene sedimentary units up to 6 km thick cover the basement (Fig. 2). The Drava Basin is a well-known hydrocarbon prospecting area, where the lacustrine basin-center turbidites and the overlying stacked deltaic successions could form good reservoirs (Lučić et al. 2001; Saftić et al. 2003; Vrbanac et al. 2010). The architecture of the basin-fill successions, their sequence stratigraphy and structural features have been widely studied (Ujszászi & Vakarcs 1993; Sacchi et al. 1999), thus it is accepted that differential uplift of the Mecsek and large-scale regional folding at about the Miocene/Pliocene boundary resulted in a regionally important unconformity. The turbidites of the Drava basin have recently been studied by Uhrin & Sztanó (2011), but no detailed core analysis is publicly available.

In addition to parts of the Neogene cover, Permo-Triassic sediments, rhyolites and Carboniferous granites also crop out along the western margin of the Mecsek (Fig. 3; Chikán & Budai 2005). These are overlain by Lower to Middle Miocene sediments dominated by siliciclastics, which occasionally acted as the source of the locally derived Late Miocene lacustrine transgressive shoreface sands (Kálla Formation, Kleb 1973). Both of these Miocene sediment packages crop out in the vicinity of the Paleo-Mesozoic basement rocks. Further from the mountains lacustrine deltaic deposits have been mapped on the surface (Kleb 1973; Chikán & Budai 2005). These used to be called the Somló Formation, however, according to the recently developed lithostratigraphic schemes they belong to the Újfalu Formation as a member (Magyar 2010; Sztanó et al. 2013a). The speciality

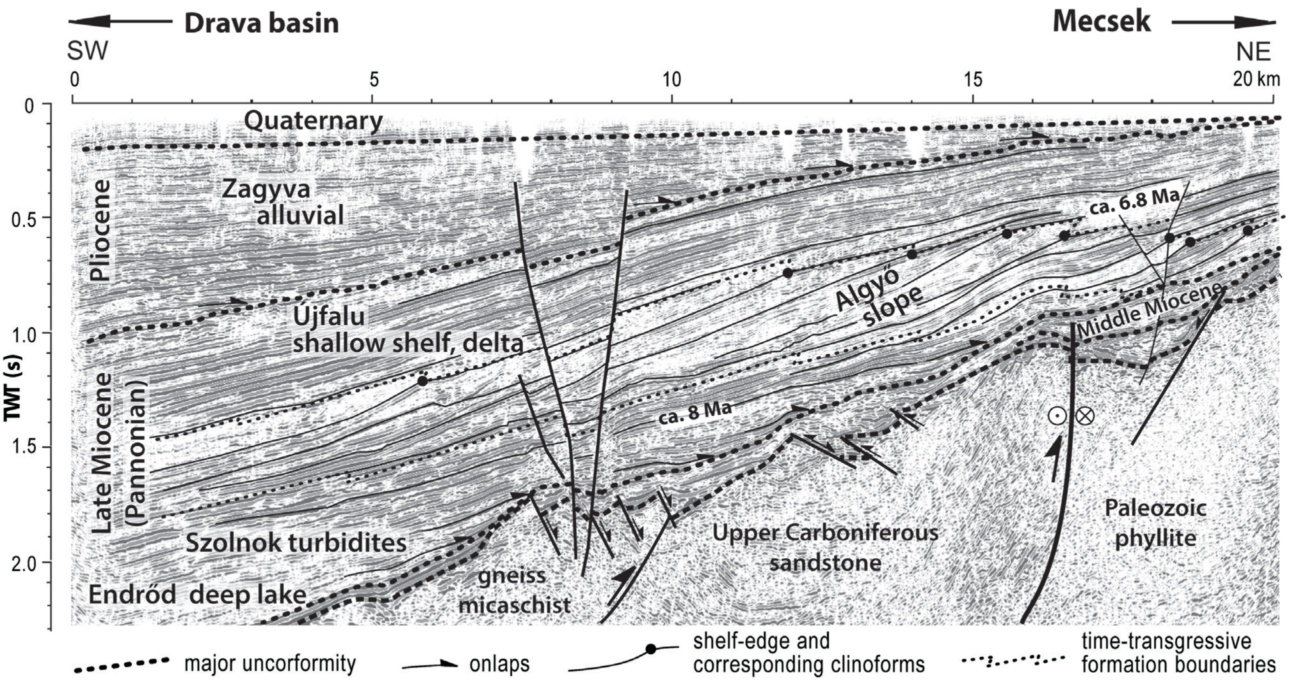


Fig. 2. Seismic profile from the Drava Basin showing the typical basin fill succession: profundal marls (Endrőd Formation), basin-center turbidites (Szolnok Formation), slope shales (Algyő Formation), stacked deltaic successions (Újfalu Formation) and alluvial deposits (Zagyva Formation). For location of section see Fig. 11.

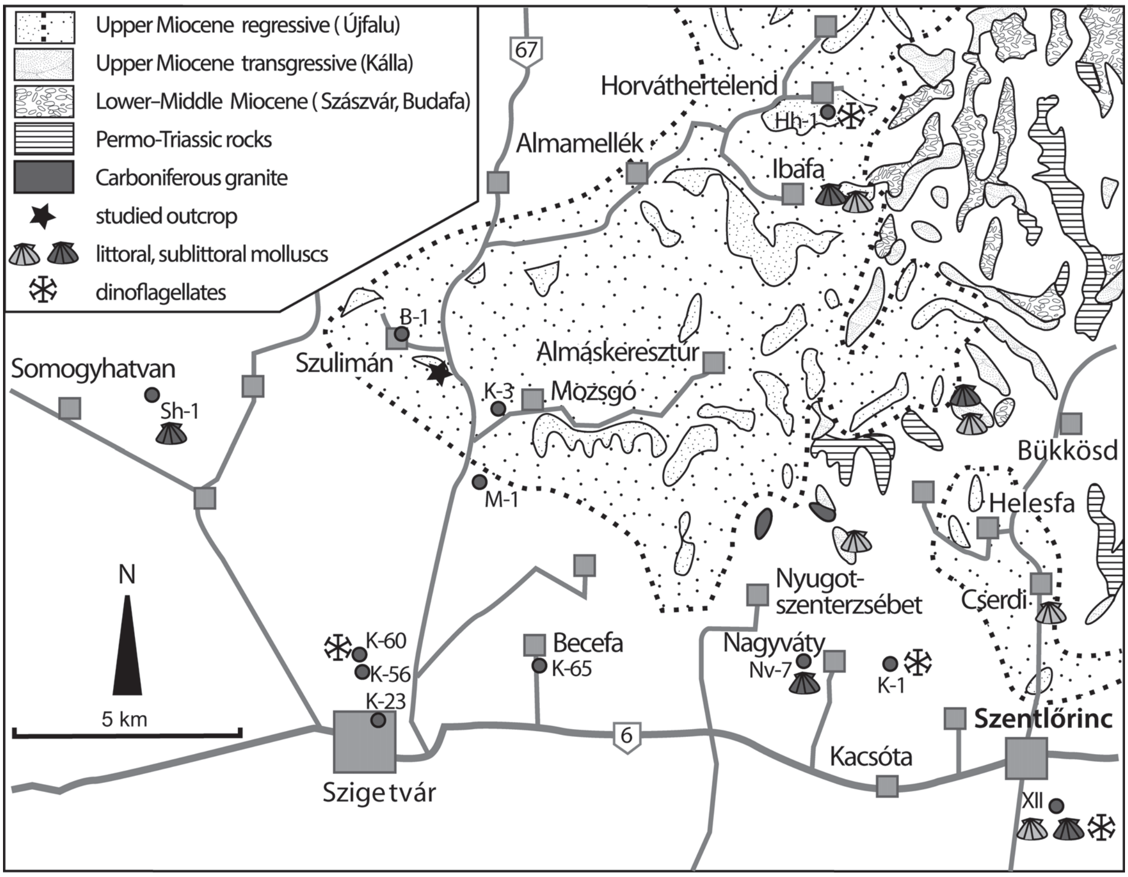


Fig. 3. Simplified geological map of the study area (based on Budai & Gyalog 2010), also showing the inferred surface extension of Upper Miocene sediments.

of the western Mecsek area is that the locally derived transgressive sands might be overlain directly by the regressive ones originating from Alpine-Carpathian sources, without shales in between.

The other local speciality is that the sediments of the above distant provenances are mixed with some locally derived material, indicating denudation and thus a subaerial or shallow flooded position of the Mecsek (Thamó-Bozsó et al. 2014). At the westernmost margin of the surface occurrences of the lacustrine beds, next to the village of Szulimán, an abandoned brickyard exposes a succession that does not fit into the shallow-water deltaic environment.

Sedimentology

The abandoned brickyard is located SE of the village (46°7′24.47″ N, 17°48′47.15″ E). It is approximately 100 m long and 6–7 m high, a large proportion of its surface is covered by debris at present. It exposes Lake Pannon sediments unconformably overlain by about 2 m of red clay-dominated Quaternary deposits. The Upper Miocene lacustrine sediments comprise alternations of muddy and sandy facies units. Above clay to clayey silt beds a 4 m thick series of fine to very fine friable sandstone occurs, and is overlain by thin beds of silts (Fig. 4).

Muddy facies unit — it consists of 0.1–0.4 m thick laminated to fully bioturbated grey clay, clayey silt, silt and white calcareous marl to limestone layers with intercalations of centimeter thick graded, graded-laminated very fine or fine-grained sand beds. Some siltstone beds are fully bioturbated, others contain very small simple vertical mud-filled burrows. The maximum 0.1 m thick limestone beds show very strange characters. Laterally these may interfinger with “ordinary” siltstones. They also produced roundish boudins or ball structures with deformed laminations of the under- and overlying clays. In thin section they are made up of micrite, no fossils other than some unindentified round features, which may be reminiscent of algae, have been found (J. Haas ex verb.)

Sandy facies unit — it is made up of 0.2–0.3 m thick, sharp-based, fine to very fine-grained sandstone beds intercalated with mudstones less than 0.1 m thick. Cementation is very poor, except for two thin limestone beds also appearing between sandstones. Most sandstones do not show any scouring or sole marking. A few have an erosional base, but their relief is not more than 0.2 m (Fig. 5). Some beds are massive, graded to silt with large ball and flame structures at the bottom. Others are graded, massive to parallel-laminated with sharp contact towards the overlying silts. Very thin cross-lamination may also occur together with the parallel lamination. In the uppermost beds hummocky cross-lamination or rather low in-phase wave lamination was observed.

Some beds contain cm-sized rip-up mud-clasts. Secondary soft-sediment deformations are common: not only load balls and flames, but also dish and pipe structures occur. The most spectacular features are, however, synsedimentary folds and related small reverse faults (Fig. 5).

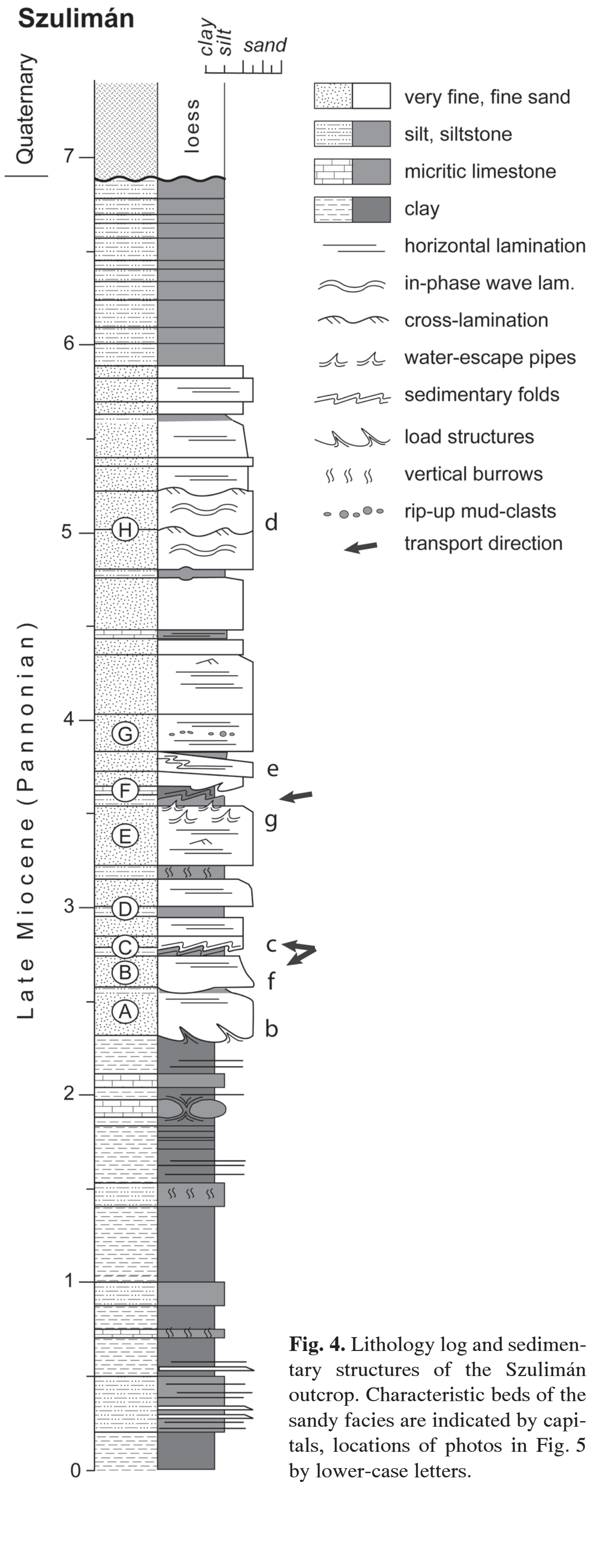


Fig. 4. Lithology log and sedimentary structures of the Szulimán outcrop. Characteristic beds of the sandy facies are indicated by capitals, locations of photos in Fig. 5 by lower-case letters.

Synsedimentary deformations occur at three levels, in beds C, F and G (Figs. 4, 6), in thicknesses below 0.1 m. Those in bed C show a very regular geometry laterally extending for several meters. The other two are more irregular, although the same vergence of folds and faults is observed. These latter are both underlain by beds with water escape features. In all cases the core of the deformed body is a silt or limestone layer, but the overlying sand bed was also included in the deformation. Folded beds show a sharp base which served as a detachment surface; above they are sharply overlain by undeformed layers. The main elements in the deformed beds are reverse faults soling in the basal detachments plane. Fault spacing is variable: more regular folds developed above faults separated by longer distances (Fig. 6a,b), while more closely spaced thrust planes co-occur with intensely deformed, elongated, flame-like layer fragments (Fig. 6c), possibly a result of less cohesive material. Fold morphology can be upright, inclined or overturned. Back folds commonly occur above the thrust planes. Along fault planes material can be dragged into even nearly isoclinal folds (e.g. in Fig. 6a, between 2nd and 3rd thrust planes). Thrust vergences range WSW–WNW, pointing to an overall westerly transport direction.

Interpretation

Depositional processes

The mudstones were formed by suspension settling in quiet waters, below wave base, most likely even below storm wave base. The limestone layers are rather enigmatic. Due to lack of any evidence of subaerial exposure or pedogenesis interpretation as calcretes is excluded. Based on analogy of other carbonates of the lacustrine succession (Magyar et al. 2004; Cziczer et al. 2009) and the algae-like structures it is speculated that these beds formed when carbonate mud accumulated after the bloom of some calcareous algae in the photic zone of the lacustrine water mass. Siltstones point to increased suspension input from distal sources. The thin graded sandstone beds were formed from turbulent flows. These may have been diluted, small-density turbidity currents, or storm-induced density flows.

Although the sandy facies unit is just a few meters thick package, its unusual sedimentological character among the outcropping Lake Pannon deposits in the Pannonian Basin makes it important. A combination of massive, graded, parallel- and cross-laminated structures point to waning flows with decreasing density and a change from turbulence to traction. Therefore this association evokes Bouma sequences (Tabd, Tbc, Tbc´; Fig. 5), and refers to deposition from high-density sandy turbidity currents. Parallel lamination indicates that the currents were supercritical (cf. Southard & Boguchwal 1990) during deposition. The occurrence of hummocky or low in-phase wave lamination also indicates rapid flows of somewhat less velocity or higher thickness (cf. Cheel 1990; Prave & Duke 1990). The predominance of these rapid flow deposits can also be related to hyperpycnal flows, and both may point to proximal-to-source character. Load and water-escape structures may indicate event-like

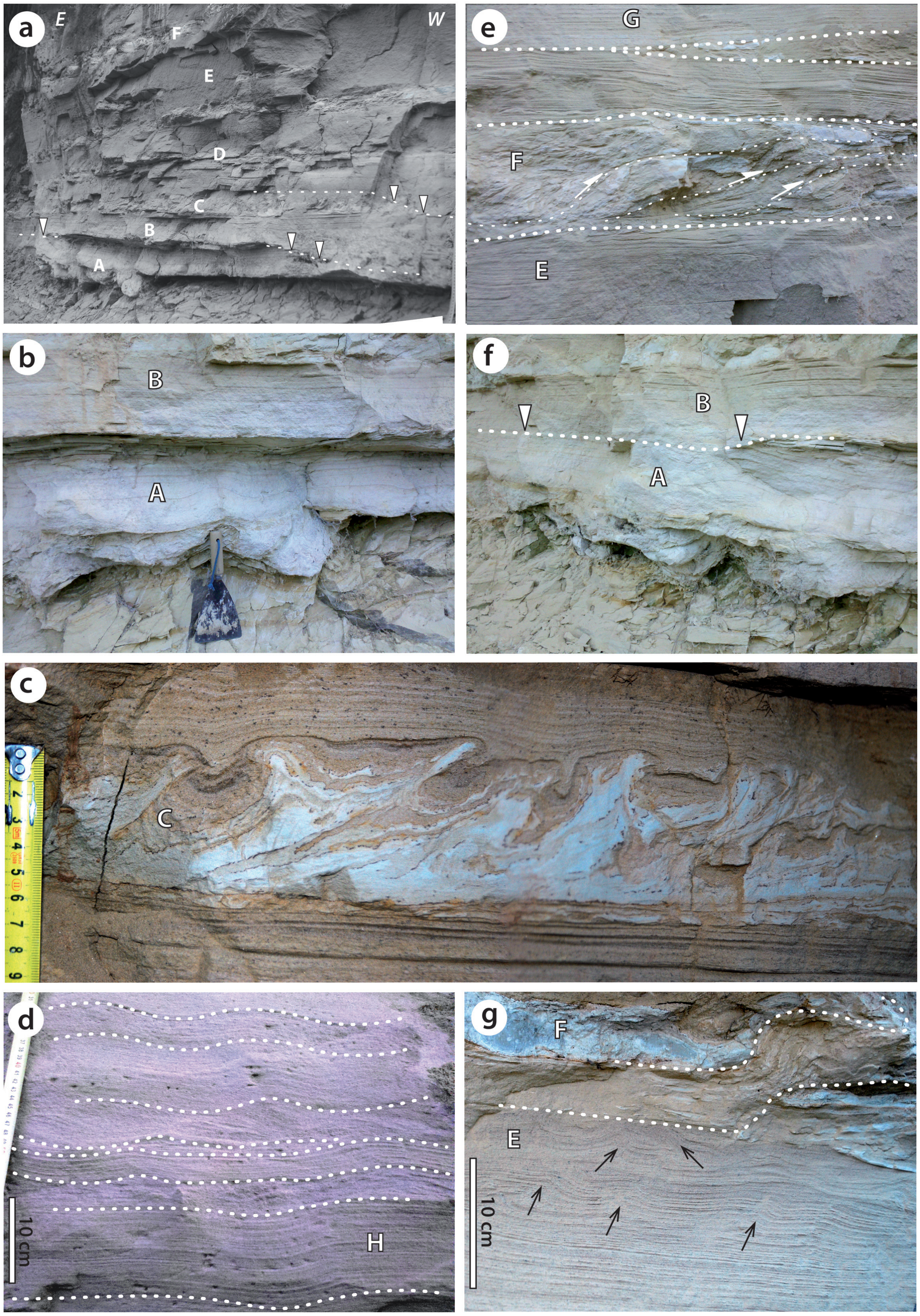


Fig. 5. Sedimentary structures. For location of the photos within the succession see Fig. 4. **a** — overview of sand beds and their erosional contacts; **b** — large load balls at the bottom of massive, graded bed A, overlain by parallel-laminated sands of bed B; **c** — silt bed C with sharp base, synsedimentary folds and reverse faults; **d** — low in-phase wave lamination in bed H; **e** — folded and faulted silt is followed by bed G with various types of water-escape structures; **f** — amalgamation of beds A and B due to erosion; **g** — dish and pipe structures in upper part of bed E, overlain by chaotically folded silt bed F.

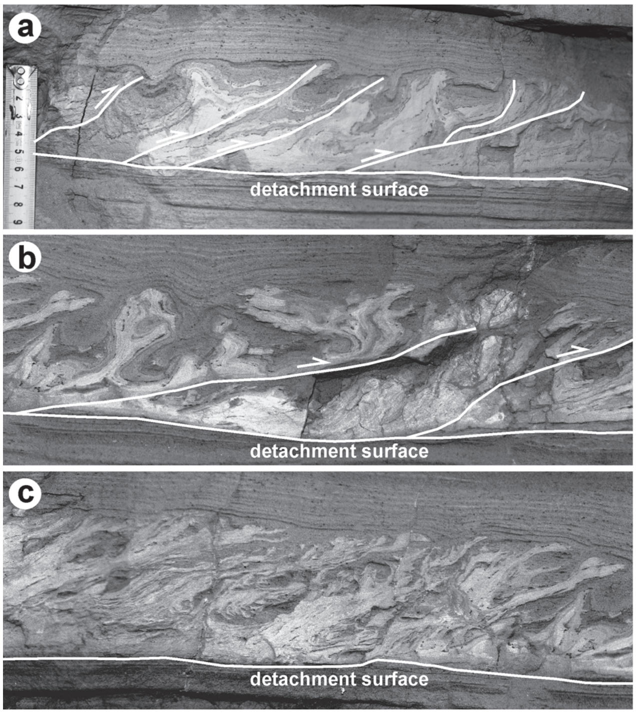


Fig. 6. Syn-sedimentary folds and faults, with increasingly incoherent folds from (**a**) to (**c**). Sediment movement happened from left to right in all photos.

deposition and rapid burial. In deep basin interiors these types of structure with the same thickness of individual beds are reported from deep-water lobes (Bérczi & Phillips 1985; Juhász 1994; Sztanó et al. 2013b), which usually comprise sand-bodies several tens of meters thick. On the margin of these, not only the overall thickness of the lobe deposits, but also bed thickness decreases drastically. On the other hand, the closest occurrence of these deep-water turbidite systems in the Drava Basin is in a distance of 10 km and at a depth of over 1000 m. This environmental interpretation for the Szulimán turbidites can be excluded based on its facies and the geological setting of the area. Another interesting turbidite locality is situated 150 km to the west in Croatia (Kovačić et al. 2004). The succession contains large channel-fills, levee and lobe deposits clearly of deep basin origin, uplifted to the surface by large reverse faults (Tomljenović & Csontos 2001), so they are not regarded as analogues to Szulimán either. The topmost part of the succession, however, contains horizontally to cross-laminated sands, according to the description similar in facies to our locality. These strata were interpreted as "peculiar type" mouth-bar deposits in extremely shallow waters.

In the uppermost strata of the Szulimán outcrop the low in-phase wave lamination to combined wave ripple cross-lamination may point to waning flows generated by storms (cf. Dott & Bourgeois 1982). This current type is rather common on flat-lying surfaces of shelves between fair-weather and storm wave bases. Except for the synsedimentary folds and faults, the other sedimentary structures do not contradict this possibility. Beds, however, are relatively thick, shell lags, post-storm bioturbation or post-storm mudstones are missing, as well as well-developed wave-ripples, which were described elsewhere in the lacustrine setting (Magyar et al. 2006). Therefore it is not excluded that storm-induced currents may have influenced deposition (cf. Myrow et al. 2002), but were not of primary importance.

However, the most important observation which helps the interpretation is the presence of small-scale folds and faults verging consistently in the same direction, which indicates an inclined depositional surface. Vergence of thrusts and folds is downslope, namely in a westerly direction. The deformation represented by the faults and folds is a result of intense shortening, thus the deformed beds are regarded as compressional lower parts of slumps. The above described structures resemble those displayed and interpreted in detail by Alsop & Marco (2011) from the Dead Sea region, though the role of faults is much higher here. Their geometry may reflect late phase slump translation and slump cessation, where non-coaxial downslope deformation took place. The dominance of incoherent folds (sensu Alsop & Marco (2013)) indicates strong deformation, also typical of the lower portions of slumps.

Depositional environment

Based on the recognition of co-existent features of turbidites and inclined topography, the following depositional setting is suggested. Models of experimental turbidity currents predict that in the case of large initial volumes, high-concentration flows on a 1–1.5° slope are likely to deposit such sand beds near the base of the slope, after 3–20 km long transport (Zeng & Lowe 1997). Thus the succession might have been formed on a slope, but in water depths still shallow enough for storm-induced currents to play some role.

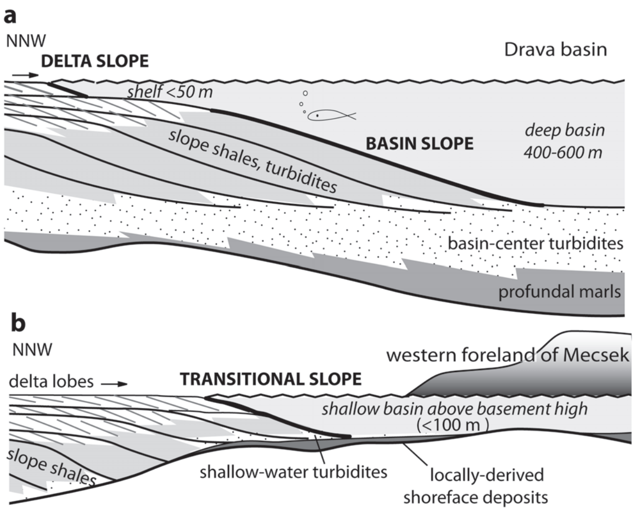


Fig. 7. Sketch of different slope types developed in Lake Pannon. **a** — few 10 s of m high delta slope prograding on the lacustrine shelf towards the basin slope of several hundred meter height; **b** — transitional slope of shelf-edge deltas in shallow-water areas near elevated basement highs. The two sections are parallel, and represent coeval deposits in the Drava Basin and in the area west of the Mecsek Mts shown as an island in the background. The foreland gently dips towards the Drava Basin as well (compare with Fig. 11).

Therefore several hundred meter high basin-margin slopes leading to the Drava Basin are excluded. However, the beds in the sandy facies are far too thick and may reveal much larger volumes of mass wasting than is expected in prodelta of 20–30 m high delta slopes on the shelf of Lake Pannon. Consequently, these small delta slopes are not regarded as the loci of deposition either. Instead it is speculated that a third type of slope had evolved. As the study area is transitional between the deep Drava Basin and the elevated Mecsek area, it is possible that here a water depth in the range of only 100 m developed, therefore the slope of the prograding shelf with shelf-edge deltaic feeder systems on top could not be higher either (Fig. 7). If the usual slope angle of 1–2° was maintained, this transitional type of slope may have been about 3–5 km long. It was smaller in all directions than the usual basin slope, but could still provide large enough inertia for gravity flows, ponding on the shallow western foreland of the Mecsek Mts. Gradually, but rather rapidly, the width of the shallow-water area in front of the prograding slope decreased and was exceeded by the transport distance of turbidity currents. As a result, sediments bypassed and were dumped into the deep basin in the south-southwest; this inhibited the accumulation of thick turbiditic successions in these transitional areas.

Evolution of deltas and slopes in the region

The depositional model of the transitional slope is also supported by available well data. Unfortunately no cores, but gamma-ray, standard potential and resistivity curves and archive reports on cuttings and cores are available. A geological section partly parallel to the progradational direction of the shelf-slope system (cf. Fig. 1b) was constructed (Fig. 8; from cluster A to well K-65). Coarsening up-units of 20 m

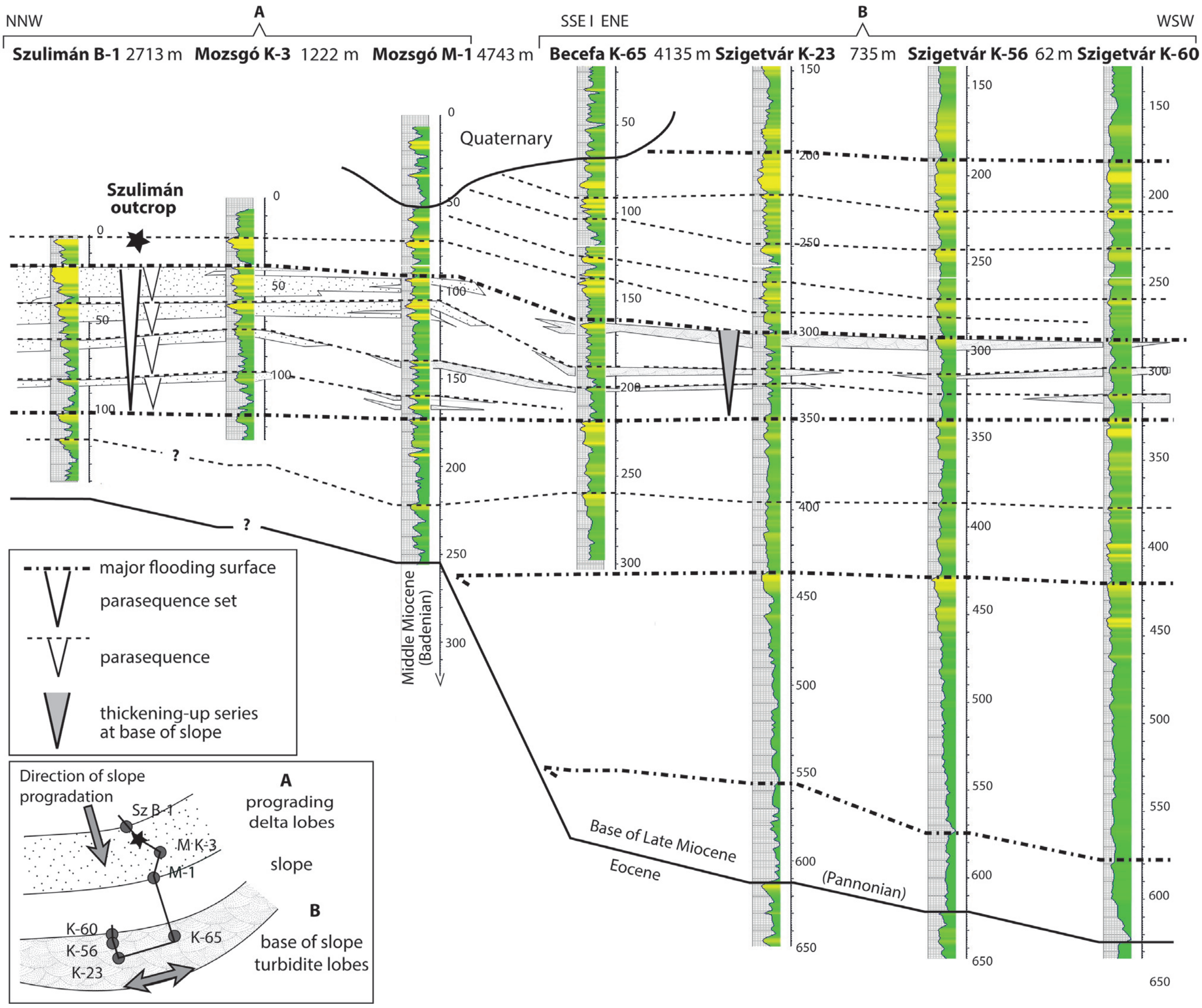


Fig. 8. Correlation panel through gamma-logs of shallow wells. The progradational parasequence set of cluster **A** is interpreted as delta progradation on the shelf-edge. The correlative set in cluster **B** is interpreted as deposits of the transitional slope, with thin turbiditic lobe deposits mimicking the shelf-edge progradation. Note that wells of cluster A to K-65 are in dip direction, while wells of cluster B are in strike with respect to the supposed slope. The base of the progradational set was used as a datum, measured depth is shown along the wells. GR logs are shaded according to facies, yellow is sand, green is shale.

thickness are interpreted as deltaic lobes, based on analogous successions of the Balaton region (cf. Sztanó et al. 2013a). They can also be regarded as parasequences that comprise a progradational parasequence set. This is best imaged between 20–100 m measured depth in the well Szulimán B-1 situated only a few hundred meters from the studied outcrop. They are sharply overlain by "hot" shales locally producing the highest count on the logs. In the following 4 km distance the progradational parasequence set is still obvious in gradually increasing depth (40–120 m MD of Mozsgó K-3 and 85–170 m MD of Mozsgó M-1). The coarsening/thickening-up set, as well as the overlying thin shales are recognized on logs of cluster B in an even larger measured depth, at about 300 m in Szigetvár K-23, 56 and K-60. It means that a regional, post-sedimentary tilt of 2.5–3° towards the S must have happened, which is reasonable if the depth of base Late Miocene or the position of the pre-Neogene is considered (Fig. 2) (Kőrössy 1989; Haas et al. 2010).

Delta front to mouth bar sandstones of the youngest parasequences reached the area of Mozsgó-1, but those of the oldest ones did not develop here. Instead their time-equivalents, several thin sandstones appear in the lower part of the parasequence set. Similar sandstones forming sheet-like bodies are present in the wells of cluster B. Their areal extent may have attained 5 km, while their thickness is less than 10 m, usually 2–5 m only. It should be emphasized that this portion of the section (Fig. 8) is parallel with the strike of the slope. These sandstone sheets are interpreted as small series of turbidites near the base of the slope, in front of the prograding shelf-edge deltas. The mostly upwards increasing thickness of the sandstone sheets, namely their thickening-up stacking pattern mimics the progradational character of the deltaic parasequence set.

The overlying shales, widespread all over the study area mark a regional flooding event. As a result of this transgression, the coastal region of Lake Pannon might have stepped back to the north, so far that for a short interval only condensed deposition took place here. After some time a new prograding slope of about the same ca. 80–100 m height might have evolved, thus base of slope turbidite sheets could form again. The studied succession at Szulimán might be part of this system. The prograding deltas of this new phase might have reached the Szigetvár area when sand bodies at about 190–200 m MD in wells K-23 and K-56 were deposited.

Actually the turbidites at Szuliman indicate a transitional slope, which did not develop just because of the marginal position but also because the lake level rose. Unfortunately no biostratigraphic evidence exists on the time span of these major floodings and the related progradational parasequence sets. However, based on regional studies of the aggradational to progradational character of the shelf slope of Lake Pannon (Sztanó et al. 2013b), it is supposed that they occurred regularly in about 100 kyr intervals.

Fossils

The sediments of Lake Pannon often contain fossils of endemic molluscs, ostracods and dinoflagellates, providing a basis for the interpretation of the depositional environment and age of the enclosing sediments (Magyar & Geary 2012). For instance, distinct mollusc assemblages characterize the littoral, sublittoral, and profundal zones of the lacustrine environment, thus offering a tool to estimate the paleo-water depth (Juhász & Magyar 1992; Magyar 1995; Geary et al. 2000).

Our efforts to recover either macro- or microfossils from the Szulimán outcrop have been unsuccessful so far. Fossiliferous outcrops and borehole sequences, however, were documented in the vicinity (Fig. 3). Although these data cannot be applied directly to the Szulimán succession, they give us a general understanding of the age and environmental conditions of Lake Pannon deposits in the region.

Environmental interpretation of molluscs

Two different mollusc assemblages occur in the area (Figs. 9, 10). The first one is characterized by *Congeria zagrabiensis*, "*Pontalmyra*" *otiophora*, *Valenciennius reussi*, and may also include *Congeria rhomboidea*, *C. croatica, Lymnocardium majeri*, *L. cristagalli*, *L. hungaricum*, other cardiids and pulmonate snails (Fig. 9). This assemblage is found in fine-grained sediments, such as marl, clay or silt, and is interpreted as a sublittoral fauna. It has been found, for instance, in surface outcrops near Ibafa (collected by G. Chikán), Bükkösd, and in the Sh-1 borehole, at 551 m depth, close to the bottom of the Lake Pannon sequence.

The other mollusc assemblage consists of cardiids, such as *Lymnocardium ferrugineum*, *L. pelzelni*, *L. schmidti*, *Prosodacnomya* sp., etc., dreissenids, such as *Congeria triangularis*, *C. balatonica* and *Dreissenomya*, and prosobranch gastropods, such as *Viviparus* and *Melanopsis* (Fig. 10). These fossils are found in sands, and are interpreted as representing a shallow-water, littoral mollusc fauna. This assemblage is found, for example, in Nyugotszenterzsébet (Bujtor 1992), Cserdi, Ibafa and Bükkösd.

However distinct the two assemblages are, they do not display any clear geographical separation within our study area. Instead, they repeatedly occur above each other in some sequences, for example in Bükkösd and Ibafa, making the paleoenvironmental interpretation a challenge.

An unusual mixture of the two assemblages in a thin gravel layer was recorded in borehole Nagyváty-7, at 290 m depth. Littoral forms, such as *Prosodacnomya*, *Melanopsis* and *Theodoxus*, occurred together with sublittoral *Lymnocardium majeri* here, obviously as a result of reworking and redeposition. This process implies the presence of a morphological gradient, possibly similar to the one that initiated the Szulimán turbidites.

Biochronostratigraphy

Apart from the deep Szentlőrinc-XII borehole where almost all Lake Pannon dinoflagellate and mollusc biozones were identified ((Sütő-Szentai 1989, 1991, 1995, 2000), Korpás-Hódi in (Wéber 1982)), the fossiliferous deposits of the study area belong to the youngest biozones of Lake Pannon sediments. These include the Spiniferites validus, Spiniferites tihanyensis, and Galeacysta etrusca microplankton zones, some freshwater algal ecozones, and the Congeria rhomboidea sublittoral and Prosodacnomya littoral mollusc zones (Magyar & Geary 2012).

According to the distribution of microplankton zones, Sütő-Szentai (1995) suggested that the study area was first flooded by Lake Pannon at the end of the Spiniferites validus chron, which is considered to be slightly older than 8 Ma (Magyar & Geary 2012). NE (updip) of Szulimán, in Horváthertelend (Fig. 3) and beyond, only the validus and tihanyensis zones were found, whereas S (downdip) of Szulimán, in the boreholes Szentlőrinc-XII, Kacsóta-1, and Szig-K-60 (indicated as "Szigetvár-III" in the original publications), the youngest brackish-water biozone, the etrusca zone was also identified above the validus and tihanyensis zones. According to Sütő-Szentai (1995), this pattern suggests that the etrusca zone was partly eroded from above the uplifted north-eastern part of the study area.

The bases of the Congeria rhomboidea, Prosodacnomya and Galeacysta etrusca zones roughly correspond to each other, and can be dated as ca. 8 Ma (Magyar & Geary 2012). The age of the outcropping Lake Pannon sedimentary sequences discussed in this paper, including the Szulimán outcrop, can thus be estimated as 7 to 8 million years.

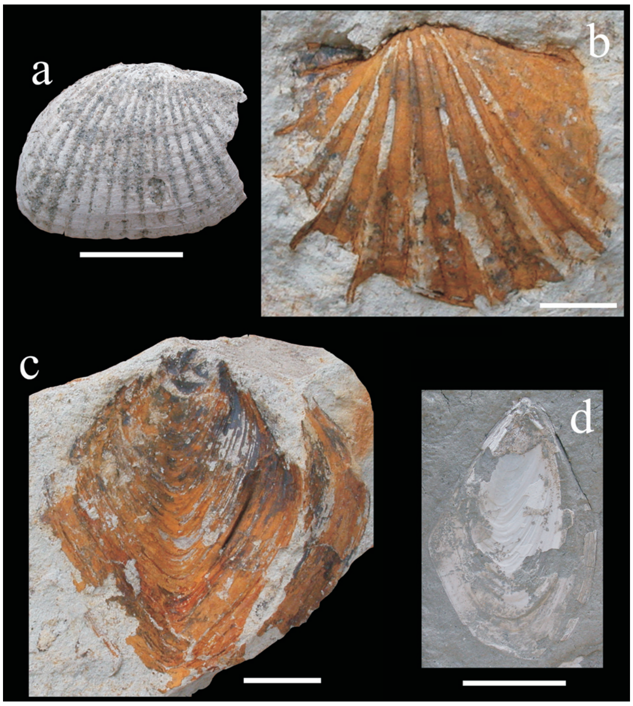


Fig. 9. Sublittoral molluscs from the study area. **a** — *Lymnocardium majeri*, borehole Nagyváty (Nv)-7, 290 m; **b** — *Lymnocardium cristagalli*, Ibafa (collected by G. Chikán); **c** — *Congeria zagrabiensis*, Ibafa (collected by G. Chikán); **d** — *Congeria zagrabiensis*, borehole Somogyhatvan (Sh)-1, 551 m. Scale bars: 1 cm.

Regional stratigraphy and flooding events: a discussion

A regional geological cross-section (Fig. 11) was compiled in order to show the position and connection of various lacustrine formations and to understand how the deposits of the transitional slopes can be classified. In the Drava Basin the deep lacustrine marls and the following thick turbiditic series, overlying and onlapping on the Middle Miocene synrift sediments, occur at a depth of at least 2 km. The Szolnok Formation pinches out towards the basin margin. Therefore the next Algyő Formation — slope-related turbidites and slope shales — overlies either the Szolnok Formation or older Paleogene to Neogene or the basement near the basin margins. It is important to keep in mind that the clinoforms in

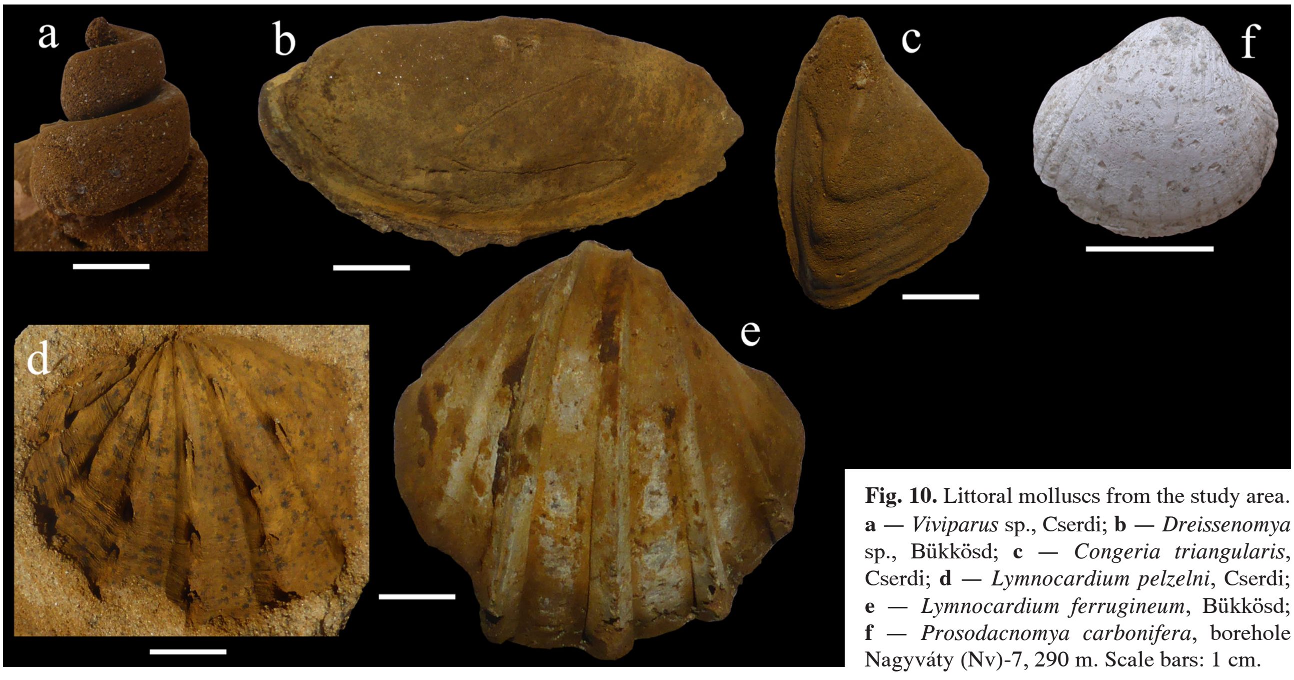


Fig. 10. Littoral molluscs from the study area. **a** — *Viviparus* sp., Cserdi; **b** — *Dreissenomya* sp., Bükkösd; **c** — *Congeria triangularis*, Cserdi; **d** — *Lymnocardium pelzelni*, Cserdi; **e** — *Lymnocardium ferrugineum*, Bükkösd; **f** — *Prosodacnomya carbonifera*, borehole Nagyváty (Nv)-7, 290 m. Scale bars: 1 cm.

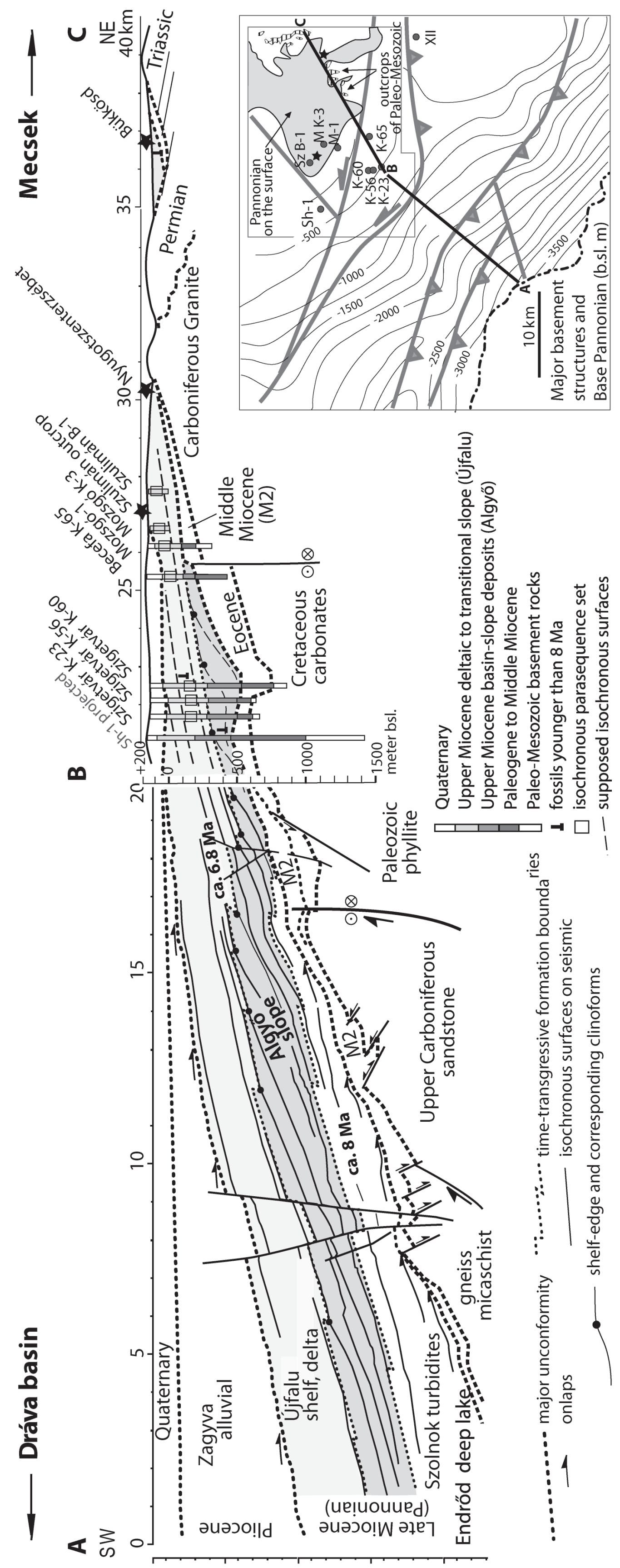


Fig. 11. Regional cross-section constructed from a seismic profile (see Fig. 2), well data and geological maps shows various basement and their Cenozoic cover units from the Drava Basin to the Mecsek Mts. The Upper Miocene lacustrine succession is tilted and thickening towards the SW. It is erosionally truncated by the Mio/Pliocene and the Quaternary unconformities. Progradation of the shelf edge, namely the boundary of the Algyő and Újfalu Formations, is clear from seismic profiles but is difficult to reconstruct from well data. The positions of sediments indicating the transitional slope discussed in this paper are marked on the wells by the isochronous parasequence set. Basement map after Haas et al. 2010.

this figure are not imaged in dip direction but in an oblique view. Based on a net of 2D sections (Uhrin 2011) slope progradation happened from NNW to SSE. It is evident from the section that as the slope advanced towards the deeper parts of the Drava Basin the height of slope increased, from about 150 m at the NE margin to 300–400 m in the basin interior. The decompacted thickness of slope shales indicates minimum estimates of water depth up to 610 m (Balázs et al. 2015).

The black dots on the seismic profile represent the shelf-slope break marking the boundary between the steeply dipping slope of the Algyő Formation and originally horizontally deposited Újfalu Formation, the product of repeated delta progradation on the lacustrine shelf. Connecting these points indicates the shelf-edge trajectory, which shows the development of the lacustrine base level, a result of subsidence and climatically-driven lake level oscillations. At the NE side of the seismic profile first a flat trajectory is evident with the shelf edge of only 150 m high slopes near the basin margin. This relatively small water depth is the result of the first flooding of the basin margin. It is followed by a steeply ascending trajectory, which points to a period when the lacustrine base level was rising. Aggradation of the shelf was about 200 m accompanied by only a modest rate of progradation. About 6.8 Ma ago (Magyar et al. 2013) the situation gradually changed, the trajectory became almost flat, indicating that aggradation ceased and progradation became dominant. After that only minor rises of the shelf-edge trajectory are visible. The ultimate cause of the major base-level rise somewhat before 6.8 Ma is unknown. The structural evolution of the Mecsek area is complex enough to involve a local increase in the rate of subsidence. Climatic impact on lake-level rise is also equally possible, but if it played a significant role it must be evident elsewhere in Lake Pannon sediments of the same age.

Actually this base-level rise roughly correlates in time with a significant backstepping of the shelf margin in eastern Hungary, where a second set of large clinoforms appears on seismic sections of the eastern Great Plain (Magyar & Sztanó 2008; Magyar 2010). This event also resulted in the accumulation of extremely thick successions of the deltaic sediments shown by well data (Juhász 1992, 1993), reported not only from the western and eastern parts of the Great Plain but also from the Drava Basin (Juhász 1998). The differences in the basin fill architectures in the Great Plain and rates of backstepping were explained by variations in sediment supply (Vakarcs et al. 1994; Csato et al. 2007). As this unit is overlain by a locally important unconformity, much effort was done to explain the origin of the latter, but much less attention has been paid to the reasons for lake-level rise below. Mostly it has been emphasized that the main driving force is differential subsidence and uplift of the region (Juhász et al. 2007). During about the same time interval economic lignite seams also formed on the northeastern margin of Lake Pannon, worked in huge open pit mines on the foothills of the Mátra and Bükk Mts. Their repeated successions up to a thickness of over 200 m can be followed close to the area where the deltaic succession is thick in the Great Plain (Magyar 2010). These pieces of evidence, together with multiple paleobiological data point to a period characterized by increased rates of precipitation from ca. 7.2 Ma (Magyar 2010).

Whatever the trigger was, this period of intense base level rise is particularly interesting because successions NE of the seismic profile, shown in detail by well data (Fig. 8), were deposited during this time. As a result of flooding, lacustrine deposits could extend as far as Nyugotszenterzsébet or Bükkösd. During this transgression wave-eroded coasts and small deltas of locally derived material developed. As was discussed before, fossils at several sites near the base of the lacustrine succession — in well Sh-1, Bükkösd, Ibafa — are not older than 8 Ma, therefore all these events must have occurred after 8 Ma. As the base level repeatedly rose, these local transgressive deposits partly got reworked and eroded. The stacked parasequence sets discussed formerly indicate that the base level was rising continuously, without noticeable falls.

The lower, dominantly shaly portion of the lacustrine succession in the Szigetvár wells (Fig. 8) may still be formed on the same, roughly 150 m high slope which is depicted on the seismic section, thus this part can be named the Algyő Formation. The overlying cyclic sequences were formed above the shelf edge, so if they appeared on seismics they would be called Újfalu, without doubt. Well data, with coarsening and thickening up series alternating with shales are also features frequently reported from the Újfalu Formation. The speciality of the area, however, is that it can be demonstrated that not only delta slopes, but longer and taller transitional slopes developed, and allowed the accumulation of turbidites at their base. Although the studied outcrop is unique so far, it must be emphasized that sedimentary characters are very different from other facies of the Újfalu Formation. The development of this third type of slope is obviously connected to the intense lake level rise, manifested in the form of stacked progradational parasequence sets. The successions clearly demonstrate that high rate of sediment supply is overwhelmed by base-level rise in the short term.

Conclusions

The unique succession at Szulimán with turbidites and synsedimentary folds with possible minor modifications by storm-induced currents clearly indicates that an inclined paleotopography, meaning some sort of slope, must have existed within a few km of the location. The usually 20–30 m high delta slopes in Lake Pannon as well as the several hundred meter high basin margin slopes can be excluded. The Szulimán turbidites formed on the flank of a basement high, namely in a transitional position between "real" deep basin slopes and sublacustrine basement highs, where no slopes developed at all. However, the formation of these slopes of transitional height is only partly a result of their spatial location, rather it can be explained by an interplay of their marginal location and of lake level rise. As lake-level rise continued, the development of transitional slopes was repeated. As there are locations where the transgressive coastal sands are directly overlain by regressive deltaic successions, without intercalations of thick sublittoral/profundal clays, it is supposed that only a small temporal difference may have existed between the flooding of some elevated areas and the arrival of the prograding deltaic feeder system. The amplitude of lake-level rise seems to be unusually high with respect to similar events in Lake Pannon, but it is not possible to determine so far if the climatic or structural signal was stronger. As the turbidites near Szulimán are associated with alternating deltaic and open-water transitional slope deposits, they are assigned to the Újfalu Formation.

Reasons explaining why these turbidites have gone unrecognized so far include the poor outcrop conditions in low-relief landscapes surrounding the mountains and the lack of industrial exploration in these areas. Even if they exist, the resolution of industrial seismic profiles is usually inadequate to detect clinoforms of less than 100 m thickness, as is the quality of archive core or borehole documentations, where sedimentary structures are scarcely mentioned. It is a question yet to be examined whether similar sediments have a wider distribution among Lake Pannon "shallow-water" successions. A potential tool for their detection is the investigation of well logs and available cores, particularly from areas where extremely thick successions of Újfalu Sand have been reported. The hardly studied outcrops of piedmont areas also deserve attention. The presence of turbidites linked to transitional slopes can also be indicated by co-existing or mixed sublittoral and littoral faunas.

Last but not least, large-scale reconstructions of basin evolution or paleogeography can benefit from detailed process-based sedimentological studies and a combination of paleontological and limited well-log data even if only very small outcrops are present.

Acknowledgments: This research was supported by the European Union and the State of Hungary, co-financed by the European Social Fund in the framework of TÁMOP 4.2.4. A/2-11-1-2012-0001 'National Excellence Program' (Zoltán Magyary Grant to KS), and by the Hungarian Scientific Research Fund (OTKA) Projects PD104937 and K81530. We thank Zoltán Lantos (MFGI) and Lilla Tőkés (ELTE) for helping in visualization of well-log data. Two anonymous reviewers are acknowledged for thought-provoking comments and suggestions. MOL Hungarian Oil and Gas Company is thanked for permitting the publication of seismic data in Figure 2. This is MTA-MTM-ELTE Paleo contribution No. 215.

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